

THE GEOLOGY OF THE TANTALITE VALLEY

MAFIC-ULTRAMAFIC COMPLEX AND THE

KUMKUM METAMORPHIC-IGNEOUS MASSIF

NEAR WARMBAD, SOUTH WEST AFRICA (NAMIBIA)

BY

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v. 1

Thesis submitted in fulfillment of the requirements  
for the degree of Doctor of Philosophy

University of Cape Town

March, 1979

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## ABSTRACT

An area south of Warmbad, South West Africa (Namibia) consisting of medium-to high-grade rocks of the Namaqua Metamorphic Complex has been mapped. It is established that this original dominantly sedimentary region has undergone polyphase metamorphism and deformation and has been intruded by a variety of gabbroic rocks. These are dominated by the Tantalite Valley and Kumkum Complexes which are associated with the major Tantalite Valley Shear Zone.

Metamorphism took place under low-P facies series conditions characterized by high geothermal gradients ( $25-50^{\circ}\text{C}/\text{km}$ ). The Tantalite Valley Shear Zone axis defines the boundary between high-and medium-grade metamorphic rocks which occur to the north and south of this line respectively. Pelitic gneiss (Black Gneiss) and granite gneiss are restricted to the high grade domain where muscovite is not found and migmatites are common. The Black Gneiss is interpreted as a restite after subtracting a granitic liquid. Black Gneiss parageneses formed during M1 ( $P = 5-6 \text{ kb}$  ( $P_{\text{tot}} > P_{\text{H}_2\text{O}}$ ),  $T = 750 - 860^{\circ}\text{C}$ , low  $f\text{O}_2$ ). During this event the Inequigranular granite gneiss and the Kumkum and Eselruh gabbronorites were emplaced in that order. Partial remobilization of granite gneiss by these gabbronorites gave rise to the hybrid Contaminated gabbronorite. The Tantalite Valley Shear Zone was initiated during M2, which was a retrogressive metamorphic event with  $P = 4-5,5 \text{ kb}$  ( $P_{\text{tot}} > P_{\text{H}_2\text{O}}$ ),  $T = 550 - 760^{\circ}\text{C}$  (from south to north of the Lineament) and low  $f\text{O}_2$ . Further Migmatites and conformable pegmatites were produced.

In Tantalite Valley the proto-mottled metagabbro was emplaced along the axis of the Lineament. This led to <sup>partial</sup> remobilization of granite gneiss and metasomatic alteration by the latter. Proto-metagabbro, probably a gabbronorite, was then intruded into relatively dry rocks producing a narrow contact aureole and generating granitic partial melts and cordierite-anthophyllite/pelitic hornfels restites. Temperatures in the aureole were from  $575 - 820^{\circ}\text{C}$ , pressure was constant at about  $5,5 \text{ kb}$  ( $P_{\text{tot}} > P_{\text{H}_2\text{O}}$ ) and  $f\text{O}_2$  low. The proto-mottled metagabbro and proto-metagabbro then underwent thorough hydration metamorphism to

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metagabbros. Gabbronorite was then intruded into metagabbro at temperatures of  $\sim 1200^{\circ}\text{C}$  as plugs, sheets and dykes and cooled slowly.

Following this, ultramafic plugs were emplaced into the Complex and in the enclosing gneisses. A drop in pressure, then occurred, probably as a result of rapid uplift, reactivating the Shear Zone in the process. Shearing accompanied by the introduction of water led to some low-grade metamorphism and partial melting to produce discordant pegmatites.

The Tantalite Valley Complex is regarded here as a variety of layered complex. Successive tapping of a magma chamber in which ultramafic cumulates sank to the floor leaving cumulus-enriched basic liquids gave rise to the emplacement sequence observed in the Complex. The magmas from which these rocks were derived, were evolved, aluminous tholeiites depleted in some lithophile elements. Element concentrations in mafic rocks from the Tantalite Valley and Kumkum areas rules out the possibility that these suites can be related by crystal fractionation but rather that lateral inhomogeneity in the source area is involved. All gabbroic rocks in these areas were derived from unusual magma-types which suggest an origin from a depleted source area, as a result of tectonism associated with the Tantalite Valley Shear Zone.

The charnockite hypothesis for the mafic-ultramafic rocks of Jackson (1976) and Toogood (1976) is rejected.

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DEDICATION

This work is dedicated to the loving  
memory of

SYLVIA ELLEN KARTUN (née Kruger)

(1928 - 1977)

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## PREFACE

Definitions

- Mixed gneiss - gneiss constituted by materials of originally sedimentary and magmatic origins (Symposium on migmatite nomenclature, 1960)
- Metabasite - a metamorphosed mafic rock of igneous or sedimentary origin (Miyashiro, 1973)
- Granolite - a rock diagnostic of the granulite facies (Winkler and Sen, 1973)
- Granulite - refers to a metamorphic facies or metamorphic zone
- Mylonite - coherent rocks with a bimodal grain-size distribution in which recrystallization (including neomineralization) of the finely comminuted portion is almost complete and in which larger lensoid augen are porphyroclastic and show evidence of strain (modified after Knopf, 1931)
- Metatexis - the conversion of metamorphic rocks of suitable composition to heterogeneous migmatites by selective melting of their quartzo-feldspathic component. (Mehnert, 1968). Metatexis is characterized by metatexitic styles of migmatization ie blastic, stromatic, agmatitic and phlebitic (see definitions in Mehnert, 1968)
- Diatexis - the conversion of metamorphic rocks of suitable composition to nearly homogeneous granitoids by melting of their quartz-feldspar-biotite component (Mehnert 1968). Diatexis is characterized by diatexitic styles of migmatization ie schlieric, nebulitic and homophanous (see definitions in Mehnert, 1968)
- Venitic - refers to neosome of veined migmatite arising in situ by partial melting of a rock (Mehnert, 1968)
- Arteritic - refers to neosome of veined migmatite which is formed elsewhere and is injected into the rock, usually as a melt (Mehnert, 1968)
- Neosome and Palaeosome - the term neosome is used for the migmatitic component of the rock which shows evidence of being younger than the palaeosome or host rock (Mehnert, 1968)

Agmatite        - a migmatitic breccia with angular fragments  
                  and a crystalline "granitized matrix"

.

Mineral Abbreviations Used

|     |                 |     |                 |
|-----|-----------------|-----|-----------------|
| Ad  | - Andalusite    | Ky  | - Kyanite       |
| Al  | - Almandine     | Ma  | - Magnetite     |
| An  | - Anorthite     | Ms  | - Muscovite     |
| Ap  | - Apatite       | Ol  | - Olivine       |
| At  | - Anthophyllite | Op  | - Orthopyroxene |
| Bi  | - Biotite       | Ox  | - Opaque oxides |
| Cc  | - Calcite       | Ph  | - Phlogopite    |
| Ccp | - Chalcopyrite  | Pi  | - Pinite        |
| Cd  | - Cordierite    | Pl  | - Plagioclase   |
| Ch  | - Chlorite      | Po  | - Pyrrhotite    |
| Cm  | - Cummingtonite | Pr  | - Prehnite      |
| Cp  | - Clinopyroxene | Px  | - Pyroxene      |
| Cz  | - Clinozoisite  | Py  | - Pyrite        |
| En  | - Enstatite     | Qz  | - Quartz        |
| Ep  | - Epidote       | Sau | - Saussurite    |
| Fa  | - Fayalite      | Se  | - Sericite      |
| Fo  | - Forsterite    | Si  | - Sillimanite   |
| Fs  | - Ferrosilite   | Sn  | - Sphene        |
| Ga  | - Garnet        | Sp  | - Spinel        |
| Gd  | - Gedrite       | Sr  | - Serpentine    |
| Hb  | - Hornblende    | St  | - Staurolite    |
| Hd  | - Hedenbergite  | Tr  | - Tremolite     |
| Il  | - Ilmenite      | Zc  | - Zircon        |
| Kf  | - K-feldspar    | Zo  | - Zoisite       |

Some Symbols Used

|                          |                                            |
|--------------------------|--------------------------------------------|
| $P$ or $P_{\text{tot}}$  | = total pressure                           |
| $P_f$                    | = fluid pressure                           |
| $P_{\text{H}_2\text{O}}$ | = partial pressure of $\text{H}_2\text{O}$ |
| $P_{\text{CO}_2}$        | = partial pressure of $\text{CO}_2$        |
| $X_{\text{H}_2\text{O}}$ | = mole fraction of $\text{H}_2\text{O}$    |
| $X_{\text{CO}_2}$        | = mole fraction of $\text{CO}_2$           |
| $T$                      | = temperature                              |
| $\bar{X}$                | = sample mean                              |
| $S$                      | = sample standard deviation                |
| $n$                      | = number of observations in sample         |
| $M$                      | = melt                                     |



## CHAPTER I

### INTRODUCTION

#### 1.1 General Statement

The area is situated in the extreme southeast of SWA (Namibia) and is bounded by meridians  $18^{\circ}35'$  and  $19^{\circ}05'$  east, and latitudes  $28^{\circ}30'$  and  $28^{\circ}50'$  south (Figs 1 A and B). It covers an area of  $760 \text{ km}^2$ , comprising the Namaqua Metamorphic Complex (see Beukes 1973; Moore 1975; Toogood 1976; Jackson 1976; Blignault 1977) which is overlain by rocks of Nama and Karroo age to the north of the study area (Fig 1 B). The Namaqua Metamorphic Complex forms part of the Namaqua Mobile Belt, which includes all the rocks lying to the west of the Kaapvaal craton older than the Nama, Gariep, Sinclair, Koras and younger sequences (see Jackson 1976).

The north of the area is characterized by flat plains with well developed dendritic drainage and scattered inselbergs. Relief drops gradually to the south until the rugged mountainous strip along the Orange River is reached. This averages 800 m in elevation, with Signaalberg (1 069 m) forming the highest point in the area. The Orange River has incised a gorge some 600 m deep cutting across all

structures in the metamorphic rocks. It is the only perennial river in the area and all other streams are tributary to it.

The northern plateau is correlated with the Bushmanland plateau to the south of the Orange River. The latter is of Tertiary age and forms part of the African erosion cycle (King 1963). The mountainland nearer the river was formed from an even younger (800 m) erosion surface as a result of Tertiary rejuvenation of the Orange River (King 1963). The well-developed drainage system owes its origin to the wetter climate prevailing during Tertiary times. In the Quaternary the climate became progressively more arid and tributaries became choked with detritus.

The climate is arid to semi-arid with the rainfall averaging 85 mm per annum at Warmbad over the last ten years. Most of this falls in March as thunder storms. The area is characterized by the highest summer temperatures in Southern Africa. Winter temperatures are mild but nights can be extremely cold due to the low humidity.

The vegetation is typically xerophytic and sparse and consists mainly of succulents in the rocky areas, and grasses and bushes of euphorbia on the plains. Dry river beds and the banks of the

Orange River are marked by acacia trees especially, and shrubs.

Most of the area is sparsely populated and is too arid to be used for anything but karakul and goat farming. Where water is available fruit and vegetables are produced in small quantities. Tantalite Valley has long been known for the mineral wealth of its pegmatites which have been successfully exploited since the 1940's.

#### 1.2 Regional Setting of the Tantalite Valley Complex and the Kumkum Massif

The Tantalite Valley Complex (Fig 6) comprises in order of decreasing age: metagabbro, gabbro-norite and plugs of troctolite, peridotite, pyroxenite and norite or their altered equivalents. It is surrounded by a thin discontinuous zone of hornfels and rare bodies of tonalite and granodiorite. So called "Mottled metagabbro" and Inequigranular granite gneiss or orthogneiss predate the Complex, the former occurring as a narrow tail-like body extending West-north-west of the Main Body for some 7 km and the latter occupying large areas to the north of the Complex and as a thin envelope around the mottled metagabbro. The Complex proper is oval-shaped with a long axis 7 km long, trending

108°. It crops out on the farm Umeis 110 and Kinderzitt 132 and occupies a surface area of approximately 46 km<sup>2</sup>. The areas occupied by the various components of the Complex are given in Table (1). The Complex is situated in the core of the Tantalite Valley Lineament (Blignault 1972); a major dextral shear-zone which extends from Lüderitz to north of Pofadder and may extend beneath Karroo cover south-east of Pofadder as far as Van Wyksvlei (Joubert 1974a) a total distance of more than 500 km. This shear zone varies in width from 20 to 40 km and the core contains a 2 to 7 km wide belt of mylonites (Moore 1975; Toogood 1976). A minimum right lateral displacement of 85 km has been estimated for it by Toogood (1976) using the method of Ramsay and Graham (1970). The rocks surrounding the Tantalite Valley Complex are characterized by multiple deformation, metamorphism and migmatization. The grade of metamorphism increases abruptly as the axis of the Tantalite Valley Lineament (Tv1) is crossed ranging from medium-grade south of the Tv1 to high grade along and north of the Tv1 (eg Toogood 1976).

The Kumkum Massif is the name given here to an area dominated by black-weathering gabbroic and pelitic rocks whose topographic expression is as a rectangular, north-west trending body occupying an

area of about 200 km<sup>2</sup> and having an impressive relief above that of the surrounding plains. Its dimensions are 22 km by 8 km and it is situated north of the axis of the Tv1, about 12 km east of the Tantalite Valley Complex. Its boundaries are clearly defined on aerial photographs and ERTS imagery (Fig 2 ) and its similarity in appearance to the Tantalite Valley Complex in the latter led Viljoen et al (1975) to interpret it as a major gabbroic intrusion. The core of the Massif is dominated by Grey Gneiss which weathers to form an amphitheatre of low relief within the Massif. Inequigranular granite gneiss underlies large areas north of the axis of Tv1, especially the plains, and extends to north of Warmbad. According to this report it is the oldest intrusive rock in the Kumkum area and the second oldest in the Tantalite Valley area.

The area as a whole is characterized by abundant pegmatite and quartz-vein intrusion and forms part of the Namaqualand pegmatite belt (Hugo 1965).

### 1.3 Previous Work around Warmbad

Haughton and Frommurze (1936) carried out a reconnaissance of the Warmbad area and produced three maps covering the area between Nakop and the Fish

River. Wiid, in the period 1942 and 1944, mapped an area of 900 km<sup>2</sup>, which included the Tantalite Valley area. His unpublished report is of a very general nature. The discovery of economic mineralization in the Tantalite Valley pegmatites by Weidner during the Second World War led to the area becoming increasingly well-known. Cameron (1955) described and mapped several of the zoned pegmatites in detail and later drilling of the pegmatites led to the discovery of Cu-Ni mineralization in the mafic and ultramafic rocks of the Tantalite Valley Complex (Von Backström 1970). Intensive investigation by various mining companies followed which resulted in extensive drilling, mapping and sampling of the Complex. Unfortunately most of these data are not available. Beukes (1973) mapped the regional geology of the Warmbad area and Käch (1973) and Moore (1975) mapped the Complex in detail. Toogood (1976) mapped an area of 5 000 km<sup>2</sup>, bordering on Onseepkans and overlapping the present study area, and described the structure and metamorphism. Moore (1975) and Von Backström (1976) provided detailed descriptions of the Tantalite Valley Complex.

#### 1.4 Previous Work in Adjacent Areas

The earliest workers in the Namaqua Mobile

Belt were chiefly concerned with the exploitation of mineral wealth, particularly copper and pegmatite minerals. Gevers et al (1937) provided the first detailed description of the pegmatites south of the Orange River. Later workers were able to map the distribution of these pegmatites and to define a pegmatite belt extending from Vioolsdrift across Bushmanland and southern South West Africa into Gordonia (eg Hugo 1965).

After the early major contributions to the study of the Namaqua area by Coetzee (1941, 1942), Von Backström and De Villiers (1972)\*, De Villiers and Söhnge (1959), Von Backström (1964) and Martin (1965), it became clear that the Namaqualand Mobile Belt is made up of three distinct lithostratigraphic units:

1. The metalavas and metasediments of the Kheis Group in the extreme east. These were studied by Vajner (1974) who considered that they were older than 2,9 b.y., although Cornell (1977) found that the Marydale lavas yielded ages of approximately 1,9 b.y. This lends support to the conclusion reached by Botha et al (1976) that the Kheis and Matsap stratigraphic sequences are equivalent, since the latter was probably laid down between 1,8 and 1,9 b.y.

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\* This report is based on field work done in the early 1940's.

(Truswell 1970). The area is characterized by low pressure, high temperature metamorphism (Vajner 1974).

2. The Calc-alkaline Haib metalavas and the Violsdrift Igneous Suite of the Richtersveld province in the extreme west. These were studied by Beukes (1973), Ward (1974), Reid (1977) and Blignault (1977). Reid (1977) showed that these rocks are characterized by 1 800 m.y. ages, greenschist facies metamorphism and can be related to one another by models involving differentiation. Bertrand (1975) described metamorphosed equivalents of the Violsdrift rocks as well as older biotite gneisses from the Goodhouse area.
  
3. Elsewhere the Mobile Belt is made up of high grade metamorphic rocks and associated granite gneisses known as the Namaqualand Metamorphic Complex. Poldervaart and Von Backström (1949) and Von Backström (1964) described the occurrence of charnockitic adamellite porphyry from an area around Keimoes. Joubert (1971) presented a detailed metamorphic and structural history of a 12 000 km<sup>2</sup> area around Springbok. Geringer (1973), Joubert (1974) and Moore (1977) working



in the north-west Cape and Beukes (1973), Jackson (1976), Toogood (1976) and Blignault (1977) working in the southern part of South West Africa showed the wide distribution of this unit. Joubert (1974a) discussed wrench faulting in the Namaqualand gneisses and paid special attention to the Tantalite Valley Lineament which he called the Pofadder Lineament. Rocks of the metamorphic complex are characterized by 1 000 m.y. ages (Nicolaysen 1962; Nicolaysen and Burger 1965) and metamorphism ranging between Barrovian (Joubert 1971) and Abukuman type (eg Beukes 1973).

#### 1.5 Present Investigation

The aims of the present project were:

1. To prepare detailed geological maps of the Tantalite Valley and Kumkum areas.
2. To unravel the history and origin of the mafic rocks in the area by examining their field relationships, petrology, mineralogy and chemistry and to test whether the charnockite/granolite origin proposed for some of them by other workers (eg Toogood 1976; Jackson 1976) is

valid.

3. To use quantitative thermodynamic methods to determine the physical conditions prevailing during metamorphism of garnet plus cordierite-bearing gneisses, and to test whether the grade of metamorphism increases across the Tantalite Valley Lineament as proposed by Moore (1975) and Toogood (1976).
4. To determine the metamorphic history of the area and the bulk chemical constraints of the mineral parageneses of some of the metamorphic rocks.
5. To re-examine the relationship between the various granitic rocks and the mafic rocks.

The Tantalite Valley and Kumkum areas were mapped in a period of five months spread between 1976 and 1977. The results are presented as Map 1 (Scale 1: 12 666) and Map 5 (Scale 1: 38 000) respectively (see Folder). A working knowledge of the rock-types found in the areas was obtained prior to going into the field, from the samples collected and preliminary mapping by Dr AC Moore. The Tantalite Valley Complex proved to be the most difficult to map due to the large amounts of scree present and because of the similarity in hand

specimen of some of the rock types, particularly the metagabbro and gabbro-norite. To overcome these problems and to make the map as objective as possible it was decided to make thin sections from a large number of samples, taking care that these were in situ. Sampling stations were marked on a three times enlarged standard 1 : 38 000 monochrome aerial photograph of the Complex and the results of thin section work were superimposed. Eventually samples could be confidently assigned to one or the other rock-type based on the hand-specimen alone and these too were added to the photo. Finally, contouring the sampling points as well as further field mapping was used to produce the final map.

Mapping of the Kumkum Massif was hampered by the general lack of roads and by the ruggedness of the terrain. Differentiating the various mafic rock-types proved comparatively easy after a large number of samples, collected during an earlier reconnaissance of the area, had been examined in thin section. Mapping was done on an aerial photograph mosaic at a scale of 1 : 38 000 and the results were directly transferred on to an overlay. Locations of samples collected from the Tantalite Valley and Kumkum areas are given on Maps 2 and 6 respectively (see Folder).

In addition, drill core produced during exploration of the Tantalite Valley Complex by various mining companies was kindly made accessible to the author by Mr P Weidner Sr. Some of the core has unfortunately been lost, but logging and sampling of what remained proved useful in obtaining a three dimensional picture of parts of the Complex, in particular the ultramafic rocks (Fig 4 in folder).

A number of scheelite prospects were investigated in the Tantalite Valley area with Dr AC Moore. A detailed map (pace and compass) of deposit S (Map 4) and a sketch map showing the regional setting of the main scheelite prospects (Map 3) are presented. Samples were collected for Tungsten analyses and for more detailed examination.

Representative samples of the mafic rocks of Tantalite Valley and Kumkum were analysed for their major and trace elements in order to discuss their origins and mutual relationships. Microprobe data on the constituent minerals of these rocks were used in conjunction with these data to put constraints on the models proposed for their origins and to define the physical conditions prevailing during their history.

Bulk chemical analyses and microprobe data on metamorphic rocks were aimed at giving a more

detailed metamorphic history of the area. Representative bulk analyses of the granitic rocks were also carried out.

#### 1.6 Acknowledgements

My grateful thanks go to Dr AC Moore who suggested the project and supervised it throughout. Dr Moore generously made available much unpublished data and constantly inspired the author to greater heights both in the field and in the laboratory. Without his encouragement and selfless assistance it is doubtful whether this report could have been written.

Funding of the project was by the C.S.I.R. to whom go special thanks.

I am indebted to all my colleagues, too numerous to mention, in the Department of Geology and Geochemistry at the University of Cape Town who provided assistance with this project. Dr D Waters, Mr AP Le Roux and Dr AR Newton were prevailed upon to read parts of the manuscript and their constructive criticism is much appreciated.

Mr RS Richard is thanked for his help in the operation of the electron microprobe.

Mrs P McPherson, Miss J Barr and Mr AP Le Roux made significant contributions to the analytical work.

Dr JW Von Backström kindly supplied the locations of several samples collected by him from Tantalite Valley.

Mrs L Brown, Mrs S Strachan, Mrs F Camden-Smith and Miss I Heinrich assisted with the draughting.

My thanks go to Miss S Swift and Mrs J Ward who did an admirable job in deciphering and typing the manuscript.

Mr and Mrs P Weidner of Tantalite Valley and Mr and Mrs C Weidner of Swartkoppies are thanked for their warm hospitality during several field trips.

Finally I would like to express my gratitude and appreciation to my parents for the sacrifices they made and for the help and encouragement they gave during the course of my studies.

## CHAPTER II

## LITHOSTRATIGRAPHY AND TECTONIC CATEGORY

The stratigraphic classification of rocks affected by processes of deformation, metamorphism and anatexis (ie rocks of the meso- and katazone) is highly problematical. Various approaches have been tried but none has as yet achieved wide acceptance. A method, however, based on metamorphic-structural criteria appears to be the most promising approach (Hedberg 1970). Wynne-Edwards (1969) introduced the concept of "tectonic categories" as a result of his work in the Grenville Tectonic Province. The terms pre-tectonic, syntectonic and post-tectonic were used by him to divide the rock units according to how they had been affected by metamorphism and deformation during a single orogeny (The Grenville Orogeny). However, the criteria he used to classify rock units into these categories were inadequate as pointed out by Jackson (1974, 1976). He proposed a modified and more vigorous tectonic classification as follows:

- (i) Pre-tectonic: Rock units containing the structural and metamorphic imprint of the earliest recognizable period of deformation;

hence assumed to predate the onset of orogeny.

- (ii) Syntectonic: Rock units unaffected by the earliest period of deformation but showing the effects of later deformation.
- (iii) Post-tectonic: Rock units that are unfoliated and undeformed; hence assumed to post-date all tectonic activity.

Because of the controversial nature of the tectonic history interpreted for the Namaqua belt by various workers, Jackson (1976) arbitrarily assumed the entire tectonic history as a single orogenic event; a condition which is retained here.

In the present study area the Pink Gneiss, the Grey Gneiss Sequence including the hornfels zone around the Tantalite Valley Complex, rare granulite and Black Gneiss, all of which probably represent an original, dominantly sedimentary sequence, belong to the Pre-tectonic Group. The Mottled metagabbro, Inequigranular granite gneiss, Tantalite Valley Complex metagabbro and tonalite/granodiorite, concordant pegmatites and aplites are foliated and are probably syntectonic. The gabbro-norites (Tantalite Valley, Kumkum, Contaminated and Eselruh) and the mafic-ultramafic rocks of the Tantalite Valley Complex



and vicinity are poorly foliated but have undergone shearing and are probably late-tectonic.

Discordant pegmatites, quartz veins and Metasomatic hornblendite are post-tectonic.

On the basis of the structural succession repeated in a large number of basin and dome structures, Beukes (1973) proposed that the pre-tectonic gneisses in the Warmbad area could be divided into three basic units: (1) A basal Houms Rivier Formation (Pink Gneiss) comprising quartzo-feldspathic gneisses overlain by (2) the Umeis Formation (Grey Gneiss Sequence) dominated by quartz-feldspar mica gneiss but including several rock-types eg calc-silicate, amphibolite, which in turn is overlain by (3) the Arus Formation (Black Gneiss) comprising a monotonous sequence of garnet-cordierite-sillimanite gneisses.

Beukes (1973, p 107) stated, however, that the possibility exists that there is more than one Umeis Formation and more than one Houms Rivier Formation. Toogood (1976) tested the above model for the Onseepkans area where similar rock-types and large basin and dome structures occur and found that there is indeed more than one Pink Gneiss horizon in the area. Furthermore, he found that basin and dome

structures are late-deformational features of the area and concluded that no structural succession based on these folds can be considered meaningful. In areas where early structures were unaffected by later deformation he found that the three units defined above were completely intercalated. This conclusion was confirmed by the present author, although it is possible using the "main" or best developed part of each pre-tectonic gneiss unit to define the lithostratigraphy, and the original sequence proposed by Beukes (1973) is retained for convenience. Terms such as 'Pink Gneiss' 'Grey Gneiss' are used informally in this report but will be rigorously defined in Chapter III to avoid confusion.

## CHAPTER III

## LITHOLOGY

3.1 Introduction

This report is mainly concerned with the mafic rocks of the Warmbad area with attention only being given to the pre-tectonic gneisses where mineral parageneses in the latter were suitable for defining the conditions prevailing during metamorphism. Rock-types such as the Pink Gneiss and the quartz-feldspar-biotite gneiss (Grey Gneiss) in the Grey Gneiss Sequence were therefore not studied in any detail. Descriptions and petrogeneses of these rocks are mainly based on the previous work of Beukes (1973), Moore (1975) and Toogood (1976).

3.2 Pre-tectonic Group3.2.1 Pink Gneiss (Quartzo-feldspathic gneiss)

The term "Pink Gneiss" is used here to describe a comparatively massive rock consisting essentially of quartz, K-feldspar and plagioclase with minor amounts of mafic minerals. Toogood (1976) arbitrarily defined the quantity of dark mineral present as being less than 10%, otherwise the rock is called a Grey Gneiss (see 3.2.2.1).

This rock-type has a wide distribution in the Namaqua Metamorphic Complex and for this reason, it has acquired a confusing proliferation of names eg leptite, leptynite, aplogneiss. In the Warmbad and Onseepkans areas, rocks here termed Pink Gneiss have been called the Houms Rivier Formation and the Austerlitz quartzo-feldspathic gneiss formation by Beukes (1973) and Toogood (1976) respectively.

In the study area, Pink Gneiss occurs mainly in the cores of late-stage basin and dome structures (Map 1) and as lenses and layers in the Grey Gneiss Sequence.

The Pink Gneiss is a homogeneous unit except for rare lenses rich in sillimanite, cordierite, micas and quartz, fibrolite-quartz-mica knots which may occur on foliation surfaces, or scarce migmatite. Contacts with other rocks are generally sharp, although gradational contacts exist with Grey Gneiss and Augen Gneiss. The rock crops out as friable, reddish-brown, poorly foliated boulders (see Fig 10) and usually weathers to form flat lying plains, but also massive rounded ridges. Fibrolite ellipsoids where found, define a mineral lineation on foliation planes.

The rock is composed predominantly of quartz (av. 43%), perthitic microcline (av. 43%) and albite/oligoclase (av. 10%) with biotite ( $\gamma = \beta$  greenish brown,  $\alpha$  pale olive brown), muscovite, opaque oxides and accessory minerals making up the remainder (Moore 1975). The accessory minerals include sillimanite, garnet, zircon, apatite, sphene and hornblende. Muscovite, chlorite, calcite, epidote and limonite are secondary (Beukes 1973; Moore 1975; Toogood 1976). The rock is medium-grained, seriate to equigranular with interlobate to polygonal texture.

Metasomatism and/or metamorphism of a pre-existing sedimentary rock has been invoked in the past to explain the origin of these rocks (eg Lamont 1947; Jansen 1960; Poldervaart and von Backström 1949). Following geochemical studies by Kröner (1968, 1971a) and Beukes (1973), backed up by new field evidence (Geringer 1973), the Pink Gneiss is now regarded as having formed as a result of isochemical metamorphism of arkosic sediments. The rôle played by metasomatism is still, however, controversial. Kröner (1968) argued that there was no chemical evidence for large-scale metasomatic granitization, whereas Beukes (1973) observed that some of the Pink Gneiss found in the Warmbad area is enriched in feldspar over that found in arkose sensu stricto (ie from 25% to 40%) and he ascribed the latter to generation

of K-feldspar during the change from medium to high grade metamorphism of a rock containing muscovite and quartz.

### 3.2.2 The Grey Gneiss Sequence

This unit (Fig 10) is made up mainly of quartz-feldspar-biotite gneiss (Grey Gneiss) but also includes amphibolites, calc-silicate rocks, marble, quartz-muscovite schist, augen gneiss, hornblende-gneiss, aluminium-rich horizons, migmatites, mylonites, rare metaquartzites and the hornfels around the Tantalite Valley Complex.

The Grey Gneiss Sequence or the Umeis Formation (Beukes 1973), is the major rock unit found in the Warmbad area. Beukes (1973) found it impossible to establish any order of succession of the various rock-types in this unit, although he concluded that calc-silicate is always found just above the Pink Gneiss at the base of the Grey Gneiss Sequence. Lithological contacts between Pink Gneiss and quartz-feldspar-biotite gneiss (Grey Gneiss) are invariably gradational and Toogood (1976) has separated the two groups on the basis of their biotite contents, the latter having greater than 10% biotite by definition. Layers and lenses of Pink Gneiss are also found within the Grey Gneiss Sequence and vice versa. Grada-

tional contacts also exist among the members of the Grey Gneiss sequence itself: ie between quartz-feldspar-biotite gneiss and quartz-feldspar-hornblende gneiss and between quartz-feldspar-biotite gneiss and augen gneiss. Amphibolite (including hornblendite) augen gneiss, mylonite and migmatite are described under the Syntectonic Group.

### 3.2.2.1 Quartz-feldspar-biotite gneiss (Grey Gneiss)

These are typically well-foliated, micaceous, well-banded, greyish rocks which weather rusty-brown.

The main minerals found are quartz (av 47,5%, S=15,31), biotite and chlorite (av 9,68%, S=7,56), oligoclase/andesine (av 15,5%, S=12,34) and microcline/microcline perthite (av 15%, S=10,48) (Moore 1975).

Some modes of Grey Gneiss given in Table ( 2 ).

Muscovite, sericite, chlorite and epidote may be secondary. The texture varies from lepidoblastic to granoblastic and the grain size is fine to coarse.

The Grey Gneiss contains lenses rich in biotite, cordierite and/or sillimanite, layers of Pink Gneiss, hornblende gneiss, augen gneiss, amphibolite and boudinaged layers of calc-silicate. In addition, migmatization and mylonitization are severe, especially along the axis of the Tantalite Valley Lineament and in places these serve to obscure the pre-tectonic

characteristics of the rock.

Beukes (1973) called these rocks 'Schistose Biotite Gneisses' whereas Toogood (1976) named them the Orangefall biotite gneiss Formation. Jackson (1976) describes similar rocks from the Aus area as 'semi-pelitic metamorphites'. Rocks of this type are found throughout Namaqualand and are subordinate only to Pink Gneiss in extent of outcrop area.

Beukes (1973) proposed that these rocks were derived by regional metamorphism of clay-rich sediments enriched in iron and magnesium.

#### 3.2.2.2 Hornblende gneiss

Hornblende gneiss is very similar in all respects to the Grey Gneiss except that hornblende is present in amounts exceeding 5% (volume). With increasing hornblende content the rock grades into amphibolite. Hornblende occurs as the sole mafic mineral as porphyroblasts up to 2 cm long or as prisms aligned parallel to the foliation, or together with biotite.

The rock is composed mainly of quartz, plagioclase, biotite and hornblende, with minor pyroxene, sphene, epidote and microcline (Toogood 1976) and was probably originally a clay-rich sediment (Beukes 1973).



This rock-type is referred to as the Pelgrimsrust hornblende gneiss/schist formation by Toogood (1976) and is included under semi-pelitic metamorphites by Jackson (1976).

#### 3.2.2.3 Quartz-muscovite schist

Quartz-muscovite schists occur as layers in the Pink Gneiss or as conformable schistose pegmatites in the Grey Gneiss Sequence, south of, or along the axis of the Tantalite Valley Lineament. They comprise mainly quartz and muscovite with some biotite, sillimanite, cordierite, garnet, plagioclase, microcline and secondary chlorite and epidote. Muscovite imparts to the rock a characteristic lustre.

They were referred to by Toogood (1976) as the Kambreek quartz-muscovite schist formation.

#### 3.2.2.4 Metaquartzite

This rock-type is rare in the study area and is restricted to thin lenses in the Pink Gneiss and Grey Gneiss Sequence. Toogood (1976) states that its predominance south of the Tantalite Valley Lineament is due to displacement along the latter. The rock is equigranular, coarse-grained (av 3 mm) and is flecked by biotite and reddish-brown blotches,

probably caused by the oxidation of opaque minerals.

The mode of sample K175 is given in Table ( 2 ).

According to Toogood (1976) the rock may contain up to 20% opaque minerals giving it a strong brown colouration.

Toogood (1976) called this unit the Pella meta-quartzite and tentatively correlated it with the Kaaiaen quartzite of the Upington area.

#### 3.2.2.5 Calc-silicate and Marble

Metacarbonates are rare in the Grey Gneiss Sequence and occur as discontinuous lenses, zoned boudins (Toogood 1976) or bands. Beukes (1973) postulated that calc-silicate marks the contact between Pink Gneiss and the Grey Gneiss Sequence but Toogood (1976) working in the Onseepkans area, found calc-silicate in the Pink Gneiss. The calc-silicates are omitted from the map of the Geology of the Tantalite Valley Complex (Map 1) because of their rarity but are included in the equivalent map for the Kumkum area (Map 5) because they provide a useful marker horizon there.

Calc-silicate rocks crop out as dark-brown to black-weathering jagged ridges. They commonly have bands up to 5 cm wide, consisting of plagioclase, quartz-

rich layers and amphibole, pyroxene, garnet-rich layers both of which may be complexly folded. Differential weathering of these layers gives the rock a furrowed, rough surface. Moore (1977) states that the banding probably represents primary layering which has been subsequently enhanced by metamorphic recrystallization. In places calc-silicate can be seen to grade into amphibolite and marble. The presence of agmatitic neosomes (Mehnert 1968) consisting of quartz and microcline, indicates that the rocks have undergone some partial melting. Joints are commonly lined with abundant yellow-green epidote.

The fresh surface varies from pale green to dark greyish-green. Macroscopically grossular garnet, diopside, magnetite and quartz may be visible.

The rock is fine-to medium-grained and generally has a granoblastic texture. Modes for this rock-type are given in Table ( 6 ).

Quartz occurs as amoeboid to polygonal grains occasionally with strain shadows and lamellae developed. It is found as discrete grains, in veins cutting the rock and as rounded inclusions in plagioclase and may have a well-defined lattice preferred orientation.

Plagioclase ranges in composition from  $An_{36}$  to  $An_{80}$ .

Grains are commonly polygonal, sericitized and poikilitic with inclusions of quartz, opaque minerals, epidote, clinozoisite and sphene. Glide twins, bending of grains and strain shadows may be present.

Diopside (av  $C \wedge \gamma = 44^\circ$ ) occurs as pale green to colourless, disseminated, anhedral grains. It shows twinning and alteration to actinolite parallel to cleavages and grain boundaries.

Hornblende is intergrown with garnet or epidote or occurs as discrete, anhedral grains. It has iron-staining along cracks and may have inclusions of plagioclase and epidote. In sample TV335 (Table 6 ) it is partly replaced by biotite, sphene and epidote.

Tremolite showing faint pleochroism (Table 6 ) occurs as anhedral, disseminated grains.

Garnet, probably grandite, is orange-brown, anhedral and poikilitic with inclusions of opaque minerals, chlorite, clinozoisite, epidote, hornblende and plagioclase.

Epidote occurs as discrete, rounded to amoeboid grains with inclusions of sphene. It may show multiple twinning and a distinct lattice preferred orientation. Discrete grains are commonly rimmed by colourless or pale-green clinozoisite. The latter also occurs rimming garnet and hornblende, as in-

clusions in plagioclase, and as discrete laths.

Light brown to orange, faintly pleochroic sphene occurs as disseminated, rounded grains surrounding opaque minerals and hornblende as inclusions in garnet, epidote and biotite and may itself have inclusions of plagioclase.

Biotite occurs as flakes which are altered in part to chlorite. Prehnite may be found parallel to the cleavage in the latter.

Disseminated magnetite grains are subhedral to anhedral and may have inclusions of epidote and sphene and show alteration to haematite.

Marble is made up predominantly of calcite with variable quantities of serpentized olivine, diopside and calcic amphibole. Modes for this rock-type are given in Table ( 7 ).

Mineral assemblages of calc-silicate and marble are given in Tables 3 and 5 respectively, and the relative frequency of samples belonging to each class into which these can be divided is given in Table 4 for calc-silicates and Table 5 for marble.

In the Namaqua Mobile Belt calc-silicates are widespread but are never abundant (eg Beukes 1973; Joubert 1971, 1974; Toogood 1976; Moore 1977) except in the Aus area (Jackson 1976) and in the area west

of Upington (Geringer 1973). Most authors favour a metasedimentary origin for these rocks and Jackson (1976, p 64) postulates that the association of psammitic gneiss (Pink Gneiss) and calc-silicates corresponds to the classic limestone-sandstone shelf facies of sedimentary petrology. Jackson (1976) favours the name granofels, in place of calc-silicate for carbonate-free rocks described above but this term has not been used here because of its applicability to a large number of different rock-types.

#### 3.2.2.6 Sillimanite-rich Rocks

The Sillimanite-rich components of the Grey Gneiss Sequence are found as thin, discontinuous lenses or as fibrolite-bearing knots both of which are rarely encountered. The former may appear similar to the Black Gneiss in the field, consisting of dark brown-weathering outcrops which are usually rich in biotite and sillimanite, the latter characteristically weathering to a pink, clayey material. It is distinguished from the Black Gneiss by its friable nature and its field relationships. The aluminous rocks are dark brown to dark green on fresh surfaces and contain mainly biotite, sillimanite, quartz, cordierite and muscovite (secondary and primary) (Table 2 ). Beukes (1973) also identified

corundum, anthophyllite and andalusite in his samples, the latter being occasionally found with sillimanite. He regards these rocks as being pelitic intercalations in the metasedimentary Grey Gneiss Sequence.

Quartz-fibrolite-muscovite nodules (Fig 30) were found in the Grey Gneiss near sample locality AM161 (Map 2) and at sample locations K135 and K176 (Map 6). The knots, which are ellipsoidal or pancake-shaped, stand out in positive relief on the foliation planes of the rock, their long axes forming a distinct lineation parallel with that found in surrounding rocks. Nodule-bearing rocks may be found near rare layers of biotite schist and quartzite in Grey Gneiss or they may occur as thin, random layers in the latter.

Pancake-shaped nodules are flattened parallel to the foliation in the rock and along strike they form bluish-grey pencil-shaped bodies. The average dimensions obtained from the measurement of eight pancake-shaped nodules at K135 was 3,5 cm for the long axis, 2,4 cm for the short axis and a thickness of 0,5 cm. Ellipsoidal knots found near AM161 were recorded as having an average long-axis length of 3 cm with an average short-axis length of 2 cm.

Nodules weather yellowish brown to dark grey and can

be pulled off the surface of the rock. In hand specimen quartz, biotite, sillimanite and occasionally garnet are visible.

Under the microscope the nodules are seen to be zoned with biotite together with plagioclase, quartz and microcline around the circumference of a core of muscovite, quartz and fibrolite. Biotite and fibrolite define a foliation in the nodule parallel to that found in the host, and together with quartz are the main minerals present. The nodules have an equigranular interlobate texture with quartz averaging 1-2 mm in diameter.

The modes of several nodule samples are given in Table ( 2 ).

Muscovite occurs as optically continuous, poikilitic mats, some of which may measure 4 mm square or as discrete, disseminated flakes which may cross-cut biotite. Inclusions are mainly fibrolite and quartz, the latter having amoeboid grain boundaries or "myrmekitically" intergrown with the host as if it had reacted with the latter (Fig 34H). Muscovite may be bent and stained by limonite along cleavages and grain boundaries.

Quartz shows strain lamellae, sub-grain development and may have a well-defined lattice preferred



orientation. It has randomly oriented fibrolite and rutile inclusions, the latter being rare. It occurs as anhedral, disseminated grains or "intergrown" with muscovite.

Fibrolite occurs as discrete, brown fibrous bundles and needles some of which are bent and wrap around other minerals, or it occurs as inclusions in muscovite and quartz.

Biotite is found as disseminated laths concentrated around the circumference of nodules. It may have opaque minerals parallel to cleavage, inclusions of zircon surrounded by pleochroic haloes and be twinned (sample K176A).

Sericitized microcline, which may have flame-like albite lamellae, occurs interstitially or as inclusions in quartz. Plagioclase is found as sericitized, anhedral, disseminated grains or myrmekitically intergrown with quartz.

Quartz-fibrolite-muscovite nodules similar to those described above have been reported as occurring in Pink Gneiss (Beukes 1973; Toogood 1976; Moore 1977) semi-aluminous gneiss (Blignault 1977) and in aluminium-rich, banded paragneiss (Von Backström and De Villiers 1972, p 25) from the Namaqualand Metamorphic Complex. Moore (1977) found that the nodules

occur in certain bands in the upper part of the Pink Gneiss unit in the Namiesberg, near Pofadder and stated that the same was generally true for the rest of Bushmanland. It is interesting to note that Von Backström and De Villiers (1972) as well as Beukes (1973) report the association of nodular schists with pelitic and quartzitic lenses and Beukes (1973) used this as evidence for a sedimentary origin for Pink Gneiss in the Warmbad area.

#### 3.2.2.7 Hornfels around the Tantalite Valley Complex

Hornfels occurs as a thin, discontinuous band around the metagabbro component of the Tantalite Valley Complex (Fig 5 ). It is best developed along the south-east contact with the metagabbro where it reaches a maximum width of 50 m; normally it does not exceed 10m. In places the hornfels may resemble either the gneissic country rock or the gabbroic intrusives but is distinguished from these by its mineralogy, its texture and field relationships. Hornfels tends to be more massive than the well-foliated country rock and is texturally distinct from the metagabbro which has largely retained an ophitic texture. Staurolite and rare kyanite are only found in rocks along the contact with the intrusive.

The main types of hornfels recognised are pelitic, cordierite-anthophyllite, metabasite, orthopyroxene-bearing and psammitic.

#### 3.2.2.7.1 Pelitic Hornfels

Pelitic hornfels is massive, (Fig 24) but may have leucocratic and melanocratic layers which occur complexly folded. (Fig 25). It is fine-grained (av 2 mm) and equigranular polygonal, although porphyroblasts of almandine up to 6 mm in diameter are found. It crops out as dark-brown boulders which superficially resemble the igneous rocks of the Complex except for the characteristic spongy, pock-marked, weathering surface (Fig 25). Layering is found but it is usually widely spaced (up to several centimeters) compared with that developed in the surrounding gneisses.

On fresh surfaces it is mottled, with bronze-coloured biotite flakes set in a dark green to buff matrix. Sulphides may occur along fracture planes or disseminated in the matrix.

Mineral assemblages and modes of the rocks are given in Tables 8 and 9 respectively.

Idioblastic to hypidioblastic pink garnet occurs disseminated through the rock as porphyroblasts or is found occasionally along healed fractures. The latter

is taken to indicate that these planes of weakness were present before thermal metamorphism and either provided suitable sites for garnet nucleation (eg by providing channelways for migration of appropriate fluids) or were already lined with minerals which reacted to form almandine.

The garnet contains inclusions of quartz, sillimanite and opaque oxides and is partly altered to greenish biotite and chlorite along cracks and grain boundaries. Electron microprobe analyses of cores and rims of garnet showed that they are almandine-rich and that two types of zoning are present. The garnets of one sample (TV 4) shows a decrease in Mn, Ca and  $\text{Fe}^{2+}$  from core to rim, while Mg shows an increase. In the other sample (T19) zoning is reversed with Mn, Ca and  $\text{Fe}^{2+}$  increasing, and Mg decreasing from the core to rim (see section 6.6).

Sillimanite is found in leucocratic bands or patches or, in rare cases, forms a lineation. Crystals are acicular to fibrous (av length  $\sim 0,4$  mm) and may occur in radiating clusters. Sillimanite is transgressive to staurolite and biotite and also occurs as inclusions in the latter as well as in cordierite and almandine.

Andalusite which may occur together with sillimanite, characteristically has abundant inclusions of biotite, opaque minerals and staurolite and may itself occur

as inclusions in biotite, quartz and cordierite.

Large anhedral grains of kyanite showing good cleavage and parting and with alteration to chlorite and pinite occur only in sample TV4 (Fig 34A).

Magnesium cordierite of a limited compositional range (Table 43) is interstitial and may be partly or completely replaced by pale green pinite. It has inclusions of opaque minerals, staurolite, sillimanite, quartz, biotite and probable zircon which is surrounded by pleochroic haloes.

Staurolite is found as small, anhedral grains (avg grain size 0,2 mm) which are included in cordierite, andalusite or biotite. Idioblastic staurolite has been found only in sample TV276 (Table 9) where it is in apparent equilibrium with cordierite (Fig 33H). This particular sample is unusual in that it occurs as a block between mottled metagabbro and metagabbro at the Western end of the Complex. In general staurolite may be clear or have inclusions of opaque oxides, hercynite, quartz or plagioclase and in one rock (TV4, Table 2) it has been partially altered to chloritoid.

Plagioclase is andesitic, shows normal or reverse zoning and is partly altered to sericite and epidote. It is usually interstitial and polygonal.

Biotite occurs as laths (av. length  $\sim 0,4$  mm) and porphyroblasts having inclusions of opaque minerals, sillimanite, staurolite, andalusite, quartz and zircon and shows partial alteration to chlorite. Prehnite may occur along  $\{001\}$  in biotite. Usually this feature has been reported from basic rocks (Phillips and Rickwood 1975; Moore 1976) but this is apparently the first reported occurrence in a pelitic host.

