THE TASMANIAN ULTRAMAFIC-GABBRO
AND
OPHIOLITE COMPLEXES

by

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HOBART

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This thesis contains no material which has been accepted for the award of any other degree or diploma in any University and, to the best of my knowledge and belief, contains no copy or paraphrase of material previously published or written by another person, except where due reference is made in the text of the thesis.

M. Rubenach

M.J. RUBENACH
University of Tasmania
May, 1973.


Frontispiece - view looking north up the Heazlewood River valley from the Waratah-Savage River road.

ABSTRACT

A string of ophiolite and ultramafic gabbro complexes in western and northern Tasmania constitutes the Main Belt, while a second group in the Adamsfield area are composed entirely of ultramafics. These complexes are associated with Eocambrian and Cambrian sequences of mudstones, wackes, conglomerates and volcanic rocks lying between or flanking Proterozoic blocks. The Heazlewood River Complex, the largest and least dismembered in the Main Belt, is an ophiolite consisting of layered ultramafics intruded by gabbros and overlain by extrusive rocks. Dolerite dyke swarms intrude the ultramafics and part of the volcanic sequence, while tonalites, trondhjemites and granophyres intrude all other rock types. The rock contacts of the ultramafics with the Eocambrian Luina Beds are faulted and/or have formed by solid flow of serpentinite. Contacts between the extrusive rocks and the Luina Beds are poorly exposed, but on the basis of established faults and melange zones it is tentatively concluded that the ophiolite is an allochthon.

The Nineteen Mile Creek Dunites occurring along the NW margin of the Heazlewood River Complex are tectonites, some of which may be deformed cumulates. The rest of the ultramafics in the Heazlewood River Complex and the other Main Belt bodies are dominantly layered orthopyroxenites, harzburgites, lherzolites and dunites, many of which contain interstitial plagioclase. Considered together, the layering styles, textures, and compositional ranges (Fo79-89, En79-89) of these rocks are consistent with a cumulative origin. It is suggested that nucleation (as opposed to mechanical sorting processes) and postcumulus diffusion are important controls on the layering.

The layered ultramafics and gabbros of the Main Belt are believed to have formed by crystal accumulation from tholeiitic or high-Mg tholeiitic magmas. Lack of olivine-plagioclase reactions suggest crystallization occurred at pressures less than 6 kb.

The extrusive rocks of the Heazlewood River Complex are probably mainly basalts, but high-Mg basalts as well as some intermediate and acid types also occur. The dolerites, which have suffered less than the volcanics from the pervasive hydrothermal or burial metamorphism, are quartz tholeiites which grade into granophyres and tonalites. Although similar to Cainozoic mid-ocean ridge basalts in Cr and Ni contents and being relatively low in the incompatible elements, the dolerites and volcanic rocks are dominantly quartz tholeiites rather than olivine tholeiites and have unusually low TiO_2 contents. Nevertheless it is concluded that the Heazlewood River Complex (and perhaps the other Main Belt complexes) may have formed as Eocambrian lithosphere at a mid-ocean ridge, marginal sea, or similar spreading environment. Detritus of ophiolite rocks in Cambrian sedimentary rocks in a number of localities suggests that the Main Belt complexes were tectonically emplaced prior to the commencement of Middle Cambrian sedimentation. It is suggested that the lenses of foliated amphibolites which occur along the contacts of several complexes, may have formed by metamorphism of ophiolite rocks prior to, or in the early stages of the initial tectonic emplacement.

As a result of a series of faulting, serpentinization and tectonic re-emplacement events, several of the Tasmanian Complex now occur as serpentinite lenses strung out along major fault zones, and in contact with sedimentary rocks as young as Silurian.

The first serpentinization phases affecting the complexes preferentially replaced dunites and harzburgites by massive lizardite serpentinites, and altered gabbro layers and dykes to rodingites or amphibole-prehnite rocks. Characteristic massive and sheared varieties of green waxy-lustred serpentinites formed in later phases associated with deformation. These green serpentinites are lizardite-chrysotile mixtures, and in many localities contain cross-fibre asbestos veins which typically surround residual pyroxenite kernals.

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(Separate Volume)

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INTRODUCTION

Scope of study

Having been granted a Commonwealth Postgraduate Scholarship, the writer began work on "the petrology and chemistry of the ultramafic rocks of Tasmania" in June 1967. This work followed on from the study of the Serpentine Hill Complex as an honours B.Sc. project in 1966. It was decided that a broad approach would be taken, covering such aspects as the origin of the layered ultramafics and gabbros, the origin and tectonic significance of the ophiolite complexes, and the dismembering of the ultramafic-gabbro and ophiolite complexes by tectonic emplacement, faulting and serpentinization. The main reasons for such an approach were that no recent synthesis of the geology and tectonic significance of these rocks had been made, and that the various aspects are interwoven and cannot be satisfactorily studied in isolation.

Research was concentrated on the Heazlewood River Complex, the largest and least dismembered of the Tasmanian complexes. Of a total of 16 weeks of field work, 11 were spent mapping the Heazlewood River Complex while 2 weeks (in addition to 12 weeks in 1966) were spent on the Serpentine Hill Complex. Periods ranging from 1 to 7 days were spent on each of the other complexes. About 700 thin sections were examined, while 64 rocks were analysed for major and some trace elements by X-ray fluorescence, 21 pyroxenes were analysed by electron microprobe, and over 70 diffractograms of serpentinites and rodingites were examined.

The writer was also given the opportunity of working with Dr. R. Varne on geology and petrology of Macquarie Island, which we believe to be an uplifted slice of oceanic crust (Varne and Rubenach, 1972, 1973). About 15 weeks were spent mapping the island and additional time was spent on trace element analyses and petrographic studies. Besides providing invaluable experience, the Macquarie Island work considerably influenced the writer's ideas on ophiolites.

Location, access and field conditions

The distribution of the ultramafic-gabbro and ophiolite complexes is shown in Figure 2, while Figure 1 gives the location of the main towns and highways in Tasmania. With the exception of the Wilson River, Noddy Creek, Spero Bay and Boyes River bodies, the larger complexes are crossed by roads or highways. The Noddy Creek Complex is crossed by a number of tracks, which are generally passable only with tracked vehicles; these tracks are accessible from Asbestos Point or Birchs Inlet in Macquarie Harbour. The Spero Bay Complex could be reached by attempting a landing in Spero Bay. Most of the Wilson River Complex can be reached overland with considerable difficulty; the writer was flown by helicopter to field camps set up by Aberfoyle N.L., but these are now abandoned.

Mapping in Western and Southwestern Tasmania is inhibited by the high rainfall (80-150 inches per annum) and the dense forest or scrub covering much of these parts of the state. A large part of the Heazlewood River Complex covered by dense vegetation (tea-tree, Banksia, rain forest with thick undergrowth) would have been very difficult to map except for lines cut by Amax Mining Ltd and Comstaff Pty Ltd. Exposure of the ultramafics is generally fair, but the dolerites and basalts are usually poorly exposed. A frustrating feature of the field work was the lack of exposed contacts between major rock types, for field relationships are critical to any study of alpine ultramafic or ophiolite complexes.

Previous literature

The occurrence of asbestos at Andersons Creek was documented by 1820*, and briefly referred to in Strzelecki (1845) and Gould (1866). In 1876, osmiridium (incorrectly called palladium) was discovered in the Wilson River area, but the identity of this metal was established much later (Montgomery, 1894). An "osmiridium rush" began at the beginning of the 20th Century, and stream gravels and sands were extensively worked in the Nineteen Mile Creek-Savage River, Mt. Stewart, Wilson River, and Adamsfield areas. The reports of Twelvetrees (1914), Reid (1921) and Nye (1929) describe the osmiridium deposits. The report of Reid is particularly impressive with regard to accurate observations and petrographic descriptions concerning the ultramafics and origin of the osmiridium deposits. With regard to the asbestos deposits, the report of Taylor (1955) is significant.

The work of Green (1959) on the Andersons Creek Complex is the only systematic study of the geology and petrography of an ultramafic-gabbro complex to be published in more recent times.

* Historical Rec. Australia Series III, Vol.III, p.38. *Library Committee of the Commonwealth Parliament*, Sydney, 1921.

Specimens and thin sections

All specimens and thin sections referred to in the text are housed in the Geology Department, University of Tasmania, unless otherwise stated. A list of specimens and their localities is included as Appendix 2.

Acknowledgments

The writer is particularly indebted to Dr. R. Varne for his supervision and the many stimulating discussions concerning all facets of the research project. Thanks are also due to other members of staff and former colleagues at the Geology Department, University of Tasmania, in particular Professor S.W. Carey, Dr. M. Solomon, Mr. M. Banks, Mr. R. Ford, Dr. D. Groves, Dr. K. Corbett, Dr. C. Gee, and Mr. J. Griffiths. The assistance of Mr. R. Ford with respect to X-ray work is acknowledged. The electron microprobe analyses were done at the Department of Geophysics and Geochemistry, Australian National University, and Mr. N. Ware is thanked for his assistance. Messrs. C. Palethorpe and R. Close of Broken Hill Proprietary are thanked for providing unpublished analytical and other data.

The writer wishes to thank the following companies for their invaluable assistance in providing transport and accommodation at field camps during parts of the field investigations: Aberfoyle N.L., Amax Mining Pty Ltd, and Broken Hill Proprietary.

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CHAPTER 1

OUTLINE OF THE PROTEROZOIC AND LOWER PALAEOZOIC GEOLOGY OF TASMANIA

INTRODUCTION

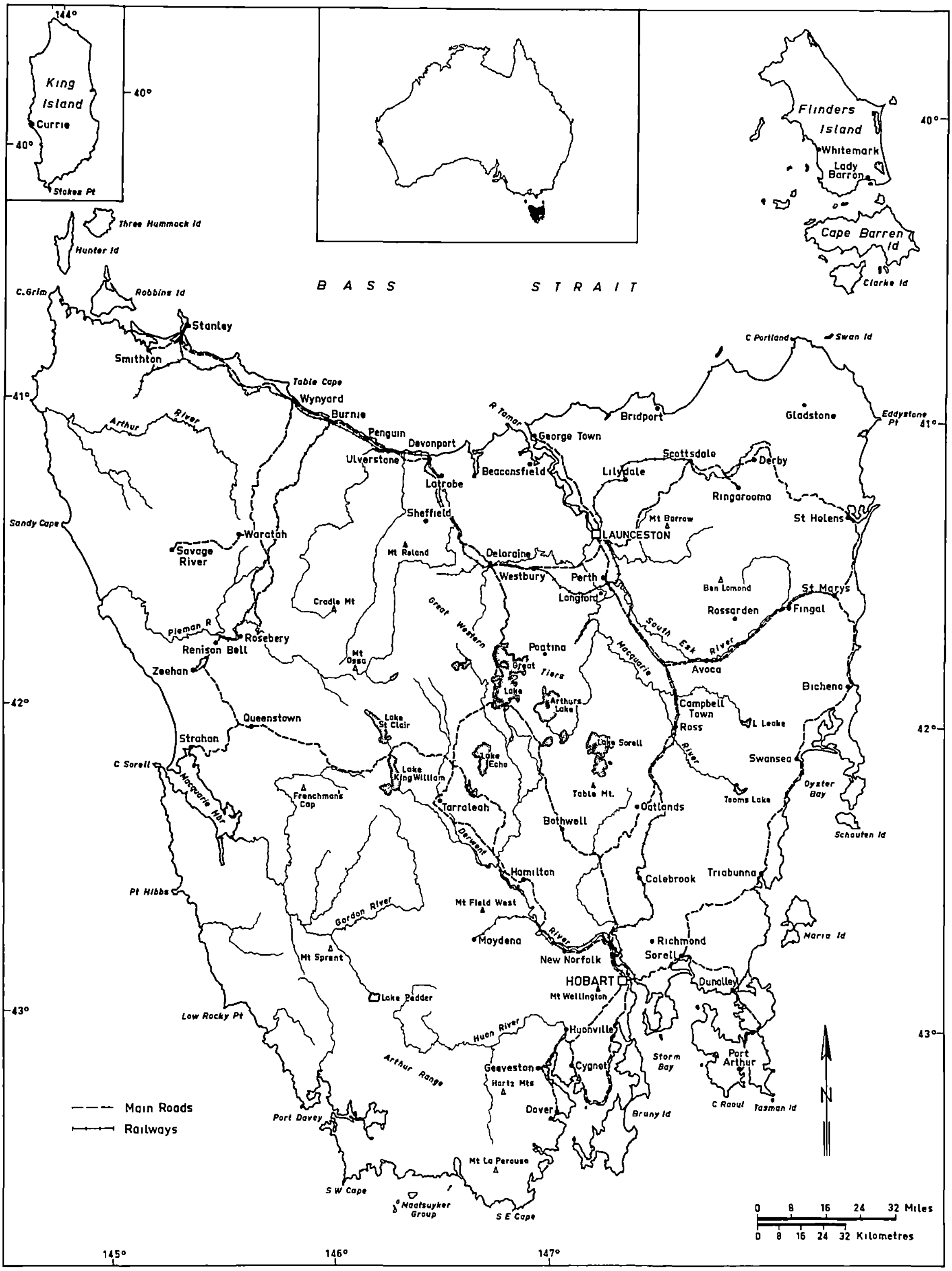
The purpose of this chapter is to set the geological framework within which the ultramafic-gabbro and ophiolite complexes can be described. More detailed summaries can be found in Spry and Banks (1962) and Solomon (1965), while a number of reconstructions according to the plate tectonics model are given in Solomon and Griffiths (1972) and Corbett *et al.* (1972).

THE PROTEROZOIC

The largest areas of Proterozoic rocks are referred to as the Rocky Cape Geanticline and the Tyennan Block, so-named because of their apparent influence on Lower Palaeozoic sedimentation and folding (Carey, 1953; Solomon, 1962).

The Proterozoic rocks can be subdivided into a metamorphosed group consisting of dominantly greenschist facies schists, quartzites, and phyllites, and an "unmetamorphosed" group of shales, quartzites and dolomites. Within the Tyennan Block these groups are separated by an unconformity and the so-called "Frenchman Orogeny" (Spry, 1962a). Metamorphosed rocks in the Rocky Cape Geanticline are confined to a narrow belt, the Arthur Lineament, which is convincingly argued by Gee (1967) to be a zone along which deformation was more intense than in the essentially contemporaneous rocks flanking it. The deformation phase in which the Arthur Lineament formed comprises at least part of the Penguin

Figure 1. Locality map of Tasmania


King Island
Currie
Stokes Pt
Flinders Island
Whitemark
Lady Barron
Cape Barren Id
Clarke Id
Three Hummock Id
Hunter Id
Robbins Id
C. Grim
BASS STRAIT
Stanley
Smithton
Table Cape
Wynyard
Burnie
Penguin
Devonport
Ulverstone
Latrobe
Sheffield
R Tamar
George Town
Beaconsfield
Bridport
Gladstone
Scottsdale
Derby
Lilydale
Ringarooma
C Portland
Swan Id
Eddystone Pt
St Helens
Mt Barrow
LAUNCESTON
Arthur River
Sandy Cape
Waratah
Savage River
Mt Roland
Deloraine
Westbury
Perth
Longford
Ben Lomond
St Marys
Fingal
Rossarden
Cradle Mt
Great Western Tiers
Pieman R
Rosebery
Renison Bell
Zeehan
Mt Ossa
Poatina
Great Lake
Arthurs Lake
South Esk River
Macquarie
Avoca
Bicheno
Queenstown
Strahan
Lake St Clair
Campbell Town
Ross
L Leake
Lake Sorell
C Sorell
Macquarie Hbr
Lake King William
Frenchmans Cap
Lake Echo
Swansea
Oyster Bay
Tarraleah
Table Mt.
Oatlands
Tooms Lake
Bothwell
Schouten Id
Derwent
Hamilton
Colebrook
Triabunna
Pt Hibbs
Gordon River
Mt Field West
Maria Id
Maydena
Richmond
Mt Sprent
New Norfolk
Sorell
HOBART
Dunalley
Lake Pedder
Mt Wellington
Low Rocky Pt
Huon River
Huonville
Arthur Range
Geeveston
Cygnet
Storm Bay
Port Arthur
Hartz Mts
Main Roads
Railways
Port Davey
Dover
Bruny Id
C Raoul
Tasman Id
N
Mt La Perouse
SW Cape
Maatsuyker Group
SE Cape
0 8 16 24 32 Miles
0 8 16 24 32 Kilometres
144°
40°
41°
42°
43°
145°
146°
147°

Movement (Gee, 1967), which was originally defined on an unconformity between Middle Cambrian sedimentary rocks and deformed Precambrian rocks near Ulverstone (Burns, 1964). K-Ar dating of a sample of syntectonic Cooee Dolerites as about 700 m.y. (Spry, 1962a), more recently confirmed with other samples (R. Coleman, *pers. comm.*), gives an Upper Proterozoic age to at least part of the Penguin Movement.

CAMBRIAN ROCKS

In the Cambrian, Tasmania formed part of the Tasman Orogenic Zone and "eugeosynclinal-type" sedimentation began. In many areas, the "Cambrian" sedimentary rocks can be subdivided into two groups, one of which contains fossils giving Middle or Upper Cambrian ages (Banks, 1962; Quilty, 1970; J. Jago, *pers. comm.*), and unfossiliferous rocks which in this thesis are labelled "Eocambrian" as their ages could be anywhere in the range from Upper Proterozoic to Lower Cambrian. In some areas where fossils have not been found, the sedimentary rocks can be tentatively placed in either group on their lithologies. A third important Cambrian element is the Mt. Read Volcanics, which flank the Tyennan Block to the west and north and which interdigitate with both Eocambrian and Middle to Upper Cambrian sequences.

1. Eocambrian rocks

The best known sequence of Eocambrian rocks is the Crimson Creek Formation, which occurs in the Dundas-Serpentine Hill area and extends NW as far as the Meredith Granite. This formation consists mainly of argillites and lithic wackes with some basic volcanics, and conformably overlies an inlier of unmetamorphosed Proterozoic rocks at Renison Bell

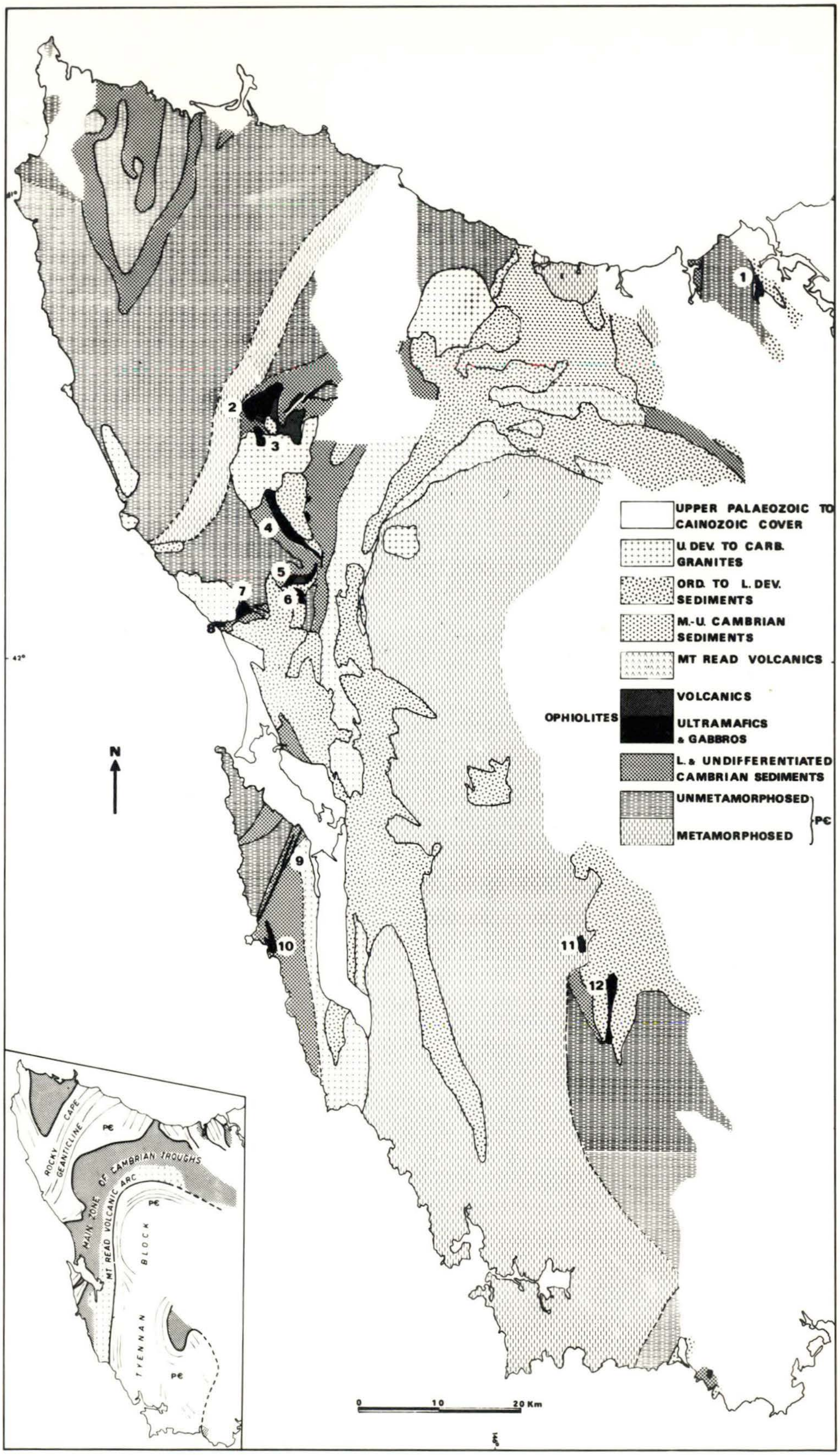
Figure 2. Geologic map of western Tasmania, showing distribution of the main ultramafic-gabbro and ophiolite bodies.

Main Belt complexes:

1. Andersons Creek Complex
2. Heazlewood River Complex
3. Mt. Stewart Serpentinite
4. Wilson River Complex
5. Serpentine Hill Complex
6. Dundas Serpentinite
7. McIvor Hill Gabbro
8. Trial Harbour Serpentinite
9. Noddy Creek Complex
10. Spero Bay Complex

Adamsfield Belt

11. Boyes River Complex
12. Adamsfield Complex


UPPER PALAEOZOIC TO CAINOZOIC COVER
U. DEV. TO CARB. GRANITES
ORD. TO L. DEV. SEDIMENTS
M.-U. CAMBRIAN SEDIMENTS
MT READ VOLCANICS
OPHIOLITES
VOLCANICS
ULTRAMAFICS & GABBROS
L. & UNDIFFERENTIATED CAMBRIAN SEDIMENTS
UNMETAMORPHOSED
METAMORPHOSED
PЄ
N
1
2
3
4
5
6
7
8
9
10
11
12
42°
0
10
20 Km
ROCKY CAPE GEANTICLINE
MAIN ZONE OF CAMBRIAN TROUGHS
MT READ VOLCANIC ARC
TYENNAN BLOCK

(Blissett, 1962). It is overlain by tectonic slices of the ultramafic-gabbro or ophiolite complexes, which are in turn overlain by Middle to Upper Cambrian sequences. To the north of the Meredith Granite, the Luina Beds, which disconformably or unconformably overlie an inlier at Waratah (Groves, 1968) are correlated on lithology with the Crimson Creek Formation. In the Adamsfield area, an unfossiliferous sequence of mudstones, turbidites, conglomerates, cherts and basic volcanics underlying fossiliferous Cambrian sediments (Corbett, 1970) can be placed in the Eocambrian group.

The Tasmanian ophiolite and ultramafic-gabbro complexes are associated with the Eocambrian sedimentary rocks, except where tectonically emplaced to higher levels. Other gabbros and basalts, which are abundant in the Eocambrian sequences in certain areas, differ petrographically and chemically from rocks in the ophiolites and ultramafic-gabbro complexes.

2. Middle to Upper Cambrian sedimentary rocks

The best known sequence of Middle to Upper Cambrian sedimentary rocks is the Dundas Group (Elliston, 1954; Banks, 1956, 1962; Blissett, 1962). Northeast of Zeehan, this unconformably overlies the Serpentine Hill Complex, and consists of mudstones, chert breccia conglomerates, paraconglomerates, wackes and acid volcanics. The Huskisson River Group to the north correlates very well with the Dundas Group (Banks, 1962; Blissett, 1962). Middle to Upper Cambrian rocks are apparently absent from the Heazlewood River-Waratah area, whereas in the northern part of the state the majority of the exposed Cambrian rocks probably belong to the Middle to Upper Cambrian group. In the Adamsfield area, there is a Middle or Upper Cambrian sequence of mudstones, dolomitic sandstones and conglomerates, and several Upper Cambrian shallow marine sequences (Corbett, 1970).

3. The Mt. Read Volcanics

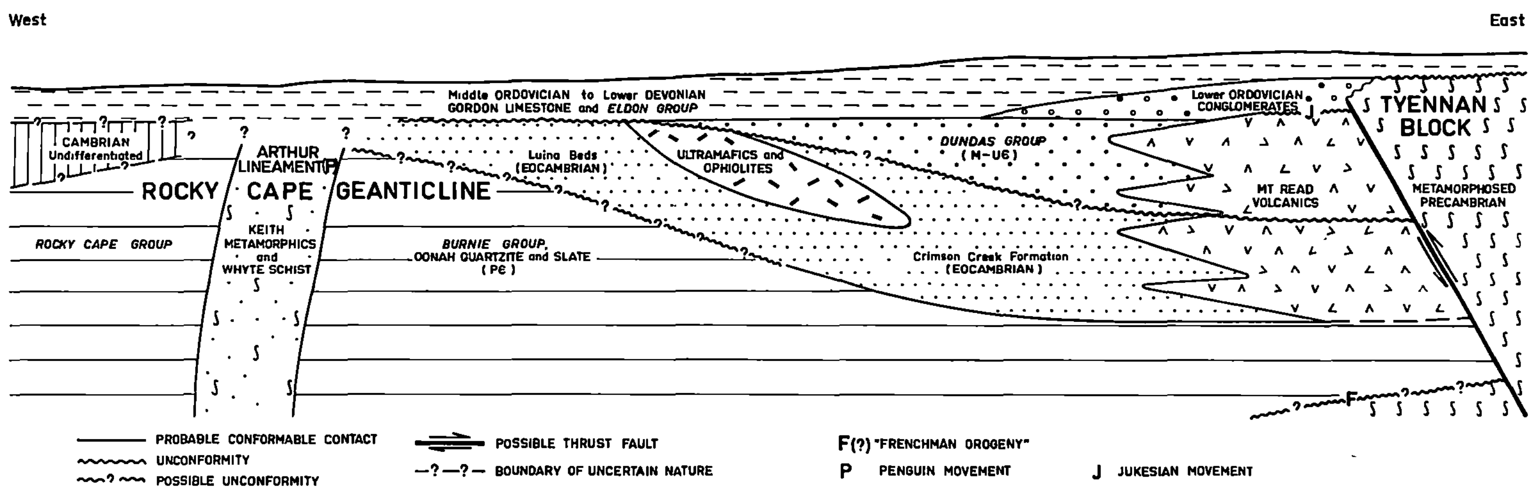
The Mt. Read Volcanics are pyroclastics and lavas, mainly of acid to intermediate compositions. They interfinger with the Eocambrian rocks near Rosebery (Brathwaite, 1969), and even more extensively with the Middle to Upper Cambrian sequences in the west and north of Tasmania. In the Queenstown area, it has now been established (Jago *et al.*, 1972) that the Mt. Read Volcanics can be subdivided into an older mineralized and more deformed sequence, unconformably overlain by a late Middle/Upper Cambrian sequence.

The Mt. Read Volcanics probably represent a Cambrian volcanic arc, which would imply the existence of a subduction zone according to the plate tectonics model. More recently, Anderson (1972) has shown that the chemistry of Mt. Read Volcanics is more consistent with Andean-type continental arcs than oceanic island arcs.

4. Cambrian tectonics

Previously, it was thought that Cambrian sedimentation took place in a series of troughs between or flanking the Precambrian blocks, and that the Middle Devonian Tabberabberan Orogeny was the main period of deformation occurring after the Proterozoic (Solomon, 1962). However, the presence of unconformities at the base of the Dundas Group, within

Figure 3. Diagrammatic section through the Proterozoic and Lower Palaeozoic of western Tasmania.


West
East
Middle ORDOVICIAN to Lower DEVONIAN
GORDON LIMESTONE and ELDON GROUP
Lower ORDOVICIAN
CONGLOMERATES
TYENNAN
BLOCK
CAMBRIAN
Undifferentiated
ARTHUR
LINEAMENT
ROCKY CAPE GEANTICLINE
Luina Beds
(EOCAMBRIAN)
ULTRAMAFICS and
OPHIOLITES
DUNDAS GROUP
(M-U6)
MT READ
VOLCANICS
METAMORPHOSED
PRECAMBRIAN
ROCKY CAPE GROUP
KEITH
METAMORPHICS
and
WHYTE SCHIST
BURNIE GROUP,
OONAH QUARTZITE and SLATE
(P€)
Crimson Creek Formation
(EOCAMBRIAN)
PROBABLE CONFORMABLE CONTACT
UNCONFORMITY
POSSIBLE UNCONFORMITY
POSSIBLE THRUST FAULT
BOUNDARY OF UNCERTAIN NATURE
F(?) "FRENCHMAN OROGENY"
P PENGUIN MOVEMENT
J JUKESIAN MOVEMENT

the Mt. Read Volcanics, and in the Cambrian sequences in the Adamsfield area, together with abundant detritus from the ultramafic-gabbro and ophiolite complexes in Cambrian sediments in a number of localities, shows that the Cambrian was indeed a tectonically active period, at least with prespect to faulting or thrusting. However whether there was a period of folding in the Cambrian has not been established with certainty (see Chapter 2).

On a broader scale, the ophiolites and the Mt. Read Volcanics provide the keys for the interpretation of the Cambrian geology in terms of plate tectonics. In later chapters an attempt is made to synthesise the formation of the ophiolites and Mt. Read Volcanics, together with the sedimentation, unconformities, periods of deformation etc., according to the plate tectonics model.

5. Ordovician to Devonian sedimentation

At the end of the Cambrian, the "eugeosynclinal phase" in the development of the orogenic zone had ceased. Uplift of the Tyennan Block resulted in thick wedges of Lower Ordovician fanglomerates forming on top of the Cambrian rocks bordering the block (Corbett, 1970). Following a marine transgression, the Gordon Limestone (Lower to Upper Ordovician) was deposited. In the Silurian and Lower Devonian, deposition of the Eldon Group (mainly sandstones and mudstones) occurred in the western half of Tasmania, which by this time had stabilized as part of a large shelf.

6. Mid-Devonian deformation

Sedimentation had ceased by the Middle Devonian when the main deformation phase of the Tabberabberan "Orogeny" set in. The Tabberabberan deformation is essentially of the paratectonic style in the sense of Dewey (1967), as there is no regional metamorphism and the folds are generally broad and upright. The structural trends were controlled to a fair degree by the Precambrian blocks (Carey, 1953; Solomon, 1962).

7. Upper Devonian to Carboniferous granites

A number of high level granitic plutons were emplaced in the Upper Devonian and Lower Carboniferous following the cessation of the Tabberabberan folding.

CHAPTER 2

GEOLOGY OF THE OPHIOLITE AND ULTRAMAFIC-GABBRO COMPLEXES

INTRODUCTION AND DEFINITIONS

In this chapter, the geology of each of the Tasmanian ophiolite and ultramafic-gabbro complexes is briefly described. Conclusions regarding such problems as the origin of the layering and solid intrusion of serpentinite, which are discussed in detail in the following chapters, are at this stage presupposed.

The term "ophiolite", originally defined by Steinmann (1927), is used in a number of different ways in the literature. The present writer largely concurs with the definitions agreed to by those present at the Geological Society of America Penrose Conference on Ophiolites in 1972; essentially, an ophiolite is a complex in which ultramafics and gabbros are overlain by mafic sheeted dykes, which in turn are overlain by mafic volcanics. That a large proportion of the ultramafics necessarily have to be tectonites is not considered essential by the present writer. When only parts of the original sequence are represented, the term "dismembered ophiolite" can be used. If, however, both the sheeted dykes and volcanics are missing, the terms "ultramafic gabbro complex" or simply "ultramafic complex" are preferred, since there is no way of showing that the plutonic rocks were originally part of an ophiolite. Thus, in this thesis, the Heazelwood River Complex is referred to as an ophiolite, the Serpentine Hill Complex as a dismembered ophiolite, and the others, which originally may or may not have formed parts of ophiolites, are called ultramafic-gabbro or ultramafic complexes.

THE TASMANIAN "ULTRAMAFIC BELTS"

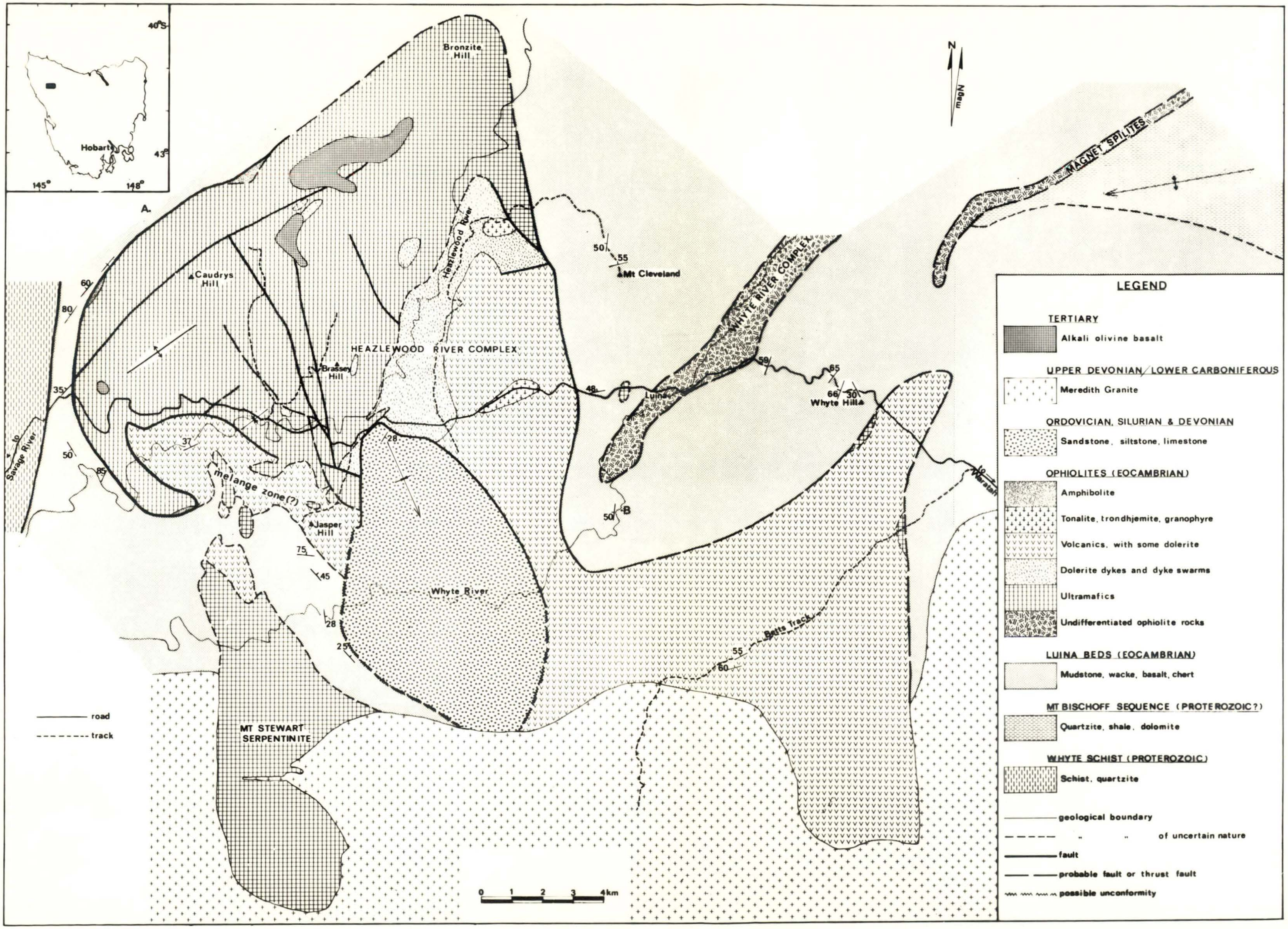
A string of ultramafic-gabbro and ophiolite complexes occurs in western Tasmania, from the Spero Bay Complex to the Heazlewood River Complex (Figure 2). The Andersons Creek Complex in northern Tasmania, is very similar to the western complexes, and together they comprise the Main Belt. In contrast, the complexes at Adamsfield, Boyes River, and perhaps at Rocky Boat Harbour in the far south, are ultramafic bodies and form a separate belt (the "Adamsfield Belt") flanking the eastern side of the Tyennan Block. Longman and Leaman (1971) postulate a series of subsurface ultramafic bodies to explain gravity highs flanking the north-east margin of the Tyennan Block. Assuming these exist, they could belong to either the Main Belt or the Adamsfield Belt.

THE HEAZLEWOOD RIVER AREA

The geology of the Heazlewood River-Waratah area is shown in Figure 4. Access to the area is provided primarily by the Waratah-Savage River road. The road to the summit of Mt. Cleveland, and with difficulty the tracks to Jasper Hill, the Lord Brassey Mine, and along Roaring Meg Creek (see Fig. 11), are accessible to four-wheel drive vehicles. Parts of Betts Track are at present open to four-wheel drive vehicles, while the tracks to Mt. Stewart, the Godkin area and the Magnet track are accessible only by foot (see Groves, 1965; and Jack and Groves, 1964, for details). Except for the relatively open area underlain by pyroxenites, the ultramafics are generally covered by dense scrub, the volcanics by eucalypt forests with undergrowth of varying density, and the sedimentary

Figure 4. Geology of the Heazlewood River - Whyte River area. The western part of this area was mapped in detail by the writer, and is included herein as figure 11. The geology of the rest of the area is compiled from Nye (1923), Groves (1968), Cox (1969) and Dr. Ransom (pers. comm.), together with several traverses and reconnaissance mapping by the present writer.

The base map was prepared using the provisional Savage River - Waratah 4 in. = 1 mile and Pieman 4 in. = 1 mile topographic maps of the Tasmanian Lands and Surveys Department, together with Balfor - Valentine air photos.



40°S
Hobart
43°
145°
148°
A.
Bronzite Hill
Caudrys Hill
Brassey Hill
Jasper Hill
HEAZLEWOOD RIVER COMPLEX
Heazlewood River
Mt Cleveland
WHYTE RIVER COMPLEX
MAGNET SPILITES
Luina
Whyte Hill
Whyte River
melange zone(?)
MT STEWART SERPENTINITE
Savage River
Waratah
Betts Track
B
N
magN
road
track
0 1 2 3 4km
LEGEND
TERTIARY
Alkali olivine basalt
UPPER DEVONIAN/LOWER CARBONIFEROUS
Meredith Granite
ORDOVICIAN, SILURIAN & DEVONIAN
Sandstone, siltstone, limestone
OPHIOLITES (EOCAMBRIAN)
Amphibolite
Tonalite, trondhjemite, granophyre
Volcanics, with some dolerite
Dolerite dykes and dyke swarms
Ultramafics
Undifferentiated ophiolite rocks
LUINA BEDS (EOCAMBRIAN)
Mudstone, wacke, basalt, chert
MT BISCHOFF SEQUENCE (PROTEROZOIC?)
Quartzite, shale, dolomite
WHYTE SCHIST (PROTEROZOIC)
Schist, quartzite
geological boundary
" " of uncertain nature
fault
probable fault or thrust fault
possible unconformity

rocks by rain forest in which the undergrowth is commonly quite dense. Lines cut by mining companies covering part of the ultramafics and the volcanics in the Jasper Hill area, together with logging tracks in the Jasper Hill area and covering parts of the dolerites, were of considerable help during the mapping program. Many of these lines and tracks are probably now overgrown.

1. Stratigraphy

The stratigraphy of the area is described in Nye (1923) and Groves (1968). The geology of parts of the area is discussed in Groves and Solomon (1964), Groves (1966), Cox (1968) and Cox and Glasson (1971).

In the western part of the area, Proterozoic schists, phyllites and quartzites are faulted against the Eocambrian sediments. These rocks are called the Whyte Schist and form part of the Arthur Lineament (Gee, 1967). To the east, unmetamorphosed Proterozoic sandstones, shales and dolomites are exposed in an ENE-WSW anticlinorium (Groves and Solomon, 1964; Groves, 1968).

The Luina Beds are a monotonous series of wackes, mudstones, basic volcanics and cherts. Except for some localized horizons of cherts there are no marker horizons. The Arthur River sequence of Groves (1968), which disconformably or unconformably overlies Proterozoic rocks at Waratah, forms part of the Luina Beds. Local subdivisions by Cox (1968) and Cox and Glasson (1971) have little regional significance as the present writer has formed similar lithologies interbedded with each other throughout the entire area.

The writer agrees with Groves and Solomon (1964) that the Luina Beds are similar to the Eocambrian Crimson Creek Formation and unlike the Middle to Upper Cambrian sequences.

A NW-trending synclinal structure of orthoquartzites, siltstones, and some limestone lenses is in part faulted against and in part unconformably overlying the Heazlewood River Complex and Luina Beds. Ordovician fossils occur in the limestone lenses (Groves, 1966), while the other rock types are similar to the Eldon Group (Silurian to Lower Devonian) found elsewhere in western Tasmania.

Flows of alkali olivine basalts of Tertiary age cap many of the hills throughout the area. They are probably relicts of extensive flows which have now been largely removed by erosion. Thin poorly consolidated lacustrine and fluviatile sediments of Tertiary age underlie the basalts in some localities; they are briefly described in Reid (1920), Nye (1923) and Groves and Solomon (1964).

2. The Meredith Granite

The Meredith Granite is one of the largest granitic plutons in western Tasmania. It is described in Groves (1968) and Stockley (1972), and is dated as Late Devonian-Early Carboniferous by Brooks (1966).

3. Structure

Deformation of the Whyte Schist was achieved during the Penguin Movement. Groves (1968) tentatively concludes that his F1 folds present in the Proterozoic rocks at Waratah, but not in the adjacent Eocambrian rocks, belong to the Penguin Movement, while F2 folds present in both groups represent Tabberabberan deformation. The Proterozoic and Eocambrian rocks at Waratah would probably be separated by a disconformity or unconformity.

Further to the west, the Luina Beds are noted for their lack of mesoscopic folds. A cleavage is sporadically developed in the mudstones, but is refracted near contacts with wackes. In the Cleveland Mine area, Cox and Glasson (1971) have deduced that the dominant folds are tight NNE to NE trending folds with axial surfaces dipping steeply NW and with easterly limbs overturned to the east. Superimposed on these are mesoscopic folds of variable orientation but with a dominant NE trend. The last phase of folding affecting the Luina Beds in this area are open folds plunging steeply NW. All this folding is assigned to the Tabberabberan by Cox and Glasson (1971). Although the field observations of the writer are consistant with this structural picture, the regional geology and structure are at present under review by geologists from Cleveland Tin Mining Company N.L. In this thesis, then, the structure of the Luina Beds as outlined in Cox and Glasson (1971) is tentatively accepted.

In contrast to the dominant NE to NNE trending folds in the Luina Beds, the Ordovician to Devonian sediments form a more open NW trending syncline. Since this fold is almost certainly Tabberabberan, the question arises as to whether the dominant folds in the Luina Beds are also Tabberabberan or represent a period of Cambrian deformation. The apparent lack of tight NE folds in the Ordovician-Devonian sedimentary rocks does not necessarily imply that such folds in the Luina Beds are Cambrian; the two groups of rocks are quite different and so the situation could simply be an example of disharmonic folding, all the deformation taking place in the Tabberabberan.

4. <u>The Heazlewood River Complex</u>

The Heazlewood River Complex is composed of layered ultramafics, dolerites and volcanics with some gabbros, tonalites, trondhjemites and granophyres (Fig. 4). Bodies such as the Mt. Stewart Serpentinite, the Whyte River Complex and the Magnet Spilites are not included in the definition, even though they, together with the Heazlewood River Complex, may have been derived by dismembering of a larger ophiolite. Likewise, the numerous isolated bodies of volcanics occurring in the Luina Group are not included in the definition of the Complex; these differ from volcanics within the ophiolites in having significantly higher iron oxides and TiO_2 contents.

The ultramafics of the Heazlewood River Complex occupy about 64 km^2, while the dolerites and volcanics outcrop over an area of about 66 km^2. The depth to which these rocks extend could not be predicted from the surface geology. However, a gravity survey along the Waratah-Savage River road was interpreted by both Johnson (1972) and the present writer. These interpretations place at least some limitations on the depth of the complex. The interpretations (see Appendix 3) are limited by the fact that a two-dimensional model was used, and more particularly by the fact that variations in the degree of serpentinization can produce rock densities over a range of 2.5 to 3.3 gm/cm^2, thus giving rise to an unlimited number of possible solutions. However, relatively unserpentinized ultramafics (i.e. the Caudrys Hill Pyroxenites) must extend for at least 3-4 km below the surface in order to explain the 38 mgal anomaly. The model given in Appendix Figure 31 contains a block of volcanics and dolerites 5 km thick, but the real thickness could be much greater or less depending on whether ultramafics underlie these rocks and whether or not they are serpentinized. So except for there being at least 3-4 km of relatively unserpentinized ultramafics beneath part of the Heazlewood River Complex, very little else can be said of its subsurface shape.

4.1 Ultramafics and gabbros

The ultramafics forming part of the Heazlewood River Complex are dominantly layered orthopyroxenites, harzburgites and dunites. Over much of the area they are plagioclase bearing. Where they have suffered serpentinization, the composition and textures of the primary rock type can generally be recognized. The description of the ultramafics, together with gabbroic rocks intruding them, is left to the following chapter.

4.2 Dolerites

Tabular dykes of dolerites and microgabbros intrude the eastern part of the ultramafics, and are particularly abundant in road cuttings in the Duffs Hill area (see Fig. 11). The dykes range from 2 cm to several metres in thickness, have trends ranging from NW to NE, and in general dip to the east. Locally they are fairly closely spaced forming sheeted dyke swarms as in Plate 1. Larger patches of dolerite occurring within the ultramafics, such as the one on Burgess Hill, are poorly exposed and it is not known whether such bodies are single or composite intrusions.

Dolerites underlying an extensive area flanking the ultramafics on the eastern side are poorly exposed, and were mapped mainly from surface float. The contact with the ultramafics was observed only in a cutting on the main road, where the dolerite appears to intrude the ultramafics. Several patches of ultramafics and amygdaloidal rocks which are almost certainly volcanics have been found enclosed in the dolerites, which are interpreted as a dyke swarm intrusive into both the ultramafics and parts of the volcanics.

Many of the patches of more coarsely textured basic rocks scattered throughout the volcanics are almost certainly dolerite dykes, but exposure is generally too poor to establish field relationships.

4.3 The volcanics

These are probably largely of basaltic compositions, but silicification accompanying some of the metamorphism of these rocks has made it difficult to judge their compositions. Some "andesitic" and "dacitic" types do occur. Of interest are the high-Mg basalts which occur to the SE of Jasper Hill and in a number of localities in the volcanics SW of Mt. Cleveland (particularly in a road cutting at 34190E/89330N). These rocks contain abundant euhedral pyroxenes pseudomorphed by chlorite or tremolite-actinolite.

Exposure of the volcanics is very poor, and most mapping was done using surface float. As a consequence, very little structural data could be obtained. In general, the rocks were called volcanics in the field if they were fine-grained, and in particular if they were obviously fragmental or contained abundant amygdales. Pillow lavas were not observed in outcrop, although a number of pillows or parts of pillows were found as surface float. Most of the lavas would appear to be massive.

Characteristic agglomerates interbedded with tuffs form prominent outcrops at 33540E/89260N, NE of Jasper Hill, and in several localities along Betts Track close to the Meredith Granite. The blocks in these agglomerates are characteristically equidimensional and subangular to rounded (Plate 2).

Thin beds or small lenses of clastic sedimentary rocks occur interbedded with the volcanics in a number of localities. These include cherts, jaspers, mudstones and micaceous wackes. It is concluded that the volcanics forming part of the Heazlewood River Complex are probably subaqueous. The lack of pillow lavas would be unusual in comparison with other ophiolites, but the writer has observed thick sequences of massive lavas, in part interbedded with pillow lavas, on Macquarie Island where they certainly formed under marine conditions (Varne and Rubenach, 1972).

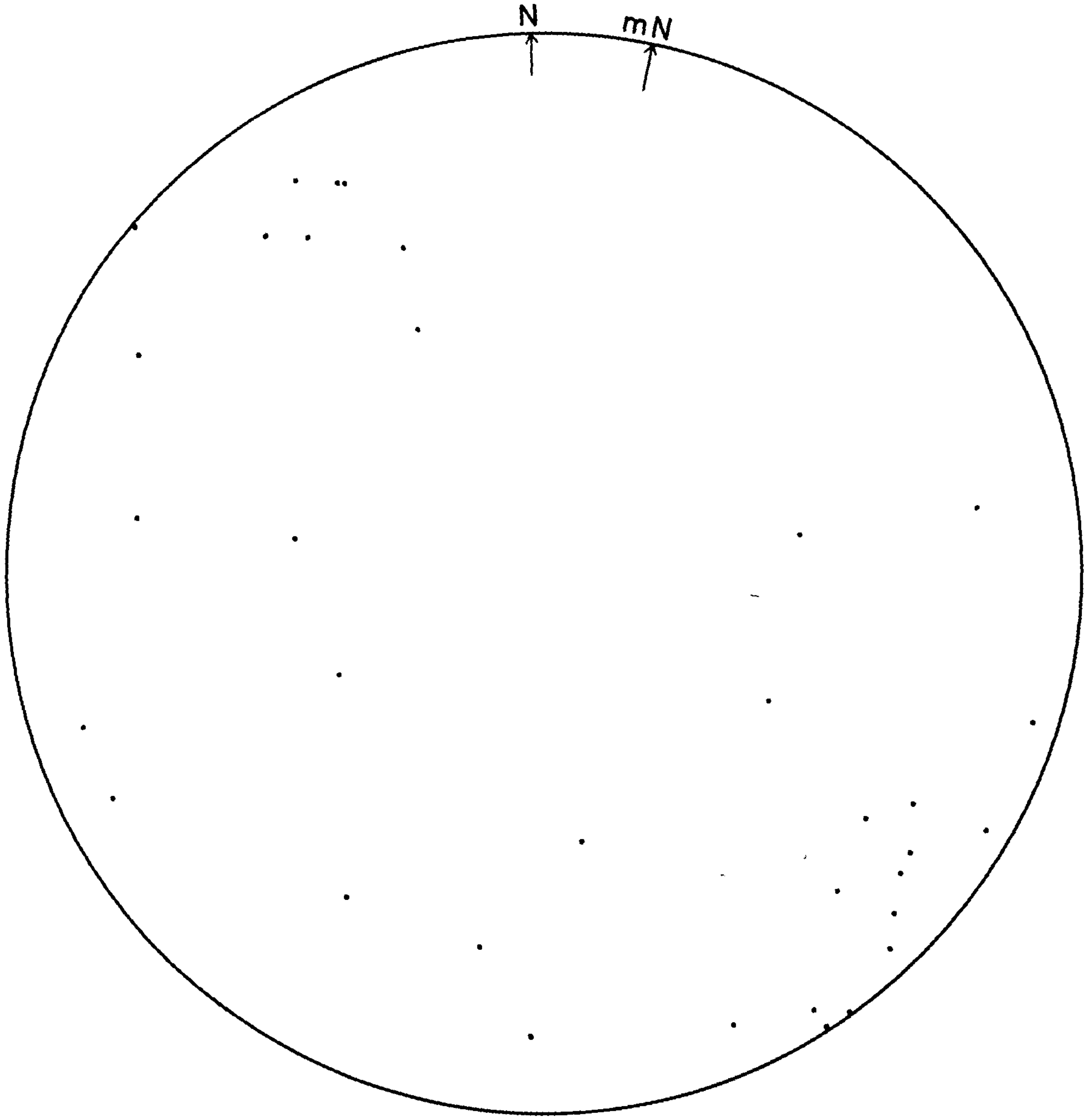
4.4 Tonalites, trondhjemites and granophyres

These rocks occur as thin tabular dykes or as very small stocks intruding ultramafics, dolerites and volcanics. Some are described in Nye (1923) and Groves (1966), who call then "syenites". They are generally poorly exposed.

4.5 Structure and contact relationships

The effect of folding on the rocks of the Heazlewood River Complex has not been established with any certainty. No mesoscopic folding of the layering has been observed, and the complexity of variations in interlayered rock types together with the faulting has made the characterization of larger folds extremely difficult. A plot of 34 layering poles in Figure 5 shows a scattering which may be in part a reflection of the fact that a lot of measurements were taken near faults. The data in Figure 5 are not inconsistent with upright folds of NE to NNE trends (i.e. parallel

Figure 5. Poles to layering in the ultramafics of the Heazlewood River Complex, plotted on a Wulff net.


N
mN

to the dominant trends in the Luina Beds), but cannot be used to demonstrate such folding. Two tentative folds in the Caudrys Hill Pyroxenites are shown in Figures 4 and 11. Because of their poor exposure, data obtained from the volcanics was insufficient for any structural interpretations.