Disseminated opaque oxides include ilmenite and magnetite which do not occur together; the former being found characteristically as plates and the latter as anhedral grains. Haematite and limonite are secondary. Chalcopyrite, pyrrhotite and minor pyrite occur mostly along fractures.

Quartz occurs as anhedral grains (av. dia.  $\sim 0,1$  mm) with amoeboid, interlobate or polygonal boundaries where in contact with other matrix minerals.

Hercynite occurs predominantly as pale green, anhedral grains included in other minerals, particularly cordierite. It has been shown qualitatively with the electron microprobe to contain large quantities of zinc.

Beukes (1973) identified potassium feldspar in his samples of pelitic hornfels but none was observed in the present study.

Secondary muscovite and chlorite occur cutting across biotite (with or without prehnite), and in rare cases staurolite and garnet. Muscovite may occur along the  $\{001\}$  cleavage of biotite where it is distinguished from prehnite by its optical properties and by its not being lensoid in shape. Minerals show signs of deformation, eg bending of biotite and chlorite; undulose extinction, sub-grain development and lattice preferred orientation of quartz and deformation twinning in plagioclase.

#### 3.2.2.7.2 Cordierite-anthophyllite hornfels

These rocks are found only as disconnected plug-like bodies, approximately 30 m in diameter, within the Grey Gneiss Sequence near the eastern contact of the Tantalite Valley metagabbro (Fig 26). The Grey Gneiss here is highly migmatized and has layers of pelitic hornfels. The latter may be interbedded with cordierite-anthophyllite hornfels as discontinuous layers up to a few metres thick. Sillimanite and anthophyllite are mutually exclusive, a feature also noted by Lal and Moorhouse (1969) and Ramsay and Davidson (1974), and when the latter is absent from the rock, eg AM170A (Table 9 ) the product is indistinguishable from pelitic hornfels sensu stricto. Cordierite-anthophyllite hornfels is

massive and medium-to coarse-grained and weathers to reddish-brown boulders.

The rock is equigranular with anthophyllite prisms having a decussate relationship with cordierite which is characteristically idiomorphic to polygonal (Fig 36E).

For mineral assemblages and modes see Tables 8 and 10 respectively.

Orthoamphibole has a low aluminium content (Table 43 ) and is hence anthophyllite as distinct from gedrite. It occurs as decussate or radiating prisms up to 6 mm in length (av 2 mm); a texture which has been explained as being due to the difficulty in nucleating orthoamphibole (Spry 1969). Anthophyllite has inclusions of phlogopite and cordierite, shows alteration to phlogopite, chlorite and very fine flakey crystals which are probably talc and may be bent.

Cordierite is more magnesian than in pelitic hornfels sensu stricto ( see Table 43 ) and occurs as polygonal grains ranging in diameter from 0,2 mm to 1,5 mm (av. 0,6 mm). Grains show simple twinning, pinitization and inclusions of phlogopite, anthophyllite, quartz and probable zircon, the latter producing pleochroic haloes. Garnet in trace amounts ( < 0,5% ) was found in sample TV102 (Table 10 ).



In spite of its strong pleochroism (Table 10 ), the brown mica is phlogopite (Table 43 ) and this may be related to the relatively high Ti content. It occurs as anhedral laths which may be bent and altered to chlorite and commonly contains inclusions of probable zircon surrounded by dark pleochroic haloes.

Quartz occurs interstitially as clear or yellow-brown, iron-stained, strained grains.

Plates of ilmenite partly altered to limonite may be bent where the rock has been crenulated and form the main opaque mineral found.

Chlorite (ripidolite) (Table 43 ) and muscovite are found as alteration products or grown across other minerals.

Beukes (1973) has reported the presence of K-feldspar and staurolite in these rocks but none was observed in the samples examined both in the field and in thin section in the present study.

### 3.2.2.7.3 Metabasite Hornfels

Metabasite hornfels superficially resembles metagabbro in hand specimen but differs from it in that it may contain almandine and abundant quartz and has an equigranular polygonal to interlobate texture. Metabasite hornfels grades into pelitic hornfels and

is intercalated with it. In some outcrops, it shows mafic and felsic bands a few centimetres wide which give the rock a gneissic appearance. The rock may show small folds on weathered surfaces, stellate tourmaline along fractures and intrusion by venitic quartz-feldspar neosome.

Modal analyses are given in Table (11).

Plagioclase (av.  $\sim$  An<sub>40</sub>) occurs as anhedral or polygonal grains which are altered in part to sericite, clay, calcite and epidote minerals and may be strained. Grains show slight normal zoning, deformation twinning and inclusions of calcic amphibole. Pale brown patches of high relief material in AM31B were found under the electron microprobe to be highly altered anorthite (Table 43). Antiperthite was observed in several samples.

Pink, idioblastic to hypidioblastic almandine porphyroblasts up to 5 mm in diameter occur sparsely distributed. They have inclusions of quartz with chlorite and hornblende occurring along grain boundaries.

Amphiboles include hornblende, actinolite, anthophyllite and cummingtonite. The last named (Table 43) is generally present as colourless twinned prisms (av. length 0,7 mm) which may have exsolution blebs of green, calcic amphibole and usually has a

decussate relationship with poikiloblastic actinolite or hornblende mats in the groundmass. The latter may have exsolution of colourless amphibole, probably cummingtonite.

Polygonal to anhedral quartz grains occur in clusters and show strain shadows and sub-grain development.

Biotite may have inclusions of zircon and epidote, both of which produce pleochroic haloes, and be discrete or associated with opaque minerals.

K-feldspar, showing carlsbad twinning and alteration to clay and sericite, is rare.

Sphene is honey-brown and anhedral and is either disseminated or found along cleavages in mica.

Opaque minerals include pyrrhotite, chalcopyrite, minor pyrite and pentlandite, magnetite and ilmenite, the last two never being found together. In sample TV9 215.3 (Table 11) pyrrhotite occurs along fractures and forming large (10 mm) anhedral crystals. Pentlandite is found in the latter as anhedral inclusions and flame-like blebs or along fractures and grain boundaries. Chalcopyrite surrounds pyrrhotite and is evenly disseminated in the matrix together with rare subhedral pyrite. The latter may also occur along fractures.

Muscovite and chlorite (ripidolite) (Table 43) occur cutting across and altering biotite.

The rock generally has an equigranular interlobate to polygonal texture, excluding the amphiboles, with an average grain size of 0,6 mm. Lattice and shape orientation is developed to varying degrees in quartz, plagioclase, micas and amphibole and strain features are common in these minerals.

#### 3.2.2.7.4 Orthopyroxene-bearing Hornfels

This closely resembles the gabbro-norite of the Complex in hand specimen but is distinguished by its texture, finer grain-size and its field relationships. The rock is fine-grained except for poikiloblastic porphyroblasts of bronzite (up to 4 mm long) and generally has a granoblastic texture. The texture of sample TV225 (Table 12) is unusual, in that it consists of leucocratic patches of about 7,5 mm in diameter, containing grains averaging 0,7 mm, set in a darker, finer-grained matrix (av 0,1 mm). (Fig 33G). The minerals making up both domains are the same ie bronzite, cordierite, plagioclase and phlogopite with the latter occurring in lesser amounts in the leucocratic areas, which have a texture dominated by poikiloblastic plagioclase which encloses rounded grains of cordierite. The inclusions are in optical continuity with one another and with cordierite in the matrix and this and the presence of corroded boundaries where plagioclase and cordierite are in

contact suggests a reaction relationship between the two. The overall effect of this texture is to give the rock a maculose appearance in hand specimen.

Spry (1969) attributes such textures to disequilibrium promoted by low ionic mobility, favoured by low stress and the absence of intergranular fluid. Hornblende and pyroxene may define a lattice preferred orientation.

Mineral assemblages and estimates of modes are given in Tables 8 and 12 respectively.

Poikilitic bronzite (Table 43 ) or hypersthene (av diameter 3 mm) contain rounded inclusions of cordierite and plagioclase and shows partial alteration to hornblende and less frequently orthoamphibole, along fractures.

Polygonal, magnesian cordierite (Table 43 ), (av diameter 0,1 mm) shows twin lamellae and abundant inclusions of biotite and ilmenite.

Anhedral flakes of mica present are phlogopite (Table 43 ), which is distinguished from the biotite of pelitic hornfels by its paler pleochroic colours and is usually less abundant than the latter.

Small polygonal plagioclase grains (0,1 mm av diameter) showing little twinning make it difficult to estimate compositions. In plagioclase-bearing samples the

latter varies between  $An_{32}$  and  $An_{55}$ . These grains are altered in part to epidote and sericite, may have inclusions of cordierite and show evidence of strain (eg strain shadows, sub-grain growth).

Hornblende occurs enclosing pyroxene or exists as discrete polygonal grains (av diameter 0,1 mm).

Plates of ilmenite and rare anhedral sulphide occur surrounded by epidote and phlogopite, as disseminated grains or along the schistosity in the rock.

Muscovite and chlorite are secondary after hornblende and phlogopite.

Mica and the inosilicates may have a distinct lattice and shape orientation, giving the rock a lepidoblastic texture in part.

#### 3.2.2.7.5 Psammitic Hornfels

This is distinguished from the other hornfels types by the absence of calcium amphibole and by the presence of large amounts of K-feldspar in some specimens and from the Grey Gneisses by its predominantly seriate (range 0,05 - 3 mm, av 1,5 mm) polygonal to interlobate granoblastic texture. However, the texture observed in AM30A (Fig 33A) superficially resembles that of a typical clastic rock, being composed of closely packed, rounded grains

of quartz and plagioclase between whose interstices is found a dark-brown clayey material. This texture is not uniformly developed in the rock and occurs in patches and lenses which rarely exceed 3 cm in width and 20 cm in length. The plagioclase is irregularly zoned and has a composition of  $An_{42}$  (Table 43 ). Anthophyllite (Table 43 ) as thin needles and lesser quantities of biotite and ilmenite are disseminated through the rock with minor sericite/epidote and chlorite occurring as alteration products of plagioclase and biotite respectively. It is most unlikely that such a texture could survive the degree of metamorphism and deformation undergone by these rocks (Moore 1975). Similar textures to those described above can however be derived by small degrees of partial melting of an albite-quartz-orthoclase mixture as shown by Mehnert et al (1973) (see section 7.1.5.1.).

Modes for samples of psammitic hornfels are given in Table (13 ).

K-feldspar shows simple and cross-hatch twinning and is partly altered to kaolin and sericite.

Anhedral or polygonal quartz may show strain shadows, sub-grain growth and in AM102 (Table 13 ) myrmekitic intergrowths with plagioclase.

Muscovite is generally secondary and occurs as randomly oriented laths, which may be bent, or as felted masses. However, in TV13 198, a drill core sample from the south-eastern contact of the Complex, muscovite forms subhedral laths in textural equilibrium with quartz and K-feldspar and is interpreted as primary. This is the only sample from all hornfelses examined where muscovite is apparently primary. Biotite is found as anhedral, irregular laths intergrown with chlorite and has inclusions of sphene, opaque minerals and probable zircon which is surrounded by pleochroic haloes.

Interstitial plagioclase ranges from  $An_{26}$  to  $An_{42}$ , and shows albite, pericline, carlsbad-albite and glide twinning.

Magnetite occurs together with pyrite but is not found with ilmenite.

Minerals may have a strong lattice preferred orientation and may be segregated into mafic (garnet and biotite) and felsic bands.

### 3.2.3 Black Gneiss

This formation is made up mainly of sillimanite-garnet-cordierite gneisses and is the equivalent of the Arus Formation (Beukes 1973) and the Jerusalem



Formation (Toogood 1976). Black Gneiss crops out over about forty percent of the surface area of the Kumkum Massif and is also found on both sides of the Houms River on Sandfontein 131 (Fig 3 ). Minor layers and lenses of this rock occur within the Grey Gneiss Sequence and in the Pink Gneiss near the contact of the latter with Grey Gneiss (Beukes 1973; Toogood 1976). Layers of Black Gneiss in Grey Gneiss become more abundant as the contact with the main body of Black Gneiss is approached and there appears to have been no break during the deposition of this sequence. Some of the layers in the Grey Gneiss near the contact may contain abundant biotite, pink-weathering sillimanite plus K-feldspar bands and idioblastic garnets and, although mineralogically akin to Black Gneiss, they have the friability typical of Grey Gneiss.

On the basis of structural relationships, Beukes (1973) postulated that the Black Gneiss represents the youngest metasedimentary formation in the Warmbad area. It has been intruded by the Inequigranular granite gneiss, gabbro-norites (Kumkum and Eselruh), amphibolites, pegmatites, aplites and quartz veins and has intercalations of granulite, Pink Gneiss and Grey Gneiss.

In outcrop and on aerial photographs (Fig 2 )

this rock superficially resembles the mafic intrusives and has sometimes been mistaken for such by previous workers, eg Viljoen et al (1975). The crystalline, compact nature of the Black Gneiss makes it resistant to erosion resulting in the formation of dark-weathering ridges (Fig 12) whose slopes are mantled by large, rounded boulders. The rock is characteristically banded and has a vitreous lustre on fresh surfaces. Leucocratic and melanocratic layers up to 2 cm thick are made up respectively mainly of microcline, sillimanite, quartz, plagioclase and garnet, cordierite, biotite. Pinch and swell structures and ptigmatic folding may occur in the light bands.

The Black Gneiss is fine-to medium-grained and may have rounded, flesh-coloured almandine porphyroblasts up to several centimeters in diameter, especially where the rock is in contact with the Inequigranular granite gneiss and Kumkum gabbro-norite. The garnets weather positively and give the rock a warty appearance. Biotite-rich horizons in the Black Gneiss have a spongy, dark-brown weathering surface. Sillimanite occurs as shiny light-grey patches or layers on the surface of the rock. Foliation planes are usually widely spaced, undulating and steeply dipping, making it difficult to get unequivocal

structural readings. Intrafolial folds and complex similar folds are found in some outcrops.

A microcline-rich variety of Black Gneiss, the Feldspathic Black Gneiss (FBG) occurs scattered throughout the Kumkum Massif (Map 5). These rocks weather to form hills made up of massive to poorly foliated, rounded, pink-coloured boulders. It superficially resembles Pink Gneiss except that the FBG may have almandine and cordierite and is crystalline instead of friable. The FBG apparently occurs as layers which are seen to grade into microcline-poor Black Gneiss and is intercalated with it.

The modes and mineral assemblages of Black Gneiss, including varieties, are given in Tables 16 and 14 respectively. Based on the minerals present a number of classes of Black Gneiss were established (Table 14) and the relative frequency of each class is given in Table (15) as a percentage of the total number of samples. Secondary minerals include muscovite, sericite, pinitite, chlorite and hornblende (Table 16 ).

Quartz is usually interstitial or occurs as inclusions in other phases and is found as clear, strained, amoeboid, interlobate or polygonal grains

which may have a weak lattice preferred orientation. Cordierite either rims garnet (Fig 36F) or occurs as interstitial anhedral or polygonal grains crowded with inclusions of opaque minerals, biotite, sillimanite, plagioclase, probable zircon, hercynite and in KK6, hypersthene. Pinitization is developed mainly along grain boundaries and along irregular cracks which may be stained yellow by limonite. Sector and polysynthetic twinning as well as strained grains can be noted. Zoning was not observed in grains examined by electron microprobe.

Garnet as porphyroblasts and/or in the matrix is almandine-rich (Table 43 ), usually pink, anhedral to hypidioblastic and may have inclusions of quartz, biotite, cordierite, plagioclase, zircon, magnetite, hercynite and sillimanite. Biotite, magnetite, haematite, limonite and rarely chlorite are found along cracks. It is commonly unzoned or very slightly zoned but samples exist where normal-(K35) and reverse-zoning (K101) are well-developed (see section 6.6). In K62 large sillimanite prisms cross-cut garnet (Fig 34E). Garnet rimmed by cordierite (Fig 36F) and opaque mineral (magnetite?) - garnet symplectites (Fig 36G) may be observed.

Two generations of sillimanite are developed in some

thin sections (a) coarse laths up to 2 mm long and 0,4 mm thick with inclusions of opaque minerals, biotite and cordierite and occasionally bent, and (b) tiny acicular and granular crystals found commonly along grain boundaries of plagioclase, K-feldspar, and cordierite and as inclusions in cordierite and garnet. These grains may also cross-cut garnet and biotite or define the outlines of a crenulation foliation. Sillimanite may be partially altered to sericite.

Two generations of biotite were distinguished optically and using the electron microprobe. The apparently older variety is anhedral, characteristically pale-brown, lower in titanium and is commonly cross-cut by sillimanite. The apparently younger biotite occurs as hypidioblastic, dark-brown, often glomeroporphyroblastic laths with a high titanium content, which are rarely cross-cut by sillimanite. In all other respects the two generations of biotite are identical ie they have a preferred orientation, may be bent, are unzoned as shown by the electron microprobe and have inclusions of apatite and zircon, the latter producing pleochroic haloes.

Plagioclases analysed range in composition from  $An_{12}$  to  $An_{53}$ . (Table 43 ). Grains are anhedral, disseminated interstitially, and are commonly sericitized

and clouded with small opaque mineral or spinel inclusions. Sillimanite may occur parallel to cleavages as in sample KK10D (Fig 34G). Twinning, including glide twins, and some zoning may be observed but the latter is not common. Myrmekite is frequently developed along the contacts between plagioclase and quartz. Plagioclase grains may be bent and show strain shadows.

Microcline, identified by its cross-hatch twinning, may be perthitic with flame-like albite lamellae (Fig 34F). Grains occur as porphyroblasts, disseminated in the groundmass or concentrated in bands. They are anhedral, sericitized and may have inclusions of sillimanite, biotite and quartz.

Hypersthene has only been observed in K237 and KK6. In the latter it occurs as poikilitic porphyroblasts up to 1 cm long with inclusions of zircon, plagioclase and biotite. Haematite staining and secondary hornblende occur along cracks and grain boundaries. In K237 hypersthene occurs as rare anhedral grains with biotite developed along its cleavages and epitaxially around grain boundaries. An analysis of pyroxene in KK6 is given in (Table 43 ).

Sample K101 contains gedrite (Table 43 ) which occurs as prisms commonly associated with biotite.

Magnetite usually occurs as anhedral, disseminated grains whereas ilmenite, with which it may coexist, typically is found as plates. Magnetite also occurs as inclusions in other minerals and commonly forms symplectic intergrowths with garnet. It is altered in some samples to red, transluscent flakes of haematite and yellow limonite.

Emerald green hercynite occurs as anhedral, disseminated grains, as inclusions in other minerals or associated with magnetite. In K101 green spinel, reddish-brown spinel and magnetite occur in intimate association often grading into one another (Table 43 ).

Secondary muscovite with small inclusions of quartz occurs near sillimanite and K-feldspar in some slides (Moore 1975).

Chlorite is a rare alteration product of biotite and garnet and may, where developed, have prehnite parallel to its cleavage.

Grains in these rocks are seriate to inequigranular, with a range from 2 mm to 0,03 mm being usual.

Average grain size of the groundmass is about 0,5 mm although fine-grained varieties like K66B average 0,08 mm. Garnets are the commonest porphyroblasts and average about 2 mm in diameter. Other porphyroblastic phases include microcline, which may occur as

glomeroporphyroblasts, hypersthene and sillimanite. Grain contacts are amoeboid to interlobate although polygonal contacts may be developed in some areas. Sillimanite and biotite are normally elongate and define the preferred orientation. Other minerals like cordierite, garnet, quartz and microcline may show a lattice preferred orientation. The texture of typical Black Gneiss can be seen in Fig (34D).

Rocks of the type described above have variously been described as aluminous gneisses, metapelites, khondalites and kinzigites by other workers.

Beukes (1973) ascribes the origin of these rocks in the Warmbad area to high-grade regional metamorphism of pelitic rocks of a definite chemical composition. Toogood (1976), however, pointed out that they cannot be described as pelitic in the Onseepkans area because they do not contain more than 90% combined micas, sillimanite, cordierite and garnet, a pre-requisite for the rock to be of unquestionable sedimentary origin.



### 3 - 3 The Syntectonic Group

#### 3.3.1 Introduction

Included in this section is a discussion of most of the intrusive rocks in the area, as well as rocks which were formed during metamorphism. Some of the rocks are of doubtful origin, eg granulite and amphibolite, but are provisionally included here.

All of the mafic rocks and some of the pre-tectonic gneisses (eg Black Gneiss, pelitic hornfels ) are characterized by a shiny black veneer which is popularly called "desert varnish". This coating, which is probably the result of chemical weathering, is composed mainly of iron and manganese oxides. For a detailed discussion on "desert varnish" as related to rocks of the Namaqua Metamorphic Complex, see Von Backström (1964).

#### 3.3.2 Syn-tectonic components of Pre-tectonic Gneiss

##### 3.3.2.1 Augen Gneiss

Augen gneiss is often developed in the Grey Gneiss Sequence (Fig 27) and occasionally in the Pink Gneiss where these units enclose layers of Inequigranular granite gneiss. Toogood (1976) postulated that there is a genetic relationship

between the orthogneiss and blastesis in some of the pre-tectonic gneisses. The rock is often extremely difficult to distinguish from deformed Inequigranular granite gneiss, but the former usually has a more uniform megacryst population, being composed mainly of pink microcline. Jansen (1960) described similar rocks from the Nuwerus area and ascribed a metasomatic origin to both granite gneiss and augen gneiss.

#### 3.3.2.2 Amphibolite including Hornblendite

Amphibolites occur as irregular layers, (Fig 11) lenses, pods, dykes and plugs within the metasedimentary rocks, particularly the Grey Gneiss Sequence, and in the Inequigranular granite gneiss and are common in the study area.

Most of the outcrops are too small to be shown on the scale of the Geological maps presented here and only the larger bodies are represented. The smallest bodies, of the order of centimeters in length, are characteristically "augen"-shaped, probably as a result of disruption and boudinaging of more extensive layers whereas at the other end of the scale, bands up to hundreds of meters long and tens of metres wide exist (Moore 1975). Amphibolite is usually conformable with the foliation in the host rock but

transgressive relationships have been observed with a decrease in grain-size being noted as the contact with the latter is approached. Gradational contacts with metasedimentary rocks, particularly calc-silicates, have been noted by Beukes (1973) and Moore (1975), whereas contacts with quartzofeldspathic gneiss is usually sharp (Moore 1975). The rock is characterized by a strong lattice preferred orientation defined by plagioclase and hornblende, the former mineral occasionally being concentrated as feldspathic bands. Amphibolite layers which have been deformed by isoclinal-, open- and crenulation-folds are useful in interpreting the structural history of the area. Amphibolite is commonly developed along the boundaries of large mafic bodies where it owes its origin to retrogressive metamorphism and shearing of the original rock. Younger and older pegmatites, aplites and quartz veins intrude amphibolite. Neosome frequently occurs in the form of stockworks producing agmatite structure with amphibolite.

The rock is fine- to coarse-grained and has a dark-brown to reddish-brown, weathering surface. Fresh surfaces are green to dark grey with hornblende, plagioclase, quartz and rarely sulphide being macroscopically visible.

The modes of amphibolite samples are given in Table ( 17 ).

Plagioclase varies in composition from  $An_{25}$  to  $An_{95}$  with an average of  $An_{52}$  ( $n = 17$ ), and may show zoning. It occurs as laths, polygonal grains and rarely as glomeroporphyroblasts showing alteration to sericite, cancrinite, clay, epidote minerals, calcite and prehnite. Albite, pericline, and carlsbad-albite twinning was observed as well as strain features such as glide-twinning, sub-grain development and undulose extinction. It may have inclusions of hornblende and opaque minerals and display a lattice preferred orientation.

Hornblende ( $\text{av } \angle \gamma = 21^\circ$ ) occurs as anhedral, polygonal or prismatic poikilitic grains with inclusions of quartz, plagioclase, apatite, biotite, zircon, epidote, sphene and possibly rutile; the last three minerals probably having been produced by exsolution. It is also an alteration product of pyroxene and occurs along cracks in plagioclase.

Grains may be zoned, with paler green cores compared to rims and show simple twinning parallel to  $\{100\}$ . Colourless amphibole (either grunerite or tremolite) may be intergrown with hornblende. The amphiboles analysed in samples AM76 and AM77 were found to be Fe-pargasitic hornblende and Mg-hornblende respec-

tively (Table 43 ).

Quartz occurs as disseminated anhedral or polygonal, strained grains which may show sub-grain growth and lattice preferred orientation or as clear, rounded inclusions in hornblende or in transgressive and concordant veins.

Biotite is found as anhedral, interstitial laths which may, be extensively altered to chlorite, have inclusions of zircon and apatite and have exsolution of rutile. Grains can be bent and strained and have sphene, chlorite and prehnite along the cleavage (Fig 32B).

Hypersthene occurs in two samples as relic grains surrounded by hornblende.

Clinopyroxene is pale green, anhedral or polygonal and is altered along cleavages and grain boundaries to hornblende. It may show twinning, exsolution of orthopyroxene and opaque minerals parallel to cleavages and inclusions of plagioclase and chlorite.

Sphene occurs as disseminated anhedral or rounded grains, as clusters and glomeroporphyroblasts associated with epidote and opaque minerals and as inclusions in hornblende, biotite and chlorite which may be surrounded by pleochroic haloes.

The epidote minerals occur included in plagioclase,

presumably as alteration products and may be surrounded by expansion cracks. They also occur in hornblende and biotite; as discrete anhedral grains, which may be twinned, associated with sphene and opaque minerals and as cross-cutting veins together with calcite and plagioclase.

Opaque minerals are magnetite and ilmenite which show alteration to haematite and limonite and occur as disseminated anhedral or skeletal grains which may be rimmed by sphene, epidote minerals, biotite and hornblende.

Interstitial, anhedral microcline, showing cross-hatch twinning and alteration to clay, occurs in TV113B , which is also cut by fine K-feldspar veins.

Disseminated apatite exists as stumpy prisms and colourless needles up to 0,7 mm long. Beukes (1973) and Toogood (1976) both found samples of amphibolite which contain garnet but this variety is scarce.

The grain size of the rock ranges from 4 mm to 0,04 mm with an average of 0,8 mm. Remnant ophitic textures were observed in some samples but the overall texture is equigranular to inequigranular polygonal to interlobate with a well developed preferred

orientation of hornblende, plagioclase and occasionally biotite and quartz.

Beukes (1973) postulated that the amphibolites in the Grey Gneiss Sequence in the Warmbad area were formed as a result of metamorphism of calcareous shales or marls which were interbedded with greywackes. However, on the basis of the present study, the possibility that some of these were derived from an igneous parent cannot be discounted.

#### 3.3.2.3 Granolite

Rare granolite (Winkler and Sen 1973) occurs in the Kumkum Massif as conformable layers intercalated with Black Gneiss and Grey Gneiss with which it has been folded (Moore 1975). It also occurs as xenoliths within the Inequigranular granite gneiss where the latter intrudes Black Gneiss. Along its contact with other rocks it is commonly sheared and amphibolitized and it has been intruded by quartz-tourmaline veins and pegmatites.

The rock is medium-grained, equigranular and has a weak but distinct schistosity defined by the mafic minerals. On fresh surfaces it is dark grey, holocrystalline and plagioclase, quartz and biotite can be identified. The texture of the rock can be described as equigranular (av. gr. size 0,7 mm)

polygonal to interlobate (Fig 36B). Plagioclase and hornblende may have a strong lattice preferred orientation.

The modes of granulite samples are given in (Table 18).

Plagioclase ranges in composition from  $An_{48}$  to  $An_{70}$  (av.  $An_{59}$ ,  $n = 8$ ).

Laths and polyhedra, some of which have undulose extinction, display albite, carlsbad-albite, pericline and glide-twinning. Grains are commonly sericitized and may have inclusions of plagioclase, quartz, hornblende, opaque minerals and apatite. Amphibole in K58 was analysed as Mg-hastingsitic hornblende (Table 43). It occurs replacing pyroxene and is generally inclusion-free although inclusions of quartz, plagioclase, apatite, biotite and opaque minerals do occur. The latter may be found as discrete rounded grains rimming hornblende. Schiller texture is observed in some samples.

Clinopyroxene (av.  $C \wedge \gamma = 42^\circ$ ) is weakly pleochroic in shades of green, and in K58 was found to be salite (Fe-bearing diopside) (Table 43). Grains are anhedral and poikilitic with inclusions of plagioclase, hornblende and opaque minerals; the latter taking the form of Schiller texture or tiny randomly



oriented grains which gives the mineral a cloudy appearance.

Orthopyroxene is strongly pleochroic and is ferrohypersthene in K58, (Table 43 ). Grains are anhedral with inclusions of plagioclase, hornblende and opaque minerals although these are less abundant than in the case of clinopyroxene. Schiller texture may be observed.

Quartz occurs as disseminated polygonal to amoeboid grains having strained extinction or as tiny inclusions in hornblende.

Ilmenite and magnetite are anhedral and disseminated and may have inclusions of hornblende, pyroxene and plagioclase and show alteration to haematite which, when it occurs along cracks in pyroxene, may increase the relief of that mineral dramatically.

#### 3.3.2.4 Mylonite

Mylonite, as defined in the preface, occurs predominantly along the length of the Tantalite Valley Lineament in a belt 4 to 7 kilometres wide (Moore 1975; Toogood 1976) and has been reported as occurring along the Kanabeam shear zone in the Aiais area (Blignault 1977) and the Kuckaus shear zone in the Aus area (Jackson 1976) both of which are

considered as north-westerly extensions of the Tantalite Valley Lineament. The south-easterly extension of the latter to Pofadder and beyond, under Karroo cover probably as far as Van Wyksvlei, has been called the Pofadder Lineament by Joubert (1974a). From the above, it therefore seems likely that mylonite occurs along the lineament for a total distance of about 500 kilometres.

Smaller, mylonite zones, up to a few meters in width mimicking the trend of the main shear zone are particularly prevalent south of the Tantalite Valley Shear Zone. (see Joubert 1974a; Toogood 1976 and Jackson 1976) but also occur north of the latter. The mylonites occupying the core of the main shear zone are not continuously developed, being completely intercalated and having sharp contacts with apparently unaffected gneisses. (see Toogood 1976, Jackson 1976). In the Onseepkans area, Toogood (1976) has noted at least two generations of mylonites. Mylonite in the present area commonly dips towards the north-east at near vertical angles and weathers to give tombstone-like, dark brown to pinkish slabs. It is characteristically inequigranular, with ovoid porphyroclasts set in and wrapped around by a fine-grained matrix, giving the rock a pseudo-amygdaloidal appearance (Fig 34B). Mylonite

is formed in the pre-tectonic gneisses and mottled metagabbro and to a lesser extent in metagabbro, In-equigranular granite gneiss, gabbro-norites, ultra-mafic rocks and concordant pegmatites. It has been deformed by three generations of folds, starting with rootless intrafolial folds, followed by tight, angular, isoclinal folds having fold axes parallel to the axis of the Tantalite Valley Lineament and finally open folds with fold axes at almost right angles to the axis of the Tantalite Valley Lineament. The rock comprises ovoid porphyroclasts of quartz, plagioclase, microcline and rarely muscovite set in a fine grained matrix of biotite, quartz, plagioclase, chlorite and muscovite which appears to have rotated the porphyroclasts and swirled around them. Accessory minerals include disseminated zircon, apatite and opaque oxides and sphene which is associated with chlorite.

Microcline appears to be found only as porphyroclasts which are either perthitic and contain inclusions of biotite, zircon and muscovite or are exsolution-free and may be replaced along grain boundaries by myrmekitic intergrowths with quartz (Fig 34C). Shearing stress along the boundaries of the original megacryst has produced a high energy at this site and has generated

abundant randomly oriented plagioclase nuclei, a situation favouring myrmekite precipitation (Phillips and Carr 1973). Toogood (1976) noted that upon increasing deformation in the rock the amount of perthite exsolution increases in microcline.

Quartz commonly shows mortar structure; with porphyroclasts, showing strain features such as undulose extinction, deformation lamellae and sub-grain development, being surrounded by tiny anhedral, unstrained grains. Where this process has gone to completion quartz is no longer found together with microcline and plagioclase as a porphyroclastic phase. (Toogood 1976).

Plagioclase in porphyroclasts is commonly sericitized, has deformation twins and undulose extinction and may have inclusions of biotite which are much coarser than that found in the matrix (Fig 34B).

Muscovite grew after biotite as is seen by cross-cutting relationships with the latter. It occurs both in the matrix, growing mostly parallel to the foliation defined by biotite and chlorite, and as rare, strained porphyroclasts. Matrix laths have opaque minerals parallel to cleavage and show bending and undulose extinction.

Biotite (  $\gamma = \beta$  dark olive brown  $\propto$  pale brown) occurs mainly in the matrix as laths which may be bent and altered to chlorite or as inclusions in porphyroclasts (Fig 34B). Hornblende, when it occurs, as in mylonitized mottled metagabbro and metagabbro, is commonly streaked out, chloritized and may show kinking.

The amount of retrograde chlorite and muscovite increases with severity of mylonitization; the limit of the latter process producing a rock consisting of alternating light and dark bands made up entirely of fine rock flour with porphyroclasts being rare to absent (Jackson 1976, Toogood 1976).

Toogood (1976) postulated that the mylonites occurring along the Tantalite Valley Lineament formed by a process of dislocation and recovery under conditions of ductile deformation characterised by moderate strain rates (  $\sim 10^{-5}$  to  $10^{-6}$ /sec). The origin of the mylonites is directly linked with that of the Tantalite Valley Lineament which formed under conditions of high P-T (Toogood 1976) and remained active for a long period after this (at most  $\sim 1\ 000$  my) when eventually comparatively low crustal levels favoured low-grade metamorphism characterised by chlorite and muscovite which are found in the mylonites at present.

### 3.3.2.5 Migmatite

These rocks, generally believed to be of anatectic origin (eg Mehnert 1968; Fyfe 1971; Winkler 1974) are developed in pre-tectonic and syn-tectonic rocks over most of the present study area. In the brief description of them which follows, the concept of 'migmatitic grade' as outlined by Jackson (1976) is used as well as the migmatite terminology of Mehnert (1968).

Migmatitic grade is a measure of the degree of homogenization achieved between palaeosome and neosome, ie degree of anatexis. Mere quantity of neosome is insufficient to indicate grade of migmatization because the former is largely dependent on bulk composition of the rock undergoing melting. (This is shown in the case of Grey Gneiss and Pink Gneiss where migmatite is common in the former and scarce in the latter, even when the two are intercalated). Nor is migmatitic grade equivalent to metamorphic grade although conditions of the Upper Amphibolite facies must be reached before melting will occur in rocks (Winkler 1970). Migmatites can be broadly divided into two types, metatexites and diatexites, based on their characteristic migmatitic structures, but more importantly for the concept of migmatitic grade, by the type of neosome produced

(cf Preface). For rocks of suitable composition, there is a complete migmatitic sequence starting with blastesis (metatexites) and ending with homogeneous granitoids (diatexites) with the change from metatexites to diatexites probably representing a response to prograde metamorphism (Mehnert 1968; Jackson 1976). In the Aus area, Jackson (1976) found that diatexites follow and tend to obliterate metatexites.

In the study area migmatite is common in the Grey Gneiss Sequence, particularly the Grey Gneiss proper, is less well developed in Black Gneiss and Inequigranular granite gneiss and is rare in Pink Gneiss. Mafic rocks not forming part of the metasedimentary sequence generally have most migmatization along their contacts with other rocks.

Metatexitic migmatite is common over the whole area, as is seen from venitic (Fig 29) and less commonly arteritic neosome as well as blastite development in parts of the Grey Gneiss Sequence (the Augen Gneiss) and in country rock in contact with Inequigranular granite gneiss. Agmatitic structures are especially common in mafic rocks, particularly those forming layers in pre-tectonic gneisses such as the amphibolites and in calc - silicates.

The neosome generated in metatexitic migmatites is generally as quartzofeldspathic segregations which are often complexly folded together with the palaeosome or occurs as rootless folds in the latter. The neosome may be rendered flecky by the presence of garnet, the latter mineral being a conspicuous feature of the Kumkum migmatites.

Diatexitic migmatites are confined to the core of the Tantalite Valley shear zone and to parts of the Kumkum Massif. Schlieric and nebulitic structures, where palaeosome and neosome are no longer distinct are characteristic, although disrupted mafic bodies may still be identifiable in the rock (Fig 28).

The origins of the migmatites, aplites and old (conformable) pegmatites are probably linked and the pre-Tantalite Valley Complex age of the latter is confirmed by the presence of foliated pegmatite xenoliths in the tonalite and grandiorite and by the common association of concordant pegmatite with the Inequigranular granite gneiss (Beukes (1973)). The migmatite, according to Moore (1975), originated as a result of partial melting of gneisses during medium-grade regional metamorphism prior to the emplacement of the Tantalite Valley Complex. The metagabbro component of the latter produced a hornfels zone around itself upon intrusion and conditions were such that partial



melting could take place to produce a second generation of migmatite which was probably of local extent. The reaction zone around the Main Kumkum body is probably of similar origin as is the partial remobilization observed to have occurred in Inequigranular granite gneiss intruded by Kumkum gabbro-norite. As the Kumkum gabbro-norite is fairly widespread in the Kumkum Massif, so the migmatization induced by its intrusion would be expected to be widespread. These migmatites would probably be approximately contemporaneous with most of those found in the Tantalite Valley area. A third generation of migmatites was produced by local partial melting of the gneisses induced by lowering of the melting temperature by the introduction of water along shear zones, giving rise to young pegmatites and probably to quartz veins (Moore 1975).

#### 3.3.2.6 Old pegmatite and aplite

Old pegmatites are commonly schistose, unzoned, and intrude the pre-tectonic gneisses concordantly. They are generally uneconomic and are themselves intruded by young pegmatites and quartz veins. These pegmatites contain mainly quartz, feldspar and muscovite. Their extremely sheared nature in certain outcrops suggests that they

probably existed prior to the development of the Tantalite Valley shear zone. In the Kumkum area these pegmatites are commonly associated with the Inequigranular granite gneiss and may be beryl-bearing (Beukes 1973).

Aplite is fairly common in the Kumkum area where it is found intruding all rock-types except the young pegmatites and quartz veins but is absent from the Tantalite Valley area. It is particularly prevalent in the Inequigranular granite gneiss where it occurs as sheared, garnetiferous veins and dykes. Aplite may be so abundant that the granite gneiss appears to occur as xenoliths in the former (Toogood 1976).

Where the rock is foliated, which is usually the case, it looks superficially like the Pink Gneiss but is distinguished from the latter by its intrusive features and presence of garnet megacrysts. Aplite is characteristically white or pink weathering and may be folded.

The rock is composed of microcline and strained quartz in roughly equal proportions with accessory amounts of biotite (  $\gamma = \beta$  dark brown  $\propto$  light brown), plagioclase, pink garnet, opaque minerals and rounded, disseminated zircon. Microcline occurs as rare perthitic megacrysts up to 2 mm in diameter and in the groundmass. Both plagioclase and microcline are

sericitized. The garnet (av. dia. 0,3 mm) is anhedral to subhedral, disseminated and has inclusions of quartz and opaque minerals. The texture of the rock is generally equigranular interlobate (av. grain size 0,2 mm) with a distinct foliation defined by elongated opaque minerals, biotite laths and lattice preferred orientation of microcline and quartz. Quartz may be concentrated in layers.

### 3.3.3 Syntectonic Pre-Tantalite Valley Complex intrusions

#### 3.3.3.1 Mottled Metagabbro

The mottled metagabbro crops out mainly to the west of the Tantalite Valley Complex but small fragments of it are found in the latter as xenoliths and lenses (Map 1). The "tail" of mottled metagabbro comes to an abrupt end about a kilometer west of the Houms River where it forms the core of a tight, symmetrical, east-south-east plunging ( $40^{\circ}$ ) synform (Moore 1975). It weathers more readily than the metagabbro of the Complex and thus at the contact with the latter there is a sudden drop in the topography; the mottled metagabbro forming low rounded hillocks. These rocks, together with the enclosing gneisses, developed a penetrative schistosity during defor-

mation. Mottled metagabbro is discontinuously surrounded by an envelope of Inequigranular granite gneiss and foliation planes in both of these rock-types dip consistently to the north-east (Map 1). The mottled metagabbro is intruded by Inequigranular granite gneiss, producing large areas occupied by hybrid rocks, and by small bodies of tonalite which commonly occur along prominent north-south fractures and along foliation planes in mottled metagabbro. Mottled metagabbro xenoliths in tonalite contain no hornblende and the major mafic minerals are biotite, epidote and sphene (Moore 1975). A large number of discordant (young) pegmatites, some of which have been exploited for beryl and lithium minerals, intrude the mottled metagabbro.

The rock is inequigranular, fine-to medium-grained with rounded megacrysts of milky plagioclase and bluish quartz giving it a characteristic mottled appearance (Fig 16). The matrix is dark green and massive and consists mainly of hornblende and plagioclase (Fig 35E) which may have a relic ophitic texture. The modes of mottled metagabbro samples are given in Table (19). Chlorite, sericite, prehnite, epidote and hornblende are secondary.

The amphibole is magnesian hornblende (Table 43 ). It occurs as poikilitic laths (av length  $\sim$  0,15 mm)

with inclusions of quartz, plagioclase, opaque minerals, sphene and, rarely, zircon which occurs as anhedral to rounded grains up to 0,1 mm long and produces pleochroic haloes. The amphibole may show twinning and zoning, with greenish actinolitic cores surrounded by blue-green hornblendic rims.

Plagioclase occurs both as rounded glomeromegacrysts and megacrysts (dia range 0,7 - 4 mm) and as laths (0,15 - 0,7 mm long) in the groundmass. Detailed work indicates that plagioclase forming the megacrysts and that in the groundmass have the same composition. Analyses reveal normal zoning in megacrysts from  $An_{79}$  (cores) to  $An_{72}$  (rims), while Moore (1975) reported values of about  $An_{70}$  (cores) to about  $An_{30}$  (rims). The average value obtained using the Michel-Lévy Method is  $An_{47}$  and it is not clear why there is such a big difference between these and the electron microprobe results. In plane polarized light the centres of grains may be cloudy with alteration to sericite and epidote while the rims are clear. Glide twins and bending of grains may be found.

Biotite occurs as disseminated flakes, but may be found surrounding or as inclusions in the opaque minerals. Biotite is characterized by the highest  $FeO^*/MgO$  ratios of the mafic rocks of the Tantalite Valley area (Table 43 ). Lensoid prehnite occurs

along cleavages (Fig 32A) and the biotite may show alteration to chlorite (ripidolite) (Table 43). Where mottled metagabbro is in direct contact with Inequigranular granite gneiss, as in the case of sample AM256 (Table 19 ) all of the hornblende in the rock has been converted to chlorite and biotite. Chlorite is also developed along the schistosity planes in mottled metagabbro.

Ilmenite is the dominant opaque mineral present but ferritchromite was also identified (Table 43). The former occurs as plates or as disseminated grains with inclusions of apatite and biotite, and is commonly surrounded by epidote, zoisite and sphene. Ilmenite in sample TV275A is rimmed successively by sphene and biotite, the latter containing prehnite parallel to its cleavage (Fig 32A).

Quartz occurs mainly as rounded inclusions within hornblende but also as disseminated, amoeboid grains and as rare, rounded megacrysts.

Von Backström (1976) described this rock as a porphyritic, chilled-facies equivalent of the metagabbro while Käch (1973) mapped it as porphyritic amphibolite.

### 3.3.3.2 Inequigranular granite gneiss

This rock is apparently restricted to the

areas north of the axis of the Tantalite Valley Lineament, and is the dominant granitoid in the area. It weathers to form rocky plains and low-lying rounded hills made up of piles of brown exfoliating boulders, and has a well-developed rectangular joint pattern which has been exploited by the drainage. It is commonly foliated parallel to the regional foliation, with elongate augen of pink to white microcline and plagioclase or aggregates of these with quartz, averaging 3 cm in length, wrapped around by biotite which defines the foliation. Mineral elongation forms the lineation which is well-developed. Some parts of the granite gneiss are unfoliated and preserve the original granitic texture with rounded and/or rectangular megacrysts present. Fresh surfaces are grey to dark brown depending on the amount of biotite present. Garnet-bearing varieties are restricted to the Kumkum area where granite gneiss intrudes Black Gneiss. The intrusive relationship is evident from the large number of Black Gneiss xenoliths found in certain outcrops of granite gneiss and the occurrence of cordierite and sillimanite in the latter, (Toogood 1976) in addition to garnet, is here attributed to contamination from the Black Gneiss. The granite gneiss has also intruded Pink Gneiss and mottled metagabbro (Fig 14). Contacts are seldom sharp and are usually highly foliated.

Where layers of granite gneiss occur in the Grey Gneiss Sequence as a result of lit-par-lit intrusion (Beukes 1973) the surrounding gneisses commonly have developed in them feldspar megacrysts and it is difficult in the field to map the contacts between the two rock types. In addition, the enclosing gneisses frequently show signs of having undergone migmatization. Beukes (1973, p 205) ascribes the presence of megacrysts in the granite gneiss and in enclosing Grey Gneisses to potassium metasomatism, although Toogood (1976) pointed out that a large proportion of the megacrysts are plagioclase. Xenoliths of Grey Gneiss in granite gneiss are friable and have probably been hydrothermally altered.

The granite gneiss has been intruded by the Kumkum gabbro-norite which has partially remobilized it to produce a contaminated gabbroic rock (Section 3.3.5.2). All gradations exist between granite gneiss and Kumkum gabbro-norite. What Beukes (1973) terms "irregular mafic inclusions in Eendoorn granite" in his Fig 56 is actually an agmatite structure probably produced as a result of remobilization of granite gneiss into fractures in mafic rocks as a result of partial melting (Mehnert 1968). These structures were also observed in the Kumkum area and it is postulated here that most of the xenoliths of Kumkum granolite (Kumkum gabbro-norite)



found by Toogood (1976) in the Beenbreek granite (Inequigranular granite gneiss) can be explained by this mechanism. The granite gneiss of the Kumkum Massif is also intruded by aplite veins. Lenses of amphibolite and bands of mica, as well as abundant concordant and discordant pegmatites are also found in the granite gneiss.

Modes of granite gneiss samples are given in (Table 20 ). Secondary minerals are chlorite, epidote and sericite.

Quartz occurs usually between biotite layers as elongated, strained grains and where the rock has undergone mylonitization these are disrupted and streaked out with the development of sub-grains. Some augen are made up entirely of aggregates of quartz with interlobate to polygonal grain boundaries. It may form myrmekite with plagioclase and microcline.

Plagioclase, with an average composition of  $An_{30}$  ( $n = 9$ , Michel-Lévy Method) may be slightly zoned, have strain shadows, show glide twinning and is occasionally myrmekitically intergrown with quartz and microcline particularly along grain boundaries. It occurs as sericitized augen, polygonal grains or bent laths in the matrix.

Microcline is found predominantly as porphyroblasts

which have inclusions of quartz, plagioclase, biotite and opaque minerals.

Biotite laths (av length  $\sim$  0,15 mm), commonly with zircon inclusions, together with some secondary chlorite and muscovite, define the foliation of the rock.

Garnet is found in the biotite-rich layers in the Kumkum area only as subhedral to anhedral, poikilitic grains with an average diameter of about 0,3 mm, having inclusions of biotite and plagioclase. Toogood (1976) found that cordierite rims occur around garnet in two of his samples of granite gneiss from the Onseepkans area. He observed that the cordierite only occurs when small quantities of sillimanite is present in addition to quartz and garnet, indicating a reaction of the type  $Al + Si + Qz \rightleftharpoons Cd$  has taken place.

The Inequigranular granite gneiss is equivalent to the syntectonic Eendoorn granite of Beukes (1973) and the Beenbreek granite of Toogood (1976). Other possible correlates are the Nababeep gneiss, which Joubert (1971) regards as the basement on top of which the para gneisses of the Namaqualand Metamorphic Complex were deposited, the Backrivier granite of the Upington area (Geringer 1973), the porphyroblastic granite gneiss of Namaqualand and Bushmanland (Joubert 1974) and the Tsirub gneiss of the Aus area (Jackson 1976).

### 3.3.3.3 Alkali-feldspar granite gneiss

This rock-type is restricted to a small area on the western bank of the Houms River where it occurs to the south of the culmination of the mottled meta-gabbro extension to the Tantalite Valley Complex. It crops out as an ovoid plug with its major axis of about 50 m in length trending parallel to the Tantalite Valley Lineament. It weathers to give bright pink, pitted boulders. Pits are filled with limonite and give the rock an almost gossanous appearance.

Specimens are pink, leucocratic, medium-grained with K-feldspar and quartz megacrysts with an average diameter of 0,7 mm, and possesses a lineation but no obvious foliation. The matrix of the rock is dominated by polygonal grains of clouded, usually untwinned plagioclase and microcline, in approximately equal proportions and lesser amounts of anhedral quartz all of which are surrounded by pink clayey material. Muscovite, haematite, apatite and sphene are accessories. The rock plots in the alkali-feldspar granite field of Streckeisen (1967).

Analyses (atomic absorption) reveal that Cu, Pb, Zn and Ag are present in concentrations of less than 100 ppm.

This rock-type is here tentatively correlated with

the Nautsis alkali-feldspar granite of the Onseepkans area (Toogood 1976). It is possible however that the rock may be the product of some sort of fracture filling process or may represent rare acid volcanics similar to those found around Goodhouse.

### 3.3.4 Syntectonic to late-syntectonic intrusions making up the Tantalite Valley Complex.

#### 3.3.4.1 Metagabbro

In terms of surface exposure area this is the second most abundant rock-type found in the Complex (Table 1 ) and consists dominantly of plagioclase and calcic amphibole with a remnant ophitic texture. It weathers to form black, conical hills with slopes covered by abundant boulders of scree, eg Signaalberg (Fig 8).

The rock is generally massive except along its contact with the pre-tectonic gneisses where a foliation has been developed. Some mineral-graded layering is found and is restricted to the south-east of the Complex. The layering, shown in Fig (22), is of three types: (i) isomodal (I) (ii) normal grading (N) and (iii) reverse grading (R) and is developed over a vertical distance of about 50 cm after which it is replaced by massive metagabbro.

Its lateral extent is uncertain because of scree cover. The following sequence is observed going from the bottom to the top of Fig (22):-

N        I        R        I        N        I        R        I.

ie an alteration of N-I and R-I pairs. These layers strike parallel to the contact of the Complex and dip steeply towards its centre.

Within the metagabbro are schlieren of coarse-grained metagabbro (hornblende up to 4,5 cm long and plagioclase up to 6 mm long) (Fig 20). Where layering is developed in the metagabbro the schlieren are generally parallel to it.

In the vicinity of shear zones, the metagabbro is often mylonitized and chlorite becomes abundant.

The metagabbro has developed a thin envelope of hornfels where it is in contact with the country rocks, proving its intrusive origin. It has itself been intruded by gabbro-norite, mafic-ultramafic plugs and discordant pegmatites. Veins of tourmaline and anhydrite may cut the rock and the former is common along joints. Near the west end of the Complex xenoliths of metagabbro occur in finer-grained metagabbro (Fig 21) indicating that there may have been more than one phase of intrusion.

On fresh surfaces the metagabbro is dark green or

grey depending on whether it is hornblende- or plagioclase-rich. In the latter case, the rock may closely resemble the gabbro-norite of the Complex but is distinguished from the latter by lack of visible pyroxene and greater abundance of macroscopic sulphides, particularly chalcopyrite and pyrrhotite. Plagioclase in the metagabbro is dark grey or milky depending on the presence of inclusions and degree of alteration.

The modes of metagabbro samples are given in (Table 21 ). Chlorite, sericite, calcite, prehnite and epidote minerals are secondary.

Analysed plagioclase varies between  $An_{62}$  and  $An_{77}$  (Table 43 ) while optical determinations by the Michel-Lévy Method for 42 samples give an average value of  $An_{60}$ . It may show normal zoning.  $An_{75-30}$ , or less commonly reverse zoning,  $An_{35-40}$ , and deformation features such as strain shadows, sub-grain development, glide twins and bending and microfracturing of grains. These are normally clouded with alteration and minute opaque mineral and spinel inclusions but have clear rims. Sericite, epidote minerals and calcite occur as alteration products along cleavages and twin planes. Laths (av. length 4 mm) are usually randomly arranged but may show a

weak shape-and lattice-preferred orientation in the mineral-graded layers (Moore 1975).

Calcic amphibole is found in the following forms:-

(1) as large mats made up entirely of tiny, acicular actinolitic-hornblende laths (Fig 32D) (2) as mats made up of pale-green actinolitic-hornblende cores, rimmed by bluish-green Mg-hornblende; the latter often forming at the expense of the surrounding plagioclase as is seen from embayments of hornblende in plagioclase (Fig 32C) and (3) as poikiloblastic laths made up entirely of blue-green hornblende containing inclusions of plagioclase and quartz. Second generation Mg-hornblende and actinolite also occur along cracks and cleavages in plagioclase. In extremely rare cases hornblende has relic cores of hypersthene and augite.

Sub-grain development, twinning, bending of grains and Schiller texture were observed in several specimens. Grains average 4 mm in length and show alteration to chlorite, biotite and calcite.

In some samples complex coronas are found in amphibole in contact with plagioclase. The following parageneses were found in sequence from the cores to the rims of coronas:-

(a) anthophyllite plus disseminated green spinel

chlorite plus minor biotite, hornblende, green  
spinel and opaque symplectite

hornblende

spinel/hornblende symplectite

(b) chlorite plus minor biotite

hornblende

spinel/hornblende symplectite

(c) actinolite plus minor chlorite

hornblende

spinel/hornblende symplectite

(d) anthophyllite

actinolite

spinel/hornblende symplectite

The symplectite zone adjacent to plagioclase is up to 0,7 mm wide and forms at the expense of plagioclase. A symplectitic intergrowth of opaque minerals and amphibole (Fig 34J) appears to be restricted to rocks in which orthoamphibole is found. The above coronas are probably the hydrous equivalent of coronas described later (section 4.1) between olivine and plagioclase in gabbro and in ultramafic rocks. Thus it is probable that the metagabbro may have originally contained some olivine.

Quartz, usually interstitial and amoeboid, occurs also as tiny, rounded inclusions in hornblende and



myrmekitically intergrown with plagioclase. It is not restricted to samples from the south and south-east of the Complex (Moore 1975).

Anthophyllite and cummingtonite are usually present as accessory minerals which have formed at the expense of ferromagnesian minerals.

Biotite is found mainly around opaque minerals. It shows alteration to chlorite (ripidolite) (Table 43) and lensoid prehnite commonly occurs parallel to  $\{001\}$  planes together with opaque minerals.

Sphene is yellow-brown, brown or colourless and is found associated with opaque minerals or as discrete, anhedral grains.

Enhedral to subhedral prisms of apatite are associated with opaque minerals or occur as inclusions within the other minerals. It can be greenish, clouded and may be bent.

Anhedral, amber-coloured grains of rutile may occur disseminated through the rock.

Opaque oxides (ilmenite and magnetite) are found along leavages in biotite and chlorite, as Schiller texture in hornblende and plagioclase or disseminated as anhedral grains. They show alteration to limonite and haematite, both of which serve to accentuate the relief of neighbouring amphiboles, and are often

surrounded by epidote minerals, chlorite, biotite, hornblende and green spinel.

The main sulphides present are pyrrhotite and chalcopyrite. The former is found either as small anhedral grains, interstitial to silicates or intergrown with chalcopyrite particularly along grain contacts. The latter occurs as anhedral, interstitial grains up to 5 mm in diameter, unevenly disseminated throughout the rock. Greater proportions in some areas suggest a possible layering. Pyrite is found as rare subhedral to anhedral grains within and interstitial to silicates but may occur as veins. Pentlandite, in sub-economic amounts, is found as "flames" and anhedral lamellae in pyrrhotite. Chalcopyrite-pyrrhotite-pyrite bearing quartz-tourmaline veins may be responsible in some localities, eg Melkboomfontein, (Map 1) for introducing sulphides into the metagabbro. Where these veins are in contact with metagabbro the latter is converted in parts to biotite- and muscovite-rich rocks.

The metagabbro (Beukes 1973; Moore 1975) has been called an amphibolite by Käch (1973) and Von Backström (1976). The latter suggested that the metagabbro represents altered gabbro-norite but it is clear from the present

study and the work of Moore (1975) that the gabbro-norite was intruded after the metagabbro and is virtually unaltered. The metagabbro was probably formed from an original gabbroic rock which underwent complete uranization as a result of the introduction of large amounts of water (Moore 1975).

#### 3.3.4.2 Metagabbro (metadolerite) dykes

These north-west to north-east-trending dykes, varying from 0,5 - 10 m wide and 7 - 50 m along strike, are found only in the metagabbro. Dips range from vertical to near horizontal. Metadolerite is characterized by a well-defined preferred orientation parallel to and near its contacts, but towards the centre of a dyke the texture is ophitic. It has an average grain size of about 0,3 mm.

Mineralogically (Table 22) it is almost identical to the metagabbro and probably represents a late intrusive phase of the latter.

#### 3.3.4.3 Tonalite and Granodiorite

Tonalite is only found on the south-west side of the Complex where it intrudes the mottled metagabbro (Fig 15) as small plugs or dykes along foliation planes and major north-south fractures.

It contains numerous xenoliths of Inequigranular granite gneiss and mottled metagabbro and is enriched in biotite for distances up to 4 cm away from the latter probably as a result of partial assimilation (Fig 18).

A small lens of granodiorite occurs near the workers' quarters to the east of the Complex. Its intrusive nature is clear from the presence of abundant angular xenoliths of foliated amphibolite (av length  $\sim$  8 cm) biotitite and rounded xenoliths of Grey Gneiss and foliated pegmatite, the latter being up to 0,5 m in diameter with an average of about 2 cm (Fig 19).

The tonalite and granodiorite are leucocratic, medium-grained rocks with a well-developed foliation. They have a pinkish weathering surface, which may be pitted. Quartz and plagioclase occur as phenocrysts as well as in the matrix.

Modes for these rocks are given in Table (23). They plot in fields 4 and 5 of the I.U.G.S. classification diagrams of Streckeisen (1967) and can justly be called granodiorites and tonalites respectively.

Plagioclase (oligoclase or andesine) may show normal zoning, ie  $An_{53-35}$ , myrmekitic intergrowths with quartz, and extreme alteration to sericite, clay and epidote.

Microcline occurs as discrete grains or as anti-perthitic blebs in plagioclase.

Anhedral to polygonal quartz grains may be strained and have rutile inclusions.

Biotite (av length 1 mm) occurs in layers which define the foliation. It may be altered to pale-green chlorite and have inclusions of zircon.

Sphene may be found together with biotite and chlorite around opaque minerals, as discrete grains or around biotite.

#### 3.3.4.4 Gabbronorite

These unmetamorphosed rocks form the major component of the Complex in terms of surface exposure area (Table 1 ). A small plug of gabbronorite is also found on Sandfontein 131 (Fig 3 ) where it apparently intrudes Black and Grey Gneiss. Gabbronorite crops out either as conical black hills with scree giving the impression of lava flows (Fig 7 ) or it tends to form sheets and sills which weather to give gently undulating topography. The gabbronorite is represented by a

wide range of compositions ranging from quartz-bearing to olivine - bearing varieties, and from norite to plagioclase - rich rocks. True gabbronorite is the most common type. Gabbronorite is often similar in appearance to metagabbro and is distinguished from the latter by its having, large poikilitic mats of pyroxene, virtually no sulphide and a well - defined preferred orientation.

Intrusive gabbronorite has sharp contacts against the metagabbro and is intruded by the ultramafic rocks and discordant pegmatites.

Where the rock contains olivine, the weathering surface is usually pitted. On fresh surfaces, it is dark grey to black, holocrystalline, medium - to coarse - grained (range  $< 0,5 - 10$  mm,  $av \sim 2$  mm) and has a well-developed preferred orientation defined by laths of plagioclase. Olivine, pyroxene, plagioclase and sulphide can be observed microscopically. The dark colour of the plagioclase is due to abundant fine inclusions which are probably magnetite because rock fragments appear to be more magnetic than amounts of opaque minerals estimated in thin

section suggest. Olivine-and quartz-bearing samples occur throughout the Complex.

The modes of gabbro-norite samples are given in Table ( 24 ). Chlorite, serpentine, actinolite and actinolitic-hornblende, epidote minerals, calcite, prehnite and cancrinite are secondary.

Analyses of plagioclase give values of between  $An_{64}$  and  $An_{77}$  (Table 43 ). The Michel-Lévy method gave an average value of about  $An_{60}$  ( $n = 34$ ) and slight normal zoning, ie  $An_{65-45}$  was observed. Plagioclase laths are clouded except for their rims, have an average length of about 2 mm, and are generally fresh, although some alteration to clay, sericite, cancrinite, calcite and epidote was noted. Plagioclase is a cumulus phase which occurs as inclusions in olivine and pyroxene, as laths having a sub-ophitic texture with respect to pyroxene and defining the preferred orientation in the rock. It shows deformation features, eg glide twins, strain shadows, bending of grains, micro-fractures and sub-grain development. Where plagioclase and olivine are in contact they always display a reaction relationship which gives rise to corona structures (Fig 33B). These are developed to varying degrees depending on the complete-

ness of the reaction. At their best developed they consist of, from the boundaries of olivine grains:-

1. A continuous rim of orthopyroxene made up of small prisms arranged at right angles to the olivine grain boundaries with a thin layer of magnetite developed either along its inner or outer circumference.
2. a discontinuous rim of clinopyroxene or pale green hornblende for hydrated samples, which at a discrete radius from the centre of olivine grains develops green spinel to form a symplectite which has scalloped contacts with plagioclase.
3. Tiny spinel needles occur in the plagioclase adjacent to the symplectite.

The coronas form at the expense of plagioclase which remains as irregular laths having scalloped grain boundaries.

Orthopyroxene occurs as subhedral to anhedral grains of hypersthene or bronzite (Table 43). It is generally unaltered but may have reaction rims of actinolitic hornblende and cummingtonite. Along shear zones in the rock the pyroxene may be completely altered to amphibole and be bent. Orthopyroxene commonly has exsolution lamellae of augite.



parallel to  $\{100\}$  and Schiller texture as a result of opaque mineral exsolution. Orthopyroxene is cumulus or intercumulus and poikilitic, with inclusions of olivine, and may form embayments into augite.

Augite (Table 43) ( $\text{av } C \wedge Y = 42^\circ$ ) is pale green, pink or neutral and occurs as an intercumulus phase which may show zoning and twinning. Blebs of

orthopyroxene and haematite occur parallel to  $\{100\}$  while in TV187 exsolution lamellae form "herringbone-structure". Schiller texture and alteration to actinolitic hornblende is better developed in augite than in orthopyroxene.

Olivine occurs as subhedral to rounded, cumulus grains up to 4 mm in diameter but average about 1.5 mm. Its composition varies only slightly:  $\text{Fo}_{80.0}$ - $\text{Fo}_{75.8}$  (Table 43). Ubiquitous cracks in olivine are lined with magnetite and calcite and form the locus for alteration to serpentine and green bowlingite. Olivine may be enclosed in orthopyroxene, embayed into by augite and have inclusions of biotite, chlorite and plagioclase. Where olivine is in contact with pyroxene, grain boundaries are sharp compared with olivine-plagioclase contacts where the reaction noted earlier has taken place.

Quartz occurs as tiny inclusions in uralitized pyroxene grains or as a rare intercumulus phase which may have strain shadows.

Brown mica is either an intercumulus phase or is epitaxially developed around pyroxene, having grown parallel to the  $\{ 100 \}$  cleavage. Analyses (Table 43 ) indicate a phlogopitic composition. Opaque exsolution occurs along  $\{ 001 \}$  and appears to become more abundant as the phlogopite becomes increasingly altered to chlorite.

Primary, brown hornblende may occur as a late-magmatic intercumulus mineral. Secondary hornblende is distinguished from the former by its blue-green colour, and by the fact that it occurs as an alteration product of pyroxene and along irregular cracks in plagioclase.

Magnetite, which is the main opaque mineral present, is found either as an intercumulus phase, as anhedral to euhedral disseminated grains, along cracks in olivine and as inclusions in other silicates, particularly plagioclase. Opaque minerals may be surrounded by phlogopite, and epidote minerals and may have inclusions of plagioclase and apatite. Opaque oxides with inclusions and exsolution lamellae of emerald-green pleonaste may be rimmed by hornblende-

spinel symplectites in contact with plagioclase (Fig 32E) and they may form a symplectic intergrowth with orthopyroxene or olivine (Fig 34I). Small, disseminated, anhedral grains of pyrrhotite and chalcopyrite may also occur.

The gabbronorite has an orthocumulus texture with plagioclase, olivine and occasionally orthopyroxene making up the cumulus phases (Fig 35A).

#### 3.3.4.5 Gabbronorite dykes

These intrude the metagabbro, the meta-gabbro dykes and the gabbronorite and may extend beyond the borders of the Complex. According to Beukes (1973, p 245) these dykes have a predominantly north-west strike.

The rock weathers dark brown to buff and may have glomerophenocrysts of plagioclase (av. dia. 2 mm) set in a fine-to medium-grained matrix. Along contacts plagioclase and augite have a strong preferred orientation which becomes less pronounced towards the centres of dykes.

These rocks are mineralogically almost identical (Table 25 ) to the gabbronorite and are probably a late stage component of the gabbronorite intrusive period.

### 3.3.4.6 The Tantalite Valley Satellite Body and Related Rocks

This plug of medium-to coarse-grained gabbro-norite is oval-shaped (700 m long, 30 m wide) (Fig 17). It has a narrow, west-north west trending, tail-like extension on the west end, made up of sheared hornblende; the body as a whole mimicking the trend and shape of the Tantalite Valley Complex. It is situated a few hundred metres south of the Witkop pegmatite (Map 1) and is surrounded by quartz-feldspar-muscovite gneisses which dip steeply to the north-east with lineations which plunge in the direction  $080^{\circ}$  at  $60^{\circ}$ . The gabbro-norite is concordantly intrusive into the gneisses with sharp contacts but without producing recognizable contact metamorphic effects.

The fresh rock is dark greenish-grey and speckled with interstitial patches of milky plagioclase, the latter feature serving to distinguish it from gabbro-norite and, where the rock is uralitized, from the metagabbro of the Complex.

Other small outcrops of this rock-type occur intruding the Grey Gneiss near the Tantalite Valley Scheelite Prospect (Map 4) and intruding Grey Gneiss on the east and west banks of the Girtus River near

the scheelite prospect there (Map 3). In the former locality it was also found as rare xenoliths in the ultramafic rocks. Although the rock is characteristically massive, isomodal layering, defined by varying proportions of plagioclase and mafic minerals was observed both in the Satellite body and in the Girtus bodies (Fig 23); the latter contained cross-bedding.

The modes of these rocks are given in Table ( 26 ).

Plagioclase, occasionally deformed, is intercumulus and occurs as anhedral crystals which are partly altered, especially in their cores, to sericite, cancrinite and epidote. It varies in composition from  $An_{20}$  to  $An_{58}$  and may show normal zoning.

The pyroxenes in AM123A are bronzite and diopside (Table 43 ). Bronzite occurs as large, subhedral, cumulus grains which are altered to actinolitic - and Mg-hornblende (Table 43 ), have inclusions of phlogopite and have clinopyroxene exsolution lamellae parallel to  $\{100\}$  . Diopside is pale green, anhedral and intercumulus. It shows twinning, exsolution lamellae of orthopyroxene parallel to  $\{100\}$  , inclusions of phlogopite and alteration to actinolite or actinolitic-hornblende (Table 43 ).

Phlogopite (Table 43 ) occurs intergrown with

hornblende, as oriented inclusions in pyroxene and as discrete, interstitial, disseminated laths. It may contain inclusions of zircon around which are pleochroic haloes.

Large, discrete, anhedral grains of hornblende may have inclusions of quartz and exsolved rutile and opaque minerals. Simple twinning and complex intergrowths with cummingtonite (Fig 35F) may be observed (Moore pers.com). Analyses of the coexisting amphiboles are given in Table 43. Hornblende also occurs along irregular fractures in the rock.

Anhedral quartz with inclusion trains and showing strain shadows occurs interstitially.

Opaque minerals occur as disseminated, anhedral crystals or as "dustings" and inclusions in amphibole and pyroxene.

The grain-size is seriate (range 3 mm - 0,05 mm, av. about 1 mm) and the texture of the fresh rock (Fig 35G) can be described as mesocumulus with orthopyroxene cumulus and clinopyroxene and plagioclase intercumulus minerals.

#### 3.3.4.7 The Mafic-Ultramafic Rocks including those found in the Grey Gneiss Sequence

The metagabbro and gabbro-norite are intruded by mafic-ultramafic rocks. Three large bodies have been identified, although several minor plugs and lenses occur. The largest body, 0,58 km<sup>2</sup> in area, referred to here as the Main Plug, is situated just south-west of the Signaalberg trigonometrical beacon and was discovered by Wiid in 1943. It has been referred to as the 'Eastern Plug' by Moore (1975). The entire body is surrounded by metagabbro and contains xenoliths of the latter along contacts which are sharp and lack evidence of metasomatism or metamorphism. The Main Plug has been extensively drilled by Rio Tinto and Tantalite Valley Minerals (Pty, Ltd) but as yet copper-nickel mineralization has been shown to be sub-economic. Sections across this body, as derived from drill-core, are given in Fig (4). Some layering defined by varying proportions of olivine and plagioclase is apparently present, a conclusion which is supported by rare outcrops in the field. Layers vary in thickness from a few feet to several hundred feet and are usually horizontal or have a gentle dip. The rock is commonly altered down to depths of around 30 feet from the surface and along shear zones and adjacent to pegmatites. The increasing depth at which country rock is intersected in boreholes in the

direction from west to east indicates that the plug is probably funnel-shaped. Small bands of metagabbro and gabbro-norite in the core-sections are interpreted as xenoliths. Pyrrhotite and chalcopyrite may fleck the surface of the rock or occur along fractures.

The second largest plug, situated in the extreme north-east corner of the Complex was called the 'north-east' plug by Moore (1975) and was not included in Käch's map of the Complex (Von Backström 1976). No borehole data are available from this body which on the surface occurs as a massive amphibole-chlorite rock which was probably a norite or pyroxenite before alteration. Moore (1975) attributed the high degree of alteration to late-stage shearing which is particularly severe in the area.

The third large ultramafic body is found near the south-west border of the Complex and was termed the 'southern plug' by Moore (1975). It is roughly elliptical with a long axis of about 0,75 km parallel with the trend of the Complex. Xenoliths of metagabbro, up to 2 m in diameter occur near its contact with metagabbro. This plug, of which a borehole section is included in Fig (4), can more correctly be described as a narrow dyke-like body. It



consists mainly of troctolite with subsidiary pyroxenite although on the surface it is altered to an amphibole-chlorite rock.

A small roughly circular plug occurs in the centre of the Complex near the northern contact with the gneisses and is referred to here as the northern plug. Its surface expression is as a reddish-brown weathering, highly altered dark-green amphibole-chlorite rock which is often highly sheared. Contacts with the enclosing gabbro-norite are sharp and lack evidence of metamorphism and metasomatism. Fresh samples of this rock have not been found and no borehole data are available.

A narrow, somewhat disjointed dyke a kilometer in length made up of reddish-brown weathering amphibole-chlorite rock is found on the northern slope of a hill of metagabbro to the south of the centre of the Complex (Map 1). It is arbitrarily called a metapyroxenite as no fresh samples are available.

Metagabbro when intensely sheared may occur as a reddish-brown weathering amphibole-chlorite rock which may closely resemble the ultramafic rocks in outcrop. They are distinguished from the latter by the way that they grade gradually into "normal" metagabbro.

Several bodies of ultramafic rock occur outside the Complex. Beukes (1973) found ultramafic rocks on both sides of the Houms River on the farm Sandfontein 131 (Fig 3). On the western bank there is a north-west trending elongated body of troctolite (eg sample GBW 421) and peridotite (eg sample GBW 380) which intrudes Grey Gneiss. The occurrence on the eastern bank was found in the present investigation to be more closely akin to the Tantalite Valley gabbro-norite rather than to an ultramafic rock. Altered ultramafic rocks, originally probably troctolite and olivine-pyroxenite, intrude the Grey Gneisses along the axis of the Tantalite Valley Lineament to the south of the Complex (Maps 1, 3 and 4) and are here believed to be equivalent to the ultramafic rocks found in the latter. Although the ultramafic rocks were emplaced before major shearing had occurred, they are massive except for a foliation along their margins. Important exceptions are the outcrops at W-X-Y (Map 3) which are schistose.

The ultramafic rocks weather reddish-brown to brown which serves to distinguish them from the gabbroic rocks (Fig 9). They weather more easily than the latter and tend to form rounded valleys and low-lying hills. The rock is generally massive but some layering can

be observed in rare outcrops. Where the rock is sheared there is often a widely spaced cleavage developed. There is a wide range in rock composition from mica peridotite, to troctolite, to norite, to olivine pyroxenite to pyroxenite or altered equivalents of these but the dominant rock-types found are troctolite and norite. The peridotites and troctolites are still recognizable as such on surface in contrast with pyroxenitic rocks which are altered to amphibole-chlorite assemblages. Borehole samples show that the amount of fresh pyroxene increases with depth. Olivine-rich varieties are strongly magnetic. Unpublished geochemical sampling maps by Rio Tinto show that Cu-Ni anomalies in the Complex occur around the main ultramafic plugs and have configurations which follow the mapped contacts of these bodies indicating that the ultramafic rocks can be distinguished from the gabbroids purely on their Ni and Cu values. Sulphides are rarely visible in surface samples of ultramafic rock but malachite staining may occur in some outcrops.

For convenience the whole range of ultramafic rocks have been divided rather arbitrarily into two groups in order to simplify the description: Group 1 comprises mafic-ultramafic rocks with  $\geq$  30%

olivine and group 2 comprises mafic-ultramafic rocks with  $< 30\%$  olivine.

3.3.4.7.1 Group 1-Troctolites, picrites, mica-peridotites

These rocks are characterized by dark-brown to reddish-brown pitted weathering surfaces which may be covered with layers of calcrete. On fresh surfaces they are blue-green to dark green, holocrystalline, medium - to coarse - grained rocks which may be transected by veins of chrysotile.

Milky plagioclase occurs between the interstices of rounded, serpentized olivine grains in troctolite and becomes gradually less as the rock becomes more peridotitic until eventually the rock is largely composed of greasy black serpentized olivine grains mottled with bronze-coloured phlogopite.

Modes of these rocks are presented in Table ( 27 ). Serpentine, calcite, chlorite, sericite, haematite, magnetite, hornblende and cancrinite are secondary.

Olivine occurs as rounded to subhedral cumulus grains (av. dia. 1,5 mm, range 0,05 - 5 mm) and has a composition of about  $Fe_{82}$  (Table 43 ). It is commonly altered along cracks to serpentine, calcite and magnetite, leaving fresh cores. Expansion cracks occur in the surrounding plagioclase as a

result of the increase in volume on conversion of olivine to serpentine. Reaction rims are invariably developed along olivine/plagioclase contacts. These vary in thickness for different samples depending on the completeness of the reaction. The following corona sequences were observed:-

- 1) olivine - cummingtonite - plagioclase
- 2) olivine - cummingtonite - green clinoamphibole/  
spinel symplectite-plagioclase (Fig 33F)
- 3) olivine - orthopyroxene partly altered to  
cummingtonite - clinoamphibole/spinel symplec-  
tite - plagioclase (Fig 32G).

An analysis of the clinoamphibole/spinel symplectite zone in troctolite sample TV49 is given in (Table 43 ). Comparing this with the analysis given by Sapountzis (1975) it probably represents Tshermakitic hornblende intergrown with spinel. Microprobe scans from olivine into plagioclase show that Mg and Fe remain approximately constant and that Al increases abruptly in a series of steps until plagioclase is reached while Si behaves in just the opposite way to Al in the reaction zone but increases once it reaches plagioclase. The plagioclase is shown to

be unzoned.

Plagioclase is a highly sericitized, intercumulus phase and this distinguishes the ultramafic rocks from the gabbro-norites where plagioclase is cumulus. A value of  $An_{77.9}$  was obtained from troctolite sample TV49 using the electron microprobe (Table 43).

The conventional Michel-Lévy method was difficult to apply due to the extreme sericitization of the plagioclase but an average value of about  $An_{60}$  was obtained for several specimens. Other alteration products are cancrinite, calcite and chlorite.

Orthopyroxene was analysed as bronzite in TV49 (Table 43). It is intercumulus in troctolite but cumulus in peridotite where it occurs as anhedral mats up to 3 mm in diameter.

Alteration products are cummingtonite, actinolitic-hornblende and chlorite and blebs and lamellae of diopside occur parallel to  $\{100\}$  which may also be lined by calcite.

Clinopyroxene ( $C \wedge \gamma = 40^\circ$ ) is greenish and slightly pleochroic and occurs as an intercumulus phase in peridotite but not in troctolite. It is diopside or diopsidic augite (Table 43) and occurs as anhedral mats up to 5 mm in diameter which may be altered to actinolitic hornblende, chlorite and

calcite and is usually dusty in appearance due to opaque exsolution. Very fine exsolution lamellae of orthopyroxene are common parallel to  $\{100\}$  and serpentine may occur along cracks.

Intercumulus phlogopite (Table 43) is altered along grain boundaries to chlorite and may have haematite and magnetite along  $\{001\}$ . It commonly occurs around or included in magnetite.

Magnetite and haematite are secondary, the former occurring as anhedral to euhedral disseminated grains, along cracks in olivine and along fractures.

Ferritchromite (Table 43) is found as inclusions in spinel and as discrete, anhedral grains which formed probably at the expense of cumulus chromite. Probe scans show that Fe and Cr vary irregularly over the surface of a grain, while Ti remains approximately constant.

Opaque mineral/plagioclase contacts may show green amphibole/spinel symplectites.

Pyrrhotite is interstitial and anhedral and is associated with granular pentlandite.

Green spinel-hercynite (Table 43) with patches of brown picotite (Table 43) occur as anhedral disseminated grains (Fig 33E) or associated with opaque minerals. Picotite may contain rare, rounded.

inclusions of olivine.

Apatite is subhedral to anhedral, lilac coloured and occurs included in primary brown amphibole. The latter is an intercumulus phase which is often associated with opaque minerals and may be altered to chlorite.

Secondary blue-green clinoamphibole occurs as scattered laths and as an alteration product of pyroxene. It is commonly twinned, associated with chlorite and may enclose plagioclase.

Chlorite occurs as green to turquoise-green (Cr-rich?) decussate laths or around phlogopite as an alteration product.

Texturally these rocks can be described as olivine <sup>+</sup> bronzite + chromite mesocumulates. (Fig 36D).

The average grain size is 1,5 - 2 mm. The probable sequence of crystallization as deduced from textural relationships is olivine → chromite → bronzite → clinopyroxene → plagioclase → phlogopite.

#### 3.3.4.7.2 Group 2 - Norite, olivine - pyroxenite and altered equivalents

Hand specimens range from olive-green to fleshy-brown with the grain size varying from coarse, for fresh specimens, to fine-grained for



highly altered varieties. Samples are weakly magnetic and may show layering, caused by variations in the amount of plagioclase present. Fresh samples are virtually unknown on surface but occur abundantly in drill-core. The dominant rock-type on surface consists of chlorite and calcic-amphibole which may be partly due to weathering in addition to alteration.

Modes for these rocks are given in Table ( 28 ).

Orthopyroxene is bronzitic (Table 43) and occurs as subhedral to euhedral grains (av length about 3,5 mm, range 2 - 7 mm) which commonly have blebs and lamellae of endiopside or diopside (Table 43 ) along  $\{100\}$  . Olivine occurs as inclusions in orthopyroxene which may show alteration to cummingtonite, actinolitic-hornblende, tremolite and chlorite and be overgrown by plagioclase and phlogopite.

Pale-green clinopyroxene (  $C\wedge\gamma$  39-45° ) is diopside or augite (Table 43 ) and occurs as amoeboid to anhedral, intercumulus grains which commonly show a reaction relationship with orthopyroxene and have fine bronzite (Table 43 ) exsolution lamellae parallel to  $\{100\}$  . It is commonly twinned and dusty in appearance due to the exsolution of opaque minerals and may be altered to

actinolitic-hornblende, chlorite and calcite.

Intercumulus plagioclase was analysed as  $An_{69,95}$  in norite specimen TV6 129, which falls in the range of  $An_{51}$  to  $An_{73}$  obtained optically for several samples. The latter values were obtained with some difficulty due to the often extreme alteration of plagioclase to sericite, epidote and calcite. Expansion cracks occur adjacent to olivine undergoing serpentinization. In fresh samples plagioclase is clouded with opaque minerals, grain margins excepted (Fig 32F).

Intercumulus phlogopite (Table 43) may have inclusions of apatite and show exsolution of rutile in the form of sagenitic webs (Fig 33D). It occurs growing epitaxially around pyroxene and as discrete flakes up to 1,5 mm long and may be altered along its borders to bright green, Cr-rich? chlorite. Prehnite may occur along  $\{001\}$ .

Olivine is found as discrete, rounded, slightly deformed cumulus grains up to 2 mm in diameter or as inclusions in orthopyroxene. It is altered to serpentine, magnetite or green bowlingite and may have oriented magnetite inclusions. Where in contact with plagioclase coronas as described in section (3.3.4.7.1) are developed.

Apatite (pinkish brown, pink, neutral) occurs as euhedral to subhedral, disseminated grains which may be clouded and occur as inclusions in opaque minerals.

Magnetite, altered to haematite and limonite, occurs along cracks in olivine and as discrete, rounded grains with inclusions of phlogopite and apatite. Opaque minerals are found along cleavages in clinopyroxene, clinoamphibole and chlorite. Some dark brown, translucent, cumulus chromite was also observed. Up to 2 volume % chalcopyrite and pyrrhotite may occur in large anhedral crystals varying from 2 to 5 mm in diameter.

Chlorite (pale green, turquoise green) (av. length 0,1 mm, range 0,7 - < 0,05 mm) occurs in nearly all samples as an alteration product of pyroxene and phlogopite. It is found as mats made up of decussate laths which are occasionally twinned, bent and show undulose extinction.

Laths of actinolitic-hornblende, tremolite (Table 43) and blue-green hornblende (av. length 0,2 mm) are the main alteration products of pyroxene. In sample TV8 173 (Table 28) brown hornblende grains (primary?) are included in orthopyroxene. Laths are commonly twinned and show a good basal parting.

Cummingtonite (neutral, pale green, brown) occurs as decussate, acicular to lath-like crystals (av. grain size 1,5 mm, range 5 - 0,05 mm) which are usually curved, have lamellar twinning and have abundant opaque minerals parallel to cleavage. The grain-size of the rock averages about 0,15 mm for altered specimens and 3,5 mm for fresh specimens. Texturally the fresh rock can be termed a bronzite  $\pm$  olivine  $\pm$  chromite mesocumulate (Fig 36C). The probable sequence of crystallization as deduced from textural relationships is olivine  $\rightarrow$  spinel  $\rightarrow$  orthopyroxene  $\rightarrow$  plagioclase  $\rightarrow$  clinopyroxene  $\rightarrow$  phlogopite.

### 3.3.5 Syn-tectonic intrusives of the Kumkum Massif

#### 3.3.5.1 Kumkum gabbro-norite

This rock-type occurs as sills, plugs, sheets and less commonly dykes intruding in equigranular granite gneiss, Grey Gneiss and Black Gneiss. It is found over large areas of the Kumkum Massif as conical hills whose slopes are covered by black scree boulders as typified by the Kumkum body situated on the farm Kumkum 105 (Fig 13). Also included in this group is the prominent

hill which lies north-east of Warmbad, called the Einsiedler body by Beukes (1973) and a prominent mafic body which lies 8 km north of the Tantalite Valley Complex and 5 km east of the Houms River (represented by sample AM242). The association of Kumkum gabbronorite and granite gneiss is conspicuous throughout the Kumkum Massif and further afield. Contacts between sills and dykes of Kumkum gabbronorite with granite gneiss are sharp whereas large plugs commonly have a contaminated zone around them. Dykes of gabbronorite intrude granite gneiss parallel to its schistosity with the grain-size of the former decreasing as the contact with granite gneiss is approached although shearing and alteration also increases. Evidence of partial melting of the surrounding rock can occasionally be observed. The main Kumkum gabbronorite body where in contact with Grey Gneiss has produced a rheomorphic, quartzo-feldspathic reaction zone in addition to Black Gneiss hornfelsic rocks (Map 5). Xenoliths of granite gneiss in the gabbronorite are rare, but near K65 (Map 6) a body of about 4 m in diameter was observed in the latter. More feldspathic and finer-grained varieties of gabbronorite occur, the latter apparently intruding the coarser gabbronorite. The gabbronorite is itself intruded by aplites, quartz veins and dis-

cordant pegmatites, the latter causing some amphibolitization of gabbro-norite along contacts.

The rock has a dark brown, often spongy and pitted weathering crust and is dark grey to dark greenish grey on fractured surfaces.

It is generally medium-to coarse-grained and equigranular except for the presence of rare plagioclase phenocrysts. Laths of plagioclase commonly have a preferred orientation. The minerals visible in hand specimen, apart from plagioclase, are biotite, olivine, hornblende and sulphide including pyrite and pyrrhotite. Modes for this rock type are given in Table (29).

Plagioclase ranges in composition from  $An_{20}$  to  $An_{77}$  (av.  $An_{55}$ ,  $n = 21$ ) and occurs as rare phenocrysts, which commonly have inclusions of pyroxene arranged randomly or parallel to zoning and in the groundmass as cumulus laths. The phenocrysts are generally more calcic than the matrix plagioclase; with values of  $An_{77}$  and  $An_{72}$  respectively having been obtained from K37A (Table 43). The plagioclase is usually clouded with tiny opaque mineral and silicate (hornblende?) inclusions but possesses clear rims. Normal zoning and sericitization are commonly observed. Coronas are developed where olivine and plagioclase grains are in contact.

The zone surrounding olivine is made up of bronzite (Table 43 ) or cummingtonite followed by a discontinuous zone, in contact with plagioclase, of olive-green to brown Fe-pargasite (Table 43 ) - spinel symplectite (Fig 33C). Where this process has gone to completion, olivine has been completely pseudomorphed. Plagioclase shows albite, pericline, baveno, carlsbad and carlsbad-albite twinning.

Pyroxenes occur as anhedral, intercumulus crystals or optically continuous mats which are embayed by laths of plagioclase. They are altered to olive-green hornblende and biotite along grain boundaries, cleavages and fractures and have inclusions of olivine, plagioclase and opaque minerals, the latter giving the pyroxene a dusty appearance. Exsolution lamellae parallel to  $\{100\}$  are common in both ortho- and clino-pyroxene except that that found in the former is usually coarser. The orthopyroxene in K37A is bronzite (Table 43 ). The clinopyroxene is characteristically pale green and is either diopside as in the case of K37A (Table 43 ) or diopsidic augite.

Euhedral to anhedral olivine occurs as unzoned, cumulus phenocrysts which in K37A have a composition

of Fo<sub>66</sub> (Table 43 ) with grains surrounded by coronas having the same composition as those included in pyroxene. Inclusions of plagioclase, which have generally been replaced by corona minerals, and rounded, brown and green spinel grains occur.

Primary, brown Fe-pargasite (Table 43 ) occurs in scattered, anhedral patches enclosing flakes of biotite, opaque minerals and green spinel. Olive-green hornblende occurs as an alteration product of pyroxene, in olivine/plagioclase coronas and along cracks in plagioclase.

Biotite (Table 43 ) is commonly associated with opaque minerals and green spinel and is generally enclosed by brown hornblende. Flakes may show strain shadows, bending, inclusions of apatite and in rare cases zircon and an intercumulus relationship with respect to plagioclase. Biotite may occur along cleavages in pyroxene and grown epitaxially around the grain boundaries of the latter. Quartz is rare and occurs as interstitial, amoeboid grains which may show strain shadows.

Emerald-green hercynite (Table 43 ) occurs as discrete, anhedral grains associated with opaque minerals, as rare coarse exsolution lamellae in



opaque minerals and as vermiiform inclusions in magmatic hornblende. Olive-brown to light brown picotite is rare and occurs as rounded, disseminated grains and as inclusions in olivine and in pyroxene (Beukes 1973). Magnetite and ilmenite occur as anhedral, disseminated grains which are not commonly found in association although Beukes (1973) noted that ilmenite exsolution is common in magnetite. Opaque minerals have inclusions of plagioclase, biotite and in rare cases pyroxene and are generally intercumulus apart from chromite which may occur as cumulus, rounded grains. Magnetite may be concentrated in areas formerly occupied by olivine which have since been replaced. Haematite and limonite staining occur as alteration products of magnetite. Symplectic intergrowths of opaque mineral and orthopyroxene are only found in one sample (K250) unlike in the case of the Tantalite Valley gabbro-norite where this texture is common. Schiller texture and opaque mineral exsolution parallel to cleavage in pyroxenes is common. Beukes (1973) observed pyrrhotite, ilmenite, pentlandite, chalcopyrite, pyrite, magnetite and marcasite (secondary?) in these rocks. He found that pyrrhotite, chalcopyrite, pentlandite and occasionally pyrite are intimately associated

and form irregular aggregates smaller than 1 mm in diameter. Pyrite also occurs as euhedral, disseminated grains.

Apatite occurs as disseminated colourless or pink needles up to 0,4 mm long, with a length to width ratio of 13,33 indicating a high-temperature magmatic origin (Wyllie et al 1962). It occurs as inclusions in biotite, pyroxene and opaque minerals.

The average grain size of the gabbro-norite, based on plagioclase laths, is 0,5 mm to 1,0 mm. Plagioclase and olivine phenocrysts may reach up to 2 mm and 1,5 mm in diameter respectively. Plagioclase laths in gabbro-norite dyke sample K44 decreases in grain size from an average of 1 mm in the centre of the dyke to about 0,2 mm where the latter is in contact with Inequigranular granite gneiss. Texturally the rock can be termed a plagioclase-olivine orthocumulate with plagioclase commonly having a well developed igneous lamination (Fig 35B). Toogood (1976) interpreted the latter as foliation and banding and used it as the main criterion for distinguishing between this rock-type which he called the Kumkum granolite and the Stolzenfels norite.

### 3.3.5.2 Contaminated gabbronorite

This, the commonest gabbroid found in the Kumkum area, weathers to form low-lying hills. It occurs as plugs, dykes and sills in granite gneiss and along contacts of the Kumkum gabbronorite with the granite gneiss. The Contaminated gabbronorite is regarded here as a textural and mineralogical variety of the Kumkum gabbronorite because the latter is seen to grade into the former as the contact with granite gneiss is approached. Toogood (1976) also noted that Stolzenfels norite (Contaminated gabbro-norite) and Kumkum granulite (Kumkum gabbronorite) grade into each other and distinguished between the two on the basis of their texture; the Stolzenfels norite being generally massive and the Kumkum granulite being 'foliated'. Texture is indeed one of the best criteria for distinguishing these two rock types, followed by mineralogy. The texture of Kumkum gabbronorite is however here considered as being igneous in contrast to the metamorphic interpretation given to it by Toogood (1976). This texture becomes increasingly "metamorphic" as the rock grades into Contaminated gabbronorite. Secondary hornblende, biotite, quartz, apatite and zircon are more abundant while olivine and primary hornblende are absent in Contaminated gabbronorite.

compared with Kumkum gabbronorite. The contaminated rock may have megacrysts of quartz, plagioclase and K-feldspar giving it an appearance similar to the mottled metagabbro of Tantalite Valley. As the contact with gabbro is approached megacrysts in granite gneiss become finer grained and the rock becomes darker in colour in response to increasing amounts of almandine-garnet and biotite being present. Contact zones are commonly sheared and retrograded, thus obscuring the original relationships. Xenoliths of Contaminated gabbronorite up to 0,75 m in diameter, surrounded by leucocratic haloes in granite gneiss and Contaminated gabbronorite agmatite, composed of angular fragments of gabbroid surrounded by granite gneiss, are taken as being evidence of local remelting of the granite gneiss. The rock is medium- to coarse-grained and equigranular except where megacrysts of plagioclase, quartz and in rare cases hornblende and K-feldspar occur. The rock has a dark-brown to buff, slightly pitted weathering crust. Mafic minerals weather reddish-brown and limonite and haematite staining are found along irregular cracks. Fresh surfaces vary in colour from brown to dark grey-green to black with plagioclase, quartz, K-feldspar hornblende, pyroxene,

biotite and sulphide being macroscopically visible. The rock is usually massive but samples exist where the plagioclase defines a preferred orientation.

Modes of samples of this rock-type are given in Table (30).

Plagioclase is found in the groundmass as laths and polyhedra and as megacrysts, commonly containing inclusions of pyroxene and quartz. The groundmass and megacrystic plagioclase in K45A have quite similar compositions ie  $An_{49}$  and  $An_{47,5}$  respectively (Table 43). The range of groundmass plagioclase compositions determined optically is  $An_{23}$  to  $An_{65}$  (av.  $An_{56}$ ,  $n = 30$ ). Grains may show zoning, bending, strain shadows and alteration to sericite. Myrmekite is commonly found along the contacts of quartz and plagioclase grains. Carlsbad, carlsbad-albite, pericline and glide-twinning are observed.

Clinopyroxene is anhedral or prismatic and poikilitic with inclusions of plagioclase, opaque minerals, hornblende and biotite. It is pale green and is altered to olive-green hornblende and has biotite along grain boundaries and cleavage. Clinopyroxene in K45A was found to be augite (Table 43 ). It is occasionally twinned, shows Schiller texture, bending of grains, limonite staining along cracks

and commonly has opaque minerals and exsolution lamellae of orthopyroxene parallel to  $\{100\}$ .