Faults in the Heazlewood River area are shown in Figures 4 and 11. NNW and NW faults are dominant. Many small sheared serpentinite zones occurring in the ultramafics may be a reflection of faulting, but are not included on the maps.

Contacts between the ultramafics and the Luina Beds are exposed in a number of localities along Nineteen Mile Creek, in a quarry just north of the road in the extreme west of the complex, and to the SE of the latter locality. In all cases, the contact consists of highly sheared serpentinite against sedimentary rocks. The sedimentary rocks adjacent to the contact show no thermal metamorphism, but those immediately in contact with serpentinite are deformed. Specimen 38795 is a sericite schist formed from a siltstone, and has a crenulation cleavage. Specimen 38799, which was also collected immediately adjacent a serpentinite contact, is a deformed basalt in which anastomosing cleavages wrap around relict fragments of the original rock. Specimen 38798 is a deformed "conglomerate" occurring at a contact: this rock may be tectonic in origin, produced by deformation of interbedded mudstones and fine sandstones. The lack of magmatic thermal aureoles and sheared nature of the rock types at the contacts show that the ultramafics did not crystallize *in situ* but rather the contacts are a result of later faulting and/or solid flow of serpentinite.

Three amphibolite bodies occurring at contacts are thought to be the result of metasomatism accompanying deformation and serpentinization, and are discussed in a later chapter.

Since the contact areas are faulted or not exposed, the writer could not establish with certainty the original structural relationships between the ultramafics and the volcanics within the Heazlewood River Complex. It can be argued, however, that since the volcanics are compositionally very similar to dolerites intruding the ultramafics (Chapter 6), that at least some of the volcanics formed later than, and originally overlay, the ultramafics. The picture of dolerite dyke swarms intruding the ultramafics and continuously feeding a growing pile of lavas overlying the ultramafics is quite consistant with the observed relationships between ultramafics, dolerites and volcanics to the east and northwest of Brassey Hill. Such a model has been proposed for other ophiolite complexes (Gass, 1967; Moores and Vine, 1970), and for the Macquarie Island rocks (Varne and Rubenach, 1972).

South of the main body of ultramafics, the relationships between ultramafics, volcanics, dolerites and sedimentary rocks are most confusing. Outcrop is poor and contacts are not exposed. In this area it appears that blocks of cumulate ultramafics (the Caudrys Hill Pyroxenites) and Luina Beds sedimentary rocks are randomly mixed up with the volcanics. The individual blocks range in size from metres to hundreds of metres. It is postulated that area represents a tectonic melange in which blocks of the various rock types were jumbled together during complex thrusting or gravity sliding.

Southwest of Mt. Cleveland, the volcanics of the Heazlewood River Complex are probably faulted against the Luina Beds. Recent mapping by Cominco Exploration Pty Ltd has established that in the area SE of Luina marked as volcanics in Figure 4, a number of roughly conformable sheets composed of ultramafics, dolerites and volcanics are interlayered with the Luina Beds. These sheets are thought to be a series of thrust slices rather than have formed *in situ* during deposition of the Luina Beds (Dr. D. Ransom, *pers. comm.*).

In no place has a normal sedimentary contact between the Luina Beds and the volcanics of the Heazlewood River Complex been observed; either the contacts are thought to be tectonic or exposure so poor that any interpretation of the nature of the contact can be considered possible. Unfortunately, the nature of these contacts is critical with regard to the origin of the ophiolite; is the Heazlewood River Complex allochthonous, as is considered likely for most ophiolites (Coleman, 1971), or is there a possibility that it could have formed *in situ* during deposition of the Luina Beds? Both possibilities cannot be excluded from the field evidence, but the writer tentatively concludes that the Heazlewood River Complex, together with the associated bodies such as the Whyte River Complex, the Magnet Dyke, and the Mt. Stewart Serpentinite, are more likely to be tectonic slices or allochthons.

THE MT. STEWART SERPENTINITE, WHYTE RIVER COMPLEX AND MAGNET SPILITES

These bodies occur in the Heazlewood River-Waratah area. They are possibly fault blocks or slices formed by dismembering of a large ophiolite of which the present Heazlewood River Complex is also a part.

The writer did not map these in any detail, and it is not known for certain whether the volcanics and dolerites comprising large parts of the Whyte River Complex and Magnet Spilites are related to those within the Heazlewood River Complex.

1. The Mt. Stewart Serpentinite

This is accessible via an old pack track which can be picked up at Jasper Hill. The body, which is described in Reid (1921), is composed largely of serpentinized dunites containing some orthopyroxenite layers, and is similar to the Nineteen Mile Creek dunites in the Heazlewood River Complex. Thermal metamorphism by the Meredith Granite has affected the serpentinites which now largely consist of felted serpentine minerals and tremolite-actinolite with no trace of the original igneous textures (for example, specimen 38684).

2. The Whyte River Complex

The rocks of this body are briefly described in Nye (1923), Cox (1968), Groves and Solomon (1964) and Groves (1965). Cox (1968) describes it as a sheet-like body broadly conformable with the Luina Beds. The ultramafics are generally serpentinized and sheared, and contain lenses of cherts as well as bodies of dolerite and gabbro which may be tectonic inclusions. Specimens of relatively unaltered ultramafics collected by the present writer from a quarry in Luina are very similar to the medium-grained plagioclase pyroxenites in the Heazlewood River Complex. The dolerites may consist of sheeted dykes as is probably the case for the Heazlewood River Complex.

3. The Magnet Spilites

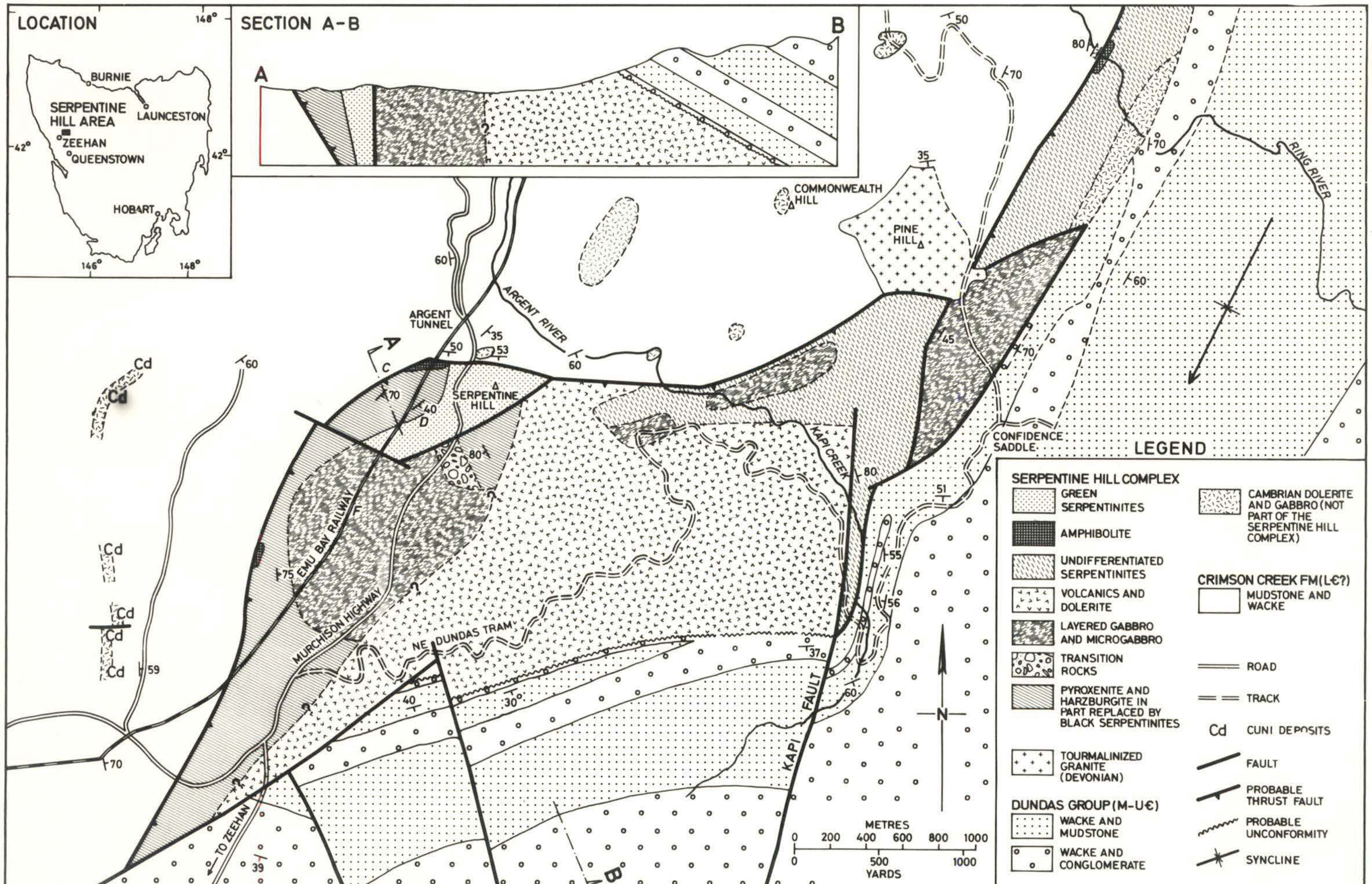
This body is described in Twelvetrees (1900), Nye (1923), Scott (1954) and Groves and Solomon (1964) who refer to it as the "Magnet Dyke". It is largely composed of altered basaltic rocks which originally may have been extrusive, intrusive or both. Several lenses of websterite occur within the body, and in places these have unusual orbicular textures (Twelvetrees, 1900). The rocks are all very poorly exposed, and no specimens of the websterite were found by the present writer. Blocks of Proterozoic rocks are included in the body. The writer believes that the Magnet Spilites more probably represent a fault slice rather than having formed by intrusion and extrusion *in situ*.

THE SERPENTINE HILL COMPLEX

The geology of the Serpentine Hill Complex is described in detail in a paper submitted for publication, included herein as an appendix. A summary is given below.

The Serpentine Hill Complex is an ophiolite which has been dismembered during tectonic emplacement and later faulting. It is essentially a sheet-like body, in the main lying between the Eocambrian Crimson Creek Formation and the Middle to Upper Cambrian Dundas Group and tapering to the NE within the Crimson Creek Formation (see Figs. 6 and 7). Clasts from both the volcanic and plutonic rocks occur in basal conglomerates of the Dundas Group, which unconformably overlies the complex. It is suggested, then, that the complex was emplaced as a thrust slice prior to Middle Cambrian sedimentation.

Figure 6. Geology of the Serpentine Hill area.



LOCATION
BURNIE
LAUNCESTON
SERPENTINE HILL AREA
ZEEHAN
QUEENSTOWN
HOBART
SECTION A-B
A
B
ARGENT TUNNEL
ARGENT RIVER
SERPENTINE HILL
COMMONWEALTH HILL
PINE HILL
RING RIVER
CONFIDENCE SADDLE
KAPI CREEK
KAPI FAULT
EMU BAY RAILWAY
MURCHISON HIGHWAY
NE DUNDAS TRAM
TO ZEEHAN
Cd
LEGEND
SERPENTINE HILL COMPLEX
GREEN SERPENTINITES
AMPHIBOLITE
UNDIFFERENTIATED SERPENTINITES
VOLCANICS AND DOLERITE
LAYERED GABBRO AND MICROGABBRO
TRANSITION ROCKS
PYROXENITE AND HARZBURGITE IN PART REPLACED BY BLACK SERPENTINITES
TOURMALINIZED GRANITE (DEVONIAN)
DUNDAS GROUP (M-U€)
WACKE AND MUDSTONE
WACKE AND CONGLOMERATE
CAMBRIAN DOLERITE AND GABBRO (NOT PART OF THE SERPENTINE HILL COMPLEX)
CRIMSON CREEK FM (L€?)
MUDSTONE AND WACKE
ROAD
TRACK
CUNI DEPOSITS
FAULT
PROBABLE THRUST FAULT
PROBABLE UNCONFORMITY
SYNCLINE
METRES
YARDS
N

The important features of the geology of the Serpentine Hill Complex are summarized below:

1) The formation of the ophiolite and its subsequent tectonic emplacement occurred before the Middle Cambrian.

2) The layered ultramafics grade through a transition zone into layered hypersthene gabbros. Layered gabbroic rocks (i.e. with more than 20% plagioclase) are quite rare in the Heazlewood River Complex.

3) The relationships between a number of types of serpentinites were established in this complex. Massive, relatively undeformed black or greenish-black serpentinites were the earliest to form, preferentially replacing the harzburgites and dunites. These serpentinites are composed essentially of "lizardite" and probably formed prior to tectonic emplacement. Highly deformed chrysotile-antigorite serpentinites occurring only at contacts with sedimentary rocks may have formed during tectonic emplacement or later faulting. Characteristic green serpentinites ("lizardite-chrysotile" composites) with a somewhat waxy lustre are associated with localized deformation which probably occurred after tectonic emplacement in the Cambrian. They cut across the primary layering and the other serpentine types, and range from massive to highly sheared varieties. Veins of cross-fibre asbestos are associated with these serpentinites, particularly as shells concentrically surrounding relict kernals of pyroxenites. Counterparts of the black "lizardite" serpentinites and the sheared contact serpentinites occur in most of the complexes, while green serpentinites very similar to those in the Serpentine Hill Complex occur in the Wilson River Complex, the Dundas Serpentinite, and the Noddy Creek Complex.

THE DUNDAS SERPENTINITE

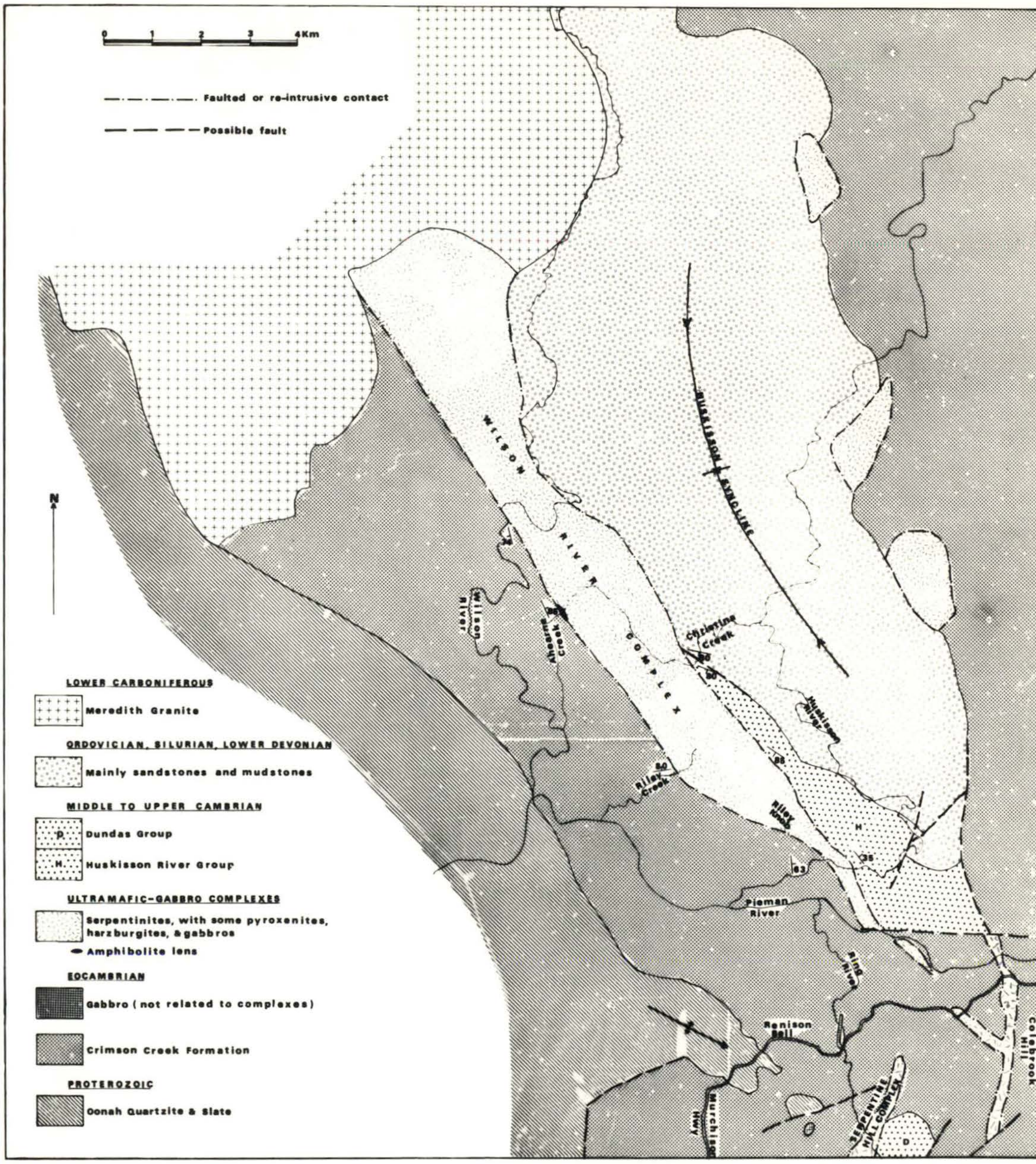
This body occurs just south of the Serpentine Hill Complex. It is briefly described in Taylor (1955) and Blissett (1962). The main body, together with two very small satellite patches to the south, is composed of deformed green serpentinites very similar to the green serpentinites in the Serpentine Hill Complex. The Dundas Serpentinite is in contact with the Dundas Group, the Crimson Creek Formation and the Proterozoic Oonah Quartzites and Slates. The contacts, which locally are silicified or are marked by a zone of sheared talc, are probably faulted and/or the result of solid intrusion of serpentinite.

Several residual kernals of layered pyroxenites and harzburgites occur at the northern end of the Dundas Serpentinite, and are surrounded by veins of cross-fibre asbestos, as at Serpentine Hill (Taylor, 1955).

THE WILSON RIVER COMPLEX

The geology of the general area including the Wilson River Complex is described in Waterhouse (1914), Reid (1921), and Taylor (1954). A geological map is included as Figure 7. Except for its narrow southern extension where it crosses the Pieman River and the Murchison Highway, the complex is accessible overland only with great difficulty. The writer spent a week on the geology of the complex, having been flown in by helicopter and working from exploration camps set up by Aberfoyle N.L. Tracks cut by this company allowed reconnaissance mapping between the Wilson River and Riley Knob, and contacts exposed in Ahearne Creek, Christina Creek, and Riley Creek, together with the spectacular layering

Figure 7. Geology of the Wilson River area.



0
1
2
3
4Km
Faulted or re-intrusive contact
Possible fault
N
WILSON RIVER COMPLEX
HUSKISSON SYNCLINE
Wilson River
Ahearne Creek
Christine Creek
Huskisson River
Riley Creek
Riley Road
Pieman River
Ring River
Renison Bell
Murchison Hwy
SERPENTINE HILL COMPLEX
Colebrook Hill
H
D
LOWER CARBONIFEROUS
Meredith Granite
ORDOVICIAN, SILURIAN, LOWER DEVONIAN
Mainly sandstones and mudstones
MIDDLE TO UPPER CAMBRIAN
Dundas Group
Huskisson River Group
ULTRAMAFIC-GABBRO COMPLEXES
Serpentinites, with some pyroxenites, harzburgites, & gabbros
Amphibolite lens
EOCAMBRIAN
Gabbro (not related to complexes)
Crimson Creek Formation
PROTEROZOIC
Oonah Quartzite & Slate

on Riley Knob, were examined in detail.

The complex is essentially composed of serpentinites, generally sheared, with scattered lenses of primary ultramafics. These lenses vary in size from less than a metre to larger masses forming ridges which are generally parallel to elongation of the complex. Most lenses observed by the writer are composed of orthopyroxenites or interlayered orthopyroxenites and harzburgites. Waterhouse (1914) and Reid (1921) describe a variety of orthopyroxenites, websterites, norites, two-pyroxene gabbros and peridotites, especially from the area north of the Wilson River. Much osmiridium was recovered from streams in the area (Reid, 1921). In the Heazlewood River and Mt. Stewart areas, this mineral can be associated with dunites, so perhaps some of the serpentinites in the Wilson River Complex were derived from similar dunites.

The Wilson River Complex is underlain by the Crimson Creek Formation, consisting mainly of argillites and wackes with some cherts and basic volcanics. These are folded, with the main folds trending NW. The contact between the ultramafics and these sediments is quite straight and is marked by a sharp change in vegetation from myrtle forest covering the sediments to eucalypts with dense undergrowth covering the ultramafics. The contact dips very steeply NE and generally consists of highly sheared serpentinite against unmetamorphosed sedimentary rocks. Waterhouse (1914) and Blissett (1962) describe local silicification and slight hornfelsing of the country rocks immediately adjacent to the contact, and the present writer interprets this as due to emplacement of warm serpentinite and/or some serpentinization occurring after tectonic emplacement. Several lenses of amphibolites with well-developed foliation occur along the contact, and one of these, which is particularly well-exposed in Ahearne Creek, is described in detail in a later chapter. A thin zone of deformed and weathered serpentinite occurs between the amphibolites and the sedimentary rocks.

The contacts with the overlying sedimentary rocks are also marked by sheared serpentinites. In the Huskisson River area, the complex is overlain by the Huskisson River Group consisting of mudstones, wackes, conglomerates and acid volcanics (Taylor, 1954). These rocks are very similar to the Dundas Group and also contain Middle and Upper Cambrian fossils (Banks, 1962; Blissett, 1962). Further north, the Huskisson River Group is wedged out between the complex and the Ordovician, Silurian and Devonian sedimentary rocks of the Huskisson Syncline. It is not known whether the latter rocks overlie the Huskisson River Group unconformably; they are faulted against each other in Christina Creek, and much of the contact SE of this locality may be faulted. North of Christina Creek, sheared serpentinites are in direct contact with the Ordovician and Silurian rocks. These contacts may be the result of faulting and/or solid intrusion of serpentinite. Lenticular bodies of conglomerates and limestones are probably Ordovician in age, while the sandstones, siltstones and mudstones comprising most of Huskisson Syncline are Silurian and Devonian (Taylor, 1954; Banks, 1962). The Wilson River Complex, therefore, has been placed in contact with rocks as young as Silurian, either by faulting or solid intrusion as serpentinite.

South of the Pieman River, the Wilson River Complex tapers to a narrow body enclosed within the Crimson Creek Formation, and ends in a

Y-shaped body of sheared green serpentinites near Colebrook Hill, which is only 2 km east of the NE part of the Serpentine Hill Complex. Kernals of pyroxenites and asbestos veins occur in the serpentinites near the Murchison Highway (Taylor, 1955).

The Wilson River Complex is part of a larger ultramafic-gabbro complex which has been subjected to serpentinization and tectonic re-emplacement. A high level aeromagnetic survey over Tasmania (Finney and Shelley, 1967) shows that the largest anomaly in the state (over 600 gammas) is centred on the Huskisson Syncline rather than the Wilson River Complex. Either the Wilson River Complex is folded underneath the syncline, or, more likely, a larger body occurs beneath the syncline. The latter idea is supported by the three isolated lenses of serpentinized ultramafics occurring along the eastern margin of the syncline - these bodies, together with the Wilson River Complex, may have all been produced by dismembering of a large ultramafic-gabbro complex. Perhaps they were all emplaced into faulted anticlines which originally flanked the Huskisson Syncline.

THE McIVOR HILL GABBRO AND THE TRIAL HARBOUR SERPENTINITE

Both these bodies are crossed by the Zeehan-Trial Harbour road. The McIvor Hill Gabbro, which is poorly exposed and highly altered, is generally referred to as a "hornblende gabbro" (Waterhouse, 1916; Blissett, 1962). Only a small number of specimens were studied by the present writer, who found no convincing evidence of primary hornblende; the samples consist essentially of saussuritized plagioclase and uralitic amphiboles and chlorite replacing clinopyroxene, orthopyroxene, or both. In other ultramafic-gabbro complexes, the writer has observed non-uralitic secondary hornblende replacing clinopyroxene, and it is possible that the hornblendes in some of the specimens described by Waterhouse (1916) are of this origin.

Fine-scale layering, with small cross-cutting bodies of pegmatitic gabbros, were observed in one locality. From field and petrographic observations, it is concluded that the McIvor Hill Gabbro is more related to gabbros forming part of ultramafic-gabbro complexes, in particular the Serpentine Hill Complex, than to smaller gabbro bodies occurring outside these complexes. This is supported by a chemical analysis given in Spry (1962b); the TiO_2 contents of rocks from the ultramafic gabbro complexes (including the McIvor Hill Gabbro) are less than 0.6%, whereas Cambrian gabbros and dolerites which do not form part of these complexes have TiO_2 contents greater than 1.5%.

The McIvor Hill Gabbro appears to be chilled against rocks of the Crimson Creek Formation near the South Comstock Mine and sediments are hornfelsed adjacent to the contact (Blissett, 1962). Other contacts are faulted. Of all the Tasmanian ultramafic-gabbro complexes and ophiolites, the McIvor Hill Gabbro is therefore the only one that can be shown to have crystallized *in situ* in Eocambrian sedimentary rocks.

A date of 518 ± 133 m.y. using the Rb-Sr method is given in Brooks (1966). The age obtained is of little use, for besides the large error involved, Brooks points out that isotopic redistribution could have occurred during the strong alteration suffered by the gabbro.

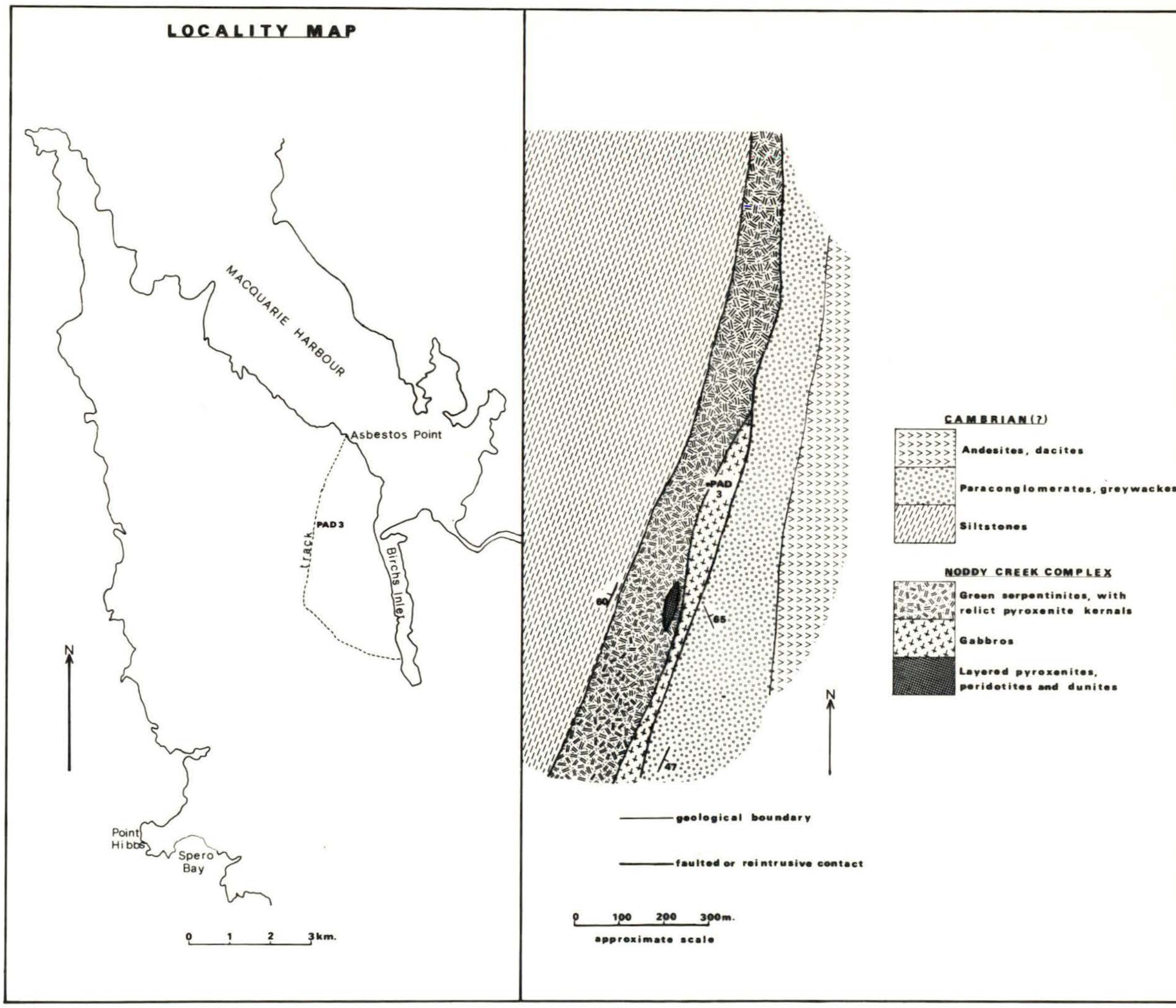
The geology of the Trial Harbour Serpentinite is described in Green (1966a). The ultramafics have been hornfelsed by the Heemskirk Granite (Green, 1966b) and very little can be said about the original nature of the rocks.

THE NODDY CREEK AND SPERO BAY COMPLEXES

The Spero Bay Complex and the serpentinites at Asbestos Point, which is in Macquarie Harbour at the extreme north ot the Noddy Creek Complex, are briefly described in Taylor (1955). Mapping by geologists from Broken Hill Proprietary established that the ultramafics form an almost continuous belt from Asbestos Point through to Hibbs Lagoon. A magnetic anomaly running from Hibbs Lagoon to the Spero Bay Complex suggests that the two complexes may in fact be connected (Flood, 1972).

At Asbestos Point, the ultramafics consist mainly of green, massive to highly sheared serpentinites. Small residual kernals of pyroxenite are common, and veins of cross-fibre asbestos surround these in a similar fashion as in the Serpentine Hill Complex. Flood (1972) states that the ultramafics at Asbestos Point are faulted off about 0.8 km inland, and reappear as two separate belts 3 km south of Macquarie Harbour. These converge 11 km further south, and continue for 10 km to Hibbs Lagoon. In width the belt varies from 50 to 1000 m. The dominant rock types are massive and sheared varieties of green serpentinites. Scattered lenses of primary ultramafics and gabbros are enclosed in the serpentinites; orthopyroxenites are the most abundant, but interlayered pyroxenites and harzburgites or dunites, pyroxenites grading to norites, and various gabbroic rocks also occur.

Figure 8. Geology of the Pad 3 area, Noddy Creek Complex.


LOCALITY MAP
MACQUARIE HARBOUR
Asbestos Point
PAD 3
track
Birchs Inlet
N
Point Hibbs
Spero Bay
0 1 2 3km.
PAD 3
60
65
47
N
CAMBRIAN (?)
Andesites, dacites
Paraconglomerates, greywackes
Siltstones
NODDY CREEK COMPLEX
Green serpentinites, with relict pyroxenite kernals
Gabbros
Layered pyroxenites, peridotites and dunites
geological boundary
faulted or reintrusive contact
0 100 200 300m.
approximate scale

The Noddy Creek Complex is elongate NNE, essentially parallel to the regional structures. Along most of its length it is bordered by Cambrian sedimentary rocks, but along part of its western contact it is juxtaposed against a thin elongate fault slice of Ordovician rocks (M. Hall, B.H.P., *pers. comm.*). The Cambrian rocks are mainly mudstones and siltstones on the western side while on the eastern side they are mainly mudstones, wackes, conglomerates and andesitic volcanics. The andesites could represent a western extension of the Mt. Read Volcanics.

The area around the Broken Hill Proprietary camp at Pad 3, which is accessible by tracks passable to tracked vehicles from Asbestos Point or Birch's Inlet, was the only part studied by the present writer. A sketch map is included as Figure 8. The ultramafics in this area are largely massive or sheared green serpentinites. In many localities, the massive green serpentinites are obviously replacing pyroxenites. In the massive types, stichtite occurs as disseminated blebs in some localities, while in some deformed varieties stichtite occurs as small lenses. Residual orthopyroxenites, ranging from small kernals a few cm in diameter to lenses up to several tens of metres in size, are abundant. In some of these, interstitial plagioclase is obvious in hand specimen. One lens included in the serpentinites consists of orthopyroxenites interlayered with harzburgites and dunites, the layering being on a scale of several centimetres. A long strip of gabbros occurs along the eastern margin. These are medium-grained and although the pyroxenes define a foliation there is no obvious mineralogical layering. The gabbros are more similar to gabbro and microgabbro bodies which intrude the layered rocks in the Serpentine Hill and Heazlewood River bodies than to gabbros forming part of the layered sequences. Small dykes of pegmatites and trondhjemites intrude the gabbros and in places appear to have caused metasomatic alteration.

The contacts of the ultramafics and gabbro and the Cambrian sedimentary rocks are near vertical and are probably the result of both faulting and solid intrusion of serpentinite. Contact zones filled with friable clays and iron oxides may be faults. In other localities, finely interbanded and crenulated talc and serpentines marks the contact zone, while in other places serpentinite which may be very sheared or only slightly deformed occurs as the contact (see Plates 4 and 5). At one locality, chrysotile cross-fibre veins occur within unmetamorphosed siltstones near the contact, suggesting that at least some of the asbestos formation occurred after tectonic emplacement of the complex into its present position.

Small clasts of a soft-green mineral, together with rarer small fragments of serpentinite, occur in several localities in conglomerates east of the complex. X-ray diffractions of the green mineral show that it is a mica, while a partial analysis gave of the order of 2000 ppm Cr (M. Hall, B.H.P., *pers. comm.*). It appears, then, that this mineral is the chrome mica fuchsite. A relatively large pebble of fuchsite (specimen 38835) shows that it has replaced a gabbro very similar to those occurring in the complex just west of the conglomerate. It is concluded, then, that the ultramafics and gabbros were exposed to erosion during the deposition of the conglomerates, and that by later faulting and solid intrusion of serpentinite, the complex was placed in contact with Cambrian sedimentary rocks which previously disconformably or unconformably overlay it. Other than they are Cambrian, the exact age of the sedimentary rocks east of the Noddy Creek Complex is uncertain. They are lithologically similar to Middle and Upper Cambrian rocks elsewhere, but correlations on this basis over such large distances cannot be justified.

Part of the Spero Bay Complex is briefly described in Taylor (1955). It consists of layered ultramafics, of which orthopyroxenites, lherzolites and perhaps dunites are the most abundant types. The olivine-bearing layers have been partly or largely replaced by massive serpentinites, and Taylor (1955) observes that the best development of asbestos veins occurs at lherzolite-serpentinite contacts. More recently a large area of gabbros was found by B.H.P. geologists on the northern side of the ultramafics.

THE ANDERSONS CREEK COMPLEX

The first detailed description of the Andersons Creek Complex is by Taylor (1955). The geology of the general area is discussed in some detail in Green (1959), who established that the ultramafics are pre-Ordovician in age on the occurrence of chromite in several horizons of Ordovician conglomerates. Green (1959) describes the petrography of the ultramafics and gabbros in detail, and concludes that the Andersons Creek Complex consists mainly of partly to completely serpentinized orthopyroxenite with lesser peridotite and clinopyroxene-rich rocks. He argues that alteration of intrusive gabbros to rodingites probably occurred during serpentinization, and that albitite dykes occurring in some localities may be linked to the "hydrothermal" processes involved in serpentinization and the formation of rodingites. The origin of the rodingites is also discussed by Baker (1958).

The present writer, who spent only two days looking at parts of the Andersons Creek Complex, largely is in agreement with Gee and Legge (in press) concerning the origin of the complex. These authors, who mapped the layering and structures in some detail, believe that the complex was essentially a layered ultramafic-gabbro complex which has been in part serpentinized and tectonically re-emplaced. The gabbroic rocks, some of which are interlayered with pyroxenites, include plagioclase pyroxenites, norites, and two-pyroxene gabbros. Some of these show fine-scale layering. Other gabbros, including hornblende gabbros, intrude the layered rocks, and have been converted into rodingites where the primary ultramafics have been serpentinized. Contrary to Green (1959), Gee and Legge (in press) argue that the albitite dykes were emplaced before the formation of the rodingites and probably belong to some magmatic event prior to serpentinization. Three septa of metamorphosed greywackes included in the complex are thought to have been hornfelsed during the magmatic stage of the ultramafic-gabbro complex (Gee and Legge, in press). The septa have suffered only low grade metamorphism, however, for the presence of albite and biotite suggests albite-epidote-hornfels facies (Turner, 1968). Cambrian slates along the western boundary of the complex have been altered to andalusite hornfelses, and Gee and Legge (in press) believe this metamorphism is due to re-intrusion of the complex as a warm serpentinite body.

In summary, the Andersons Creek Complex is believed to have crystallized essentially as a layered ultramafic-gabbro complex. Following serpentinization, re-emplacement occurred essentially as a warm serpentinite body into early Upper Cambrian sedimentary rocks, thermally metamorphosing slates at the contacts. Exposure to erosion occurred during deposition of Lower Ordovician conglomerates. During the Tabberabberan deformation, the sedimentary pile was pushed against the Precambrian block to the west, and a series of folds and imbricate thrusts developed. Gee and Legge (in press) believe that "the participation of Andersons Creek Complex in the Tabberabberan Orogeny is minimal."

ULVERSTONE-FORTH AREA

Several very small bodies of deformed green serpentinites occur in the Ulverstone-Forth area, and are briefly described in Taylor (1955) and Burns (1964). This occurrence is unusual in that they occur entirely within metamorphosed Precambrian rocks in the centre of the Forth Nucleus. They have probably been emplaced as solid serpentinites into their present locations, but it is not known whether or not they were originally derived from a Cambrian ultramafic-gabbro complex.

THE ADAMSFIELD AND BOYES RIVER COMPLEXES

The Adamsfield Complex is described in Reid (1921) and Nye (1929). More recently, the Adamsfield area was mapped as part of a large project on the Cambrian and Ordovician sedimentation (Corbett, 1970) and the Adamsfield Complex studied in detail as an honours B.Sc. project (Brown, 1972). The present writer spent only 5 days mapping rocks of the Adamsfield Complex.

The geology of the area is shown in Figure 9. The Adamsfield Complex, which has been shown by Brown (1972) to be discontinuous along the Saw Back Range (Fig. 10), occupies a faulted anticline and is in contact with an Older Cambrian* greywacke-mudstone-chert sequence, an Upper Cambrian deltaic sequence, and Lower Ordovician fanglomerates and shallow marine sedimentary rocks. The Lake Edgar Fault, along which the Adamsfield Complex is situated, appears to have influenced tectonics and sedimentation as far back as the Proterozoic (Corbett, 1970). The Boyes River Complex is in contact with similar rocks, and as well Proterozoic metamorphics. Some of the contacts are faulted, but Corbett (1970) interprets the relationship of the Upper Cambrian rocks to the Boyes River Complex as an unconformity.

The Adamsfield Complex consists of sheared and massive dark-coloured serpentinites with inclusions of partially to completely serpentinized orthopyroxenites, olivine pyroxenites and dunites (Brown, 1972). The serpentinites are unlike the green waxy-lustred serpentinites

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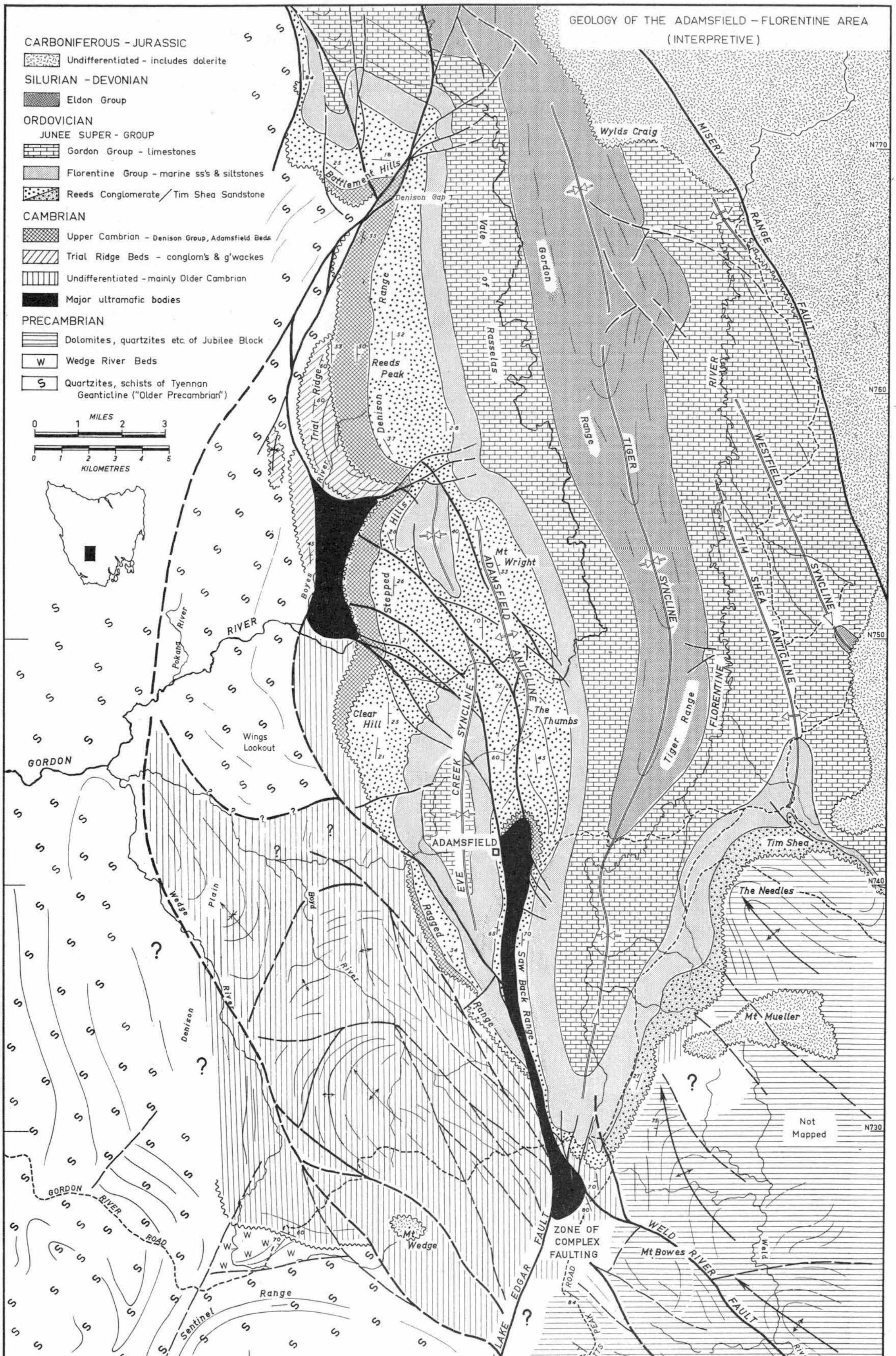
* Equivalent to the Eocambrian in this thesis.

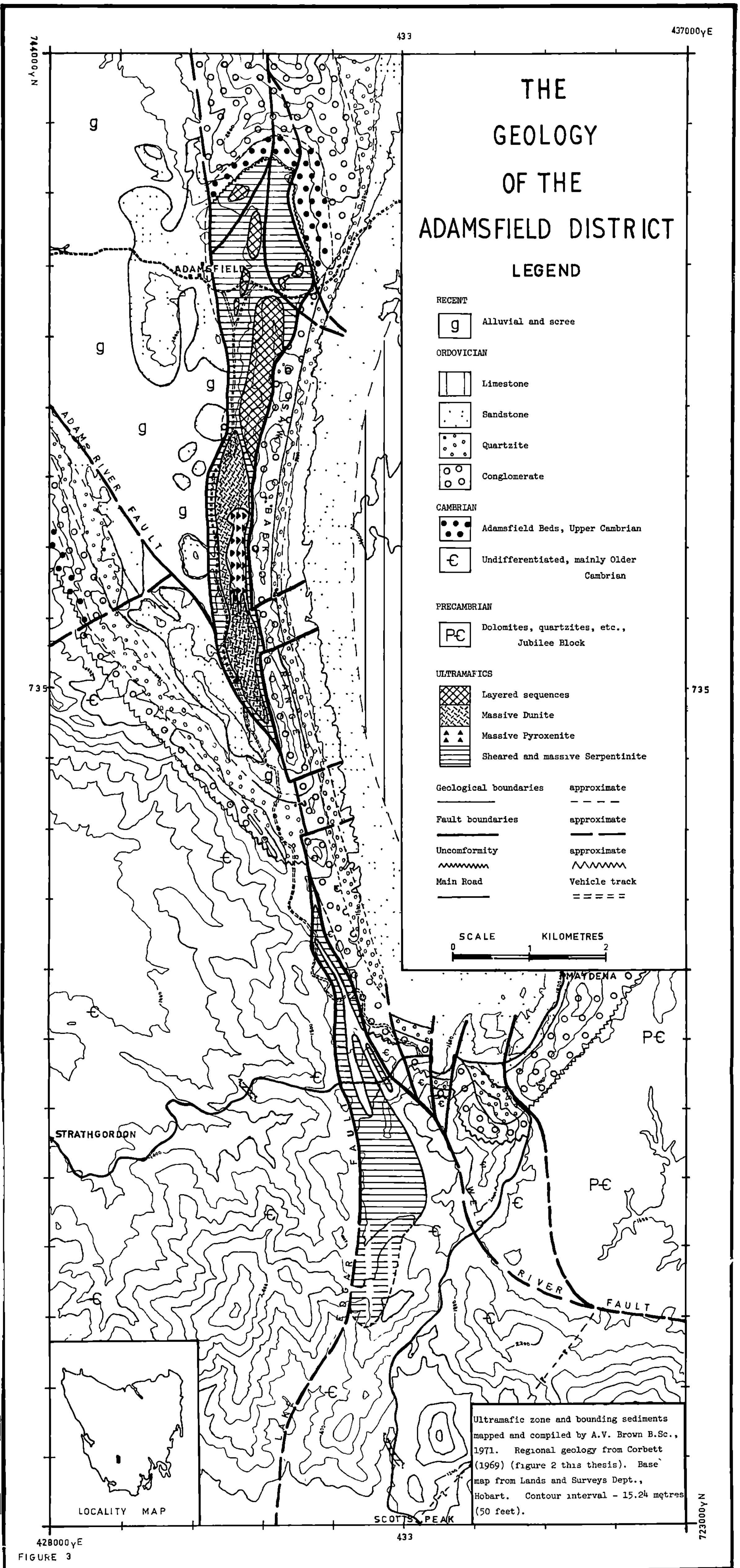
Figure 9. The geology of the Adamsfield - Florentine area. This figure has been taken without alteration from Corbett (1969).


GEOLOGY OF THE ADAMSFIELD – FLORENTINE AREA
(INTERPRETIVE)
CARBONIFEROUS - JURASSIC
Undifferentiated - includes dolerite
SILURIAN - DEVONIAN
Eldon Group
ORDOVICIAN
JUNEE SUPER - GROUP
Gordon Group - limestones
Florentine Group - marine ss's & siltstones
Reeds Conglomerate / Tim Shea Sandstone
CAMBRIAN
Upper Cambrian - Denison Group, Adamsfield Beds
Trial Ridge Beds - conglom's & g'wackes
Undifferentiated - mainly Older Cambrian
Major ultramafic bodies
PRECAMBRIAN
Dolomites, quartzites etc. of Jubilee Block
W Wedge River Beds
S Quartzites, schists of Tyennan Geanticline ("Older Precambrian")
MILES
0 1 2 3
0 1 2 3 4 5
KILOMETRES
Wylds Craig
MISERY RANGE FAULT
Battlement Hills
Denison Gap
Vale of Rasselas
Gordon Range
TIGER SYNCLINE
Tiger Range
FLORENTINE RIVER
WESTFIELD SYNCLINE
TIM SHEA ANTICLINE
Denison Range
Reeds Peak
Trial Ridge
Stepped Hills
Mt Wright
ADAMSFIELD ANTICLINE
EVE CREEK SYNCLINE
Clear Hill
The Thumbs
ADAMSFIELD
Ragged Range
Saw Back Range
Tim Shea
The Needles
Mt Mueller
Not Mapped
Boyes River
Pokana River
RIVER
GORDON
Wings Lookout
Wedge Plain
Boyd River
Denison River
GORDON RIVER ROAD
Mt Wedge
Sentinel Range
LAKE EDGAR FAULT
ZONE OF COMPLEX FAULTING
WELD RIVER FAULT
Mt Bowes
Weld River
SCOTTS PEAK ROAD
N770
N760
N750
N740
N730
N720
E420
E430
E440
Geology by K.D. Corbett

which are common in most of the other complexes. Some of the primary ultramafics are massive and have deformation textures, while some lenses consist of interlayered olivine pyroxenites and dunites, similar to those in other Tasmanian complexes. However, some of the ultramafics are tectonites containing such structures as intrafolial folds (commonly isoclinal and in places refolded by later folds), cross-cutting relationships between pyroxenites and dunites, boudinaged pyroxenites in dunites, and finely layered mylonitized ultramafics which lie between the massive and the layered ultramafics, and may have been derived from the latter during tectonic emplacement of the complex. Small ridges of quartzites near the northern end of the Adamsfield Complex may be tectonic inclusions. An unusual jasperoid rock with patches of a quartz-riebeckite rock occur along part of the NE contact zone is described in Corbett (1970) and Brown (1972); no satisfactory hypothesis for the origin of these rocks can be offered by the present writer.

Abundant serpentinite detritus occurs in an unusual sequence of dolomitic sandstones, mudstones and conglomerates cropping out along the Scotts Peak Road just SE of the Adamsfield Complex (Corbett, 1970). Fossils collected from this sequence indicate a Middle or perhaps Upper Cambrian age (Quilty, 1971). Some of the conglomerates contain more than 50% serpentinite detritus, with some rounded fragments of serpentinite over 10 cm long, as well as pebbles and boulders of dolerites and other altered igneous rocks. At the northern end of the Adamsfield Complex, abundant serpentinite detritus occurs in the lowermost beds of an Upper Cambrian shallow marine sequence. The ultramafics originally were probably unconformably overlain by this sequence (Banks and Carey,

Figure 10. The geology of the Adamsfield district. This figure has been taken without alteration from Brown (1971).



THE GEOLOGY OF THE ADAMSFIELD DISTRICT
LEGEND
RECENT
Alluvial and scree
ORDOVICIAN
Limestone
Sandstone
Quartzite
Conglomerate
CAMBRIAN
Adamsfield Beds, Upper Cambrian
Undifferentiated, mainly Older Cambrian
PRECAMBRIAN
Dolomites, quartzites, etc., Jubilee Block
ULTRAMAFICS
Layered sequences
Massive Dunite
Massive Pyroxenite
Sheared and massive Serpentinite
Geological boundaries
approximate
Fault boundaries
approximate
Uncomformity
approximate
Main Road
Vehicle track
SCALE KILOMETRES
0
1
2
ADAMSFIELD
ADAMS RIVER FAULT
STRATHGORDON
MAYDENA
SCOTTS PEAK
LAKE EDGAR FAULT
WEST RIVER FAULT
LOCALITY MAP
433
437000yE
428000yE
744000yN
723000yN
735
Ultramafic zone and bounding sediments mapped and compiled by A.V. Brown B.Sc., 1971. Regional geology from Corbett (1969) (figure 2 this thesis). Base map from Lands and Surveys Dept., Hobart. Contour interval - 15.24 metres (50 feet).

FIGURE 3

1954; Banks, 1962; Corbett, 1970), but the contacts are now faulted and/ or are the result of solid intrusion of serpentinite. Detrital chromite occurs throughout the Upper Cambrian shallow marine sequence, and less commonly in the Lower Ordovician fanglomerates; the chromite may have been derived directly from exposed ultramafics or by reworking of older sediments.

The Adamsfield Complex is therefore in contact with a number of sedimentary successions which contain detritus originally derived from it and in some cases which unconformably overlay it. It is thought by Corbett (1970), Brown (1972) and the present writer that the complex was probably emplaced into the Lake Edgar Fault zone early in the Eocambrian and then throughout the Lower Palaeozoic was subjected to successive cycles of exposure to erosion and re-emplacement in the Lower Palaeozoic.

The Boyes River Complex is poorly exposed, and appears to be composed mainly of massive serpentinites similar to those in the Adamsfield Complex (K. Corbett, *pers. comm.*).

ROCKY BOAT HARBOUR

A small body of ultramafics is recorded from the Rocky Boat Harbour area, just east of New River Lagoon in the far south of Tasmania (Twelvetrees, 1915). A probable Cambrian sequence of conglomerates, siltstones and mudstones in places containing serpentinite fragments, unconformably overlies Precambrian dolomites in this area. (Banks, 1962).

SUMMARY AND CONCLUSIONS

Ultramafic-gabbro and ophiolite complexes occur in the Cambrian belts flanking Proterozoic blocks. The Heazlewood River Complex is an ophiolite, the Serpentine Hill Complex a dismembered ophiolite, while the other bodies of the Main Belt in western and northern Tasmania are best referred to as dismembered layered ultramafic-gabbro complexes. The Adamsfield and Boyes River bodies are ultramafic complexes and form a separate belt on the eastern side of the Tyennan Block.

With the possible exception of the McIvor Hill Gabbro, all the Tasmanian complexes have been tectonically emplaced into their present locations. The presence of unconformities between several of the complexes and the overlying Cambrian sedimentary rocks, together with the abundance of ultramafic detritus in sedimentary rocks (probably Middle or Upper Cambrian in most cases) suggests that the magmatic events which formed the complexes occurred before the Middle Cambrian. Tectonic re-emplacements appear to have occurred throughout the Lower Palaeozoic, so that many of the bodies are severely dismembered, largely serpentinized, and have been placed in contact with sedimentary rocks as young as Silurian. The disruptive effects of these tectonic and serpentinization processes have been more extreme in the Noddy Creek and Adamsfield areas, where series of narrow elongate serpentinite bodies are strung out along major fault zones.