Orthopyroxene occurs as disseminated, anhedral or occasionally polygonal grains which have a faint pleochroism and in K45A is found to be hypersthene (Table 43). It is altered to olive-green hornblende and biotite is found along cleavages and grown epitaxially around grain boundaries. Blebs and exsolution lamellae of clinopyroxene and opaque minerals occur parallel to  $\{100\}$ . Inclusions of plagioclase, bent grains, Schiller texture and limonite staining along cracks may be observed.

Hornblende is secondary and occurs as an alteration product of pyroxene. It is often poikilitic with inclusions of plagioclase, biotite, opaque minerals, pyroxene and quartz, the latter occurring parallel to the cleavage in sample K41. It may be twinned and zoned, with the cores of grains appearing paler green than their rims. In K52 hornblende occurs as large poikilitic mats up to 0,5 cm in diameter.

Biotite is found as discrete, disseminated flakes or grown epitaxially around hornblende and pyroxene or along cracks and cleavage in pyroxene, hornblende and in rare cases, plagioclase. It is commonly poikilitic with inclusions of apatite, plagioclase, orthopyroxene, opaque minerals, hornblende, and

zircon, the latter producing prominent pleochroic haloes. Biotite is often associated with opaque minerals and contains inclusions of the latter along its grain boundaries. Grains may be bent and have prehnite, parallel to  $\{001\}$ . In K46 the biotite wraps around K-feldspar megacrysts but away from these it assumes a random disposition.

Quartz is anhedral, amoeboid or polygonal and is either interstitial or occurs as inclusions in hornblende and plagioclase or as discrete megacrysts up to 1 mm in diameter.

Sericitized orthoclase with inclusions of clinopyroxene, plagioclase and quartz occurs as megacrysts surrounded by a discontinuous zone, depleted in biotite, containing perthite, plagioclase, quartz and magnetite set in a gabbroic matrix.

Magnetite and ilmenite, which are rarely associated, occur as anhedral, disseminated grains which have inclusions of biotite and limonite and haematite alteration.

Apatite occurs as colourless hexagonal crystals or needles up to 0,45 mm long, having an average length to thickness ratio of 11,40, which according to Wyllie et al (1962) is typical of rocks crystallized from a high temperature magmatic state. Apatite is

found included in biotite and orthopyroxene as well as cross-cutting the latter.

The groundmass has a grain size ranging from 0,04 mm to 2 mm, with an average of 0,7 mm. K-feldspar megacrysts attain a length of up to 5 cm while quartz and plagioclase megacrysts average about 0,5 cm in diameter.

The rock probably originally had an ophitic texture which has since been modified to equigranular to inequigranular interlobate or polygonal. (Fig 36A).

#### 3.3.5.3 The Eselruh gabbro

The Eselruh gabbro is a rare gabbroic variety found in the Kumkum Massif. Most of the samples were collected near the farm Eselruh but isolated outcrops occur to the east. It crops out as small, scattered plugs and sills which intrude Black Gneiss and Grey Gneiss conformably along schistosity. The margins and some internal zones of these bodies may be sheared and amphibolitized. They are intruded by aplite, pegmatites and quartz veins.

The rock has a dark brown, pitted weathering surface, although orthopyroxene may weather positively and give the surface a warty appearance. It is



massive, medium-to coarse-grained and equigranular except for rare plagioclase phenocrysts. Fresh surfaces are dark grey to dark green and have flecks of milky plagioclase and phlogopite.

Modes of these rocks are given in Table (31).

Intercumulus plagioclase is generally highly sericitized and clouded by tiny green silicate inclusions which may be hornblende or spinel. It ranges in composition from  $An_{50}$  to  $An_{88}$  with an average of  $An_{70}$  for six samples. Zoning may occur. Where in contact with olivine typical coronas consisting of bronzite or cummingtonite around olivine followed by a discontinuous zone of pargasite-spinel symplectite in contact with plagioclase occur.

The pyroxenes are altered to olive-green hornblende and are characterised by extreme opaque exsolution which gives them a cloudy appearance. They are cumulus, euhedral and have rounded inclusions of olivine. Clinopyroxene shows twinning, zoning and hourglass structure defined by opaque exsolution. (Fig 35C and D). In K4 the pyroxenes were found to be diopside (Table 43 ) and bronzite (Table 43 ). Ortho- and clino-pyroxene both have exsolution lamellae of clinopyroxene and orthopyroxene respectively parallel to  $\{100\}$  .

Hornblende is of two types; yellow-brown, primary pargasitic hornblende (Table 43 ) and secondary blue-green hornblende occurring as an alteration product of pyroxene.

Cumulus olivine was originally more plentiful in the rock and has often been completely pseudomorphed by corona minerals (Fig 35C) or remains as small anhedral relics in the centre of coronas. Tiny, rounded grains are included in pyroxene where they have been armoured from reacting with plagioclase. These grains were analysed as Fo<sub>72</sub> in sample K4. (Table 43 ).

Phlogopite (Table 43 ) occurs as anhedral, disseminated laths which may be surrounded by olive-green hornblende.

Anhedral emerald-green hercynite and brown picotite occur associated with the opaque minerals which include magnetite, ilmenite, chromite and sulphide.

The average grain size of the rock varies from 1 mm to 3 mm for different samples with a range of 0,1 mm to 5 mm. The rocks are meso- or adcumulates with orthopyroxene, clinopyroxene, minor olivine and minor chromite being the main cumulus minerals. Plagioclase and phlogopite are intercumulus (Fig 35C).

### 3.3.6 Post-tectonic Group

#### 3.3.6.1 Young Pegmatite and Quartz Veins

Young pegmatites are distinguished from the old pegmatites in that the former are zoned, coarse-grained, garnet bearing, have economic concentrations of Rare Earth Elements, are relatively undeformed, are commonly sheet-like and discordant and are usually much larger (up to 400 m in length) than the Old pegmatite. Those bodies intruding the Complex or occurring adjacent to the latter in the surrounding gneisses or mottled metagabbro are by far the most important economically. The best known of these called White City, Witkop, Homestead, and Lepidolite (Lithium) (Map 1) continue to be mined and have yielded large quantities of beryl, microlite, tantalite-columbite, bismuth and lithium minerals in the past. The geology of these and other bodies in the area have been studied in detail (Cameron 1955; Beukes 1973, Von Backström 1976) and the interested reader is referred to these authors for further information.

Young pegmatites found about 2 km south-east of the Signaalberg Trig. Beacon (Map 1) are noteworthy in that they are scheelite-bearing and have in part

been responsible for the concentration of this mineral in altered ultramafic rocks which they intrude as flat-dipping sheets (Fig 31). The relationships of the young pegmatite to the ultramafic rock is shown in Map ( 4 ).

Moore (1975), after noting the distribution of the economically important pegmatites around the Tantalite Valley Complex, that is to the east, west, north and south of the latter, suggested that these owed their origin to the migration of acid melts to low-pressure areas around a rigid body acted upon by 'a stress directed from the north-north-east and south-south-west and with a component of shear which gave right lateral displacement.' He postulated that these pegmatites formed as a result of partial melting induced in pre-tectonic gneisses by the introduction of water into the latter from the shear zone, using a mechanism similar to that proposed by Beach (1974). The pegmatites then intruded zones of weakness in the country rock and began to crystallize from the walls inward, producing the characteristic zoning observed (Von Backström 1976).

Young pegmatites are also common in the Kumkum Massif where their outcrops provide a striking

contrast against the dark weathering rocks.

In both the Tantalite Valley and Kumkum areas, this is the youngest intrusive found apart from rare quartz veins containing epidote and albite which occur along mylonite zones and fractures

### 3.3.6.2 Metasomatic hornblendite

This rock type has thus far only been found in a small area to the south-east of the Tantalite Valley Complex. It occurs in zones up to several centimeters wide along the contacts between young andradite-bearing pegmatites, which occur as transgressive sheets (Fig 31) or dyke-like bodies and amphibolitized olivine-picrites and norites found in the Grey Gneiss Sequence. Metasomatic hornblendite also occurs as pods along shear zones in the ultramafic rocks where pegmatitic rocks are apparently absent and near the crest zone of a domed pegmatite sheet (Map 4).

The rock is characteristically massive and coarse-grained with phlogopite (Table 43) and decussate laths of blue-green Mg - hornblende (Table 43) up to 2,5 mm in length, being the main minerals present (Fig 36H). Modes of this rock are given in Table ( 32 ). Anhedra greenish-grey scheelite

granules can often be discerned in hand specimen but mineralization is unevenly disseminated through alteration zones. The mineralization has probably been introduced by milky, plagioclase-rich, scheelite-bearing, veins which form off-shoots from the young pegmatite. Biotitite may also occur in alteration zones and is included here as a mineralogical variety of the Metasomatic hornblendite. The former commonly contains less visible scheelite than the Metasomatic hornblendite proper.

The area in question contains four known scheelite deposits, labelled S, T, U and W-X-Y on Map 3, one of which (deposit S) was mapped in detail by pace and compass by Dr A.C. Moore and the author (Map 4). Mafic rocks in other areas (labelled Z on Map 3) were also examined for possible tungsten mineralization. One of the deposits, deposit S, has been successfully exploited in the past by Mr P Weidner Snr. It was concluded that the association young-pegmatite/Metasomatic hornblendite/amphibolitized ultramafic rock is critical to the development of scheelite mineralization which does not occur to any great extent where these conditions are not met. The metasomatic process was probably important in concentrating scheelite, as this mineral occurs dominantly in rock where this

process has been most efficient, ie in Metasomatic  
hornblendite proper.

## CHAPTER IV

## TEXTURES

"You know my method. It is  
founded on the observance of  
trifles."

The Boscombe Valley Mystery - Sir Arthur Conan-Doyle

#### 4.1 Olivine-plagioclase Coronas

Synantectic corona structures (as defined by Spry 1969 p 104) are commonly found where olivine and plagioclase are in contact in mafic and ultramafic rocks from the study area. These coronas all bear a close similarity to one another with respect to constituent minerals, and order of development. They generally consist of olivine surrounded by successive rims of bronzite or cummingtonite, with or without the development of a layer of magnetite along either its inner or outer (Fig 33B) circumference. There may be a narrow, discontinuous clear zone beyond this (Fig 33F). The clear zone is as yet unidentified but has the optical properties of clinopyroxene. Then follows a rim of green amphibole which abruptly develops fine blebs of spinel ( $\sim 10$  volume %), which also occurs in the adjacent plagioclase (Fig 33B). Each of the zones has essentially constant width, with the symplectite



zone being the widest (Fig 33F ). They follow the outline of olivine grains continuously except where these are in contact with minerals other than plagioclase, where no corona-minerals are developed. The bronzite in the coronas occurs as tiny parallel prisms perpendicular to the olivine interface and may be altered to cummingtonite as in the troctolites of Tantalite Valley. The amphibole-spinel symplectite layer consists of oriented amphibole rods, which fan out from discrete centres located on the contact with the previous layer. The contact of the amphibole layer with plagioclase is characteristically scalloped. In some of the Kumkum and Eselruh gabbro-norites, olivine may be entirely replaced by anhedral coronas consisting of a core of pale pink bronzite, surrounded by olive-green hornblende plus spinel symplectites (Fig 33C). The development of coronas in most samples has not altered the cumulus texture of the rock but where the former structures are abundant, as in the Eselruh gabbro-norite, the rock can be termed a coronite (Spry 1969). The coronas found in the present area can be classed as complex and partly hydrous (Spry 1969). None contain garnet. The absolute range in olivine and plagioclase compositions in rocks found to have coronas are  $\text{Fo}_{66}$  to  $\text{Fo}_{83}$ , and  $\text{An}_{64}$  to  $\text{An}_{85}$ . It is found that

the olivine surrounded by coronas is identical in composition to unreacted grains found as inclusions in minerals from the same rock (Table 43 , compare analyses of olivines from TV263 and K37A.)

Orthopyroxene in the corona is generally bronzite and differs from that in the rest of the rock by having a higher Mg/Mg + Fe ratio, and a lower  $\text{TiO}_2$  content (Table 43 , compare analyses 2,4 with 3,5 ). The amphibole in the coronas from Eselruh and Kumkum gabbro-norite is pargasite (Table 43 ). These amphiboles have lower  $\text{TiO}_2$ ,  $\text{FeO}^*$  and  $\text{K}_2\text{O}$  and higher  $\text{Al}_2\text{O}_3$ ,  $\text{MgO}$ ,  $\text{CaO}$  and  $\text{Na}_2\text{O}$  than the amphibole in the host rocks. The amphibole in the coronas in a troctolite (sample TV49, Table 43) is chemically very similar to sample 520 described by Sapountzis (1975, Table 1). They are essentially Tschermakitic hornblendes. Sapountzis ascribed the high  $\text{Al}_2\text{O}_3$  (22,81%) of the amphibole to a symplectically intergrown aluminous-rich phase, probably spinel. This was also found to be the case for sample TV49 but the fineness of the intergrowth prevented the analysis of separate phases. The amphibole in the coronas in TV49 is low in  $\text{TiO}_2$ , whereas the rare amphibole found in the rest of the rock is dark brown and is Ti-rich. The same observation was made in samples of Kumkum and Eselruh gabbro-norite.

Qualitative electron microprobe scans across coronas from olivine cores to the plagioclase, show trends for  $\text{Al}_2\text{O}_3$ ,  $\text{SiO}_2$ ,  $\text{FeO}^*$  and  $\text{MgO}$  similar to those represented in Fig 34 (Spry 1969).  $\text{Al}_2\text{O}_3$  increases sharply outside olivine and continues increasing until plagioclase is reached, with a slight peak as the amphibole layer is crossed.  $\text{SiO}_2$  increases from olivine to orthopyroxene, decreases abruptly in amphibole and increases again in plagioclase.  $\text{FeO}^*$  and  $\text{MgO}$  decrease steadily from olivine to amphibole and levels off in unzoned plagioclase.

Olivine-plagioclase coronas have been widely reported in the literature; they have been found in gabbros (eg Sapountzis 1975; Grieve and Gittins 1975; Starmer 1969) metagabbros (eg Buddington 1952; Whitney and McLelland 1975) metadolerites (eg Gjelsvik 1952; England 1974) troctolites (eg Mason 1967; Huang and Merritt 1954) and along the contacts between peridotite and anorthosite layers (Griffin 1971b). The above intrusives are generally found in terrains characterized by amphibolite to granulite facies metamorphism although it is interesting to note that olivine-plagioclase coronites have also been found in lunar rocks such as troctolite (Gooley et al 1974) and "dunite" (Dymek et al 1975). The reaction of olivine and plagioclase generally gives assemblages of pyroxenes

and an aluminous phase, the latter being spinel or garnet depending on the P-T conditions (eg O'Hara 1967). Numerous attempts have been made to determine the upper stability limits of coexisting olivine and plagioclase but as yet the results are at variance with one another due to the use of differing starting materials, seeding problems, low reaction rates, imprecision in measurement of pressure and temperature, difficulties in identifying stable reaction products and the narrow range of temperatures at which experiments can be run. There is agreement amongst experimental workers as to the slope of this reaction (ie  $\frac{dP}{dT}$  is positive) (ie Green and Ringwood 1967b; Irving and Green 1970; Kushiro and Yoder 1966) but disagreement as to the position. This is due mainly to differences in starting material used, ie pure, artificial as opposed to natural systems. The anorthite molecule in plagioclase coexisting with olivine is partly stabilized to higher P if albite is present as a major component (Green and Ringwood 1967b). Also if iron is present in the system as a major component then the pressure at which garnet will appear decreases. Thus, the intermediate assemblage pyroxenes plus spinel will be suppressed and the assemblage garnet plus pyroxenes plus plagioclase will form instead. The experimental results on the system olivine<sup>+</sup>

plagioclase which are mutually consistent and are in agreement with hypotheses postulated for the origin of olivine-plagioclase coronas have been synthesized by Griffin and Heier (1973 Fig 3), part of which is reproduced here as Fig ( 37 ).

Various proposals for the origin of this type of corona have been made:

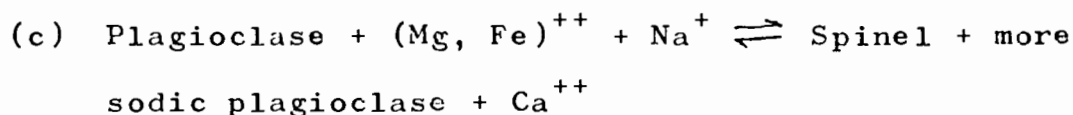
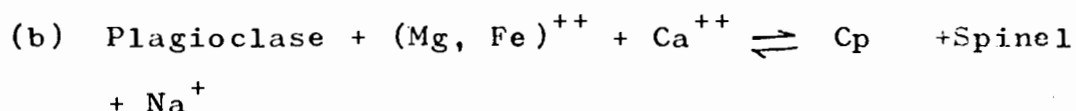
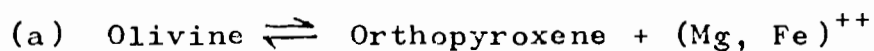
1. Prograde regional metamorphism (Murthy 1958; Whitney and McLelland 1975; England 1974).
2. Thermal Metamorphism (Huang and Merritt 1954; Shand 1945; Friedman 1955).
3. Metasomatism during Regional Metamorphism (Reynolds and Fredrickson 1962).
4. Sub-solidus cooling of magmatic rocks at moderate to high P (Mason 1967; Starmer 1969; Griffin and Heier 1973; Gardner and Robins 1974; Gooley et al 1974; Sapountzis 1975; Esbensen 1978).

Griffin and Heier (1973) reject the metasomatic hypothesis because it is possible to write balanced equations to express each of the reactions taking place in corona formation using only the observed mineral constituents of the rocks. In the case of hydrous coronas it is necessary to add only water to these reactions. The corona-forming reaction is essentially isochemical according to these authors.

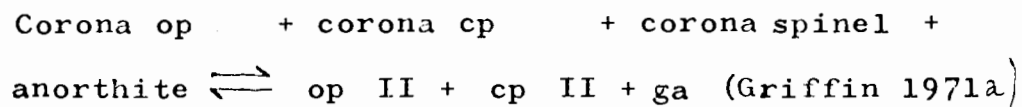
The fact that corona phases are stable at lower temperatures and/or higher pressures than the primary phases, since  $dP/dT > 0$  for the corona-forming reaction, makes it unlikely that the coronas have formed by thermal or regional metamorphism. Also regional metamorphism usually involves penetrative deformation which would be expected to destroy delicate corona textures. Griffin and Heier (1973) conclude that olivine-plagioclase coronas are most likely to have developed during cooling of intrusive rocks from igneous temperatures at moderate to high pressure. They state that the cooling will tend toward the local geotherm and will be considerably faster than uplift. This hypothesis has the advantage of being simple, is in agreement with experimental work and explains, "the ubiquitous presence of coronas in the igneous rocks of deep-seated terranes, as well as the preservation of delicate corona textures and of relic igneous textures in the metamorphosed rocks" (Griffin and Heier 1973, p 327).

The mechanism usually invoked for corona formation is two way sub-solidus diffusion of material across the olivine-plagioclase interface with, or without, the presence of an aqueous medium (eg Mason 1967; Sapountzis 1975; Esbensen 1978). Reactions take

place under essentially closed system conditions although small amounts of water may have been introduced from the country rocks (Mason 1967; England 1974). Whitney and McLelland (1975) proposed three partial reactions for the production of olivine coronas, namely:



Reaction (a) occurs adjacent to olivine, reaction (b) occurs adjacent to plagioclase and reaction (c) occurs in the surrounding plagioclase and explains both the presence of spinel inclusions in the latter and its depletion in Ca in the vicinity of coronas (Starmer 1969; Griffin and Heier 1969; Goode 1970; Whitney and McLelland 1973; Grieve and Gittins 1975). Under conditions of further cooling or increased pressure, garnet forms according to the reaction.



The presence of amphibole in place of clinopyroxene in some coronas is considered to be due to hydrous equivalents of the above reactions.

The above process involves redistribution of certain

elements such as Mg and Fe into the outer rim and Si into the inner rim. Small increases in Mg/Fe ratios from olivine to the coronas were obtained by Murthy (1958), Mason (1967), Reynolds and Frederickson (1962), Grieve and Gittins (1975), and Griffin and Heier (1973).

In summary, it appears that during the formation of corona orthopyroxene and amphibole that some iron was liberated in excess of that combined with Mg in these minerals and could give rise to rims of magnetite along interfaces of corona minerals, or clouding in adjacent plagioclase (Starmer 1969). There is a similarity in compositions of olivine in coronas and unreacted olivine included in other minerals in the same rock from the study area.

This is the result of diffusion of Fe, Mg and Si being limited to that part of olivine which has reacted. Also Mg/Fe bronzite is greater than Mg/Fe olivine which is in agreement with the findings of Murthy (1958) and others.

The replacement of plagioclase by clinopyroxene (or amphibole) required a reduction in the amount of Al present, but since the latter is comparatively immobile, it remained to form spinel of the spinel-hercynite series (Reynolds and Frederickson 1962; Starmer 1969). Ti is predictably absent or



very minor in corona minerals (Grieve and Gittins 1975) because of the low contents of Ti in olivine and plagioclase. The high Na and K of amphibole in some coronas (eg Eselruh and Kumkum gabbro-norites) is difficult to explain by purely isochemical reaction of olivine and plagioclase and it is probable that these ions may have been introduced together with hydrous solutions. Esbensen (1978 p 131) also explained this problem by proposing that the corona-forming process involves the rock as a whole. The assumption that the reaction is isochemical is only correct to a first approximation.

Orthopyroxene-amphibole interfaces do not represent the original plagioclase-olivine contacts because boundaries migrate in solid-state reactions as a result of differing diffusion rates, the Hartley-Kirkendall effect, (Grieve and Gittins 1975). Diffusion of Si from plagioclase into olivine is slower than that of Fe and Mg from olivine into plagioclase so that the movement of the reaction boundary is towards olivine

(Sapountzis 1975). *This is in contrast with most of the earlier workers and Griffin and Heier (1973) who take the pyroxene-symplectite interface as the original olivine-plagioclase boundary.*

The corona minerals are formed in a semi-restricted environment ie one in which interference

and competition between growing crystals is common, which tends to favour fast growing nuclei and explains the orientation of the orthopyroxene fibres with (001) perpendicular to the boundary with olivine, this being the fastest growth direction (Goode 1970). If fewer nuclei were involved in developing the symplectite zone than the orthopyroxene zone, the decreased lateral competition between nuclei in the former would explain the fanning out of the wormy intergrowths away from the interface with orthopyroxene. The scalloped nature of contacts between the symplectite zone and the plagioclase is probably as a result of some areas growing faster or for longer periods than others (Goode 1970).

Toogood (1976) reported (olivine)-orthopyroxene-clinopyroxene/green spinel-(plagioclase) coronas in mafic rocks showing ophitic textures from the Onseepkans area (Toogood 1976, Plate 11) and attributed their origin to prograde regional metamorphism. He interpreted the mafic rocks of the Kumkum area, in which similar coronas exist, as charnockite and granulite in spite of their lack of deformation or alteration and the fact that they have a cumulus texture, and a lamination defined by plagioclase. The corona textures found in the Tantalite Valley gabbro-norite and ultramafic

rocks, and Kumkum and Eselruh gabbro-norites are here interpreted as being due to cooling of these intrusives from igneous temperatures of approximately  $1\ 200^{\circ}\text{C}$  under conditions of medium to high pressure in order that the reaction olivine + plagioclase +  $\text{H}_2\text{O} \rightleftharpoons$  bronzite + clinopyroxene + amphibole + spinel could take place. The absence of garnet from the coronas indicates that the reaction Amphibole + spinel + plagioclase  $\rightleftharpoons$  garnet + pyroxene which is favoured at higher pressures has not taken place.

Coexisting pyroxenes from the Tantalite Valley gabbro-norites and mafic-ultramafic rocks give temperatures of around  $1\ 000^{\circ}\text{C}$  using the method of Wood and Banno (1973), whereas those from the Kumkum and Eselruh gabbro-norites which are texturally and mineralogically very similar to the former give temperatures of around  $800^{\circ}\text{C}$ . The  $1\ 000^{\circ}\text{C}$  temperature is interpreted as the intrusion temperature, whereas the  $800^{\circ}\text{C}$  figure is seen as the temperature of the country rocks to which the Kumkum mafics cooled after intrusion. Thus it seems likely that the Tantalite Valley gabbro-norites and mafic-ultramafic rocks were intruded and cooled in rocks which were at a comparatively low temperature, whereas those in the Kumkum area were intruded and cooled when the rocks were undergoing

high grade metamorphism. This conclusion is supported by the fact that corona textures are better developed in Kumkum rocks than those from the Tantalite Valley.

Coexisting olivine and plagioclase are apparently stable in "ultramafic charnockite" from the Aus area (Jackson 1976). However, the presence of the paragenesis 'hypersthene + diopside + plagioclase + spinel + magnetite' does indicate that a reaction between olivine and plagioclase has occurred in granulites which generally only contain coexisting ortho- and *clino*-pyroxene with no olivine.

Moore and Reid (pers. comm.) have recorded the presence of coronas of (olivine) - orthopyroxene rimmed by black spinel-green calcic amphibole/green spinel symplectite - (plagioclase) from the olivine gabbro-norite on the farm Nouzees, south of Pofadder. The rocks consist of gabbro-norites and olivine gabbro-norites showing cumulus textures, igneous lamination defined by plagioclase, and marginal alteration to amphibolite where in contact with the country rock. This body has been termed a charnockite by Jackson (1976) but the presence of

corona structures and igneous textures implies cooling of an intrusive body under conditions of high to medium pressure.

If an intrusion temperature of around  $1\ 000^{\circ}\text{C}$  is assumed, then a pressure ranging from 6 to 8 kb is necessary for the reaction:-

Olivine + plagioclase  $\rightleftharpoons$  Pyroxenes + spinel to move to the right, depending on which of the three equilibrium curves presented in Fig ( 37 ) is used. Green and Ringwood (1967b) used natural rocks for their experiments compared with the other workers who used pure Fo - An systems so that their results would be expected to be more applicable. However, the figure of 8 kb (at  $1\ 000^{\circ}\text{C}$ ) obtained from their curve is certainly too high since the gabbro-norites in which the coronas are found were intruded after the metagabbro which was emplaced at a pressure of around 5,5 kb (section 7.1) and staurolite breakdown in the aureole of the metagabbro indicates that pressures decreased after contact metamorphism (section 7.1.1). The experimental curve of Kushiro and Yoder (1966) gives the result in best agreement with the 5,5 kb predicted, ie 6 kb at  $1\ 000^{\circ}\text{C}$ .

#### 4.2 Opaque mineral-plagioclase coronas

Gabbro-norites from the Tantalite Valley and

Kumkum areas may develop coronas where inter-cumulus opaque minerals, (titanomagnetite) with which may be associated primary green spinel, is in contact with plagioclase. These single coronas consist of greenish-brown amphibole symplectically intergrown with elongate green spinel (Fig 32E). Some phlogopite or biotite may occur along contacts with opaque minerals. Spinel fans out in the amphibole away from, and at right angles to, the opaque mineral interface and contacts with plagioclase are usually scalloped.

Reports of synantectic coronas formed between opaque minerals and plagioclase are less common in the literature than for those formed between olivine and plagioclase. Whitney and McLelland (1975) described ilmenite, magnetite and/or brown-green spinel surrounded by a rim of calcic amphibole which developed radiating intergrowths of green spinel in contact with plagioclase. Some flakes of biotite were observed along the contacts of the opaque mineral and the amphibole. Similar coronas were described by Grieve and Gittins (1975) and Esbensen (1978) for ilmenite in contact with plagioclase, and by Reynolds and Frederickson (1962) for magnetite in contact with plagioclase. In the latter occurrence biotite is absent and it is likely that this mineral does not play any part in

the corona reactions at all (see also Starmer 1969). Brøgger (1934), Friedman (1955) and Starmer (1969), reported garnet as forming the outer rim of coronas about magnetite, titanomagnetite, ilmenite and, rarely, pyrite with which may be associated primary spinel. It appears from these authors that such coronas do not occur in rocks where olivine is absent, and, in fact, are usually associated with rocks containing olivine-plagioclase coronas. Grieve and Gittins (1975) and Esbensen (1978) found that the amphiboles of coronas around opaque minerals in contact with plagioclase have a high Ti content in contrast to those found in olivine-plagioclase coronas.

It is generally accepted by most authors that these coronas form in the same way and under similar circumstances as the olivine-plagioclase coronas. Hornblende could not have originated by direct reaction of oxide and plagioclase since no MgO is present in this system and some FeO must also have been introduced (Esbensen 1978). The opaque mineral shows little or no corrosion (Starmer 1969; Esbensen 1978) and the corona forming reaction involves mainly plagioclase and the introduction, by diffusion, of Fe and Mg ions arising from the conversion of olivine to bronzite in olivine-plagioclase coronas. The opaque mineral/plagioclase interface serves only

as a good nucleation site for migrating ions.

Biotite only develops on this interface if there is sufficient potassium in the rock.

#### 4.3 Vermicular Orthopyroxene-magnetite symplectites

Orthopyroxene-magnetite symplectites are common in the olivine gabbronorites of Tantalite Valley but this texture occurs in only one sample from the Kumkum gabbronorite (ie K250). Vermiform intergrowths of magnetite with green or colourless amphibole are also occasionally observed in the metagabbro from the Tantalite Valley Complex (Fig 34J).

In the fresh rock only a very small proportion of the olivine grains are replaced by symplectites but where this process has gone to completion large symplectite mats up to several millimeters in size are found with magnetite being either randomly oriented or aligned in orthopyroxene (Fig 34I). Symplectites in various stages of development are generally very similar in appearance to the examples presented as Figs 3 to 10 in Ambler and Ashley (1977). Symplectites appear to pseudomorph olivine, preserving original olivine grain boundaries in the process. Olivine compositions are the same in grains being altered as in those which are un-



affected. Symplectites may be surrounded by primary igneous minerals. Orthopyroxene in symplectites is commonly in optical continuity with surrounding magnetite-free orthopyroxene rims.

Orthopyroxene-opaque mineral symplectites are relatively common in olivine-bearing mafic and ultramafic igneous rocks and their metamorphosed equivalents (Ambler and Ashley 1977). They have been reported from terrestrial olivine-bearing cumulates (eg Goode 1974; Haselton and Nash 1975; Ambler and Ashley 1977) and from similar lunar rocks (eg Bell et al 1975; Dymek et al 1975) as well as inclusions from South African kimberlites (Haselton and Nash 1975).

The symplectites commonly take the form of vermicular intergrowths of orthopyroxene and spinellids (eg spinel, pleonaste, magnetite and chromite) in the case of terrestrial olivine-bearing cumulates or vermicular intergrowths of ilmenite and orthopyroxene in the case of kimberlites and lunar olivine-bearing cumulates (Ambler and Ashley 1977). Generally the orthopyroxene to opaque mineral ratio is relatively constant, ie about 60:40, and the symplectite orthopyroxene and opaque minerals are almost identical in composition to orthopyroxene and opaque minerals found in the rest of the rock. Theories which have been proposed to explain the

origin of such symplectites include reaction of olivine with an interstitial late stage magmatic fluid phase (eg Gardner and Robins 1974; Ambler and Ashley 1977) and solid state reactions involving subsolidus oxidation of olivine (Goode 1974). Ambler and Ashley (1977) demonstrated that symplectites are unlikely to have been produced by the mechanism proposed by Goode (1974). Instead they suggest that orthopyroxene-magnetite symplectites result from "simultaneous coprecipitation of orthopyroxene and magnetite at dispersed locations in a crystal mush where olivine was reacting with late stage liquids." (p 172). They envisage that at this stage 80 - 90% of the volume of the system was solid; which would explain why this texture is developed only in parts of a rock. This hypothesis is accepted for the formation of similar symplectite textures found in rocks from the study area on the grounds of mineralogical, textural and chemical similarities. The inferences that can be made from the presence of relic symplectites in metagabbro are two-fold: first, that some of the metagabbro originally contained olivine and second, that late magmatic reactions took place between solidus phases and a magmatic liquid.

#### 4.4 Clouding in Plagioclase

Clouding in plagioclase is a common feature of the metagabbros, gabbro-norites, olivine gabbro-norites and ultramafic rocks of the Tantalite Valley Complex and the Kumkum gabbro-norite. Inclusions mainly consist of opaque oxides but hercynite and/or hornblende may also occur, especially near coronas where plagioclase is in contact with other minerals, including other grains of plagioclase, in olivine-bearing and olivine-free rocks, it has rims which are clear and more sodic than the cores. The intensity of clouding developed is extremely variable, with plagioclase in ultramafic rocks showing the least clouding.

Clouding is a common feature observed in plagioclases from gabbros and dolerites from metamorphic terranes and in plagioclases from volcanic rocks (Armbrustmacher and Banks 1974). Inclusions in plagioclase which are commonly too small to be identified using conventional microscopic methods, are usually crystallographically oriented parallel to (010) and (001). They have been identified by different workers as magnetite, maghemite, haematite, spinel, rutile and hornblende. Burns (1966) attributed the degree of clouding in plagioclase in the Scourie dykes of Scotland to the higher than normal  $H_2O$  content of the feldspar while

Bridgewater (1966) showed that the dark colour of feldspars in some charnockites and metagabbros is due to microscopic structural defects and high water content. Margins of clouded plagioclase grains are commonly inclusion-free and are more sodic than the rest of the grain, observations which are confirmed for the present study area. The reason for these features is still a problem. Either material making up the inclusions migrated towards the centre of the grain or it was lost to outside minerals. The first hypothesis is here considered unlikely because the iron content in clouded plagioclase was probably originally similar to that in unclouded plagioclase (Armbrustmacher and Banks 1974) and for any diffusion process to occur in a grain requires that a concentration gradient exist for the diffusing elements. In the case of the possibility of material being lost to outside minerals, it is interesting to note that clouded plagioclase with clear rims is commonly present near coronites in the rocks under discussion, and have been reported in the literature (Friedman 1955; Poldervaart and Gilkey 1954; Murthy 1958). Elements found in inclusions in the plagioclase and some calcium from the rims of plagioclase may have been used to produce the amphibole found in coronas. Where plagioclase is in contact with other

grains of plagioclase, the clear, sodic rims are probably due to redistribution of elements by intergranular fluids or by the formation of hornblende along grain margins, a feature which is relatively common in coronites (Starmer 1969).

Clouding in plagioclase was for a long time taken as evidence for a rock having been metamorphosed, but Poldervaart and Gilkey (1954) showed that this is not always the case. According to Spry (1969), the requisites for clouding in plagioclase appear to be an elevated temperature over a long period, the presence of water and a supply of iron, the common constituent of inclusions. Armbrustmacher and Banks (1974) showed that the source of the iron is the crystal itself and that slow cooling causes exsolution of this element which is stabilized in the plagioclase lattice at high temperatures. Similarly, expulsion of Ti from the plagioclase lattice coupled with the introduction, by diffusion, of such elements as Mg and Fe, arising from corona-forming reactions, is probably responsible for minerals such as ilmenite and spinel causing clouding of the plagioclase. The need for slow cooling probably explains the common association of plagioclase clouding with rocks intruded into metamorphic terrains. The presence

of water, which probably acted as a catalyst, is inferred from the common association of hydrous phases, eg hornblende with development of clouding. The difference in intensity of clouding observed in plutonic and hypabyssal mafic rocks (Armbrustmacher and Banks 1974) is probably due to the fact that the former have cooled more slowly than the latter and there has been more time for the development of clouding.

A theory of sub-solidus exsolution under slow cooling conditions for clouding observed in the plagioclase in mafic rocks of the study area is compatible with the conclusions reached on the development of coronas in these rocks and the re-equilibration observed in coexisting pyroxenes in the Kumkum gabbro-norite.

#### 4.5 Sector Zoning in Clinopyroxenes from Eselruh gabbro-norite

The Eselruh gabbro-norite is characterized by large, euhedral, cumulus crystals of diopside displaying well-developed hour-glass structures defined by opaque mineral (possibly magnetite) exsolution in the host. Sector zoning is apparently absent from the less abundant coexisting bronzite.

Sector zoning has been recorded in staurolite, chloritoid, quartz, clinopyroxene and orthopyroxene (Johannsen 1931; Goode and Krieg 1967; Hollister 1970; Moore 1971; Hollister and Gancarz 1971).

This phenomenon has been observed for pyroxenes from volcanic, hypabyssal and cumulus rocks (eg Goode and Krieg 1967; Strong 1969; Hollister and Gancarz 1971; Gray 1971).

The hourglass structure in pyroxene can be thought of as being formed by the intersection of pyramids, each one identifiable by the crystal face that would form its base, whether this face is developed or not. All pyramids that have for their bases faces of the same crystal form are collectively named after that form. For the four forms  $\{100\}$ ,  $\{010\}$ ,  $\{110\}$  and  $\{\bar{1}11\}$  in clinopyroxene there are four corresponding sets of pyramids, which are designated the (100), (010), (110) and ( $\bar{1}11$ ) sectors, (Hollister and Gancarz 1971) which differ slightly from one another chemically. All four kinds of sectors were identified in various samples of Eselruh gabbro-norite (Fig 38 ).

The presence of compositional sector-zoning in clinopyroxene is a probable indication that that mineral has grown relatively rapidly and in an unrestricted crystallization environment (Goode and Krieg 1967; Hollister and Gancarz 1971).

In slowly cooled rocks (eg Eselruh gabbronorite) a preferential adsorption hypothesis is favoured to explain compositional sector zoning in augite (Strong 1969) and orthopyroxene (Goode and Krieg 1967). Strong (1969) proposed that hourglass shaped skeletal crystals ('hoppers') were initially formed as a result of restricting the growth of certain crystal faces by selective adsorption of impurities on to these. Later crystallization would "fill out" these hoppers by adding augite enriched in  $\text{FeO}$ ,  $\text{Na}_2\text{O}$ ,  $\text{TiO}_2$  and  $\text{Al}_2\text{O}_3$  and depleted in  $\text{MgO}$  and  $\text{CaO}$  compared to the hopper, which would, on exsolution of opaque minerals, form the dark part observed in a sector. Compositional sector zoning involving elements in the same relationship as above, was also found in a study of clinopyroxenes from volcanic rocks made by Hollister and Gancarz (1971). Thus it seems that Strong's hypothesis for the generation of hourglass structure, at least for clinopyroxenes, is applicable for widely different environments and is therefore the one favoured here to explain the structures observed in the diopsides of the Eselruh gabbronorite.

#### 4.6 Exsolution in pyroxenes of the Tantalite Valley Complex



Ortho- and clino-pyroxenes of the gabbro-norites and norites of the Tantalite Valley Complex are characterized by exsolution parallel to (100) of clino- and ortho-pyroxene respectively. Fig ( 39 ) shows a qualitative electron microprobe scan for Fe, Mg, Al and Ca on adjacent bronzite and diopside grains in a norite sample (TV6 129), <sup>(Table 43, p 163)</sup> which is regarded as illustrating the representative type of exsolution found in pyroxenes of these rocks. From the variation in the  $\text{CaK}_{\alpha}$  X-ray intensity, it is seen that lamellae of calcium-poor pyroxene in diopside are very closely spaced and have approximately equal widths (  $\sim 10 \mu\text{m}$  ), whereas exsolution cannot be detected in bronzite using this method, except for a slight increase in  $\text{CaK}_{\alpha}$  intensity near the  $120 \mu\text{m}$  mark (Fig 39 ). This is in agreement with the microscope observations, that exsolution in the orthopyroxene takes the form of scattered blebs of clinopyroxene, whereas in clinopyroxene there are abundant fine lamellae of orthopyroxene parallel to (100). The fine exsolution lamellae observed in the clinopyroxene probably indicate a relatively large interval between crystallization and initiation of exsolution *after* which time slow re-equilibrium of all major and minor elements took place between

host and lamellae (Boyd and Brown 1966, 1967, 1969). Boyd and Brown (1966) showed that there is a definite proportionality between the width of lamellae and the apparent calcium content, taken as being a function of the peak height. They showed that it is practical to carry out electron microprobe analyses of lamellae only where these are wider than about  $6\ \mu\text{m}$ . In sample TV6 129, the lamellae in clinopyroxene are similar in width to those found in the clinopyroxene from Bushveld gabbros (Boyd and Brown 1966) and can be completely resolved using the electron micro-  
*where a finely focussed (1-2  $\mu$  diameter) beam was used*  
 probe. Exsolved clinopyroxene and discrete clinopyroxene in sample TV6 129 have similar compositions, a feature which is commonly observed in pyroxenes from layered complexes. This implies that equilibrium was probably maintained during the cooling period (Boyd and Brown 1967).

Exsolution in pyroxenes from layered igneous complexes usually show orthopyroxenes with clinopyroxene lamellae parallel to (001) in addition to (100), which is the result of inversion from pigeonite. This feature does not occur in any of the orthopyroxene from mafic intrusives in the Kumkum or Tantalite Valley areas. It is concluded that pigeonite was not present in these rocks. Instead, it is proposed that pyroxenes of pre-

dominantly bronzitic and augitic composition crystallized directly from the magma at about 1 200°C (Boyd and Brown 1966) and that the temperatures obtained from coexisting pyroxenes in these rocks (see section 5.1) probably represents that at which subsolidus re-equilibration between lamellae and host took place.

#### 4.7 The Mottled Metagabbro

The mottled metagabbro (see section 3.3.3.1) has been intimately intruded by coarse-grained Inequigranular granite gneiss and to a lesser extent by tonalite. Blue quartz ocelli occur scattered through the mottled metagabbro and, together with plagioclase megacrysts, give the rock its characteristic mottled appearance. Ocelli are surrounded by mantles of hornblende which may, in turn, be rimmed by plagioclase.

Quartz ocelli have been recorded only from basic rocks, particularly volcanics intruded by granitic bodies (see Angus 1971). An ocellus usually is composed of a large, rounded quartz grain which is rimmed by hornblende or diopside arranged tangentially around its margins. In the immediate vicinity of an ocellus hornblende or diopside may be depleted relative to elsewhere in the same rock. Rocks with

abundant ocelli have been called ocellar hybrids by Angus (1962, 1971). He examined various hypotheses proposed for their origin and concluded that ocelli form as a direct result of metasomatic alteration of basic rocks by silica-rich fluids.

Angus (1962) described the stages in the development of quartz ocelli in a basalt, consisting of hornblende, labradorite (mostly in the groundmass but also as rare microphenocrysts), accessory opaque minerals and sphene, which has been intruded by a coarse-grained granodiorite. Since there is a striking similarity between the rocks as described above and the mottled metagabbro which is intruded by the Inequigranular granite gneiss, it is useful to list the stages of ocelli development given by Angus (1962):

1. Quartz is intimately intergrown with plagioclase and some hornblende, and is surrounded by an irregular mantle of hornblende.
2. Ocelli become typically porphyroblastic with a distinct central area of quartz, surrounded by a quartz-plagioclase intergrowth, and mantled by hornblende.
3. The intermediate quartz-plagioclase zone all but disappears.

In the ocelli-forming process the groundmass tex-

ture and mineralogy are considerably modified and the grain size becomes coarser. Quartz has a tendency to separate from plagioclase and become concentrated in ocelli; K-feldspar and biotite appear in minor amounts and apatite becomes more abundant. Groundmass plagioclase becomes more sodic and megacrysts of plagioclase develop sodic rims.

The growth of quartz ocelli was probably achieved by the tendency of incoming siliceous fluids to dissolve plagioclase and at the same time to expel Fe, Mg and Ca. These elements, together with contributions from the groundmass could lead to the fixation of minerals such as hornblende along the rim of the proto-ocellus, which at the same time causes depletion of the same mineral (ie hornblende) in the groundmass. Eventually even the plagioclase, in the form of Ca, Na and Al ions, is displaced from the quartz which remains behind as a fully developed ocellus. In the process Si, Na and lesser amounts of K and P enter the rock and some Fe, Mg, Ca and Ti are removed. Transportation of ions is presumably achieved by permeating intergranular fluids (Angus 1971).

The mottled metagabbro is enriched in P, Na, K, Si and Ti (Fig 70) relative to other mafic rocks of the Tantalite Valley Complex, and it contains

biotite instead of phogopite. These features, together with the presence of ocelli, indicate that the mottled metagabbro formed by metasomatic introduction of granitic liquids from the Inequigranular granite gneiss, in the manner described by Angus (1962). The high Ti content of mottled metagabbro indicates that this rock was already enriched in this element before metasomatism because this process would be expected to deplete Ti (Angus 1971).

The origin of plagioclase megacrysts in mottled metagabbro is difficult to explain. The normal zoning points to an igneous origin, whereas the presence of inclusions of hornblende and opaque minerals indicates a metamorphic origin. Megacrysts are particularly abundant along zones of weakness, eg joints, in the rock and are rare where the rock is massive, indicating that their distribution may be a function of the accessibility of metasomatic fluids. It is unlikely, however that megacrysts were formed directly from such fluids since the Ca content of the megacrysts is so much greater than that found in the plagioclase in granite gneiss. The sodic rims of the megacrysts, on the other hand, are probably a manifestation of metasomatism. Enlargement of pre-existing microphenocrysts acting as nucleation sites in parts of

the rock affected by metasomatism could perhaps have produced the distribution of megacrysts observed in outcrops. In the samples examined, sodic rims around megacrysts are narrow, making this hypothesis unlikely.

It is tentatively postulated here that the plagioclase megacrysts formed as a result of the addition of Ca, Na and Al ions, originating from the sites of ocelli formation in the host rock itself, on to pre-existing microphenocrysts. Subsequently, sodic rims developed around these megacrysts following further introduction of granitic fluids. It is also possible that the megacrysts are a primary feature of the rock which owe their present distribution to flow differentiation. The scattered nature of megacrysts in outcrop and their preference for occurring along joints, makes this unlikely.

#### 4.8 The origin of small-scale layering in Metagabbro

Small-scale layering is best developed near the contact of the metagabbro with the country rocks, and away from these the metagabbro is characteristically massive. Plagioclase and hornblende, (which are secondary after plagioclase and pyroxene), have an ophitic relationship with respect to each other. The layering dips at  $45^{\circ}$  or more towards the centre of the Complex and strikes parallel to the contact

between metagabbro and the country rock. Layers are isomodal (predominantly either hornblende or plagioclase), or they show reverse and normal density-grading, all of which follow one another in regularly repeated sequences (Fig 22 ). The layering dies out abruptly both laterally and vertically, and is replaced by massive metagabbro. It is stressed that layering in the metagabbro is rare.

The origin of steeply dipping layering showing features typical of gravity settling in the Tantalite Valley Complex is difficult to explain. Possible mechanisms proposed by Goode (1977) for the formation of such layering are:-

- (a) Tectonic layering
- (b) Wall crystallization
- (c) Folding of gravity-settled layering
- (d) Flow differentiation
- (e) Mush flow.