The Tasmanian bodies can be referred to as alpine complexes in that they occur in an orogenic belt and have been tectonically emplaced into their present locations (Benson, 1926; Turner and Verhoogen, 1960).

It is important to realize that tectonic re-emplacement began at a relatively early stage in the development of the Tasman Orogenic Zone in Tasmania; the tectonic significance of this will be discussed in following chapters.

CHAPTER 3

PETROLOGY OF THE ULTRAMAFICS AND GABBROS OF THE HEAZLEWOOD RIVER COMPLEX

INTRODUCTION

In this chapter the layering and petrography of the ultramafics and intrusive gabbroic rocks of the Heazlewood River Complex is discussed in detail. Brief comparisons are made with similar rocks in the other Tasmanian complexes and the origin of the textures and layering in discussed.

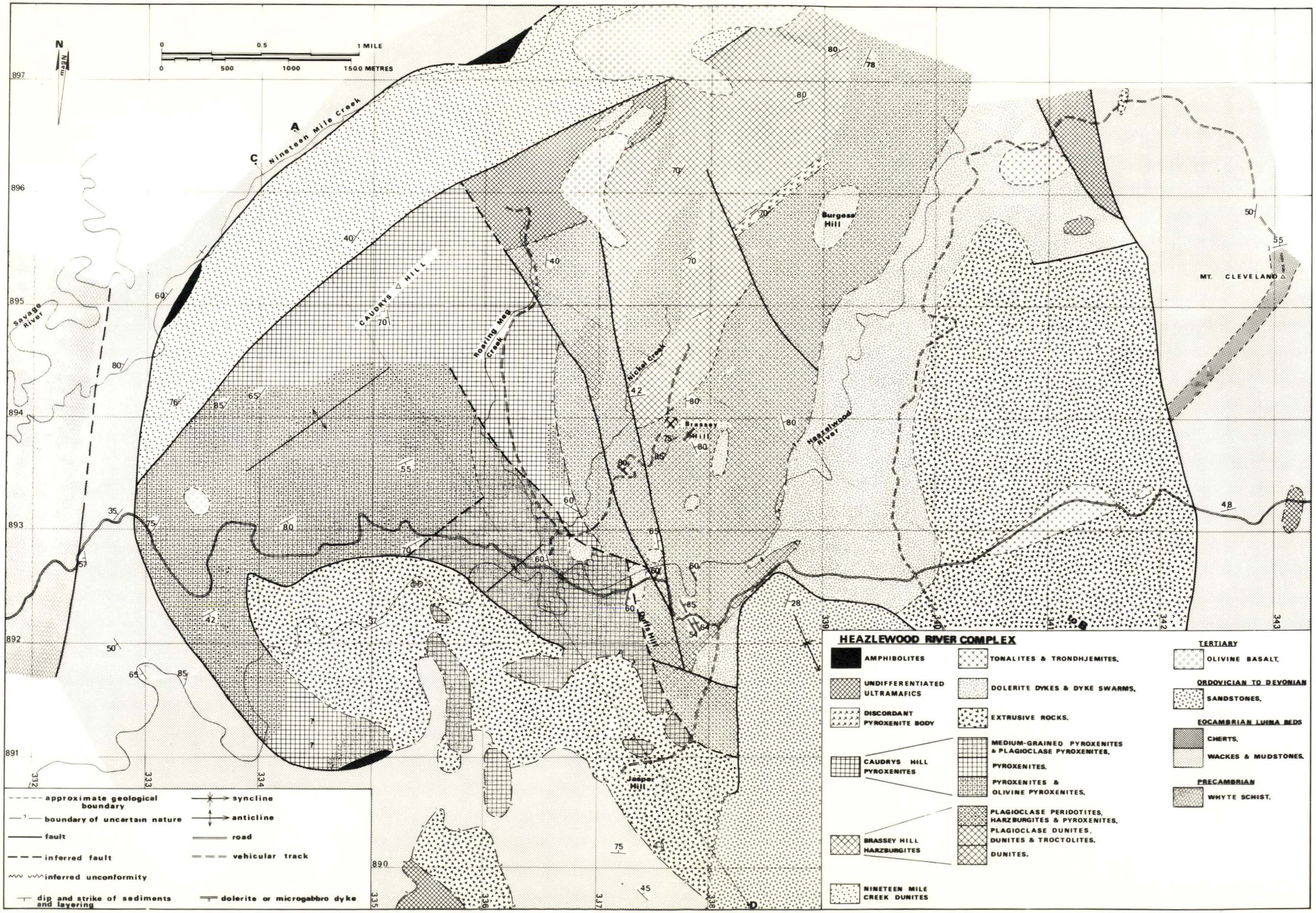
The geology and petrography of these rocks is described in Twelvetrees (1913) and Reid (1921). Allowing for the fact that Reid's is a sketch map with little survey control, and that the present writer prefers to call most of the layered plagioclase-bearing rocks "plagioclase ultramafics" rather than gabbros, there is a fair degree of similarity between the map of Reid (1921) and that of the present writer (Fig. 11). Reid also observed the layering, noted the association of the osmiridium with the plagioclase-free dunites along the NW margin of the complex, and proposed what is essentially a crystal settling hypothesis for the origin of the rocks. Together with other authors writing in the first part of this century, Reid (1921) was influenced by the ideas of Bowen (1915) and thought that the ultramafics and the granites (now known to be Devonian-Carboniferous) were differentiates from the same magmas.

As previously mentioned, much of the area, with the exception of the parts underlain by pyroxenites, is covered in dense scrub. Thus

Figure 11. Geology of the western part of the Heazlewood River Complex.

The provisional Savage River - Waratah 4 in. = 1 mile topographic maps of the Tasmanian Lands and Surveys Dept. were used in preparing the base map.

Mapping done by M. Rubenach, in the period 1967-1971.


HEAZLEWOOD RIVER COMPLEX
AMPHIBOLITES
UNDIFFERENTIATED ULTRAMAFICS
DISCORDANT PYROXENITE BODY
CAUDRYS HILL PYROXENITES
BRASSEY HILL HARZBURGITES
NINETEEN MILE CREEK DUNITES
TONALITES & TRONDHJEMITES.
DOLERITE DYKES & DYKE SWARMS.
EXTRUSIVE ROCKS.
MEDIUM-GRAINED PYROXENITES & PLAGIOCLASE PYROXENITES.
PYROXENITES.
PYROXENITES & OLIVINE PYROXENITES.
PLAGIOCLASE PERIDOTITES, HARZBURGITES & PYROXENITES.
PLAGIOCLASE DUNITES, DUNITES & TROCTOLITES.
DUNITES.
TERTIARY
OLIVINE BASALT.
ORDOVICIAN TO DEVONIAN
SANDSTONES.
EOCAMBRIAN LUINA BEDS
CHERTS.
WACKES & MUDSTONES.
PRECAMBRIAN
WHYTE SCHIST.
approximate geological boundary
boundary of uncertain nature
fault
inferred fault
inferred unconformity
dip and strike of sediments and layering
syncline
anticline
road
vehicular track
dolerite or microgabbro dyke
0 0.5 1 MILE
0 500 1000 1500 METRES
N
Nineteen Mile Creek
Savage River
CAUDRYS HILL
Roaring Meg Creek
Nickel Creek
Burgess Hill
Brassey Hill
Heazlewood River
MT. CLEVELAND
Jasper Hill
Dt's Hill
to B
A
C
D
897
896
895
894
893
892
891
890
332
333
334
335
336
337
338
339
340
341
342
343

mapping was difficult except where lines were cut by mining companies. The NE part of the complex proved very difficult to reach, and the writer did not map it. The outcrop varies from poor to fairly good, but single outcrops are usually quite small and reasonably continuous outcrop is restricted to a few creeks and some road cuttings.

SUBDIVISION OF THE ULTRAMAFICS

The subdivision of the ultramafics into the Nineteen Mile Creek Dunites, the Caudrys Hill Pyroxenites, and the Brassey Hill Harzburgites is shown in Figure 11. This subdivision was made possible by the fact that the rock types are easy to recognize even when serpentinized. Post-serpentinization deformation, during which primary textures are generally destroyed, is mainly restricted to localized zones such as faults. Totally serpentinized rocks NE of Caudrys Hill form the only substantial area where identification of primary rock types proved very difficult, and these are labelled as "undifferentiated" in Figure 11.

The Nineteen Mile Creek Dunites do not contain plagioclase and have altered to massive serpentinites of lighter shades of green than dunites in the Brassey Hill Harzburgites. Distinction between the Brassey Hill Harzburgites and the Caudrys Hill Pyroxenites was generally very easy, except in the faulted area between these groups, where rock types characteristic of both are interlayered in some outcrops.

Near the NE contact of the complex in Figure 11, serpentinites, coarse-grained pyroxenites, and some medium-grained pyroxenites and plagioclase lherzolites are left as "undifferentiated". In the NE part of the complex not mapped by the present writer, the ultramafics are dominantly orthopyroxenites according to Reid (1921).

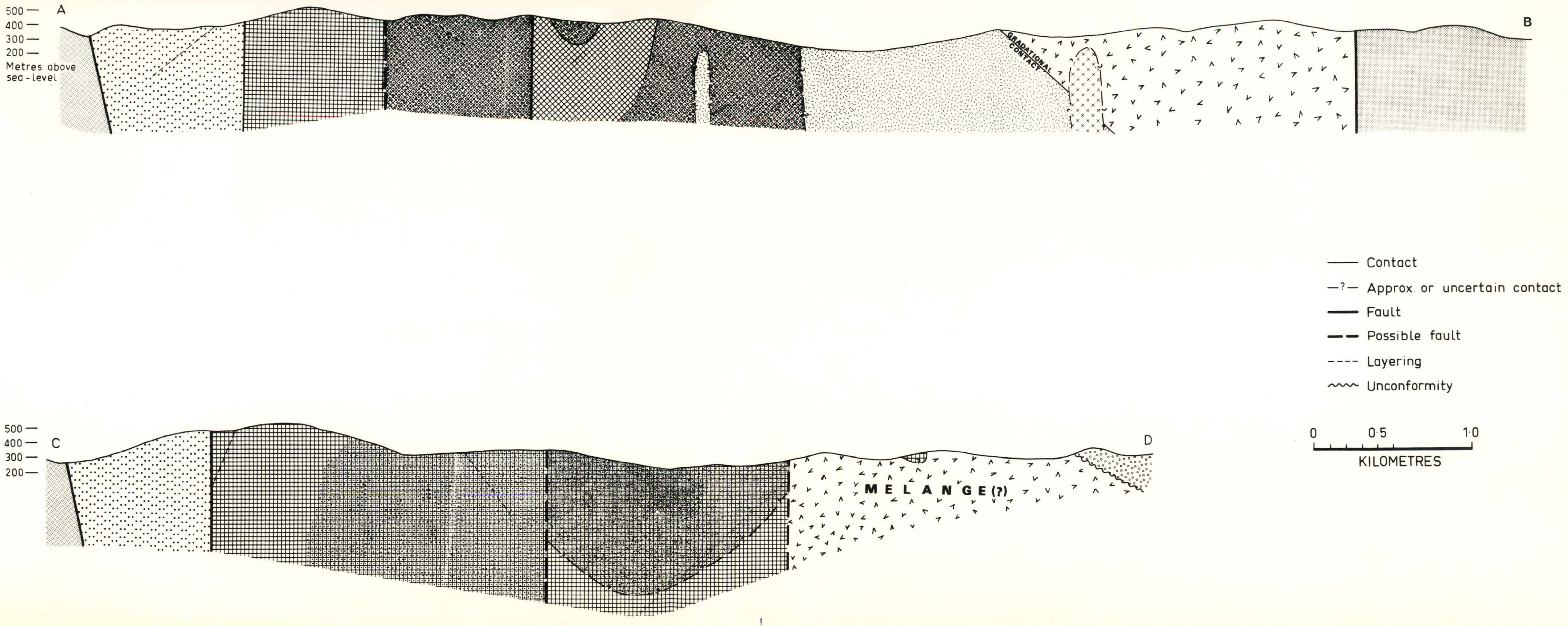
Besides the layered ultramafics, several ultramafic dykes intrude the Brassey Hill Harzburgites just west of Burgess Hill. The largest of these is shown in Figure 11.

ROCK TYPES AND LAYERING

1. The Nineteen Mile Creek Dunites

These consist mainly of dunites largely replaced by massive green serpentinite. Some pyroxene dunites and harzburgites also occur but are fairly rare. Interlayering of dunites and orthopyroxenites on the scale of 0.5-5 cm occurs in some localities, especially near the faulted contacts with the Brassey Hill Harzburgites and the Caudrys Hill Pyroxenites. Orthopyroxenite layers of the order of several metres thick are less common. Where exposed, the layering is similar to the interlayered pyroxenites and harzburgites in the Brassey Hill Harzburgites; no intrafolial folds or cross-cutting relationships between pyroxenites and dunites were observed, which is surprising considering the deformed fabrics developed in these rocks (see discussion in the following sections). Some larger bodies of pyroxenite appear to be surrounded by deformed dunites, and may be "tectonic fish" or boudins. Schlieren of chromitite, described from one locality by Reid (1921, Plate III), may be boudinaged chromitite layers. The present writer found numerous small boulders of chromitite in gravels in creeks about 2 km NE of the summit of Caudrys Hill, but none *in situ*.

Figure 12. Sections through the Heazlewood River Complex. The position of section AB is shown in Fig. 4, while CD is shown on Fig. 11. The legend is the same as for Fig. 11.


500
400
300
200
A
Metres above sea-level
GRADATIONAL CONTACT
B
Contact
—?— Approx. or uncertain contact
Fault
Possible fault
Layering
Unconformity
500
400
300
200
C
MELANGE(?)
D
0
0·5
1·0
KILOMETRES

As previously mentioned, near the contacts with the Luina Beds the serpentinized dunites are sheared and no thermal metamorphism of the sedimentary rocks has been observed.

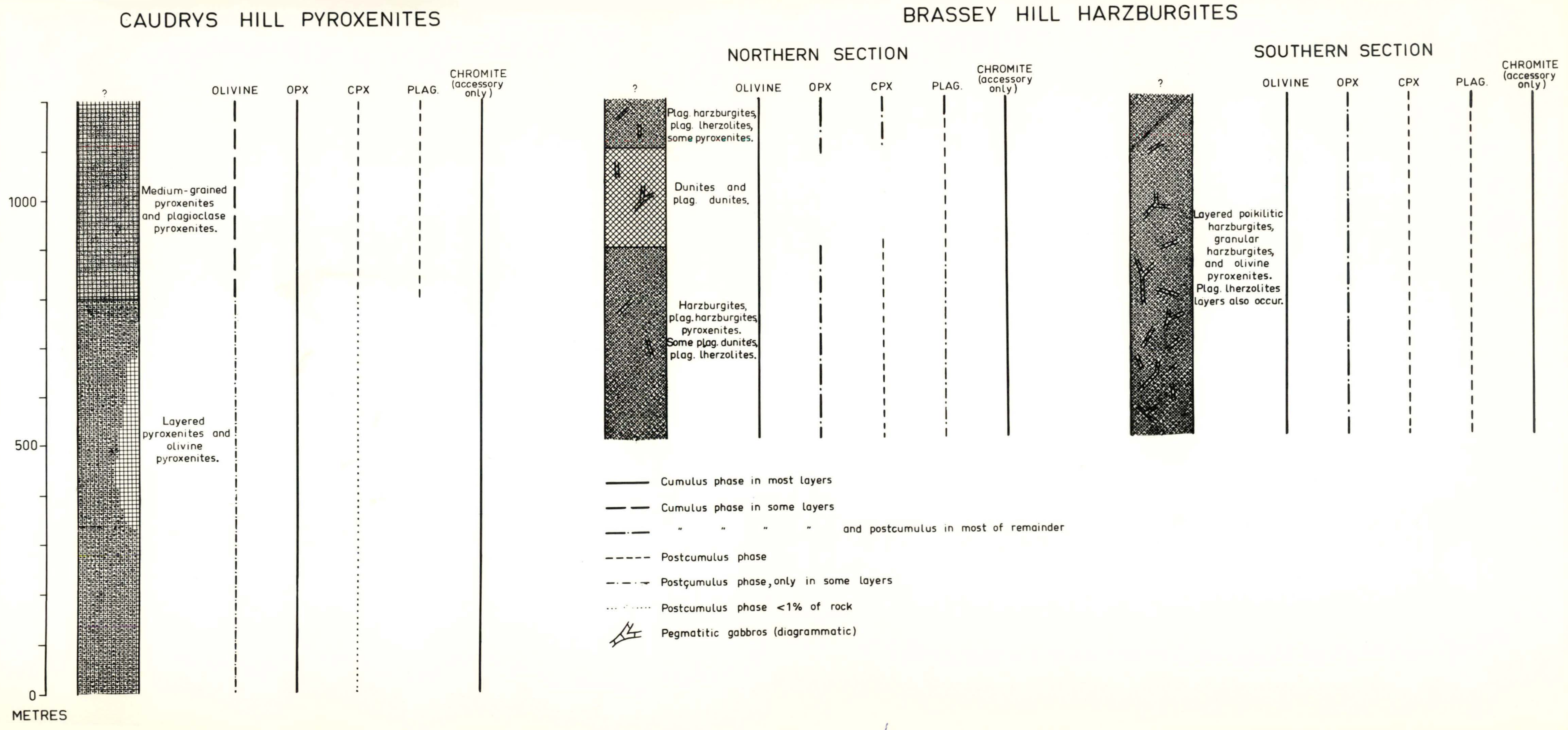
2. The Caudrys Hill Pyroxenites and Brassey Hill Harzburgites

The main rock types occurring in these groups are summarized in Figures 11 and 13. It must be emphasized that the stratigraphic sections in Figure 13 are quite tentative, as the structural data necessary to establish the sequence in the layered rocks are too few. The cryptic variation in chemistry of the rocks is consistent with the sequence deduced on structural grounds (Fig. 15). However, this may be fortuitous as there could be multiple intrusions. Also, layered intrusions do not necessarily show a regular cryptic variation (see Hawkes, 1967). Of the stratigraphic sections, the Caudrys Hill Pyroxenites section is probably the most reliable and the northern Brassey Hill Harzburgites the least reliable.

Most of the lower part of the Caudrys Hill Pyroxenites is composed of interlayered olivine pyroxenites and orthopyroxenites, except in the east and NE where most pyroxenites are olivine-free. In contrast, the upper part of the sequence is composed mainly of medium-grained pyroxenites, plagioclase pyroxenites, with less abundant layers of norites, and rare plagioclase lherzolite and wehrlite layers.

Little evidence of serpentinization is visible in the typical hillside outcrops of the Caudrys Hill Pyroxenites, but in the more continuous exposures in road cuttings and along Roaring Meg Creek, partial serpentinization along joints is obvious. The pyroxenites are fully

Figure 13. Hypothetical columnar sections through the ultramafics of the Heazlewood River Complex. The legend is the same as for Fig. 11.


CAUDRYS HILL PYROXENITES
?
OLIVINE
OPX
CPX
PLAG.
CHROMITE (accessory only)
Medium-grained pyroxenites and plagioclase pyroxenites.
Layered pyroxenites and olivine pyroxenites.
1000
500
0
METRES
BRASSEY HILL HARZBURGITES
NORTHERN SECTION
?
OLIVINE
OPX
CPX
PLAG.
CHROMITE (accessory only)
Plag. harzburgites, plag. lherzolites, some pyroxenites.
Dunites and plag. dunites.
Harzburgites, plag. harzburgites, pyroxenites. Some plag. dunites, plag. lherzolites.
SOUTHERN SECTION
?
OLIVINE
OPX
CPX
PLAG.
CHROMITE (accessory only)
Layered poikilitic harzburgites, granular harzburgites, and olivine pyroxenites. Plag. lherzolites layers also occur.
Cumulus phase in most layers
Cumulus phase in some layers
" " " " and postcumulus in most of remainder
Postcumulus phase
Postcumulus phase, only in some layers
Postcumulus phase <1% of rock
Pegmatitic gabbros (diagrammatic)

serpentinized only adjacent to faults and along the sheared contacts with the Luina Beds.

In the southern part of the Brassey Hill Harzburgites, the most abundant rocks are harzburgites, both granular and poikilitic varieties being represented. The olivine/pyroxene ratio in these is quite variable, while interstitial altered plagioclase is present in amounts ranging from 0-2%, and very rarely is as much as 10%. Interlayered with harzburgites are olivine pyroxenites, medium-grained pyroxenites and plagioclase lherzolites in which the plagioclase content rarely rises above 20%. Northeast of the Lord Brassey Mine, some dunites and plagioclase dunites are interlayered with harzburgites.

In the northwestern part of Brassey Hill Harzburgites, the main rock types are plagioclase harzburgites and harzburgites interlayered with pyroxenites and plagioclase dunites. These are apparently overlain by a monotonous sequence of interlayered plagioclase dunites and dunites in which pyroxenes are very rare. Locally, the plagioclase dunites grade into troctolites, but rocks with greater than 30% plagioclase are not common. The dunites and plagioclase dunites appear to be overlain by a sequence of plagioclase lherzolites, plagioclase harzburgites and some pyroxenites. Flanking the plagioclase dunites to the west is a sequence of harzburgites and plagioclase harzburgites, but these were not studied by the writer.

In the western part of the Brassey Hill Harzburgites, particularly in the complex fault zone separating them from the Caudrys Hill Pyroxenites, layers of pyroxenites, medium-grained pyroxenites, and norites are fairly abundant. Just west of the fault zone, some layers of poikilitic harzburgites and rare plagioclase dunites occur in the Caudrys Hill Pyroxenites. There may be, then, a lateral grading of the Brassey Hill Harzburgites into the Caudrys Hill Pyroxenites, but the complex fault zone has made it impossible to establish this with any certainty.

The olivine grains in rocks comprising the Brassey Hill Harzburgites are commonly partly to entirely replaced by serpentine minerals, which are dull in lustre, and are coloured black, greenish black, or, in the case of many dunites and plagioclase dunites, a dark brownish green. Plagioclase is commonly replaced by hydrogrossular or saussurite. Only along shear zones are pyroxenites replaced by serpentinite.

2.1 Layering

In the lower part of the Caudrys Hill Pyroxenites, layering is on a scale of 1 cm to 1 m (Plates 10-12). Outcrops in which no layering is obvious are probably parts of more massive layers. Adjacent layers may differ in the ratio of olivine to pyroxene, and/or textures (for example, degree of preferred orientation, grainsize, grain shape, etc.). In the upper part of the sequence, layering is commonly the expression of textural variations, but compositional variations also occur. Within some plagioclase pyroxenites, a diffuse layering, which is a function of variations in the interstitial plagioclase, may be subparallel to the main layering (Plate 13). In other plagioclase pyroxenites, the distribution of interstitial plagioclase appears to be quite patchy. In some rock lenses of plagioclase, several cm in length, represent single oikocrysts which include numerous small orthopyroxene grains.

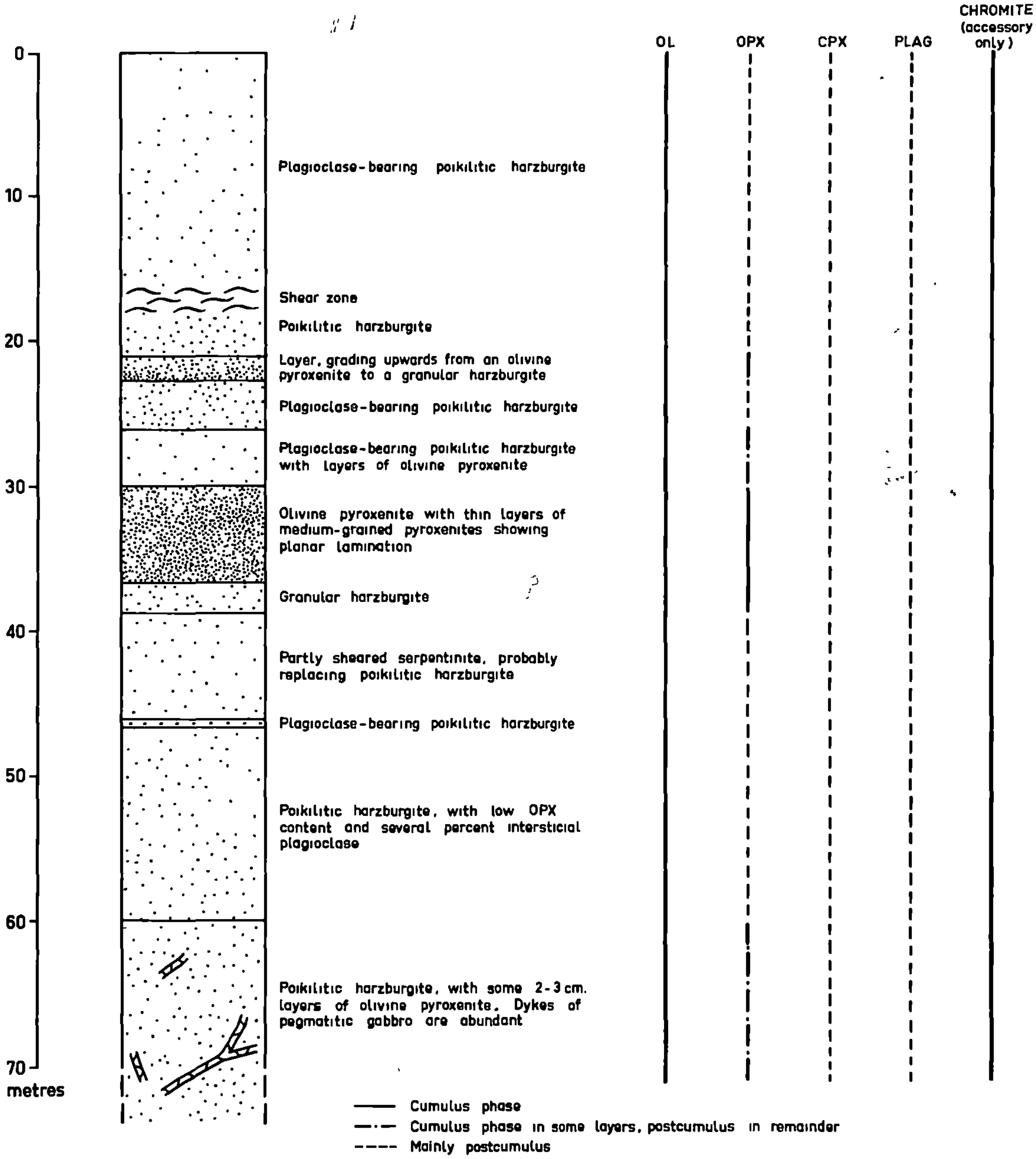
Layering in the Brassey Hill Harzburgites is quite variable in scale, from several millimetres to the order of tens of metres. The layering is more obvious in the field when it reflects abrupt changes in rock types, such as pyroxenites interlayered with poikilitic harzburgites. A reconstructed section (Fig. 14) through a drill-hole put down by Amax Mining at the Lord Brassey Mine shows the typical layering developed in the southern part of the Brassey Hill Harzburgites. Harzburgites may occur as massive layers or as units broken by numerous olivine pyroxenite layers on a spacing as short as 2-3 cm. In some massive harzburgites there is a "sub-layering" on a scale of 2 cm or greater, due to variations in the olivine/pyroxene ratio, while others appear to be quite uniform throughout. The average olivine/pyroxene ratio and the interstitial plagioclase content varies from unit to unit, as is illustrated in Figure 14. Units of olivine pyroxenites and medium-grained pyroxenites interlayered with harzburgites are commonly themselves layered on a scale as fine as several millimetres. The core of a drill-hole put down at 8944N/3376E shows a greater proportion of pyroxenites, olivine pyroxenites, medium-grained pyroxenites and norites interlayered with harzburgites, but the hole passes through a fault zone and much of the core is serpentinized and deformed.

Layering in the plagioclase dunites varies from massive to very fine-scale (Plate 14). Contacts between layers can be quite sharp (Plate 16) or diffuse (Plate 15). The diffuse layering is probably a reflection of the plagioclase being interstitial, and is discussed in a later section.

Figure 14

RECONSTRUCTED SECTION THROUGH DRILL HOLE AT THE

LORD BRASSEY MINE



Because of the lack of good outcrop of these rocks, it is not known to what extent the plagioclase lherzolites are discrete layers or lenses and irregular patches within plagioclase harzburgite or plagioclase dunite layers.

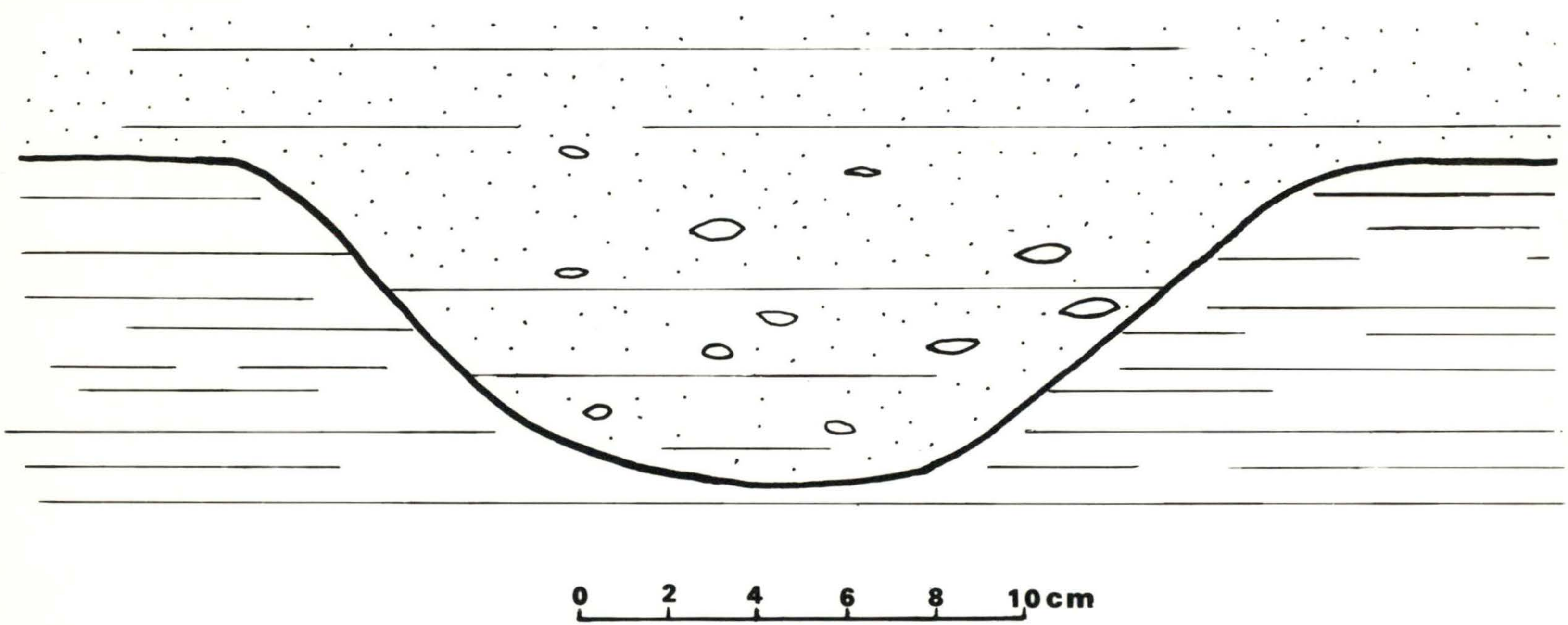
A quantitative study of the modal variations within layers was not done. Thus the proportion of layers showing rhythmic variations in modal composition is not known with certainty. However, it is usually obvious in the field whether or not rhythmic layering is abundant, and very few definite cases were observed by the writer. One example is a layer grading from an olivine pyroxenite to a granular harzburgite shown in Figure 14. Some possibly graded layers also occur in the troctolites and plagioclase-rich dunites.

2.2. Other structures

There is some evidence for current activity during the crystallization of the Brassey Hill Harzburgites. The best evidence includes cross-layering in plagioclase dunites in one locality (Plate 16) and a scour-and-fill structure in olivine pyroxenites (Fig. 16, Plate 17). The latter occurs at 89410N/33805E, and consists of a small trough cut into finely-layered medium-grained olivine pyroxenites which is filled by the overlying coarser-grained olivine pyroxenites. Small elliptical nodules of plagioclase dunites are included in the coarser-grained pyroxenites in or near the trough; these were presumably derived from the "erosion" of a plagioclase dunite by currents within the magma. Besides being caused by currents acting in the magma during the settling of crystals, the writer can think of no other satisfactory explanation for these two structures. They provide some of the best evidence that the layered

Figure 15. "Scour-and-fill" structure in layered pyroxenites.

SCOUR-AND-FILL STRUCTURE IN LAYERED PYROXENITES



ultramafic rocks formed by crystal accumulation from a magma.

In general, the style of the layering in the Brassey Hill and Caudrys Hill Pyroxenites is consistent with a cumulative origin. No intrafolial folds or other structures characteristic of tectonites were found. Some local deformation of primary rocks, such as the fracture or spaced cleavage shown in Plate 17 is superimposed on the primary layering and is probably related to nearby faults or may be axial surface to folds. Post-serpentinization deformation (i.e. local serpentinite shear zones) are fairly common and are probably largely a reflection of faults.

3. Intrusive ultramafics

Several bodies of ultramafics intrude the Brassey Hill Harzburgites west of Burgess Hill. The smallest of these is a plagioclase harzburgite dyke shown in Plate 21. The largest body is a pyroxenite locally grading and interlayered with harzburgite, and is shown in Figure 11. The cross-cutting contact of this body against plagioclase dunites, is shown in Plate 20. Unfortunately the contact zone is largely serpentinized. The significance of these intrusive bodies regarding textures of ultramafic rocks is described in a later section.

In other localities where layering is not well developed, it is difficult to tell if some of the pyroxenite bodies, such as the one in Plate 19, are part of the layered sequence or later intrusives.

TEXTURES OF THE ULTRAMAFIC ROCKS

In this section, the textures of the layered ultramafic rocks and the ultramafic dykes are described in detail. It is shown that some textures are ambiguous, since those developed in ultramafic dykes and in annealed tectonites can be quite similar to certain cumulate textures. However, considering the style of layering and variations in chemical composition of the layered ultramafics as well as the textures, a cumulative hypothesis is preferred for the Brassey Hill Harzburgites and Caudrys Hill Pyroxenites as it is the most consistent with all the observed data.

In describing cumulates, the nomenclature of Jackson (1967) is used in preference to that of Wager and Brown (1968).

1. <u>Intrusive ultramafic bodies</u>

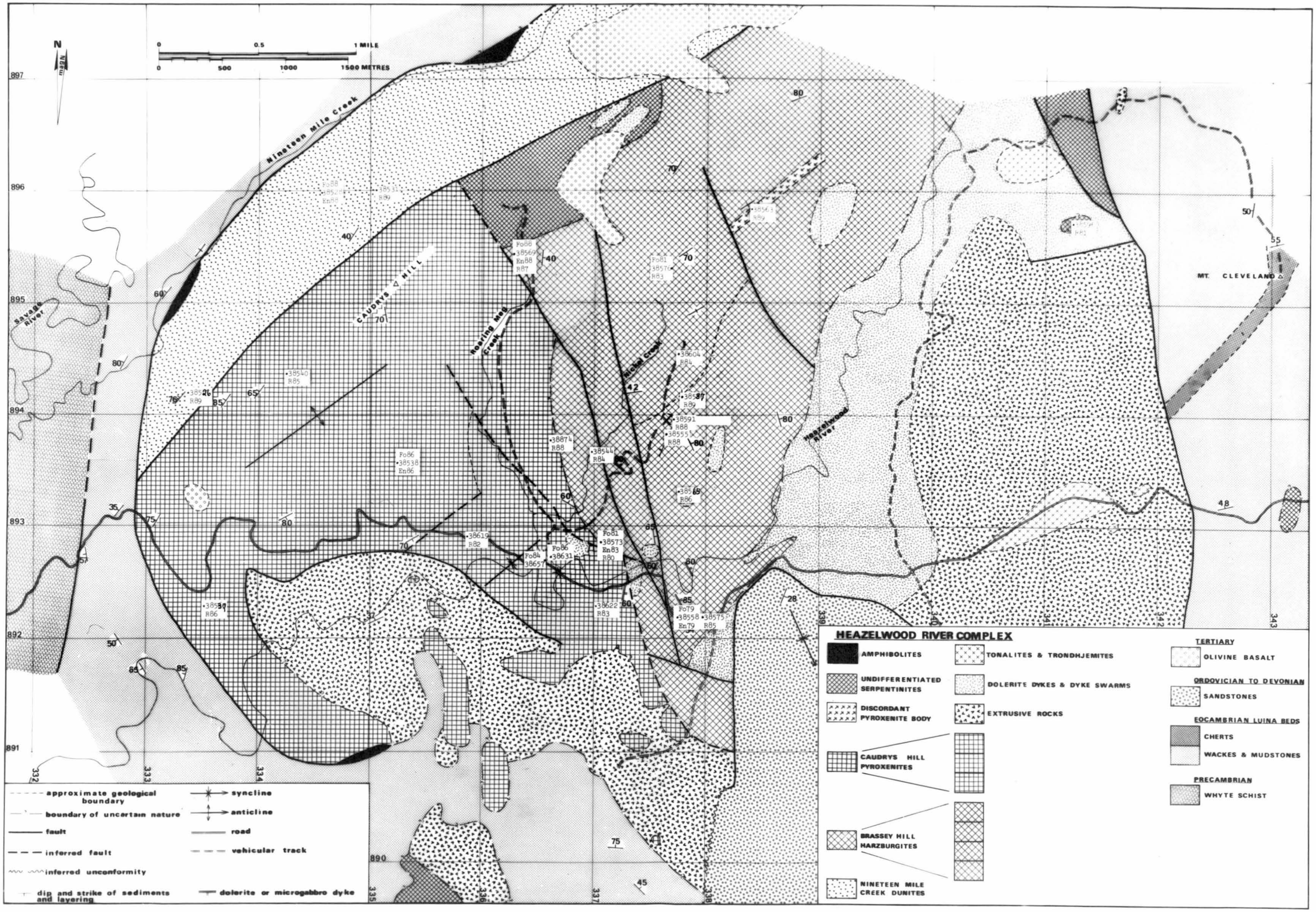
The petrography of these rocks is described first because of the significance of the textures.

Specimen 38916 is from the harzburgite dyke shown in Plate 20. It is composed of approximately 55% orthopyroxene, 35% olivine, 5% each of clinopyroxene and saussuritized plagioclase, and accessory chromite. The grainsize averages 1 mm. Olivine grains are anhedral and partly moulded against subhedral orthopyroxenes which define a foliation parallel to the dyke walls. Where in mutual contact, orthopyroxene grains have triple-point grain boundaries. The altered plagioclase and most clinopyroxene is interstitial.

The larger ultramafic dyke shown in Figure 11 is composed mainly of granular orthopyroxene with a little olivine and clinopyroxene (38564).

Figure 16. Locations and compositions of analysed samples of ultramafics, Heazlewood River Complex.

Fo and En values are for olivines and pyroxenes respectively while "R" is the Mg/Mg+Fe (mole percent) for the whole rock.



N
0 0.5 1 MILE
0 500 1000 1500 METRES
Nineteen Mile Creek
Savage River
CAUDRYS HILL
Roaring Meg Creek
Nickel Creek
Heazelwood River
MT. CLEVELAND
HEAZELWOOD RIVER COMPLEX
AMPHIBOLITES
UNDIFFERENTIATED SERPENTINITES
DISCORDANT PYROXENITE BODY
CAUDRYS HILL PYROXENITES
BRASSEY HILL HARZBURGITES
NINETEEN MILE CREEK DUNITES
TONALITES & TRONDHJEMITES
DOLERITE DYKES & DYKE SWARMS
EXTRUSIVE ROCKS
TERTIARY
OLIVINE BASALT
ORDOVICIAN TO DEVONIAN
SANDSTONES
EOCAMBRIAN LUINA BEDS
CHERTS
WACKES & MUDSTONES
PRECAMBRIAN
WHYTE SCHIST
approximate geological boundary
boundary of uncertain nature
fault
inferred fault
inferred unconformity
dip and strike of sediments and layering
syncline
anticline
road
vehicular track
dolerite or microgabbro dyke

However, near the contact with plagioclase dunites (Plate 19) it consists of alternating layers of olivine pyroxenite and serpentinized pyroxene dunite (38563). The orthopyroxenes define a foliation parallel to the contact, and clinopyroxene is interstitial to orthopyroxene and in places poikilitically enclose them.

The textures developed in these dykes are similar to some of the textures of rocks from the ultramafic zones of the Bushveld and Stillwater Complexes which are believed to be cumulative in origin (Hess, 1960; Jackson, 1961; Wager and Brown, 1968), demonstrating very well the ambiguity of certain textures in ultramafic rocks.

The mode of emplacement of the dykes is uncertain. They may have been emplaced as magmas, as crystal mushes lubricated by interstitial magma, or as hot solid bodies by plastic flow. If the last of these is true, annealing must be postulated to explain the lack of deformation textures. Emplacement occurred before serpentinization, as no thermal metamorphism of the serpentinized country rocks has occurred. Replacement of the walls of the smaller dyke by rodingite is also consistent with serpentinization occurring after emplacement.

2. Brassey Hill Harzburgites

These rocks show every gradation from dunites to plagioclase dunites and plagioclase lherzolites, and from dunites to harzburgites and to orthopyroxenites. Compositionally, they fall in the range Fo79-89 for olivines and En79-89 for orthopyroxenes (electron microprobe analyses). Plagioclase is commonly altered, but where it could be determined optically, a range of An60-65 was obtained from albite twins.

2.1 Dunites and plagioclase dunites

These rocks are generally serpentinized, the olivines being replaced by mesh-textured "lizardite" and the plagioclase by hydrogrossular, saussurite, chlorite, or fine-grained white mica.

Dunites free of plagioclase (38572) are quite rare, for commonly where they appeared so in outcrop they were found to contain a little altered plagioclase in thin section (38654, 38904).

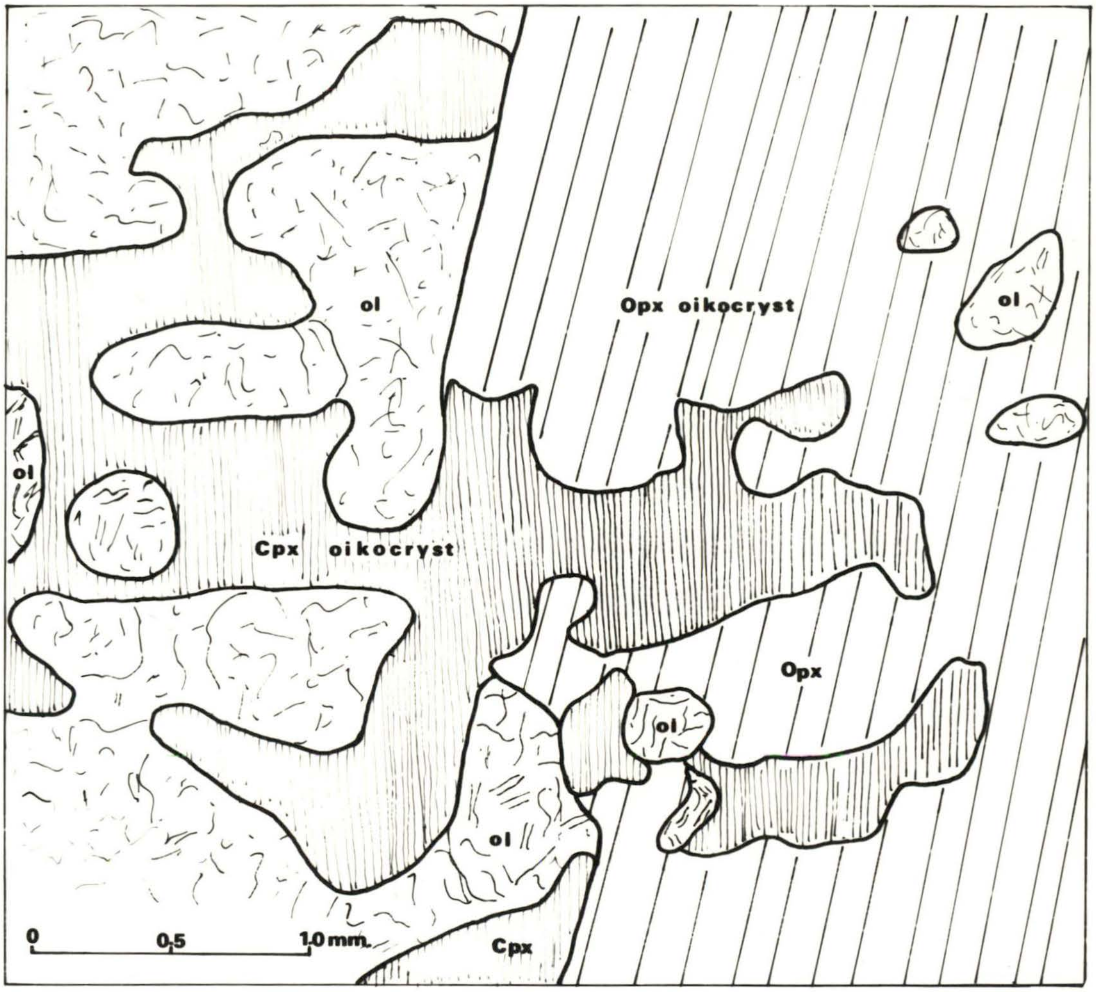
Specimens 38601, 38602, 38604 and 38605 are typical plagioclase dunites and troctolites. Where the outlines of the original olivines are obvious, they are generally ovoid or equant and range in size from 1-3 mm. The plagioclase is now pseudomorphed by fine-grained isotropic hydrogrossular with veins and patches of a cloudy fine-grained mineral (Plate 23), or less commonly by chlorite. When only a little pseudomorphed plagioclase is present, it occurs as cuspate patches interstitial to the pseudomorphed olivines. Where more abundant, the pseudomorphed plagioclase encloses subhedral serpentinized olivines, as in Plate 24. Accessory chromite occurs in most dunites and plagioclase dunites.

The textural relations in the dunites and plagioclase dunites are consistent with olivine being a cumulus phase while plagioclase was in most cases postcumulus. In some of the more plagioclase-rich dunites and troctolites, the plagioclase could be a cumulus phase also. Chromite is probably cumulus.

2.2 Poikilitic harzburgites and lherzolites

Harzburgites and lherzolites in which the pyroxenes are less than 50% and the plagioclase content less than 20% are generally poikilitic. Exceptions include specimen 38554 in which the pseudomorphed ortho-

Figure 17. Texture of lherzolite, 38565


ol
Opx oikocryst
ol
ol
Cpx oikocryst
Opx
ol
ol
Cpx
0
0.5
1.0 mm.

pyroxenes (but not the clinopyroxenes) are non-poikilitic, and 38562 in which the clinopyroxenes are non-poikilitic. In the poikilitic rocks, olivine (generally partly to entirely replaced by mesh-textured "lizardite") and accessory chromite can be interpreted as cumulus phases, while the poikilitic and interstitial pyroxenes would be essentially postcumulus, and the interstitial plagioclase (in most cases replaced by hydrogrossular, saussurite, or chlorite) also postcumulus. In most cases, orthopyroxene is more abundant than clinopyroxene, and some rocks (e.g. 38592) contain no clinopyroxene. Wehrlites, such as 38562, are very rare.

Orthopyroxene oikocrysts are generally in the range 2-30 mm in size. They may be fresh, partly replaced by uralite or talc, or pseudomorphed by single "lizardite" plates (the so-called "bastite" pseudomorphs). They contain (100) exsolution lamellae and in some cases blebs of clinopyroxene. The oikocryst boundaries tend to be moulded against cumulus olivines, while olivine inclusions (which may or may not be serpentinized) are rounded and on the average half the size of those outside the oikocrysts (for example, in 38592, olivine inclusions average 5 mm while those outside the oikocrysts average 10 mm). This could be only partly due to protection of olivine against postcumulus enlargement, but is probably largely a result of reaction between the cumulus olivines and the interstitial magma from which the oikocrysts grew. In contrast to many clinopyroxene oikocrysts, the volume of olivine inclusions never exceeds the volume of the host orthopyroxene oikocryst.

Clinopyroxenes rarely comprise more than 20% of harzburgites and lherzolites. They are usually unaltered, but may be partly replaced by uralite or pseudomorphed by light green or brown hornblende. They commonly occur as interstitial cuspate grains or thin rims surrounding cumulus olivines. In many clinopyroxene oikocrysts, the volume of olivine inclusions greatly exceeds the volume of pyroxene (Plate 25). Cuspate grains and rims may be optically continuous over a large area of thin section, showing that they comprise a single oikocryst enclosing up to 20 cumulus olivines. The interstitial magma crystallizing clinopyroxene did not, therefore, react with the cumulus olivines to the same extent as in the case of orthopyroxene oikocrysts.

In many cases (especially 38554), clinopyroxenes rim orthopyroxene oikocrysts, suggesting that the latter crystallized first. However in 38565, and to a less extent in 38555, parts of clinopyroxene oikocrysts extend into large orthopyroxene oikocrysts. In these specimens, orthopyroxene oikocrysts continued to grow after the clinopyroxene oikocrysts formed, replacing the olivine inclusions in the clinopyroxene, but not reacting with the clinopyroxene itself (Fig. 17).

Although the pyroxene content of poikilitic rocks is generally less than 50%, in some olivine pyroxenites, most (38556) or all (38573) the olivine is poikilitically enclosed in large orthopyroxenes. It is not known whether the orthopyroxenes in such rocks are entirely postcumulus or partly cumulus and partly postcumulus.

2.3 Plagioclase-rich harzburgites and lherzolites

The textures of the rocks are quite variable where the plagioclase content is greater than 10%. In all cases, olivines and accessory chromite could be interpreted as cumulus phases. The olivines, which are partly replaced by serpentine, are usually anhedral equant or ovoid grains, but surrounded by plagioclase they are subhedral stubby prisms or plates (38565, 38568 and 38575).

Orthopyroxenes could be interpreted as cumulus in some rocks (38569 and 38575) while in others they are poikilitic and are probably for the large part postcumulus. In 38558 and 38576, orthopyroxene oikocrysts up to 1 cm long contain anhedral inclusions of cumulus olivine and clinopyroxene and euhedral inclusions of cumulus plagioclase (Plate 26). It is probably significant that the latter two specimens are some of the most differentiated rocks of the exposed Brassey Hill Harzburgites, with olivine compositions of Fo81 and Fo79 respectively.

Clinopyroxene never comprises more than 25% of these rocks. It occurs as interstitial or poikilitic grains in 38565, 38566, 38568 and 38575, while in 38558, 38570 and 38576 it occurs as anhedral grains which may be cumulus.

The plagioclase in most of these rocks is partly or largely altered to saussurite or fine-grained white mica, and never comprises more than 40% of the rock. Albite and pericline twins are common. Compositions in the range An61-64 were measured, and although such compositions are rather low for ultramafic rocks they were all obtained from the more differentiated rocks. When it occurs as polygonal mosaics between the other minerals it is difficult to judge if plagioclase is essentially a cumulus or postcumulus phase.

2.4 Granular harzburgites

Specimen 38571 is an unusually fresh granular harzburgite consisting of 60% orthopyroxene, 40% olivine, and 1% chromite, with an average grainsize of 3 mm. The texture is unusual, with odd-shaped orthopyroxenes moulded against subhedral olivines (Plate 27), perhaps due to interference during postcumulus enlargement of both minerals. Specimen 38560 is a typical granular harzburgite, with cumulus orthopyroxene and olivine and interstitial postcumulus plagioclase and some clinopyroxene.

2.5 Olivine pyroxenites and pyroxenites

Specimens 38591 and 38594 are typical olivine pyroxenites. They are composed of subhedral orthopyroxene and anhedral olivines with some interstitial saussuritized plagioclase and accessory chromite grains. Clinopyroxene occurs as interstitial rims and cuspate patches, and less commonly as large oikocrysts in which the olivine and orthopyroxene inclusions have a much smaller size than grains outside the oikocryst (see Plates 28 and 29). These can be interpreted as having classic cumulate textures in which the orthopyroxenes and olivines "settled", grew larger by postcumulus addition, and in which postcumulus plagioclase and clinopyroxene crystallized interstitial to and, in the latter case partly reacting with, the cumulus grains.

Specimens 38586 and 38587 were collected from the "scour-and-fill" exposure (see Fig. 15 and Plate 17). The former is a olivine pyroxenite of average grainsize less than 1 mm collected from the lower scoured horizon. Specimen 38587, collected from inside the scour, is an olivine pyroxenite averaging 2 mm in grainsize and contains dunite and plagioclase dunite nodules. The origin of the structure is described in a previous section. The fine-scale layering developed in both the olivine pyroxenites is due to variations in interstitial plagioclase content.

Specimen 38544 consists mainly of granular anhedral clinopyroxenes with scattered large orthopyroxenes. It has the dubious distinction of being the only clinopyroxene-dominated ultramafic rock found by the writer in the Tasmanian complexes.