The layering is clearly not tectonic in origin as is seen from the presence of igneous textures and compositional layering. By varying physico-chemical conditions during crystallization a compositionally layered sequence could be built up along a contact wall. Wall crystallization is unlikely to have taken place in the metagabbro since



comb textures characteristic of this form of crystallization are totally absent. Further it is difficult to imagine local and repetitive changes in crystallizing conditions as would be required to produce the rare, rhythmic layering observed. The Tantalite Valley metagabbro is foliated and deformed mainly along contacts with the country rock so that (c) is a likely possibility. However, drilling and geophysical evidence point to the Complex having near vertical contacts with the surrounding rocks. It is possible though, that the contact relationship was modified by later shearing, since in the core of the shear zone, dips are close to vertical. This may explain why the layering is preserved some distance away from the country rock contact.

McClay and Campbell (1976) reported the presence of rhythmic layering and cumulate textures on near-vertical surfaces from the marginal series of the Jimberlana Intrusion, West Australia. They ascribed the layering and cumulate features to a combination of convection and flow with the role of gravity being replaced by that component of the flow which pushes the cumulus grains towards the vertical layered series developing in the centre of the intrusion. The flow differentiation hypothesis however, suffers from a major disadvantage in that it cannot explain the formation of a large number of bands (Goode 1977).

Also such features as reverse fractionation and a decreasing concentration of phenocrysts towards the walls are not observed in the metagabbro layering, as might be expected if the layering formed by flow differentiation processes (McClay and Campbell 1976).

Layering can also be produced by differential viscous flow of a crystal mush along a steep contact. (Goode 1977). This mechanism has been invoked only to explain banding and isomodal layering found along steeply dipping contacts and it is hard to see how mush flow could produce reverse and normal grading as observed in the metagabbro.

It is concluded here that the layering observed in the metagabbro was formed on a horizontal or shallowly dipping surface which subsequently underwent folding. However, evidence that this has occurred is lacking. Slumping of initially horizontal layering as a result of contraction of the core of the magma chamber by eg crystallization, could also have produced the present disposition of the layering.

The presence of density graded layering in metagabbro suggests that gravity settling has occurred. Separation of plagioclase from pyroxene in the layering is probably as a consequence of differential settling.

In general, the degree of separation of simultaneously crystallizing phases will depend on the settling distance, the relative settling velocities, and the concentration of crystals or the thickness of the crystallization zone, which controls the rate of crystal accumulation (Goode 1976). The parameters  $L$ , the distance above the top of the crystal pile, and  $L_c$ , the critical distance below the nucleation zone at which complete separation of pyroxene/olivine and plagioclase occurs (Fig 40 ) were defined by Goode (1976, p 391).

According to him, massive gabbroids would form as a result of continuous crystallization whereas isomodal and graded layering would result from repeated bursts of discontinuous crystallization when  $L > L_c$  and  $L < L_c$  respectively. Continuous nucleation is favoured in a slow cooling, near-supersaturated environment which explains why this phenomenon is observed only in large plutons. The metagabbro component of the Tantalite Valley Complex, which according to Goode (1976) qualifies as a large pluton, it being greater than 100m in diameter, must consequently have undergone mostly continuous crystallization to form massive metagabbro and only near contacts with country rocks were conditions right for discontinuous crystallization to form layering. Fluctuations in the

degree of saturation of a melt caused by, for example, intermittent seismic tremors (Hoffer 1965) could result in conditions of discontinuous crystallization and the development of isomodal and graded layering.

It is proposed here that sporadic movement along the TV lineament could have been transmitted into the cooling metagabbro near contacts with the country rock, inducing conditions suitable for discontinuous nucleation and the production of density graded layers and isomodal layers as proposed by Goode (1976). Reverse graded layering probably formed as a result of settling overlap of plagioclase by pyroxene in later nucleation cycles (Goode 1976).

#### 4.9 Pegmatitic schlieren in Metagabbro

Pegmatitic metagabbro is found in Tantalite Valley metagabbro either as irregular veins and lenses or as thin selvages concordant with the layering (Fig 20 ). It is rare and apparently randomly distributed. Hornblende crystals may attain several centimeters in length. Plagioclase is commonly clouded but has clear rims. These minerals show an ophitic relationship towards each other.

Pegmatitic schlieren, up to a metre in thickness, have been described from the upper parts of large dolerite intrusions from tholeiitic provinces (eg Walker and Poldervaart 1949). They are found towards the tops of sills, approximately parallel to the contact with the country rock, or they develop in the vicinity of xenoliths.

Walker (1953) proposed that they are formed as a result of crystal fractionation where fugitive constituents and late-stage melts were trapped beneath a virtually impermeable roof, leaving the derivative magmas to crystallize under conditions of reduced viscosity. Decreased cooling rates at lower temperatures within a cooling body and depression of freezing points were probably responsible for developing a pegmatitic texture (Eales and Booth 1974). The coarse grain-size suggests crystallization under high  $P_{H_2O}$  and possibly of increased  $P_{O_2}$  over that found in the host rock. Eales and Booth (1974) and Eales and Robey (1976) showed that pegmatitic dolerite is enriched in  $SiO_2$ , alkalis and Rb at the expense of  $Al_2O_3$ , MgO and CaO compared with gabbros, olivine-gabbros and dolerites, and that the Ca-pyroxene in the schlieren is more iron-rich than that found in olivine-gabbro.

The pegmatitic mafic rocks at Tantalite Valley are

here attributed to the confluence, as a result of diffusion, of pockets of intercumulus, volatile-rich magmas distributed in a crystalline matrix. When such magmas are unable to penetrate the overlying rocks they will probably make room for themselves vertically and horizontally and a pegmatitic layer will result on crystallization. If a zone of weakness, eg a joint, exists in the cover rocks, then the volatile-enriched magmas can escape and form pegmatitic dykes.

#### 4.10 Intergrowths of Prehnite and Biotite

Prehnite intergrown with biotite occurs in mottled metagabbro, metagabbro, calc-silicate rocks, quartzite, feldspathic Black Gneiss, amphibolite, Inequigranular granite gneiss and pelitic and metabasite hornfels. In one sample of troctolite (AM89B) prehnite is also found along cracks in plagioclase. All rocks containing prehnite in biotite are characterized by the presence of sericitized plagioclase, free quartz (except troctolite) and brown mica with or without the presence of calcic amphibole. Prehnite commonly occurs as colourless, lens-shaped pods, parallel to the (001) cleavage in biotite or its chloritized equivalent. The bending of biotite cleavage around these inclusions suggests that prehnite

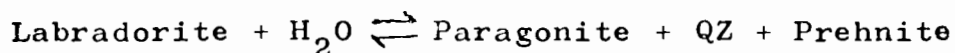
formation involved an increase in total volume which is the reverse of what is expected if prehnite metasomatically replaces biotite (Moore 1976).

The occurrence of prehnite within biotite is probably fairly common (see Phillips and Rickwood 1975) but the fact that it has not been more widely reported in the literature is possibly due to the difficulty in identifying prehnite using routine petrographic methods.

The occurrence of prehnite within biotite, and chlorite, has been long recognised in mafic and dioritic rocks (Osborne and Lovering 1953; Wells and Bishop 1955; Hall 1965; Field and Rodwell 1968; Maggetti and Nickel 1973; Phillips and Rickwood 1975; Moore 1976) but this is the first record of it occurring in quartzite, calc-silicate rocks and pelitic hornfels.

There is an apparently ubiquitous association of biotite-prehnite with altered plagioclase which has led some authors to propose that prehnite formed in biotite as a result of the release of Ca ions into the system caused by the breakdown of plagioclase by late-stage, K-rich, hydrothermal fluids (eg Field and Rodwell 1968). Wells and Bishop (1955) suggested that biotite and prehnite formed simultaneously from hornblende as a result of low temperature, K-metasomatism. The need for addition

of K has been questioned by Rickwood and Phillips (1976), who note that biotite is commonly chloritized, implying removal of K from the system. Phillips and Rickwood (1975) suggested a simple deuteric origin for prehnite and showed that it forms independently of any reactions involving biotite, amphibole or plagioclase. Moore (1976) also accepted a deuteric hypothesis but proposed that the conversion of pyroxene to amphibole could have led to the production of prehnite, although he discarded plagioclase as a possible source on the grounds that epidote and not prehnite occurs in the plagioclase. His suggestion that biotite cleavages merely acted as nucleating sites for prehnite was accepted by Rickwood and Phillips (1976), who also re-examined the question of whether the breakdown of plagioclase was involved in prehnite formation. They proposed the reaction:-



For this reaction to proceed, only addition of water is required and the products have all been identified in altered mafic rocks. They posed the question that if this reaction is valid, which seems highly likely, why does prehnite not occur more commonly in rocks. This condition they ascribed to the scarcity of rocks in which the minerals calcic



plagioclase, quartz and a platy mineral (eg biotite), needed for nucleation purposes, are found. That prehnite and biotite are found in a large variety of rock types is clear from the previous discussion. Also prehnite might not have been recognized by many workers.

Prehnite is a mineral diagnostic of the prehnite-pumpellyite facies (Turner 1968, page 52). A range of temperature from about 250°C to 380°C at 3 kb is considered by Liou (1971) to be a reasonable estimate for the stability field of this facies. On the basis of his experimental results, Liou (1971) showed that the high temperature stability limit of prehnite is 403°C at 3 kbar and 393°C at 5 kbar ( $P_{H_2O} = P_{total}$ ). This stability limit will occur at lower temperatures in nature where  $P_{H_2O}$  is less than  $P_{total}$ .

The prehnite found in rocks of the study area was probably formed late in the metamorphic history of the region, possibly during uplift and late stage shearing. During this period, pressures were probably lower than about 3 kb, as deduced from the breakdown of staurolite in the Tantalite Valley aureole, so that the prehnite formed at temperatures of less than 403°C. Its formation by alteration of plagioclase would explain the presence of this mineral in a large variety of rock-types.

## CHAPTER V

## GEOTHERMOMETRY AND GEOBAROMETRY

5.1 Two-pyroxene geothermometer

Equilibrium temperatures have been calculated for coexisting ortho- and clino-pyroxene from various mafic and ultramafic rocks (Table 33 ) using equations derived by Wood and Banno (1973) and Wells (1977) (Appendix 6 ).

Wood and Banno (1973) have assumed all iron is  $\text{Fe}^{2+}$ . However if some of the iron is  $\text{Fe}^{3+}$ , then the calculated equilibrium temperature would be increased because of the decreased  $X_{\text{Fe}}^{\text{opx}}$  value in the equation. For calculations on the pyroxenes here all iron was assumed to be ferrous. Some analyses (Table 33 ) were recalculated after estimating the proportion of ferric iron on the basis of charge balance and stoichiometry. For samples TV263 and TV233 (discrete grains) (Table 43 ),  $\text{Fe}^{3+}$  is significant in clinopyroxene and orthopyroxene respectively, with the result that the calculated temperatures are increased by 18 to 25°C above those calculated using  $\text{Fe}^{2+}$  only. The correction for  $\text{Fe}^{3+}$  for the other results makes little difference and has been neglected. Gener-

ally, the agreement between the temperatures obtained using the equations of Wood and Banno (1973) and Wells (1977) is quite good and within the  $\pm 70^{\circ}\text{C}$  error limit set by these authors. Also the temperatures obtained, using analyses from exsolution lamellae are in agreement with those obtained from discrete phases and it has been postulated elsewhere (section 4.6 ) that these temperatures represent sub-solidus re-equilibration temperatures.

In order to make meaningful comparisons of temperatures derived from pyroxenes in different rock types, it is necessary that the rocks be chemically and mineralogically similar, a condition which is met in samples from the present area. From Table ( 33 ) the absolute range of temperatures recorded from the rocks of Tantalite Valley is  $917 - 1108^{\circ}\text{C}$  whereas a range of  $762 - 914^{\circ}\text{C}$  is obtained for the Kumkum samples.

Toogood (1976 Fig 24) and Jackson (1976 Fig 61) classify the gabbro-norite of Tantalite Valley as charnockite. Co-existing orthopyroxene-clinopyroxene pairs of the charnockite series of Madras give temperatures of  $841 \pm 11^{\circ}\text{C}$  (see Hewins 1975) while those from Kondapalli yield temperatures of  $859 \pm 9^{\circ}\text{C}$  (Leelanadam 1967). The temperatures

obtained from Tantalite Valley gabbro-norite are therefore too high to fall into the range of mafic charnockites but are too low, due to a degree of sub-solidus re-equilibration, to represent the igneous intrusion temperature (see section 4.6 ). A similar conclusion was reached by Hermes (1970) for the gabbros of the Mecklenburg Complex. He calculated temperatures of approximately 900°C for coexisting pyroxenes. It is unlikely in the case of the Tantalite<sup>Valley</sup> gabbro-norite or ultramafic rocks that the temperatures recorded from pyroxenes were modified by the prevailing metamorphic conditions since at the time of the emplacement of these rocks the country rocks were probably relatively cool. This is based on evidence that the metagabbro intruded rocks which had already undergone metamorphism and was subsequently altered under lower T, medium P conditions prior to the emplacement of the gabbro-norite.

In the Kumkum area, the temperature of the peak of metamorphism<sup>M1</sup>, as indicated by coexisting pyroxenes in granulite and<sup>some</sup> pelitic mineral assemblages, is similar to temperatures obtained from coexisting pyroxenes from mafic intrusives. A better estimate of the actual intrusion temperature of the latter can be gained from such features as exsolution in pyroxenes, sulphide and silicate parageneses,

textures and the fact that sapphirine-bearing assemblages have been found in the area which suggest temperatures greater than 1000°C were realized, probably along Black Gneiss-mafic intrusive contacts. (see section 7.2.3.1).

From the above it is concluded that the mafic rocks of the Kumkum area were intruded into rocks at the peak of metamorphism (high T, medium P) and subsequently underwent slow sub-solidus re-equilibration until they attained the physical conditions in the surrounding rocks, after which the system was "frozen" possibly by rapid uplift.

## 5.2 Olivine-Spinel Geothermometry

Jackson (1969) devised a geothermometer utilizing coexisting olivine and chromite, the theoretical basis for which is set out in appendix (7) based on Irvine (1965). Temperatures for olivine coexisting with picotite, spinel and ferritchromite in troctolite sample TV49 from Tantalite Valley and for olivine coexisting with hercynite from Kumkum gabbro-norite sample K37A were calculated using the equation (see appendix 7):-

$$T = \frac{5580 \alpha + 1018 \beta - 1720 \gamma + 2400}{0,90 \alpha + 2,56 \beta - 3,08 \gamma - 1,47 + 1.987 \ln K_D^{ol-sp_{Mg-Fe^{2+}}}}$$

and the results presented in Table ( 34 ). The temperature for the spinel-olivine pair in TV49 is thought to approximate equilibrium temperatures because, in spite of the uncertainty involved in this method, it is of similar magnitude to the temperature derived from the pyroxenite sample TV6 129 using coexisting pyroxenes, after Wood and Banno (1973). The ferritchromite in this rock is almost certainly of secondary origin (section 6.5.1). This is also probably the case for coexisting picotite which gives a temperature of  $849^{\circ}\text{C}$ , compared with the  $1205^{\circ}\text{C}$  calculated using spinel. The temperature of  $863^{\circ}\text{C}$  obtained using olivine and spinel in the Kumkum gabbro-norite sample K37A (Table 34) is in good agreement with that of  $808^{\circ}\text{C}$  (Table 33) obtained for this rock using coexisting pyroxenes (Wood and Banno 1973). The temperature of  $914^{\circ}\text{C}$  obtained using the expression of Wells (1977) is thought to be too high to represent the peak of metamorphism in the area since Hewins (1975) has placed the upper limit of hornblende-bearing granulites at  $860^{\circ}\text{C}$  and no hornblende-free granulites are found in the Kumkum area.

Evans and Frost (1975) applied the olivine-spinel geothermometer to a wide range of rock types and found that the temperatures obtained are usually

too high. They proposed instead a plot of normal-  
 ized values of  $\ln K_D^{\text{ol-sp}} \text{Mg-Fe}^{2+}$ , ie  $\ln K_D^{\text{ol-sp}} \text{Mg-Fe}^{2+}$   
 versus  $\alpha$ , where these terms are related by the  
 expression:-

$$\ln K_D^{\text{ol-sp}} \text{Mg-Fe}^{2+} = \ln K_D^{\text{ol-sp}} \text{Mg-Fe}^{2+} + (0,05 - \gamma) \quad 4.$$

0,05 is the value of  $\gamma$  which is used to normalize  
 chromites from high grade rocks. In a diagram  
 illustrating this relationship, Evans and Frost  
 (1975 Fig 8 ) were able to estimate the positions  
 of two "isotherms" at 700 and 1 200°C. These were  
 based on natural chromite-olivine assemblages for  
 which equilibrium temperatures have been fairly  
 well established. The diagram is reproduced as Fig  
 (41) and it can be seen that spinel s.s. - olivine  
 from troctolite sample TV49 equilibrated at between  
 700 and 1200°C, whereas the hercynite-olivine pair  
 from the Kumkum gabbro norite and the picotite-  
 olivine and ferritchromite-olivine pairs from sample  
 TV49 all plot close to the 700°C "isotherm" on Fig  
 ( 41). These temperatures are of the right order  
 of magnitude when compared with those obtained using  
 the Jackson geothermometer.

### 5.3 The Stroh Geobarometer for coexisting olivine, orthopyroxene and spinel

The geothermometer of Stroh (1976) (Appendix 8, equation b) suffers from the disadvantage that it is calibrated using the experimental work of MacGregor (1974) for the system  $\text{MgO-SiO}_2\text{-Al}_2\text{O}_3$ . The effect of introducing additional components such as  $\text{FeO}$ ,  $\text{Fe}_2\text{O}_3$ ,  $\text{Cr}_2\text{O}_3$  and  $\text{CaO}$  to the system will cause a decrease in the pressures calculated using this method so that such pressures may only be considered nominal. Since the compositions of spinel, orthopyroxene and olivine from troctolite sample TV49 from Tantalite Valley most closely approximate those for the ideal system it was decided to test the Stroh geobarometer for this rock. A temperature of  $1049^\circ\text{C}$ , which was obtained from coexisting pyroxenes in a pyroxenite sample TV 6 129 by the method of Wood and Banno (1973), was assumed for the purposes of the calculation. This is not unreasonable because a maximum temperature of  $1205^\circ\text{C}$  was calculated from coexisting spinel sensu stricto and olivine from sample TV49 using the geothermometer of Jackson (1969). Using the analyses of spinel, olivine and bronzite from sample TV49 (Table 43 ) a pressure of 1,9 kb was obtained using equation (b) from Appendix ( 8 ) while the analyses of ferritchromite and picotite from the same rock, when utilized in this expression, gave negative pressures.



The explanation for this is that the ferritchromite and picotite contain large amounts of  $\text{Cr}_2\text{O}_3$ ,  $\text{FeO}$  and  $\text{Fe}_2\text{O}_3$  so that the equation (b) which works best for systems containing mainly  $\text{MgO}$ ,  $\text{SiO}_2$  and  $\text{Al}_2\text{O}_3$  breaks down. The test nominal pressure calculated, ie 1,9 kb, is also considered to be too low as a result of departures from the ideal system. From the presence in TV<sup>49</sup> of olivine-plagioclase coronas it can be inferred that pressures during the cooling of the rock were at least about 5 kb using the experimental work of Kushiro and Yoder (1966). The Stroh geobarometer may eventually prove useful when it is calibrated using experimental data on systems which include the common additional components found in nature but at present the pressures that are calculated can only be taken as minimum estimates.

#### 5.4 Garnet-Cordierite Geothermometry-Geobarometry

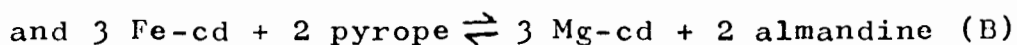
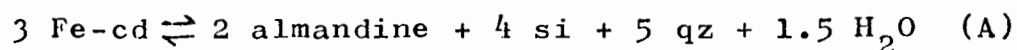
The divariant assemblage cordierite-garnet-sillimanite-quartz is commonly found in metapelites from intermediate pressure high-grade metamorphic terrains but may also occur in some thermal aureoles (eg Dostal 1975). The partition of iron and magnesium between garnet and cordierite in the above assemblage is especially suited to the determination of P and T of metamorphism since equilibria

can be written which involve large values of  $\Delta V^0$  and  $\Delta H_1$  bar, T respectively. It would be easy to obtain values for P and T for rocks containing the above divariant assemblage using calibrations derived from experimental data on pure and natural phases were it not for the conflicting results of various workers. For instance, on the basis of experimental studies on synthetic model pelitic compositions (Hensen and Green; 1971, 1972, 1973) and on studies involving the compositions of natural coexisting garnet and cordierite (Thompson 1976 b), it was shown that  $K_D^{\text{Ga-cd}}_{\text{Mg-Fe}^{2+}}$  decreases with increasing temperature. Currie (1971, 1974), however, using experimental and theoretical methods, proposed exactly the reverse. More recently Hensen (1977) was able to confirm that the findings of Thompson (1976 b) and Hensen and Green (1971, 1972, 1973) are correct. Uncertainties concerning the data on garnet and cordierite are also due to deviations from the ideal  $\text{FeO-MgO-Al}_2\text{O}_3\text{-SiO}_2$  system as a result of the presence of additional components, such as MnO and CaO in garnet and  $\text{H}_2\text{O}$  in cordierite, and by the presence of zoned minerals in nature.

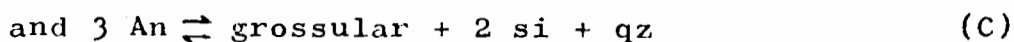
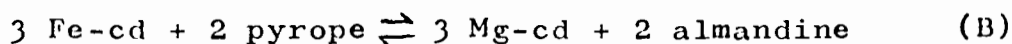
In the present study the temperature of metamorphism has been determined for a number of metapelite samples containing coexisting garnet, cordierite,

and biotite. This was done by using the graphs of Thompson (1976b) which relate  $K_D^{Ga-cd}$   $Mg-Fe^{2+}$  and  $K_D^{Ga-bi}$   $Mg-Fe^{2+}$  to T. Values obtained from the low temperature end of Thompson's curve relating  $K_D^{Ga-cd}$   $Mg-Fe^{2+}$  to T were increased by 30°C following the recommendations of Holdaway and Lee (1977).

Equilibrium curves for the reactions:-



were calculated for the garnet-cordierite pairs in Tantalite Valley hornfels sample T19, for various values of  $X_{H_2O}$  (ie 1, 0.6, 0.2) using the method outlined in appendix (9). The intersection of these curves gives an estimate of pressure in P.T space (Fig 61) which, however, varies according to the value of  $X_{H_2O}$  which is chosen. In order to circumvent this difficulty, pressure estimates were made from the intersection of two equilibria not involving a vapour phase, (Appendix 9, eqn iii and Appendix 10, eqn i) derived from the reactions:-



In this way, assuming that temperature is accurately known, it is also possible to estimate the value of  $X_{H_2O}$  from the intersection of the equilibria representing the reactions (A), (B) and (C).

In the present study grains of cordierite, garnet, biotite, plagioclase and, if present, potassium-feldspar, in close spatial association, were selected for analysis (Table 43 ). Of these minerals zoning was detected only in garnet, hence both cores and rims of garnet have been analysed. For the other minerals analyses of rims are regarded as being representative.

Biotite may be present both as light-brown, Ti-poor and dark-brown, Ti-rich flakes with all compositional gradations between these two extremes. In some rocks it has uniform composition throughout. A similar irregular variation of biotite composition was observed by Ashworth (1975) from garnet plus cordierite-bearing pelitic rocks from the Scottish Caledonides. He noted that garnet coexisting with biotite was zoned, with iron increasing and magnesium decreasing towards the rims. He suggested that the variations in garnet and biotite are due to retrograde diffusion reactions involving these minerals. In the present samples irregularity in

biotite composition is not matched by an irregularity in rim compositions in coexisting cordierite and garnet, although the garnet may be zoned. Thus, on account of the variability in biotite composition encountered in some rocks, temperatures derived from garnet-cordierite pairs are preferred to those obtained from coexisting garnet and biotite.

The results for determinations of T, P and estimated  $X_{H_2O}$  are given in Table (35) and graphically on Fig (42). Calculations were made using both garnet rim and core zones, and unzoned cordierite since the modal content of cordierite usually greatly exceeds that of garnet. <sup>(see App. 9)</sup> Thus, the cordierite will effectively act as an infinite reservoir for iron and magnesium during ionic migration. Where garnet is modally in excess of biotite (eg T19) the use of garnet cores and biotite (unzoned) in calculations is questionable, however trial calculations were carried out.

Two Black Gneiss samples A3 and A5 from the Sandfontein 131 area (Fig 3 ) give P 4,38-5,20kb, T 577-630°C and  $X_{H_2O} < 0,4$ . Sample A3 contains discrete cordierite and garnet whereas in A5 garnet has partly broken down to form cordierite along rims. Therefore it is probable that the values of P and T for A3 and A5 are representative of conditions prevailing during retrograde metamorphism.

Most samples from a number of different environments in the Kumkum area give a fairly narrow range of pressures (from 4,24 to 4,90 kb) taking both rim and core compositions into account. There is one low value, (3,96 kb, sample K35 P core) and one high value (6,10 kb, sample K66B). The low value is obtained from the core of a garnet porphyroblast which shows normal zoning and the high value is obtained from the rim of a garnet which shows slight reverse zoning. Both samples K35 P and K66B are hornfelsic varieties of Black Gneiss. The anomalous pressures calculated for these samples are probably due to the respective minerals not being in equilibrium with one another.

The garnet in K19A is partly replaced by cordierite which does not occur as a discrete phase in the rock. This texture has been attributed to a decrease in pressure (Toogood 1976). It is interesting to note that the pressure determined from K19A falls in the normal 4 to 5 kb range determined from most of the other samples and it is likely that these pressures are representative of those prevailing during retro-grade metamorphism (see also Moore 1975, Fig

13 ). The temperatures calculated for the Kumkum samples range from 610 - 760°C, which also probably reflects the temperatures of retrograde metamorphism for both the reasons given above and because co-existing pyroxene in mafic rocks in the area indicate that at the peak of metamorphism temperatures were approximately 800 to 900°C (section 5.1 ). The value of  $X_{H_2O}$  was certainly less than 1 and probably less than 0,5 (Table 35).

A pelitic hornfels from Tantalite Valley (sample T19) gives T 650°C, P 5,5 kb and  $X_{H_2O}$  about 0,6 (Fig 61 ) (see Moore et al in press) using the core zones of garnet, which is justified in this case because cordierite is modally in excess (see App. 9). The lower P and T obtained from garnet rims is interpreted as being due to re-equilibration of these rims during later retrograde metamorphic conditions. This conclusion is supported by the observation that staurolite is unstable in the contact aureole (Moore et al in press). Temperatures obtained from co-existing garnet and biotite using the calibration of Thompson (1976b) give a temperature of 620°C for biotite and garnet core zones, and 520°C for biotite and garnet rims. These results are in good agreement with the experimental

calibration of Ferry and Spear (1977) which, for an assumed pressure of 5,5 kb, give equivalent temperatures of  $623^{\circ}\text{C}$  and  $521^{\circ}\text{C}$ .

In conclusion, it seems that the metamorphic conditions deduced from the pelitic rocks within and north of the Tantalite Valley Lineament, (ie samples from Sandfontein 131, the Tantalite Valley Complex and the Kumkum area) represent a widespread, high-grade metamorphic event characterized by pressures of approximately 4 kb to 5 kb and temperatures of  $577$  to  $630^{\circ}\text{C}$  in the Sandfontein area, and increase northwards to between  $610$  and  $760^{\circ}\text{C}$  in the Kumkum area. This event (M2) followed an earlier high-grade event (M1) characterized by higher pressures and higher temperatures with the temperature also showing a tendency to increase northwards from the axis of the Tantalite Valley Lineament (see Toogood 1976). The Kumkum gabbro was emplaced earlier than the metagabbro of the Tantalite Valley Complex, since the former was emplaced during the M1 event whereas the Tantalite Valley metagabbro was intruded during or just after the M2 event. The value of  $X_{\text{H}_2\text{O}}$  in pelitic rocks was less than 1 for both events and probably less than 0,5. Since biotite is found with cordierite and almandine  $X_{\text{H}_2\text{O}}$  was unlikely to have been lower



than 0,4 (Holdaway and Lee 1977, p 196).

#### 5.5 Plagioclase-Garnet-Sillimanite-Quartz Geobarometer

Where plagioclase, garnet, aluminosilicate and quartz are in equilibrium in a rock it is possible to determine the pressure of metamorphism provided the temperature is known using the method of Ghent (1976) and Schmid and Wood (1976) (see Appendix 10).

The assemblage plagioclase-garnet-sillimanite-quartz is common in the Black Gneiss so that this geobarometer may be used. Zoning in garnets, when present, does pose a problem and the assumptions made in the previous section are valid here. Core compositions were used to estimate the P prevailing during the earlier metamorphism. The temperatures used in the calculations were obtained from co-existing cordierite and garnet and co-existing cordierite and biotite from the same samples using the calibrations of Thompson (1976b).

The temperatures derived from co-existing garnet-cordierite and garnet-biotite are usually significantly different, but both have been utilized in the calculations to give a range in P. The temperature obtained from garnet-cordierite pairs are

preferred for the reasons given in the previous section. Sample K101 which is a cordierite-garnet-gedrite rock from the Kumkum area contains rare sillimanite included in garnet, so that P estimates made for this rock are only tentative. Calculated pressures for various pelitic rocks in the area are listed in Table (36). For Black Gneiss samples from Sandfontein 131 calculated pressures are in the range 4,2 to 6,0 kb, depending on the temperature used and a similar range of 4,1 to 6,1 kb is found for the Black Gneiss samples from the Kumkum area.

The interpretation of results from hornfelsic Black Gneiss sample K35, which was collected from along the contact between the main plug of Kumkum gabbro-norite and the Black Gneiss, is complicated by the presence of both porphyroblastic and groundmass garnets, with the former showing normal zoning. The porphyroblast cores give a pressure range of 3,8 to 5,4 kb whereas the rims of porphyroblasts and groundmass garnets were equilibrated at a pressure of between 4,3 and 6,5 kb. Black Gneiss sample K66B is unusual in that the garnets are flesh-coloured in place of the normal pink and occur as porphyroblasts in a fine-grained matrix. The rock was intruded by Inequigranular granite gneiss and may be termed a hornfelsic variety for

the purposes of this discussion.

Garnets from the Pelitic hornfels (sample T19) from Tantalite Valley give pressures ranging from 4,8 kb for the core zones to 5,3 kb for the rims, with temperatures decreasing outwards. The rims here have been affected by re-equilibration as a result of retrograde metamorphism and it seems unrealistic that pressures would have increased during the latter event particularly in view of the conclusions reached concerning breakdown of staurolite in the aureole, which probably took place at pressure of around 2,5 kb. If the intersection of the garnet-cordierite exchange reaction and the plagioclase-garnet- $\text{Al}_2\text{SiO}_5$ -quartz reaction is used to obtain the pressure at which the cores and rims of T19 equilibrated, then it is indeed found that the cores formed at higher pressures than the rims.

#### 5.6 Plagioclase-Alkali Feldspar Geothermometer

This geothermometer originally proposed by Barth (1951) is based on the distribution of albite-component between plagioclase and K-feldspar. Stormer (1975) derived an expression relating T, P and composition of coexisting feldspars using

modern thermodynamic data. He produced a series of isotherms for pressures of 1, 2, 5 and 10 kb in order to solve this rather complex equation (Stormer 1975; equation 18).

Since coexisting plagioclase and microcline occur in several of the Black Gneiss samples examined, it was decided to check temperatures obtained by the method of Thompson (1976 b) for coexisting garnet, cordierite and biotite against those derived using Stormer's geothermometer assuming a pressure of 5 kb. The results are given in Table ( 37 ).

In general the temperatures fall within the range of those calculated using the garnet-cordierite geothermometer of Thompson (1976 b) although sample K66B gives temperatures which are apparently too low. This discrepancy may be due to exsolution having taken place in the microcline analysed. Generally, however, perthite-free microcline was chosen for analysis.

The advantage of the Stormer geothermometer is that it places a constraint on the temperature of metamorphism which may have a wide range if the garnet-cordierite and garnet-biotite geothermometers of Thompson (1976 b) are used. Powell and Powell (1977) have criticised the Stormer geothermometer because it does not take into account the Ca-content

of K-feldspar. If Ca is present in significant amounts then the temperatures derived using this method will be too high. Fortunately, the Ca-content of the microcline from the samples examined is too low to make any difference to the temperatures given in Table ( 37 ).

#### 5.7 Iron-Titanium Oxide Geothermometry-Geobarometry applied to Black Gneiss

Buddington and Lindsley (1964) showed that it is possible to determine the temperature and  $f_{O_2}$  of formation of co-existing pairs of titaniferous magnetite and ilmenite in a variety of rocks.

In the Black Gneiss from the present study area, magnetite and ilmenite commonly occur together so that the method of Buddington and Lindsley (1964) can be applied. In practice it was found difficult to obtain reliable electron microprobe analyses for these opaque minerals. This may be due to the occasional presence of fine ilmenite lamellae in magnetite, which are revealed by qualitative scans. When magnetite is free of these features it is characterized by a complete lack of Ti (Table 43). The ilmenite, unlike magnetite, is always homogeneous and may have up to 6% by weight haematite in solid

solution (Table 43 ). Since the magnetite is titanium-free, compositional contours for co-existing magnetite and ilmenite on Fig 5 of Buddington and Lindsley (1964) will be essentially parallel so that there will be no unique solution for  $f_{O_2}$  and T. Buddington and Lindsley (1964) showed that textures similar to those described here, ie ilmenite lamellae in magnetite, are formed in metamorphic rocks during cooling under oxidizing conditions. This could be interpreted as oxidation spinodal decomposition of magnetite. These conditions may have been met in the present area, when rocks which had been metamorphosed at the peak of metamorphism (M1) subsequently underwent cooling and some hydration during the retrogressive metamorphism event, (M2). A corollary of this is that the first metamorphic event involved more reducing conditions than the second, although  $f_{O_2}$  in the Black Gneisses must have been low during both events, as can be seen by the presence of almandine and hercynite (see Section 7.2.3).

## CHAPTER VI

## MINERAL CHEMISTRY AND METAMORPHISM

6.1 Olivines

The CaO content of olivine from mafic and ultramafic rocks from the Tantalite Valley-Kumkum area is less than 0,10 weight percent (Table 43 ) which is typical for those from plutonic environments (Simkin and Smith 1970). They are virtually unzoned and range in composition from  $Fe_{66}$  in the gabbroic rocks to  $Fe_{82}$  in ultramafic rocks (Fig 43 ).

Oriented plates of an opaque mineral (either chromite or magnetite/haematite) occur as inclusions in olivine from troctolite samples TV45 and TV36. This texture is interpreted as having formed by exsolution of excess chrome or iron from the olivine lattice on cooling from magmatic temperatures. Goode (1970) described the same texture for iron-rich olivine from the Kalka intrusion, Australia.

The composition of olivine can be used to determine the magnesium to ferrous iron ratio of the liquid from which it crystallized by the method of Roeder and Emslie (1970). Assuming  $K_D^{ol-liq} Fe^{2+}-Mg$  is equal to 0,30 for olivine tholeiite and the average composition of olivine from the ultramafic rocks and the gabbro-norite of the Tantalite Valley Complex

is about  $\text{Fo}_{82}$  and  $\text{Fo}_{77}$  respectively, gives the result that  $X_{\text{MgO}}^{\text{liq}} / X_{\text{FeO}}^{\text{liq}} = 1,37$  for the ultramafic rocks and  $X_{\text{MgO}}^{\text{liq}} / X_{\text{FeO}}^{\text{liq}} = 1,00$  for olivine gabbro-norite.

The most magnesium-rich olivine found in the Skaergaard has a composition of  $\text{Fo}_{81}$  (Wager and Brown 1968, p 38), which according to Roeder and Emslie (1970) must have crystallized from a liquid with  $X_{\text{MgO}}^{\text{liq}} / X_{\text{FeO}}^{\text{liq}} = 1,28$ . Likewise using the most magnesium-rich olivine composition from the Bushveld Igneous Complex, ie  $\text{Fo}_{88}$  (Wager and Brown 1968, p 388),  $X_{\text{MgO}}^{\text{liq}} / X_{\text{FeO}}^{\text{liq}}$  from which this olivine crystallized was equal to 2,20. Olivine phenocrysts in kimberlite, taken as representative of mantle olivine, was found by Boyd and Clement (1977) to have an average core composition of  $\text{Fo}_{90}$ . The liquid from which it crystallized must have had a  $X_{\text{MgO}}^{\text{liq}} / X_{\text{FeO}}^{\text{liq}}$  ratio of 2,70. From the above discussion, it is clear that the liquid from which the olivine found in ultramafic rocks of Tantalite Valley formed was less magnesian than mantle-derived and Bushveld liquids and more magnesian than Skaergaard liquids. Thus the olivine in the Tantalite Valley ultramafic rocks must have been derived from a liquid which had undergone some differentiation after its formation in the mantle.



Kesson (1973) defined the quantity Mg Number ie  $\left[ 100 \text{ Mg} / (\text{Mg} + \text{Fe}^{2+}) \text{ atomic ratio} \right]$ . For primary basaltic magmas in equilibrium with upper mantle source regions (approximated by upper mantle olivine compositions ie  $\text{Fo}_{87.5}$  to  $\text{Fo}_{92}$ ) Mg Numbers range from 67 - 68 to 78. The Mg Number obtained from the olivine ( $\text{Fo}_{82}$ ) in ultramafic rocks from Tantalite Valley is 59 which supports the conclusion reached in the previous paragraph that this olivine could not have crystallized from a primary basaltic magma.

## 6.2 Pyroxenes

### 6.2.1 Pyroxene trends

Trends obtained from plotting pyroxene compositions in the Di-Hd-En-Fs quadrilateral are particularly useful in establishing the extent and continuity of fractionation within an intrusion, and the character of the various fractionation stages in terms of co-existing pyroxene assemblages. These compositions are expressed in terms of molecular proportions of Wo, En and Fs ( $\text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mn}$ ). Tholeiitic differentiates are distinguished from alkali-basalt differentiates by the presence of co-existing calcium-rich and calcium-poor pyroxenes in the former in contrast to only calcium-poor

pyroxene being found in the latter. Generally in tholeiites Fe enters both Ca-rich and Ca-poor pyroxenes in increasing amounts relative to Mg with fractionation and Ca first decreases to a minimum and then increases with fractionation in the Ca-rich pyroxenes (eg Skaergaard and Bushveld trends).

*Point analyses of*

Ca-rich and Ca-poor pyroxenes from a number of different rock types from the present study area have been plotted in Fig (43). <sup>*These have all undergone some degree of exsolution*</sup> The crossing tie-lines in the diagram are probably due to incomplete subsolidus re-equilibration having taken place between cumulus or intercumulus orthopyroxene and clinopyroxene during the cooling of the rocks. The unaltered igneous rocks have clinopyroxenes which show a general trend of decreasing Ca-content with increasing iron, whereas orthopyroxenes show an apparent trend parallel to the En-Fs join ie Ca is constant as Fe increases. The anomalously high Ca-content of discrete orthopyroxene from TV6 129 (ie 2A in Fig 43) is due to the influence of clinopyroxene lamellae on the electron microprobe analyses.

Pigeonite is absent.

Pyroxenes from granulite, Contaminated gabbro-norite and Black Gneiss are also shown in Fig (43) for comparison with igneous pyroxenes. The metamorphic pyroxenes are seen to be enriched in iron relative

to igneous pyroxenes.

Generally the immiscibility gap between Ca-rich and Ca-poor pyroxenes appears greater for pyroxenes from the present study area than for those from the Skaergaard over a corresponding range of pyroxene compositions. This may be explained by depression of the solidus surface relative to the solvus surface in the pyroxene system, causing the immiscibility gap between co-existing ortho- and clino-pyroxene for the former to be wider compared with Skaergaard pyroxenes of equivalent composition (Atkins 1969). In general the compositions of pyroxenes from the igneous rocks of the Kumkum and Tantalite Valley areas closely resemble those from the Bushveld Igneous Complex, as can be seen by comparing analyses in Table ( 43 ) with those given by Atkins (1969) and Markgraaff (1976).

The Bushveld Igneous Complex was emplaced at a greater pressure and lower temperature than the Skaergaard (Atkins 1969) and for this reason, pyroxenes from the former are more closely comparable to those from Kumkum and Tantalite Valley. The depression of the solidus surface in the pyroxene system is thought by Atkins (1969) to be due to lower temperatures, (as a result of higher  $P_{H_2O}$ ) or higher pressure operating during pyroxene crystallization. Evidence for a fairly high  $P_{H_2O}$

in the Tantalite Valley and Kumkum unaltered igneous rocks can be seen by the presence of some dark brown magmatic hornblende and phlogopite or biotite. The presence of olivine-plagioclase coronas in these rocks indicates that the rocks cooled at pressures of about 5 kb which is about 5 times greater than that envisaged for the Skaergaard intrusion.

Another possibility is that exsolution effects shift the composition of the host and lamellae phases to more Ca-rich compositions on the clinopyroxene side and to more Ca-poor compositions on the orthopyroxene side in the pyroxene quadrilateral (Boyd and Brown 1967). Slower cooling would be expected in rocks of the Bushveld Igneous Complex and Tantalite Valley - Kumkum areas compared to those of the Skaergaard due to the greater depth at which the former were emplaced. Consequently, the two pyroxene field is expanded in the Bushveld and Tantalite Valley rocks relative to those of the Skaergaard and exsolution, which is more pronounced in the former than in the latter will cause a shift away from the Skaergaard trend as observed.

Further, the Skaergaard trend is based mainly on wet chemical analyses which do not distinguish between discrete and lamellae phases; thus clinopyroxene analyses will be diluted by a certain amount of orthopyroxene lamellae and vice versa. However, the part played by exsolution in changing the width of the pyroxene immiscibility gap is insignificant compared with factors like P and T. The lower temperatures and higher pressures prevailing during the emplacement of the Tantalite Valley - Kumkum igneous rocks and the Bushveld Complex compared with the Skaergaard Intrusion can explain the observed pyroxene trends.

#### 6.2.2 Pyroxene End-Members

Wilson (1976) has shown that for the purpose of detecting differences in physical conditions under which pyroxenes have equilibrated, hypothetical pyroxene end-members are preferable to plots involving  $Al^{IV}$  and  $Al^{VI}$  contents in pyroxenes.

Various schemes exist for the calculation of

pyroxene end members, but in most it is necessary to accurately know the  $\text{Fe}_2\text{O}_3$  content of the pyroxene which cannot be achieved using the electron microprobe. Even when determined using wet chemical methods, the result may not give the original oxidation state of the primary pyroxene. Other schemes like that of Essene and Fyfe (1967) do not place any great emphasis on  $\text{Fe}_2\text{O}_3$  and ignore  $\text{TiO}_2$  and  $\text{K}_2\text{O}$  content of pyroxene, making them of limited use.

Cawthorn and Collerson (1974) evolved a scheme which is suited to calculating pyroxene end-members from electron microprobe analyses and this is the method used here, with modifications for orthopyroxene as suggested by Wilson (1976). The  $\text{Fe}_2\text{O}_3$  content of the pyroxenes have been estimated by assuming ideal stoichiometry even though the effect of unknown  $\text{Fe}_2\text{O}_3$  content on the Cawthorn-Collerson method is minimal. Calculated end-members for ortho- and clino-pyroxenes from the study area are listed in Tables 38 and 39. Jadeite and Ca-tschermak's molecule are the most important end members in clinopyroxenes for estimating

physical conditions (Cawthorn and Collerson 1974).

Both increase on increasing pressure, (Kushiro 1962; Aoki 1964), but since the amount of each is dependent on the iron content of the pyroxene, pyroxenes in different rocks must have approximately equal ferrosilite contents if comparisons between rocks are to be made (Wilson 1976). This effect is probably more important than pressure in giving the trend observed in Fig ( 45 ) where jadeite and ferrosilite in clinopyroxene are plotted against each other.

It could be argued that ultramafic rocks and rocks of the satellite body of Tantalite Valley as well as the unaltered igneous gabbro-norites of the Kumkum area formed at a higher pressure than the gabbro-norites of Tantalite Valley because pyroxenes from these rocks have higher jadeite contents than those found in the Tantalite Valley gabbro-norites. However, the distinct negative correlation observed points to a chemical rather than a physical control. A similar negative correlation can be seen when Mg tschermakite + Ca tschermakite (a pressure sensitive index) in orthopyroxene is plotted against

Ferrosilite in orthopyroxene (Fig 46 ). This conclusion is supported by the observation that some samples which appear to indicate a higher pressure in Fig (45) show the reverse tendency in Fig (46). If jadeite is plotted against Ca-tschermak for clinopyroxene Fig(44) an overall positive correlation would be expected if pressure was the controlling factor since both end-members increase on increasing pressure. However, what is found is a completely random spread of points. Tantalite Valley gabbro-norite sample AM41 is unusual in that it has the lowest jadeite content and the lowest Mg-tschermakite plus Ca-tschermakite content of all the clino- and *ortho*-pyroxenes respectively. This sample is fine-grained and may belong to the dyke component of the gabbro-norite. The pyroxenes from this rock have the highest Fe-contents of any analysed from Tantalite Valley (Fig 43 ). Further, the fact that the pyroxenes from AM41 plot at the extremes of the trends in Fig ( 45 ) and ( 46 ) indicates that this rock may indeed be a late-stage phase of the gabbro-norite.

From the above it would appear that Na and  $Al^{VI}$  in the clinopyroxenes and total aluminium content of orthopyroxene (see Wilson 1976) increases with increasing Mg content. This is in agreement with



Wilson (1976) who found that there is a close correlation between  $(\text{MgO} + \text{FeO total} + \text{MnO})$  in weight percent and  $\text{Al}_2\text{O}_3$  in orthopyroxenes and clinopyroxenes. Further it is unlikely that the gabbro-norites and ultramafic rocks of Kumkum and Tantalite Valley formed at very different pressures because of the presence of similar olivine-plagioclase reaction coronas in all of them (section 4.1). However, it is to be stressed that the Tantalite Valley rocks were emplaced at a later time than the Kumkum intrusives into rocks which had cooled down considerably from the high temperature which had prevailed during the intrusion of the Kumkum rocks. During this time pressure must have remained essentially constant.

#### 6.2.3 Clinopyroxene and Igneous Parentage

Kushiro (1960) and Le Bas (1962) have shown that the aluminium content of a clinopyroxene from basaltic rocks may be used as an aid in determining the magmatic parentage of the host rock. Trial plots were carried out for igneous clinopyroxenes found in mafic and ultramafic rocks from the Kumkum and Tantalite Valley areas, bearing in mind the limitations when applying this technique to plutonic rocks. The clinopyroxenes must be soda-poor and fall in the range  $\text{Ca}_{35-53} \text{Mg Fe}_0 - 18$  before they

can be tested in the diagrams devised by Le Bas (1962). These conditions are met for all analysed clinopyroxenes of igneous origin from the study area.