2.6 Medium- and fine-grained pyroxenites

Pyroxenites similar to some of the medium- and fine-grained pyroxenites in the upper part of the Caudrys Hill Pyroxenites occur interlayered with the harzburgites in some localities. These range in size from less than 1 mm to several millimetres, and in many cases have a well-developed planar lamination parallel to the layer contacts. Fabric studies, which would determine the degree of alignment of crystallographic axes, were not done. The orthopyroxenes commonly have developed triple-point junctions, except where interstitial plagioclase is present. Some specimens (e.g. 38567 and 38584) contain scattered anhedral clinopyroxenes up to 2 mm in size. The textures in many of these rocks, although consistent with a cumulative origin, are nevertheless ambiguous, since they are similar to those of observed pyroxenite dykes (see 38916 in section on intrusive ultramafic bodies). This problem is discussed in more detail in another section.

2.7 Chromitites

No chromitite layers were observed within the Brassey Hill Harzburgites, but a lens of chromitite, partly replaced by the chlorite kammererite, was found in a serpentinite shear-zone on the SW slopes of Brassey Hill. The texture of this rock (Plate 30) is consistent with a cumulative origin. Massive chromitite at a locality 3385E/8958N occurs in a pegmatitic gabbro dyke and is described in another section.

2.8 Summary and conclusions

Despite the ambiguity of some of the textures a cumulative origin is preferred for the Brassey Hill Harzburgites. In accordance with this hypothesis, most of the rocks would be classified as olivine cumulates, in that cumulus olivine occurs in most layers, the only exceptions being olivine-free pyroxenite or norite layers. In plagioclase dunites, the interstitial plagioclase could be interpreted as postcumulus, but could also be a cumulus phase in some troctolite layers. In poikilitic harzburgites and lherzolites, orthopyroxenes, clinopyroxene and plagioclase would be postcumulus. In granular harzburgites and olivine pyroxenites, both olivine and orthopyroxene would be essentially cumulus phases. Plagioclase lherzolites, in which olivine, clinopyroxene and plagioclase would be cumulus phases and orthopyroxene essentially postcumulus, happen to be the most differentiated rocks and could be representative of the top of the exposed sequence. Accessory subhedral chromite, which is slightly more abundant in the dunites and harzburgites than in pyroxenites, is probably cumulus throughout.

The origin of the layering is discussed in another section.

Another question that needs to be considered is the exact relationship of the northern and southern parts of the Brassey Hill Harzburgites. As illustrated in Figures 11, 13 and 16, the northern and southern sections differ in the relative abundance of rock types, but each shows a similar differentiation range according to their Mg/Mg + Fe contents. But are these sections lateral equivalents of each other, having crystallized from a single magma chamber, or did they crystallize from separate magmas? The former hypothesis is preferred, as the more differentiated rocks (plagioclase lherzolites) are similar in both cases, and the differences in other rock types are largely a function of which is the more abundant postcumulus phase, plagioclase or pyroxene.

3. Caudrys Hill Pyroxenites

3.1 Pyroxenites and olivine pyroxenites, lower zone of Caudrys Hill Pyroxenites

Most of these rocks are coarse-grained, but average grain sizes fall in the range 2-20 mm. They are granular, and consist dominantly of orthopyroxenes (anhedral, but subhedral in some samples), with 0-20% olivine, 0-2% interstitial clinopyroxene, and approximately 1% accessory chromite. Plagioclase is generally absent, but close to the faulted contacts with the Brassey Hill Harzburgites, some samples of plagioclase pyroxenites, poikilitic harzburgites and one possible wehrlite (38542) were found.

Specimens 38537 and 38538, collected from two adjacent layers from the outcrop in Plate 12, are typical of these rocks. The former is an orthopyroxenite containing a little interstitial clinopyroxene, less than 1% chromite (including several larger anhedral grains up to 2 mm in size), but no olivine. Specimen 38538 contains 2% chromite and about 20% olivine which occurs largely as grains with curved boundaries which are smaller than, and tend to be interstitial to, the orthopyroxenes (Plate 30).

Specimen 38539 is another olivine pyroxenite similar to 38538, while 38533 and 38534 are olivine-free pyroxenites. In all these rocks, the orthopyroxenes are commonly anhedral to subhedral stubby prisms which commonly form mutual interference and triple-point grain boundaries where interstitial clinopyroxene or olivine are absent (Plates 31 and 32). Such textures are commonly developed in rocks believed to be cumulates (see Jackson, 1961, especially Fig. 62; Wager and Brown, 1968). In the extreme case where no interstitial minerals are present, Hess (1960) postulates that continuous diffusion between magma interstitial to the settled crystals and the immediately overlying body of magma allowed the cumulus grains to continue growing at the expense of other possible post-cumulus phases. However, similar textures in metamorphic rocks are believed to have formed by deformation followed by annealing (Spry, 1969; Vernon, 1970), so cannot be considered as being diagnostic of a cumulative origin for ultramafic rocks.

Specimen 38545, which contains interstitial plagioclase, is very similar to pyroxene cumulates in which some secondary enlargement has occurred (see, for example, Fig. 61 in Jackson, 1961).

3.2 Medium-grained pyroxenites and plagioclase pyroxenites

These rocks consist essentially of orthopyroxene and accessory chromite, with varying amounts of plagioclase and clinopyroxene. Some contain scattered large olivine grains. The grainsize generally averages 1-2 mm, but a range of 0.3-8 mm has been recorded. In some specimens, for example 38622, the grainsize is quite variable, some patches averaging 0.3 mm while others average 1.5 mm.

Many specimens (e.g. 38633) show a distinct planar lamination due to alignment of platy orthopyroxenes parallel or subparallel to the layering. In other rocks, especially coarser-grained varieties, the orientation is apparently random (no petrofabric studies were done).

Orthopyroxenes vary in shape from stubby laths to thin elongate plates. They probably fall in the range En82-86. The amount of orthopyroxene in most of the rocks ranges from 50% (e.g. 38613) to 95% (e.g. 38633). Some rocks (e.g. 38637) contain scattered phenocrysts of orthopyroxene.

Clinopyroxene is ubiquitous, ranging from 1-15%. It occurs interstitial to the orthopyroxenes, commonly as thin rims which poikilitically enclose numerous orthopyroxenes. Some specimens (e.g. 38633) contain scattered large anhedral clinopyroxene grains, up to 3 mm in size. These poikilitically enclose orthopyroxenes at their margins, so it is difficult to decide if they are in part cumulus or entirely postcumulus. In some rocks, such as 38613, the anhedral clinopyroxene grains are not interstitial and are probably in part cumulus.

Plagioclase is present in most rocks, in amounts up to 50%, and is commonly partly altered to a cloudy saussurite. Compositions in the range An51-64 were determined from albite twins. Its distribution is quite variable, usually occurring as diffuse patches which may or may not be subparallel to the main layering. These patches are composed of numerous grains which on the whole are interstitial to or surround the orthopyroxenes (Plate 34). Usually plagioclase has crystallized after clinopyroxene (e.g. Plate 35). In some rocks (e.g. 38621), lenticular grains of plagioclase up to several centimetres long poikilitically include 15 or more small platy or lath-shaped orthopyroxenes. In some rocks (e.g. 38613 and perhaps 38888) plagioclase may be a cumulus as well as a postcumulus phase.

Several pyroxenites and plagioclase pyroxenites (38607, 38631, 38632) contain scattered large anhedral grains of olivine, up to 5 mm in size. These may be quite irregular in shape (Plate 37). Perhaps they are large cumulus grains.

Specimen 38610 is from one of the rare peridotites interlayered with the pyroxenites and plagioclase pyroxenites. It consists mainly of subhedral or euhedral olivines poikilitically surrounded by orthopyroxene and clinopyroxene, with accessory chromite and a little interstitial plagioclase.

In these rocks, plagioclase is commonly partly altered to saussurite (dark cloudy patches of indeterminate Ca-Al silicates). Orthopyroxene may be partly replaced along grain boundaries and cracks by fibrous tremolite-actinolite (uralite) and/or talc. Clinopyroxene in many sections is partly or entirely pseudomorphed by light green hornblende (Plate 35); this is thought to be a later hydrothermal replacement rather than part of the original magmatic processes. Olivines are partly replaced by veins of serpentine and dust-like magnetite.

With increasing alteration, the pyroxenites and plagioclase pyroxenites are converted to rocks composed of uralite, saussurite, talc, serpentine, and magnetite dust. Most of the small enclaves of pyroxenites within the volcanics are completely altered. The original textures are generally well preserved (38635, 38638, 38934) but in some cases (e.g. 38636) they have been obliterated despite the fact that the rock is undeformed. In many samples, the plagioclase is replaced by chlorite rather than saussurite.

The textures of the upper part of the Caudrys Hill Pyroxenites show a range from polygonal types with well-developed triple points to interstitial and poikilitic types where plagioclase and clinopyroxene have crystallized later than the orthopyroxenes.

3.3 Deformation of the Caudrys Hill Pyroxenites

Rocks from the lower part of the Caudrys Hill Pyroxenites have suffered deformation in localized areas, particularly adjacent to faults. This deformation has been superimposed on the "primary" textures and layering described above, and is thought to be a result of the faulting and/or an axial surface cleavage. The expression of the deformation in outcrop is the development of a fracture or spaced cleavage, the best example of which is shown in Plate 18.

Specimens 38535, 38536 and 38543 are pyroxenites and olivine pyroxenites that have suffered protoclasis. Granulation has occurred along the grain boundaries and some cracks in the original orthopyroxene, producing mortar texture (in the sense of Spry, 1969). The finer-grained orthopyroxenes surrounding the larger remnant grains are only partly annealed in 38536 and 38543 (Plate 38), while in 38535 they have formed a better-developed polygonal texture and are probably annealed to a greater extent.

Specimen 38541 has a rarer well-developed polygonal texture (Plate 39) which differs from the previously described textures with abundant triple-points in that a far greater proportion of the grain boundaries are straight. The rock also has the distinction of being the only rock in which exsolution lamellae of clinopyroxene are absent in the orthopyroxenes. The texture of this rock is interpreted as being due to annealing after it had been deformed. The kink bands present in some grains may be relicts of the deformation.

3.4 Summary

The Caudrys Hill Pyroxenites are orthopyroxene-dominated ultramafic rocks. In the lower part of the sequence, they are mainly pyroxenites and olivine pyroxenites in which olivine and minor clinopyroxene have crystallized interstitially. The upper part of the sequence consists of medium- to fine-grained pyroxenites and plagioclase pyroxenites in which clinopyroxene and plagioclase have largely crystallized interstitially to subhedral orthopyroxenes. Many of these rocks show planar lamination of orthopyroxenes. The orthopyroxenes have developed mutual interference or polygonal textures where in contact with each other. Protoclastic deformation of some of the pyroxenites and olivine pyroxenites occurred after the layering and the above textures had formed.

The textures in the Caudrys Hill Pyroxenites are consistent with a cumulative origin, the only unusual textures being the interstitial olivines in many of the olivine pyroxenites, which the writer has not seen described elsewhere. Other possibilities are they they are annealed tectonites or the result of multiple intrusions of ultramafics. The last two hypotheses are considered the least likely since no isoclinal or intrafolial folds, or no cross-cutting relationships between ultramafic rock types, have been observed. These points are discussed in more detail in a later section.

4. Nineteen Mile Creek Dunites

These are largely serpentinized dunites with some pyroxene dunites and harzburgites. In many localities orthopyroxenites are interlayered with dunites. Most specimens collected are deformed.

4.1 Dunites and pyroxene dunites

These are replaced by a serpentinite with a fine mesh texture. Brucite and fine-grained magnetite are also present, and in some samples talc occurs as scattered flakes concentrated in bands or rimming serpentinized orthopyroxene augen. Minute residual cores of olivine are usually scattered throughout the serpentine mesh, and the shapes and sizes of the original olivine grains are commonly defined by simultaneous extinction of these cores in patches throughout the thin section.

Specimens 38522, 38523 and 38524 are typical serpentinized dunites and pyroxene dunites. The original olivines were quite elongate (up to several centimetres in length) and show a well-defined foliation. No petrofabric studies were done in order to determine the degree of lattice preferred orientation. Scattered chromite grains occur throughout the slide. Specimen 38524 also contains scattered augen of serpentinized orthopyroxene (Plate 40). In 38526, the section is cut almost parallel to the foliation plane and shows that the olivines do not define a dimensional lineation. A single large deformed chromite grain occurs in the latter thin section.

4.2 Harzburgites

Specimens 38527 and 38528 are composed of strained anhedral orthopyroxenes surrounded by a finer-grained granoblastic mozaic of anhedral olivines (Plates 41 and 42). Kink bands are common in the olivines (Plate 43). Accessory anhedral chromite is scattered throughout the rocks.

Sections 38933a, b, c were cut from a deformed ultramafic rock consisting of interlayered pyroxenites, harzburgites and dunites. The olivines form a granoblastic mozaic, whereas most of the orthopyroxenes are thin, long, lenticular "strips" which are essentially parallel to the layering, which in this case is almost certainly tectonic in origin. Some of the orthopyroxene augen have been necked or boudinaged (Plate 44). It is thought that this rock has suffered plastic deformation, perhaps at a higher temperature than when protoclastic deformation of pyroxenites occurs.

4.3 Pyroxenites

Specimens 38521 and 38530 are coarse pyroxenites in which the anhedral orthopyroxenes are strained and kinked. Finer-grained orthopyroxenes produced during protoclasis surround some of the larger grains and also along the kink surfaces in some grains (Plate 45).

Specimen 38531 is a granoblastic pyroxenite in which a polygonal texture (i.e. triple-point junctions and straight boundaries) is poorly developed. Some of the grains are undulose and kinked. Elongate undulose orthopyroxene augen and chromite schlieren provide the best evidence that this rock has been deformed, the present granoblastic texture probably being due to partial annealing. Specimen 38525 has a similar granoblastic/xenoblastic texture as in 38531, but few grains are undulose and scattered deformed chromite grains provide the only evidence to suggest that the rock may have suffered deformation.

4.4 Chromitite

A chromitite sample (38905), collected from a loose boulder in a creek, probably came from one of the lenses or layers in the dunites (Reid, 1921). The rock consists mainly of euhedral chromite grains enclosed in pyroxenes and olivines replaced by serpentinite (Plate 46). In places, pseudomorphed orthopyroxenes poikilitically enclose the chromites. The only evidence of deformation are spaced cleavages along which the chromites have suffered cataclasis, probably prior to serpentinization (Plate 47). The texture of this rock is very similar to chromitites of cumulative origin (for example, see Jackson, 1961).

4.5 Origin of the textures

With the exception of the chromitite, the textures in the other rock types of the Nineteen Mile Creek Dunites are all thought to be the result of deformation. The distinct foliation due to the preferred orientation of olivines in the dunites is probably due to a relatively high temperature plastic deformation. Fabric studies of naturally and artificially deformed peridotites are discussed by a number of authors, including Raleigh (1967), Carter (1970), Ave'Lallemant and Carter (1970), and Lappin (1971). However, because of the serpentinization, it would be extremely difficult to compare the fabrics in the Nineteen Mile Creek Dunites with the patterns obtained by other workers.

The textures of the harzburgites are similar to textures produced in the transitional stage during the tectonic emplacement of the Twin Sisters Dunite (Ragan, 1963, 1967). The rocks have obviously been deformed, but a granoblastic mozaic of recrystallized olivines has been produced rather than an obvious preferred orientation, which possibly forms during deformation at higher temperatures. The pyroxenite layers in 38933 developed a strong dimensional preferred orientation and necking rather than protoclasis, apparently under similar conditions as the formation of the granoblastic mozaic of recrystallized olivines. The textures in the other pyroxenites may be the result of deformation followed in most cases by partial annealing.

Associated with such plastic deformation, one would expect to find intrafolial folds or isoclinally-folded layering, such as the examples described in Thayer (1963), O'Hara (1965), Ragan (1967) and Brown (1972). Where the dunites are massive and no other rock types occur, lack of such structures is not surprising. However, where dunites, pyroxenites and harzburgites are interlayered on a fine scale, no folding or cross-cutting relationships were observed. Undoubtedly at least some of the layering observed in the Nineteen Mile Creek Dunites is tectonic in origin (due to metamorphic differentiation or perhaps transposition), but perhaps some could be relict cumulate layering. Some of the larger pyroxenite bodies may be tectonic fish or boudins surrounded by plastically deformed dunites.

The presence of cumulate-like textures in the chromitite specimen is significant, for it suggests that at least some of the Nineteen Mile Creek Dunites could be deformed cumulates. The chromitite possibly acted as a competent block or layer. With the exception of one dunite, the compositions of the rocks and minerals (Fo88-89, En88-89) fall in the range of cumulates reported in Jackson (1967) and Wager and Brown (1968). However, they also fall in the range of alpine ultramafics believed to be of residual upper mantle origin (see Chapter 5).

5. Undifferentiated ultramafics

Specimens 38546-38552 are serpentinites from an area NE of Caudrys Hill marked as "undifferentiated" in Figure 11. Many are weathered and most have suffered post-serpentinization deformation, so that deducing the composition of the primary ultramafic is very difficult. Specimen 38546 is the least deformed, and the presence of some bastite pseudomorphs suggests that it is a serpentinized harzburgite.

Specimen 38596 was collected from an inlier of medium-grained pyroxenites and lherzolites occurring within dolerites about 2 km due east of Burgess Hill. It is composed of olivines poikilitically surrounded by plagioclase (altered to saussurite and white mica), orthopyroxene and clinopyroxene. Where surrounded by plagioclase the olivines are subhedral to euhedral plates (Plate 48), but the olivine grains included in the pyroxenes are smaller and more rounded. This rock is similar to the rare lherzolites occurring in the upper zone of the Caudrys Hill Pyroxenites and its texture could be interpreted as a classic cumulative type.

Specimens 38683 and 38685 were collected from small serpentinite bodies occurring within the volcanics of the Heazlewood River Complex SE of Whyte Hill (Fig. 4). The former is a coarse-grained peridotite largely replaced by serpentine and talc, and has been deformed after alteration. Specimen 38685 is an altered plagioclase dunite, with euhedral olivines replaced by mesh serpentine and the interstitial plagioclase by a fine-grained chlorite (?). The latter rock is extremely similar to the plagioclase dunites of the Brassey Hill Harzburgites and raises the possibility of layered ultramafics underlying large parts of the volcanics of the Heazlewood River Complex.

Specimen 38653, collected from a quarry in the town of Luina, is a plagioclase pyroxenite very similar to those in the upper zone of the Caudrys Hill Pyroxenites. This rock comes from the Whyte River Complex, which is thought to be a fault slice of rock types similar to the Heazlewood River Complex. Both of these complexes originally may have formed part of the same ophiolite.

INTRUSIVE GABBROS AND MICROGABBROS

Dykes of pegmatitic gabbros and coarse-grained gabbros are very common throughout the Brassey Hill Harzburgites, and are particularly abundant on Brassey Hill and extending to the NW. They are generally less than a metre thick, and are quite irregular in shape and size; they curve, form sharp bends, occur as complex interconnections (Plate 49), and show abrupt thickness variations and terminations. Very few can be traced for more than 10 m. Those that have intruded poikilitic harzburgites and plagioclase dunites have altered to rodingites, probably during serpentinization of the host ultramafic.

Medium-grained gabbros tend to form straight tabular dykes, generally thinner than 2 m. These also could not be traced for more than 10 m. There is a gradation from the medium-grained gabbros into microgabbros, which occur in the same areas as, and are subparallel to, sheeted dolerite dykes intruding ultramafics. The medium-grained gabbros and microgabbros are restricted to the southern half of the Brassey Hill Harzburgites and to the upper zone of the Caudrys Hill Pyroxenites. A single exception is a microgabbro dyke intruding pyroxenites on the SE side of Caudrys Hill.

As is discussed below, all the gabbro types appear to grade into each other. In some places this can be observed in outcrop; for example, Plate 50 shows a "microgabbro" dyke with irregular patches of typical pegmatitic gabbro.

Gabbro and irregular pegmatitic gabbro intrusives are abundant in ultramafic-gabbro and ophiolite complexes elsewhere (for example, Thayer, 1967; Thayer and Himmelberg, 1968; Moores, 1969). Similar rocks, together with microgabbro dykes, intrude harzburgites on Macquarie Island (Varne and Rubenach, 1972).

Petrography of the intrusive gabbroic rocks

The intrusive gabbros, pegmatitic gabbros and microgabbros are essentially composed of plagioclase and clinopyroxene. Orthopyroxene (bronzite to hypersthene) occurs in most medium- to coarse-grained varieties, but is rarer in the microgabbros. Plagioclase compositions determined from albite twins fall in the range An60-68 for coarser-grained rocks, while unaltered microgabbros have compositions in the range An50-60. A very small amount of accessory spinel (probably magnetite) occurs in most of these rocks, while large irregular grains of magnetite occur in some of the coarser-grained rocks such as 38644 (Plate 51).

Specimens 38645-7 are typical pegmatitic gabbros composed of orthopyroxene, clinopyroxene and plagioclase, with grain sizes varying from 1-10 mm. These are altered in that the plagioclase is partly saussuritized and the pyroxenes partly replaced by amphiboles.

Specimens 38639, 38640 and 38642 are typical medium- to coarse-grained gabbros with a more uniform grainsize. Most of the grains are anhedral, but in 38639 the plagioclase-rich patches form a polygonal mozaic (Plate 52). Some inclusions of plagioclase occur in pyroxenes of 38639, which on texture alone could easily be mistaken for a cumulate if its field relationships were not known (compare Plate 52 with Fig. 154 of Wager and Brown, 1968).

Texturally, 38608 is a transition between the typical medium-grained gabbros and the microgabbros such as 38705 and 38891. Microgabbros commonly have a uniform grainsize of around 0.5 mm and are composed essentially of augite and labradorite with a little iron oxide. More rarely they contain a little hypersthene. The augite grains are anhedral and commonly define a foliation parallel to the dyke walls. The overall texture (Plate 53) is "granoblastic", similar to many coarser-grained gabbros, but also very similar to some granulites. It is suggested that these "granoblastic" textures in the gabbros and microgabbros formed instead of subophitic textures because the magmas were injected into layered ultramafics that were still quite hot. So in cooling more slowly than a normal dolerite dyke, granoblastic textures resulted from subsolidus adjustment of grain boundaries, as would occur during the formation of a basic granulite or hornfels (Spry, 1969).

Specimen 38897 (Plate 54) is an unusual microgabbro containing a thin lens and a "nodule" of coarser-grained augite which show a polygonal granoblastic texture (i.e. similar to some cumulate pyroxenites).

Specimen 38917 is from a block of massive chromitite occurring in a pegmatitic gabbro dyke at 3385E/8958N. The texture of the chromitite is shown in Plate 55. The chromite grains, some of which appear to have been resorbed, are surrounded by hydrogarnet and saussurite replacing plagioclase and tremolite-actinolite replacing pyroxenes. A partial analysis of the hydrogarnet, which is light green in colour, yielded about 5% Cr_2O_3, suggesting that it may be a hydro-uvarovite/hydrogrossular mixture instead of the usual hydrogrossular replacing plagioclase in rodingite. (Massive chromitite was also found by the writer in a pegmatitic gabbro dyke intruding harzburgites on Macquarie Island.)

Most of these gabbros and microgabbros are altered. Usually, the plagioclase is replaced by cloudy saussurite, fine-grained mica, or more rarely, chlorite. The pyroxenes have been partly to entirely replaced by amphiboles; orthopyroxenes are usually replaced by uralitic tremolite-actinolite, while the clinopyroxenes may be replaced by uralite or pseudomorphed by single crystals of light green or light brown hornblende. Some of the altered rocks in which the textures are preserved include 38609, 38657, 38673, 38892 and 38898. Specimens 38652 and 38662 consist of a mess of random tremolite-actinolite grains and interstitial saussurite, the original igneous textures being entirely destroyed. Specimens 38675, 38680 and 38681 are altered gabbros containing abundant anhedral quartz which is probably of metasomatic origin.

As mentioned above, gabbroic dykes in highly serpentinized ultramafics have commonly been replaced by rodingite. This process almost certainly occurs during serpentinization, and commonly involves Ca and Al metasomatism (Bilgrami and Howie, 1960; Coleman, 1966, 1967). Specimen 38669 is a rodingitized gabbro in which the plagioclase has been replaced by fine-grained hydrogrossular, the orthopyroxenes partly replaced by uralite and chlorite, while the clinopyroxene and magnetite have suffered slight alteration. The texture of the original rock is very well preserved. Specimen 38661 is a rodingite in which the pyroxenes are relatively more altered to uralite, but in which the igneous texture is still fairly well preserved. A more complete discussion of the origin of the rodingites is given in another chapter.

Specimens 38651 and 38671 are altered gabbros consisting of uralitic amphiboles, saussurite, chlorite, and patches and veins of prehnite.

ORIGIN OF THE TEXTURES AND LAYERING IN THE ULTRAMAFICS AND GABBROS

1. Cumulate textures of the Brassey Hill Harzburgites and Caudrys Hill Pyroxenites

As has been indicated, the textures of the undeformed layered ultramafics are very similar to those developed in complexes believed to be of cumulative origin, in particular to the ultramafic zones of the Stillwater and Bushveld complexes and the Great Dyke (Hess, 1960; Jackson, 1961; Wager and Brown, 1968). In accordance with this hypothesis, most layers in the Heazlewood River Complex can be classed as orthopyroxene and/or olivine cumulates. Accessory chromite is ubiquitous and is probably cumulus in most or all layers. Where the cumulus grains continued to grow after settling, presumably by continuous diffusion between the interstitial magma and the immediately overlying main body of magma, granular orthopyroxenites, dunites or harzburgites with mutual interference textures were formed. If such diffusion was incomplete or did not take place, the interstices between the settled crystals were filled by varying proportions of plagioclase, orthopyroxene, clinopyroxene, and olivine (?) with little or no secondary enlargement of the cumulus grains. Such interstitial minerals may form isolated grains or may be optically continuous over patches, so forming oikocrysts enclosing a number of cumulus grains. The minerals forming oikocrysts are orthopyroxene and clinopyroxene (in poikilitic harzburgites and lherzolites) and to a less extent plagioclase (in some plagioclase pyroxenites). The interstitial magma may react with cumulus grains, so that the oikocryst grows at the expense of the cumulus grains. This commonly occurred during the growth of orthopyroxene oikocrysts in poikilitic harzburgites. Reaction with the cumulus grains during the formation of clinopyroxene oikocrysts was generally more limited, but more extensive reaction has occurred in some layers of lherzolite and pyroxenite.

In comparison with other cumulates, the only unusual rocks are the olivine pyroxenites from the lower part of the Caudrys Hill Pyroxenites. In accordance with the cumulate hypothesis, the olivine in these rocks would be interpreted as an interstitial postcumulus phase The writer does not know of the occurrence of similar rocks from other cumulates. The plagioclase dunites and troctolites, which are abundant in the northern part of the Brassey Hill Harzburgites, are not very common in the ultramafic zones of the Bushveld and Stillwater complexes and the Great Dyke. However, the writer observed abundant plagioclase dunites interlayered with gabbros and troctolites in the layered gabbros on Macquarie Island.(Varne and Rubenach, 1972).

It has been argued in a previous section that some of the deformed Nineteen Mile Creek Dunites could have originally been cumulates, mainly on the basis of a single specimen of chromitite. The composition of these rocks also does not preclude a cumulative origin. However, this hypothesis must be regarded as quite tentative for the positive evidence for a cumulative origin is very slight.

2. Multiple intrusion and tectonite hypotheses

On textural grounds, there is no reason why the Brassey Hill Harzburgites and the Caudrys Hill Pyroxenites cannot be cumulates. However, it has been shown that textures of ultramafic and gabbroic dykes, and also of rocks which have been annealed following deformation, can be fairly similar to certain cumulate textures, namely the granular or polygonal textures and interstitial textures. The poikilitic textures would be more difficult to explain: they could perhaps form by slow cooling of a magma or crystal mush lubricated by magma, but it is more difficult to explain how poikilitic textures could form during annealing of a tectonite.

If most of the layering in the Caudrys Hill Pyroxenites and Brassey Hill Harzburgites is the result of multiple injections of magmas many more cross-cutting relationships would be expected (for example, as occurs in the dolerite dyke swarms on Macquarie Island - Varne and Rubenach, 1972). Also the hypothesis of abundant multiple intrusions is far more complex than the cumulative hypothesis for the variety of rock types and their bulk compositions would be even more difficult to explain.

If part or all of the Caudrys Hill Pyroxenites and Brassey Hill Harzburgites are annealed tectonites, the layering would have to be explained by metamorphic differentiation, involving mechanical separation of grains into layers and/or diffusion. However, as has been already discussed, isoclinal or intrafolial folds, together with cross-cutting relationships, would be expected if the rocks are tectonites, and these are not observed. Also, some relict deformation textures would be expected to be preserved, but the only deformation textures observed are cataclastic rather than "plastic", and formed later than the textures under discussion.

It is very difficult to explain the observed scour-and-fill structure and the outcrop showing truncated cross bedding according to the tectonite hypothesis, but these structures are easily explained in terms of the cumulate hypothesis.

From all this textural and structural evidence, it is concluded that the cumulate hypothesis best explains the observed data. In the next chapter it is argued that the chemical data are also consistent with crystal accumulation.

3. The convergence of textures of cumulates, dykes and high grade metamorphics*

Given that certain textures developed in cumulates, gabbro dykes, ultramafic dykes, and high-grade regional or thermal metamorphics are similar, what is the explanation of this convergence? The common factor in all cases is that they all cooled slowly through the moderate or high temperature range. In metamorphic rocks, textures are believed to tend towards the lowest total surface energy by grain boundary migration (Spry, 1969). The polygonal granoblastic textures (i.e. straight grain boundaries and abundant triple-points) in many regional and thermal metamorphic rocks are thought to reflect such a low surface energy state. There is no reason why similar textures cannot form either as crystals slowly grow from a magma or by subsolidus grain boundary migration during slow cooling.

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* After preparation of this section, it was discovered that many of the ideas expressed here are discussed in detail in Vernon (1970).

It is suggested then that some, and perhaps all, the cumulate textures satisfy the lowest total surface energy requirement. It is not known whether the present texture of a rock formed in the final stages of magmatic crystallization, or is small or major textural adjustments occurred by subsolidus grain boundary migration, as is suggested by Cameron (1969) for some plagioclase cumulates.

Whether a particular grain forms rational (i.e. is euhedral) or irrational grain boundaries in contact with another mineral can also be linked to the surface emergy requirements (Spry, 1969). Similar explanations may be applied to such observations as (1) olivine inclusions in orthopyroxene iokocrysts are invariably rounded; (2) contacts between olivines and orthopyroxenes in granular harzburgites and olivine pyroxenites are usually curved; (3) olivines surrounded by plagioclase are commonly euhedral; (4) where in mutual contact, plagioclase grains in both cumulates and gabbros commonly form polygonal textures in which triple-point junctions and straight boundaries are abundant.

To explain why the gabbro and microgabbro dykes in the Heazlewood River Complex have granoblastic textures similar to basic granulites and hornfelses, it is suggested they intruded the layered rocks while the latter were still hot. As previously described, coarser-grained gabbros and pegmatitic gabbros form very irregular dykes, so perhaps they were emplaced while the ultramafics were still relatively plastic. It is possible that some of these formed by the channelling of interstitial magma trapped between cumulus grains, perhaps by a filter press mechanism as described in Hess (1960). The medium-grained gabbros and microgabbros tend to form tabular dykes, so when these were emplaced the ultramafics had cooled sufficiently for brittle fracture to take place. It is also interesting that the microgabbro dykes intruding ultramafics occur in the same areas and are subparallel to sheeted dolerite dykes. The dolerites have subophitic textures and lath-shaped grains rather than the granoblastic textures in the microgabbros, and perhaps intruded the ultramafics when they were cooler than when the microgabbros intruded.

The picture, then, is one of continuous intrusion of basic dykes into the layered ultramafics as they gradually cooled. In chapter 6, it is argued that to some degree there is a chemical as well as textural continuum in these rocks.

4. Origin of the layering in the ultramafics

The origin of layering in cumulates is discussed in detail by a number of authors, including Hess (1960), Jackson (1961, 1967), Hawkes (1967), and Wager and Brown (1968). Since the present writer did not do modal analyses or study sections of layered rocks in a systematic way, the layering in the Heazlewood River Complex is only briefly compared with that in other complexes and conclusions drawn must be regarded as tentative.

4.1 Comparison with layering in other layered intrusions

At first sight, the layering in the Brassey Hill Harzburgites and Caudrys Hill Pyroxenites would appear to be very similar to layering in the ultramafic zones of the large layered intrusions. This could be expected in view of the similarities in textures and most of the rock types. The southern part of the Brassey Hill Harzburgites, where the most common rocks are granular and poikilitic harzburgites, commonly with thinner layers of olivine pyroxenites and pyroxenites, would appear to be the most similar to the ultramafic zones of other layered intrusions. However, a comparison outlined below, using the nomenclature of Jackson (1967) reveals some differences.

As far as can be judged without detailed modal analyses, most of the layers in the Brassey Hill Harzburgites and Caudrys Hill Pyroxenites appear to be essentially isomodal with respect to the cumulus phases, as is the case for the large layered intrusions. Observed mineral-graded layers include a single olivine pyroxenite-granular harzburgite layer (see Fig. 14) and some layers of plagioclase dunite and troctolite.

The types of contacts between adjacent layers (see Jackson, 1967, pp. 22 and 35) are summarized below:

1) Phase contacts - horizons marked by the appearance or disappearance of a cumulus mineral. Many of the contacts in the Brassey Hill Harzburgites are of this nature, especially contacts between poikilitic harzburgites and granular harzburgites or olivine pyroxenites. Such contacts are quite rare in the Caudrys Hill Pyroxenites, being restricted to contacts between plagioclase pyroxenites and the rare lherzolite layers in the upper part of the sequence.

2) Ratio contacts - horizons marked by a sharp change in the proportion of two cumulus minerals. Again, such contacts are quite common in the Brassey Hill Harzburgites (especially contacts between granular harzburgites and olivine pyroxenites), but very rare in the Caudrys Hill Pyroxenites.

3) Form contacts - horizons marked by sharp changes in such properties as grainsize, shape and preferred orientation of cumulus grains. Many phase and ratio contacts are commonly, but not necessarily, also form contacts. Form contacts are abundant in the Caudrys Hill Pyroxenites, common examples including sharp changes in grainsize and/or sharp changes from layers composed of stubby prismatic grains of a more random orientation to layers composed of platy grains defining a distinct igneous lamination.

4) A type of layering contact, which is common in the Heazlewood River Complex but not mentioned in Jackson (1967) is a reflection of variations in the interstitial postcumulus minerals. Examples of this include most of the layering in the plagioclase dunites and some of the layering in the pyroxenites and olivine pyroxenites of the lower part of the Caudrys Hill Pyroxenites. "Sublayering" due to variations in interstitial plagioclase content within layers of plagioclase pyroxenites, especially in the upper part of the Caudrys Hill Pyroxenites sequence, is of this type. The contacts in this type of layering may be sharp but commonly are diffuse. There is every gradation from a well-defined "sublayering", which is parallel to other local layering types, into irregular patches of interstitial plagioclase within a larger plagioclase pyroxenite or plagioclase dunite layer.

According to Jackson (1967, p. 35), over 70% of the layer contacts in the ultramafic zones of the large layered intrusions are phase contacts between adjacent isomodal layers, each of which contains a single cumulus phase (excluding accessory chromite). Such layering also occurs in the

Heazlewood River Complex, but in contrast, phase contacts involving two cumulus minerals, ratio contacts, contacts exclusively of the form type, and layering defined solely on variations in the proportion of interstitial phases, are also fairly abundant.

Because of structural complications and lack of detailed studies by the writer, it is not known if cyclic units such as those present in the ultramafic zones of the large layered intrusions (Jackson, 1961, 1967; Wager and Brown, 1968) occur in the Heazlewood River Complex.

4.2 Origin of the layering

Discussions of the possible factors involved in the formation of the smaller scale layering in cumulates (as opposed to the larger scale cryptic variations) can be found in Hess (1960), Jackson (1961), Hawkes (1967) and Wager and Brown (1968).

As a first approximation, some of the main factors which may directly or indirectly produce layering are as follows:

1) Nucleation. This can also be linked with undercooling.

2) Hydrodynamic factors. The rate of settling of crystals in a magma can be expressed as $V = \frac{2gr^2}{9} \frac{dc-dm}{n}$ (Hess, 1960, p. 139), where V is the settling velocity, g the gravity constant, r the radius of the grains, dc the density of the crystals, dm the density of the magma, and n the viscosity of the magma. So in a magma chamber at any given time, the densities and sizes of grains should be the important factors in crystal settling rates. The presence of convection currents in a magma would provide an additional hydrodynamic factor.

Hess (1960) has shown that olivines and orthopyroxenes of similar Mg/Mg+Fe have nearly identical densities. Thus sorting of olivine and orthopyroxene grains of similar sizes would not be expected. The relative lack of mineral-graded layers in the dominantly orthopyroxene and olivine cumulates of the Heazlewood River Complex and the large layered intrusions is consistant with this observation.

Because of the dependence of settling rates on the square of the radii of the grains, size-graded layers would be expected to be common if the crystals fell through an appreciable column of magma before accumulating on the floor of the magma chamber. Unfortunately lack of time prevented detailed studies of grainsize variations in layers, such as those done by Jackson (1961) and Macdonald (1967). However, a large number of layers less than 1 m thick were closely examined in the field and in thin section. If size-grading occurs it is certainly quite subtle, as it is not at all obvious in thin section. It would be quite insignificant compared with very sharp grainsize changes across layer contacts. Jackson (1961, pp.35-37) discusses the lack of size-graded layers in the ultramafic zone of the Stillwater Complex, and on the basis of this and other arguments postulates that crystallization occurred relatively close to the bottom of the magma. The present writer believes that similar arguments could apply in the case of the Heazlewood River Complex.

The present writer also agrees with Jackson (1961) regarding the origin of planar lamination of platy grains. If grains grow with such a habit, it would seem likely that they would define a planar lamination simply by falling and coming to rest on top of other crystals; convection currents are therefore probably not necessary to explain planar lamination.

As in the case of the Stillwater Complex, there are some structures which can be explained by convection currents during crystal accumulation, but these are relatively rare in comparison with the Skaergaard Complex (Wager and Brown, 1968).

In view of the fact that the layering in the ultramafics cannot be satisfactorily explained in terms of hydrodynamic sorting processes, it is suggested that undercooling and nucleation could be important controls in producing phase, ratio and form contacts. It is proposed that after considerable undercooling of stagnant magma near the bottom of the chamber, nucleation, growth and settling of cumulus olivine and/or orthopyroxene occurred. The temperature of the magma at the bottom of the chamber would rise due to this crystallization (due to the "latent heat of crystallization"), so that crystallization would cease until there occurred sufficient undercooling for another nucleation burst.

In order to explain rhythmic layering, Hawkes (1967) suggests undercooling and nucleation may cause oscillatory variations in the composition of a magma. He illustrates his model with reference to the Di-Ab-An system, and shows how undercooling could produce oscillatory changes in the composition of the liquid about the cotectic. It is suggested that a similar mechanism could apply in the case of the Heazlewood River Complex, but in contrast to the formation of rhythmic layering in the Freetown Basic Complex as discussed by Hawkes, it is suggested that relatively short nucleation bursts near the bottom of the stagnant magma may have produced the fine-scale layering with isomodal phase and ratio contacts in the Heazlewood River Complex. The relevant cotectic in the case of the Heazlewood River Complex would be olivine-orthopyroxene in the system Mg_2SiO_4-$CaMgSi_2O_6$-SiO_2 of Kushiro (1964).

Hawkes (1967) also argues that the total number of crystals per unit volume of rock is a measure of the degree of undercooling. If this is so, then in the case of the Heazlewood River Complex, the average grainsize of a layer could reflect the degree of undercooling. Thus the sharp changes in grainsize which are commonly concomitant with ratio or phase contacts between adjacent layers, could perhaps be explained by cycles of undercooling, nucleation and sudden cessation of nucleation due to release of latent heat. The shapes of grains depend to some extent on nucleation and especially the rate of growth, so these factors may also explain the abrupt changes from layers comprised of platy grains showing planar lamination to layers composed of stubby prismatic grains of apparent random orientation.

Jackson (1961) suggests that intermittent variable depth convection could periodically replenish the magma from which crystallization is occurring at the bottom of the chamber.

Hess (1960) suggests that a relatively slow accumulation of grains would enable a more complete diffusion to take place between the magma interstitial to cumulus grains and the immediately overlying main body of magma, thus promoting overgrowth of cumulus grains rather than interstitial crystallization of other minerals. Conversely, a relatively rapid accumulation would tend to trap the interstitial magma so that interstitial and poikilitic grains of minerals other than the cumulus phases will crystallize. Wager and Brown (1968) point out that undercooling may also influence whether or not the cumulus grains continue to grow so as to completely fill the interstices (i.e. form adcumulates);

if the temperature of the interstitial magma is sufficiently raised by the growth of the cumulus grains then secondary enlargement of the cumulus grains would temporarily cease. By the time undercooling sufficient for continued crystallization of the interstitial magma has taken place, too many additional cumulus grains could have settled on top of the horizon under consideration so that the diffusion necessary for continued enlargement of cumulus grains in this horizon is impossible. So localized undercooling and diffusion of ions may go hand in hand in determining the relative importance of secondary enlargement of cumulus grains or interstitial growth of other minerals during the formation of any particular layer. Perhaps such processes also help in determining the relative abundance of the various postcumulus phases (usually orthopyroxene, clinopyroxene and plagioclase) in any particular layer.

The type of layering which is a function of variation in interstitial minerals may therefore be controlled by such processes as rate of settling of cumulus grains, diffusion between the interstitial magma and the immediately overlying main body of magma, together with temperature changes, undercooling, and nucleation processes acting within the interstitial magma. The diffuse nature of much of this type of layering would be expected from such processes, as would the irregular patches of interstitial plagioclase in many plagioclase pyroxenites and dunites. It is also possible that diffuse layers and irregular patches defined on interstitial plagioclase content could have formed by postcumulus diffusion entirely confined to the interstitial magma trapped between the cumulus grains.

4.3 Conclusions

It is concluded that the Caudrys Hill Pyroxenites and the Brassey Hill Harzburgites are essentially orthopyroxene and/or olivine cumulates which may have formed by crystallization at the bottom of the magma chamber. Cycles beginning with undercooling, followed by nucleation and growth which raised the temperature sufficiently to stop crystallization, may explain the observed layering in which the contacts are of the phase, ratio, and form types. Layering defined by interstitial minerals may depend on similar processes acting at or near the interface between the settled grains and the main body of magma, as well as diffusion processes acting between the interstitial magma and the overlying main body of magma and/or entirely within the magma trapped interstitially between the cumulus grains.

Chapter 4

THE PETROLOGY OF THE LAYERED ULTRAMAFICS AND GABBROS OF THE OTHER TASMANIAN COMPLEXES IN COMPARISON WITH THE HEAZLEWOOD RIVER COMPLEX

In this chapter, the rock types, layering, petrography and intrusive relationships of the primary ultramafics and gabbros in the other Tasmanian complexes are compared with their equivalents in the Heazlewood River Complex. In making such comparisons, it must be realized that these bodies have all been dismembered to varying degrees and in no case is a complete stratigraphic sequence represented.

1. The Serpentine Hill Complex

The petrology of the layered ultramafics, layered gabbros and intrusive gabbroic rocks in this complex is described in detail in the Appendix. The ultramafics are mainly orthopyroxenites, harzburgites, with rarer dunites, lherzolites and wehrlites. Such rock types as plagioclase dunites, plagioclase harzburgites, and plagioclase lherzolites are extremely rare or absent, and rocks similar to the Nineteen Mile Creek Dunites were not observed. The pyroxenites, harzburgites and dunites occur interlayered with each other, commonly on a relatively fine scale. Both poikilitic and granular harzburgites occur. Interstitial plagioclase occurs in some rocks, but it is not nearly as abundant as it is throughout most parts of the Brassey Hill Harzburgites.

As a generalization, the layered ultramafic rocks in the Serpentine Hill Complex resemble the Brassey Hill Harzburgites to the west and SW of Brassey Hill itself, where abundant pyroxenites occur interlayered with harzburgites. The textures are extremely similar to those in the equivalent rock types in the Brassey Hill Harzburgites. Many of the orthopyroxenite layers do not contain olivine, however, and are also similar to olivine-free layers in the Caudrys Hill Pyroxenites.

The medium- to fine-grained pyroxenites or plagioclase lherzolites, which typify the most differentiated rocks in the Heazlewood River Complex, are missing from the Serpentine Hill Complex. Instead, there is an apparent gradation from the layered ultramafics into layered two-pyroxene gabbros, through a transition zone consisting of interlayered pyroxenites, plagioclase pyroxenites, and two-pyroxene gabbros. Olivine is totally absent from both the layered gabbros and the transition rocks. The exact relationships between the three groups are obscured by swarms of intrusive microgabbros and faulting. There are no equivalents of the transition rocks and layered two-pyroxene gabbros in the Heazlewood River Complex.

The pegmatitic gabbro and microgabbro dykes are very similar in form and composition to their counterparts in the Heazlewood River Complex, except that the microgabbros are very abundant in parts of the Serpentine Hill Complex.

The textures in the layered ultramafics of the Serpentine Hill Complex are very similar to the equivalent rocks in the Heazlewood River Complex. Observed differences, such as the presence of some very coarse-grained pyroxenite layers in the former complex, are relatively minor. The layering is also very similar, with phase, ratio and form contacts all being fairly abundant, as in the Heazlewood River Complex. Layering defined entirely by the proportion of interstitial minerals is not as common in the Serpentine Hill Complex but has been observed within some coarse-grained pyroxenites which contain "sub-layers" with interstitial plagioclase. As described in the Appendix, the transition rocks have classic cumulate textures, while plagioclase, hypersthene and augite can all be interpreted as cumulus phases in the two-pyroxene layered gabbros. Layering in the latter is commonly on a fine scale, and planar lamination is fairly common.

Several residual lenses of interlayered pyroxenites and harzburgites occurring in the extreme north of the Dundas Serpentinite are very similar to those occurring in the Serpentine Hill Complex.

2. The Wilson River Complex

As previously mentioned, this body has suffered more than the Serpentine Hill Complex from dismembering and serpentinization. The writer looked in detail at well exposed layered ultramafics on Riley Knob, and also observed massive orthopyroxenites near the eastern contact just north of Christina Creek. Small pods of massive and layered pyroxenites occur in the Colebrook Hill area (Fig. 7).

At Riley Knob, the main rock types are orthopyroxenites, granular harzburgites and some dunites. Plagioclase pyroxenites and some norites (in which plagioclase is interstitial) also occur. The layering is on a fine scale, and the pyroxenites either alternate with granular harzburgites or occur as successive pyroxenite layers with form contacts (i.e. sharp changes in grainsize and grain shape, and in the degree of preferred orientation) (see Plates 58-60). Some pyroxenite layers show size-grading, being the only examples of this phenomenon observed in the Tasmanian complexes. Many of the finer-grained pyroxenite layers are composed of platy orthopyroxenes showing planar lamination. The rock types, layering style and textures developed in the layered ultramafics are very similar to some parts of the Serpentine Hill succession, and are quite consistant with a cumulative origin (see specimens 38860 and 38878).

Some gabbroic and ultramafic rocks occurring north of the Wilson River are described in Waterhouse (1914). One rock particularly well described is an olivine websterite in which anhedral clinopyroxene is interstitial to subhedral orthopyroxenes.

A considerable quantity of alluvial osmiridium was obtained from the Wilson River area (Reid, 1921). In the Heazlewood River Complex, the occurrence of this mineral is obviously associated with the Nineteen Mile Creek Dunites, and also occurs in association with serpentinized dunites in the Mt. Stewart Serpentinite. It is suggested then, that some of the highly serpentinized rocks in the Wilson River Complex may be similar to the Nineteen Mile Creek Dunites.

3. The McIvor Hill Gabbro

As previously mentioned, layering (Plate 3) and intrusive bodies of pegmatitic gabbros have been observed in some localities in the McIvor Hill Gabbro, which in many ways resembles the layered gabbros in the Serpentine Hill Complex. In thin section (for example, specimens 38935 and 38878), the gabbro consists of saussuritized plagioclase, orthopyroxene replaced by bastite or uralite, and clinopyroxene partly pseudomorphed by hornblende. The texture is consistent with, but not diagnostic of, a cumulative origin. The replacement of augite by hornblende in believed to have occurred during alteration, and the existence of primary hornblende, as reported in Waterhouse (1916) and Blissett (1962) could not be confirmed.

4. The Noddy Creek Complex

Relict cores of unserpentinized ultramafics are not very common in the Noddy Creek Complex. In the area examined by the writer (Fig. 8), most of the primary ultramafics are massive orthopyroxenites, but one kernal of interlayered pyroxenites, dunites and harzburgites occurs (Plate 61). The orthopyroxenites locally grade to plagioclase pyroxenites in which plagioclase and clinopyroxene are interstitial to the orthopyroxenes (Plate 62). Most of the gabbros shown in Figure 8 are very similar to those intruding ultramafics in the Heazlewood River Complex, but some (for example, specimen 38834) could be part of a layered ultramafic-gabbro sequence. Most are highly altered to saussurite and amphiboles.

5. <u>The Spero Bay Complex</u>

Partly serpentinized ultramafics consisting mainly of orthopyroxenites and lherzolites interlayered on a relatively fine scale (1.5-17 cm) are briefly described in Taylor (1955, p.57). It is not known whether the large gabbro body immediately adjacent to the ultramafics is composed of layered gabbros, intrusive gabbros or both.

6. <u>The Andersons Creek Complex</u>

The petrography of this body is discussed in detail in Green (1959) and Gee and Legge (*in press*) and only a brief summary is given here.

Gee and Legge (*in press*) interpret the primary rocks as layered ultramafics passing up into layered gabbros, and intruded by various gabbros and hornblende gabbros. The ultramafics have been serpentinized and the only unaltered cores remaining are orthopyroxenites. However, Green (1959, p.15) argues that at least some of the partly or entirely serpentinized rocks are peridotites. Gee and Legge (*in press*) observed gabbros interlayered with pyroxenites on a scale of metres. Some pyroxenites have interstitial feldspar. The layered gabbros are mainly two-pyroxene gabbros with some norites, and are laminated on a scale of millimetres to centimetres, adjacent layers differing in feldspar content. No textures typical of tectonites have been observed, although many of the ultramafics have suffered post-serpentinization deformation. Some of the thin sections of Gee and Legge (*in press*) and Green (1959) were examined by the present writer; the rocks are very similar to the pyroxenites and layered gabbros in the Serpentine Hill Complex, and are quite consistant with a cumulative origin. The layering is very similar to that in the Serpentine Hill Complex, and as is discussed in the following chapter, the rocks of both complexes are very similar in chemistry.

Intrusive gabbros in the Andersons Creek Complex include pegmatitic gabbros, and also some hornblende gabbros which were not observed by the writer in other complexes. Many of these have altered to rodingites where they are in contact with serpentinites (Green, 1958; Baker, 1959; Gee and Legge, *in press*); this alteration is discussed in a later chapter. Albitite and quartz-albite dykes are also fairly common. Gee and Legge (*in press*) argue fairly convincingly that these are intrusive bodies that were emplaced before serpentinization. Such dykes do not appear to be common in other complexes, although the present writer found one in the Serpentine Hill Complex (see the albite-prehnite rock in the Appendix) and several highly altered quartz-albite rocks in both the Heazlewood River and Noddy Creek bodies.

7. The Adamsfield Complex

The Adamsfield Complex is described in detail in Brown (1972), and the discussion below is mainly a summary of his conclusions, with which the present writer is largely in agreement.

The Adamsfield Complex consists entirely of ultramafics, none of which are plagioclase-bearing. The primary ultramafics include layered series of orthopyroxenites, olivine pyroxenites and dunites interlayered on a fine scale. Brown (1972) indicates that these rocks have textures consistent with (but not diagnostic of) a cumulative origin, and their chemical compositions fall in the range of cumulates from the large layered intrusions. The rest of the ultramafics are tectonites, which include plastically deformed massive dunites and protoclastically deformed massive orthopyroxenites. Between the massive and layered ultramafics is a complex deformation zone of dunites, pyroxenites and mylonitized pyroxenites. Irregular layering, boudins of pyroxenite dunite, and isoclinal and intrafolial folds (several phases in places) are typical of this zone. The folds have plastically deformed the cataclasites and mylonites which presumably must have been produced by an earlier deformation.

Brown (1972) postulates that all the primary ultramafics may have formed part of a layered complex of cumulative origin, and that most of the deformation probably occurred during tectonic re-emplacement. In the earlier stages the hot ultramafics, in particular the dunites, deformed plastically but as the rocks cooled down brittle or cataclastic deformation occurred. Serpentinization was later still and facilitated tectonic re-emplacements to still higher levels. In the essential details, the tectonic re-emplacement of the Adamsfield Complex is believed by Brown (1972) to be analogous to that of the Twin Sisters Dunite (Ragan, 1967).

The layered ultramafics at Adamsfield are similar to some of the rocks in the Heazlewood River Complex, and in particular to the finely interlayered pyroxenites and dunites exposed at Riley Knob in the Wilson River Complex (cf. Plate 58). However, mylonites, isoclinal and intrafolial folds, boudins, and irregular transecting relationships between ultramafics have not been observed in the Main Belt complexes. The textures in the deformed Nineteen Mile Creek Dunites are similar to those of the massive dunites and pyroxenites of the Adamsfield Complex, suggesting that the deformation perhaps occurred under similar conditions. It is also significant that the osmiridium deposits are associated with both the Nineteen Mile Creek Dunites and the Adamsfield Complex.

8. Discussion and conclusions

In making comparisons between the Tasmanian complexes it must be remembered that all of them are dismembered to varying extents during tectonic re-emplacements and later faulting. The fact that orthopyroxenites and gabbroic rocks are the most abundant primary rocks in the more serpentinized bodies as Noddy Creek, Wilson River and Andersons Creek may be simply because these rock types are obviously the most resistant to serpentinization. Bearing these facts in mind, the following generalizations can be made:

1) With the exception of the Nineteen Mile Creek Dunites, the textures and layering in the primary ultramafics and layered gabbros in the Main Belt complexes are consistent with these rocks being formed by crystal accumulation from magmas. They can be classed as olivine and/or orthopyroxene cumulates*, with clinopyroxene and plagioclase being additional cumulus phases in the layered gabbroic rocks.

2) Probable gradations from layered ultramafics upwards into layered two-pyroxene gabbros occur in the Serpentine Hill and Andersons Creek bodies. It is not known whether or not such layered gabbro sequences

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* Accessory chromite present in most rock types is probably an additional cumulus phase.

originally formed part of such complexes as Heazlewood River, or whether layered ultramafics originally formed part of the McIvor Hill Gabbro. Similarly, it is impossible to tell if rock types peculiar to one complex (for example, the plagioclase dunites of the Heazlewood River Complex) were present in other complexes before serpentinization, tectonic re-emplacement and faulting.

3) Intrusive gabbroic rocks, in particular pegmatitic and coarser-grained gabbros, occur in all the large Main Belt complexes. Microgabbros have been observed only in the Serpentine Hill and Heazlewood River Complexes.

4) The Adamsfield Complex differs from those in the Main Belt in that no plagioclase-bearing rocks are present. A series of layered ultramafics is, however, similar in textures and layering characteristics to some outcrops in the Main Belt complexes. Deformation textures in massive dunites and pyroxenites in the Adamsfield Complex are similar to those in the Nineteen Mile Creek Dunites, but in contrast no counterparts of the mylonites or the isoclinally folded and boudinaged tectonites have been observed in the latter.

In subsequent chapters, differences in chemical compositions and serpentinite types in the Main Belt complexes and the Adamsfield Complex are discussed. The Boyes River body is composed of similar serpentinites and is probably related to the Adamsfield Complex. However, despite the differences in location, petrology and chemistry between the Main Belt and the Adamsfield ultramafics, it is nevertheless interesting that osmiridium is associated with both. (The upper mantle under Tasmania in the Eocambrian was presumably enriched in osmium and iridium relative to other platinoid elements.)

CHAPTER 5

CHEMICAL COMPOSITIONS OF THE TASMANIAN ULTRAMAFICS AND GABBROS

INTRODUCTION

In this chapter, the chemistry of the primary ultramafic and gabbroic rocks and their constituent minerals is discussed in detail. Comparisons are made with such rocks as the large layered intrusives, alpine ultramafics and ophiolite complexes, and conclusions are reached on the nature of the magmas from which the layered ultramafics and gabbros crystallized.

The methods of analysis are discussed in the Appendix.

BULK COMPOSITIONS OF THE ULTRAMAFICS AND GABBROS

The chemical compositions of most of the ultramafics analysed by the writer are presented in Table 1, while analyses of rocks from the Serpentine Hill Complex are given in the Appendix. Additional analyses of Tasmanian ultramafics and gabbros are given in Taylor (1955), Spry (1962b), Palethorpe (1972), Brown (1972) and Gee and Legge (*in press*).

1. The affect of serpentinization on chemical composition

As can be seen from the ignition losses, the olivine-bearing ultramafics are partly or largely serpentinized. The question as to whether metasomatism has accompanied serpentinization must be considered if the serpentinized samples are to be taken as approximately representing the primary composition, allowance being made for the H_2O added.

Table 1

Analyses of ultramafic rocks.

	38526	38531	38595	38530	38540	38622	38619	38555	38565	38587	38591	38569
SiO_2	36.71	56.36	54.36	54.55	54.81	54.56	52.50	40.11	38.57	52.39	50.78	38.44
Al_2O_3	0.95	0.18	0.39	1.12	1.78	3.02	7.49	2.14	6.35	1.95	3.87	11.55
Fe_2O_3	1.53		1.80		0.60	0.80	0.59	4.38	2.16	1.60	0.47	0.53
FeO	7.71	7.73*	6.22	9.02*	9.60	9.79	8.58	4.11	6.91	5.38	6.52	5.29
MgO	41.97	34.14	33.39	32.23	31.75	27.99	22.81	33.65	31.60	30.97	29.08	22.17
CaO	0.04	0.12	1.28	0.79	1.21	1.96	3.98	2.14	3.77	3.47	5.35	13.10
Na_2O	0.01	0.02	0.01	0.01	0.02	0.32	0.84	0.01	0.11	0.01	0.22	0.05
K_2O	n.d.	0.00	0.00	0.00	0.00	0.00	0;21	0.00	0.00	0.00	0.00	0.00
TiO_2	0.07	0.60	0.11	0.09	0.14	0.14	0.13	0.09	0.11	0.12	0.13	0.07
P_2O_5	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.02	0.00	0.00	0.00	0.00
MnO	0.11	0.10	0.17	0.13	0.18	0.17	0.16	0.17	0.13	0.12	0.13	0.08
NiO	0.36	n.d.	0.16	0.28	0.29	0.15	0.20	0.27	0.23	0.13	0.21	0.28
Cr_2O_3	0.64	0.97	0.56	0.73	0.74	0.48	0.48	0.62	0.66	0.57	0.62	0.47
Loss	10.02	0.84	2.29	0.68	0.79	1.22	1.09	11.82	9.57	2.86	2.11	8.16
	100.12	100.56	100.74	99.63	101.91	100.60	99.07	99.53	100.17	99.57	99.49	100.19

* total Fe as FeO.

38526 serpentinized dunite, Nineteen Mile Ck Dunites
38531 pyroxenite, Nineteen Mile Creek Dunites
38595 olivine pyroxenite, Nineteen Mile Ck Dunites
38539 olivine pyroxenite, Caudrys Hill Pyroxenites
38540 pyroxenite, Caudrys Hill Pyroxenites
38622 plagioclase pyroxenite, Caudrys Hill P'ites
38619 " " " " "

38555 poikilitic harzburgite, Brassey Hill Harzburgites
38565 " " "
38587 olivine pyroxenite, "
38591 " " "
38569 plagioclase lherzolite, "

Table 1 cont.

	38604	38573	38576	38874	38573	38544	38596	38563	38853	38862	38863
SiO_2	35.43	42.47	41.69	43.12	49.48	52.26	43.14	41.74	52.39	53.02	57.39
Al_2O_3	9.59	7.46	14.91	0.81	2.84	1.41	5.77	1.06	1.95	1.54	0.18
Fe_2O_3	3.06	1.71	0.64	0.93	2.40		2.82	4.92	1.60		0.00
FeO	3.51	6.90	6.03	7.76	10.01	8.30*	9.98	3.55	5.38	7.59*	5.31
MgO	28.10	27.30	18.30	35.62	27.60	23.61	27.27	35.38	30.97	30.21	36.53
CaO	7.14	5.72	11.08	0.47	2.74	10.71	4.85	1.35	3.47	4.07	1.16
Na_2O	0.02	0.17	0.46	0.02	0.03	0.20	0.37	0.02	0.01	0.00	0.08
K_2O	0.00	0.03	0.05	0.00	0.02	0.00	0.02	0.00	0.00	0.04	0.00
TiO_2	0.10	0.11	0.11	0.10	0.16	0.12	0.21	0.12	0.12	0.15	0.10
P_2O_5	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00
MnO	0.06	0.12	0.10	0.14	0.16	0.12	0.10	0.12	0.12	0.13	0.09
NiO	0.27	0.19	0.14	0.21	0.24	0.12	0.19	0.23	0.13	0.22	0.15
Cr_2O_3	0.52	0.61	0.35	0.66	0.75	0.41	0.80	0.72	0.57	0.67	0.57
Loss	11.39	7.61	6.80	9.61	3.88	1.93	5.28	10.74	2.86	2.74	0.00
	99.19	100.40	100.66	99.45	100.31	99.19	99.81	99.95	99.57	100.38	101.38

* total Fe as FeO.

38604 Plagioclase dunite, Brassey Hill Harzburgites
38575 plagioclase lherzolite, Brassey Hill Harz'ites
38576 ol-opx-cpx-bearing gabbro, " " "
38874 harzburgite, Caudrys Hill Pyroxenites
38573 olivine pyroxenite, Brassey Hill Harzburgites
38541 pyroxenite, " " "
38596 plagioclase lherzolite, "undifferentiated", Heazlewood River Complex
38563 harzburgite, ultramafic dyke, Heazlewood River Complex
38853 pyroxenite, Wilson River Complex
38862 pyroxenite, Andersons Creek Complex
38863 pyroxenite, Adamsfield Complex

In chapter 7, it is established that metasomatism has occurred where the ultramafic rock has suffered post-serpentinization deformation or alteration to the green waxy-lustred serpentinites; such serpentinites are excluded from discussion in this chapter. It is also quite likely that Ca metasomatism was involved in the alteration of the plagioclase dunite 38604, in which the plagioclase has been replaced mainly by hydrogrossular.

For the serpentinite derived from a dunite with olivine of about Fo90-93, the MgO/SiO_2 ratio should be about 1.23 if the reaction was of the constant composition type (Coleman, 1971). Palethorpe (1972) analysed a partially serpentinized dunite (BH1) from the Nineteen Mile Creek Dunites, and obtained a MgO/SiO_2 ratio of 1.23. The serpentinized dunite 38526 has a lower MgO/SiO_2 ratio of 1.14, but this is perhaps because it has a higher FeO content. Both these dunites contain brucite, as is expected for constant composition serpentinization of dunite (Coleman, 1971). Serpentinized or partly serpentinized harzburgites from the Main Belt complexes have MgO/SiO_2 ratios in the range 0.70 - 0.90. This is consistant with constant composition serpentinization, since harzburgites should have ratios in the range 0.55 (orthopyroxenite) to 1.23 (dunite), depending on the relative proportions of olivine and pyroxene. However, this could only be confirmed by doing detailed modal analyses and analysing the constituent minerals, as done by Green (1964). Such work was not done by the present writer, but whether or not metasomatism has occurred during serpentinization does not affect the main conclusions of this chapter, as unaltered pyroxenites show the same differentiation ranges as the serpentinized harzburgites.

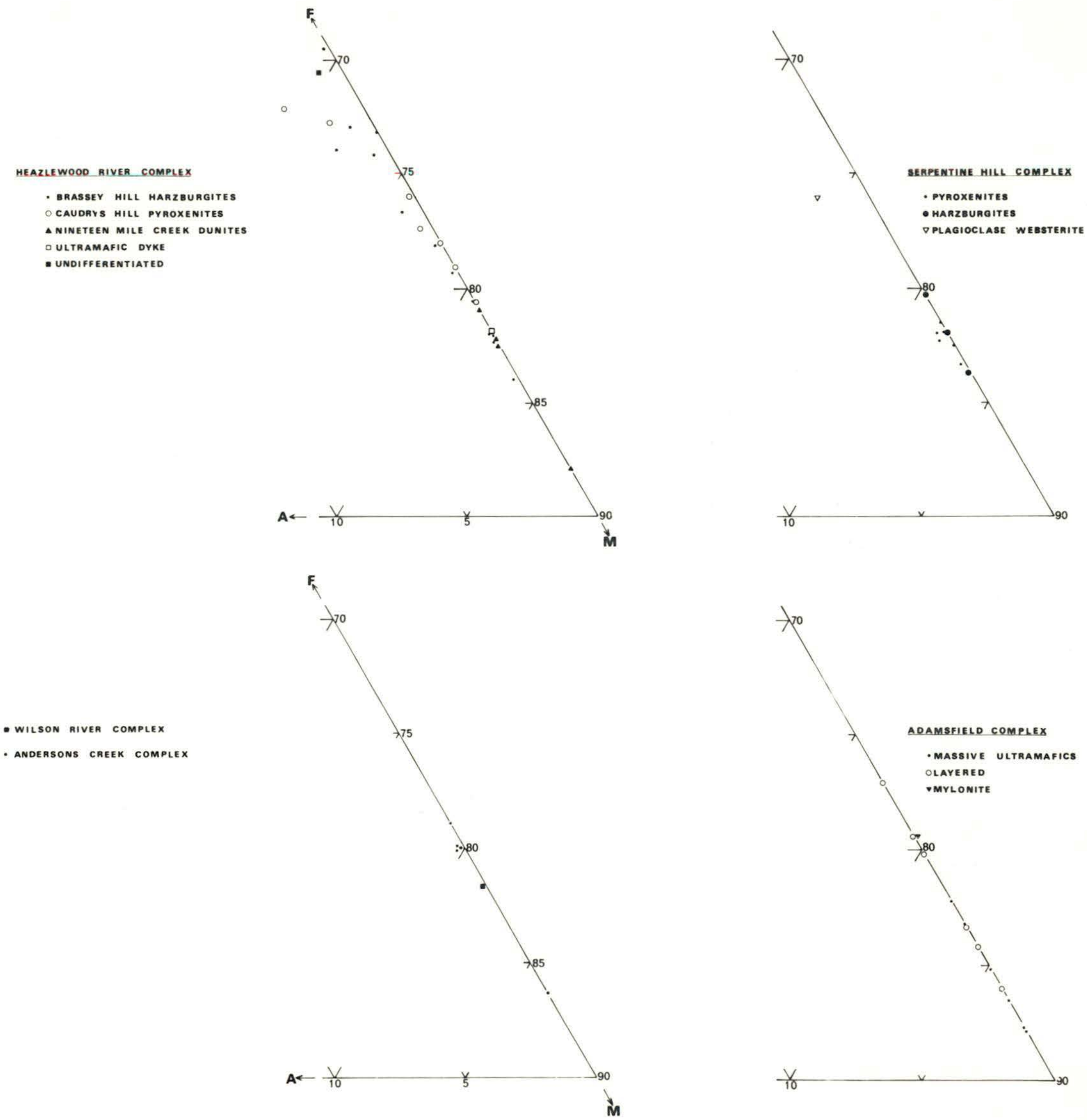
2. Comparisons between the Tasmanian complexes

It must be emphasized that the complexes are dismembered and were not sampled systematically. In other words, compositional ranges quoted probably do not represent the ranges of the original layered sequences.

Rocks from the Brassey Hill Harzburgites and the Caudrys Hill Pyroxenites show a similar range in chemical composition. This is illustrated in the AFM plot of Figure 18, which shows a continuum in weight percent MgO/MgO+'FeO' ratios from 69.0 to 82.2 for both groups. One troctolite analysed by Palethorpe (1972) falls outside the range with a value of 83.7. There are no obvious differences in the compositions of pyroxenites, which occur both in the Caudrys Hill Pyroxenites and the Brassey Hill Harzburgites. In contrast, the Nineteen Mile Creek Dunites have MgO/MgO+FeO ratios between 80.9 and 82.5, with one partly serpentinized dunite (BH1) with a value of 87.8. (This sample is referred to above in relation to constant composition serpentinization.)

There is no obvious distinction in composition between the pyroxenites and harzburgites of the Serpentine Hill Complex and similar rock types in the Brassey Hill Harzburgites and the Caudrys Hill Pyroxenites, even with regard to the NiO and Cr_2O_3 contents and trace V and Co values given in Palethorpe (1972). The pyroxenites and harzburgites of the Serpentine Hill Complex show a more restricted range of MgO/MgO+FeO rock and the plagioclase websterite is the only analysed ultramafic rock with a ratio of less than 80. However, the Serpentine Hill Complex is more dismembered than the Heazlewood River Complex, and more of the original sequence is probably missing.

Figure 18. AFM diagrams for Tasmanian ultramafic rocks. Oxide values are in weight percent.


HEAZLEWOOD RIVER COMPLEX
• BRASSEY HILL HARZBURGITES
○ CAUDRYS HILL PYROXENITES
▲ NINETEEN MILE CREEK DUNITES
□ ULTRAMAFIC DYKE
■ UNDIFFERENTIATED
SERPENTINE HILL COMPLEX
• PYROXENITES
● HARZBURGITES
▽ PLAGIOCLASE WEBSTERITE
■ WILSON RIVER COMPLEX
• ANDERSONS CREEK COMPLEX
ADAMSFIELD COMPLEX
• MASSIVE ULTRAMAFICS
○ LAYERED
▼ MYLONITE
F
A
M
70
75
80
85
90
10
5

Pyroxenites from the Andersons Creek Complex and one from the Wilson River Complex are similar to those of the Serpentine Hill Complex and to those of the Caudrys Hill Pyroxenites and Brassey Hill Harzburgites, except in that one sample from Taylor (1955) has a MgO/MgO+'FeO' ratio of 86.3.

The layered ultramafics from the Adamsfield Complex show a range of MgO/MgO+FeO ratios from 77 to 86, while the massive rocks range from 82 to 88. (Note: in Fig.6 of Brown, 1972, cation percentages for Adamsfield rocks are compared with weight percentages from other complexes.)

3. <u>Layered gabbros and intrusive gabbros</u>

Compositions of intrusive gabbros from the Heazlewood River Complex are given in Table 2, while microgabbros and a layered gabbro from the Serpentine Hill Complex are given in the Appendix.

In terms of chemical composition, there appears to be an overlap between the layered gabbros, coarser-grained intrusive gabbros, and micro-gabbros. Plotted on an AFM diagram, most of these fall in a cluster removed from the ultramafics, with one intrusive gabbro occurring in the intervening space. As is discussed in chapter 3 and the Appendix on the Serpentine Hill Complex, there also appears to be a mineralogical and textural gradation between all these gabbro types. Two rocks from the McIvor Hill body are more differentiated than the other gabbroic rocks.

It is not known whether the apparent bimodal distribution of layered gabbros and ultramafics on the AFM plot is real or is due to insufficient sampling. Structurally there seems to be a gradation from

Table 2

Compositions of gabbros and microgabbros from the Heazlewood River Complex.

	38639	38642	38705	38891	38893	38897
SiO_2	49.15	49.51	52.05	49.66	48.74	49.03
Al_2O_3	15.90	15.10	15.27	15.03	15.89	15.19
Fe_2O_3	0.90		1.44			
FeO	5.85	9.13*	7.87	9.83*	7.58*	9.86*
MgO	11.15	11.39	8.66	9.83	9.89	9.53
CaO	13.42	13.16	11.67	13.52	11.70	12.52
Na_2O	0.79	1.01	2.54	1.13	1.63	1.28
K_2O	0.20	0.20	0.00	0.02	0.70	0.20
TiO_2	0.14	0.24	0.12	0.24	0.18	0.31
P_2O_5	0.00	0.00	0.00	0.00	0.00	0.00
MnO	0.10	0.14	0.08	0.15	0.11	0.11
NiO	0.02	0.03	0.05	n.d.	0.02	0.01
Cr_2O_3	0.06	0.17	0.01	n.d.	0.12	0.02
Loss	1.52	0.48	1.02	0.61	3.68	2.35
	100.22	100.56	100.78	100.02	100.24	100.38

38639, 38642 gabbros

38705, 38891, 38893, 38897 microgabbros

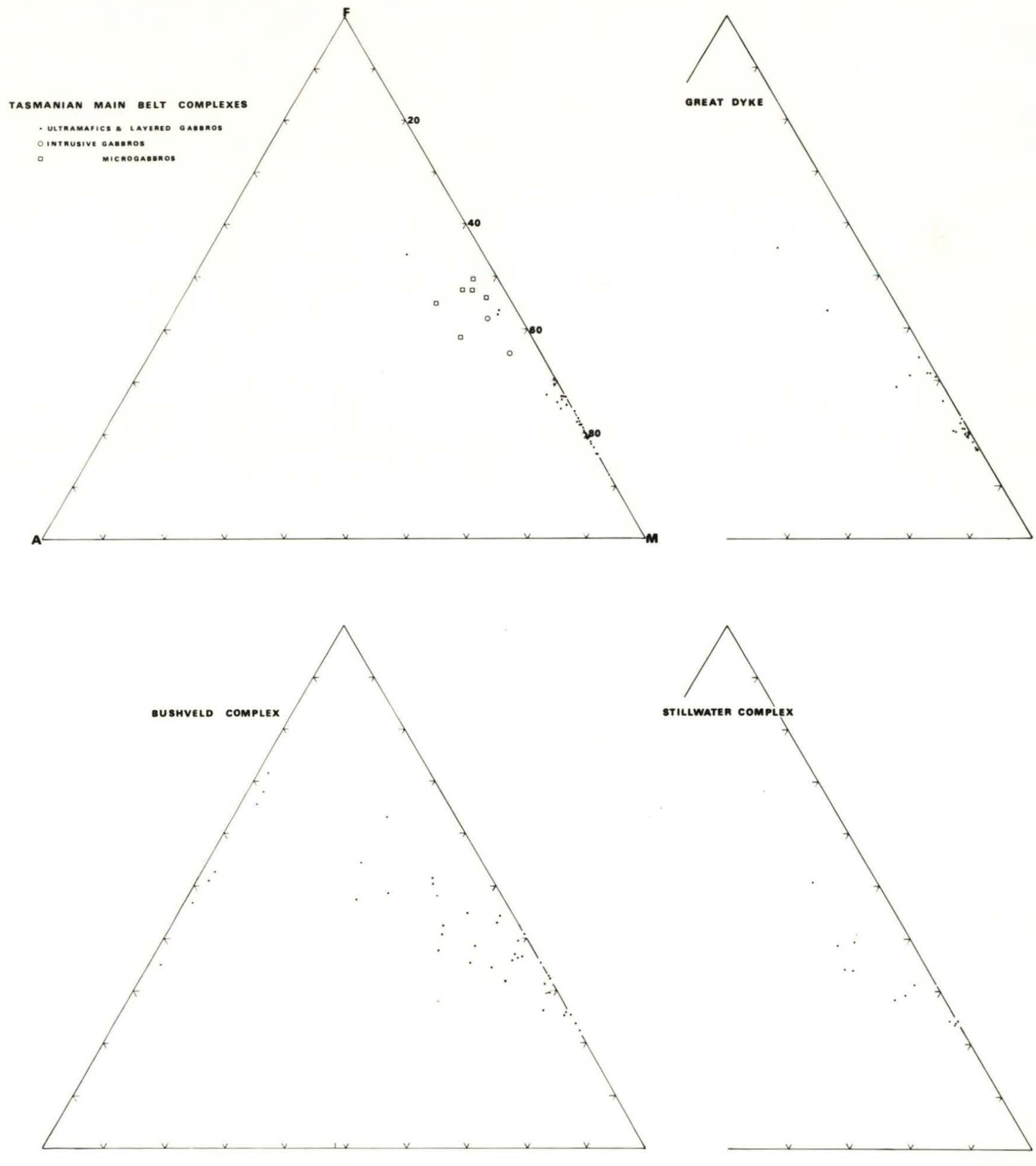
* total Fe as FeO.

layered ultramafics into layered ultramafics in both the Andersons Creek and Serpentine Hill Complex, but it is possible that the layered gabbros in both these complexes crystallized from a separate magma (possibly similar in composition to the other intrusive gabbros) which intruded the ultramafics. On the other hand, the chemical differences do not preclude crystallization of the layered ultramafics, the transition rocks and the layered gabbros all from the one magma.

4. Comparison of the Tasmanian rocks with the large layered intrusions

In Figure 19, the Main Belt ultramafics and layered gabbros are compared with the layered rocks of the Bushveld, Stillwater and Great Dyke intrusions, and show a closer resemblance to the Great Dyke. The earliest differentiates of the Stillwater and Bushveld intrusions have MgO/MgO+FeO ratios of 78 or less, while those from the Great Dyke have ratios of as much as 83; with the exception of one dunite which may not be part of the layered sequence, the Main Belt ultramafics have MgO/MgO+FeO ratios of up to 84. The plagioclase-bearing differentiates of the Great Dyke and Main Belt complexes tend to hug the FM tieline of the AFM diagram than do gabbroic rocks of the Stillwater and Bushveld intrusions; in the case of the Main Belt complexes, this reflects the lack of leucocratic gabbros and anorthosites rather than the plagioclase compositions, which if anything are less anorthite-rich than in the other layered intrusions. The Stillwater, Great Dyke and Main Belt complexes do not show the upper extremes of differentiation as in the Bushveld Complex, but the Bushveld intrusion is the only one of these with its top exposed.

Figure 19. AFM diagrams comparing Tasmanian ultramafics and layered gabbros with the Bushveld Complex (Hall, 1932; Lombaard, 1934) the Stillwater Complex (Hess, 1960; Bowes and Skinner, 1969) and the Great Dyke (Worst, 1958). Values are in weight percent.


F
TASMANIAN MAIN BELT COMPLEXES
· ULTRAMAFICS & LAYERED GABBROS
○ INTRUSIVE GABBROS
□ MICROGABBROS
20
40
60
80
A
M
GREAT DYKE
BUSHVELD COMPLEX
STILLWATER COMPLEX

Other comparisons between the Great Dyke and the Main Belt complexes are also interesting. The Hartley area of the Great Dyke is about 3 km thick, with a 2 km ultramafic zone (Worst, 1958; Wager and Brown, 1968). Layered ultramafics in the Heazlewood River Complex could extend for as much as 4 km below the surface, but it is not known how many separate intrusions are present or whether the same sequence is repeated by faulting and/or folding. The ultramafic zone of the Great Dyke is composed of olivine and orthopyroxene cumulates while the gabbro zone, in which plagioclase, augite and orthopyroxene are all cumulus phases, contains no olivine; this compares very well with the Main Belt complexes. There is a disproportionate amount of ultramafics relative to layered gabbros in the Tasmanian complexes and in the Great Dyke in comparison with the Bushveld Complex which is known to have crystallized from a basic magma (Wager and Brown, 1968). This can be explained in either of three ways.

1) Much of the gabbro zones are not exposed. This is particularly relevant in the case of the Tasmanian complexes considering their degree of dismemberment.

2) Crystallization occurred in an open magma chamber, with a considerable part of the parent magma being removed, perhaps tapped off through overlying volcanoes (Wager and Brown, 1968, p.467). This could also be relevant to the Tasmanian complexes considering the large volumes of volcanics associated with the layered rocks in the two ophiolite complexes, and will be discussed in detail in another section.

3) The Great Dyke and Tasmanian complexes may have crystallized from picritic or high-Mg basaltic magmas rather than ordinary basaltic

magmas. This is quite possible in the case of the Great Dyke, as the olivines (Fo94) and orthopyroxenes (En93) are more magnesian than in the layered basic intrusions which have also crystallized at relatively low pressures (Wager and Brown, 1968, p.467).

There are numerous differences of course between the layered rocks of the Main Belt complexes and the Great Dyke when more detailed comparisons are made. For example, layers of chromitite appear to be quite rare and postcumulus pyroxenes more abundant in the olivine cumulates of the Tasmanian complexes in comparison with the Great Dyke. The geological setting is also quite different - the Great Dyke is a post-tectonic intrusion in a stable area while the Tasmanian complexes occur in an orogenic belt.

5. Comparisons with alpine ultramafics and ophiolites

5.1 Alpine ultramafics and their origin

Alpine ultramafics are generally thought of as occurring in orogenic belts and being tectonically rather than magmatically emplaced into their present locations (Benson, 1926; Turner and Verhoogen, 1960; Wyllie, 1967). Emplacement can be achieved in various ways, of which cold solid intrusion as serpentinite, cool to hot emplacement by plastic flow of peridotite, and thrust emplacement being the more common mechanisms postulated. Peridotites, commonly partly to entirely serpentinized, are the most abundant rock type. The primary rocks may be classified as plagioclase-peridotite, spinel peridotite, or garnet peridotite facies (O'Hara, 1967), although low-alumina peridotites which do not contain plagioclase or garnet but contain low-alumina chromite as the spinel phase, are difficult to classify on this basis. The present writer prefers not to include in the definition of alpine ultramafics either the zoned ultramafic complexes such as those described in Taylor (1967), or Archaean ultramafics such as those described in Viljoen and Viljoen (1970) and Williams (1972).

A commonly held hypothesis is that alpine peridotites are tectonically emplaced slices of upper mantle (Hess, 1955; de Roever, 1957; Ragan, 1963; Green, 1964, 1970). In this hypothesis, it is usually postulated that the peridotite is residual upper mantle from which a basaltic fraction has been removed (Green and Ringwood, 1967). Other alpine ultramafics are thought to be cumulates which crystallized in the crust (for example, Challis, 1965, 1969), and some authors take the more extreme view that a relatively large proportion of alpine ultramafic bodies are dismembered re-emplaced cumulates (McTaggart, 1971). Thayer (1967, 1969) argues that many of the deformed alpine ultramafic-gabbro complexes were originally cumulates that were tectonically re-emplaced, perhaps as crystal mushes.

It is quite possible that there is at least an element of truth in each of the current hypotheses concerning the origin of alpine ultramafics. As long as the alpine ultramafic bodies are defined in terms of tectonic emplacement in an orogenic belt, then all the various hypotheses concerning the origin of the rocks before tectonic emplacement must be kept open, and each case examined on its own merits.

5.2 Chemistry of the Tasmanian complexes in comparison with alpine ultramafic bodies.

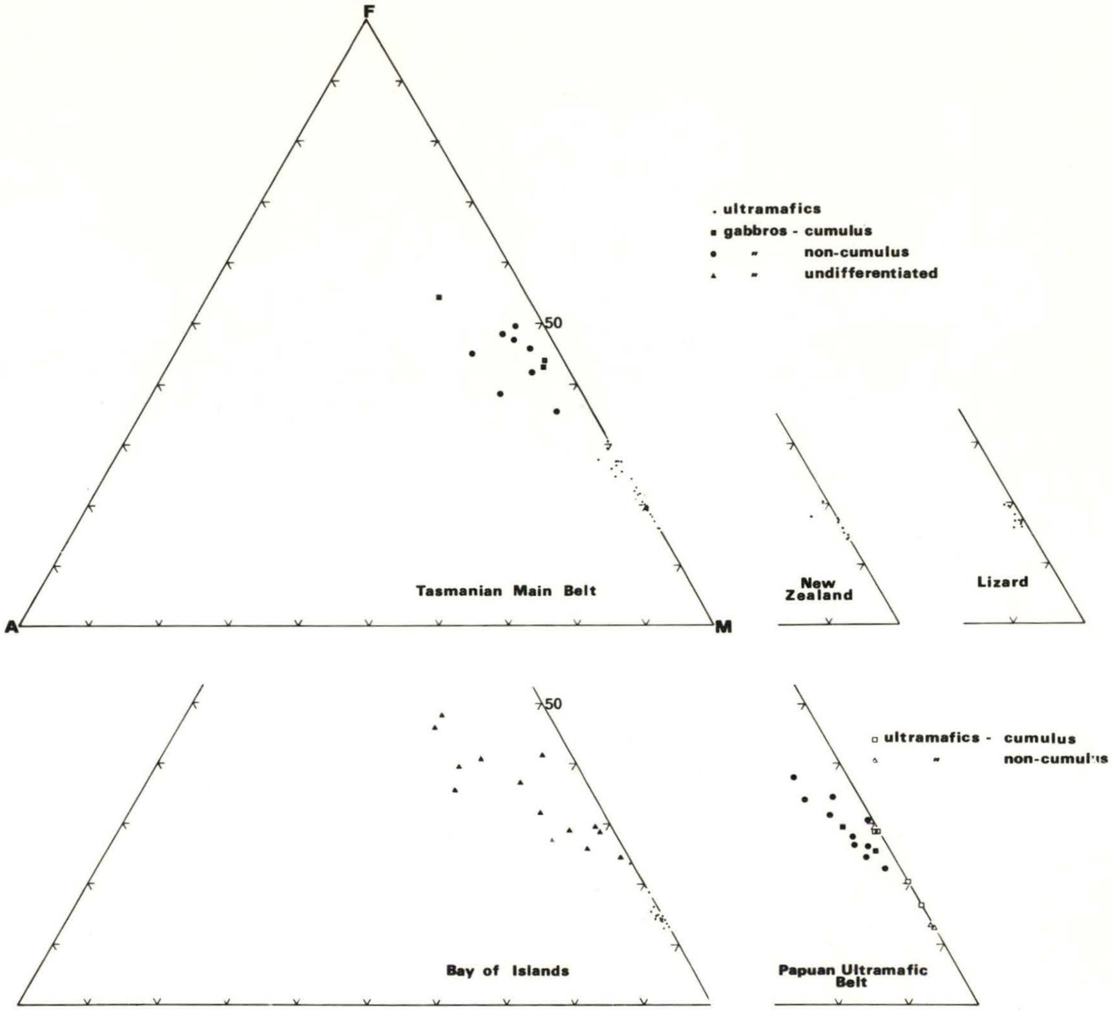
A large number of the "typical" alpine ultramafics show a restricted range in composition. Rocks from the Lizard (Green, 1964) are typical of these, and are plotted on an AFM diagram (Fig. 20). There is an overlap between the bulk compositions of such alpine peridotites and the cumulates from the large layered intrusions, in particular the Great Dyke. However, in such bodies in which the spinels are aluminous, a residual upper mantle origin is preferred (Green, 1967). It is also possible that alpine peridotites with aluminous spinels could have crystallized from magmas within upper mantle, as suggested for the Burro Mountain peridotite by Loney *et al.* (1971).

Other alpine peridotites of restricted bulk compositions include the New Zealand Permian ultramafics, which Challis (1965, 1969) believes to be largely cumulates. However, Walcott (1969) shows that the Red Hill Complex has deformed or tectonite fabrics, so the origin of the New Zealand ultramafics is still controversial.

5.3 Comparison of the Main Belt complexes with alpine ultramafic-gabbro and ophiolite complexes

Thayer (1967) points out that in many complexes the alpine ultramafics are intimately associated with gabbros which might be cogenitic. This is particularly true of the ophiolite complexes which (if complete) contain dolerite dyke swarms and basic volcanics as well as ultramafics and gabbros. The various rock types in an ophiolite bear definite structural relationships to each other, as is discussed in detail in chapter 7.

Figure 20. AFM diagrams, comparing Tasmanian ultramafics and gabbros with alpine ultramafics and ophiolite complexes. Values are in weight percent. Data for the Lizard are taken from Green (1964), for the New Zealand ultramafics from Challis (1965), the Papuan Ultramafic Belt from Davies (1961), and the Bay of Islands complex from Smith (1958) and Bowes _et al_ (1970).


F
A
M
50
. ultramafics
■ gabbros - cumulus
● " non-cumulus
▲ " undifferentiated
Tasmanian Main Belt
New Zealand
Lizard
50
Bay of Islands
□ ultramafics - cumulus
△ " non-cumulus
Papuan Ultramafic Belt

Thayer (1967) plots rocks from some of the classic Tethyan ophiolites on AFM diagrams, and in general terms the distribution is similar to that obtained for the Main Belt complexes in Tasmania. He suggests the various rock types may form an "Alpine Mafic Magma Stem". However, it must be emphasized that trends on variation diagrams do not necessarily mean that the rocks are genetically related. If genetic connections exist, they may include fractional melting as well as fractional crystallization sequences.

It is now believed by a number of authors that the majority of ultramafics in complete ophiolite sequences are non-cumulative and are probably residual upper mantle, while layered ultramafics and gabbros of cumulative origin comprise a smaller proportion (Moores, 1969; Moores and Vine, 1971; Coleman, 1971; Davies, 1972). Abundant intrusive gabbros also occur in ophiolites. In contrast to these ophiolites, most of the ultramafics of the Main Belt complexes are of cumulative origin.

In Figure 20, data of Davies (1971) from the Papuan Ultramafic Belt in which satisfactory distinction is made between cumulative and non-cumulative ultramafics, as well as between layered and non-layered gabbros, are compared with the ultramafics and gabbros from the Main Belt. The four cumulus ultramafics from the Papuan Ultramafic Belt show a range of MgO/MgO+FeO ratios from 70 to 80, compared with a range of 69 to 84 for the Main Belt cumulates. As in the Main Belt complexes, layered and non-layered gabbros from the Papuan Ultramafic Belt are very similar in chemistry, but in contrast they are more magnesian than the Tasmanian gabbros and overlap with the cumulate ultramafics. Rocks from the Troodos and Vourinos complexes (Moores and Vine, 1971; Brunn, 1956) show a similar distribution as in the Main Belt complexes, but in these cases distinctions between cumulates and non-cumulates, and between layered and intrusive gabbros are generally not made in the presentation of the analytical data.

The ultramafics and gabbros from the Bay of Islands Complex (Smith, 1958; Bowes *et al.*, 1970) are also plotted on Figure 20. This complex is now known to be an ophiolite or a series of ophiolites (Williams and Malpas, 1972). Wager and Brown (1968), together with Bowes *et al.* (1970), believe the ultramafics and gabbros to be cumulates. The present writer, however, believes that comparisons based on chemical data, such as those made by Bowes *et al.* (1970), are meaningless unless the geological situation and textures of the specimens analysed are studied in fair detail.

6. Conclusions

In terms of bulk chemical compositions, the ultramafics and layered gabbros from the Main Belt complexes show a variation similar to the ultramafic zones and lower gabbro zones of the Bushveld, Stillwater and, in particular, the Great Dyke intrusions. However, the Tasmanian complexes are alpine bodies in that they have suffered tectonic re-emplacement, and occur in an orogenic belt (in fact, were initially emplaced in the sedimentation phase in the development of the Tasman Orogenic Zone in Tasmania). In comparison with other alpine bodies, they show a greater degree of affinity to the ophiolite complexes than other varieties of alpine peridotites which show a more narrow range of compositions. In the following chapter, discussion of the chemistry, origin and tectonic significance of ophiolites is dealt with in more detail.

In this section, little has been said concerning the Adamsfield ultramafics, some of which are quite magnesian in composition. This body will be discussed in the following section on mineralogical compositions.

COMPOSITIONS OF CONSTITUENT MINERALS

1. Olivines

Electron microprobe analyses of olivines from the Heazlewood River Complex were done concurrently with pyroxenes. Unfortunately, in contrast with the pyroxenes, the olivine analyses gave poor totals (around 93%), thought to be due to a poor carbon coating on the olivine standard used. The analyses are not tabulated here, but the Fo values given below may be approximately correct.

38528	Fo88
38569	Fo88
38538	Fo86
38631	Fo86
38657	Fo84
38576	Fo81
38573	Fo81
38558	Fo79

As can be seen from Figure 16, the Fo values are similar to En values for coexisting orthopyroxenes.

Table 3

Compositions and structural formulae of orthopyroxenes.

	Orthopyroxenes from ultramafic rocks								from microgabbros		
	38528	38569	38538	38576	38573	38657	38558	38887	34747	38808	38891
SiO_2	56.91	57.35	57.05	57.21	56.93	57.26	55.91	57.80	54.18	52.35	53.56
TiO_2	0.06	0.06	0.06	0.12	0.08	0.09	0.28	0.05	0.16	0.17	0.12
Al_2O_3	2.89	2.40	1.73	1.39	1.75	1.81	1.29	2.22	1.30	1.06	1.05
FeO	6.53	6.66	8.21	10.25	8.20	9.22	9.94	6.06	17.34	19.89	16.38
Fe_2O_3*	0.72	0.82	0.78	1.36	2.87	0.91	3.63	0.50	2.73	3.13	4.08
MnO	0.12	0.16	0.14	0.24	0.21	0.15	0.29	0.12	0.38	0.46	0.44
MgO	31.17	32.44	31.13	30.03	29.78	30.43	28.36	32.90	22.72	19.91	22.87
CaO	1.19	1.20	0.90	0.79	0.88	0.94	0.91	0.90	1.69	1.30	1.06
Na_2O	0.04	0.00	0.05	0.00	0.02	0.05	0.00	0.00	0.00	0.00	0.06
K_2O	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cr_2O_3	0.54	0.50	0.30	0.27	0.25	0.21	0.11	0.58	0.21	0.10	0.00
NiO	0.12	0.11	0.17	0.00	0.14	0.21	0.11	0.10	0.13	0.19	0.00
	100.29	101.70	100.52	101.71	101.12	101.29	100.83	101.23	100.85	99.56	

* Fe_2O_3 values were taken from the bulk composition of the same specimen or the closest rock type.

Table 3 cont.

	-Orthopyroxenes from ultramafic rocks								from microgabbros		
	38528	38569	38538	38576	38573	38657	38558	38887	34746	38808	38891
Si	1.963	1.958	1.979	2.008	1.973	1.980	1.968	1.967	1.952	1.982	1.970
Al(Z)	0.037	0.042	0.021	0.000	0.027	0.020	0.032	0.033	0.048	0.018	0.030
Al(WXY)	0.081	0.050	0.050	0.043	0.045	0.054	0.022	0.056	0.010	0.020	0.016
Ti	0.002	0.002	0.002	0.003	0.002	0.002	0.007	0.013	0.005	0.005	0.003
Fe^{3+}	0.019	0.021	0.020	0.036	0.075	0.024	0.096	0.001	0.077	0.089	0.113
Fe^{2+}	0.188	0.190	0.238	0.298	0.238	0.266	0.293	0.173	0.541	0.630	0.504
Mn	0.004	0.003	0.004	0.007	0.006	0.004	0.009	0.003	0.012	0.015	0.014
Mg	1.603	1.650	1.609	1.556	1.537	1.568	1.488	1.671	1.263	1.124	1.254
Ca	0.044	0.044	0.033	0.030	0.033	0.035	0.034	0.033	0.068	0.053	0.042
Na	0.002	0.000	0.003	0.000	0.001	0.003	0.000	0.000	0.000	0.000	0.002
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Cr	0.015	0.014	0.008	0.008	0.007	0.006	0.003	0.016	0.006	0.003	0.000
Ni	0.003	0.003	0.005	0.000	0.003	0.006	0.003	0.003	0.004	0.006	0.000
Z	2.000	2.000	2.000	2.008	2.000	2.000	2.000	2.000	2.000	2.000	2.000
WXY	1.960	1.975	1.973	1.980	1.947	1.988	1.981	1.957	1.985	1.971	1.947
mg	88.6	88.7	86.2	82.3	83.1	84.4	79.3	90.6	67.2	61.0	67.0
'Mg'	86.5	86.6	84.6	81.1	81.7	82.8	77.9	89.0	64.8	59.3	65.6
'Ca'	2.4	2.3	1.8	1.5	1.7	1.2	1.8	1.8	3.5	2.8	2.2
'Fe'	11.1	11.1	13.6	17.4	16.6	16.0	20.3	9.3	31.7	37.9	32.2

Electron microprobe analyses done by M. Rubenach, at A.N.U.

mg $Mg/Mg+Fe^{2+}+Fe^{3+}$, cation ratios.

Table 4

Compositions and structural formulae of clinopyroxenes.

	Clinopyroxenes from ultramafics					from microgabbros			dolerite	tonalite
	38569	38576	38558	38887	38807	38808	38891	38644	38700	38786
SiO_2	52.14	50.49	51.45	53.34	51.78	49.89	51.26	49.86	51.06	50.91
TiO_2	0.15	0.20	0.22	0.12	0.19	0.17	0.24	0.20	0.21	0.20
Al_2O_3	3.46	2.75	2.45	2.71	2.12	1.29	2.03	2.34	2.63	1.85
FeO	2.37	4.17	5.31	2.23	4.39	8.98	7.36	9.35	8.82	10.04
Fe_2O_3*	0.82	0.55	0.70	0.19	0.36	1.41	1.83	1.89	1.12	1.46
MnO	0.09	0.12	0.11	0.06	0.14	0.20	0.22	0.27	0.23	0.27
MgO	16.83	15.99	16.61	17.22	16.53	13.66	14.63	13.21	16.86	14.66
CaO	23.50	23.24	22.49	23.83	23.01	22.75	20.56	21.95	19.55	20.38
Na_2O	0.32	0.35	0.30	0.29	0.24	0.27	0.15	0.22	0.25	0.31
K_2O	0.02	0.03	0.03	0.00	0.00	0.00	0.03	0.06	0.00	0.04
Cr_2O_3	0.92	0.58	0.39	0.92	0.44	0.11	0.05	0.00	0.22	0.00
NiO	0.03	0.13	0.12	0.12	0.11	0.14	0.12	0.14	0.12	0.14
	100.11	98.60	100.17	101.03	99.31	98.67	98.48	99.49	101.07	100.26

* Fe_2O_3 values were taken from the bulk compositions of the same specimen or the closest rock type.

Table 4 cont.

	Clinopyroxenes from ultramafics					from microgabbros			dolerite	tonalite
	38569	38576	38558	38887	38807	38808	38891	38644	38700	38786
Si	1.850	1.891	1.898	1.993	1.918	1.909	1.934	1.896	1.884	1.911
Al(Z)	0.150	0.109	0.102	0.077	0.032	0.058	0.046	0.104	0.114	0.082
Al(WXY)	0.003	0.012	0.005	0.038	0.011	0.000	0.024	0.001	0.000	0.000
Ti	0.004	0.007	0.006	0.003	0.005	0.005	0.007	0.006	0.006	0.006
Fe^{3+}	0.023	0.015	0.020	0.005	0.010	0.041	0.052	0.054	0.031	0.041
Fe^{2+}	0.075	0.131	0.164	0.067	0.136	0.287	0.232	0.297	0.272	0.315
Mn	0.003	0.004	0.003	0.002	0.005	0.006	0.007	0.009	0.007	0.009
Mg	0.944	0.893	0.913	0.925	0.913	0.779	0.823	0.747	0.927	0.820
Ca	0.948	0.933	0.889	0.920	0.914	0.933	0.831	0.893	0.773	0.820
Na	0.024	0.025	0.021	0.020	0.018	0.022	0.011	0.003	0.018	0.023
K	0.001	0.001	0.001	0.000	0.000	0.000	0.001	0.001	0.000	0.002
Cr	0.028	0.017	0.012	0.026	0.013	0.003	0.001	0.000	0.006	0.000
Ni	0.004	0.004	0.004	0.004	0.003	0.004	0.004	0.004	0.004	0.004
Z	2.000	2.000	2.000	2.000	2.000	1.967	2.000	2.000	1.998	1.993
WXY	2.056	2.041	2.037	2.011	2.027	2.081	1.993	2.014	2.044	2.039
mg	90.6	86.0	83.3	92.8	86.2	70.4	74.3	68.1	75.4	69.7
'Mg'	47.5	45.3	46.0	48.2	46.3	38.2	42.4	37.5	46.3	41.1
'Ca'	47.6	47.3	47.8	48.0	46.3	45.7	42.9	49.9	38.6	41.1
'Fe'	4.9	7.4	9.2	3.8	7.4	16.1	14.7	17.6	15.1	17.8

Electron microprobe analyses done by M. Rubenach at A.N.U.

mg - Mg/Mg + Fe^{2+} + Fe^{3+}, cation ratios.

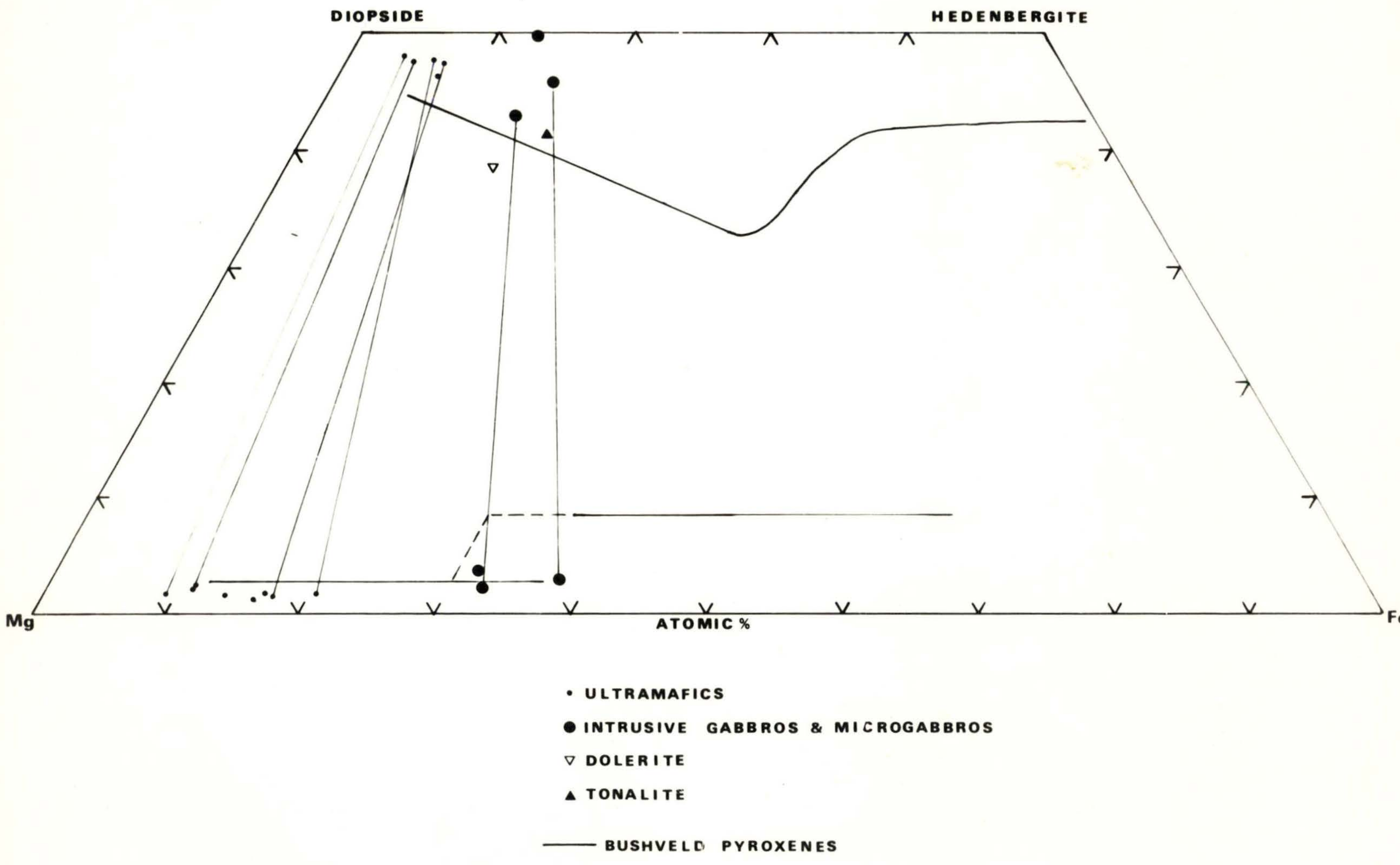
2. Pyroxenes

Analyses of pyroxenes from layered ultramafics and intrusive gabbros are given in Tables 3 and 4. Exsolution lamellae were not properly accounted for, producing likely small errors (except for the clinopyroxenes of 38700 and 38786, which do not contain lamellae). The errors would be larger for the ultramafic and gabbro clinopyroxenes as these contain thicker and more widely-spaced lamellae than the orthopyroxenes, and may account for the position of the clinopyroxenes from the Tasmanian complexes in comparison with the Bushveld trend on the pyroxene quadrilateral (Fig. 21).

The compositions of the pyroxenes are fairly similar to those from the ultramafic zones of the Bushveld and Stillwater Complexes (Hess, 1960; Wager and Brown, 1968; Atkins, 1969), even with regard to such components as TiO_2, Cr_2O_3, NiO and Al_2O_3. The orthopyroxenes range from En90.6 to 79.3 in the layered ultramafics, similar to the ultramafic zone of the Bushveld intrusion. No pyroxenes from layered gabbros have been analysed, and the values quoted above probably do not include the upper and lower limits of the layered rocks exposed in the Main Belt complexes.

Plotted on an Al_2O_3/SiO_2 diagram (Fig. 22), the clinopyroxenes fall in the tholeiitic field empirically determined by Challis (1965, 1969). The clinopyroxenes are mainly diopsides and augites according to the nomenclature of Poldervaart and Hess (1951).

Figure 21. Pyroxene compositions plotted on the pyroxene quadrilateral. Coexisting pyroxenes are joined by tie-lines.


DIOPSIDE
HEDENBERGITE
Mg
Fe
ATOMIC %
• ULTRAMAFICS
● INTRUSIVE GABBROS & MICROGABBROS
▽ DOLERITE
▲ TONALITE
—— BUSHVELD PYROXENES

2.1 Exsolution in pyroxenes

In most ultramafics, layered gabbros and intrusive gabbros in the Tasmanian complexes, the pyroxenes contain exsolution lamellae, with calcium-poor pyroxene lamellae in host calcium-rich pyroxenes and vice versa. In the host orthopyroxenes, the lamellae are parallel to (100)* and are rather thin and closely spaced, whereas the (100) lamellae in clinopyroxenes are thicker and more widely spaced. Rocks containing little or no clinopyroxene, in particular the lower zone of the Caudrys Hill Pyroxenites, contain very few obvious lamellae.

At first sight, the occurrence of (100) lamellae in both calcium-rich and calcium-poor pyroxenes suggests that the latter crystallized as orthopyroxene rather than pigeonite (Poldervaart and Hess, 1951, Hess, 1960). However, in lherzolites in the Brassey Hill Harzbürgites where postcumulus clinopyroxenes are abundant (for example, 38568, Plate 64), (001) as well as (100) lamellae occur in clinopyroxenes suggesting that they originally coexisted with pigeonite rather than orthopyroxene. Also the orthopyroxenes in rocks containing fairly abundant clinopyroxene (either postcumulus or cumulus) commonly contain abundant blebs of calcium-rich pyroxenes as well as (100) lamellae (Plate 65). Wager and Brown (1968, pp. 42-48) suggest that the presence of abundant blebs of calcium-poor pyroxenes implies that the host orthopyroxenes crystallized as pigeonite close to the inversion temperature, exsolution occurring almost contemporaneously with inversion.