If  $(\text{Ca} + \text{Na} + \text{K})$ , Mg and  $(\text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mn})$  in atomic percent are plotted, then three fields are established by the clinopyroxenes according to the host-magma type (see Le Bas' definition), ie Peralkaline (Field I), normal alkaline (Field II) and nonalkaline (Field III), which includes the tholeiitic, high alumina and calcalkaline series. Clinopyroxenes from rocks studied here fall into fields II and III on such a diagram (Fig 47).

A plot of  $\text{SiO}_2$  against  $\text{Al}_2\text{O}_3$  reveals a similar separation into three fields. When applied to the present samples clinopyroxenes plot in the non-alkaline field (Fig 48). Generally there is a negative correlation between the two elements (see also Le Bas 1962 and Kushiro 1960) and the  $\text{Al}_2\text{O}_3/\text{SiO}_2$  ratio varies in sympathy with that of the host rock (Le Bas 1962).

The proportion of Al in the tetrahedral site in clinopyroxene ( $\text{Al}_z$ ) is about 70 percent of total Al in clinopyroxene from non-alkaline rocks, which is what is found for the samples studied here (Table 43). In the case of normal-alkaline and peralkaline series 90 and 100 percent respectively

of the Al is in the tetrahedral site. Further, the amount of total Al in clinopyroxene closely reflects the degree of silica saturation of the host magma (Le Bas 1962).

Clinopyroxenes from non-alkaline rocks are characterised by less than 1 percent  $\text{TiO}_2$  and this is also found for all the samples examined here (Table 40 ). Further, if  $\text{Al}_Z$  is plotted against  $\text{TiO}_2$  they fall into the field of non-alkaline rocks.

A disadvantage of the Le Bas method is that it cannot distinguish between clinopyroxenes from calcalkaline, tholeiitic and high alumina rocks and groups them all as non-alkaline. Kennedy (1955) and Osborn (1959) have suggested that a high  $\text{Fe}_2\text{O}_3/\text{FeO}$  ratio (as determined from pyroxenes and/or host rock) distinguishes the calc-alkaline series from the tholeiitic and high alumina series which both have low ratios.

The  $\text{Fe}_2\text{O}_3/\text{FeO}$  ratio for analysed orthopyroxenes and clinopyroxenes from the present study area are normally much less than 0,58. A further drawback is that the amount of  $\text{Al}^{\text{IV}}$  in clinopyroxene can be decreased by increasing load pressure.

From the criteria given above it is concluded that the host magma-type from which the clinopyroxenes formed was probably non-alkaline. Further, on the

basis of  $\text{Fe}_2\text{O}_3/\text{FeO}$  ratios in pyroxenes it appears that the field can be further narrowed down to tholeiites or high alumina series as being the most likely host magma type. The anomalously high (Ca + Na + K) content of these clinopyroxenes (Fig 47 ) may be due to the ability of plutonic clinopyroxene to incorporate more Ca into its lattice as a result of the greater pressure prevailing compared to the volcanic environment.

#### 6.2.4 The distribution co-efficient $K_D^{\text{opx-cpx}}_{\text{Mg-Fe}}$

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$K_D^{\text{opx-cpx}}_{\text{Mg-Fe}}$  (as defined in Table 41 and hereafter referred to as  $K_D$ ) is dependent on temperature, pressure and solid solution constituents other than Mg and  $\text{Fe}^{2+}$  (Ramberg and De Vore 1951; Kretz 1961, 1963). It is generally found that igneous two pyroxene assemblages have  $K_D \cong 0,73$  while metamorphic ones have a  $K_D \cong 0,57$  (Kretz 1961) illustrating the different physico-chemical conditions prevailing in these two environments. In the co-existing pyroxenes examined from the Tantalite Valley Complex the  $K_D$  shows a large spread of values from 0,61 to 1,04 (Table 41 ). The large range in  $K_D$  values from 0,38 to 0,84 calculated using analyses from norite sample TV6 129 (Table 41 ) is a function of the difficulty in

analyzing lamellae in pyroxenes with the electron microprobe (see section 4.6 ) and the conclusions drawn from these results are tentative. Lamellae would be expected to equilibrate readily with its host and the  $K_D$  values obtained for discrete and exsolved phases ie TV6 129A and B respectively (Table 41 ) should be similar. However, as is seen in Table (41) they are not.

The  $K_D$  range is 0,38 to 0,48 in samples of granulite, Contaminated gabbro-norite, Eselruh gabbro-norite and Kumkum gabbro-norite from the Kumkum area (Table 41 ).

Therefore the  $K_D$  values of the Tantalite Valley samples have an igneous affinity whereas those from the Kumkum area have a metamorphic one.

It is unlikely that the difference in  $K_D$  values between the two areas is due to compositional factors because the rocks are on the whole texturally, mineralogically and chemically similar.

The pressure in both areas was probably very similar as is indicated by the presence of coronas between olivine and plagioclase. That temperature was dominant in controlling the  $K_D$  values can be seen by comparing the temperatures calculated from co-existing pyroxenes from Kumkum and Tantalite Valley. In general, it can be said that the  $K_D$

value probably increases with increasing temperature and varies according to the degree of subsolidus re-equilibration which the rock has experienced. The large range in  $K_D$  values in the Tantalite Valley rocks is probably due to incomplete re-equilibration as a result of rapid freezing of the system in contrast with the older Kumkum rocks where re-equilibration occurred over a long period during high grade metamorphism.

Atkins (1969) proposed that the lower  $K_D$  values of pyroxenes from the eastern Bushveld Complex compared to Skaergaard pyroxenes is due mainly to the greater depth at which the Bushveld rocks were emplaced ie the  $K_D$  value decreases as the pressure increases (see also Kretz 1963). He calculated that the crystallization temperatures for the Bushveld Complex are about 200°C lower than that for the Skaergaard Intrusion but did not consider that such a large difference was likely due to the similarity in primary magma composition and mineralogy of the two intrusions. Naldrett and Kullerud (1967) found that  $K_D$  values decrease with increasing iron-enrichment of pyroxenes (ie decreasing temperatures). Atkins (1969) however, found that  $K_D$  is constant over much of the differentiation sequence in the eastern Bushveld Complex and ascribed this to decreasing pressure acting on more differen-

tiated rocks occurring higher in the magma chamber compensating for the effect of decreasing temperature. Crystallization of pyroxenes under high pressures can give rise to high  $K_D$  values (Moore 1971) or low  $K_D$  values (Hossain 1977) for high  $R_2O_3$  contents.

### 6.3 Plagioclase Composition in Mafic Rocks

It is well known that the anorthite content of plagioclase tends to increase with increasing metamorphic grade <sup>providing that there is a co-existing Ca-rich phase present</sup> (Deer et al 1971, p 336). For plagioclase from metabasites in the present study area there is no well-defined trend of increasing An-content on going from the Tantalite Valley to the Kumkum areas, ie with increasing grade of metamorphism (see Toogood 1976). Both areas are in fact characterized by a wide-range of plagioclase compositions from  $\sim An_{20}$  to  $\sim An_{90}$ . Compositions determined with the microprobe on plagioclase in mafic rocks from the Kumkum and Tantalite Valley areas are represented in an An-Ab-Or ternary diagram (Fig 49). The maximum range in An-content in plagioclase from a single rock is about 15 percent with an average variation of about 4%. This variation is due to zoning and difference in composition between groundmass and megacrystic plagioclase. From Fig (49) it can be

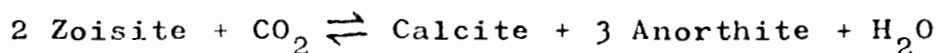
seen that there is no tendency for plagioclases from metabasites from both areas to separate out into two distinct groups. All the plagioclase analyses are characterised by Or less than 1% over the range  $An_{95}$  to  $An_{47}$ .

The range of plagioclase compositions in the area in general indicates grades of metamorphism from the lower part of the amphibolite facies to the granulite facies. No co-existing Ca-poor and Ca-enriched plagioclase, characteristic of the transition from the greenschist to the amphibolite facies (eg Cooper 1972; Winkler 1974), was observed. Retrograde metamorphism of plagioclase to albite, sericite and epidote, is however, ubiquitous and is characteristic of the greenschist facies. The unusually high An content of plagioclase from amphibolite sample AM76 (Table 43) needs explanation.

This sample was collected from the core of the Tantalite Valley Shear Zone where most amphibolites contain plagioclase of intermediate composition, which is in keeping with the medium grade metamorphism postulated for this domain. The composition of the plagioclase in sample AM76 is therefore due to some influence other than increased grade of metamorphism. Local increases in  $X_{CO_2}$  at constant T and  $P_f$  can lead to reactions

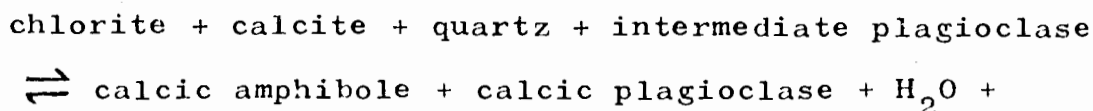


which produce increases in anorthite-component in plagioclase. A possible reaction is:-



(Winkler 1974, Fig 10-2)

However, the absence of zoisite and calcite in sample AM76 is a problem. It is possible that zoisite reacted completely and that calcite has been removed from the rock, since veins of calcite do occur. Another likely reaction is:-



$\text{CO}_2$  (Ferry 1976), which occurs at the amphibolite grade of metamorphism. However, none of the reactant phases were identified in sample AM76 so that if this reaction did occur then the metamorphic conditions needed for reaction must have been exceeded.

#### 6.4 Calcic Amphiboles

##### 6.4.1 Classification and Structure

Calcic amphiboles include all those amphiboles which have  $\text{Ca} = 1,5 - 2,0$  cations per 23 oxygen anions and are best classified using the scheme of Leake (1968) which is given here as Fig (50). Generally, the names for individual samples are

taken from that of the field into which the analysis plots on Leake's diagram (Fig 50 ). He distinguished two main groups of calcic amphiboles, ie those with ionic proportions, on the basis of 23 oxygens, of  $(Ca + Na + K)$  greater than 2,5 and those with  $(Ca + Na + K)$  less than 2,5. These he called calcic and sub-calcic amphiboles respectively. In the Tantalite Valley area all analyzed amphiboles with  $Ca = 1,5 - 2,0$  plot in the field of sub-calcic amphiboles with the exception of that from amphibolite sample AM76. The  $Mg/(Mg + Fe)$  ratio of these samples is equal to or greater than 0,5, and increases with increasing Si content (Fig 50 ). In contrast, all the analyzed amphiboles from the Kumkum Massif with  $Ca = 1,5 - 2,0$  including sample AM76, plot in the field of calcic amphiboles (Fig 50 ) with  $Mg/(Mg + Fe)$  mostly greater than 0,5.

There is no systematic variation of the ratio  $Mg/(Mg + Fe)$  with metamorphic grade in calcic amphiboles and the latter cannot be used as a grade indicator (Binns 1965; Leake 1968; Jackson 1976).

The structure of calcic amphibole can be represented as  $A_{0-1}X_2Y_5Z_8O_{23}$  (ignoring water and fluorine content, a condition which is necessary in recalculating amphibole analyses obtained from the electron microprobe) where  $A = Na, K, Ca$ ;  $X = Ca, Na$ ;  $Y = Mg, Fe^{2+}, Fe^{3+}, Mn, Al^{VI}, Ti^{4+}$  and  $Z = Si, Al^{IV}$ .

(Kisch and Warnars 1969).

#### 6.4.2 Edenitic substitution

Increasing occupation of the A site by alkalis with rising grade of metamorphism is a well-defined trend, particularly in low pressure metamorphic terranes (Shido 1958; Engel and Engel 1962; Binns 1965; Bard 1970; Jackson 1976). Generally the accommodation of sodium and potassium in the vacant A site of hornblende is known as edenitic substitution but Binns (1965) emphasized that only the alkali enrichment in the A site ie 'edenitic alkalis' defined as  $E = (Na + K) - \left\{ (2 - [(\sum Y - 5) + Ca]) \times 2 \right\}$  is controlled by the metamorphic grade. This is because alkali cations can be divided into those which fill, one for one, the X and A sites of the hornblende structure ('richterite alkalis') and those additional cations that enter the A site and produce an overall charge excess ('edenite alkalis'). for the calcic amphiboles from the Kumkum and Tantalite Valley areas, the overflow from the Y site and Ca go into the X site and this generates excess Ca which together with Na and K go into the A site. Thus all of the alkalis in these amphiboles are edenitic and the total (Na + K) is a gauge of the degree of edenitic substitution. Furthermore, the Na content of calcic amphiboles studied were shown to be insensitive to

the Na content of co-existing plagioclase using a triangular plot of  $\text{Na}_2\text{O}$ ,  $\text{CaO}$  and  $(\text{MgO} + \text{FeO} + \text{MnO})$  in the manner of Robinson and Jaffe (1969). This proved that the amount of Na found in the hornblende is not dependent on the slow breakdown of the albite component of plagioclase on going to higher grades. Theoretically, the generation of edenite-component (plus anorthite plus quartz) from richterite and tremolite involves a temperature increase (corresponding to the transition from the greenschist to amphibole facies) and a volume increase of about 7%, suggesting that the prevailing hydrostatic pressure must have been relatively low for edenite to form. (See Bard 1970). The latter may explain why edenitic alkalis are commonly found in hornblendes from low P facies series metamorphic terranes.

In (Fig 51), after Miyashiro (1973), the trend of increasing edenitic substitution is parallel to the  $(\text{Ca} + \text{Na} + \text{K})$  axis towards higher values of  $(\text{Ca} + \text{Na} + \text{K})$ . Ca is however independent of metamorphic grade and only Na + K need be considered in evaluating edenitic substitution. The Tantalite Valley samples have an average  $(\text{Ca} + \text{Na} + \text{K})$  of 2,2. Metagabbro sample AM88E is unusual in that it alone plots in the actinolite field and shows the lowest edenitic substitution (ie lowest grade of metamorphism) of all. The rock was collected from along

the sheared contact of metagabbro and ultramafic rock. Since shear zones are loci of retrograde metamorphism in the area, the behaviour of this sample is predictable. Hornblendes from the metasomatic rocks of Tantalite Valley, including the mottled metagabbro and the Metasomatised hornblende, have a  $(Ca + Na + K)$  value of around 2,4. There is in general a distinct hiatus between the  $(Ca + Na + K)$  values of Tantalite Valley samples and those from the Kumkum area, and it is inferred that higher grades of metamorphism were realized in the latter area compared with the former. The anomalous  $(Ca + Na + K)$  content of hornblende from amphibolite sample AM76 may be due to the fact that this rock contains the highest CaO content of all the samples from which hornblendes were analysed. Hornblende in AM76 also contains the highest Ca of any amphiboles examined. This is supported by Bard (1970) who found that Ca-content of amphibole is controlled by the bulk chemistry of the rock. He found that the sum of the alkalis in hornblende is however grade dependent. Thus the anomalously high  $(Ca + Na + K)$  content of hornblende from AM76 must be due to the high Ca content of the rock, since the grade of metamorphism at which this rock formed was no different from the others examined from Tantalite Valley.

### 6.4.3 Tschermakitite Substitution

Al substituting for Si in the Z-sites of calcic amphiboles with increasing metamorphic grade is known as tschermakitite substitution. The overall effect of this is to increase  $\text{Al}^{\text{IV}}$  and decrease  $\text{Al}^{\text{VI}}$ . The above replacement is coupled by increased entry of alkalis into the A-site giving essentially a progression from tschermakitite hornblende (lower amphibolite facies) towards edenite-pargasite varieties (granulite facies) (Raase 1974). The degree of tschermakitite substitution in calcic amphiboles is shown along the Si-axis of Fig (52) after Miyashiro (1973) with the lowest Si values indicating the highest grades of metamorphism. The Tantalite Valley samples have an average Si value of 6,6 as distinct from the Kumkum samples which have Si less than 6,4, indicating that the former were formed under lower grade metamorphic conditions than the latter. This is in agreement with the conclusion reached from edenitic substitution trends. The slight overlap in Si values for hornblendes from the Kumkum and Tantalite Valley areas could be explained by the fact that although  $\text{Al}^{\text{IV}}$  increases on increasing metamorphism, this may be offset to some extent by  $\text{Al}^{\text{VI}}$  which increases with increasing pressure (eg Leake 1965a; Bard 1970). Once more, metagabbro sample AM88E is found to be distinctly

different from the other samples and an origin for it linked with retrograde metamorphism is favoured. Leake (1965<sup>b</sup>) showed that metamorphic hornblendes have higher Si than igneous hornblendes and devised a plot of Ti versus Si to distinguish between hornblendes formed in these different environments. From Fig ( 53 ) it can be seen that all the calcic amphiboles, with the exception of samples AM41 from the Tantalite Valley area and K37A (groundmass) and possibly K4 (groundmass) from the Kumkum area, are metamorphic in origin. It is interesting to note that, as expected, hornblendes formed as a result of the reaction between olivine and plagioclase in the unaltered gabbro-norites of the Kumkum area ie samples K37A and K4 plot as metamorphic, whereas the corresponding groundmass equivalents plot as igneous in Fig (53).

Raase (1974) showed on an  $Al^{VI}$  versus Si plot that hornblendes from low-pressure and medium-high pressure facies series terranes are clearly separated out, with a line at 5 kb signifying the cross-over point between the two different pressure environments. All the analyzed samples from Kumkum and Tantalite Valley plot on the 5 kb line or on the low-pressure side of it. (Fig 52 ). From this it is concluded that all hornblendes from the present area formed under conditions of

low-pressure facies series metamorphism and that in the increase in grade from the Tantalite Valley to the Kumkum area, pressure was not an important controlling factor but rather temperature.

#### 6.4.4 Titanium content of Hornblende

It is now generally accepted that Ti content of calcic amphibole increases with increasing grade of metamorphism and is independent of the titanium content of the host rock if the latter has Ti in excess, ie sphene and/or ilmenite are present (eg Engel and Engel 1962; Leake 1965<sup>b</sup>; Binns 1965). Furthermore, there is a close correlation between the  $\gamma$  absorption colour of hornblende and its Ti content with the former varying from blue-green to green to brown-green to brown with increasing grade (eg Engel and Engel 1962; Binns 1965; Leake 1965<sup>b</sup>; Raase 1974; Jackson 1976). High Ti and low  $\text{Fe}^{3+}$  gives  $\gamma$  red-brown in hornblende whereas the reverse situation favours  $\gamma$  blue-green. (Binns 1965). It is probable therefore that  $\text{Fe}^{3+}$  decreases in hornblende with increasing metamorphism. This is substantiated by the fact that the amount of  $\text{Fe}^{3+}$  in hornblende depends on the host rock chemistry which reflects oxidation conditions during metamorphism and it is well known that the  $\text{Fe}_2\text{O}_3/\text{FeO}$  ratio decreases in a rock on increasing grade of meta-



morphism.

The average Ti content of Tantalite Valley metamorphic hornblendes ( $\sim 0,04$ ) places them into the transition zone between greenschist and amphibolite facies (see Raase 1974). In the rocks from which these samples were analyzed, sphene and/or ilmenite is present, indicating that the rock is saturated with titanium. Vejnar (1975) showed that where the latter is not the case, Ti may be preferentially partitioned into biotite if the latter co-exists with hornblende and that the Ti content of hornblende may actually decrease with increasing metamorphic grade. With the exception of hornblende sample AM41 which has  $\gamma$  brown and is interpreted as igneous (Fig 53) all the Tantalite Valley hornblendes generally have  $\gamma$  blue-green which corresponds to the low titanium contents observed and probably to high  $\text{Fe}^{3+}$ . The hornblendes from the Kumkum area have  $\gamma$  ranging from brown to olive-green and some are interpreted as being igneous in origin (Fig 53, discrete grains in samples K4 and K37A). Discrete amphibole grains have Ti contents resembling values quoted by Raase (1974) for hornblendes from granulite facies rocks but amphiboles forming part of the coronas between olivine and plagioclase in spite of having high  $\text{Al}^{\text{IV}}$  contents (indicative of relatively high

grade) have Ti contents which resemble those of hornblendes from comparatively low-grade rocks. This is explained by the fact that olivine and plagioclase from which the hornblende was formed during the corona reaction are both Ti-poor phases.

Many authors (eg Leake 1965<sup>b</sup>, 1968; Bard 1970; Raase 1974; Liou et al 1974) ascribe enrichment of Ti in hornblende with increasing metamorphic grade, to increasing temperature. The mechanism for this is uncertain. In the Y site only,  $\text{Al}^{\text{VI}}$  and  $\text{Fe}^{3+}$ , other than Ti, are grade controlled. (Bard 1970).  $\text{Fe}^{3+}$  decreases with grade and  $\text{Al}^{\text{VI}}$  is controlled by pressure, increasing with increasing pressure (Leake 1965<sup>b</sup>). With increasing temperature more of the total Al substitutes for Si so that less aluminium goes into the Y site as  $\text{Al}^{\text{VI}}$ . Thus with increasing temperature of metamorphism there will be less  $\text{Al}^{\text{VI}}$  in the Y site and probably less  $\text{Fe}^{3+}$  since the rock will be less hydrated and therefore less oxidized so that more Ti can be incorporated in this site. This interpretation is favoured here to explain the observed colour differences between hornblendes examined from the Tantalite Valley and Kumkum areas because pressure is not thought to have been significantly different between the two areas (see Fig 52 and section 4.1).

#### 6.4.5 Co-existing hornblende and cummingtonite

Cummingtonite intergrown with hornblende has been reported from a large number of metamorphosed

mafic and ultramafic rocks, particularly amphibolites (eg Vernon 1962; Leake 1972; Hietanen 1973 ), from metasedimentary rocks of suitable composition, including banded ironstones (Bonnichsen 1969) and from unmetamorphosed volcanic rocks (Klein 1968). The co-existence of hornblende and cummingtonite has been used by Ross et al (1969) as evidence for a miscibility gap existing between calcium-rich and calcium-poor amphiboles under certain geological conditions. The unmixing of hornblende and cummingtonite is usually taken to indicate cooling following a higher grade metamorphism (Stout 1972, p 109). Cummingtonite co-existing with hornblende is not found in environments of low-grade or high P metamorphism but is confined to medium-to high-temperature medium-grade metamorphic rocks (Leake 1972).

Mg-Fe distributions in cummingtonite-hornblende pairs have been shown by Kisch and Warnaars (1969) to be markedly different from the Mg-Fe distributions for cummingtonite-actinolite pairs. Actinolites are found to have a higher Mg/Fe ratio than co-existing cummingtonite whereas the reverse is true when hornblende is the co-existing calcic amphibole, ie  $K_D^{\text{actinolite-cm}}$  is usually much greater than unity, whereas  $K_D^{\text{hornblende-cm}}$  is usually equal to or less than unity.

The fact that cummingtonite is richer in iron than

co-existing actinolite is explained by the M4 position in cummingtonite being predominantly filled by  $\text{Fe}^{2+}$  and not Ca as in the case of actinolite. Filling of the remaining three octahedral sites,  $M_1$ ,  $M_2$  and  $M_3$  by comparable amounts of Mg and  $\text{Fe}^{2+}$  in co-existing cummingtonite and actinolite will then produce the observed distribution coefficient relationship. This effect will be counterbalanced slightly by  $\text{Fe}^{3+}$  which enters calcic amphibole in preference to co-existing cummingtonite and will tend to make the distribution coefficient using total iron smaller than that using only  $\text{Fe}^{2+}$ .

The relationship between hornblende and cummingtonite is explained by the non-random distribution of  $\text{Al}^{\text{VI}}$  over octahedral lattice sites in hornblende which will have the effect of lowering the overall Mg/Fe ratio in the latter, since co-existing cummingtonite remains Al-poor. Also Warnars and Kisch (1969) suggested that  $K_D^{\text{hornblende-cm}}_{\text{Mg-Fe}}$  is affected by the presence of an Al-rich phase; usually plagioclase. From 27 hornblende-cummingtonite pairs they noted that the distribution coefficient decreases as the calcium content of co-existing plagioclase decreases but did not point out that the aluminium content of the hornblende generally increases as well.

Cumingtonite exsolving hornblende and vice versa is a common feature of metabasite hornfels and satellite body metagabbro from the Tantalite Valley area. The exsolution is mostly irregular and patchy but green tschermakitic or magnesium hornblende may exsolve parallel to  $(10\bar{1})$  and  $(100)$  in cumingtonite as irregular lamellae (Fig 35F ).

Very fine hornblende lamellae may be responsible for the unusual composition observed for cumingtonite which has altered <sup>from</sup> orthopyroxene in Tantalite Valley gabbro sample AM21 (Table 43 ).

Actinolitic hornblende also occurs as an alteration product of pyroxene. In cumingtonite of metamorphic cumingtonite - hornblende pairs, the maximum contents of  $\text{CaO}$ ,  $\text{Al}_2\text{O}_3$  and  $\text{Na}_2\text{O}$  are 1,9%, 3,2% and 0,2% respectively (Klein 1968).

Since cumingtonite in AM21 has 5,71 weight %  $\text{CaO}$  and 2,29 weight %  $\text{Al}_2\text{O}_3$ , it seems highly likely that contamination, especially by exsolved hornblende, is the cause (Stout 1972). For co-existing hornblende and cumingtonite pairs from the meta-

basite hornfels and the metagabbro of the satellite body  $X_{\text{Hb-Mg}}^{\text{Hb}}$  is always less than  $X_{\text{Mg}}^{\text{cm}}$ , and the

$K_{\text{D Mg-Fe}}$  generally increases as the co-existing plagioclase in the rock becomes more sodic (Table 42).

The latter observation is the reverse of that found by Warnaars and Kisch (1969). The discrepancy can be explained by the fact that for a restricted

range of  $Mg/(Mg + Fe)$  ratios, as are involved here, the distribution coefficient increases as the Al content of the hornblende decreases. The aluminium content of hornblende varies sympathetically with the aluminium content of the whole rock for the Tantalite Valley samples. For the metagabbro of the satellite body the hornblende contains an average of 10,75 weight %  $Al_2O_3$ , while the host rock contains 10,48 weight %  $Al_2O_3$ , while the metabasite hornblende contains an average of 13,54 weight %  $Al_2O_3$  while the host rock contains in the region of 18 weight %  $Al_2O_3$ . Also calcic plagioclase contains more aluminium than sodic plagioclase and the co-existing hornblende would be expected to show a corresponding variation in aluminium content. Thus the dominant factor in controlling the magnitude of the distribution coefficient is probably the aluminium content in hornblende and not that in a co-existing aluminous phase. The  $Al_2O_3$  content of the rock and the metamorphic conditions are responsible for controlling the Al content of hornblende. The first control is difficult to evaluate because not enough data are available but is probably more important than the second. The  $Al^{VI}$  content in hornblende increases with increasing grade of metamorphism, more particularly with increasing pressure (Leake 1965a). It is not here envisaged that the metabasite horn-

fels formed at a higher pressure than the meta-gabbro of the satellite body and bulk control of aluminium content of hornblende is favoured.

The Mg-Fe distribution between co-existing cummingtonite and hornblende can be shown graphically in a triangular diagram with the apices Mg, Ca, Fe total all expressed as mole percent. (Fig 5<sup>4</sup> ).

As almost all of the hornblendes are highly aluminous (6 - 12 weight percent) their graphical representation in (Fig 5<sup>4</sup>) is only partial. The hornblende analysis points can be considered as projections onto the CaO-MgO-FeO plane from  $Al_2O_3$ . The reason why some calcic amphiboles plot above the  $Ca_2 (Fe^{2+}, Mg)_5$  line is because of cations other than Ca,  $Fe^{2+}$  and Mg in octahedral sites.

Manganese is preferentially concentrated in cummingtonite in cummingtonite-hornblende pairs (Klein 1968). This is true for the Tantalite Valley samples where MnO content of cummingtonite is consistently about two to three times larger than in co-existing hornblende. The preferential concentration of Al and Ti in hornblende rather than cummingtonite is also in agreement with the findings of Leake (1968).

#### 6.4.6 Co-existing Hornblende and Actinolite

Co-existing actinolite-hornblende usually

indicates that grade of metamorphism was transitional between greenschist and amphibolite facies (Ernst 1968) although under conditions of relatively low P and T, it may form in the epidote-amphibolite facies (Grapes 1975). This assemblage has been widely reported from both regional and contact metamorphic rocks formed under low P and is absent from Barrovian terranes (Ernst 1968). Grapes (1975) distinguished two types of co-existing hornblende/actinolite pairs in low P regional metamorphic terranes which he described as equilibrium and disequilibrium. In the former, the amphiboles exist as separate homogenous grains or intergrown with each other and this is taken to indicate the existence of a miscibility gap between actinolite and hornblende. In the latter, amphiboles commonly show a "zoning" relationship with each other and their compositions do not represent points across a miscibility gap. Co-existing hornblende and actinolite of the disequilibrium type is a common feature found in the metagabbro and amphibolites of Tantalite Valley. Pale green actinolite or actinolitic hornblende, rimmed by darker green magnesian hornblende is found as a result of incomplete uralitization of these rocks (see section 7.2.1). Grapes (1975) has suggested, for similar relationships, that shearing deformation during uplift may be responsible for the change from actinolite to hornblende not going to completion.



Hornblende is more aluminous and has  $Mg/(Mg + Fe)$  less than co-existing actinolite as is clearly shown from electron microprobe analyses from metagabbro sample AM90. These findings are in agreement with those of Klein (1969) for seven actinolite-hornblende pairs. This result is explained by the fact that alumina, which is more abundant in hornblende than actinolite, has the ability to occupy Mg-preferent octahedral sites in large amounts, decreasing the total  $(Mg + Fe)$  there, and seeing that Fe is preferred over Mg in the remaining sites, it has the effect of lowering the overall  $Mg/(Mg+Fe)$  ratio. In actinolite where aluminium is relatively less abundant, this effect is hardly noticeable. (Kisch and Warnars 1969).

#### 6.5 Compositional Changes in Spinel

If analyses of spinels from rocks from Kumkum and Tantalite Valley are plotted on a ternary  $Fe^{3+}$ , Cr, Al diagram (Fig 55) it can be seen that the distribution of points is similar to the trend noted by Evans and Frost (1975 Fig 2) for spinels metamorphosed under conditions of the amphibolite and granulite facies. They observed that such spinels may define a complete spectrum of compositions ie magnetite-ferritchromite-Al-rich chromite green hercynite/spinel, with spinel analyses from a

single sample often encompassing a large part of this range. This last observation is true also for some of the rocks studied here, since in troctolite sample TV49 from Tantalite Valley spinel, picotite and ferritchromite occur together and in Black Gneiss sample K101 from the Kumkum area hercynite, hercynite/ulvöspinel/spinel solid solution and magnetite co-exist. The mineral assemblage found in sample TV49 (Table 27) is characteristic of the Orthopyroxene-plagioclase lherzolite mineral facies in ultramafic rocks as defined by O'Hara (1968), which forms under similar P and T conditions to granulite facies rocks. Thus it is to be expected that spinels from this rock will approximate the trend given for high grade rocks in Fig (2) in Evans and Frost (1975).

For spinels from metamorphosed ultramafic rocks there is a correlation between  $100 \text{ Cr}/(\text{Cr} + \text{Al})$  and  $X_{\text{Mg}}$  as can be seen in Fig 4 of Evans and Frost (1975). In a similar diagram given as Fig (56) here it is evident that while spinels from troctolite sample TV49 demonstrate this effect well, all the other spinel analyses from metamorphosed rocks of non-ultramafic bulk composition show no correlation at all.

Metamorphosed chromites from ultramafic rocks have higher  $\text{Fe}^{2+}/\text{Mg}$  ratios than their unmetamorphosed

equivalents (Evans and Frost 1975) so that it is quite possible that the ferritchromites from sample TV49 plotted in the field of chromites from Strati-form rocks (Fig 56) before metamorphism.

The  $f_{O_2}$  at spinel formation can be estimated from Fig 11 of Irvine (1965), valid for the assemblage spinel, olivine, orthopyroxene. This method was tried for spinel sensu stricto from troctolite sample TV49 and for hercynite-spinel solid solution from Kumkum gabbro-norite sample K37A where olivine and orthopyroxene co-exist with spinel. Both spinel analyses used have almost no Cr present so that the "Al" -  $f_{O_2}$  isobars can be read for the relevant values of 100  $\gamma$  and 100  $X_{Mg}^{spinel}$  in Fig 11 (valid for temperatures of around 1 200°C) of Irvine (1965). The results are  $f_{O_2} = 10^{-7.4}$  atm and  $f_{O_2} = 10^{-9.4}$  atm respectively. It is shown elsewhere (section 5.2) that sample TV49 was crystallized between 700 and 1 200°C, probably near 1 000°C and that K37A was emplaced at temperatures near 1 000°C and re-equilibrated to the temperature of the prevailing metamorphism, ie 800°C so that while the  $f_{O_2}$  value determined for sample TV49 is probably realistic, that for sample K37A can only be considered a very rough estimate. The  $f_{O_2}$  value for sample TV49 gives an idea of the oxidation state of the troctolite early in its crystallization since spinel and

olivine were the first phases to crystallize in this rock. That the value of  $fo_2 = 10^{-7.4}$  atm for sample TV49 is realistic is seen from the experimental work of Ulmer (1969) who found that at one atmosphere and  $1300^\circ\text{C}$  that  $fo_2$  in the range from  $10^{-5}$  to  $10^{-7}$  atm is necessary to form the range of compositions observed for natural spinels.

#### 6.5.1 Ferritchromite and its origin in mafic and ultramafic rocks from Tantalite Valley

Spangenberg (1943 p 23) coined the term "ferritchromite" to describe an opaque mineral having the chemical and physical properties between chromite and magnetite. Bliss and MacLean (1975) define ferritchromite as "an alteration product of chromite that is deficient in Mg and Al but does not represent a mineral of specific composition."

The most likely origin suggested for the generation of this mineral is that proposed by Bliss and MacLean (1975) who showed that ferritchromite formation is not directly related to serpentinization, as postulated by Beeson and Jackson (1969), but occurs as a result of the rock being metamorphosed to at least lower amphibolite facies grade. Serpentinization may however bring into contact the

reactants for producing ferritchromite by depositing rims of secondary magnetite on magmatic chromite and the ferritchromite forms from the latter by means of a replacement process in which  $\text{FeO}$  and  $\text{Fe}_2\text{O}_3$  substitute for  $\text{MgO}$ ,  $\text{Al}_2\text{O}_3$  and to a lesser extent,  $\text{Cr}_2\text{O}_3$  (Weiser 1967; Beeson and Jackson 1969; Bliss and MacLean 1975). The  $\text{MgO}$  and  $\text{Al}_2\text{O}_3$  released may then be taken up in chlorite (Beeson and Jackson 1969; Bliss and MacLean 1975). It should be noted, however, that ferritchromite alteration is not confined to serpentized rocks and occurs in chromite-bearing mafic rocks generally that have been subjected to high temperatures, eg xenoliths and xenocrysts (Frisch 1971). Further, ferritchromite formation occurs under a relatively high  $f_{\text{O}_2}$  and variations in  $\text{Mg/Fe}^{2+}$  during this process are buffered by co-existing mafic silicates (Bliss and MacLean 1975).

In the Tantalite Valley area, <sup>discrete grains of</sup> ferritchromite have been identified in the mottled metagabbro (TV275) and in one troctolite (TV49) by means of the electron microprobe. Sample TV275 was collected close to the contact with metagabbro which is intrusive into the mottled metagabbro and the emplacement of the former could have provided the heat required for the alteration of primary chromite to ferritchromite in the latter. An optical

examination of mottled metagabbro samples found away from the contact did not reveal the presence of ferritchromite. The troctolite of Tantalite Valley is characteristically serpentized and chloritized which indicates that water was entering into the rock at the same time as the area was experiencing the effect of waning low-grade regional metamorphism. Thus under these conditions and with the troctolite originally containing magmatic chromite, it is hardly surprising that ferritchromite is found.

The ferritchromite in the mottled metagabbro is devoid of MgO and that in the troctolite has spinel  $X_{Mg}$  in the range 0,078 to 0,091 compared with 0,085 - 0,112 for ferritchromite from the Stillwater Intrusion. The Cr content of ferritchromites from Tantalite Valley, (ie  $\alpha$  = 0,298 to 0,331) is lower than those for Stillwater ferritchromite (ie  $\alpha$  = 0,619 to 0,751).

#### 6.6 Compositional Zoning in Garnets

Garnets showing a regular decrease in Mn content from cores to rims along a typical bell-shaped profile are well known from low-to medium-grade metamorphic environments (eg Hollister 1966; Atherton 1968; Cooper 1972). Less commonly, the

situation arises in which the rims of garnet are enriched in Mn with respect to their cores, as has been reported by several authors (eg Grant and Weiblen 1971; de Béthune et al 1968; Okrusch 1971). These two types of zoning are, for convenience, referred to in the proceeding discussion as "normal" and "reverse" respectively.

The most widely accepted model to explain the production of normal zoning in garnet is equilibrium segregation or Rayleigh fractionation and the entire garnet is regarded as having grown during prograde metamorphism (eg Atherton 1968; Hollister 1969). In low-to medium-grade rocks FeO and MgO increase and MnO decreases from the core towards the rim of the garnet, which corresponds to the trend which might be expected under conditions of increasing metamorphic grade (Miyashiro 1953; Atherton 1968). CaO either decreases towards the rim (eg Atherton 1968) or it has a uniform concentration even in grains from low-grade rocks (Woodsworth 1977). At temperatures above about 600°C zoned garnets generally tend to become homogenized as a result of increased efficiency of ionic diffusion thus explaining the virtual absence of zoned garnets from high-grade metamorphic terranes (Evans and Guidotti 1966; Woodsworth 1977; Anderson and Olimpio 1977). In order to explain reversed zoning it is necessary

to account for the Mn-enriched rims. The latter may be formed as a result of the breakdown of a Mn-rich phase such as ilmenite or as is usually the case, by unstable garnet releasing some Mn which rediffuses back into the garnet if no other mineral phase will accept it. (de Béthune et al 1975; Hollister 1977). Such zoning is normally regarded as being the result of retrogression (eg Woodsworth 1977; Ashworth and Chinner 1978).

In the present study area, normal zoning was detected in porphyroblastic garnets from Tantalite Valley pelitic hornfels sample TV4 and hornfelsic Black Gneiss sample K35 (Table 43 ). In both specimens Fe, Ca and Mn decrease and Mg increases on going from core to rim. The porphyroblasts in sample TV4 are regarded as having grown entirely during the contact metamorphic event. In sample K35, taken from the aureole of the main Kumkum gabbro-norite body, the situation is somewhat different. Here in addition to porphyroblasts occurring, there are also groundmass garnets which appear to be unzoned. These have slightly lower Mn, Ca and Fe and higher Mg than the rims of the porphyroblasts. The porphyroblasts are interpreted as garnets which formed during the regional metamorphism and which have re-equilibrated under conditions of prograde contact metamorphism. The groundmass garnets are



interpreted as having formed entirely during the later event when the groundmass was somewhat depleted in Mn, Ca and Fe and correspondingly enriched in Mg. In the present area, therefore, normal zoning is apparently restricted to contact metamorphic rocks and is generally quite scarce.

Significant reverse zoning was noted in garnets from a sample of the Black Gneiss (K101) containing the assemblage cordierite+garnet+gedrite and from a sample of the Tantalite Valley pelitic hornfels (T19, Table 43). A small degree of reverse zoning was also detected in sample K66B from the Kumkum area which was collected close to the contact with Inequigranular granite gneiss and has an hornfelsic texture. Garnets from these rocks show an increase of Fe, Mn and Ca and a decrease of Mg in going from core to rim, which is interpreted as indicating re-equilibration of rim compositions in response to retrogressive metamorphism, probably in the form of a decrease in both pressure and temperature. If garnet + cordierite + gedrite-bearing Black Gneiss formed as a restite after some partial melting had taken place, as seems highly likely, (section 7.2.3), then under cooling conditions, reverse zoning could have formed in garnet. Likewise the same pattern of zoning could be formed in garnets from a hornfels which cooled after the initial high temperature of contact metamorphism.

In most of the garnet examined from the Black Gneiss proper no zoning of any consequence was detected (eg compare core and rim analyses of A3, Table 43 ). This is probably as a result of homogenization of garnet compositions under conditions of high-grade metamorphism as discussed earlier. However the P and T conditions which are calculated using garnet rims are more compatible with estimates for conditions prevailing during retrograde metamorphism. This is borne out by the fact that where garnet has partially broken down to form cordierite, a common occurrence in the area, the P,T conditions calculated from these grains are the same as for discrete garnet and cordierite grains in contact with each other.

The temperature at which retrogressive metamorphism took place was around  $600^{\circ}\text{C}$  (section 5.4 ) and thus further rehomogenization under the new P - T conditions could probably take place. (In pyroxenes from the Kumkum gabbroids re-equilibration diffusion ceased at around  $800^{\circ}\text{C}$  which is taken as a maximum estimate for the temperature of the main metamorphism). The system must then have been rapidly frozen, possibly as a result of rapid uplift, to prevent the formation of widespread reverse zoning in garnet following further retrograde metamorphism.

## 6.7 Biotite

Although the composition of biotite is largely dependent on the bulk chemistry of the host rock, and the nature of co-existing ferromagnesian phases present, Engel and Engel (1960) showed that biotite from the Adirondacks, U.S.A., shows a decrease in Mn,  $\text{Fe}^{2+}$  and  $\text{Fe}^{3+}$  and an increase in Ti and Mg, with increasing metamorphic grade. Binns (1969) found that the Ti-content and the  $\text{Mg}/(\text{Mg} + \text{Fe})$  ratio for biotites from metapelites in the Broken Hill area, New South Wales, increases with increasing metamorphic grade. The colour of biotite changes from green through greenish-brown, brown and reddish-brown with increasing grade of metamorphism and is controlled primarily by a decrease of  $\text{Fe}^{3+}$  and an increase in Ti (Engel and Engel 1960; Miyashiro 1973; Jackson 1976).

Two valuable grade dependent quantities for biotite will therefore be  $\text{TiO}_2$  and the  $\text{FeO}^*/\text{MgO}$  ratio and these are plotted against each other for biotites from the present study area in Fig ( 57 ). It can be seen that biotites from the Black Gneiss have higher  $\text{TiO}_2$  contents and a relatively lower  $\text{FeO}^*/\text{MgO}$  ratio than for biotites from Tantalite Valley pelitic hornfels even though the average bulk rock  $\text{TiO}_2$  and  $\text{FeO}^*/\text{MgO}$  values for Black Gneiss

and Tantalite Valley pelitic hornfels are approximately the same (compare A and B in Fig 57 ). This would imply that biotites from the Black Gneiss have been metamorphosed at a higher grade of metamorphism than those from the Tantalite Valley pelitic hornfels. This is also reflected by the  $\gamma$  colour of biotite, which is pale brown, olive-brown, reddish-brown for biotite from Black Gneiss and is green, greenish-brown and reddish-brown for biotite from pelitic hornfels. However, the possibility that the biotites from the Tantalite Valley pelitic hornfels have been subjected to retrograde metamorphism cannot be ruled out, since biotite may be altered to chlorite (eg sample TV276). Generally, the biotites from the Tantalite Valley hornfels have a constant  $\text{TiO}_2$  value of between 1 and 2% for a large range of  $\text{FeO}^*/\text{MgO}$  values.

An examination of Fig ( 57 ) shows that biotites from mafic rocks of the Kumkum area are characterized by higher  $\text{TiO}_2$  contents than those from mafic rocks in the Tantalite Valley area. This is surprising because biotite from the gabbro-norite, satellite body gabbro-norite and ultramafic rocks of Tantalite Valley are regarded as igneous and would be expected to contain more  $\text{TiO}_2$  than biotites from the Kumkum mafic rocks which can be considered as metamorphic (see section 6.2.4). The

explanation for this is that the igneous biotites exsolved  $\text{TiO}_2$  on cooling, evidence for which is seen from rutile forming crystallographically oriented needles in phlogopite in ultramafic rocks (Fig 33D) and the occurrence of ilmenite parallel to the cleavages of biotite.

The biotites from the mottled metagabbro may be produced as a result of metasomatic introduction of potassium into these rocks from the neighbouring Inequigranular granite gneiss as discussed in section ( 4.7 ). The similarity of biotite compositions in the mottled metagabbro to those of primary igneous biotite from <sup>granitic rocks as represented by</sup> the Violsdrift granodiorite and adamellite (Reid 1977) which can be seen in Fig ( 57 ) lends support to this hypothesis.

## 6.8 Chlorite

Chlorite is commonly found in rocks which have undergone shearing in the area, particularly those occurring along the axis of the Tantalite Valley Shear Zone. Its development appears to be related to reactivation of shear zones late in the tectonic history of the area.

Using the nomenclature of Hey (1954) the majority of the chlorite examined was found to be ripidolite

with the exception of pycnochlorite in some psammitic and metabasite hornfels (samples AM30A and AM31B).

Cooper (1972) found that in metabasites from the Haast Schist Group in New Zealand, the Mg/Fe ratio of chlorite increases with increasing grade. This is represented on Fig ( 58 ). Chlorites from the metabasites and metagabbros of the Tantalite Valley area all plot in the same field as the Haast River chlorites and correspond to those formed at the garnet and oligoclase zones of the greenschist facies in the latter area. Chlorites from Tantalite Valley pelitic hornfels and cordierite-anthophyllite hornfels are enriched in Al compared with the field established by Cooper (1972) for his metabasites. The high Mg/Fe ratio of chlorite from a metamorphosed sample of ultramafic rock from Tantalite Valley (Fig 58) does not imply that this rock was metamorphosed to a higher grade than the metagabbros and metabasites from the same area but rather that the high ratio was inherited from the chemistry of the host rock. The field of chlorites from metamorphosed Haib basaltic andesites (Reid 1977) is shown on Fig ( 58 ) for comparison. The Haib rocks were apparently metamorphosed to a slightly higher grade compared to rocks of a similar compositional range from Tantalite Valley.

### 6.9 The origin of sulphide minerals in rocks from Kumkum and Tantalite Valley

Sulphide minerals occur mainly in the Tantalite Valley metagabbro, ultramafic rocks and hornfels and in the Kumkum gabbro-norite. In the gabbroic and ultramafic rocks, they occur commonly as aggregates up to 2 cm in diameter, consisting of mostly pyrrhotite with lesser amounts of chalcopyrite and pentlandite which may occur as exsolution "flames" and blebs within the pyrrhotite (see also Beukes 1973). The sulphide minerals are always disseminated and are interstitial or intercumulus. Pyrrhotite, minor chalcopyrite and pentlandite and rare pyrite were mobilized during metamorphism of the mafic-ultramafic rocks of the Tantalite Valley Complex and migrated to adjacent hornfels where they concentrated along joint and shear planes.

On the basis of a study on the immiscible sulphide melt in a cooling tholeiitic lava from Hawaii, Skinner and Peck (1969) concluded that the immiscible sulphide-rich liquids which separate from cooling basic magmas are always FeS-rich and contain a significant oxygen content, which is manifested usually by the presence of magnetite in the solid rock.