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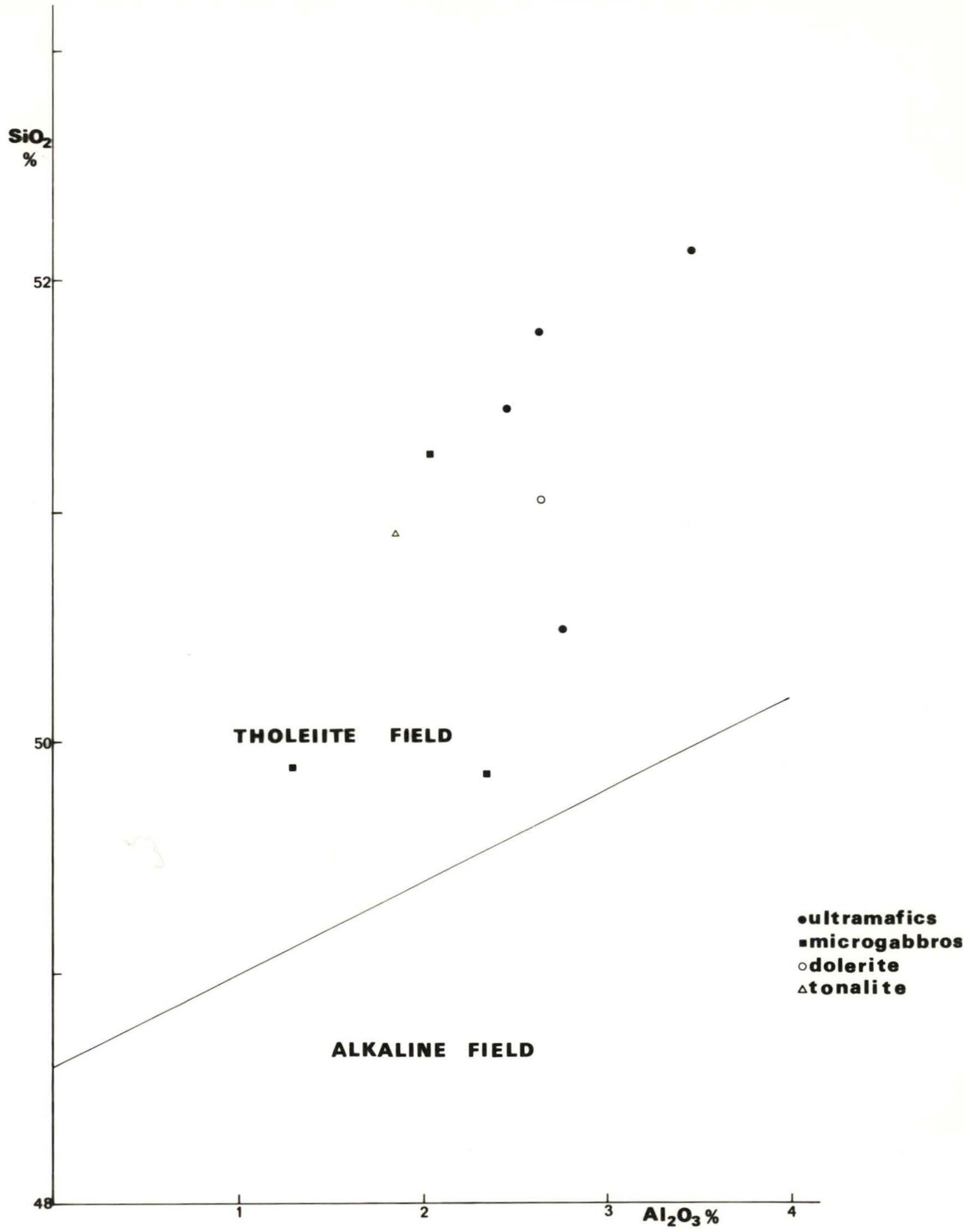
* Robinson *et al.* (1971) have demonstrated that pyroxene lamellae are not exactly (001) or (100).

Figure 22. Al_2O_3 versus SiO_2 plot for clinopyroxenes. The fields for tholeiites and alkali basalts are taken from Challis (1969).


SiO2 %
52
50
48
1
2
3
4
Al2O3 %
THOLEIITE FIELD
ALKALINE FIELD
•ultramafics
▪microgabbros
○dolerite
△tonalite

Curved irrational grain boundaries are commonly developed between orthopyroxenes and clinopyroxenes in cumulates or annealed metamorphics (see chapter 3), presumably satisfying minimum surface energy requirements (Spry, 1969). So pyroxenes exsolved at higher temperatures would perhaps tend to form blebs, while at lower temperatures the grain boundary migration would tend to be too sluggish and the normal thin lamellae would remain as such. The thin (100) lamellae which also occur in bleb-bearing orthopyroxenes may have exsolved during cooling after inversion.

If the interpretation regarding bleb lamellae is corrèct, it would imply that inverted pigeonites occur throughout the exposed Brassey Hill Harzburgites. The fact that (001) lamellae do not occur in calcium-rich pyroxenes in most of the ultramafics may be because the latter are generally postcumulus and so may have crystallized interstitially to the calcium-poor pyroxenes after inversion. The (001) lamellae are present (Plate 64) in the rare cases where large clinopyroxene oikocrysts have substantially "replaced" cumulus olivines, and perhaps this particular postcumulus reaction occurred earlier than normal, when the calcium-poor pyroxenes were still pigeonites.

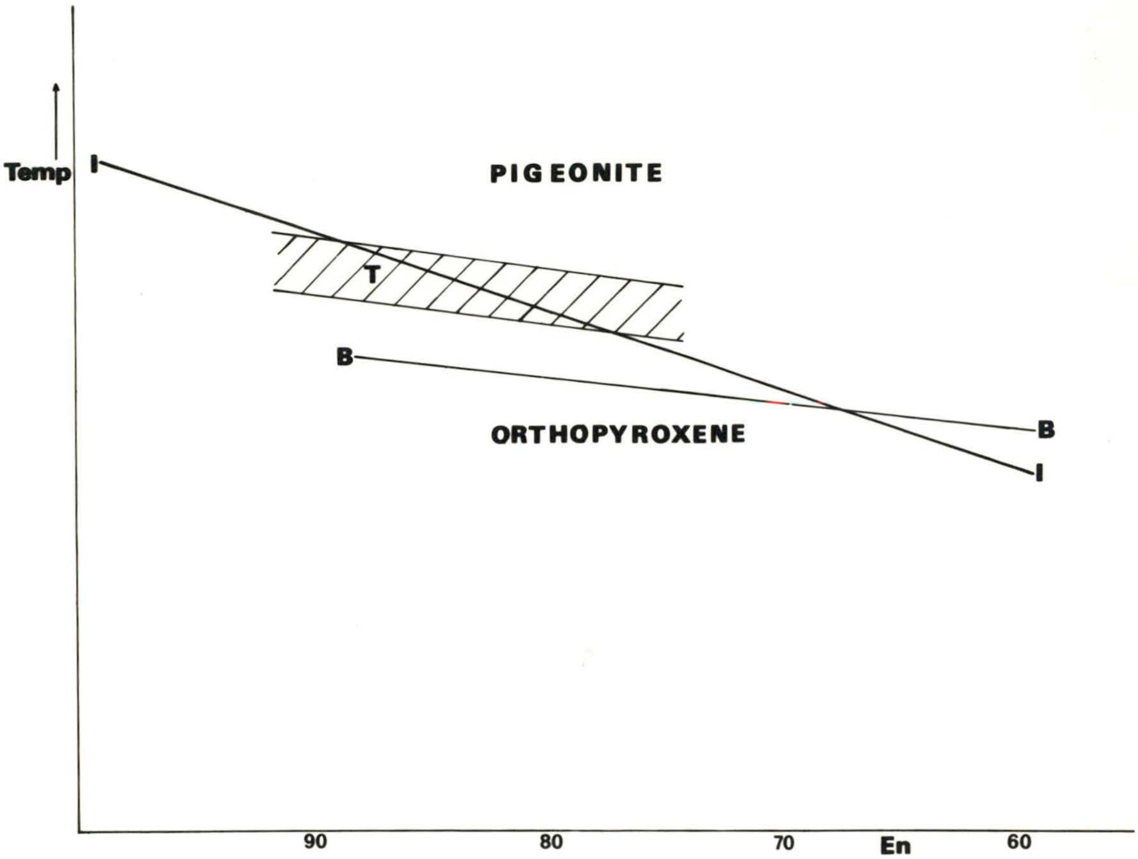
It can be argued then, that pigeonite was formed at least during the crystallization of the more differentiated Brassey Hill Harzburgites, and perhaps much earlier if the interpretation of bleb lamellae is correct. On the presence of abundant bleb lamellae, the calcium-poor pyroxenes of the upper zone of the Caudrys Hill Pyroxenites may be inverted pigeonites, and similarly the calcium-poor pyroxenes of the transition rocks of the Serpentine Hill Complex. In the layered gabbros of the Serpentine Hill

Complex, bleb lamellae are abundant in orthopyroxenes while coexisting clinopyroxenes have both (001) and (100) lamellae, indicating that they coexisted with pigeonites which subsequently inverted. Similarly most orthopyroxenes occurring in intrusive gabbros and microgabbros can be shown to be inverted pigeonites.

The occurrence of inverted pigeonites of more magnesian compositions than in the Bushveld and Stillwater intrusions could mean that the liquidus temperatures of the magmas crystallizing the Tasmanian Main Belt ultramafics were relatively higher, as is illustrated in Figure 23. (Compare with Fig. 23 of Wager and Brown, 1968, and Fig. 12 of Hess, 1960). This conclusion must, however, be regarded as tentative since the relationships between the various calcium-poor pyroxene polymorphs is poorly understood, the influence of FeO, Fe_2O_3, Al_2O_3, etc. on the pigeonite-orthopyroxene inversion is not known, and the kinetics of the inversion have not been studied. The amount of undercooling between each crystallization burst could also be important; perhaps during crystallization of parts of the Heazlewood River Complex that the inversion temperature was close to the liquidus temperature, so that after substantial undercooling orthopyroxene would directly nucleate, but with less undercooling pigeonite would crystallize and soon after invert to orthopyroxene.

Another interesting point with regard to exsolution in pyroxenes is the possibility that calcium-rich pyroxene exsolved soon after crystallization of the calcium-poor host could migrate to the grain boundaries and form part of the interstitial clinopyroxene. The writer found no clear-cut cases where this could have occurred, but the textures

Figure 23. Hypothetical relationships between liquidus temperatures and pyroxene inversion. The line I-I represents the pigeonite-orthopyroxene inversion curve, while B-B represents the Bushveld pyroxenes (Hess, 1960). The field T represents crystallization of pyroxenes from the Tasmanian Main Belt complexes.



Temp
I
PIGEONITE
T
B
ORTHOPYROXENE
B
I
90
80
70
En
60

in 38625 (Plate 66) suggest this process could be possible. In this rock, small calcium-rich pyroxene blebs grade into larger ones, which are optically continuous with interstitial calcium-rich pyroxene. If this process does occur it would imply that "equilibrium" between co-existing pyroxenes in the Tasmanian rocks could in some cases possibly be subsolidus rather than strictly magmatic, and this would have to be taken into account in any interpretations based on the chemical compositions of the pyroxenes.

In some rocks, thin rods of a dark, high relief mineral have exsolved along a number of crystallographic directions in pyroxenes. This occurrence is rare in the layered ultramafics, but has been found in lherzolites and plagioclase lherzolites such as 38558 and 38568. The exsolution rods are very abundant in pyroxenes in some of the Serpentine Hill microgabbros (Plate 67). The mineral could not be positively identified, but is thought to be rutile rather than spinel, as it is anisotropic and the rods are similar to the exsolution rods of rutile described by Moore (1968). In contrast to the Giles Complex occurrence, however, in the Tasmanian rocks the rods are more abundant in clinopyroxene rather than orthopyroxene. Moore (1968) suggests that rutile exsolution may indicate relatively high pressure conditions, but this is precluded for the Tasmanian rocks, since these probably crystallized at relatively low pressures (see below).

2.2 Alumina content of pyroxenes

Discussions concerning the dependence of the alumina content of pyroxenes on pressure can be found in Green (1964) and O'Hara (1967). It is suggested by O'Hara (1967) that more Al should substitute in pyroxenes at higher pressures within the plagioclase lherzolite facies, although the amount of substitution should also depend to some extent on temperature.

In Figure 24, the Al_2O_3 content of pyroxenes from the Tasmanian complexes is compared with pyroxenes from the Bushveld and Stillwater intrusions (which crystallized under high-level crustal conditions) and with pyroxenes from the Giles Complex, which is believed to have formed by crystal accumulation under granulite facies conditions (Nesbitt *et al.*, 1970; Moore, 1971). Since the alumina content of pyroxenes also depends on magma type (Le Bas, 1962), this must be taken into account if any comparisons are to be made. However, Nesbitt *et al.* (1970) indicate that the composition of the chilled margin of the Mt. Davies intrusions of the Giles Complex is very similar to the Stillwater chilled margin, and that on the abundance of orthopyroxene the parent magma was almost certainly tholeiitic. Clinopyroxenes from the Bushveld, Stillwater, Giles Complex and Tasmanian bodies all plot in the tholeiitic field of the Al_2O_3/SiO_2 plot of Challis (1969).

Assuming the proposition concerning the pressure dependence of Al contents of pyroxenes is valid, the plot in Figure 24 is consistant with the field observations concerning the relative depths of crystallization of the Giles Complex compared with the Bushveld and Stillwater intrusions, and suggests that the layered rocks of the Tasmanian Main Belt

Figure 24. Al_2O_3 contents of pyroxenes from the Main Belt complexes and other layered intrusions. Data from the Giles complex are taken from Moore (1971), the Bushveld Complex from Atkins (1969), and the Stillwater Complex from Hess (1960) and Wager and Brown (1968).

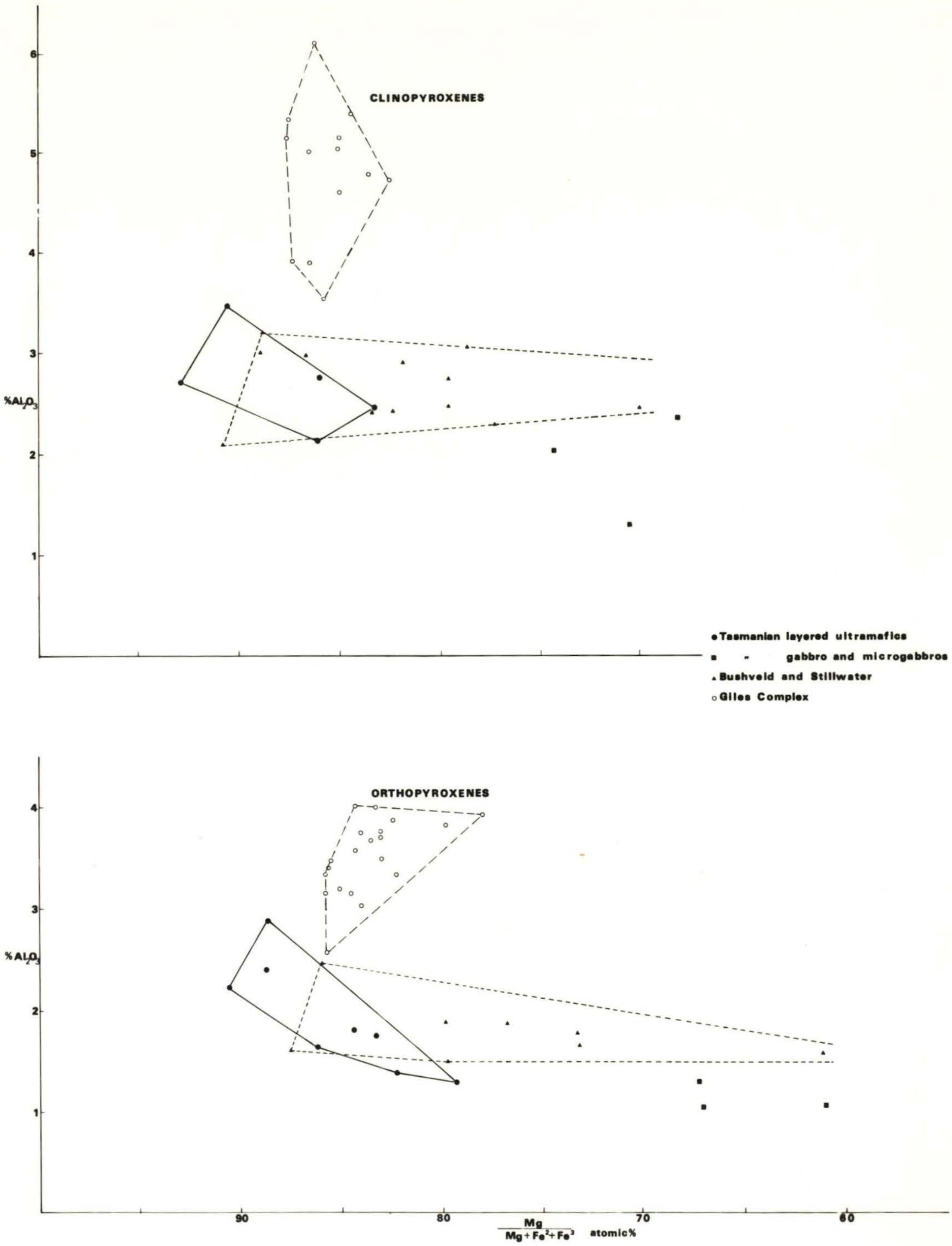
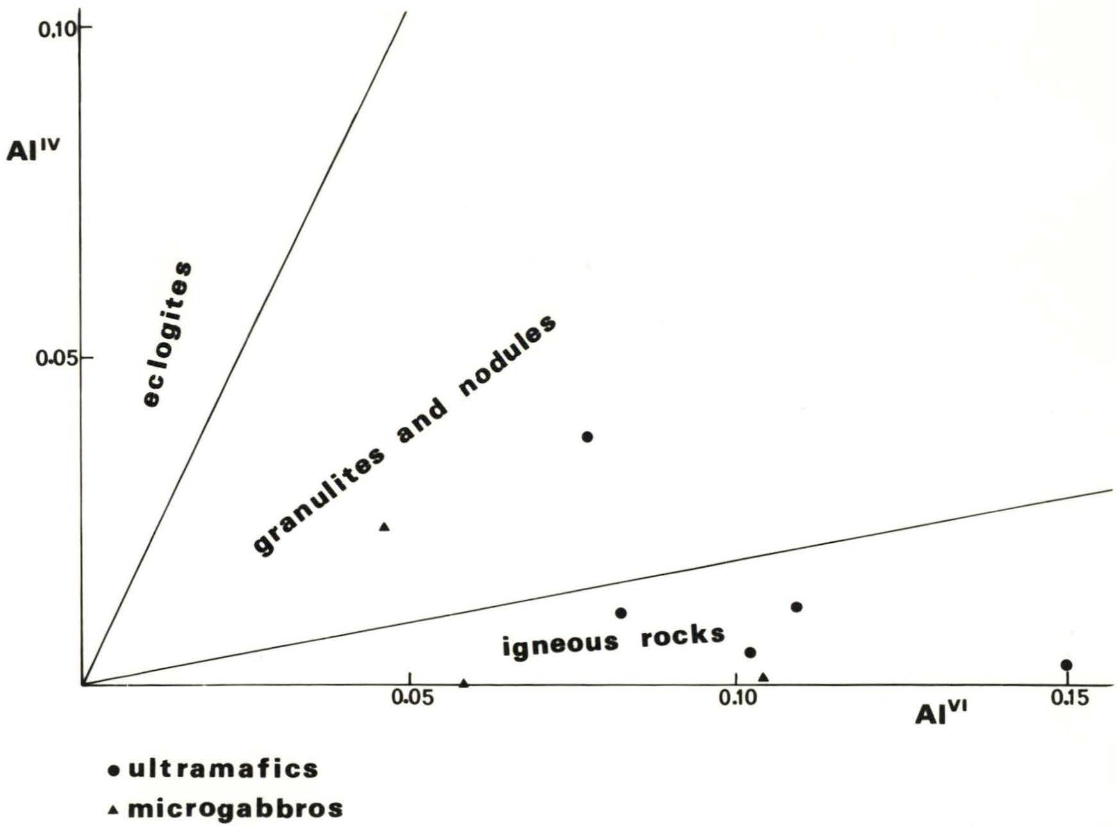

CLINOPYROXENES
%Al_2O_3
6
5
4
3
2
1
Tasmanian layered ultramafics
- gabbro and microgabbros
Bushveld and Stillwater
Giles Complex
ORTHOPYROXENES
%Al_2O_3
4
3
2
1
90
80
70
60
$\frac{Mg}{Mg+Fe^{2}+Fe^{3}}$ atomic%

Figure 25. Al^{IV} versus Al^{VI} plot for clinopyroxenes from the Tasmanian Main Belt complexes. The various fields are taken from Aoki and Kushiro (1968).



0.10
0.05
Al^{IV}
eclogites
granulites and nodules
igneous rocks
0.05
0.10
0.15
Al^{VI}
• ultramafics
▴ microgabbros

complexes crystallized at a depth comparable more with the Bushveld and Stillwater intrusions than with the Giles Complex. The relatively low Al^{IV} component in the Tasmanian clinopyroxenes is also fairly consistent with this view, for with the exception of two Serpentine Hill pyroxenes they fall within the igneous field in the Al^{IV}/Al^{VI} plot of Aoki and Kushiro (1968). The Giles Complex clinopyroxenes fall within the granulite field (Moore, 1971).

As previously mentioned, the Tasmanian pyroxenes and those from the Bushveld are similar with regard to their minor component contents (Cr_2O_3, NiO, TiO_2, etc.). The Giles Complex pyroxenes are also similar, except that the Cr_2O_3 content is higher; this might be expected if Cr and Al were competing for the same octahedral sites.

There are a number of objections and other factors that must be considered in relation to Al_2O_3 content and pressure. First of all is the effect of temperature. Anastasiou and Seifert (1972) believe on the basis of experimental work that the solid solution of Al in enstatite is a function of temperature and is not strongly pressure-dependent. The present writer has found it very difficult to reconcile this view with the empirical relationships deduced from Figures 24 and 25, and would prefer to see experimental work done on the solid solutions involved in the system orthopyroxene - clinopyroxene - plagioclase. Another factor which could be important if the temperature dependence is the more critical, is the possibility of subsolidus adjustment of Al contents of pyroxenes during slow cooling. With regard to this, no definite spinel or other aluminous phases have been observed as exsolution bodies in the Tasmanian pyroxenes, as is the case for some pyroxenes from the Giles Complex (Moore, 1971). However, it will be important to check that the rod lamellae exsolved from some pyroxenes are really rutile and not a spinel.

2.3 Partition coefficients between coexisting pyroxenes

K_D values for the partitioning of Mg and Fe between coexisting orthopyroxenes were calculated according to the equations of Kretz (1963). They fall in the range 0.75-0.95 for ultramafic pairs and 0.87 and 0.90 for the two microgabbro pairs. Bushveld pyroxenes give a range of 0.64 to 0.70, and according to the data in Kretz (1963) this could imply higher liquidus temperatures for the Tasmanian rocks. Ultramafic nodules in basalts and alpine ultramafics have K_D values in the range 0.34 to 2.00 (Kretz, 1963).

Interpretation of K_D values cannot, however, be taken very seriously. They could be affected by such factors as the Al_2O_3, Fe_2O_3, Na_2O etc. contents of the pyroxenes and by subsolidus "re-equilibration". Errors of up to 10% would be involved in the Tasmanian rocks as exsolution lamellae were not properly accounted for. Also, it is quite likely that kinetics, diffusion, and lattice defects have a strong influence on partitioning (C. Cuff, *pers. comm.*), so that interpretation in accordance with a simple thermodynamic model may not be justified.

3. Chromites

The compositions of two chromite samples are given in Table 5. Two other partial analyses from the Heazlewood River Complex, done at the Tasmanian Department of Mines laboratories, indicate similar types of spinel compositions, namely fairly high Cr_2O_3 (greater than 40%), moderate total Fe as FeO (25-35%) and relatively low Al_2O_3 (around 10%).

The compositions of the two chromite samples are plotted on Figure 26. However, a lot more analyses are necessary in order to characterize the spinels occurring in any complex (see Thayer, 1970; and Loney *et al.*, 1971).

Four chromite samples from Adamsfield (Brown, 1962) are also plotted in Figure 26. The Adamsfield chromites are characterized by unusually low Al_2O_3 and high Cr_2O_3, and their significance in relation to the origin of the Adamsfield Complex is discussed in another section.

4. Plagioclase compositions

As previously mentioned, plagioclase from the layered ultramafics and gabbros have labradorite compositions from albite twins. The only samples of ultramafics in which plagioclase grains are relatively unaltered are generally the more differentiated types, but an unaltered norite layer low in the Serpentine Hill sequence also gave a labradorite composition (An68; see Appendix 1). These plagioclase compositions are low in An compared with those from the ultramafic zones of the large layered intrusions, in which most of the plagioclase is also postcumulus (Wager and Brown, 1968). The present writer can offer no explanation for the unusual plagioclase compositions in the Tasmanian rocks.

Table 5

Chromite compositions.*

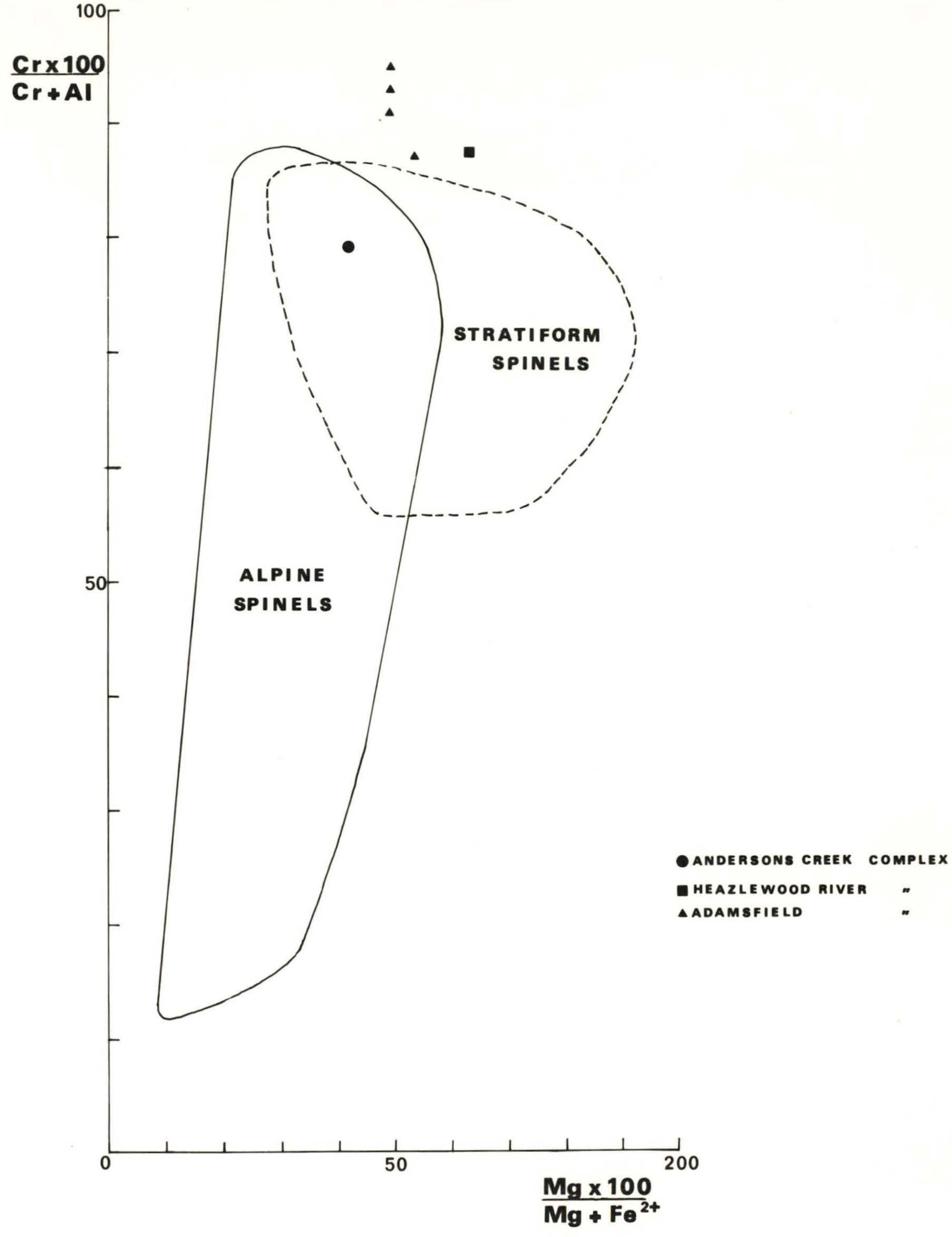
	38905**	Andersons Creek
MgO	7.1	9.4
Al_2O_3	5.5	10.0
Cr_2O_3	57.1	58.0
'FeO'	24.8	18.7
MnO		0.3
	94.5	96.4

Formulae on the basis of four oxygen atoms (total Fe was distributed between the two sites and the formulae recalculated as Fe^{2+} and Fe^{3+}).

Mg	0.383	0.479
Fe^{2+}	0.632	0.503
Mn		0.008
	1.015	0.990
Cr	1.637	1.570
Fe^{3+}	0.120	0.033
Al	0.234	0.413
	1.991	2.016

* Samples analysed by Tasmanian Department of Mines, Launceston.
** Chromitite from Nineteen Mile Creek Dunites.
'FeO' - total Fe as FeO.

Figure 26. Cr x 100 / Cr + Al versus Mg x 100 / Mg + Fe^{2+} for Tasmanian chromites. The fields for alpine and stratiform chromites are taken from Irvine (1967), and the Adamsfield data are taken from Brown (1972).



100
Cr x 100
Cr + Al
50
0
50
200
Mg x 100
Mg + Fe2+
STRATIFORM SPINELS
ALPINE SPINELS
ANDERSONS CREEK COMPLEX
HEAZLEWOOD RIVER "
ADAMSFIELD "

SIGNIFICANCE OF THE COMPOSITIONAL DATA

The iron enrichment trends shown by the bulk chemical compositions of the rocks and compositions of pyroxenes and olivines are consistent with the cumulative origin proposed on textural and structural grounds for most of the ultramafics and layered gabbros of the Main Belt complexes. The variations in bulk chemistry of the ultramafics are similar to those shown by the ultramafic zones of the Bushveld, Stillwater and in particular the Great Dyke intrusions. With regard to alpine ultramafics, the bulk compositions of the Main Belt complexes show greatest resemblance to the cumulate ultramafics and gabbros and the intrusive gabbros within ophiolite complexes.

Exsolution information indicates that pigeonite was the liquidus calcium-poor pyroxene during the crystallization of at least some or perhaps most of the Brassey Hill Harzburgites, and also during the crystallization of the upper zone of the Caudrys Hill Pyroxenites, and the transition rocks and layered gabbros of the Serpentine Hill Complex. The crystallization of pigeonites of higher Mg/Mg+Fe ratios than indicated by the Bushveld or Stillwater sequences, may mean that the parent magmas for the Main Belt layered rocks has higher liquidus temperatures than the Bushveld and Stillwater magmas.

The pyroxenes from the Main Belt layered rocks are fairly similar in composition to those from the ultramafic zones of the Bushveld and Stillwater complexes. The Al_2O_3 and Al^{IV} contents of the ultramafic pyroxenes are similar to Bushveld and Stillwater pyroxenes, and lower than for pyroxenes from the Giles Complex which is believed to have formed by crystallization under granulite facies conditions. Empirically this suggests that the Tasmanian Main Belt rocks crystallized at relatively shallow crustal depths, roughly of the same order of depth of crystallization as the Bushveld and Stillwater intrusions. However, the relative importance of temperature and pressure in determining the Al_2O_3 content of pyroxenes is not fully known as yet.

1. <u>The number of parental magmas giving rise to the Heazlewood River ultramafics</u>

It was indicated in chapter 3 that it was difficult to decide from the field evidence whether or not the main subdivisions of the Heazlewood River ultramafics (i.e. the Nineteen Mile Creek Dunites, the Brassey Hill Harzburgites, the Caudrys Hill Pyroxenites and the pyroxenites of the Bronzite Hill area) are fault blocks of a layered sequence which crystallized from a single parent magma, or whether they are blocks from different intrusions that have been tectonically juxtaposed. This problem is again looked at in the light of the compositional data.

At least some of the Nineteen Mile Creek Dunites fall in the compositional range of cumulates. Also, the chromitite 38905 has a cumulate-like texture. So some of the Nineteen Mile Creek Dunites may be cumulates, and perhaps could be representative of the basal zone of the Brassey Hill Harzburgites and/or the Caudrys Hill pyroxenites. However, it is also possible that a large proportion of the Nineteen Mile Creek Dunites are residual upper mantle rocks rather than deformed cumulates.

Compositionally, rocks from the Brassey Hill Harzburgites and the Caudrys Hill Pyroxenites show a fair degree of overlap. The rocks within and immediately adjacent to the complex fault zone between these two sequences appear to represent a transition. However, if these two groups are lateral equivalents of each other, one would have to postulate drastic diffusive and convective processes to explain why a thick sequence of orthopyroxene cumulates formed in one part of the magma chamber at the same time as a thick sequence of olivine and orthopyroxene cumulates formed elsewhere in the chamber. Another possibility is that the two groups are juxtaposed blocks representing different horizons of a sequence which crystallized within a single magma chamber; a sudden influx of a large body of fresh magma into the chamber after crystallization of one group could explain why the second group gives considerable overlap in Mg/Mg+Fe ratios. A third possibility is that the Brassey Hill Harzburgites and the Caudrys Hill Pyroxenites each crystallized from separate magmas.

It is not known, then, whether the layered ultramafics of the Heazlewood River Complex all crystallized from a single parent magma, from a series of magmas injected into a single magma chamber, or from a series of magmas each of which crystallized as a separate entity.

PARENTAL MAGMA TYPES AND DEPTH OF CRYSTALLIZATION

1. Possible parental magmas

Because of the similarities regarding rock types and textures with the ultramafic zones of the Bushveld and Stillwater intrusions, it is possible that the Main Belt ultramafics and layered gabbros also crystallized from tholeiitic magmas. Possible parental magmas for the Bushveld and Stillwater intrusions are given in Wager and Brown (1968, Table 26), and vary from a quartz-normative tholeiite to an olivine-normative tholeiite.

Because of the presence of inverted pigeonite at more magnesian compositions than in the Bushveld and Stillwater Complexes, it is possible that the liquidus temperatures for the Tasmanian rocks were higher than the latter. Thus the composition of the parent magma may have been a high-Mg tholeiite. Wager and Brown (1968) suggest on the basis of olivine compositions as high as Fo94 and the disproportionate amount of gabbroic differentiates, that the parental magma of the Great Dyke might be more picritic than the Bushveld and Stillwater magmas. The minerals analysed from the Main Belt rocks are not quite as magnesian as those at the bottom of the Great Dyke, but the bulk compositions of some rocks are (Fig. 19), and it must be remembered that the base of none of the Tasmanian ultramafic sequences is exposed. Perhaps the parental magmas for the Main Belt rocks were similar in composition to intrusive gabbros such as 38639, or even the high-Mg basalt specimen 38876 (see following chapter). Such high-Mg basaltic rocks are quite common in Archaean Greenstone belts (Viljoen and Viljoen, 1970; Hallberg and

Williams, 1972), and are known from younger rocks (Dallwitz *et al.*, 1966). All of the intrusive gabbros and microgabbros analysed have MgO/Mg+Fe ratios higher than the parental magmas of the Bushveld and Stillwater intrusions, and these compositions can also be considered as possibilities for parental magmas for the Main Belt layered rocks.

As previously mentioned, it has not been demonstrated conclusively that the layered two-pyroxene gabbros in the Serpentine Hill and Andersons Creek bodies crystallized from the same magmas as the layered ultramafics, although the field evidence strongly suggests that this is the case. Because of their close similarity in petrography and chemistry with intrusive gabbros and microgabbros, it is possible that the layered gabbros crystallized from relatively large bodies of magma which intruded the layered ultramafics.

2. Relevant experimental work

Experimental work relevant to the origin and crystallization of tholeiitic basalts is discussed in Green and Ringwood (1967), Green *et al.* (1967) and Green (1969). According to these results, olivine but not orthopyroxene occurs on or near the liquidus of quartz tholeiitic compositions at low pressures (Green *et al.*, 1967). However, at greater than 9 kb, orthopyroxene and/or clinopyroxene replaces olivine as the liquidus phase (Green and Ringwood, 1967). This would appear to be inconsistent with the fact that orthopyroxene, as well as olivine is a cumulus phase at the bottom of the ultramafic zones of the Stillwater and Bushveld Complexes, which certainly crystallized at less than 9 kb. The saturated quartz tholeiite composition studied in Green *et al.* (1967)

is in fact not very different from the compositions of chilled margins of the large layered intrusions. The fact that orthopyroxene can appear as an early liquidus phase in quartz tholeiitic magmas at low pressures is also demonstrated by the Tasmanian Jurassic dolerites, thick sills of which commonly have an orthopyroxene-bearing zone just above the lower chilled margin (Edwards, 1942; McDougall, 1962). It is possible that subtle compositional differences could determine whether or not orthopyroxene appears as an early phase in tholeiitic rocks at low pressures; for example, the compositional differences between the Tasmanian Jurassic dolerites and the Palisades Sill parental magmas are slight (Walker, 1969), but olivine, which is unknown in the Tasmanian dolerites, appears before orthopyroxene in the Palisades sequence. In large magma chambers, the thermal gradients throughout the magma are probably quite complex, and such factors as undercooling, ease of nucleation, and the effects of convection currents are undoubtedly critical in determining what mineral is nucleating and accumulating at any given time and place. In fact, it could be expected that the results obtained in an experimental capsule might not necessarily be consistant with what actually occurs in a crystallizing magma of similar composition in a far more complex natural environment. It is concluded, then, on the basis of observations of natural rocks rather than experimental work, that orthopyroxene as well as olivine can be precipitated from tholeiitic magmas at low pressures.

Tholeiitic magmas (probably including high-Mg tholeiites) are believed to form by partial melting of upper mantle material at pressures less than 9 kb, or by more extensive partial melting at pressures greater than 9 kb (Green *et al.*,1967; Green and Ringwood, 1967; Green, 1971, especially Fig. 3).

Olivine-plagioclase reactions, investigated by Kushiro and Yoder (1966) and Green and Hibberson (1970) are very important regarding the depth of crystallization of the Main Belt complexes. The latter authors included runs on Fo + An and Fo(92) + labradorite (An59) in their investigations, the second of these being particularly relevant in the case of the Tasmanian ultramafics. The results show that olivine + plagioclase is unstable about above 9 kb at 1100^{o}C, and are replaced by the assemblage olivine + orthopyroxene + clinopyroxene + aluminous spinel. The reaction has a positive slope on a T/P phase digram, so that an olivine + plagioclase assemblage would react on cooling at constant pressure, assuming that the temperature of the reaction boundary is sufficiently high for the reaction to be kinetically viable. Green and Hibberson (1970, p.220) state that "an olivine tholeiitic or picritic magma, crystallizing in the lower crust at depths corresponding to 6-10 kb, might yield accumulates containing olivine, pyroxenes and plagioclase. Cooling of such accumulates at constant pressure would result in incompatibility of olivine and plagioclase, while temperatures were quite high (800-1000^{o}C). The extent of reaction would be kinetically controlled largely by the cooling rate of the body. It is considered that these conditions would lead to corona formation, often of concentric multistage type ..... , as the olivine and plagioclase passed successively through stability fields of pyroxenes + spinel, pyroxenes + spinel + garnet, pyroxenes + garnet." No such reaction coronas exist in rocks from Tasmanian complexes in which olivine and plagioclase coexist, so they probably crystallized at pressures of around 6 kb or less. Reaction coronas around olivine in contact with plagioclase have been observed by the writer in some of the Macquarie Island non-layered gabbros (Varne and Rubenach, 1972). They also occur in the Giles Complex accumulates (Nesbitt *et al.*, 1970), which are discussed above in regard to the alumina contents of pyroxenes. The presence of these reaction coronas in the Giles Complex accumulates is consistent with the view that the relatively high Al_2O_3 contents of the Giles Complex pyroxenes (compared with the Bushveld and Stillwater pyroxenes) is due to crystallization at relatively higher pressures.

It is concluded, then, that with the possible exception of the Nineteen Mile Creek Dunites, the ultramafics and layered gabbros of the Main Belt complexes crystallized from tholeiitic or migh-Mg tholeiitic magmas at relatively shallow depths corresponding to pressures of less than 6 kb.

3. Origin of the Adamsfield ultramafics

The origin of the Adamsfield ultramafics is an intriguing problem, and is discussed in Brown (1972). The relatively undeformed layered ultramafics are quite possibly cumulates on their textures, layering style and bulk compositions. However, the layered ultramafics overlap in composition with the more deformed massive pyroxenites and dunites, many of which, however, are most probably too magnesian to form from mafic magmas at low pressures. If they crystallized from a magma at higher pressures (i.e. in the upper mantle), the magma would have to be of ultramafic composition and considerably depleted in alumina to explain the very low alumina contents of both the orthopyroxenes and the coexisting chromites. The massive ultramafics on the other hand could be residual upper mantle, as is suggested by Ragan (1963) for the Twin Sisters Dunite. The latter also has low-Al chromites (Irvine, 1966), although not as low as in the Adamsfield rocks.

CHAPTER 6

THE HEAZLEWOOD RIVER OPHIOLITE COMPLEX AND ITS TECTONIC SIGNIFICANCE

INTRODUCTION

In this chapter, consideration is given to the dolerites, volcanic rocks and acid intrusive bodies which, as well as the ultramafics and gabbros, form an integral part of the Heazlewood River Complex. The various rock types, their structural relationships and chemical compositions are compared with other ophiolites and rocks comprising oceanic crust in order to critically evaluate the hypothesis that ophiolites form at mid-ocean ridges and are tectonically emplaced into orogenic belts.

In an attempt to explain the origin and emplacement of the ophiolites and other aspects of the Eocambrian and Cambrian geology of Tasmania, a number of models based on plate tectonics concepts are discussed in the final sections of this chapter.

OPHIOLITES AND THEIR TECTONIC SIGNIFICANCE

The term "ophiolite" was coined by Steinmann (1905, 1927) in his discussion of the association of ultramafics, gabbros, diabases, pillow lavas and cherts in certain mountain belts. From the late 1920's to late 1960's, the term was given a variety of meanings by different authors. Recently, a more restricted definition of "ophiolite" has become fashionable, and the present writer concurs with most aspects of the definition as used by those present at the Geological Society of America Penrose Conference on Ophiolites (1972).

According to this definition, the lower part of a completely developed ophiolite is composed of ultramafics "... consisting of variable proportions of harzburgite, lherzolite and dunite, usually with a metamorphic tectonite fabric (more or less serpentinized)."* This is overlain by a "gabbro complex, ordinarily with cumulus textures, commonly containing cumulus peridotites and pyroxenites and usually less deformed than the ultramafic complex." The gabbros and ultramafics are in part intruded by and largely overlain by a mafic sheeted dyke complex which passes up into a mafic volcanic complex, commonly consisting of pillow lavas. Minor sodic extrusives and intrusives of acid composition are commonly present. A complete ophiolite sequence is not necessarily represented in any particular body, and dismembering by faulting is common.

Following suggestions that alpine ultramafics could be slices of oceanic crust/upper mantle (Dietz, 1963; Hess, 1964), it was realized by Gass (1968) that the Troodos massif of Cyprus (Gass and Masson Smith, 1963; Gass, 1967) may be a slice of oceanic crust and upper mantle formed at a mid-ocean ridge. As the similarities between the rock types comprising ophiolites and rocks dredged from mid-ocean ridges became more obvious (Thayer, 1969), the interpretation of ophiolites and ultramafic gabbro complexes as having formed at ridges was accepted by many workers. At the same time, the popularity of sea-floor spreading was gaining

* Passages in parentheses are quoted from the "Ophiolite Manifesto" of the Penrose Conference. See "Geotimes", *Amer. Geol. Inst.*, *17*, No. 12, pp. 24-25 (Dec. 1972).

(Vine, 1966; Sykes, 1967; Le Pichon, 1968; Oliver *et al.*, 1969), as was the realization that many ophiolites have been emplaced as entire thrust sheets (Gass and Masson Smith, 1963; Davies, 1968, 1971; Reinhardt, 1969). Thus ophiolites fitted neatly into a scheme involving formation at ridges and emplacement as tectonic slices at or near subduction zones in models explaining orogenic belts in terms of plate tectonics (Dewey and Bird, 1970, 1971; Coleman, 1971; Church and Stevens, 1971).

It would appear that the same sort of magmatic and structural processes would be involved in the formation of both ophiolites and oceanic crust at a mid-ocean ridge. Certainly, the emplacement of sheeted dykes in some ophiolites on such a density that horizons consist of greater than 90% subparallel dykes must involve crustal extension on a scale envisaged in sea floor spreading (Moores and Vine, 1971). There is, then, a similarity between the geology of ophiolites and oceanic crustal models, such as those of Cann (1968) and Miyashiro *et al.* (1970), which are based on the petrology of samples dredged from the ocean floors, various geophysical data, and structural considerations involved in the emplacement of large volumes of magma into a tensional environment. Support for these models has been provided by studies of the geology and petrology of Macquarie Island, which is believed to be a slice of oceanic crust and upper mantle (Varne *et al.*, 1969; Varne and Rubenach, *in press*). The northern part of Macquarie Island is composed of ultramafics and gabbros intruded by sheeted dyke swarms, while the rest is composed of basaltic volcanics, in places intruded by abundant dykes. The ultramafics are mainly harzburgites, and the gabbros include a layered ultramafic-gabbro complex as well as a variety of non-layered types, at least some of which intrude the ultramafics. The petrography and chemistry of the dolerites and volcanics are quite compatible with the olivine tholeiites formed at mid-ocean ridges, and a reconstructed section through the island, based on observed intrusive relationships and the metamorphic grade of the extrusive rocks and dykes, is quite consistent with oceanic crustal models.

Having established that a particular complex in an orogenic belt is allochthonous and that it consists of a complete or near-complete ophiolite sequence, *prima facie* it should be a simple matter to compare the petrology and chemistry of the ophiolite rocks with oceanic crustal rocks. In particular, the volcanic rocks and dolerites could be examined in detail as mid-ocean ridge basaltic rocks fall in a narrow compositional range which is now fairly well characterized (Engel and Engel, 1965; Kay *et al.*, 1970; Cann, 1971; Nicholls and Islam, 1971). However, there are certain problems, and these are outlined below.

1) Is the spreading environment at a mid-ocean ridge necessarily unique? If marginal seas of inter-arc basins form by crustal spreading (Karig, 1971), then perhaps ophiolites form in these environments instead of at a mid-ocean ridge. What data are available suggest that the basaltic rocks formed in marginal sea environments could be similar to mid-ocean ridge basalts (Hart *et al.*, 1972).

2) Do rocks emplaced at pre-Cainozoic ridges or in marginal seas necessarily have to have the same compositions as those formed at Cainozoic equivalents? This question raises a number of interesting problems, some of which are discussed below.

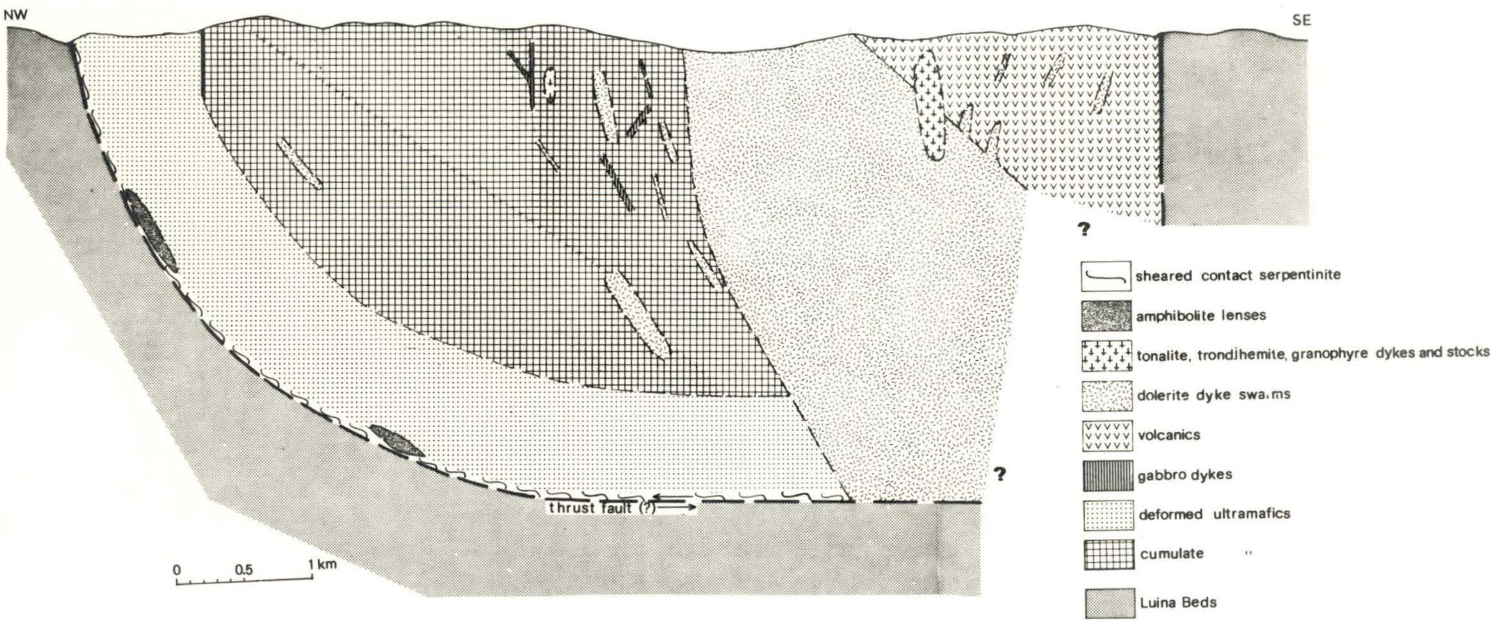
3) Ophiolite rocks are invariably metamorphosed, and changes in chemistry are likely to accompany metamorphism, even of the burial or hydrothermal types in which the rocks have suffered no penetrative deformation (Smith, 1968; Cann, 1969; Govett and Pantazis, 1971). This problem was realized by Moores and Vine (1971) who attempted to compare the dolerites and pillow lavas of the Troodos Complex with ocean floor basalts. However, because of the demonstrated relative immobility of Ti, Zr and Y during certain weathering and metamorphic processes (Cann, 1970), success has been attained in comparing ophiolite volcanics with rocks from various oceanic environments (Pearce and Cann, 1971); the relatively large and least dismembered ophiolites such as the Troodos and Oman complexes are fairly similar in Ti, Zr and Y contents to mid-ocean ridge basalts, and unlike seamount basalts and tholeiites or calcalkaline rocks from island arcs.

THE HEAZLEWOOD RIVER COMPLEX AS AN OPHIOLITE

1. Summary of important field relationships

In previous chapters it was established that the ultramafics of the Heazlewood River Complex are intruded by sheeted microgabbro and dolerite dykes, and that in the area immediately east of the Brassey Hill Harzburgites, the dolerites are probably dyke swarms intruding the ultramafics and part of the volcanic sequence. The volcanic rocks, mainly massive lavas and fragmental types, are intruded by dolerite dykes in some localities, while the other patches appear relatively free of dykes; a similar situation occurs on Macquarie Island (Varne and Rubenach, 1972)

Figure 27. Diagrammatic representation of the important structural relationships between various rock types in the Heazlewood River Complex. This section is based on the section AB of 12 and the appendix figure 3.1, but is largely hypothetical.



NW
SE
?
?
thrust fault (?)
0
0.5
1 km
sheared contact serpentinite
amphibolite lenses
tonalite, trondjhemite, granophyre dykes and stocks
dolerite dyke swa ms
volcanics
gabbro dykes
deformed ultramafics
cumulate ''
Luina Beds

and in the Troodos Complex (Gass and Masson Smith, 1963; Vine and Moores, 1971), and it is thought likely that many of the dykes were feeders for overlying extrusives. Tonalites, trondjhemites and granophyres intrude ultramafics, dolerites and volcanics.

In the areas mapped in more detail by the writer, the contacts between the Heazlewood River Complex and the Luina Beds are almost certainly tectonic. The contacts with the ultramafics are considered to be tectonic, as they consist of sheared serpentinites against sedimentary rocks which show no thermal metamorphism. The sedimentary rocks may, however, be sheared for a distance of up to one metre from the contact. To the SE of Mt. Cleveland, the contacts of the dolerites and volcanics with the Luina Beds are faulted. In the Jasper Hill area, the poorly exposed but confused jumble of volcanic rocks, sedimentary rocks, pyroxenites and dolerites could possibly represent a chaotic melange produced by thrusting or gravity sliding. East of the area mapped in detail by the writer, such bodies as the Whyte River Complex, the Magnet Spilites, and several separate bodies SE of Luina appear to be conformable sheets composed of ultramafics, dolerites and volcanics, and are thought to be thrust slices of ophiolite rocks rather than having formed *in situ*. Although the data are preliminary (see discussion below), there appear to be distinct petrographic and chemical differences between ophiolite rocks and other Eocambrian, Middle or Upper Cambrian basalts, dolerites and gabbros, which further supports the idea that the ophiolites are entirely allochthonous.