The first phase to crystallize from such a liquid will be a Cu-Ni-pyrrhotite solid solution which, when cooled below 600°C will yield pyrrhotite with chalcopyrite, pyrite and pentlandite as possible exsolved phases. However, Craig and Kullerud (1969) have shown that pyrite and pentlandite cannot co-precipitate. Craig (1966) has shown that melting in a typical massive nickeliferous-cupriferous-pyrrhotite ore as found in mafic and ultramafic rocks starts at 1105°C, and that addition of 15% magnetite to the system lowers this temperature to 1019°C. The effect of water on the melting temperature is minimal (see Craig and Kullerud 1969).

From the above discussion, it is inferred that in the ultramafic and mafic rocks from Kumkum and Tantalite Valley, an immiscible sulphide melt separated from silicate liquid filling the interstices of a crystal mush as the system cooled below about 1100°C. At still lower temperatures, a Cu-Ni-pyrrhotite solid solution crystallized out. At temperatures below about 600°C, exsolution of chalcopyrite and pentlandite could take place in pyrrhotite. The ultramafic rocks generally have a greater pentlandite:pyrrhotite ratio than the gabbroic rocks (Beukes 1973) since more nickel was present in the ultramafic magmas compared with the gabbroic magmas. The euhedral nature of pyrite in



these rocks was taken by Beukes (1973) to mean that pyrite was the first sulphide to form but experimental work has shown that chalcopyrite will form before pyrite in cooling a Cu-Ni-Fe sulphide melt and that pentlandite and pyrite cannot crystallize directly from such a melt (Craig and Kullerud 1969). Furthermore, there is considerable evidence that pyrite has a strong 'crystallizing power' so that it tends to form subhedral crystals anyway (see Stanton 1972, p 80). It is tentatively postulated here that the pyrite was the last sulphide to crystallize with its growth occurring in a solid medium. The presence of pyrrhotite-pentlandite-chalcopyrite aggregates in the Kumkum gabbro norite implies that the rock was emplaced at around  $1100^{\circ}\text{C}$  and cooled to below  $600^{\circ}\text{C}$  which is a serious objection to this rock being a charnockite. <sup>as proposed by Toogood (1976)</sup> Beukes (1973) has reported <sup>^</sup>the common occurrence of magnetite with ilmenite lamellae in the Kumkum gabbro norite which indicates that cooling of this rock took place under oxidizing conditions (Buddington and Lindsley 1964).

Addendum to Chapter 7 as required by External Examiners, 1979.

In the present study it was decided to supplement thin section data for metamorphic rocks from the Tantalite Valley-Kumkum area with those of previous workers to make the coverage as broad as possible. Many of their samples were not available for study (Beukes 1973, Toogood 1976) so that the minerals present were simply listed without regard to equilibrium relationships. Using the Phase Rule it was found that the number of phases present in most samples ruled out the possibility of an equilibrium assemblage being present. The mineral lists were then used to establish rock groupings and comparisons made with samples collected from Tantalite Valley and Kumkum.

The purpose of this classification was:-

- (i) To systematize the discussion.
- (ii) To establish the frequency of occurrence of the various classes belonging to a particular metamorphic rock-type.
- (iii) To try to establish whether reasonable reactions could be proposed to explain the present mineral assemblages, particularly where samples were not available for study. The feasibility of proposed reactions was determined by comparison with similar reactions which could be established in those rocks for which sections were available. On textural grounds two metamorphic episodes may be recognized in some rocks, based on the presence of relic minerals. Compositional variations of similar phases (e.g. garnet, cordierite, biotite) also reflect the different physical conditions of the two metamorphic episodes.

111.

In discussions of metamorphic minerals and geobarometry-geothermometry it was assumed that phases in contact with one another were in equilibrium (at least at their contacts). Armoured phases and the cores of zoned phases were considered to be out of equilibrium with the "matrix" minerals in the rock. Most of the mineralogical calculations were done with these assumptions in mind. In some cases, described in the text, trial calculations were carried out on minerals even though there were doubts as to whether they were in equilibrium.

The author acknowledges the shortcomings of the above approach but defends the use of it in the general absence of unequivocal textural relations discriminating between M1 and M2 equilibrium assemblages (parageneses), the pervasive and widespread development of M2 parageneses which served to mask the evidence of an earlier metamorphic episode and the non-availability of many of the samples used.

The minerals taken as characteristic of the metamorphic episodes, M1, M2 and M<sub>3</sub>, as inferred from detailed textural and mineralogical work are given below:-

TABLE:

| Rock Type        | M1             | M2             | M3 <sup>*</sup>        |
|------------------|----------------|----------------|------------------------|
| METAPELITES      | sillimanite    | sillimanite    | andalusite             |
|                  | cordierite     | cordierite     | muscovite              |
|                  | garnet         | garnet         | chlorite               |
|                  | biotite        | biotite        | biotite                |
|                  | sapphirine     | staurolite     | chloritoid             |
|                  | corundum       | orthoamphibole | epidote                |
|                  | hercynite      | (kyanite)      | talc                   |
|                  | hypersthene    | hercynite      | prehnite               |
|                  | orthoamphibole |                |                        |
| METABASITES      | orthopyroxene  | orthopyroxene  | chlorite               |
|                  | clinopyroxene  | clinopyroxene  | prehnite               |
|                  | hornblende     | hornblende     | epidote                |
|                  | spinel         | garnet         | tremolite-actinolite   |
|                  |                | cummingtonite  | actinolitic hornblende |
| CA-METAMORPHITES | calcite        | clinozoisite   | muscovite              |
|                  | diopside       | tremolite      |                        |
|                  | hornblende     | brucite        |                        |
|                  | scapolite      | serpentine     |                        |
|                  | zoisite        | humite         |                        |
|                  | grossularite   |                |                        |
|                  | forsterite     |                |                        |
|                  | tremolite      |                |                        |
|                  |                |                |                        |

\* M3 - refers to localized retrograde metamorphism related to contact or shearing metamorphism.

## CHAPTER 7

## METAMORPHISM

7.1 Contact Metamorphism of the Tantalite Valley Complex

It is usual for zoning to be found in contact metamorphic aureoles as a result of the temperature gradient. It has not been possible to establish consistent patterns for the Tantalite Valley aureole because of its narrow width, uneven development, tectonic disruption and superimposed metamorphic effects. Instead an attempt has been made to establish the range of conditions which prevailed during contact metamorphism using those minerals which are considered to have represented equilibrium assemblages at the peak of metamorphism.

7.1.1 Pelitic Hornfels

The textures and mineralogy of the pelitic hornfels samples reveal that, except for sample TV276, staurolite is found only as rounded inclusions within other minerals, particularly cordierite and andalusite which suggests that the staurolite has

survived only as a relic phase. Regionally metamorphosed pelitic rocks outside the aureole contain staurolite-free assemblages such as cordierite + garnet + sillimanite or garnet + sillimanite + biotite (Beukes 1973; Toogood 1976) so it is most likely that the staurolite formed during contact metamorphism.

The staurolite-forming reaction cannot readily be deduced. The phase relations between staurolite, garnet, cordierite, aluminosilicate and anthophyllite in the model system  $\text{FeO} - \text{MgO} - \text{SiO}_2 - \text{Al}_2\text{O}_3 - \text{H}_2\text{O}$  are illustrated by the Schreinemakers' bundle (Fig 62) after Robinson and Jaffe (1969). If staurolite is considered to have formed stably during contact metamorphism the assemblages belong to sector 'A' (Fig 62), but the textures which indicate the breakdown of staurolite (to cordierite or andalusite) suggest crossing of the reaction:

$$\text{staurolite} + \text{quartz} \rightleftharpoons \text{cordierite} + \text{garnet} + \text{aluminosilicate} + \text{H}_2\text{O} \quad (1)$$

into sector 'B'. This may be achieved either by an essentially isobaric temperature increase, or by a reduction in pressure. That is, the staurolite may break down under prograde conditions during development of the aureole or by a retrograde process involving uplift at some subsequent stage. The latter alternative is preferred here since the

estimates of peak metamorphic conditions fall well within the sillimanite field of Holdaway (1971) and at the edge of the sillimanite field according to Richardson et al (1969), whereas the aluminosilicate texturally associated with staurolite is andalusite. The temperatures at which this re-equilibration took place may possibly be indicated by the Mg-Fe partition between cordierite (or biotite) and the garnet rims (ie 525°C) and such retrograde changes may be associated with deformation and metamorphism in the Tantalite Valley shear zone, possibly associated with the intrusion of the younger pegmatites (Moore 1975) which are transgressive to the contact aureole and associated rocks.

An alternative explanation for the breakdown of staurolite rests on the suggestion of Compton (1960) and Turner (1968, p 210), and in the experimental work of Ganguly (1972) that the stability field of staurolite extends towards higher  $f_{O_2}$  than that of almandine and cordierite. In particular, if staurolite partitions  $Fe^{3+}$  relative to garnet and cordierite then reaction (1) may proceed isothermally with decreasing  $f_{O_2}$ .

According to Hess (1969) the staurolite + cordierite paragenesis is stable at 515 - 590°C at 1,7 - 4 kb, which is here valid only for sample TV276.

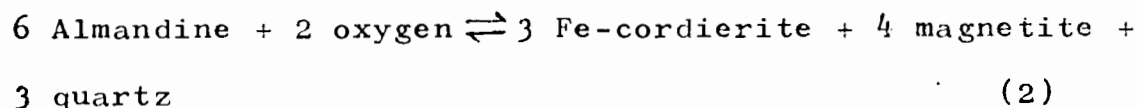
Although both andalusite and sillimanite commonly

coexist in the same thin section, the andalusite is present as poikiloblastic laths and anhedral, interlobate grains enclosing numerous small inclusions of quartz, staurolite, sillimanite, opaque minerals and biotite. This is interpreted as indicating that the andalusite growth post-dates that of the enclosed minerals which are relic phases. Sillimanite is not regarded as an unambiguous index mineral in the aureole because in many samples where it is found it shows features which indicate that it is older than the contact metamorphism (Moore 1975, Fig 15). Furthermore, sillimanite-rich varieties of the Grey Gneiss Sequence are present at a distance from the aureole (Moore 1975). Kyanite has not been found on a regional scale (Beukes 1973; Moore 1975; Toogood 1976) and even within the aureole TV4 is the only sample in which it occurs. Its development rather than that of andalusite (the more common aluminosilicate in the aureole) is difficult to explain. There may be a bulk chemical control as suggested by Pitcher (1965) and Langer (1976), since if the occurrence of kyanite in aureoles is best explained by simple equilibrium crystallization in the kyanite field (Naggar and Atherton; Atherton, Naggar and Pitcher 1975) with continued crystallization along an essentially isobaric path, this mineral would be found in



aureoles more frequently. Kröner (1971<sup>b</sup>) has shown that it is possible to produce kyanite at pressures as low as 1,2 kb at room temperature.

The evidence from zoning in garnet (see section 6.6 ) indicates that it grew during the contact metamorphism. Almandine is relatively rare in contact metamorphic rocks because of its restricted stability under low pressure conditions (Chinner 1962). However, in the Tantalite Valley aureole almandine is common, suggesting moderate P,T conditions. According to Hsü (1968) the stability of almandine is also controlled by the fugacity of oxygen in pelites undergoing metamorphism. At high values of  $f_{O_2}$  (above the QFM buffer), almandine tends to break down according to the reaction:-

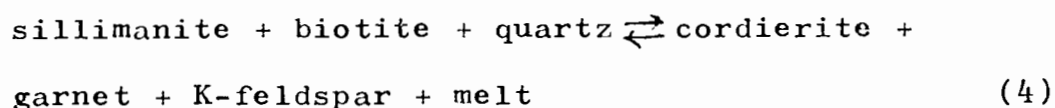
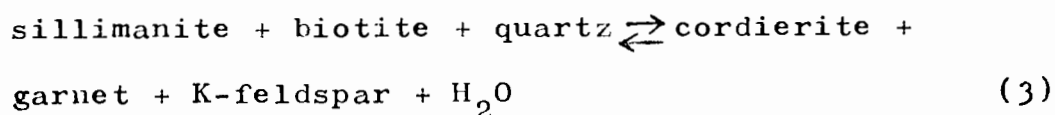


Hercynite, a mineral which may occur in pelitic hornfels, was shown by Turnock and Eugster (1962) to be stable only in a restricted field of low  $f_{O_2}$  and this, and the presence of almandine in the pelitic hornfels indicates that relatively reducing conditions were operative during contact metamorphism.

Coexisting garnet and cordierite occur in several samples together with biotite and those analyzed in sample T19 give  $P = 5,5 \text{ kb}$  and  $T = 620-650^\circ\text{C}$  with

$H_2O < 1$  using the methods outlined in Appendix 9.

This pressure was confirmed independently using the geobarometer of Ghent (1976). Almandine + cordierite assemblages can be produced by the breaking of the sillimanite-biotite tie lines on increasing metamorphic temperature by the reactions:-



Reaction (4) occurs at higher temperatures and has a much steeper positive slope in P,T space than reaction (3) (Ashworth and Chinner 1978). In the Tantalite Valley aureole, temperatures probably increased with time, while pressure remained constant at about 5,5 kb; conditions which would favour reaction (4) over reaction (3) for producing the assemblage garnet + cordierite. That partial melting was possible in the aureole, can be seen from textures in psammitic hornfels, presence of high temperature mineral assemblages like cordierite plus hypersthene and migmatitic rocks and the fact that sample T19 plots near the granite solidus in P,T space (Fig 61). Furthermore, the pelitic rocks

contain no K-feldspar and no primary muscovite, and are depleted in  $K_2O$  relative to average shales. (Shaw 1956). The latter characteristic was used by Dostal (1975) and Harris (1976) as evidence for cordierite-garnet rocks being formed as residua after removing a granitic melt, a theory which is favoured by Ashworth and Chinner (1978) to explain the production of garnet-cordierite-sillimanite hornfels around granites intruded into medium-grade gneisses. Further evidence that reaction (4) has taken place in Tantalite Valley pelitic hornfels can be seen from the textural disequilibrium of <sup>touching</sup> sillimanite and biotite in these rocks.

Generally, the pelitic hornfels of Tantalite Valley resemble rocks from the main Donegal aureole in Ireland in that both have andalusite, staurolite, sillimanite and kyanite present, for which Hess (1969) postulated a pressure of greater than 4,6 kb.

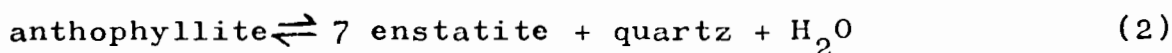
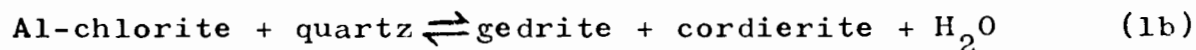
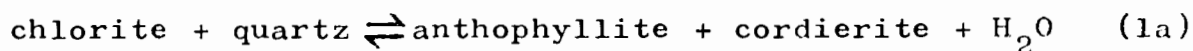
Garnet-cordierite compositions in T19 from Tantalite Valley indicate that for this rock, the pressure was near 5 kb ( $P_{H_2O} < P_{total}$ ) and the temperature near 650°C at the peak of contact metamorphism (see section 5.4 ).

#### 7.1.2 Cordierite-anthophyllite hornfels

The minerals anthophyllite, cordierite and alman-

dine are typical of the hornblende-hornfels facies (Winkler 1974) or as Reverdatto (1973) terms it, the amphibole-hornfels facies.

Univariant reactions for the formation and breakdown of orthoamphibole can be written:



These produce reaction curves which are nearly parallel to the P-axis in P,T space and are therefore well-suited for the determination of the temperature of metamorphism. Akella and Winkler (1966) have obtained equilibrium conditions for reaction (1b) at  $P_f = 2$  kb and  $562^\circ\text{C}$ , at  $P_f = 1$  kb and  $548^\circ\text{C}$  and at  $P_f = 0,5$  kb and  $530^\circ\text{C}$ . Extrapolating to  $P_f = 5$  kb gives a temperature of  $570^\circ\text{C}$  which defines the minimum temperature at which orthoamphibole can form in hornfels from Tantalite Valley. Fleming and Fawcett (1976) found that, in the system  $\text{MgO} - \text{FeO} - \text{Al}_2\text{O}_3 - \text{SiO}_2 - \text{H}_2\text{O}$  at  $P_f = 2$  kb, reaction (1a) occurs at about  $600^\circ\text{C}$  and that the upper thermal stability of chlorite + quartz is independent of the Al content and the mg ratio of the chlorite. Reaction

(2) was studied by Greenwood (1963) in the system  $\text{MgO} - \text{SiO}_2 - \text{H}_2\text{O}$ . If his results are extrapolated to  $P_f = 5 \text{ kb}$  then pure magnesium anthophyllite breaks down near  $800^\circ\text{C}$ . For conditions where  $P_f$  is less than  $P_t$ , as has been shown to be the situation in the Tantalite Valley hornfels (section 5.4), then anthophyllite will break down at a lower temperature than  $800^\circ\text{C}$  at 5 kb (Johannes 1969). Also the anthophyllite from the cordierite-anthophyllite hornfels of Tantalite Valley contains Al and has an mg ratio around 0,6 which from the work of Ravaire and Hinrichsen (1975) would indicate that it would be expected to break down nearer  $700^\circ\text{C}$  than  $800^\circ\text{C}$  at 5 kb.

Thus the temperature range of cordierite-anthophyllite rocks in the aureole of the Tantalite Valley Complex is between about  $600$  and  $700^\circ\text{C}$ .

### 7.1.3 Metabasite Hornfels

Cummingtonite which is a common constituent of Tantalite Valley metabasite hornfels, is restricted to medium-to high-temperature metamorphic assemblages formed under conditions of relatively low pressures (Leake 1972). The occurrence of garnet in metabasites may mean that metamorphic pressures were relatively high or that mg in the rocks was low

(Leake 1972). In the present area, garnet is very rare in amphibolites (see Beukes 1973; Toogood 1976) which have formed during regional metamorphism but are common in Tantalite Valley metabasite hornfels. Since the prevailing pressure was essentially the same in both domains, mg, which on average is higher in the metabasite hornfels than in the amphibolites (see Tables 44 and 48), is considered to have been the controlling factor in the development of garnet (see Leake 1972). That some hornfels are rich in iron can be seen from the development of staurolite in addition to almandine. Kushiro and Yoder (1966) experimentally determined the lower stability limit of garnet in quartz tholeiites by studying the breakdown of orthopyroxene in the presence of plagioclase to form garnet at temperatures in the range 700 to 900°C. Trial C.I.P.W. norm calculations for the metabasite hornfels show that this rock approximates a quartz tholeiite in composition, having both normative quartz and hypersthene. If a pressure of 5 kb is assumed for the contact metamorphism, then if the curve of Kushiro and Yoder (1966) is extrapolated to slightly lower temperatures, the maximum temperature at which garnet could be formed is  $\sim 670^{\circ}\text{C}$ . Liou et al (1974) showed in experiments using metabasite hornfels that the "chlorite out" isograd takes

place at about  $550^{\circ}\text{C}$  at  $P_f = 2 \text{ kb}$  and  $575^{\circ}\text{C}$  at  $P_f = 5 \text{ kb}$ . They also defined the lower boundary of the amphibolite facies as the "chlorite out" isograd. In the Tantalite Valley metabasite hornfels chlorite is only present as a secondary mineral so that metamorphic temperatures must have been in excess of  $575^{\circ}\text{C}$ .

Generally, the minerals which are found in the metabasite hornfels of Tantalite Valley are characteristic of the hornblende-hornfels or amphibolite facies and the temperature range of  $575^{\circ}\text{C}$  to  $670^{\circ}\text{C}$  (at  $P = 5 \text{ kb}$ ) which was deduced above is in agreement with the conclusion.

#### 7.1.4 Orthopyroxene-bearing hornfels

Orthopyroxene-bearing metamorphic rocks by definition belong to the granulite facies (Winkler 1974) and are termed granulites (Winkler and Sen 1973); although more specifically, orthopyroxene-bearing hornfels are grouped under the pyroxene-hornfels facies of Miyashiro (1973). The assemblage hornblende + biotite + pyroxene is common in granulites (De Waard 1965b), with hornblende and biotite being regarded as prograde minerals (eg Sen 1970; Leelanandam 1970).

In the sample TV195 from the Tantalite Valley aureole,

the minerals making up the paragenesis hornblende + phlogopite + ortho-and clino-pyroxene + plagioclase are in apparent textural equilibrium. From the phases present it is likely that a reaction of the type:-

hornblende + quartz  $\rightleftharpoons$  orthopyroxene + clinopyroxene + labradorite + magnetite + ilmenite + H<sub>2</sub>O (Sen and Ray 1971), may have occurred. Binns (1968) experimentally found that the minimum temperature at which hornblende changes to orthopyroxene is in the range 760 to 790°C, which serves as a minimum temperature estimate for sample TV195.

The assemblage cordierite + hypersthene + phlogopite with or without plagioclase was found in two samples from the Tantalite Valley aureole. Hess (1969) has estimated that cordierite-hypersthene is stable at 750°C (< 3 kb) to 820°C (5 kb). If a pressure of approximately 5 kb is assumed for the Tantalite Valley contact metamorphism, then it appears that samples TV225 and TV226 were formed at about 820°C.

From the above discussion, it is clear that temperatures were as high as 760 to 820°C in some parts of the Tantalite Valley aureole.

#### 7.1.5 Psammitic Hornfels

The parageneses H1 to H3 in Table ( 8 ) are arranged



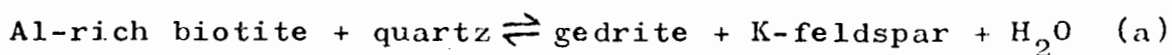
essentially in order of increasing grade of metamorphism as far as reaction (1) given below is concerned. Muscovite + quartz in the presence of plagioclase and K-feldspar (ie H1) breaks down according to the reaction: muscovite + quartz + plagioclase + K-feldspar +  $H_2O \rightleftharpoons$  melt (1)

which occurs at about  $650^{\circ}C$  (assuming a pressure of about 5 kb  <sup>$(P_{H_2O} = P_{Total})$</sup>  and the presence of plagioclase of intermediate composition) (Winkler 1970). The presence of muscovite in H1 suggests that temperatures were below  $650^{\circ}C$ , whereas its absence in H2 indicates either that reaction (1) has proceeded until all the muscovite was used up or the composition was such that it never occurred at all. With increasing temperature K-feldspar and some plagioclase will melt (reaction (1)) and biotite may react with quartz to give anthophyllite (see section 7.1.5.1). This will give the assemblage observed in sample AM30A, ie H3.

#### 7.1.5.1 Partial Melt texture in Psammitic Hornfels

In section (3.2.2.7.5) it was suggested that the texture of sample AM30A may be due to partial melting. In order to test this hypothesis, electron microprobe scans were carried out across plagioclase-quartz and quartz-quartz grain boundaries (Fig 60 ) for comparison with those made by Mehnert et al (1973) on

experimentally quenched, partially melted quartzofeldspathic rocks. The material between quartz grains gives a profile very similar to a mobilized melt between two quartz crystals in granulite which has undergone partial melting (Mehnert et al, 1973, Fig 8b) as do the results from the material between quartz and plagioclase. No concentration gradients were detected for element scans across boundaries between grains and matrix. The interstitial material is turbid and probably represents microcrystalline and altered melt. The original melt composition was probably rich in potassium, a typical incompatible element which becomes concentrated in the liquid phase. The presence of anthophyllite in sample AM30A is further evidence that partial melting has occurred since Knabe (1970) produced anatectic melts in quartz + plagioclase + biotite-bearing rocks and noted that during partial melting orthoamphibole developed according to the proposed reaction:



The K-feldspar produced enters the melt together with quartz and plagioclase.

The small degree of partial melting in sample AM30A is probably due to only local lowering of the temperature of the solidus by limited access of water into parts of the rock, eg through localized

hydraulic fracturing (Beach 1974) or migration of water along pressure gradients (Carpenter 1968). If the amount of water introduced was small, only a limited degree of partial melting could take place and it would continue only as long as water was present as a separate phase or until all of the orthoclase had disappeared (Kleeman 1965; Mehnert 1968). Quenching in situ could result if the amount of melt thus produced was insufficient for removal from the host, if the temperature was close to the solidus temperature and if the supply of water was limited.

Mehnert et al (1973) showed experimentally that a small degree of partial melting in a quartzofeldspathic gneiss, with a composition analagous to sample AM30A, takes place at temperatures of the order of 650 - 700°C. Under these temperatures and a pressure of 5.5 kb, dry melting of any muscovite present in the rock would have occurred (Fyfe 1971). This may in part explain the virtual absence of primary muscovite in the aureole of the Tantalite Valley Complex. Dry melting of biotite at a pressure of 5,5 kb occurs at temperatures of more than 750°C (Winkler 1974, p 293). This temperature serves as a maximum estimate at which biotite melted in AM30A according to reaction (a), since it is unlikely that the system was completely dry.

#### 7.1.6 Width of the Tantalite Valley Contact Aureole

The Tantalite Valley aureole is anomalously thin for contact metamorphism by a gabbroic magma intruding at pressures of approximately 5 - 6 kb (Jaeger 1957; Reverdatto et al 1970), even taking into account a pure conduction model (Jaeger 1957). However, the predominant heat transfer mechanism for aureoles around deep-seated intrusions is fluid transfer, involving both the intrusive and the country rocks (Jackson 1973). Gabbroic magma is generally low in volatiles and if the country rocks were dehydrated, as by a previous metamorphism, then there should be very little fluid available for heat transfer, resulting in the formation of a relatively narrow aureole. It is thus concluded that the thin aureole of the Tantalite Valley Complex is due to dehydration of the country rocks by the previous regional metamorphic episode. It is common in orogenic areas that contact metamorphism is superimposed upon a regional metamorphism of low- to medium-grade (Reverdatto 1973), as is suggested here.

#### 7.1.7 Chemistry of the Tantalite Valley Hornfelses

Several samples of hornfels have been analyzed for

major and trace elements (Table 44 ). These were used to calculate the classical ACF and AKF diagrams for comparison with published data. <sup>(Fig 90)</sup> Using the fields for sedimentary rocks in <sup>the</sup> ACF diagram <sup>after</sup> Winkler (1974, Fig 5 - 7) some idea may be gained as to the pre-metamorphic mineralogy of these rocks assuming essentially isochemical metamorphism. All the pelitic hornfels analyses fall within fields equivalent to clays and shales with some carbonate component. The similarity of the pelitic hornfels to shales may be seen by comparing the analyses with the average analyses of shales and pelitic rocks of Shaw (1956) given in Table ( 44 ). The major differences lie in marked depletion in  $K_2O$  and, in sample TV4, enrichment in CaO for the Tantalite Valley hornfels. The higher CaO may reflect an originally higher carbonate content in the source rock, now present in garnet. The lower  $K_2O$  suggests metamorphism has not been strictly isochemical and that there may have been a loss of more mobile elements, particularly as these rocks have undergone polymetamorphism. The relatively high copper and nickel values for sample TV4 (Table 44 ) can be ascribed to transgressive veinlets of sulphide minerals which are associated with mineralization of the metagabbro of the Tantalite Valley Complex (Moore 1975). This sample also shows a relatively

high zinc content but no zincblende has been identified. Electron microprobe analyses have shown that the spinel (hercynite) in this rock may contain up to 12% weight ZnO (other elements: MgO = 3,9%, FeO = 2,9%,  $Al_2O_3$  = 54,4%,  $Cr_2O_3$  = 0,5%). The strong affinity for zinc by hercynite has been recorded by Hollister (1969) and Loomis (1976) in contact metamorphic aureoles elsewhere. The high Y content of T19 (84 ppm) is probably due to the presence of large amounts of garnet in this sample (Leake 1972). The bulk chemistry of pyroxene-bearing hornfels sample TV226 (Table 44) when plotted in an ACF diagram<sup>(Fig 90)</sup> indicates that the rock was probably derived by metamorphism of an aluminium-enriched (chlorite alteration?) basic rock. This view is supported by the high chromium and nickel contents of this sample.

All of the analyzed metabasite hornfels samples plot in the greywacke field on an ACF diagram<sup>(Fig 90)</sup>, with the exception of sample TV13 98 (Table 44) which falls into the field of marls with 35 - 65% carbonate. Copper, nickel, cobalt and chrome in general have higher concentrations in these samples than the average greywacke (Taylor 1965). The enrichment of copper, nickel and possibly cobalt is due to the presence of veins of sulphide in the rocks. The high chrome content in all but one of

the samples ie TV13 98, indicates an original igneous parentage. If Cr is plotted against 100 Niggli mg on a diagram devised by Van de Kamp (1969), then rocks which have an igneous parentage will show a positive correlation and will fall outside the field of shale-carbonate rocks. From this diagram (Fig 83 ) it is clear that all of the metabasite hornfels samples except TV13 98, lie along the igneous trend defined by metamorphosed mafic rocks mostly from the Tantalite Valley area. Sample TV13 98 plots well away from this trend and borders on the field of shale-carbonate rocks, indicating a sedimentary parentage.

The origin of the cordierite-anthophyllite hornfels based on sample TV170 (Table 44 ) has been discussed elsewhere (Section 7.1.8).<sup>P</sup> All except the metabasite hornfels analyses of Table ( 44 ) have been plotted on a modified AFM diagram (Fig 59 ) treating biotite as the common phase as this is the only ubiquitous primary potassium-bearing mineral found in these rocks. The actual compositions of the biotites are used (from Table 43 ) so that projections through biotite differ slightly for the AFM diagram.

$$A = (Al_2O_3 - X) / (Al_2O_3 - X) + (MgO - Y) + (FeO - Z)$$

$$M = (\text{MgO} - Y) / (\text{MgO} - Y) + (\text{FeO} - Z)$$

$$F = (1 - M)$$

where X, Y and Z are the molar proportions of  $\text{Al}_2\text{O}_3$ , MgO and FeO respectively in biotite contributing to the whole rock analysis. The biotite proportions are obtained by assuming all of the  $\text{K}_2\text{O}$  in the rock is present in biotite. Also plotted on the AFM diagram are the appropriate mineral analyses where these are available. With the exception of samples TV276 and TV4 (Table 44 ) all rocks plot within their respective phase triangles or very near two phase joins. Reinhardt (1968) considered three possibilities for explaining anomalous representation of rocks and minerals in the AFM diagram:

- (a) rock specimens analyzed may have been large enough to contain more than one equilibrium assemblage,
- (b) the presence of prograde, retrograde and post-metamorphic crystallization, and
- (c) errors in chemical analyses.

These three possibilities were checked for sample TV276. The powders used for analysis were checked for the presence of garnet by oil immersion (none was found) and a second set of replicate analyses confirmed the first set. Several more thin sections



were cut and this indicated that the development of chlorite, after biotite, is both unevenly distributed through the rock and is more abundant than initially thought. If a correction is made for 13% modal chlorite (composition as in Table 43) then the analysis plots in the phase triangle cordierite-staurolite- $\text{Al}_2\text{SiO}_5$  which is in accordance with its actual mineralogy. Both andalusite and sillimanite are present but the sillimanite, which occurs as oriented laths outlining S-surfaces and as inclusions in other phases (eg biotite), is considered to be a relic mineral and not part of the stable assemblage: biotite + staurolite + cordierite + andalusite.

Sample TV276 is unusual in that it is the only pelitic hornfels sample collected in which euhedral staurolite occurs: elsewhere the staurolite is rounded and is typically found as inclusions. The reason for this may be due to the unique situation of this sample. It occurs as a sliver of hornfels between the mottled metagabbro and the metagabbro which produced the aureole (Moore 1975) where it is less likely to have been affected by later tectonic and metamorphic events in the Tantalite Valley shear zone.

The analysis of sample TV4 (Table 44) indicates

that it should plot within either the phase triangle  $\text{Al}_2\text{SiO}_5$ -garnet-cordierite or garnet-staurolite-cordierite. Unfortunately, the cordierite in this rock is too severely altered for a meaningful electron microprobe analysis. The actual mineralogy of the rock, including both staurolite and kyanite, is at variance with that predicted by the AFM diagram for an equilibrium assemblage; either kyanite or staurolite should not be present. However, the 9 cm half-core used for this analysis showed uneven development of minerals and the analysis is for the sample as a whole. Kyanite is rare (estimated 1%) and in the thin section where it occurs there is no garnet. Excluding late-stage alteration products the assemblage is biotite + staurolite + kyanite + probable cordierite (no primary cordierite survives). Elsewhere in the core, kyanite is absent and the assemblage is biotite + garnet + staurolite + cordierite (highly altered). Hence the analysis reflects this latter more abundant assemblage and the analyzed core was large enough to contain more than one equilibrium assemblage on a small scale.

Because staurolite is found commonly in the aureole but nowhere outside it is considered as an index mineral for the contact metamorphism. Kyanite either represents a relic regional metamorphic

mineral or is part of a stable assemblage which cannot be represented in terms of the AFM projection used (ie minor elements are not taken into account). Since kyanite has not been found on a regional scale (Beukes 1973; Moore 1975; Toogood 1976), the latter alternative seems preferable. The controls of kyanite development in contact aureoles are discussed in section (7.1.1).

#### 7.1.8 Origin of Tantalite Valley cordierite-anthophyllite hornfels

There are currently three main theories invoked to explain the origin of cordierite + anthophyllite assemblages:

- (i) isochemical thermal metamorphism of pre-existing basic volcanics which had undergone low temperature alteration to chlorite + quartz assemblages (Vallance 1967; Chinner and Fox 1974).
- (ii) partial melting of greywacke-type sediments yielding a refractory residua containing cordierite and anthophyllite (ie metamorphic differentiation) (Grant 1968; Lal and Moorhouse 1969).
- (iii) metasomatism of pre-existing rocks with the

mineral or is part of a stable assemblage which cannot be represented in terms of the AFM projection used (ie minor elements are not taken into account). Since kyanite has not been found on a regional scale (Beukes 1973; Moore 1975; Toogood 1976), the latter alternative seems preferable. The controls of kyanite development in contact aureoles are discussed in section (7.1.1).

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- (iii) metasomatism of pre-existing rocks with the

introduction of magnesium and iron, and loss of calcium and alkalis (Eskola 1914, 1915; Härme and Perttunen 1971).

In order to test which is most likely for the cordierite-anthophyllite hornfels of Tantalite Valley, major and trace element determinations were carried out on sample AM170 (Table 44 ).

In spite of plotting in the field appropriate to shales and clays in the ACF diagram (Fig 90 ), this analysis does not correspond to any of the usual or unusual pelitic compositions indicated by Shaw (1956) or Pettijohn (1957). The main difference is in the high MgO content, while the low values of compatible trace elements (Cr, Ni, Co and Mn) imply an original bulk chemistry outside the field for mafic-ultramafic igneous rocks, assuming isochemical metamorphism. This unusual bulk chemistry is quite characteristic of these rocks in general, whether they are from regional or contact metamorphic terranes, and it is this feature more than any other which any theory on their origin must explain.

For the Tantalite Valley cordierite-anthophyllite hornfels hypothesis (1) cannot be ruled out entirely but is considered less likely than the other alternatives because of the limited extent of these

rocks in a polymetamorphic environment. Had the original rocks been basic lavas which underwent alteration to essentially quartz + chlorite assemblages as a result of late-magmatic or diagenetic alteration, then with the later regional metamorphism one might anticipate the regional development of cordierite + anthophyllite assemblages. Furthermore, numerous amphibolites are found elsewhere in the area with no indications of cordierite and anthophyllite development.

Lal and Moorhouse (1969) discussed the origin of cordierite + gedrite assemblages in detail and rejected the hypothesis that they may form as a result of metasomatism because of the absence of replacement textures and because introduction of MgO is unlikely to be present in late-stage granitic solutions to cause metasomatism. This argument is equally applicable to the Tantalite Valley situation, but Lal and Moorhouse (1969, p 160) were dealing with metasomatism by granitic fluids when they took into account the geochemical evidence, whereas it could be argued here that the Tantalite Valley metagabbro provided a source for magnesium metasomatism (after Bugge 1943). However, there are two lines of evidence against this argument. Firstly, the rocks are found in a restricted area

(Fig 5 ) and at a distance from the actual contact. If the metagabbro had been a source for magnesium metasomatism it seems more likely that such rocks would be more common in the vicinity of the large metagabbro body in that they would be nearer the source of metasomatising fluids.

Secondly, the work of Shieh and Taylor (1969) on oxygen isotopes in pelitic rocks in an aureole surrounding a granite intrusion indicated that the outward migration of water from the intrusive into the aureole was negligible. In any metasomatic process, water would be the major transporting medium and it seems even less likely that major outward migration of water plus metasomatising ions would take place if the pluton was a gabbro.

Thus, the genesis of these cordierite + anthophyllite rocks by a process of metamorphic differentiation through partial melting is favoured. Significantly these rocks are found only at the eastern end of the igneous complex (Fig 5 ) near a large area of migmatites and early, conformable pegmatites (Moore 1975; p 243) and a small area of granodioritic igneous rocks. At the time of intrusion, the regional temperatures were already high so that the local heat from the gabbroid could increase the temperature to near the solidus for rocks with a greywacke composition. The Tantalite Valley shear

zone was already in existence at this time (Moore 1975) and if some deformation took place at this stage, then migration of water to lower pressure regions could cause local lowering of the solidus, significant partial melting and extraction of a granitoid melt. The situation of the cordierite + anthophyllite rocks near a major migmatite zone is ideal for this mechanism to have operated and their form, as discontinuous pod-like masses appears to confirm it. At the western end of the Tantalite Valley Complex, the gabbroid is in contact with the earlier mottled metagabbro, apart from a small pod of pelitic hornfels caught between the two mafic rocks, and thus there are no compositions appropriate for producing granitoid melts and cordierite + anthophyllite residua (Grant 1968).

#### 7.1.9 Retrograde Metamorphism

It is important to note that the gabbro subsequent to producing the contact metamorphism underwent complete alteration to a metagabbro (hornblende + plagioclase) which largely preserves its original igneous texture. Together with the metamorphic aureole it has acted as a single entity during later periods of deformation and metamorphism within the Tantalite Valley shear zone (Moore 1975). This



implies the later introduction of water into the gabbroid to convert the pyroxenes to amphiboles. Such an introduction may have been along shear zones in the manner suggested by Beach (1974) and could have taken place at the time that the post-uplift retrograde reactions took place in the hornfels. However, whereas conversion of gabbroid to metagabbroid has been exceptionally thorough, there is little evidence of widespread hydration of rocks in the aureole. Veins within hornfels contain tourmaline, chlorite and, in the southeast, redistributed sulphides derived from the metagabbro. These are possibly also related to this period of retrograde metamorphism and discordant pegmatite intrusion.

Even later deformation and retrograde metamorphism within the shear zone has led to the development of chlorite, sericite, pinitite, prehnite, epidote and haematite in all the rocks.

7.1.10 Summary of metamorphic conditions in the  
Tantalite Valley contact aureole.

The Tantalite Valley metagabbro was intruded into rocks of varying compositions undergoing relatively high temperature regional metamorphism and developed a narrow contact aureole with mineral assemblages characteristic of the hornblende-hornfels and pyroxene-hornfels facies of contact metamorphism (Miyashiro 1973). Temperatures in the aureole were in the range from 575 to 820°C and pressure was constant at about 5,5 kb<sup>(Fig 61)</sup> so that small degrees of partial melting could take place, provided that some water was available, giving rise to garnet-cordierite and cordierite-anthophyllite-bearing restites.

## 7.2 Regional Metamorphism

### 7.2.1 Stages of Uralitization developed in Metagabbro

Uralitization is the process whereby metamorphic or igneous pyroxene is altered to amphibole, preserving the original crystal form of the pyroxene in the process.

The Tantalite Valley metagabbro has a composition close to that of the gabbronorite (Table 45 ), and probably once contained both ortho- and clino-pyroxene. As a result of hydration metamorphism (section 7.1.9) it acquired its present mineralogy of mainly hornblende and plagioclase. The hydration was extremely thorough and only very rarely is relic pyroxene observed in the rock. The reverse situation applies to gabbronorite where uralitization is minor and is confined to the grain boundaries of pyroxene. The stages of the uralitization process can be observed in the gabbronorite and metagabbro and these bear a striking resemblance to the progressive changes noted by Piispanen and Alapieti (1977) in augite-bearing diabase and by Oliver (1951) in hypersthene-diallage bearing norite undergoing uralitization.

In the first stages, orthopyroxene is altered to the exclusion of clinopyroxene to form serpentine and uralite. Clinopyroxene and the altered

orthopyroxene are then gradually replaced from margins inward by numerous small flakes of pale green to colourless actinolitic hornblende until only a ragged "glomeroporphyroblast" of amphibole remains. Next green to blue-green hornblende begins to form along the boundaries of these glomeroporphyroblasts partly at the expense of surrounding plagioclase. This stage is responsible for giving the amphibole the appearance of being zoned and is commonly represented in samples of metagabbro. Ultimately the pyroxene and some of the neighbouring plagioclase are turned into a uniformly green hornblende which may have tiny quartz inclusions. The replacement of plagioclase has the effect of changing the texture of the rock from ophitic to equigranular.

During the change from clinopyroxene to actinolitic hornblende to hornblende, Ti, Al, Fe, Na, K and  $H_2O$  are added and Ca and Si are lost. (Piispanen and Alapieti 1977). This is in agreement with the findings of Grapes (1975) who found that the change from actinolite to hornblende involved an increase in Al, Fe, Na and K. Oliver (1951) ascribes the homogenization of orthopyroxene and clinopyroxene to hornblende of constant composition to an interchange of Ca, Mg and Fe between the two pyroxenes during alteration. He found that the chemical difference

between fresh and altered norite is so small as to be within experimental error, although he states that some water and possibly a small amount of Si and Al may have been added during the uralitization process. Piispanen and Alapieti (1977) also regard the process as being essentially isochemical but regard the loss of Si in the rock as essential in explaining the fact that uralitized rocks are commonly less dense than their unaltered equivalents. However, the common occurrence of quartz inclusions in hornblende formed by uralitization in Tantalite Valley metagabbro would indicate that Si is relatively immobile.

The water required to initiate uralitization may be present as a late-stage intercumulus phase in the crystallizing rock or it may have been introduced by deuteric action. The thoroughness of the uralitization process in the metagabbro of the Complex is probably convincing proof that large quantities of water were introduced into the system after intrusion.

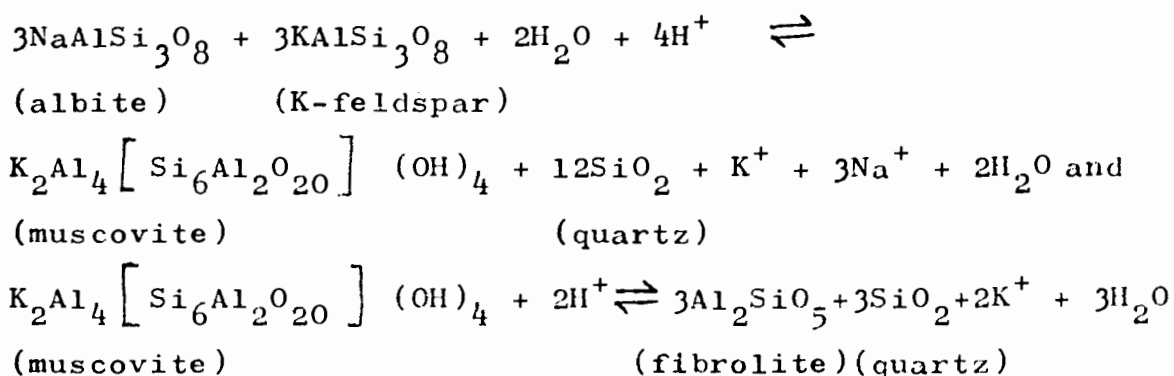
If it is assumed that the metagabbro is the uralitized equivalent of the gabbro-norite then 1.12%  $H_2O$  must have been added to the metagabbro (the difference between the average  $H_2O$  contents of metagabbro and gabbro-norite given in Table 45. Two dolerite pegmatites from Bird's River

analyzed by Eales and Booth (1974) have an average water content of 1,6%. It is assumed that this water content is taken as representative of that found in an intercumulus liquid in a gabbroic rock, because Eales and Booth (1974) regard the dolerite-pegmatite as representing expressed intercumulus liquid. To provide 1,12% H<sub>2</sub>O for uralitization between 50 and 75% of the rock must comprise intercumulus liquid. For metagabbro showing an ophitic texture this figure is unrealistically high. In the gabbro-norite the presence of trapped intercumulus liquids has been proved because of the development of orthopyroxene-opaque mineral symplectites in this rock (see section 4.3). Thus it is likely that the uralitization in the gabbro-norite is entirely due to the water from trapped intercumulus liquid since the water content of this rock indicates negligible alteration (see section 8.1.2).

#### 7.2.2 Fibrolite-bearing nodules in the Grey Gneiss

These nodules occur in Grey Gneiss along or near the axis of the Tantalite Valley Lineament and are characterized by quartz-muscovite-fibrolite assemblages, with biotite and minor plagioclase and microcline commonly occurring around the circumference.

The zoned nature of these nodules has been explained by Losert (1968) as being due to expulsion of K and Na following the breakdown of feldspar which previously occupied a place near the present centre of the nodule. Some of the K released may also have come from the breakdown of biotite which could remain in small amounts if it was originally plentiful in the matrix. Al which is relatively immobile (Carmichael 1969) tends to remain behind and combines with excess Si to form fibrolite, if the rock is within the sillimanite stability field. Losert (1968) does not say why muscovite is a common constituent of nodules generally, since de-alkalization would be expected to destroy it. An explanation can be found in the equations (rewritten here, with  $H^+$  as the mobile species) proposed by Macaudière and Touret (1969) for the nodule-forming process, ie:-



Here it is seen that in the process of feldspar breakdown to form fibrolite, muscovite is formed as an

an intermediate product. These equations also account for the reaction relationship observed between muscovite and quartz in the absence of feldspar, and the presence of inclusions of fibrolite in muscovite and quartz, in the nodules from the present area. The restriction of the latter to rocks found near the axis of the Tantalite Valley Lineament may be due to the shear zone having been responsible for introducing water, which is an essential ingredient for initiating the de-alkalization process. Alkalis released as a result of these reactions may be carried into the country rocks by migrating fluids and can be fixed in certain horizons to form rocks such as the Augen Gneiss which, according to Toogood (1976), owes its K-feldspar megacrysts mainly to blastesis. The sparse distribution of nodules in a rock can be explained by the preferential concentration of fluids along S-surfaces, such as bedding or schistosity planes. Nodule formation requires a relatively static environment so that this process occurs after the peak of metamorphism and deformation has been reached in an area (Losert 1968). The ellipsoidal form of nodules as well as a tendency to parallelism of their long axes is usually due to growth in an anisotropic medium rather than tectonic deformation. From the above discussion it is probable that the



nodules found in rocks in the core of the Tantalite Valley Lineament formed after the peak of metamorphism (M 1) and probably during the retrogressive M 2 event (see section 5.4) following the local introduction of water into feldspar-rich rocks. The liberation of alkalis during nodule formation could have induced blastesis in neighbouring rocks. \*

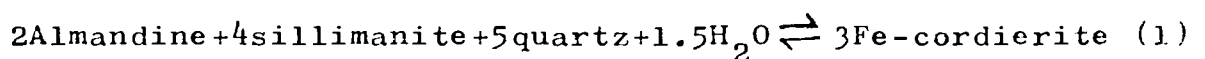
### 7.2.3 Metamorphism of the Black Gneiss

This discussion is based on 38 samples of Black Gneiss collected by the author and Dr AC Moore, and descriptions of 44 similar rocks given by Beukes (1973), Moore (1975) and Toogood (1976). They have been classified into 9 groups using their mineral assemblages (Table 14) and the relative frequency of samples occurring in each group is given in Table ( 15) as a percentage of the total.