The Serpentine Hill Complex is a dismembered ophiolite which was probably emplaced as an allochthonous thrust sheet before the Middle Cambrian (see Appendix 1). In view of their similarities in rock types, textures and chemical compositions, it is possible that many or all of the Main Belt complexes are fragments of ophiolites; this question is discussed in more detail below.

2. Petrography of dolerites, extrusive rocks and acid intrusives

2.1 Metamorphism

The extrusive rocks, together with the dolerites, tonalites, trondjhemites and granophyres, have all suffered metamorphism which is essentially of the hydrothermal or burial type. So except near fault zones, the primary materials (largely pyroxenes and plagioclase) have been replaced in such a way that in many cases the original textures are fairly well preserved. These types of metamorphism depend on the availability of water as well as the temperature, pressure and other conditions and this may explain why there are patches, in some localities less than 1 m in size, which have suffered only slight alteration. Some dolerites intruding the ultramafics show hardly any alteration, while some dolerites and tonalities elsewhere are only partly altered. However, in the majority of specimens that were sectioned replacement of the primary minerals is complete or near complete.

The style of alteration, the mineral assemblages and their distribution are in general similar to those described by Smith (1968) for burial metamorphosed Ordovician lavas from New South Wales. Much of the outcrops would correspond to the grey domains of Smith, while relatively small patches and irregular veins rich in quartz and epidote would correspond to his yellow-green domains.

The most widespread assemblage consists of albite + tremolite-actinolite, with varying amounts of chlorite, quartz, epidote, fine-grained white mica, saussurite, and iron oxides. Albite, commonly accompanied by saussurite and/or very small flakes of white mica, has replaced the plagioclase of the original rock. Tremolite-actinolite has largely replaced pyroxenes, and varies from single grains pseudomorphing pyroxenes to acicular bundles commonly referred to as uralite. In thin section it ranges in colour from very light brown to darker green-brown varieties. It generally has a low extinction angle, but it is quite likely in some rocks that the amphibole is hornblende rather than tremolite-actinolite. The quartz in most cases is probably primary, but in some specimens the quartz is partly or largely secondary. The chlorite has in general replaced pyroxenes, but in some rocks has replaced what was probably intersertal glass or mesostatis in the original rock. Epidote occurs as small grains disseminated throughout some specimens, while in others it occurs in small partly radiating acicular masses. The iron oxides present include primary magnetite in some rocks, while in others they are finer-grained and probably result from alteration of the original pyroxenes. Vesicles have been filled by quartz, quartz + epidote, albite + quartz, chlorite, or chlorite + epidote + quartz assemblages. The assemblages and textures in these rocks show that they have suffered burial or hydrothermal metamorphism, probably to greenschist facies.

Some specimens of metamorphosed dolerite have been altered to amphibolite facies. In 38714 and 38919, the pyroxenes have been pseudomorphed by a light green-brown amphibole (hornblende) while the plagioclase, although partly altered to sericite and saussurite, is still labradorite (Plate 68). In 38727 the plagioclase is still labradorite but the green-brown amphibole which has replaced the pyroxenes is somewhat uralitic. Each of these specimens was collected within the main mass of dolerites just east of the Brassey Hill Harzburgites, where otherwise the above greenschist metamorphism is more prevalent.

A third type of metamorphism has affected most of the volcanic rocks north and NW of Jasper Hill, extending for at least as far as the Heazlewood River. The essential minerals are albite, chlorite and quartz, and varying amounts of carbonate (probably calcite) are present in most samples. In contrast to the above types, amphiboles are absent. Epidote and small amounts of secondary iron oxides are present in some specimens. The albite has replaced primary plagioclase, while the chlorite has replaced pyroxenes, glass or mesostasis. The carbonates are patchy, occurring as irregular grains, vesicle fillings, and scattered large grains (Plate 69). It is very difficult to tell how much of the quartz is primary and how much is secondary. Patches and anastomosing veins consisting largely of a mosaic of granular secondary quartz are common in many rocks (e.g. 38756, 38773). In some localities, whole patches of volcanic rocks (especially tuffs) have been silicified; remnants of original textures may occur in some of these (38770, 38774, 38776), but in other specimens these have been completely destroyed (38769).

The writer is uncertain as to how much of the intersertal quartz and isolated quartz grains occurring in the volcanic lavas is primary and how much was introduced during alteration. This is of course important in naming the rocks - if the quartz is primary the rock could be called a dacite on its chemical composition, but if it is secondary then it is probably silicified basalt. It is thought that most of the volcanics suffering this albite + chlorite + carbonate metamorphism are in fact silicified basalts, but some specimens may be altered andesites and dacites. Govett and Pantazis (1971) appear to have had similar problems in dealing with altered volcanics and dolerites in the Troodos Massif.

The albite + quartz + chlorite + carbonate metamorphism would differ from the other two types in that the PCO_2/PH_2O ratio was probably higher. The situation is perhaps analogous with experimental data given in Turner (1968, pp.135-142); tremolite tends to be unstable at relatively high PCO_2/PH_2O, but in the volcanics in question an assemblage of chlorite + carbonate + quartz formed instead of carbonate + quartz in the reactions discussed in Turner. In contrast to areas containing amphiboles, the chlorite + carbonate + quartz metamorphic assemblage occurs in a restricted area of volcanics, many of which are fragmental. The same area is the "chaotic melange" zone previously described, and metamorphism and silicification have also affected to some degree the pyroxenite blocks. Perhaps the greater permeability of this area due to complex faulting and the presence of fragmental volcanics provided easier access to fluids of relatively higher PCO_2.

2.2 Petrography of dolerite dykes

As mentioned above, the affects of hydrothermal or burial metamorphism on these rocks have been quite variable, some (38706, 38803, 38805, 38895) being remarkably unaltered whereas most have suffered partial to complete alteration to minerals such as chlorite, amphiboles, albite and epidote. It can be established from the large number of specimens sectioned that most (if not all) the dolerites are quartz tholeiites. With the exception of 38723, pigeonite or hypersthene (generally but not exclusively inverted pigeonite) occurs in all rocks which contain unaltered or partly altered pyroxenes.

The textures of the dolerites are typically subophitic, with interlocking laths of plagioclase (An 58-68) and larger pyroxenes (Plates 70, 71). However, some dolerites intruding the ultramafics (in particular 38704 and 38895) are intermediate in texture between the microgabbros described in chapter 3 and the normal subophitic dolerites.

Where present, quartz tends to be interstitial. It varies from 0% in specimens such as 38706, 38803 and 38805, to over 10% in some specimens, and there appears to be every gradation from quartz-free dolerites to granophyres and tonalites such as 38786. In the more altered samples, some of the interstitial quartz could be secondary, but in many cases granophyric intergrowths between quartz and albite indicate that the quartz is igneous and was not introduced or did not form during later metamorphism (Plate 72).

Primary iron oxide grains generally comprise less than 1% of any dolerite specimen. Some specimens contain scattered skeletal grains of magnetite (Plate 72).

Dolerites collected from the Whyte River Complex (38708, 38709) and from an ophiolite sheet south of Whyte Hill (38717) are petrographically indistinguishable from the above dolerites, except that 38708 and 38709 are in part brecciated.

It is concluded that the dolerites forming part of the Heazlewood River Ophiolite Complex are mainly quartz tholeiites, grading into tonalites and granophyres. They are petrographically dissimilar from mid-ocean ridge basalts and dolerites (which very rarely contain calcium-poor pyroxenes), and are quite unlike any of the dolerites from Macquarie Island which consist essentially of augite + plagioclase + titanomagnetite.

2.3 Petrography of tonalites, trondjhemites and granophyres

Most of these rocks are altered, with the pyroxenes replaced by chlorite or uralite and the plagioclase (or albite ?) altered to saussurite and fine-grained white mica. Specimen 38786 is the only one which contains some relict pyroxene (see analysis of augite in Table 4). It consists mainly of altered stubby plagioclase prisms, small augites partly replaced by amphibole and chlorite, and a little interstitial quartz. The pyroxenes occur as inclusions in the plagioclase, rather than in subophitic relationships as in the dolerites. Specimens 38779 and 38788 are similar to 38786, but are more altered.

Specimen 38787 is from a dyke intruding volcanics. It consists of elongate laths of albite up to 5 mm in length in a finer-grained groundmass of chlorite and albite (preserving a subophitic texture) and interstitial quartz.

Specimens 38781 and 38732 are granophyres consisting mainly of quartz and altered plagioclase (which could have been originally albite), with 10% chlorite which has replaced pyroxenes or amphiboles.

2.4 Petrography of extrusive rocks

Besides the obvious tuffs and agglomerates, extrusive rocks are labelled as such mainly on their fine grainsize and on the abundance of amygdales. Neither of the latter two criteria are of course satisfactory. (For example, abundant amygdales occur in dykes intruding pillow lavas on Macquarie Island). Therefore, many of the specimens described as extrusives may be dolerites, and vice versa.

In contrast to the dolerites, no primary plagioclase or pyroxenes have been identified in the extrusive rocks. Specimen 38749 is a typical basalt consisting of tremolite-actinolite, albite, and a small amount of iron oxides. The intergrowths of albite and amphibole are probably preserving primary subvariolitic intergrowths of plagioclase and pyroxene. Specimen 38753 is a similar rock in which the subvariolitic textures have been partly obliterated during metamorphism. Specimen 38754 consists of chlorite, albite, saussurite and epidote, in which subophitic textures are preserved. It contains scattered amygdales composed of quartz and epidote.

Specimen 38751 is from an agglomerate block, whose chemical composition is more similar to an andesite than a basalt (Table VIII). It consists of tremolite actinolite, albite, chlorite (after glass ?), amygdales of quartz of epidote, and has a general subophitic to intersertal texture. It could, however, be an altered basalt, with SiO_2 added during metamorphism.

Specimen 38764 consists of laths of albite and chlorite (after pyroxene), with a well-developed flow or trachytic texture. Scattered elongate pyroxene phenocrysts pseudomorphed by chlorite and small grains of quartz also occur. This rock has the chemical composition of a dacite (Table VIII), but again it is difficult to tell how much metasomatism accompanied the metamorphism.

Specimen 38876 is one of the unusual high-Mg basalts occuring in the Heazlewood River Complex. It is composed of euhedral pyroxene grains up to 2 mm long in a groundmass of chlorite and tremolite-actinolite (Plate 75). The pyroxenes, which may have included both calcium-poor and calcium-rich types, are replaced by tremolite-actinolite or chlorite. Abundant amygdales composed of chlorite or granular albite + quartz occur in the rock. Specimen 38875 is another high-Mg basalt, but is more highly altered.

2.5 Comparison with basalts and dolerites outside of the ophiolites

Specimens 38746 and 38747 are two basalts within the Luina Beds. They consists of augite (partly altered to chlorite) and albite with sub-ophitic textures, and contain titanomagnetite and intersertal glass replaced by chlorite and iron oxides. Specimen 38750 is a similar but more altered basalt from the Cleveland Mine. These basalts differ from the ophiolite volcanics in that primary iron oxides are much more abundant. Similarly 38736, which is from a dyke intruding the Luina Beds, contains scattered large iron oxide grains. Similar titanomagnetite-bearing dolerite and gabbro dykes are common in the Crimson Creek Formation in the Serpentine Hill and Wilson River areas, and all of these are petrographically unlike rocks within the ophiolites.

The extrusive rocks and dolerites from the Serpentine Hill Complex are generally highly altered, but resemble certain types from

Table 6

Chemical compositions of dolerites and volcanic rocks from the Heazlewood River Complex.

	38876	38803	38702	38704	38700	38701	38753	38751	38764	38766	38776	Mid-ocean ridge basalt
SiO_2	50.96	50.10	49.85	52.30	53.33	50.81	52.59	56.25	64.33	66.09	77.69	49.61
Al_2O_3	7.56	14.80	15.41	14.53	14.05	14.52	14.53	14.11	12.86	15.93	10.39	16.01
Fe_2O_3	1.38	n.d.	0.94	1.49	0.91	1.43	n.d.	n.d.	1.98	n.d.	n.d.	
FeO	7.62	7.51	8.19	7.87	8.83	11.14	8.82*	7.19*	5.04	4.69*	2.80*	11.49
MgO	17.78	12.22	8.04	8.17	6.87	6.42	9.08	6.63	4.90	2.76	0.93	7.84
CaO	6.92	11.90	10.98	11.85	9.20	9.15	11.39	5.26	2.49	0.38	0.32	11.32
Na_2O	0.60	1.37	2.22	1.40	2.74	1.92	1.23	5.71	5.37	6.97	5.52	2.76
K_2O	0.00	0.29	0.21	0.15	0.18	0.43	0.08	0.02	0.03	0.15	0.10	0.22
TiO_2	0.18	0.40	0.30	0.30	0.35	0.43	0.31	0.28	0.30	0.25	0.18	1.43
P_2O_5	0.00	0.07	0.00	0.00	0.07	0.02	0.01	0.06	0.06	0.05	0.04	0.14
MnO	0.16	0.12	0.06	0.12	0.14	0.22	0.14	0.09	0.03	0.01	0.00	0.18
NiO	0.14	0.01	0.02	0.01	0.01	n.d.	n.d.	n.d.	0.07	n.d.	n.d.	
Cr_2O_3	0.47	0.08	n.d.	0.08	0.05	n.d.	n.d.	n.d.	0.01	n.d.	n.d.	
Loss	6.24	1.08	3.03	2.11	3.08	3.48	1.78	4.05	2.65	2.82	2.05	
	100.01	99.86	99.25	100.38	99.81	99.97	99.96	99.65	100.12	100.10	99.42	101.00
Rb	<6	25	11	2	9	23	10	<6	<6	11	<6	0.6-12
Sr	15	105	79	28	77	64	44	62	46	50	51	40-250
Zr	<20	32	<20	8	15	29	<20	28	49	49	118	35-165
Y	<6	14	<6	10	8	6	8	<6	<6	<6	<6	10-60
Ni	1080	220	160	110	110	n.d.	n.d.	n.d.	100	n.d.	n.d.	30-200
Cr	3190	550	n.d.	530	345	n.d.	n.d.	n.d.	470	n.d.	n.d.	200-400

38876 - high-Mg basalt; 38803, 38702, 38704, 38700, 38701 - dolerites; 38753, 38751, 38764, 38766 - volcanic rocks; 38776 - silicified volcanic rock.
The Rb, Sr, Zr and Y values of 38704 were determined by Dr. R. Varne using X-ray fluorescence at the School of Environmental Sciences, University of East Anglia. The data for the mid-ocean ridge basalt is taken from Cann (1971) and Nichols and Islam (1971).

Table 7

C.I.P.W. norms for some dolerites.

	38803	38702	38704	38700
SI	0.00	0.28	6.02	4.32
OR	1.71	1.24	0.89	1.06
AB	11.59	18.79	11.85	23.19
AN	33.38	31.46	32.92	25.51
DI	20.33	18.71	21.06	16.21
HY	22.73	22.58	22.34	24.20
OL	8.08	0.00	0.00	0.00
MT	0.00	1.63	2.16	1.32
IL	0.76	0.57	0.57	0.67
AP	0.17	0.00	0.00	0.17

the Heazlewood River Complex.

3. Chemical compositions of dolerites, extrusive rocks and acid intrusives

Table VI lists major and some minor elements for dolerites and extrusive rocks from the Heazlewood River Complex, with data for mid-ocean ridge basalts added for comparison.

The high-Mg basalt is similar in composition to the clinoenstatite-bearing basalt from Cape Vogel area, Papua, which may or may not be related to the Papuan Ultramafic Belt (Dallwitz et al., 1966). It is also fairly similar to some of the high-Mg basalts which are abundant in the Archaean greenstone belts of South Africa and Western Australia (Viljoen and Viljoen, 1970; Williams, 1972).

C.I.P.W. norms for the least metamorphosed dolerites are given in Table VII. Specimen 38803 has normative olivine, while the rest have normative silica. The norms overlap with the more silica-saturated mid-ocean ridge basalts, which are less common than the olivine normative types (Kay et al., 1970; Cann, 1971). They are also similar to the saturated continental tholeiites (Walker, 1969, p.83), except that normative OR and IL are lower for the dolerites from the Heazlewood River Complex.

The dolerites and volcanic rocks are all relatively low in TiO_2 and P_2O_5 when compared with mid-ocean ridge basalts. They all have low or fairly low K_2O and Rb, but it must be emphasized that 38803 is the only unaltered rock. The rest of the dolerites are partly altered and all the extrusive rocks completely replaced by secondary minerals, and K, Rb and Sr are particularly mobile during metamorphism of igneous rocks

<u>Table 8</u>

Compositions of tonalite, trondhjemite and granophyre.

	38786	38784	38790
SiO	60.74	78.64	66.07
Al O	15.70	11.25	16.05
Fe_2O_3	n.d.	n.d.	n.d.
FeO	6.06*	1.44*	2.92*
MgO	4.15	0.60	3.89
CaO	4.09	0.55	3.16
Na O	4.30	6.45	6.65
K O	0.34	0.09	0.22
TiO	0.22	0.24	0.38
P O	0.17	0.01	0.10
MnO	0.06	0.00	0.05
Loss	3.23	1.00	1.13
	99.06	100.27	100.60
Rb	9	<6	13
Sr	158	48	46
Zr	26	31	33
Y	13	<6	<6
Ni	140	<20	90
Cr	170	35	80

* total Fe as FeO.

38786 tonalite

38784 granophyre

38790 trondhjemite

(for example, Cann; 1970). Excluding the high-Mg basalt which has a very high Cr and Ni, the Ni contents are within the range or slightly higher than mid-ocean ridge basalts, while the Cr contents (even for the "dacite" 38764) tend to be higher than mid-ocean ridge basalts. In terms of Cr and Ni contents the volcanic rocks and dolerites from the Heazlewood River Complex are quite unlike tholeiites and calcalkaline rocks from island arc regions, which typically have Cr values in the range 0-50 ppm and Ni values in the range 0-30 ppm (Jakes and White, 1972; Jakes and Gill, 1970). Thus with regard to Cr and Ni, the formation of the Heazlewood River Complex within an island arc environment, as is suggested for ophiolites by Jakes and Gill (1970), is considered to be unlikely.

The Ni and Cr contents of the tonalite and trondhjemite (Table VIII) fall outside, while those of the granophyre lie within the ranges of island arc rocks.

Analyses of some other Cambrian basic rocks are given in Table IX. The two rocks from the Serpentine Hill Complex are similar to the dolerites and volcanics from the Heazlewood River Complex, but the other four rocks, collected outside the ophiolites, have much higher TiO_2 contents, corresponding to the relatively higher titanomagnetite contents in these rocks compared with the ophiolite rocks. The majority of the Precambrian and Cambrian basaltic rocks whose analyses are recorded in Spry (1962b, Table IX) have TiO_2 contents greater than 0.5% and/or relatively high K_2O contents (greater than 0.4%). There are exceptions, however, and it will be interesting to compare the major and minor element

Table 9

Compositions of gabbro, dolerite and basalts from outside of the ophiolites, and of a dolerite and a basalt from the Serpentine Hill Complex.

	38747	31832	38815	40130	38811	38812
SiO_2	51.75	47.18	52.12	46.05	53.92	47.20
Al_2O_3	13.90	16.31	14.25	16.08	14.07	13.62
Fe_2O_3	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
FeO	11.45*	13.27*	12.41*	11.13*	12.73*	13.45*
MgO	4.87	5.43	4.71	3.30	5.21	7.04
CaO	6.91	7.70	6.54	13.63	9.44	4.72
Na_2O	5.44	2.97	4.14	3.26	1.55	1.70
K_2O	0.15	0.56	1.04	1.63	0.00	0.00
TiO_2	1.87	1.68	1.61	1.05	0.31	0.54
P_2O_5	0.22	0.12	0.11	0.09	0.01	0.03
MnO	0.17	0.24	0.11	0.25	0.25	0.18
Loss	2.86	4.23	3.45	3.92	2.57	12.41
	99.59	99.69	100.19	100.49	100.06	100.89
Rb	<6	34	29	74	<6	<6
Sr	186	189	332	326	41	68
Zr	120	37	71	10	13	63
Y	50	30	13	21	<6	<6
Ni	60	n.d.	<20	40	70	100
Cr	120	n.d.	80	185	140	140

* total Fe as FeO.

38747 basalt from the Luina Beds.
31832 gabbro intruding Crimson Creek Formation, Ring River.
38815 dolerite intruding Mt. Read Volcanics.
40130 basalt from the Mt. Read Volcanics (Anderson, 1972).
38811 dolerite, Serpentine Hill Complex.
38812 altered basalt, Serpentine Hill Complex.

compositions of the various suites of Upper Proterozoic and Lower Palaeozoic basaltic rocks in more detail.

In Figure 28, the Ti and Zr contents of dolerites, volcanic rocks and acid intrusives from the Heazlewood River and Serpentine Hill Complexes are compared with mid-ocean ridge basalts, island arc tholeiites, calcalkaline andesites from island arcs, and seamount basalts. These results indicate the depleted nature of the Tasmanian ophiolites with respect to Ti. The Zr values show a greater range, from less than the detection limits of the analytical procedures for some specimens to 118 ppm for the silicified volcanic rock 38776. (It is possible, that Zr could have been introduced during metasomatism of this rock.) The rocks from the Tasmanian ophiolites are, therefore, unlike rocks from the various oceanic environments with regard to Ti and Zr, and unlike most of the basalts and dolerites from the Oman and Troodos ophiolites which Pearce and Cann (1971) show to be similar to mid-ocean ridge basalts.

4. The Heazlewood River Complex and oceanic crust

In previous sections it is argued that, based on the rock types and their structural relationships, a good case exists for ophiolites being tectonic slices of oceanic crust (Gass, 1968; Thayer, 1969; Moores and Vine, 1971; Bonatti et al., 1971; Coleman, 1972; Varne and Rubenach, 1972, in press). This hypothesis can therefore be examined more closely by comparing the details of the petrography and chemistry of the rocks constituting any particular ophiolite with oceanic crustal rocks.

In many aspects, the petrography and chemistry of the dolerites and volcanics from the Heazlewood River Complex are at variance with mid-ocean ridge basalts. In particular, the dolerites and volcanics from the Heazlewood River Complex commonly contain calcium-poor pyroxenes and/or quartz, and these minerals are quite rare in mid-ocean ridge rocks. Also, the TiO_2 contents of the Tasmanian rocks are unusually low. On the positive side, dolerites and volcanics from the Heazlewood River Complex are relatively low in the incompatible elements as are mid-ocean ridge tholeiites, and their Cr and Ni contents are more similar to mid-ocean ridge basalts than to island arc rocks.

Since the incompatible elements should be strongly partitioned into a basaltic melt (Green & Ringwood, 1967), it is concluded by Green (1971) that Cainozoic mid-ocean basalts are probably partial melts of upper mantle rocks already depleted in these elements during an earlier partial melting event. Strontuim isotopes data from oceanic peridotites and mid-ocean ridge basalts are consistent with this hypothesis (Bonatti et al., 1971). As the rocks from the Heazlewood River Complex are also depleted in the incompatible elements it is possible that they too were melted from upper mantle rocks which had suffered an earlier partial melting event.

Relative to other basaltic rock types, tholeiites and quartz tholeiites are thought to be produced by partial melting of the upper mantle at shallow depths (less than 10 kb) or extensive partial melting at pressures greater than 10 kb. It is postulated, then, that the dolerites and volcanics of the Heazlewood River Complex could have formed in a spreading zone such as a mid-ocean ridge, but the magmas were produced at less than 10 kb pressure, or, relative to the typical olivine tholeiites of mid-ocean ridge basalts, by a greater degree of partial melting at pressures greater than 10 kb.

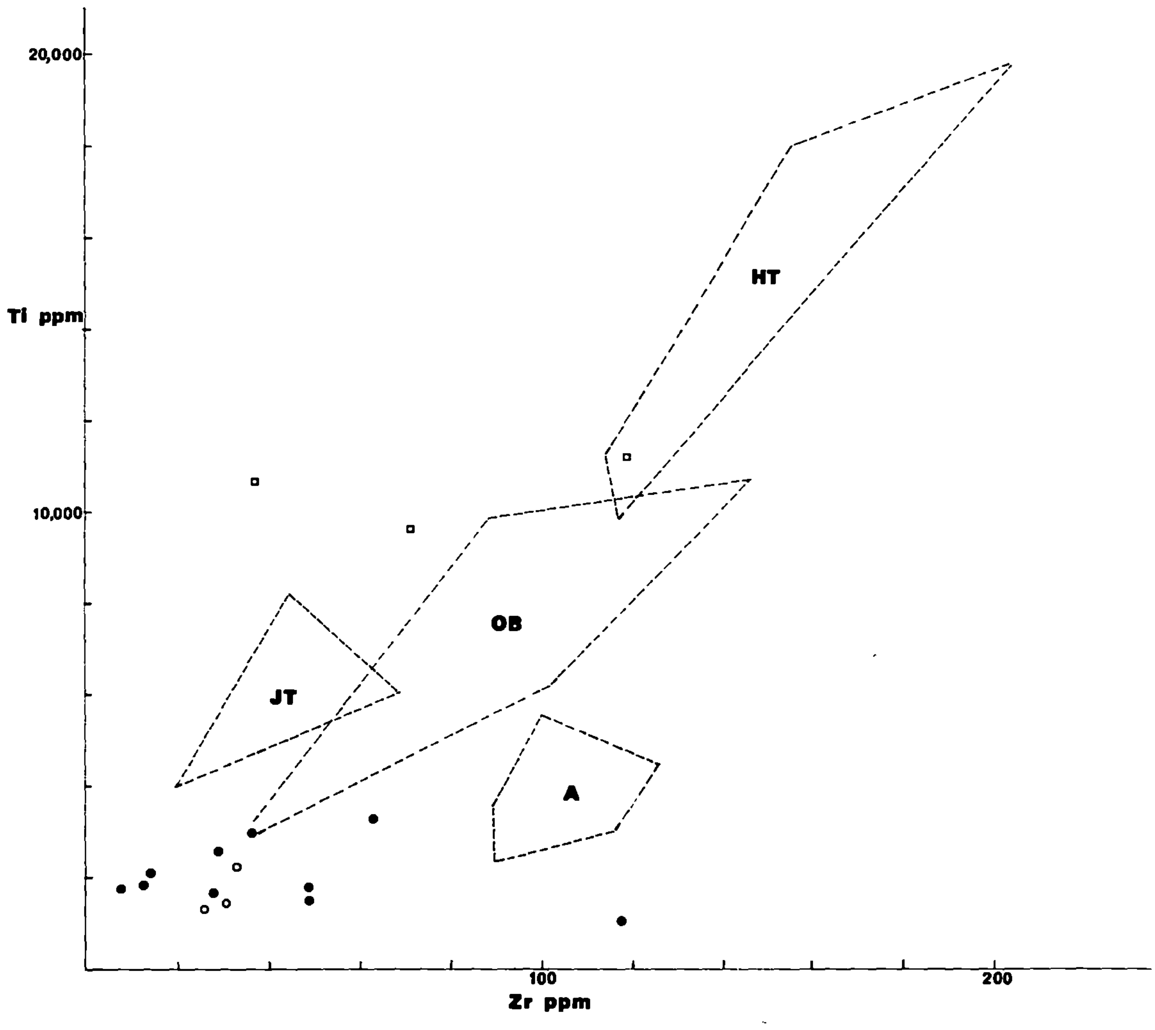

20,000
10,000
Ti ppm
HT
OB
JT
A
100
200
Zr ppm

Figure 28. Ti versus Zr for dolerites (dots), volcanic rocks (dots), sodic acid intrusives (open circles) and Tasmanian Cambrian basalts which do no belong to the ophiolites. The various fields are taken from Pearce and Cann (1971).

JT - Japanese tholeiites (island arc tholeiites)
OB - mid-ocean ridge basalts
A - island arc andesites
HT - Hawiian tholeiites

Despite the differences in petrography and chemistry of the dolerites and volcanics of the Heazlewood River Complex and mid-ocean ridge basalts, it is concluded that there is no necessary reason why the Heazlewood River Complex did not form at a mid-ocean ridge or a spreading environment similar to what is postulated for marginal seas. As very few data have been published from other ophiolites, it is not known whether or not the Heazlewood River Complex is unique in that unusual or special conditions operated during its formation in a spreading environment. It is possible that other ophiolites are appreciably different from Cainozoic mid-ocean ridge basalts, and it is possible that rocks formed in Phanerozoic spreading zones may be different from those formed in Cainozoic spreading zones.

5. Possible genetic relationships between the rock types comprising the Heazlewood River Complex

The association of extrusive rocks, dolerites, intrusive gabbros, layered gabbros and acid sodic intrusives with ultramafics in ophiolite complexes is now well known (Thayer, 1967; Davies, 1971; Coleman, 1971). Plotted on a AFM diagram these rocks fall into overlapping fields, which Thayer (1967) refers to as the "alpine mafic magma stem". That there must be some structural or tectonic control on this very common association appears to be highly likely, but are the various rock types related by fractional melting processes, fractional crystallization processes, or are they genetically unrelated?

In chapter 5 it was argued that the layered ultramafics probably crystallized at a relatively high level in the crust from tholeiitic or high-Mg tholeiitic magmas. Since most of the intrusive gabbros, microgabbros, dolerites, and at least some of the extrusive rocks are tholeiitic in composition, the parent magmas giving rise to the cumulate ultramafics could have been similar in composition to any of these rocks. If so, it is possible that some of the most differentiated dolerites could represent magmas tapped off from a chamber in which olivine and orthopyroxene (the more abundant cumulus minerals) were precipitating to form the layered ultramafics. Thus if a less differentiated gabbro (such as 38639) or the high-Mg basalt (38876) are taken as parent magmas, could fractionation of olivine and orthopyroxene from these produce typical microgabbro, gabbro, dolerite or basalt compositions? It is obvious that this cannot occur within the Heazlewood River Complex itself for the gabbros and dolerites intrude the ultramafics. However it is conceivable that the dolerites and volcanic rocks represent magmatic fractions drawn off from a chamber in which cumulates similar to the ultramafics in the Heazlewood River Complex were being formed.

Possible genetic connections between ultramafics, gabbros and basalts in the Papuan Ultramafic Belt were investigated by Davies (1971) using a graphical method devised by Powers (1958). In this method, the compositions of the postulated parent magma, the cumulus phases fractionated, and the fractionated magmas are plotted on diagrams of oxide components versus MgO. Thus if Ol and Opx are the cumulus phase, and X is the parent, vectors are drawn along the Ol-X and Opx-X lines so as to reach the composition of the fractionated magma (see Davies, 1971, pp.32-38). If this can be done for each plot then it is possible that various rocks are related by fractional crystallization.

Figure 29. AFM diagram (weight percentage, total Fe as FeO) for the Heazlewood River Complex. A basalt and a dolerite from the Serpentine Hill Complex are also included, each being marked with an "S".

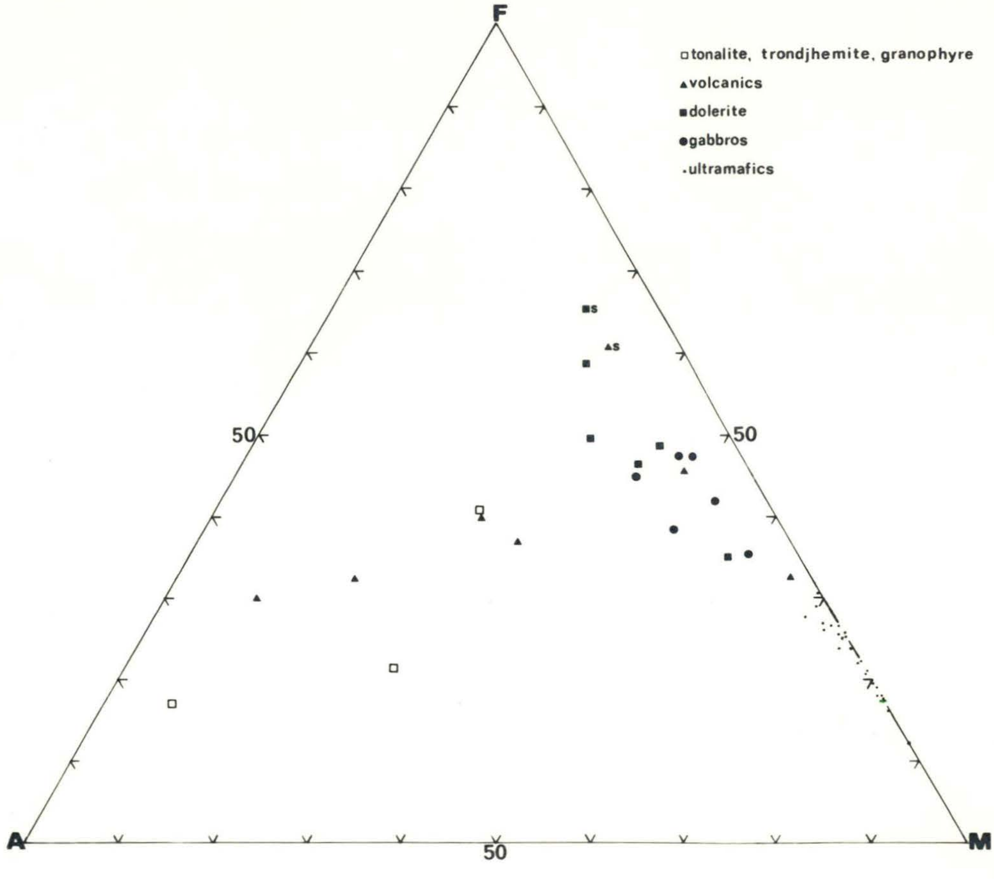

F
A
M
50
50
50
tonalite, trondjhemite, granophyre
volcanics
dolerite
gabbros
ultramafics
s
s

Table 10

Fractionation of high-Mg basalt.

	Composition of high-Mg basalt (water removed).	Composition after removal of 33% orthopyroxene.	Composition after removal of 17% orthopyroxene, 17% olivine, 2.5% plagioclase, 2.5% clinopyroxene and 0.7% chromite.
SiO_2	54.38	52.74	57.59
Al_2O_3	8.67	12.01	12.21
Fe_2O_3	1.47	1.79	1.92
FeO	8.13	8.04	7.14
MgO	18.97	12.90	9.20
CaO	7.38	10.50	10.15
Na_2O	0.64	0.92	0.85
K_2O	0.00	0.00	0.00
TiO_2	0.19	0.25	0.27
P_2O_5	0.00	0.00	0.00
MnO	0.17	0.18	0.16
NiO	0.14	0.12	0.13
Cr_2O_3	0.47	0.56	0.11

Table 11

Compositional ranges of gabbros and dolerites.

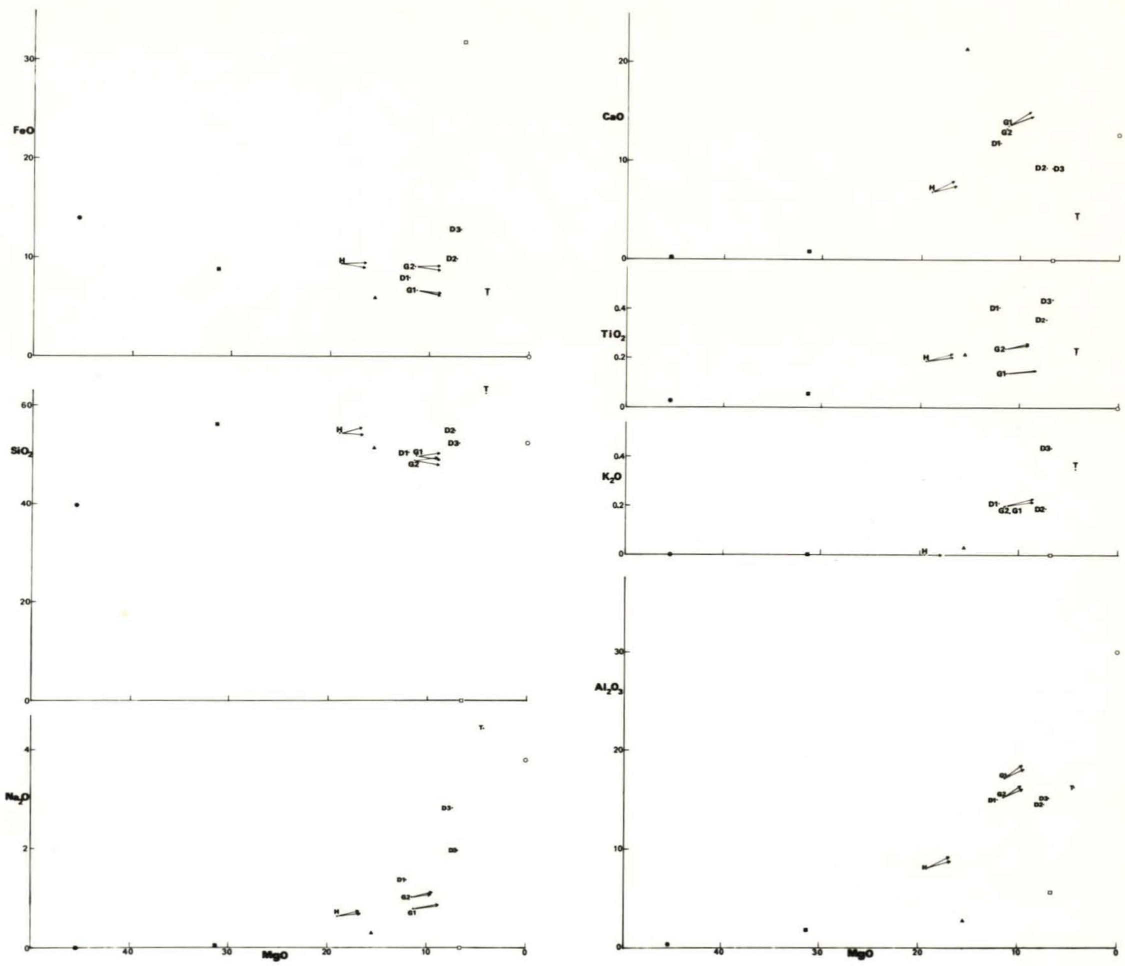
	Gabbros and microgabbros	Dolerites
SiO_2	49-52	50-54
Al_2O_3	15.0-16.9	14.0-15.5
'FeO'	6.8-9.9	7.5-12.5
MgO	8.7-11.2	6.5-12.3
CaO	11.7-13.5	9.3-11.9
Na_2O	0.8-2.5	1.4-2.8
K_2O	0.00-0.70	0.15-0.45
TiO_2	0.12-0.31	0.30-0.45
P_2O_5	0.00	0.00-0.07
MnO	0.08-0.15	0.06-0.14
Cr	80-1190 ppm	350-550 ppm
Ni	85-230 ppm	110-220 ppm
Rb	0-31 ppm	0-25 ppm
Sr	29-113 ppm	28-105 ppm

Figure 30. MgO diagrams for selected rocks. For details concerning construction and significance of this type of diagram, see Davies (1971).

The compositions used for olivine, orthopyroxene, clinopyroxene and plagioclase are "average" values for the Heazlewood River Complex. The arrows represent the direction of movement of the residual liquid composition; in this figure they do not represent vectors.

H - high-Mg basalt, 38876
G_1 - gabbro, 38639
G_2 - gabbro, 38642
D_1 - dolerite, 38803
D_2 - dolerite, 38700
D_3 - dolerite, 38701
T - tonalite, 38786

Total Fe is calculated as FeO. Values plotted are weight percent of the rocks recalculated water free.



FeO
SiO2
Na2O
CaO
TiO2
K2O
Al2O3
MgO
MgO

Figure 30 consists of MgO diagrams in which are plotted the "average" compositions of the component minerals of the ultramafics, together with the gabbros 38639 and 38642, the high-Mg basalt 38876, the dolerites 38700, 38701 and 38803, and the tonalite 38786. It was found that it is impossible to reach the compositions of the dolerites or tonalite by fractionating olivine and orthopyroxene from the gabbros, especially with regard to the FeO/MgO, Na_2O/MgO, CaO/MgO and Al_2O_3/MgO plots. Even if large amounts of plagioclase and clinopyroxene (which are not as abundant as orthopyroxene and olivine in the ultramafics) are fractionated as well, it is still impossible to reach the dolerite and tonalite compositions. It appears probable, then, that the ultramafics, intrusive gabbros and dolerites are not related by simple fractional crystallization processes. In a similar study of the Papuan Ultramafic Belt rocks, Davies (1971) showed that it was possible for many (but not all) of the plots to reach the composition of the basalts from a high-level gabbro by removal of reasonable proportions of the minerals occurring in the cumulate ultramafics and gabbros.

The plots in Figure 30 indicate that compositions fairly close to the dolerites could be produced by fractionating about 33% orthopyroxene from the high-Mg basalt. This is not surprising since the basalt contains about 25% pseudomorphed pyroxenes (Plate 75), some or perhaps most of which were orthopyroxenes. Table 10 gives the composition of a liquid obtained by fractionating 33% orthopyroxene from the high-Mg basalt. This liquid is fairly similar to the dolerite 38803 (the high Cr_2O_3 could be easily lowered by removing 1% chromite). However, removing olivine (which is the other important cumulus phase in the ultramafics) produces greater misfits, especially regarding silica. It must be remembered, though, that the basalt has been metamorphosed and probably has suffered metasomatism; the quartz comprising the amygdales might have been added, for example.

As a corollary to the latter discussion, it is interesting to note that the high-Mg basalt provides an example from within the Heazlewood River Complex of a tholeiitic magma which, having been produced by partial melting, probably stopped in a magma chamber long enough to crystallize large pyroxenes before it was tapped off as lava flow. It is exactly this process of settling crystals in a magma chamber from a tholeiitic magma that is required for the formation of the cumulate ultramafics. However, the above discussions indicate that the more fractionated dolerites and the layered ultramafics probably did not crystallize from a single magma type similar in composition to the less fractionated dolerites and gabbros.

It is suggested then, that while fractionation probably played some part in producing the variations in dolerites, basalts, and intrusive gabbroic rocks, fractional melting may have been a more important controlling factor. In chapter 3 it was argued that the textural continuum from coarser-grained gabbros through microgabbros to subophitic-textured dolerites intruding the layered ultramafics may have been the result of continuous injection of tholeiitic magmas during slow cooling of the ultramafics. Table 11 and Figure 29 indicate an overlap between chemical compositions of the gabbros and dolerites, suggesting a chemical as well as a textural continuum. The dolerite 38704 is quite similar in composition to the basalt 38753, and it is thought likely that the basalts and dolerites were derived from the same types of magmas.

The picture, then, is one of partial melting and emplacement of large volumes of tholeiitic magmas over a large period of time. Large magma chambers formed in which the cumulate ultramafics (and perhaps layered gabbros as in the Serpentine Hill Complex) formed by crystal settling from tholeiitic or high-Mg tholeiitic magmas. There are no known rocks within the Heazlewood River Complex itself which obviously represent the more differentiated liquid fractions of the parent magmas from which the ultramafics crystallized. As the ultramafics gradually cooled, basic magmas were continuously injected into them, first forming irregular coarser-grained gabbro bodies and later tabular dykes of micro-gabbros. By this time tensional fractures were forming in the layered rocks, perhaps in response to a regional tensional stress, and subparallel sheets of tholeiites were injected, in places forming swarms of over 90% dolerite dykes. The dykes possibly fed overlying basaltic extrusions, many of which were themselves intruded by dykes as a thick dolerite-extrusive pile gradually formed.

6. <u>Origin of the "calcalkaline rocks"</u>

On the AFM plot in Figure 29, the tonalite, trondhjemite and granophyre appear to form a calcalkaline trend. The intermediate and acid volcanic rocks also fall on this trend, but as previously mentioned, the writer is uncertain as to how much metasomatism has been involved during the metamorphism of the volcanic rocks. Nevertheless, the trend is real for the intrusives, and has been noted for sodic acid intrusives in other ophiolites (Thayer, 1967; Coleman, 1971). Although some of the dolerites appear to have differentiated towards an iron enrichment (for example, 38701, together with the dolerite and basalt from the Serpentine Hill Complex discussed in the Appendix 1), there also appears to be a gradation from quartz-bearing dolerites such as 38700 to tonalites such as 38786. Assuming these are fractionation trends, they are quite unlike those shown by other differentiated quartz tholeiites, such as the Tasmanian dolerites (McDougall, 1962), in which the granophyric types show appreciable iron enrichment. Another interesting point is that the acid intrusives and volcanics have relatively low Al_2O_3 and very low K_2O, and in most cases quite high Cr and Ni when compared to calcalkaline rocks from volcanic arcs or continental plutonic environments (Taylor, 1969). It appears unlikely that the ophiolite "calcalkaline" rocks were produced either by anatexis of crustal rocks, as is considered likely for granitic rocks, or by partial melting of eclogite or amphibolite as is postulated for island arc rocks (Green and Ringwood, 1968). It is possible to obtain a calcalkaline trend by fractionating magnetite from a basalt under relatively high fO_2 conditions (Osborn, 1962), but this process is thought to be unlikely in the case of normal calcalkaline rocks because of their Ni, V, Co and Cr contents (Taylor et al., 1969). However, it is possible in view of the unusual "calcalkaline" rocks from the Heazlewood River Complex that they were derived by fractionating magnetite from tholeiitic magmas. (In this regard it would be interesting to look at the transition metal contents of magnetites which occur in some gabbros and dolerites in the Complex.) It may also be possible to obtain andesitic and dacitic liquids by partial melting of upper mantle peridotite under hydrous conditions, as is suggested by Kushiro et al. (1972); however, the liquids these authors obtained are very depleted in MgO, and in contrast the "calcalkaline" rocks from the Heazlewood River Complex have relatively high MgO contents.

It must be pointed out that while sodic acid intrusives and extrusives are fairly common in ophiolites they appear to be fairly rare within Cainozoic oceanic crust (Aumento, 1969, and Hart, 1971 describe some examples). This may reflect the sampling problem involved in studying oceanic crust or perhaps different environmental conditions in the spreading zones in which Phanerozoic ophiolites formed.

7. Summary and conclusions

The main conclusions regarding the origin of the Heazlewood River Complex are outlined below.

1) The Heazlewood River Complex is an ophiolite which was quite probably tectonically emplaced into its present location.

2) The greater proportion of the ultramafics are cumulates which probably crystallized at pressures less than 6 kb from magmas of tholeiitic or high-Mg tholeiitic composition.

3) There appears to be a chemical as well as textural continuum from intrusive gabbros through microgabbros to dolerites and basalts. Most of these rocks are tholeiites or quartz tholeiites. The dolerites appear to grade into tonalites and granophyres, but the origin of the sodic acid intrusive and extrusive rocks is difficult to explain.

4) The Heazlewood River Complex may form a geochemical as well as a structural entity. However, the variations in chemistry of gabbros, dolerites and volcanics cannot be easily explained in terms of low pressure fractionation of olivine and pyroxenes (i.e. the minerals which are the dominant cumulus phases in the ultramafics) from tholeiitic magmas. It is suggested that much of the variety in chemical composition of these

rocks may be the result of fractional melting rather than fractional melting relationships.

5) The dolerites and volcanics have relatively low concentrations of the incompatible elements, but relative to Cainozoic mid-ocean ridge basalts they tend to form a suite of quartz tholeiites rather than olivine tholeiites, and are unusually low in TiO_2. It is argued, however, that this does not preclude the Heazlewood River Complex from forming in a spreading environment such as a mid-ocean ridge or marginal sea. Relative to mid-ocean ridge basalts, the magmas giving rise to the dolerites and volcanic rocks may have been produced by partial melting at low pressures or by a greater degree of partial melting at pressures greater than 10 kb.

It is suggested that upwelling of a mantle plume (Green, 1971) into the axis of spreading may have been important in producing the extension partial melting required. At first these magmas formed large magma chambers in which thick sequences of olivine-orthopyroxene-chromite cumulates formed, probably at pressures less than 6 kb. As the cumulates gradually cooled they were intruded by gabbros and microgabbros, and later, under the influence of a regional tensional stress, were fractured and intruded by subparallel dolerite dyke swarms. The dolerite dykes at least in part fed overlying fissure eruptions, and a thick sequence of dolerites and extrusive rocks was built up.

The considerable variations from area to area within the cumulates of the Heazlewood River Complex can be explained in terms of this model. As discussed in Greenbaum (1972), one could expect such variations in cumulates which formed at a mid-ocean ridge, as the temperature variations within the large magma chambers would be quite complex, and sudden influxes of fresh magmas into the chambers would probably be commonplace. It is also conceivable that a sequence of cumulates could be intruded by another large body of magma from which crystallized another sequence of cumulates. Thus it is possible that some of the complexities of the geology and petrology of the plutonic rocks of the Heazlewood River Complex reflect the complex fluctuating conditions which could be expected in a spreading environment.

The Serpentine Hill Complex is a dismembered ophiolite, and probably formed in a similar way as the Heazlewood River Complex.

LATE PROTEROZOIC AND CAMBRIAN TECTONICS OF TASMANIA

The main elements in any tectonic reconstruction of the Late Proterozoic and Cambrian geology are as follows.

1) The Precambrian blocks. The most important of these are the Rocky Cape Geanticline and the Tyennan Block, but others such as the Jubilee Block (Corbett, 1970) on the eastern side of the Tyennan Block could be important. It must be remembered, also, that the Rocky Cape Geanticline was probably continuous with Precambrian rocks now in Antarctica (Solomon and Griffiths, 1972).

2) The Eocambrian and Middle to Upper Cambrian sedimentary rocks which between the Precambrian blocks and flanking the Tyennan Block on the eastern side. West and north of the Tyennan Block, these can be subdivided into an Eocambrian group of wackes and mudstones with minor basic volcanics and cherts, and a Middle to Upper Cambrian group of paraconglomerates, wackes, mudstones and acid volcanics. East of the Tyennan Block, older Cambrian rocks are mainly sequences of mudstones and wackes,

mudstones and acid volcanics. East of the Tyennan Block, older Cambrian rocks are mainly sequences of mudstones and wackes while Upper Cambrian sedimentary rocks are shallow water (deltaic).

3) A third element is the Mt. Read Volcanics flanking the Tyennan Block to the west and north. On their major and minor element chemistry, these rocks are fairly typical calcalkaline (Solomon, 1964). The abundance of acid relative to intermediate members, together with the details of the trace element chemistry, show that these rocks are more similar to continental or Andean-type arc volcanics than island arc volcanics. Some of these rocks are known to be Upper Cambrian in age, while the underlying volcanics are probably Middle Cambrian and may or may not extend down into the Lower Cambrian (Gee et al., 1970; Jago et al., 1972). According to the plate tectonics model, a Cambrian subduction zone would have been associated with the volcanic arc.

4) The ophiolites and ultramafic gabbro complexes. These probably have a magmatic age of Proterozoic or Lower Cambrian, as several of them were tectonically re-emplaced and subjected to erosion in the Middle Cambrian. The ophiolites (the Serpentine Hill Complex, the H.R.C. and its satellite sheets such as the Whyte River Complex) could have formed as part of oceanic crust during Late Proterozoic or Lower Cambrian spreading which occurred between the Tyennan and Rocky Cape blocks, and were later emplaced as thrust slices into the Eocambrian rocks. On their similarity with the ultramafics and gabbros of the Heazlewood River and Serpentine Hill bodies, the other Main Belt complexes may also represent thrust slices of layer 3 of oceanic crust formed in this period of spreading. However, it is reported in Blissett (1962) that the McIvor Hill Gabbro has a chilled margin and has metamorphosed sedimentary rocks of the

Crimson Creek Formation. If this observation is correct, it could mean that some of the other Main Belt complexes also crystallized within the Eocambrian sedimentary rocks rather than at a mid-ocean ridge. At this stage, less is understood about the Adamsfield Complex, but this together with the Boyes River body may have been initially emplaced as a hot solid diapir of upper mantle rocks.

A number of models, which attempt to explain these and other features of the Upper Proterozoic and Lower Palaezoic geology of Tasmania, are proposed in Solomon and Griffiths (1972) and Corbett et al. (1972). The realization that Mt. Read Volcanic Arc is very similar to Cainozoic continental arcs rather than oceanic island arcs poses a major objection to the model of Solomon and Griffiths (1972). Thus the subduction zone corresponding to these volcanics would have to be located west of the volcanic arc or east of the Tyennan Block. In either case the subduction zone would face towards the volcanic arc. The two models are shown in Figure 31, and are discussed below.

Model 1

In this model, the ophiolites and ultramafic-gabbro complexes of the Main Belt formed by spreading at a mid-ocean ridge or similar tensional environment. Subduction beginning in the Lower Cambrian gradually consumed the oceanic crust between the Tyennan Block and the Rocky Cape Geanticline, and slices of oceanic crust were thrust into or on top of the Eocambrian sedimentary pile. The collision and related thrusting was accomplished before the Middle Cambrian, as Dundas Group sedimentary rocks unconformably overlie the Serpentine Hill sheet. After the collision, the Mt. Read volcanism continued into the Upper Cambrian. Perhaps the emplacement of the Adamsfield and Boyes River ultramafics (which also probably occurred before or during the Middle Cambrian) was related to this collision.

There are many objections to this model. Firstly, because of their nature it would be logical to assume that the Eocambrian sedimentary rocks could have formed at or near the trench. However, at Renison Bell the Crimson Creek Formation conformably overlies upper Proterozoic quartzites and dolomites which would appear to have been deposited in a relatively shallow marine environment. The trench itself would have to be placed west of a line between Waratah and Renison Bell, and the necessary distance between the trench and the volcanic arc would also require this. But this creates a major space problem, as the rocks of the Rocky Cape Geanticline appear far too close to this line. A possible solution is to propose that the Rocky Cape Geanticline was thrust over the trench during the collision. There is no direct evidence of this, and faulted contacts at the present margin of the Tyennan Block appear to be quite steep. However, the possibility that the Rocky Cape Geanticline was thrust over the trench in the Lower Cambrian cannot be precluded from what is currently known of the geology along its present contacts.

Another major objection to Model 1 is the intrusive contact of the McIvor Hill Gabbro with the Crimson Creek Formation. It is obvious that this body must be looked at in more detail, both from the point of view of relationship to the other Main Belt complexes, and also check the contact described in Blissett (1962).

Additional problem is the apparent lack of deformation in the Eocambrian rocks. The only reported tight folding in the Eocambrian rocks is in the Cleveland Mine area (Cox and Glasson, 1971), but this is now disputed by D. Ransom (<u>pers. comm.</u>) who has recently remapped the area. Possible records of the collision in the rocks themselves could be the thrust emplacement and dismembering of the ophiolites and ultramafic gabbro complexes, the unconformity at the base of the Dundas Group at Serpentine Hill, and the deformation of the lower part of the Mt. Read Volcanics. It is possible, though, that a collision can be accomplished without intense folding of the sedimentary rocks.

Model 2

This model overcomes the "trench problem" by placing the subduction zone in an area where conveniently there is greater cover of Ordovician and younger rocks. The Main Belt complexes would have formed by limited spreading in a small oceanic basin developed between the Tyennan Block and the Rocky Cape Geanticline. In the Lower or Lower Middle Cambrian, collision occurred between the Tyennan Block and another large block (which included the Jubilee Block now exposed in the Adamsfield area), emplacing the Adamsfield and Boyes River bodies as mantle slices. The effects of the collision were felt on the other side of the Tyennan Block, with the Main Belt complexes being tectonically re-emplaced. The collision was over by the Upper Cambrian, and deltaic sediments, fanglomerates, and shelf sediments were deposited for the rest of the Lower Palaezoic.

The main objection to this model is the location of the Mt. Read Volcanics; it seems too much a coincidence that the arc is located at the edge of the Cambrian basin on the western and northern sides of the Tyennan Block. The situation of the volcanic arc is more elegantly explained under Model 1.

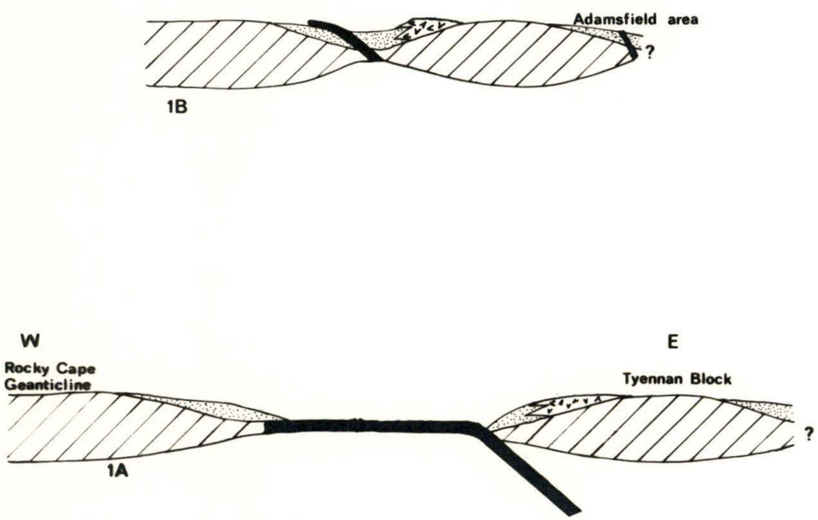
The present writer tends to favour Model 1 rather than Model 2. It is obvious, however, that a lot more work must be done in critical areas in order to test the various tectonic models.

Figure 31. Two plate tectonics models for western Tasmania (diagrammatic). The hatching represents sialic crust including Proterozoic rocks, the stipling Eocambrian flysch wedges, the v-symbols the Mt. Read Volcanics, and the black areas oceanic crust generated at spreading zones as well as ophiolite and ultramafic thrust slices.

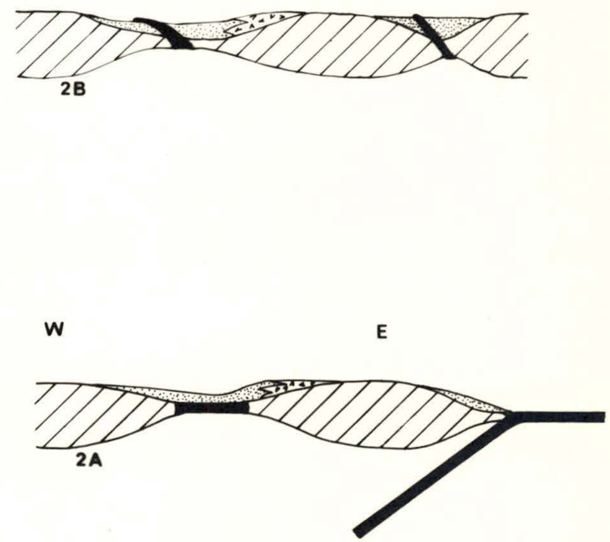
In model 1, oceanic crust generated at a spreading ridge situated between the Rocky Cape Geanticline and the Tyennan Block is consumed at an easterly-dipping subduction zone (1A). Ophiolites are emplaced as thrust slices as a result of a continent - continent collision (1B) which occurred prior to Middle Cambrian sedimentation. In model 2, the subduction zone is situated east of the Tyennan Block, and oceanic crust forms between this block and the Rocky Cape Geanticline by limited spreading in a marginal sea type of environment (2A). The ophiolites and the Adamsfield ultramafics are emplaced as thrust slices after a collision east of the Tyennan Block (2B).

Note - compare with figure 3.

MODEL 1



MODEL 2



CHAPTER 7

SERPENTINITES, RODINGITES AND AMPHIBOLITES

SERPENTINITES

A systematic study of serpentinized ultramafics was not undertaken. The most detailed work was done on serpentinites in the eastern part of the Serpentine Hill Complex, and is discussed in Appendix 1.

As is discussed in Appendix 1, the exact relationship between lizardite and chrysotile is still uncertain, especially after the observations of Chernosky (1971). In this thesis, the serpentine-group minerals are labelled according to the criteria of Whittaker and Zussman (1956); the identification of lizardite, clinochrysotile and orthochrysotile, which is based solely on diffraction peaks, must therefore be regarded as tentative.

1. Serpentinites from the Serpentine Hill Complex

These are discussed in detail in Appendix 1, and the main conclusions are summarized below.

1) An earlier phase of serpentinization preferentially attacked the harzburgite and dunite layers, replacing them by massive black to green-black dull-lustred serpentinites. In places the pyroxenites are also attacked. These black serpentinites are essentially composed of lizardite, but contain chrysotile where deformed along local shear zones. Gabbro dykes or layers in the ultramafics were altered to amphibole-prehnite rocks probably during this phase. Scattered clumps of the chrome chlorite kammererite occur in the black serpentinites in some localities; this mineral has formed at least in part from alteration of chromite.

2) A second type of serpentinite occurs only at contacts between the ultramafics and underlying country rocks. This serpentinite is highly sheared, is essentially composed of antigorite and chrysotile, and is thought to have formed during tectonic emplacement of the Serpentine Hill Complex before the Middle Cambrian.

3) A third variety of serpentinites is restricted to a shear zone, and cuts across (so is younger than) the other two types. The serpentinites are green (ranging from dark to light, but mainly a medium green) and have a characteristic waxy lustre. They range from highly sheared varieties to massive blocky varieties which are cut by widely spaced joints with slickensided surfaces. Compositionally they are lizardite-chrysotile mixtures. The green serpentinites have replaced pyroxenites and even gabbro as well as peridotites and dunites, but residual pyroxenite kernals are fairly abundant. Concentric shells of cross-fibre asbestos veins commonly surround these residual kernals where they occur in the less deformed varieties of green serpentinites. Small lenses of stichtite (a chrome-magnesium hydrated carbonate) occur in the green serpentinites, in contrast to the silicate kammererite in the black serpentinites.

2. Serpentinites from other Main Belt complexes

2.1 The Heazlewood River Complex

As in the Serpentine Hill Complex, the harzburgites, lherzolites and dunites of the Heazlewood River Complex are partly to entirely replaced by unsheared dull-lustred serpentinites. Pyroxenites are generally not replaced by serpentinites (except along shear zones, faulted contacts and adjacent to slickensided joint surfaces), but in places are partly to entirely replaced by fibrous tremolite-actinolite and/or talc.

The harzburgites and lherzolites from the Brassey Hill Harzburgites are partly replaced by black serpentinites, as in the Heazlewood River Complex. Orthopyroxenes in these rocks may remain largely altered, or otherwise are partly replaced by tremolite-actinolite and/or talc, or are pseudomorphed by large bastite plates. Clinopyroxenes are unaltered or replaced by light green-brown hornblende. Interstitial plagioclase is replaced by hydrogrossular and/or saussurite. The serpentinite replacing dunites and plagioclase dunites is generally a dull green-brown rather than black. In all cases, the olivines are replaced by mesh-textured serpentinites, and the primary textures are commonly well preserved. Fine-grained secondary magnetite occurs disseminated throughout serpentinized ultramafics. Clumps of kammererite occur in several localities.

The dunites and harzburgites from the Nineteen Mile Creek Dunites are replaced by unsheared dull-lustred serpentinites, ranging in colour from apple-green to darker green-brown.

Along faults and shear zones, the ultramafics are commonly entirely altered to deformed serpentinites, which break along polished or slickensided surfaces. These are generally green-black in colour, but are lighter green in shear zones in the Nineteen Mile Creek Dunites. Deformed magnetite veins and small amounts of kammererite may be found in these shear zones.

Counterparts of the green waxy-lustred serpentinites of the Serpentine Hill Complex have not been observed in the Heazlewood River Complex.

Table XII lists the species identified in some serpentinites from the Heazlewood River Complex. Specimen 38522 is an apple-green serpentinized dunite from the Nineteen Mile Creek Dunites, and contains

Table 12

Serpentinites from the Heazlewood River Complex.

	"Lizardite"	"Clino-chrysotile"	"Ortho-chrysotile"	Antigorite	Magnetite	Talc	Brucite	Others
Unsheared serpentinites:								
38522 (dunite)	xx				tr		xx	trace ol
38555 (harzburgite)	xx				x			
38557 (harzburgite)	xx	x			tr	x		
38565 (harzburgite)	xx	x	?		x			opx + ol
38605 (plag.dunite)	xx				tr			
38656 (harzburgite)	xx							
Deformed serpentinites:								
38541	x	?	?	xx	x			cpx + hb
38550	xx	?	?					
38559	xx	xx	?		tr			
38599	x	x	?	x	tr			
38600	xx				tr			
Sheared serpentinite at contact with sedimentary rocks:								
38597		x	?	xx	tr			

xx abundant   x fairly abundant   tr trace   ? not positively identified brucite as well as lizardite. This is consistant with the observation of Hostetler et al. (1966) that brucite commonly occurs in serpentinized dunites or olivine-rich ultramafics, and also suggests that the serpentinization reaction is closer to the "constant composition" type[x] rather than "constant volume" type (Thayer, 1966). The other unsheared serpentinite samples are black serpentinites from the Brassey Hill Harzburgites and consist mainly of lizardite. However, in contrast to the unsheared black serpentinites from the Serpentine Hill Complex, two of the samples contain some chrysotile.

The deformed serpentinites, most of which were collected near large faults within the Heazlewood River Complex, are obviously quite variable in mineralogical composition; all five contain lizardite, while two contain chrysotile and two contain antigorite.

The highly sheared serpentinite (38597) was collected from a contact with sedimentary rocks. It is composed of chrysotile and antigorite, as are similar contact serpentinites from the Serpentine Hill Complex.

2.2 Wilson River Complex

Serpentinites in the Wilson River Complex are mainly black, green-black or green-brown varieties, and are commonly deformed. Lighter green waxy-lustred varieties similar to the green serpentinites in the Serpentine Hill Complex, occur around Riley Knob, and along much of the souther Y-shaped branch in the Colebrook Hill area.

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[x] Disregarding the H_2O content.

2.3 The Dundas Serpentinite

This main serpentinite at Dundas, together with the small satellite bodies to the south, is composed almost entirely of green waxy-lustred serpentinites. These are highly sheared in places, especially near contacts. At some contacts a zone of deformed talc separates the green serpentinites from the sedimentary rocks. At one locality near the Razorback Mine, serpentinite at the contact has been replaced by silica. Asbestos veins are associated with several residual kernals of pyroxenite in the extreme north of the body.

Small blebs and rarer small lenses of stichtite are abundant in some patches in the green serpentinites on Stichtite Hill (Petterd, 1910; Twelvetrees, 1914).

2.4 The Noddy Creek Complex

In the area shown in Figure 8, and in the Asbestos Point area, the ultramafics consist largely of green waxy-lustred serpentinites which are very similar to those at Serpentine Hill. Lenses of primary ultramafics included in these are mainly peridotites partly replaced by black serpentinites (Fig. 8). Both highly sheared and "massive" varieties cut but few shear planes occur. Diffraction studies by R. Close (<u>pers. comm.</u>) and the present writer show that they are mainly lizardite-chrysotile mixtures, while some composed only of lizardite. Cross-fibre asbestos veins commonly form concentric shells surrounding orthopyroxenite kernals in patches of undeformed green serpentinites which have clearly pseudomorphed orthopyroxenites. Stichtite is quite abundant as blebs and small lenses in the serpentinites; according to R. Close (<u>pers. comm.</u>) stichtite-bearing serpentinite commonly occurs zones surrounding serpentinites in which cross-fibre asbestos veins are abundant.

2.5 The Andersons Creek Complex

The serpentinites in this body are described in Taylor (1955), Green (1959) and Gee and Legge (in press). Dark coloured serpentinites, both massive and deformed, are common, as are deformed green waxy-lustred varieties. Asbestos occurs in the latter varieties, and includes slip-fibre as well as cross-fibre types. No stichtite has been recorded from this area.

2.6 The Ulverstone area

Several small bodies of serpentinite occurring in Precambrian metamorphics in the Ulverstone area are described in Taylor (1955). These are mainly deformed green waxy-lustred serpentinites which contain asbestos as in other localities.

3. Serpentinites from the Adamsfield Complex

The serpentinites from the Adamsfield Complex are discussed in detail in Brown (1972). They are generally black, green-black or dark green in colour, although some highly sheared varieties are light green. Both massive and sheared varieties occur; the massive varieties have a dull lustre, while sheared varieties tend to be more vitreous and have slickensided joint surfaces. None of the green waxy-lustred serpentinites that are common in many of the Main Belt complexes occur in the Adamsfield area.

Brown (1972) discusses the results of 33 diffractograms from selected serpentinite samples. Regardless of the composition of the primary rock, all contain lizardite and the dunites and olivine-rich harzburgites all contain brucite. Chrysotile occurs in nine of the samples, and appears to bear no relationship to deformation as is suggested for serpentinites from the Serpentine Hill Complex (Appendix 1). Macroscopic and microscopic veins of chrysotile appear in general to be more abundant in the serpentinites from the Adamsfield Complex in comparison with dark coloured serpentinites from the Main Belt complexes.

4. Chemistry of the serpentinites

The chemical compositions of some serpentinized ultramafics are given in Table 13. Additional analyses can be found in Taylor (1955), Palethorpe (1972), Brown (1972) and Gee and Legge (in press). Some of the specimens in Table 1 (such as 38526, 38555 and 38604) are largely replaced by serpentine and other secondary minerals, and are also discussed in this section.

4.1 Oxidation during serpentinization

The serpentinized ultramafics all show appreciable oxidation of Fe^{2+} to Fe^{3+} during serpentinization, except for the serpentinized dunite 38526. It is interesting to note that all the black or green-black serpentinized rocks (whether deformed or not) have Fe_2O_3/FeO ratios greater than 1, while 38604 and 38526, both of which contain green-brown serpentinite, have Fe_2O_3/FeO ratios less than 1. Thus the colour of these serpentinites in part reflects the amount of secondary fine-grained magnetite present, although the amount of Fe substitution in serpentine minerals could also influence the colour. The green waxy-lustred serpentinites from Serpentine Hill (38806; three samples in Palethorpe, 1972; one from Taylor, 1955) have total 'FeO'[x] contents of 2.2-3.5% (calculated dry), while all other

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[x] Total Fe as FeO.

serpentinized ultramafics (regardless of colour) have 'FeO' contents of 6.5 or greater. As is discussed below, Fe was probably removed during the formation of the green waxy-lustred serpentinites.

4.2 The MgO/SiO_2 ratio of serpentinites

As is discussed in Thayer (1966), serpentinization can be a "constant composition" process, in which water alone is added, or at the other extreme it may be a "constant volume" reaction. In the latter case, some elements are partly removed so that the alteration of a dense primary ultramafic to a less dense serpentinite occurs without increase in volume of the body as a whole. The MgO/SiO_2 ratio offers a guide in considering what type of serpentinization process has occurred (Page, 1967; Coleman, 1971); the ratio for an alpine dunite is around 1.23, for an orthopyroxenite around 0.55, and for a pure serpentine around 1.0. Other factors such as density, removal or addition of other components, etc., must also be taken into account, so that it can become quite involved or even impossible to determine the exact chemical changes during serpentinization.

If the alteration is essentially isochemical (except for the addition of water), a serpentinized dunite would have a MgO/SiO_2 ratio of around 1.23, a serpentinized orthopyroxenite a ratio of around 0.55, while a harzburgite or lherzolite would vary between these extremes, depending on the relative proportions of olivine and pyroxenes. As mentioned in chapter 5, a serpentinized dunite analysed by Palethorpe (1972) gave a ratio of 1.23 while 38526 gave a ratio of 1.14. The latter sample has a greater Fe content, so the MgO/SiO_2 ratio would be expected to be slightly lower than 1.23. Most serpentinized harzburgites and lherzolites from the Serpentine Hill and Heazlewood River Complexes have MgO/SiO_2 ratios in the range 0.70-0.90, but the only way of checking if the alteration was

Table 13

Compositions of serpentinites, rodingite and amphibolites.

	38542	38599	38597	38806	38655	38689	38694	34736
SiO_2	41.46	39.82	41.13	40.60	34.32	47.76	44.79	48.05
Al_2O_3	1.02	1.61	0.11	1.16	12.20	13.62	17.56	14.90
Fe_2O_3	6.61	6.86	3.82	0.79	n.d.	n.d.	1.26	n.d.
FeO	3.71	3.05	2.63	1.05	9.71*	9.75	7.86	10.51*
MgO	30.28	36.06	39.11	40.91	11.86	11.81	10.28	9.87
CaO	6.02	0.05	0.05	0.00	22.44	8.90	12.87	10.68
Na_2O	0.00	0.00	0.00	0.00	0.00	2.80	1.48	1.50
K_2O	0.00	0.00	0.00	0.00	0.00	0.27	0.40	1.26
TiO_2	0.09	0.10	0.05	0.06	0.45	0.84	0.50	0.64
P_2O_5	0.01	0.01	0.02	0.00	0.14	0.11	0.05	0.06
MnO	0.10	0.06	0.05	0.07	0.15	0.18	0.11	0.24
NiO	0.20	0.34	0.33	0.15	n.d.	n.d.	n.d.	n.d.
Cr_2O_3	0.36	0.66	0.55	0.60	n.d.	n.d.	n.d.	0.08
Loss	9.47	11.67	12.69	15.28	8.01	4.93	3.02	2.42
	99.33	100.29	100.54	100.67	99.28	100.97	100.18	100.21

38542 partly serpentinized wehrlite or websterite, Heazelwood River Complex.
38599 deformed serpentinite, Heazelwood River Complex.
38597 sheared contact serpentinite, Heazelwood River Complex.
38806 green waxy-lustred serpentinite, Serpentine Hill Complex.
38655 rodingite, Heazlewood River Complex.
38689, 38694 amphibolites, Heazlewood River Complex.
34736 amphibolite, Serpentine Hill Complex.

* total Fe as FeO.

isochemical would be to base calculation on such factors as the chemical composition of the rock and of the primary minerals, the density of the rocks, and the model composition of the primary rock (for example, see Green, 1964).

Specimen 38597 has an MgO/SiO_2 ratio of 0.95. This is highly sheared serpentinite derived from an orthopyroxenite and so either much Mg was added or Si was removed. The latter alternative is the more likely as mudstones in contact with this serpentinite are silicified.

In Table XIV, the composition of a residual pyroxenite kernal from Serpentine Hill is compared with the immediately surrounding green waxy-lustred serpentinite which has obviously replaced pyroxenite. The results indicate that serpentinization involved total depletion of Ca, depletion of Fe, and it is thought that Si was probably removed rather than Mg added to cause the large change in the MgO/SiO_2 ratio. Two analyses of green waxy-lustred serpentinites in Taylor (1955), and four analyses in Palethorpe (1972) give the same results; loss of Ca, depletion of Fe, and change in the MgO/SiO_2 ratio from 0.55-0.70 for the pyroxenite to 1.00-1.05 for the serpentinite. It is concluded that the serpentinization processes producing the green waxy-lustred serpentinites approximate to the "constant volume" serpentinization, with excess Si, Fe and Ca having been removed in solution. The presence of talc and/or silicification at contact zones involving green serpentinites and sedimentary rocks in the Dundas Serpentinite and in the Noddy Creek Complex is consistent with this hypotheses. Perhaps some of the Fe removed during serpentinization was concentrated in localized shear zones, some of which contain abundant fibrous magnetite.

Table 14

Analyses of pyroxenite kernal and surrounding green serpentinite from Serpentine Hill quarry, recalculated water-free.

	pyroxenite	serpentinite
SiO_2	53.05	47.87
TiO_2	0.10	0.07
Al_2O_3	1.36	1.37
FeO	6.31	1.24
Fe_2O_3	1.32	0.93
MgO	32.53	48.23
CaO	3.90	0.00
Na_2O	0.00	0.00
K_2O	0.00	0.00
P_2O_5	0.00	0.00
MnO	0.27	0.08
Cr_2O_3	0.71	0.71
MgO/SiO_2	0.61	1.01

Brown (1972) discusses the chemistry of progressive serpentinization of dunite and pyroxenite from Adamsfield. He shows that the MgO/SiO_2 ratio for the dunite-serpentinite sequence decreases from 1.18 to 1.00, while for the orthopyroxenite sequence it increased from 0.75 to 1.00. From this he concludes that serpentinization processes in the Adamsfield Complex were essentially on a volume for volume basis.

4.3 Conclusions

From these preliminary studies of the chemical compositions of the Tasmanian serpentinites, the following conclusions can be made.

1) The formation of unsheared dull-lustred serpentinites in the Main Belt ultramafics was probably at least in some places an isochemical process. However, quite detailed work would be necessary to establish this with greater certainty, especially for serpentinized peridotites.

2) The serpentinization processes during which the green waxy-lustred serpentinites formed appear to be quite vigorous, and resulted in the removal of some Si and Fe and virtually all the Ca in the original rock. The process was probably close to a "constant volume" replacement type. The presence of the carbonate stichtite (in contrast to the silicate kammererite in the black serpentinites) and of fibrous dolomite in some localities in the green serpentinites suggest that the fCO_2/fH_2O ratio may have been slightly higher in comparison with the formation of the black serpentinites.

3) The colour of serpentinites in Tasmania is a function of the total Fe left in the rock at the conclusion of serpentinization and the FeO/Fe_2O_3 ratio.

5. Re-intrusion of solid serpentinite

As is characteristic of many alpine bodies, contacts between ultramafics and country rocks in Tasmania generally show no metamorphism of magmatic origin. The contacts commonly consist of serpentinite against sedimentary rocks which are either unmetamorphosed or are sheared within 1 m of the contact. An apparent exception is the presence of andalusite in pelitic sedimentary rocks along the western contact of the Andersons Creek Complex. However, the present writer agrees with Gee and Legge (in press) that this thermal metamorphism probably occurred during serpentinization, and also believes (contrary to Gee and Legge) that apparent thermal metamorphism of septa of Precambrian schists included in this body is probably due to serpentinization rather than magmatic processes. The only other alteration of country rocks noted by the writer is localized silicification of contact zones in some localities in the Dundas Serpentinite and Noddy Creek Complexes.

Since magmatic thermal aureoles appear to be absent, it is thought that the contacts between ultramafics and country rocks in Tasmania are tectonic, being produced by faulting and/or solid flow of serpentinite. (The writer knows of no field criteria which satisfactorily distinguish between these two processes, and is uncertain as to whether such a distinction is realistic).

A mechanism for solid emplacement of serpentinite is discussed in Raleigh and Patterson (1965) and Raleigh (1967). Experimental work by these authors show that serpentinite loses its ductility at about 500°C, partly dehydrates and deforms brittlely. Under these conditions, the serpentinite could be intruded as a number of independent fault-bounded blocks lubricated by the high PH_2O set free by the partial dehydration.

Another possible mechanism is proposed by Carey (1953). If a sufficiently large stress is applied for a sufficiently long period of time, serpentinite may be capable of plastic flow as a "rheid" body, analogous to the solid flow of salt in a salt dome. The temperature of the serpentinite body would have to remain relatively low, for otherwise partial dehydration and embrittlement would occur (Raleigh, 1967) and "rheid" flow would be less unlikely.

RODINGITES AND RELATED ROCKS

Rodingites were first described from the Roding River, New Zealand (Marshall, 1911; Bell et al., 1911). They consist essentially of hydrogrossular, with pyroxene and/or amphiboles, and varying amounts of other hydrous Ca-Al silicates such as prehnite, vesuvianite, zoisite and pectolite. They usually occur within ultramafic bodies or at their contacts with country rocks. It is now thought that they probably form by metasomatism of basic igneous rocks, or in some cases sedimentary rocks, during serpentinization processes (Bilgrami and Howie, 1960; Coleman, 1966, 1967). The present writer found no evidence contrary to this hypothesis.

1. Rodingites from the Andersons Creek Complex

The alteration of gabbros and hornblende gabbros in the Andersons Creek Complex to rodingites is described in Baker (1959), Green (1959) and Gee and Legge (in press). The rodingites consist essentially of fine-grained hydrogrossular, pyroxenes and/or amphiboles, and chlorite, with such minerals as vesuvianite, clinozoisite and prehnite present in many samples. Baker (1959) records an analysis of hydrogrossular with 0.99%

H_2O^+. Data given in Gee and Legge (<u>in press</u>) indicate considerable Ca-metasomatism in converting a gabbro to rodingite.

Baker (1959) argues from field evidence that the rodingite formed prior to the intrusion of serpentinite. However, the present writer agrees with Gee and Legge (<u>in press</u>) in that the rodingite probably formed from gabbro dykes during serpentinization, and that the serpentinite was subsequently deformed and "re-intruded" the rodingite.

2. Rodingites from the Heazlewood River Complex

Gabbro, microgabbro and dolerite dykes intruding what are now serpentinized or partly serpentinized ultramafics have been altered either to saussurite-uralite rocks or to rodingites. Rodingites are more common in the olivine-rich harzburgites and plagioclase dunites where serpentinization is complete or near complete. In some cases, the alteration of the dyke to rodingite is restricted to a narrow margin (Plate 21), with the plagioclase in the rest of the dyke replaced by saussurite.

The petrography of rodingitized gabbroic rocks is described in chapter 3. Hydrogrossular is an essential mineral, and others present include pyroxenes (the original igneous pyroxenes), amphiboles (replacing pyroxenes), chlorite, and iron oxides (in many cases the original igneous magnetite). Very fine-grained high relief minerals (saussurite) also occur, but no vesuvianite, clinozoisite or pectolite could be identified. Prehnite is quite rare. In some rodingites the overall texture of the gabbro is preserved, while in others the original igneous texture has been obliterated. In some cases a "reaction zone" containing amphiboles and chlorite occurs in the serpentinite immediately adjacent to the rodingite body.

The composition of a rodingite (38655) is given in Table 13. The rock consists largely of hydrogrossular, with scattered small clinopyroxenes partly replaced by chlorite and tremolite-actinolite. The original rock may have been a dolerite or a microgabbro. The analysis suggests considerable Ca-metasomatism during the formation of the rodingite from the original rock; the Al_2O_3/CaO ratio is 0.54 for 38655, compared with a ratio of 1.1 or greater for unaltered dolerites and microgabbros.

As described in detail in chapter 3, the interstitial plagioclase in most of the serpentinized plagioclase dunites and peridotites from the Brassey Hill Harzburgites is replaced by hydrogrossular and some saussurite. Specimen 38604 is such a serpentinized plagioclase dunite, and its chemical composition is given in Table I. The Al_2O_3/CaO ratio of this rock is 1.34, which is lower than calcic plagioclase (which has a ratio in the range 1.9-2.4) from which the hydrogrossular and saussurite was derived. The higher ratio for 38604 compared with the rodingite 38655 is perhaps due to the greater amount of saussurite in the former.

The formation of rodingites and the replacement of interstitial plagioclase in ultramafics by hydrogrossular and saussurite most probably occurred during serpentinization. However, the origin of the extra Ca needed for the formation of hydrogrossular is a problem. Perhaps some Ca could have been derived from the alteration of pyroxenes. However, clinopyroxenes are generally replaced by amphiboles rather than serpentine minerals, while orthopyroxenes are replaced by tremolite-actinolite or by both bastite and tremolite-actinolite, which if anything would require addition rather than loss of calcium. Some Ca was probably available from the serpentinization along faults and shear zones where pyroxenes are generally fully serpentinized. It is also quite possible that Ca was brought in by the water which produced the serpentinization.

3. Amphibole-prehnite rocks from the Serpentine Hill Complex

In contrast to the rodingites formed in very similar types of serpentinites in the Heazlewood River Complex, dykes and layers of gabbros in black serpentinites in the Serpentine Hill Complex have altered to amphibole-prehnite rocks. These are described in detail in Appendix 1. In contrast to the formation of rodingites, appreciable Ca-metasomatism does not appear to have occurred during the formation of amphibole-prehnite rocks. This is consistent with the suggestion of Coleman (1967) that the stability of prehnite relative to hydrogrossular is enhanced by higher Si activity and lower activities of Ca and Al.

CONTACT AMPHIBOLITE LENSES

Lenses of amphibolites occur at ultramafic contacts in the Serpentine Hill, Wilson River and Heazlewood River complexes. They range up to 500 m in length and 200 m in width, and are elongate parallel to the contact. With the exception of the one cropping out in the Ring River in the Serpentine Hill Complex, the lenses are surrounded by sheared serpentinite. None of the lenses are associated with the green waxy-lustred serpentinites. Sedimentary rocks immediately adjacent to the lenses show no effects of thermal metamorphism, but in some cases mudstones have been converted to sericite-bearing phyllites within a metre of the contact.

With the exception of the lens in the Ring River, most of the amphibolites show a strong foliation that roughly parallels the contact.

In many localities, a lineation (defined by the elongate amphiboles) is obvious, and is generally sub-horizontal. In some localities, the amphibolites are striped, with layers varying in plagioclase content. Elsewhere, they contain no plagioclase. The foliation is thought to represent transposition. Isoclinal folds were not observed in the field, but occur in one of the specimens (38689), in which the foliation is clearly axial surface to the folds.

1. Amphibolite lenses from the Serpentine Hill Complex

These are described in detail in Appendix 1. The amphibolite body in the Ring River differs from the others in that it is not foliated, and consist mainly of fibrous tremolite-actinolite and serpentine minerals. It contains residual olivine grains and there is no sheared serpentinite zone between the amphibolite and the sedimentary rocks. It is thought that this body may have formed by metasomatism of partly serpentinized ultramafics in a similar fashion to the tremolite-actinolite reaction zones at serpentinite contacts as described in Coleman (1966, 1967).

The other two lenses in the Serpentine Hill Complex are foliated similar to those in the Wilson River and Heazlewood River complexes.

2. Amphibolite lenses from the Wilson River Complex

An amphibolite lens on the lower contact of the Wilson River Complex in Ahearne Creek was studied in detail. Such lenses were not present at other contacts examined by the writer, but A. Jessup (pers. comm.) found several other lenses along the lower contact. The lens in Ahearne Creek is about 20 m wide. It is separated from unmetamorphosed Crimson Creek mudstones and wackes by a thin sheared serpentinite zone, while on the eastern side it is replaced by sheared serpentinite. (Specimen 38848 shows the amphibolite replaced by chlorite at the actual contact.)

The amphibolites are foliated, show an obvious lineation, and vary from fine-grained to rocks with an average grainsize of 3 mm. In 38847-38850, the hornblendes are very light brown xenoblastic prisms. They appear to have suffered some protoclasis as they are surrounded by fine-grained hornblende; annealing would have occurred after deformation, since the grains are generally unstrained. Specimen 38849 contains some hornblende porphyroblasts up to 2 mm, the groundmass hornblendes averaging 0.3 mm. Small grains of olive-green spinel are scattered throughout these rocks, while fine-grained magnetite occurs mainly interstitial to the larger hornblendes. Specimen 38849 contains small relict olivine grains, while 38850 contains several large relict porphyroblasts of olivine (Plate 81), as well as small clinopyroxenes which may or may not be of metamorphic origin. It is quite probable that those amphibolites have been derived from the metamorphism and metasomatism of ultramafic rocks, the olivines being relicts from the original peridotite.

Specimen 38851 consists of darker green-brown hornblende with 30% plagioclase replaced by sericite. It does not contain green spinel, and contains several relict clinopyroxenes partly to largely replaced by hornblende. This rock is similar to 34736 from the Serpentine Hill Complex, and may have been formed by metamorphism and deformation of a gabbro rather than an ultramafic rock.

Table XV gives analyses of an amphibole and spinel from 38850. The amphibole is obviously a hornblende rather than tremolite-actinolite, which was first suspected from the very light pleochroic colours and small extinction angle. The composition of this hornblende is fairly similar to "pargasites" from the Lizard (Green, 1964) and St. Pauls Rocks peridotites (Melson *et al.*, 1971) and less so to the hornblende from a lherzolite nodule (Varne, 1970). It is lower in alkalis than true pargasites, and is best called "hornblende" from the data in Deer *et al.* (1966; figs. 60 and 61). The green spinel is best referred to as picotite (Deer *et al.*, 1966, p.426). Although the spinel may be relict from the original ultramafic it is thought that it is more likely to be metamorphic in origin, probably forming in part from the original chromite. Green spinel formed by thermal metamorphism of the Trial Harbour Serpentinite by the Heemskirk Granite is described in Green (1966), demonstrating that green spinel can form in ultramafics by metamorphic processes.

The amphibolites are amphibolite facies, while the adjacent mudstones and greywackes show very low grade regional metamorphism (probably lower than greenschist facies). No structures corresponding to the foliation and lineation in the amphibolites occur in the adjacent sedimentary rocks. These observations show that the amphibolites did not form *in situ*, but were tectonically emplaced with the ultramafic body. Their origin is discussed in more detail below.

3. Amphibolite lenses from the Heazlewood River Complex

The northernmost is the largest of these and is exposed in a small creek flowing into Nineteen Mile Creek. The others were observed as blocks that are probably float rather than outcrop. The two northern bodies are separated from sedimentary rocks by a thin strip of sheared serpentinites, but there is no outcrop in the vicinity of the southern lens. All three lenses show a foliation and generally a lineation. The foliation in the northern lens strikes NW to WNW (parallel to the main contact in this area - see Fig. 11), and the lineation appears to be mainly sub-horizontal. Much of the amphibolite in the northern lens is striped, whereas the other two are more uniform.

Specimens 38686 and 38699 are from the southern lens. The former is a fine-grained amphibolite composed of green-brown hornblende and a little quartz, and is cut by veins of epidote and prehnite. Specimen 38699 is a fine-grained amphibolite which has been brecciated subsequent to the development of the foliation.

Specimen 38695 is from the NW lens, and is composed of green-brown amphibole and saussuritized plagioclase.

The amphibolites from the northernmost lens are composed of green-brown amphibole, with or without saussuritized or sericitized plagioclase. In 38639, the foliation is obviously axial surface to small isoclinal folds. Specimen 38693 is composed of green-brown hornblende and sericitized plagioclase, and is very similar to 38851 from the Wilson River Complex. A crenulation cleavage occurs in part of 38690, suggesting two "phases" of deformation. Specimen 38691 contains what appear to be augen of dolerite wrapped around by the foliation; the pyroxenes in the augen are replaced by hornblende and the plagioclase by saussurite, but the texture is very similar to that of a typical sub-ophitic dolerite. Specimen 38692 actually contains residual grains of hypersthene and clinopyroxene partly replaced by hornblende. A number of loose boulders of dolerite were found in the area underlain by the amphibolite, and these (e.g. 38703) are composed of labradorite, hypersthene and augite, and are indistinguishable from dolerites intruding ultramafics about 1 km to the SE. It is quite likely that these

Table 15

Compositions of hornblende and green spinel from the amphibolite 38850.

	hornblende	spinel
SiO_2	45.81	0.00
TiO_2	0.45	0.23
Al_2O_3	11.84	47.22
'FeO'	7.28	28.39
MnO	0.19	0.32
MgO	17.57	11.02
CaO	12.28	0.09
Na_2O	1.29	0.00
K_2O	0.05	0.00
Cr_2O_3	0.24	10.19
NiO	0.00	0.85
	97.00	97.51

Structural Formulae

Based on 24(OH)*			Based on 4(O)		
Si	6.432)	2.000	Al	1.588)	1.990
Al	1.568)		Cr	0.230)	
			Ti	0.005)	
Al	0.391)	5.020	Fe**	0.177)	
Ti	0.047)				
Fe	0.855)		Mg	0.469)	1.001
Mn	0.023)		Fe**	0.504)	
Mg	3.677)		Mn	0.008)	
Cr	0.027)		Ni	0.020)	
Ca	1.848)	2.207			
Na	0.351)				
K	0.008)				
OH*	2.818				

* Assuming 3.00% H_2O.

** Total Fe was distributed between the two sites.

loose boulders occurred as augen or lenses in the amphibolites, and that the amphibolites were at least in part derived from the metamorphism of dolerite dyke swarms forming part of the Heazlewood River Complex. (Hypersthene-bearing rocks have not been observed in the Cambrian rocks outside of the ophiolites.)

As in the case for the two western lenses in the Serpentine Hill Complex and the one from the Wilson River Complex, it is though t that the lenses in the Heazlewood River Complex were formed before tectonic emplacement into their present situations, as the metamorphic grade and structures in the amphibolites are incompatible with those of the immediately adjacent sedimentary rocks.

Three analyses of amphibolites are given in Table XIII.

4. Origin of the amphibolite lenses

The occurrence of metamorphic amphiboles or amphibolites in ultramafics or at their contacts may be the result of metasomatic alteration during serpentinization (Coleman, 1966, 1967), recrystallization under hydrous conditions during tectonic emplacement of an ultramafic body (as is suggested by Green, 1964a, for the hornblende-bearing ultramafics in the Lizard intrusion), or perhaps "dynamothermal" metamorphism of country rocks during emplacement of a hot periodtite diapir (Green, 1964b). As previously mentioned, the lens cropping out in the Ring River fits best into the ideas of Coleman (1966, 1967) regarding tremolite-actinolite-bearing reaction zones at serpentinite contacts. The other lens are different, and the ideas of the above authors are drawn upon in the following hypothesis to explain their origin.

It is suggested above that the foliated amphibolite lenses were derived from metamorphism, deformation and, at least in some cases, metasomatism of ophiolite rocks, namely ultramafics, gabbros and dyke swarms. That this was achieved before tectonic emplacement of the complexes into the positions they now occupy is necessitated by the disparity in metamorphic grade and structural style between the amphibolites and the country rocks. That this process is possible can be demonstrated from the Gray Creek Complex (White, 1965; Green, 1968a) situated NW of Townsville in North Queensland. In this complex, which is currently being studied by the present writer, residual lenses and boudins of clinopyroxenites, wehrlites and layered gabbros, in places intruded by abundant dykes, are wrapped around by foliated amphibolites derived from these rock types. The amphibolites, which in places are very well exposed, have suffered at least three phases of deformation, the first of which produced the dominant foliation and lineation by "stripping out" of cleavages axial surface to isoclinal folds. Many of the amphibolites are remarkably similar, both in hand specimen and in thin section, to those on the ultramafic contacts in Tasmania.

It is suggested, then, that the foliated amphibolites formed by deformation and metamorphism of the ophilites either (1) along local belts before tectonic emplacement, or (2) during deformation which occurred as an early stage in the tectonic emplacement. In the former hypothesis, it can be further postulated that the thrusts which emplaced the ophiolites developed along these local belts of deformation, so as to explain the occurrence of the amphibolite lenses at the ophiolite contacts. It is quite possible that deformation of the Nineteen Mile Creek Dunites occurred contemporaneously with the formation of amphibolites but under relatively dry conditions.

The hypotheses for the origin of the foliated amphibolite lenses are illustrated in Figure 32.

CHAPTER 8

SUMMARY AND CONCLUSIONS

GEOLOGICAL HISTORY OF THE TASMANIAN OPHIOLITE AND ULTRAMAFIC-GABBRO COMPLEXES

In this section, the geological history of the Tasmanian ophiolite and ultramafic-gabbro complexes is traced from the magmatic stage through the various re-emplacements and serpentinization phases.

1. Magmatic stage

Most of the layered ultramafics and gabbros of the Main Belt complexes probably formed by crystal accumulation from large tholeiitic or high-Mg tholeiitic magmas. This probably occurred at a relatively low load pressure (probably less than 6 kb), possibly at a mid-ocean ridge or some other spreading environment. The layered rocks were intruded by abundant gabbroic dykes and later by microgabbros at the host rocks gradually cooled. Then, as a result of the pervasive tension in the spreading zone, the layered rocks in at least some of the bodies (the Heazlewood River and Serpentine Hill ophiolites) were fractured and intruded by dolerite dyke swarms. The dykes, largely quartz tholeiites and relatively depleted in the incompatible elements, probably acted as feeders for a thick volcanic sequence which formed above the layered plutonic rocks. Bodies of tonalites, trondhjemites and granophyres also intruded the layered ultramafics and gabbros, dolerites and at least some of the volcanic rocks, and the extrusive rocks of acid to intermediate composition may have formed from similar magmas. At this stage all these rocks may have constituted layers 2 and 3 of oceanic crust formed in the spreading zone in the late Proterozoic or Lower Cambrian.

A possible objection to the hypothesis that these complexes originally formed at a mid-ocean ridge type of environment is the observation of Blissett (1962) that the McIvor Hill Gabbro is chilled against and has thermally metamorphosed sedimentary rocks of the Crimson Creek Formation. This implies that this body, and some other of the Main Belt complexes, could have formed in situ during Eocambrian sedimentation. However, all the contacts between the complexes and the sedimentary rocks examined by the present writer are tectonic, and the sharp contrast in composition between the ophiolite rocks and other intrusive and extrusive rocks in the Eocambrian sequences supports the hypothesis that the Main Belt complexes are allochthonous.

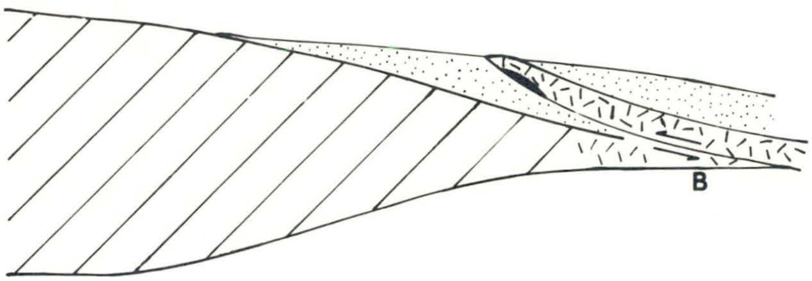
The layered ultramafics of the Adamsfield Complex may have formed by crystal settling from a magma (Brown, 1972). However, it is not known whether the massive ultramafics of this complex are cumulates for whether they represent residual upper mantle; if the former hypothesis is correct it would necessitate crystallization from an ultramafic magma within the upper mantle.

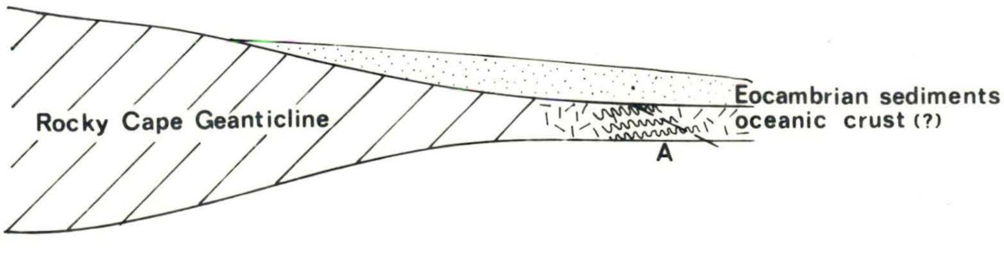
2. Tectonic emplacement in the Eocambrian

2.1 The Main Belt complexes

Most of the Main Belt complexes were probably emplaced as thrust slices into the Eocambrian sedimentary rocks. The formation of the foliated amphibolites by metamorphism of ophiolite rocks preceded or accompanied the earlier stages of this tectonic emplacement. Deformation of the Nineteen Mile Creek Dunites may have occurred penecontemporaneously with the formation of the amphibolites. Dismembering of the ophiolites (e.g. the Serpentine Hill Complex), the production of melange zones in the SW of the

Figure 32. Model for the formation of amphibolites and tectonic emplacement of the Tasmanian ultramafic gabbro and ophiolite complexes. In A, oceanic crust is locally deformed, forming lenses of amphibolites in deformed and partly serpentinized ultramafics. In B, slices of crust are thrust into the Eocambrian flyschoid sediments, the thrust initiating along the belt deformed in A. An amphibolite lens is shown in black.


B


Rocky Cape Geanticline
Eocambrian sediments
oceanic crust (?)
A

Heazlewood River Complex, and the formation of at least some of the sheared contact serpentinites, may have occurred during this initial tectonic emplacement.

The emplacement of the ophiolites and ultramafic-gabbro complexes as thrust slices may be related to a continent-continent collision. The unconformity between the Serpentine Hill Complex and the overlying Dundas Group indicates that the thrusting occurred before the Middle Cambrian sedimentation began. The abrupt change in lithologies from the Eocambrian sedimentary rocks to the Middle and Upper Cambrian sequences may also be a consequence of this tectonic event.

2.2 The Adamsfield area

The deformation of the massive ultramafics, and the formation of mylonites and their subsequent folding, may have occurred during the initial emplacement of the Adamsfield Complex as a diapir or hot slice from the upper mantle. The abundance of ultramafic detritus in Middle or Upper Cambrian sedimentary rocks exposed along the Scotts Peak road places an upper limit on the initial emplacement of these rocks.

3. Earlier serpentinization phases

3.1 The Main Belt complexes

In the Main Belt complexes, the dunites and peridotites, and less commonly the pyroxenites, have been largely replaced by massive dull-lustred serpentinites. These are generally black or greenish-black, but are green or greenish-brown where they have replaced dunites and plagioclase dunites in the Heazlewood River Complex. The replacement of interstitial plagioclase by hydrogrossular and/or saussurite, and the alteration of other gabbroic and dolerite dykes to rodingites, amphibole-prehnite rocks, or saussurite-uralite rocks, probably occurred during the formation of these massive serpentinites. These serpentinization events, then, occurred after the intrusion of gabbro and dolerite dykes, but it is not known whether they took place before of after the initial tectonic emplacement. Similarly, it is not known when the deformed serpentinites (excluding the green waxy-lustred types) occurring along faults and shear zones were produced, but it seems quite likely that they formed at various times spanning much of the geological history of the Main Belt complexes.

3.2 The Adamsfield area

The serpentinization processes in the Adamsfield and Boyes River bodies possibly began in the initial tectonic emplacement and continued throughout subsequent re-emplacements along the Lake Edgar Fault zone.

4. <u>Later serpentinization and tectonic re-intrusion events</u>

4.1 The Main Belt complexes

Most or all of these were probably emplaced as thrust slices before the Middle Cambrian, but many have been re-emplaced so that they are now in contact with Middle and Upper Cambrian sedimentary rocks (the Dundas Serpentinite), Ordovician rocks (the Andersons Creek and Noddy Creek bodies), and rocks as young as Silurian in the case of the Wilson River Complex. Tectonic re-emplacement may have been achieved by faulting or solid flow of serpentinite according to the mechanism of Raleigh (1967) or Carey (1953); several of these mechanisms could have been involved in the formation of any single contact between serpentinite and country rocks.

In the Serpentine Hill Complex it can be demonstrated that the green waxy-lustred serpentinites formed after the black massive serpentinites and the sheared contact serpentinites, and probably occurred after the initial tectonic emplacement as a thrust slice. The same green waxy-lustred serpentinites comprise large parts of the Dundas, Noddy Creek, Wilson River and Andersons Creek bodies, all of which have been placed in contact with sedimentary rocks younger than Eocambrian; thus it appears likely that re-emplacement of these complexes is related to the formation of these green serpentinites. The re-emplacement may have occurred during the Middle Devonian Tabberabberan "Orogeny", the only established significant Palaeozoic deformation phase that occurred after the Middle Cambrian.

4.2 The Adamsfield area

As mentioned above, the Adamsfield Complex must have been emplaced, partly serpentinized and exposed to erosion during the deposition of Middle or Upper Cambrian sedimentary rocks in the Scotts Peak Road area. Subsequent to this it was unconformably overlain by Upper Cambrian deltaic sediments at Adamsfield itself. Ultramafic rocks were again (or still?) exposed at the start of deposition of the Lower Ordovician fanglomerates (which in places contain chromite), but later (during the Tabberabberan "Orogeny") were placed in contact with these Ordovician rocks by faulting, solid flow of serpentinite, or both. The Adamsfield Complex, together with the Boyes River body, is situated along the Lake Edgar Fault zone, which is believed to have controlled tectonics and sedimentation throughout the Upper Proterozoic and Lower Palaeozoic (Corbett, 1970).

5. The significance of successive re-intrusions of the Tasmanian complexes

A number of conclusions regarding tectonic re-intrusions of ophiolite and ultramafic-gabbro complexes are summarized below.

1) After their initial emplacement, ophiolite and ultramafic-gabbro complexes are likely to be tectonically re-emplaced during any

subsequent movement or deformation phase in the orogenic belt. In fact, the complexes would tend to lie along a zone of weakness by virtue of their initial tectonic emplacement, and later deformation events are likely to be concentrated along such already existing weak zones.

2) The age of initial emplacement and subsequent re-intrusions can be very difficult to establish, especially in the case of an orogenic belt which has suffered more deformation and metamorphism than the Palaeozoic of Tasmania. Since successive re-intrusions can occur, the initial emplacement of a dismembered ophiolite or alpine ultramafic body cannot necessarily be determined from the ages of the country rocks with which the body at present happens to be in contact.

3) Searching for ultramafic detritus in sedimentary rocks with which ophiolites and ultramafics are associated can prove quite helpful in placing limits upon ages of tectonic emplacement and re-intrusion.

4) Detailed studies of the field characteristics and relationships, chemistry, and mineralogical compositions of serpentinites may distinguish a number of phases of serpentinization and their relationships to tectonic re-intrusive events.

5) Successive serpentinization, deformation, and tectonic re-intrusive events can convert an ultramafic-gabbro complex into a series of thin elongate lenses of serpentinite stretched out along a major fault zone and containing only scattered residual kernals of the primary rock types. In such cases, it may be difficult or impossible to unravel the geological history of the ultramafic rocks.

CONCLUSIONS - FUTURE WORK ON THE TASMANIAN COMPLEXES

A broad approach was taken in this Ph.D. project on the Tasmanian ultramafic-gabbro and ophiolite complexes; it involved studying field relationships, petrography and chemistry of the associated volcanics dolerites and acid intrusives as well as the ultramafics and gabbros, a study of the various serpentinites and associated metasomatic rocks, and an attempt at synthesizing all this data to determine the origin and tectonic significance of the complexes. Because of this approach, no particular aspect was exhaustively studied; all hypotheses and conclusions must be regarded as tentative and can be tested by more detailed studies.

Some of the more obvious areas for future investigations are outlined below:

1) A detailed study of the layered ultramafics. This would involve careful mapping and layer-by-layer chemical and petrographic studies (including modal analyses) of well exposed accessible sequences such as those at Serpentine Hill and Brassey Hill. Among other things, such investigations could evaluate the nucleation hypotheses outlined in this thesis, and should lead to a better understanding of the origin of the layering.

2) A detailed study of the structure of the Heazlewood River Complex. This would involve careful mapping of all the faults and contacts between different horizons, together with the measurement and interpretation of all structural elements, in particular the layering.

3) A study of the McIvor Hill Gabbro. Detailed investigations of the petrology would probably be dissappointing, as the rocks are extensively altered to uralite and saussurite. However, a study of the

southern contact, which Blissett (1962) describes as intrusive into sedimentary rocks, may be critical regarding emplacement of the Main Belt complexes.

4) Detailed investigations of the petrography and major and trace element chemistry of Proterozoic and Cambrian volcanics, dolerites and gabbros in comparison with the ophiolite rocks.

5) A study of the serpentinites. Excellent exposures occur in the Tasmanian complexes for research on aspects such as chemical changes during serpentinization and the origin of chrysotile asbestos. A combination of petrographic work, microprobe analyses, and X-ray investigations of various serpentinite types could illuminate the problems of serpentine "polymorphism".

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