The commonest assemblage found (Group D) consists of garnet, cordierite and sillimanite with quartz, biotite, plagioclase, K-feldspar and hercynite present in variable amounts. Such a paragenesis is especially suited for geothermometry and geobarometry (see section 5.4) but suffers from a serious disadvantage in that garnet, cordierite and biotite re-equilibrate under new, often retrograde, metamorphic conditions. Such is the case in the

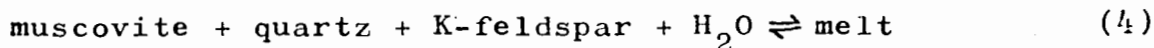
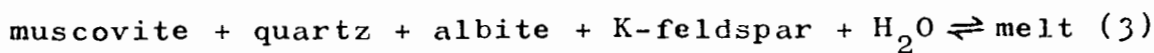
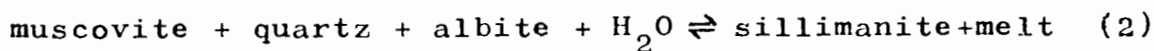
\* Note: Griffin (pers. comm. 1979) has pointed out that similar nodules found in Norwegian rocks are obviously related to strain since they are strongly elongated parallel to local fold axes and show a marked change in form relative to their position in macroscopic folds.

samples examined from the study area where results, as represented graphically on Fig ( 42 ), indicate that the retrograde metamorphic event (M 2) affected rocks over a wide area and was characterized by temperatures in the range 577 to 760°C and pressures between 4 and 5 kb. From work on co-existing pyroxenes and the presence of reaction relationships between olivine and plagioclase in mafic rocks which cooled under peak metamorphic conditions, M 1 must have occurred at 760 to 900°C and at a pressure between 5 and 5.5 kb. It is therefore envisaged that the garnet + cordierite assemblages in Black Gneiss were formed during M 1 probably by reaction (4) (see section 7.1.1) and underwent re-equilibration during M 2. During the latter event some cordierite formed at the expense of garnet by the retrograde reaction:



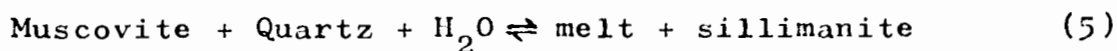
(Holdaway and Lee 1977) which occurred as a consequence of decreasing pressure after M 1 (Toogood 1976). Since natural cordierite is more magnesian than the garnet it replaces the excess iron forms a symplectic magnetite intergrowth with garnet, a texture which is commonly observed in the Black Gneiss.

The absence (except in sample DJT 376 (Table 14 )) of primary muscovite and in some samples of K-feldspar and the common association of migmatite with this rock type is explained by one of the following reactions having taken place:



Reactions (2) and (3) have been studied experimentally by Storre and Karotke (1971) and Winkler (1970) respectively, who found that they can only proceed when  $P_{\text{H}_2\text{O}}$  is greater than about 3,5 kb. At a pressure of 5 kb ( $P_{\text{H}_2\text{O}} \approx P_{\text{tot}}$ ) these reactions proceed at about 650°C (Fig 65 ). Reaction (4) is closely approximated by the work of Lambert et al (1969) who show that it only occurs at pressures above 5,5 kb and at temperatures greater than 700°C (Fig 65 ) which makes it less likely than the two previous reactions discussed. The persistence of muscovite and quartz in sample DJT 376 may be due to the absence of plagioclase and K-feldspar. Storre and Karotke (1972) showed that muscovite + quartz break down in the absence of feldspars by the

reaction:



which occurs first at about 730°C at 5,5 kb. (Fig 65). There are two generations of sillimanite in some samples; the older showing a reaction relationship with biotite, the younger occurring as stumpy prisms cross-cutting garnets.

The older sillimanite is interpreted as being one of the reactants in the cordierite + garnet + K-feldspar + melt forming reaction (see equation 4, section 7.1.1) while the younger sillimanite is regarded as being formed as a product in reactions of the type (2) and (5) given above.

The geographical distribution of the Black Gneiss can, from the above discussion, be taken as that of high grade rocks since according to Winkler (1974, p 83) 'anatexis in gneiss defines the beginning of high-grade metamorphism if  $\text{H}_2\text{O}$  pressures are larger than about 3.5 kb'. Generally the Black Gneiss is only found along or north of the axis of the Tantalite Valley shear zone in which domain primary muscovite is completely absent and migmatites are common.

The variation in composition of biotites in Black Gneiss, especially with respect to Ti content, is interpreted as being due to different degrees of reaction with sillimanite to form garnet and cordierite according to reaction (4), section (7.1.1). It is likely that Ti in the form of ilmenite was liberated during such a reaction. (see Table 16)

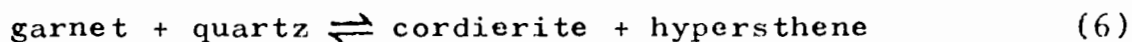
The sole aluminosilicate found in samples of the Black Gneiss is sillimanite, which serves as an indication of minimum temperature conditions ie  $500^{\circ}\text{C}$  if the triple point of Holdaway (1971) is used.

Generally green hercynite is a common accessory mineral in the Black Gneiss. The fact that it occurs in contact with sillimanite and quartz is indicative of high temperatures of metamorphism according to Richardson (1968) who experimentally investigated this assemblage, with  $P_{\text{H}_2\text{O}} = P_{\text{total}}$ , and found that it only occurs at temperatures greater than  $780^{\circ}\text{C}$ . This is in agreement with the work of Holdaway and Lee (1977) who found that at  $P_{\text{H}_2\text{O}} = P_{\text{total}}$  hercynite + sillimanite + quartz is stable above  $768^{\circ}\text{C}$ . If  $P_{\text{H}_2\text{O}}$  is less than  $P_{\text{total}}$  then this assemblage will form at lower temperature.

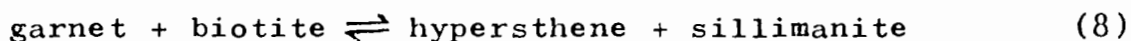
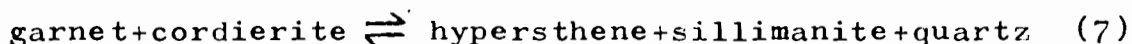
Large plagioclase grains occurring in sample KK10D (Table 16) are unusual in that they have needles of

sillimanite aligned parallel to the  $\{001\}$  and abundant inclusions of a dark green mineral with the optical properties of hercynite, filling the grid (Fig 3<sup>4</sup>G). Where spinel is least concentrated it is possible to see that it occurs as fine needles which are oriented parallel to the main cleavage directions in plagioclase. Similar textures to these were described by Wass (1973) for spinel in plagioclase from alkali basaltic rocks from the Southern Highlands of New South Wales. She attributed this phenomenon to exsolution of spinel from plagioclase as a result of a drop in pressure. It appears that under conditions of relatively high pressure plagioclase can incorporate significant FeO and MgO into its lattice, which under decreased pressure conditions diffuse together with Al, resulting in spinel exsolution. This loss of Al in the plagioclase also results in removal of calcium. To exsolve sillimanite from plagioclase in addition to spinel would probably require a large release of Al from the plagioclase and possibly introduction of Si from outside. Such a process might be expected to occur if calcic plagioclase re-equilibrated under lower grades of metamorphism (M 2?) to form relatively sodic plagioclase.

The assemblage cordierite + hypersthene (Group B), which forms by the continuous reaction:



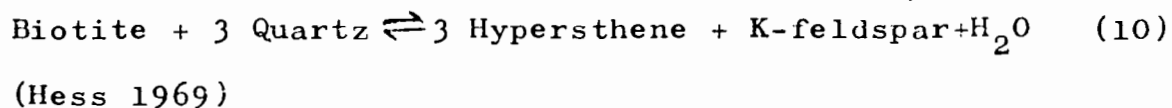
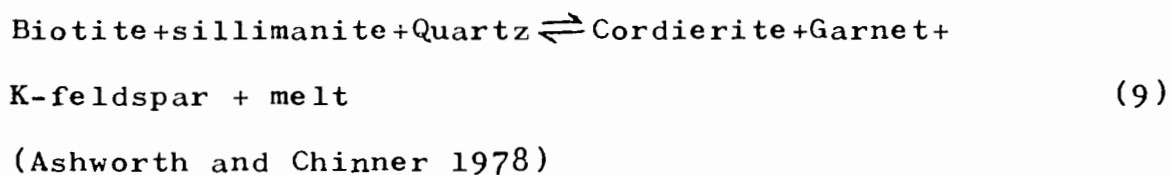
(Berg 1977), is according to Hess (1969) stable from 750° - 820°C between 3 and 5 kb. The presence of sillimanite in contact with hypersthene in sample KK6 indicates that reactions of the type:



may have occurred, although garnet is completely absent from samples belonging to Group B. Hess (1969) has estimated that the assemblage hypersthene + sillimanite is stable at pressures and temperatures greater than 4,5 kb and 790°C respectively, whereas Hensen and Green (1973), on the basis of experimental data on the system  $\text{MgO-FeO-Al}_2\text{O}_3\text{-SiO}_2$ , put the position of reaction (7) at between 12 and 9.5 kb and 850 and 1050°C, but found that natural hypersthene-sillimanite-quartz assemblages are inconsistent with their experimental data. They attributed this to the ability of natural aluminous hypersthene to incorporate some ferric iron,

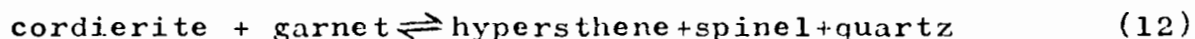
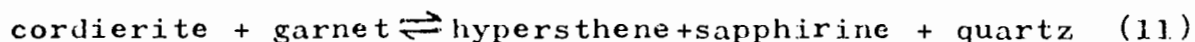
thus displacing reaction (7) to lower P and T, possibly to around 900 to 950°C at 9 kb. In hypersthene from sample KK6,  $\text{Al}_2\text{O}_3 = 5.66 \text{ wt\%}$  and  $\text{Fe}_2\text{O}_3 = 3.62\%$ , but it is not known whether the assemblage found in this rock is compatible with physical conditions prevailing during M 1 which have been deduced elsewhere.

Garnet does occur together with cordierite and hypersthene in samples from Group A and the assemblage listed in A (1) in Table ( 14 ) was probably derived by a combination of the reactions:



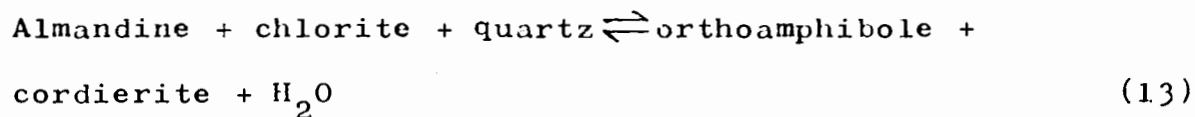
based on the fact that for this rock sillimanite occurs as inclusions in cordierite and is not in contact with hypersthene and hypersthene is altered retrogressively in part to biotite. It is tempting to ascribe the assemblage listed as A (2) in Table ( 14 ) to the reactions of Hensen and Green (1973) given as:





were it not for the absence of quartz. The assemblage made up of the minerals given in equations (11) and (12) defines an invariant point at 7,5 kb and 1 125°C (Hensen and Green 1973, Fig 1). According to Newton (1972) the presence of cordierite and corundum occurring together as they do in samples GBW54 and GBW590 (Table 14 ) is an indication of  $\text{SiO}_2$  - undersaturation. For a full discussion of assemblages with sapphirine see section (7.2.3.1).

The assemblage of cordierite + garnet + orthoamphibole (Group C) occurs in rocks that are richer in  $\text{FeO}/(\text{FeO}+\text{MgO})$  than rocks containing only cordierite and orthoamphibole (Lal and Shukla 1975). These rocks are characteristically depleted in  $\text{K}_2\text{O}$  so that the identification of K-feldspar in sample GBW292 (Table 14 ) by Beukes (1973) is puzzling. According to Lal and Shukla (1975) the assemblage cordierite + garnet + orthoamphibole forms according to the reaction



with chlorite disappearing at higher grade leaving an assemblage consisting of the products and two of

the reactants namely almandine and quartz. Sillimanite and orthoamphibole together are incompatible if garnet and cordierite are co-existing stable phases in the same rock as can be seen if the minerals are represented on an AFM diagram. The sillimanite in sample K101 is found only as inclusions in garnet.

#### 7.2.3.1 Sapphirine-bearing assemblages in Black Gneiss

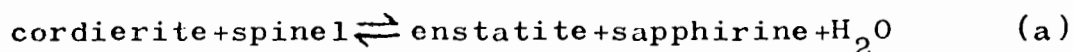
Sapphirine occurs predominantly in MgO-rich rocks, relatively poor in FeO, CaO, SiO<sub>2</sub> and with intermediate to high Al<sub>2</sub>O<sub>3</sub> contents, (Mg-rich pelitic or mafic rocks) which have been subjected to high temperatures and probably medium to high pressures as well (Seifert 1974). It has been fairly widely reported from high-grade regional metamorphic terrains but is virtually absent from high-grade pelitic hornfelses.

Beukes (1973) recorded the presence of sapphirine in two samples of Black Gneiss collected in the Kumkum Massif. Since all ferromagnesian phases which co-exist with sapphirine have high Mg/Fe ratios (Seifert 1974) it is certain that the biotite which Beukes lists as occurring with sapphirine in his samples is actually phlogopite and similarly

that orthopyroxene is probably bronzite or enstatite and garnet is pyrope-rich. Furthermore, all references found for the assemblage cordierite + spinel + corundum (see Seifert 1974) describe the cordierite to be magnesian and the spinel to be hercynitic. Thus the assemblages recorded by Beukes (1973) can be rewritten as:

1. Phlogopite+cordierite+orthopyroxene+sapphirine+spinel (GBW51), and
2. Phlogopite+Mg-cordierite+garnet+plagioclase+hercynite+sapphirine+corundum+orthopyroxene (GBW590)

Beukes gives no details of textural relations or the exact locations of these samples, however, the assemblage in (1) suggests that the reaction

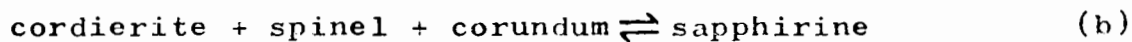


has taken place. Seifert (1974) experimentally established that the stable tie-line enstatite + sapphirine appears above 3.5 kb and at temperatures greater than 765°C in the water-saturated pure MgO-Al<sub>2</sub>O<sub>3</sub> - SiO<sub>2</sub> - H<sub>2</sub>O system. For a rock containing the assemblage sapphirine + cordierite + bronzite + phlogopite + spinel, Clifford et al (1975b) estimated a temperature close to 1 000°C, using the 7 wt%

$\text{Al}_2\text{O}_3$  in orthopyroxene-isopleth of Anastasiou and Seifert (1972), and a pressure between 6 and about 7 kb.

Thus the assemblage given in (1) probably formed at temperatures of around  $1\,000^\circ\text{C}$  and pressures greater than 3,5 kb but less than 7 kb. The equilibrium curve for reaction (a) is nearly parallel to the T-axis in  $P_{\text{H}_2\text{O}} - T$  space; this and the fact that the formation of sapphirine and enstatite involves a large volume decrease for the solids indicates that reaction (a) is favoured mainly by an increase in pressure. This reaction is only terminated at higher temperatures by the incoming of melt. (Seifert 1974).

The interpretation of assemblage (2) is more difficult because of the conflict between experimental results and what is actually observed in nature. According to Seifert (1974) the assemblage cordierite + spinel + corundum is stable at pressures below 300 bars and temperatures below  $600^\circ\text{C}$ , after which it breaks down to form sapphirine according to the reaction



However, he found at temperatures below  $750^\circ\text{C}$  at 1 kb  $P_{\text{H}_2\text{O}}$  that no reaction could be detected, and attributed this to slow reaction rates of components.

He showed that (b) has a steep negative slope in  $P_{H_2O} - T$  space and that the formation of sapphirine by this reaction involves a rather large volume decrease which implies that increasing P as well as T is necessary for its formation. The comparative abundance in nature of cordierite + spinel + corundum and the wide range of P and T over which the product of the mutual reaction of these phases, ie sapphirine, is stable, suggests that the experimental determinations are at fault. Seifert (1974) states that the assemblage cordierite + spinel is stable up to 3.8 kb in the pure system, which can be taken as a minimum pressure for the formation of assemblage (2). Furthermore, the assemblage garnet + cordierite + hypersthene found in assemblage (2) is stable at pressures from 3.8 kb to 7 kb and temperatures from 700 to 900°C (Berg 1977).

The common origin proposed for sapphirine and/or corundum-bearing pelitic rocks is that they are restites formed by partial fusion of shales (Seifert 1974; Clifford et al 1975b) either by magmatic rocks or by high-grade metamorphism.

In conclusion, it seems likely that the sapphirine rocks in the present area were formed under conditions of increasing P and T at pressures of between

3,8 and 7 kb and temperatures of around 1 000°C. Such conditions were probably met where Black Gneiss was in direct contact with intrusive mafic rocks, such as the Kumkum gabbro-norite. Conditions along these contacts were suitable for partial melting to take place leaving sapphirine-bearing restites where compositions of the host rock were suitable.

#### 7.2.3.2 Chemistry of the Black Gneisses

Major element analyses exist for whole rock samples of the Black Gneiss and trace element analyses for three of these (Table 44 ). If these analyses are plotted on a conventional ACF diagram they all lie in the field of shales which may contain minor carbonate (Fig. 90 ). However, compared with the composition of the average shale and pelitic schist given by Shaw (1956), some or all of the Black Gneiss samples show depletion in Si, Na, K and Ca and enrichment in Ti, Al, Fe and Mg, so that they could not have been derived by isochemical metamorphism of rocks of pelitic composition. Two alternatives remain:

- (a) either they were formed by metasomatic introduction of Mg and Fe with the simultaneous removal of Na and K, or
- (b) they are the solid residue after the extraction

of an anatectic melt.

The most abundant development of Black Gneiss in the area occurs in the north-west of the Kumkum Massif where mafic intrusives are absent and the only intrusives found are granitic. Metasomatic introduction of Fe and Mg is therefore an unlikely mechanism for the formation of garnet-cordierite-sillimanite gneisses, a conclusion which is supported by the lack of replacement textures in the Black Gneiss. The latter has been formed under conditions of high grade metamorphism with the formation of migmatites resulting from muscovite breakdown and the formation of the divariant assemblage garnet+cordierite+sillimanite+quartz. (See section 7.1.1). Thus the hypothesis that these rocks are the residue after removal of an anatectic melt is favoured. Dostal (1975) has observed that Al, Mg and Fe are enriched and Si, Na, and K are depleted in garnet + cordierite + sillimanite-bearing hornfelses occurring around the Loon Lake pluton, in southeastern Ontario, compared with biotite gneiss in the same area. He showed that the chemistry of the hornfelses favours them being residual after the removal of a granitic melt from biotite gneiss.

According to Lal and Moorhouse (1969) calcium tends to increase in the residue of anatexis so that depletion in Ca in Black Gneiss compared to the

average shale and pelitic schist indicates that the parental material was low in Ca.

The scarcity of garnet-cordierite-anthophyllite-bearing Black Gneiss compared to the commonest variety of Black Gneiss (Group D), both of which formed as a result of partial melting, can probably be explained by the necessity for a relatively low K, Ca and Al content in the original rock and almost complete removal of an anatectic melt before orthoamphibole-bearing Black Gneiss can form (Lal and Moorhouse 1969).

The elements Cr, Zn, Ba and Zr are enriched and Rb, Sr and Cu are depleted in Black Gneiss compared to the average values given for shale by Turekian and Wedepohl (1961) and Taylor (1965). The behaviour of Cr and Ba may be due to the tendency for Cr to increase in garnet and biotite (Engel and Engel 1960) and Ba to increase in feldspars with increasing grade (Heier 1960). The enrichment in Zn and Zr may be due to the fact that these elements occur mainly in the refractory phases, spinel and zircon respectively. It is well known that Cu is highly mobile during metamorphism and that both Sr and Rb are associated with feldspars which tend to become incorporated into an anatectic melt.



For Black Gneiss samples muscovite is absent and either K-feldspar or biotite must be used as the point of projection to represent bulk rock analyses on an AFM diagram. The calculation scheme used if K-feldspar is present is:

$$A = \frac{[Al_2O_3] - [K_2O]}{[Al_2O_3] - ([K_2O] + [FeO] + [MgO])}$$

$$M = \frac{[MgO]}{([MgO] + [FeO])}$$

where the value in brackets refers to the molar proportions of the respective oxides. For projection through biotite see section (7.1.7). Two separate AFM diagrams are given: Fig ( 63 ) is for the projection through K-feldspar and Fig ( 64 ) is for the projection through biotite. Where available, the appropriate mineral analyses are represented.

Only one bulk analysis is available from the Sandfontein area (ie sample GBW98) and this plots in the field defined by garnet-cordierite-sillimanite (Fig 64) which is consistent with its paragenesis. The mineral analyses for garnet, cordierite and biotite are from Black Gneiss sample A5, (Fig 3) which has been used because of the similar modal composition to GBW98.

Samples from the Kumkum Massif lacking in K-feldspar are also represented in Fig ( 64 ). Sample GBW302 plots within the field sillimanite-garnet-cordierite, defined by the mineral analyses from A5, while sample K101, as expected, falls in the field cordierite-garnet-gedrite. Although no bulk rock analysis is available for KK6, it would be expected to lie somewhere along the tie-line between hypersthene and cordierite (see Fig 59 ).

In Fig ( 63 ) rocks from the Kumkum Massif containing sillimanite + garnet + cordierite + biotite, in addition to K-feldspar and quartz, plot in the field sillimanite - garnet - cordierite. Sample GBW241 lies in the field defined by sillimanite - garnet - biotite (garnet and biotite compositions are taken to be the same as for KK7) which is in accord with the observed mineralogy of GBW241. From Fig ( 63 ) it is clear that the compatible mineral associations are sillimanite - garnet - biotite (Group E, Table 14 ), sillimanite - cordierite - biotite (Group F, Table 14 ), garnet - cordierite - sillimanite or garnet - cordierite - biotite (Group D, Table 14 ), (garnet) - cordierite - biotite - orthoamphibole (Group C, Table 14 ), cordierite - garnet - hypersthene - (biotite) (Group A, Table 14 ), cordierite - biotite-hypersthene (Group B, Table 14 ), Garnet-biotite (Group

H, Table 14 ), cordierite - biotite (Group I, Table 14 ) and sillimanite - biotite (Group G, Table 14 ).

Two of the assemblages above (Group A and Group C) contain four co-existing ferromagnesian minerals. On the basis of petrographic observations and reactions postulated garnet in Group C and biotite in Group A are taken to represent unstable or relic minerals. From Fig ( 63 ) it can be seen that there is a variation in the cordierite - garnet - biotite phase triangles for different rocks. Reinhardt (1968) interpreted this to mean that chemical equilibrium was achieved in the rocks under different external conditions (P, T and chemical potential of mobile components) of metamorphism. The retrograde metamorphic overprint on the samples of Black Gneiss examined makes it difficult to pin-point the exact cause of the shift in three-phase triangles.

From the theory outlined in Thompson (1976a) the effect of varying P, T and the chemical potential of a mobile component (eg  $X_{H_2O}$ ) on the three phase triangle defined by garnet - biotite - cordierite can be predicted. For the case where P increases (T,  $X_{H_2O}$  constant), the co-existing phases making up the triangle become more magnesian and the triangle will

shift to the right in Fig ( 63 ). Decreasing  $X_{H_2O}$  (P, T constant) will have the same effect. If T is increased (P,  $X_{H_2O}$  constant) the three-phase triangle shifts to the right in Fig ( 63 ) and contracts at the same time so that there will be an apparent rotation of the lines on going to higher temperatures. This is because on increasing the temperature, the co-existing phases become more magnesian and the distribution coefficients involving these phases tend towards unity. The configuration of tie-lines drawn in Fig ( 63 ) favours T as being the main factor in changing the position of the three-phase triangles. However, the correlation between increasing temperature (as calculated from  $K_{D}^{Mg-Fe^{2+}}_{ga-cd}$ ) and the shift in three-phase triangles in Fig (63) is not as perfect as would be expected if this were the case. The order of increasing temperature calculated using  $K_{D}^{Mg-Fe^{2+}}_{ga-cd}$  is 2, 6, 3, 1, 4 (using the nomenclature for samples as given in Fig(63) while the trend of decreasing  $FeO/MgO$  is 6, 2, 1, 3, 4 using almandine compositions. The fact that the garnet, biotite and cordierite, from hornfelsic Black Gneiss sample K35, have the lowest  $(FeO)/(MgO)$  ratios of all the analyzed garnet-cordierite-biotite assemblages and that this rock formed at higher temperatures than the "normal" Black Gneiss (eg sample KK7) indicates that variation in tem-

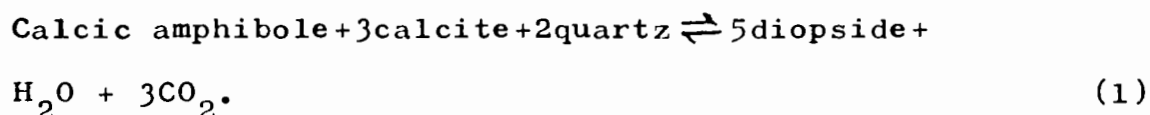
perature of metamorphism was probably significant. Reinhardt (1968) has shown that oxidation of a rock can also have the effect of changing the  $(\text{FeO})/(\text{MgO})$  ratio of minerals to lower values as a result of the increased production of magnetite at the expense of ferromagnesian silicates. Haematite occurs as a rare secondary mineral in Black Gneiss, and the predominance of ilmenite over magnetite in samples indicates that  $P_{\text{O}_2}$  was probably low so that this effect could hardly have been significant. During M 2 a decrease in pressure occurred as seen by cordierite rims around garnet, which could also partly account for the geometry of the three-phase triangles.

#### 7.2.4 Metamorphism of Calcareous Rocks

##### 7.2.4.1 Calc-silicates

Generally calc-silicates and marbles are rare in the Kumkum and Tantalite Valley areas. On the strength of 44 samples of calc-silicate collected by various workers ie Beukes (1973), Moore (1975), Toogood (1976) and the present author, the calc-silicates have been divided into six main classes on the basis of their mineral assemblages (Table 3 ) with the relative frequency of occurrence of each class being given in Table ( 4 ) as a percent.

The parageneses and metamorphic conditions which can be deduced from them are discussed below, assuming a metamorphic pressure of around 5 kb. The assemblage diopside + hornblende + quartz + plagioclase with or without the presence of microcline, sphene and epidote (Class B) is the one which is found most frequently in the Tantalite Valley and Kumkum areas. Diopside occurs in exactly 75% of the total number of samples, mostly in the absence of calcite. Diopside occurring without calcite is a characteristic assemblage of the middle to upper part of the amphibolite facies of metamorphism (Miyashiro 1973, p 260). The most likely reaction which can produce diopside considering the minerals associated with it and especially the assemblage diopside + tremolite + plagioclase + quartz + calcite in sample DT1358 recorded by Toogood (1976) (see Table 3) is:

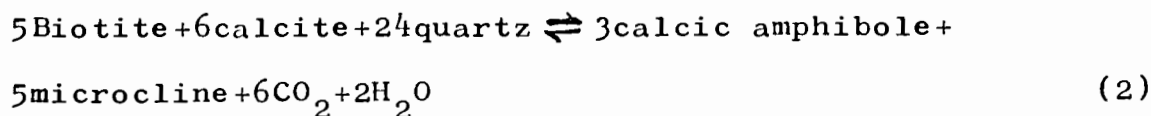


(Metz 1970, Winkler 1974, p 119; Ferry 1976).

This reaction is commonly observed in metamorphosed impure limestones and siliceous dolomites (Ferry 1976). The fact that calcite is not found in most of the diopside-bearing rocks is indicative

of the metamorphic temperatures at Kumkum and Tantalite Valley having exceeded the temperatures for the reaction. Slaughter et al (1975) investigated reaction (1) in the system  $\text{CaO-MgO-SiO}_2\text{-H}_2\text{O-CO}_2$  at  $P_{\text{total}} = 5 \text{ kb}$  and found that it occurs over a temperature range of 600 to  $650^\circ\text{C}$  depending on whether  $X_{\text{CO}_2}$  is low or high.

The presence of microcline together with hornblende and quartz with or without biotite in some samples (eg TV335, DT1304, Table 3) suggests that the reaction proposed by Hewitt (1973) ie:



may have occurred. According to Ferry (1976) this reaction occurs at lower metamorphic grades than reaction (1). In the two samples named the high potassium content as manifested either by high biotite or high K-feldspar content is evidence that these specimens were derived from sediments.

Rocks containing hornblende and plagioclase alone (ie Group F) first form in the lower part of the amphibolite facies and are stable up to the middle to upper part of the amphibolite facies where diopside gradually forms at the expense of calcic

amphibole according to reaction (1). If calcite is absent and reaction (1) cannot occur then hornblende and calcic plagioclase will adapt to higher grades of metamorphism by increased Tschermak's substitution for hornblende and exchange of NaSi for CaAl in plagioclase (Ferry 1976).

The occurrence of the assemblage epidote + quartz without the presence of garnet in many of the samples in Table ( 3 ) suggests that these specimens were formed at temperatures below 650°C at  $P_{\text{total}} = 5 \text{ kb}$  (see later discussion on equation (4)).

Scapolite and sphene are always found in rock samples of class A with or without diopside, garnet, calcite, plagioclase, hornblende, quartz and microcline. Scapolite-bearing rocks are only found in the Kumkum area. Beukes (1973) has reported the presence of muscovite but this is probably secondary after scapolite and/or feldspar. Scapolite is stable in both medium- and high-stage metamorphism (Winkler 1974, p 125) and the presence of assemblages with scapolite, calcite and plagioclase suggests that a reaction similar to

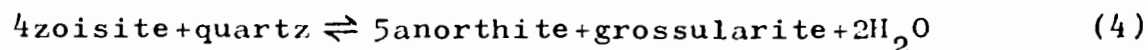
$3 \text{ Anorthite} + \text{calcite} \rightleftharpoons \text{scapolite (meionite end member)}$  (3)  
occurred. (Fyfe et al 1958; Ferry 1976).

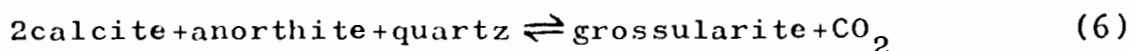
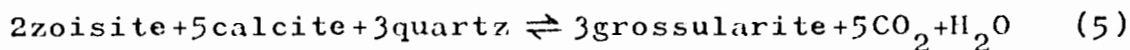
Unfortunately no experimental data are available for this reaction. (Fyfe et al 1958, p 160-161)



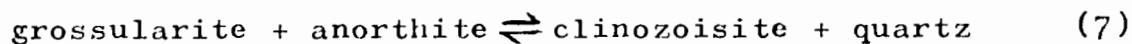
suggested that scapolite is stable up to  $800^{\circ}\text{C}$  if  $P_{\text{H}_2\text{O}}$  is high. Orville (1975) found that at  $750^{\circ}\text{C}$  and 4 kb, in the presence of excess calcite, scapolite + plagioclase is more stable than plagioclase over a wide range of bulk-equivalent An-contents (see Orville 1975, Fig 9). Further, Ferry (1976) found in an area in south-central Maine that only areas adjacent to granodiorite magmas reached temperatures high enough for scapolite to form. These granodiorites were emplaced during regional metamorphism when  $P_t \cong 3,6$  kb. Extrapolating the granite heating curves in Fig 8 - 2 from Winkler (1974) to these higher pressures gives a temperature of close to  $800^{\circ}\text{C}$  along the contact of the country rock intruded by granite magma. The high temperatures needed to stabilize scapolite justifies the conclusion that reaction (3) occurred in calc-silicates of suitable composition during the peak of metamorphism (M 1).

Grossular-rich garnet occurs in about 20% of the samples collected (ie Group C and E, Table 3 ). Considering the minerals present in these rocks the following grossularite-forming reactions are possible:



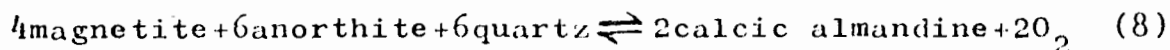


It is commonly observed in calc-silicates from Kumkum and Tantalite Valley that grossularite and anorthitic plagioclase may have broken down during retrograde metamorphism to form light green clinozoisite and quartz, the latter occurring as rounded inclusions in plagioclase. The clinozoisite has nucleated on primary yellow-green zoisite and along garnet rims. Thus the retrograde reaction can be expressed as:

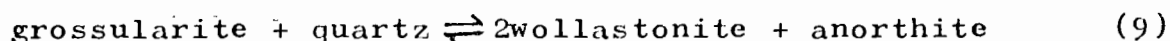


This textural evidence suggests that reaction (4) could have been responsible for the formation of grossularite in these rocks during prograde metamorphism.

Reactions (4) and (5) occur at about 650°C and low  $X_{\text{CO}_2}$  at  $P_{\text{total}} = 5$  kb (Holdaway 1972; Storre and Nitsch 1973). Reaction (6) takes place at 700 to 725°C and low  $X_{\text{CO}_2}$  at  $P_{\text{total}} = 5$  kb (Mukherjee and Rege 1972). If the garnet is calcic almandine, as reported from sample DT99A (Toogood 1976), then it can form according to the reaction:



which probably occurs under medium-grade metamorphic conditions (Winchester 1972; Tanner 1976). Textural evidence for reaction (8) having taken place was not observed by Toogood (1976). An estimate of the upper limit of prograde metamorphism can be made by considering the stability of grossularite. Boettcher (1970) has shown that grossularite breaks down in the presence of quartz at about  $725^{\circ}\text{C}$  and low  $X_{\text{CO}_2}$  at  $P_{\text{total}} = 5 \text{ kb}$  according to the reaction:



Since the product assemblage has not been recognised from the study area temperatures as high as  $725^{\circ}\text{C}$  could not have affected grossularite + quartz bearing assemblages. Where quartz is absent Boettcher (1970) has shown that grossularite will not break down until about  $875^{\circ}\text{C}$  at  $P_{\text{total}} = 2 \text{ kb}$ . The stability of garnet in the absence of quartz in the samples examined implies that temperatures could not have been as high as this.

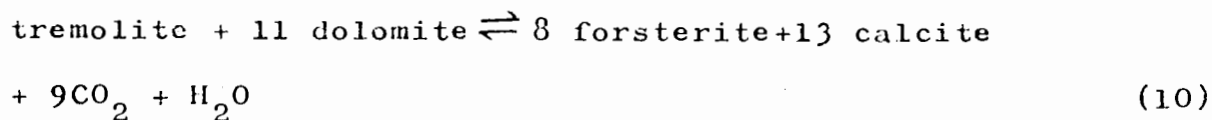
In summary, the calc-silicates of the Tantalite Valley area have been metamorphosed at grades varying from medium to high. It is unlikely that

temperatures were higher than 800°C during the peak of metamorphism in the area and in fact must have been between about 600 and 750°C, based on the prograde mineral assemblages examined in samples from the entire area. Mineral assemblages in some rocks indicate that the maximum temperature reached during retrograde metamorphism was about 650°C. Mineral reactions in calc-silicates took place under conditions of low  $X_{\text{CO}_2}$ , both during the peak of metamorphism and during the retrograde events.

#### 7.2.4.2 Marbles

This discussion is based on ten samples from the study area described by Beukes (1973) and Moore (1975). On the basis of mineral assemblages they have been divided into three groups given in Table ( 5 ).

Group A comprises half of the samples and is characterized by the assemblage forsterite + calcite, which probably formed by the reaction:

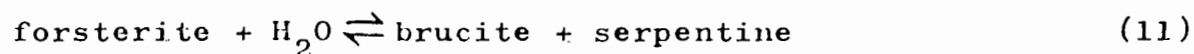


which occurs between 640 and 690°C, corresponding to

$X_{\text{CO}_2} = 0,1$  and  $0,6$  respectively, at  $P_f = 5$  kb (Metz 1973). The marble layers in the present area occur as rare, thin bands in Grey Gneiss sensu stricto and thus  $X_{\text{CO}_2}$  was probably significantly diluted by water introduced from the latter. Thus it is likely that reaction (10) occurred at temperatures of around  $640^\circ\text{C}$  although the fact that no tremolite is present probably indicates that final temperatures were much higher than this.

Spinel, chondrodite, brucite and humite are commonly observed in forsterite + calcite-bearing marbles from the Central Alps. These rocks were formed under conditions of high temperature and intermediate pressure. (Trommsdorff 1966; Trommsdorff and Schwander 1969). The same accessory minerals occur in forsterite marbles from Tantalite Valley and Kumkum which is taken to imply that these rocks were formed at roughly the same grade as those from the Alps.

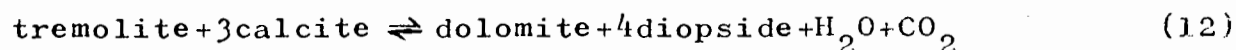
Brucite only occurs in the forsterite marbles (Group A, Table 5) and textures indicate that it probably formed as a result of the retrograde breakdown of olivine according to the reaction:



This reaction takes place at a temperature of about 450°C over a large range of pressures (Deer et al 1971, Fig 79).

The formation of diopside in marbles of Group B (Table 5) was probably according to reaction (1) which occurs at about 650°C for  $P_{\text{total}} = 5 \text{ kb}$ .

In one sample, GBW288 (Group C, Table 5) calcite and tremolite are stable, indicating that the reaction:



(Metz and Trommsdorff 1968) has not taken place. This reaction proceeds between 625 and 700°C at high  $X_{\text{CO}_2}$  assuming  $P_{\text{total}} = 5 \text{ kb}$  (Metz 1973).

The assemblage calcite + tremolite in the absence of quartz is generally stable to higher grades than the same assemblage with quartz which reacts according to equation (1).

From the above it can be concluded that the T -  $X_{\text{CO}_2}$  conditions at  $P_{\text{total}} = 5 \text{ kb}$  deduced from marble assemblages are compatible with those derived from calc-silicates.

Prograde metamorphic temperatures were between 640 and 690°C and  $X_{\text{CO}_2}$  was low. Retrograde metamorphism

occurred with temperatures of around 450°C.

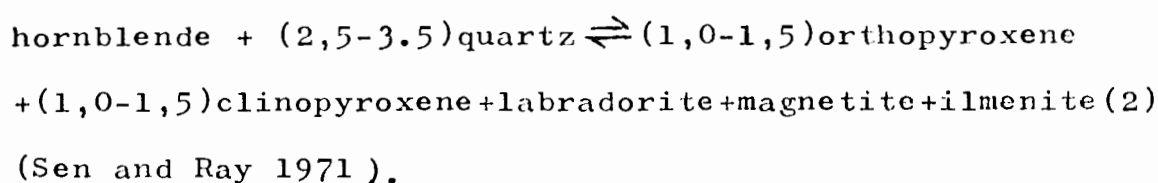
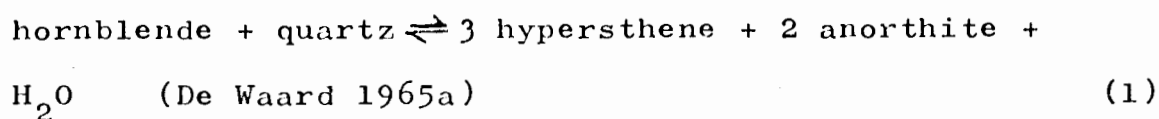
#### 7.2.5 Metamorphism of Granolites

Granolite, as defined by Winkler and Sen (1973), is a rock diagnostic of the granulite facies. Rocks from this facies are characterized by the presence of hypersthene (Winkler 1974).

The conditions necessary for the formation of granulites ie high temperatures and  $P_{H_2O} \ll P_{total}$  are met only in the deeper parts of the crust (Winkler 1974). This explains why granulites are mainly reported from deeply eroded Precambrian shield areas and cores of younger orogens (Katz 1968, Oliver 1969).

Rare outcrops of granulite are restricted to the area north of the axis of the Tantalite Valley Lineament. The Kumkum granulite is characterized by the presence of labradorite, hypersthene, hornblende and opaque minerals which may be accompanied by quartz, clinopyroxene and biotite (Table 18). The minerals are all in textural equilibrium and can be described as having an equigranular polygonal texture (Moore 1970). Hydrous minerals, (eg hornblende and biotite), are generally common in granulites and are usually taken to be prograde. (eg Binns 1965, Sen 1970; Ramaswamy and Murty

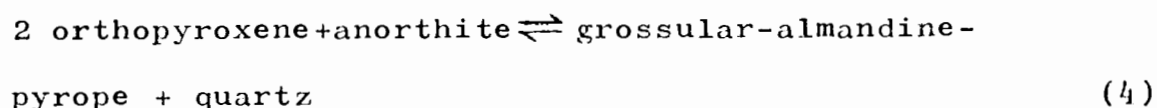
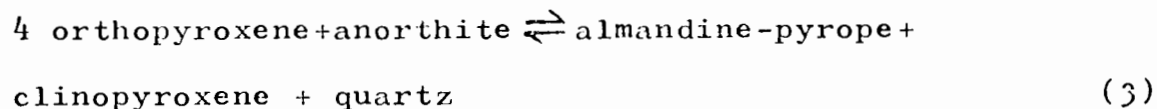
1974). From the minerals present in the Kumkum granulite it is feasible that one-or two-pyroxene-bearing assemblages were formed according to the reactions:



These mark the transition between the amphibolite and granulite facies. When quartz is used up in these reactions hornblende will be stabilized (Sen and Ray 1971). Under conditions of  $P_{\text{total}} = P_{\text{H}_2\text{O}}$ , Binns (1968) experimentally showed that the dehydration of Broken Hill hornblende to form orthopyroxene occurs at a minimum of  $760^\circ\text{C}$  but probably nearer  $790^\circ\text{C}$ . For the case where  $P_{\text{total}} \gg P_{\text{H}_2\text{O}}$  the reaction temperature will be lower but not less than  $700^\circ\text{C}$ . Hewins (1975) confirmed the accuracy of these values by applying pyroxene geothermometry (Wood and Banno 1973) to hornblende-bearing granulites and charnockites in the literature, for which he obtained temperatures in the range  $780 - 860^\circ\text{C}$ . Pyroxenes from Kumkum granulite sample K58 gave



temperatures of 790°C using the Wood and Banno method (1973) and 760°C using the Wells (1977) method. This places it in the temperature range determined for hornblende-bearing granulites in general. According to Hewins (1975) hornblende breakdown is probably complete by the time the granulite has reached about 860°C. Thus the fact that all Kumkum granulite samples contain hornblende implies that temperatures were not greater than this. An estimate of the maximum pressure at which the Kumkum granulite formed can be obtained by considering the reactions:



which were studied experimentally by Kushiro and Yoder (1966) and Green and Ringwood (1967b). The results of the former authors are more applicable to the present study since Kushiro and Yoder (1966) performed their experiments at temperatures of between 700 and 900°C which includes the temperature range envisaged for the formation of hornblende-bearing granulites. Since garnet is not found in

any of the Kumkum granulites and orthopyroxene and plagioclase are stable, the pressures at which these rocks formed must have been less than 7 kb at about  $770^{\circ}\text{C}$ , derived from pyroxene geothermometry and using the experimental work of Kushiro and Yoder (1966). An estimate of pressure can be obtained from Fig ( 52 ) after Raase (1974), where hornblende analyses from granulite sample K58 plot on the low pressure side of the 5 kb control line. The minimum pressure at which the granulite could have formed can be obtained from the Black Gneiss with which it may be intercalated. Geobarometry on samples containing cordierite and garnet gives an absolute minimum value of about 4 kb. Thus the granulites of the Kumkum area were formed at a temperature of around  $770^{\circ}\text{C}$  and a pressure between 4 and 7 kb with  $P_{\text{total}} \gg P_{\text{H}_2\text{O}}$ . Jackson (1976) has shown that for granulites in the Aus area, which formed under similar metamorphic conditions,  $P_{\text{H}_2\text{O}}$  must have been near to 2 kb and tentatively proposed that the low  $P_{\text{H}_2\text{O}}$  observed might be due to the dilution of the fluid phase by the presence of significant  $\text{CO}_2$  in the country rocks. Another mechanism which has been invoked to explain the condition  $P_{\text{H}_2\text{O}} \ll P_{\text{total}}$  needed for granulite formation is dehydration of the parent rock either during a long period of high-grade metamorphism or during several metamorphic events which progressively

depleted the rocks in water. It is also possible that if the parent rock was igneous it may have been essentially anhydrous already (Winkler 1974). In the case of the Kumkum granulite any of these explanations is possible to explain why  $P_{\text{total}} \gg P_{\text{H}_2\text{O}}$ . Since the Kumkum granulite is likely to have been derived from a mafic igneous rock (see below ), the rock must have been virtually anhydrous even before metamorphism. A full analysis of Kumkum granulite sample K15 is presented in Table (44). If this is represented on a conventional ACF diagram ( Fig. 90 ) it plots in the field of basaltic and andesitic rocks. A comparison of major elements (Table 46) shows the striking similarity of the composition of sample K15 to that of the average tholeiite of Nockolds (1954). The Co, Cr and V concentrations found in sample K15 are similar to those found in basaltic rocks (Turekian and Wedepohl 1961; Taylor 1965). Bhattacharyya (1971), on the basis of 225 orthopyroxene analyses, showed that orthopyroxenes having  $(\text{MgO} + \text{FeO} + \text{Fe}_2\text{O}_3)$  higher than 44,3 percent are metamorphic. The value of ferrohypersthene in sample K58 is 47,6% which also has  $\text{Al}_2\text{O}_3 = 0,6\%$ . For garnet-free assemblages the  $\text{Al}_2\text{O}_3$  content of co-existing pyroxenes is expected to increase from low to high pressure granulite assemblages according to the experimental work of Green and Ringwood (1967b). Since the  $\text{Al}_2\text{O}_3$  content of both ortho- and clino-pyroxene in K58 is so much

lower than that generally found in garnet-free granulites and charnockites of equivalent composition to K58, it can be inferred that sample K58 formed at relatively low pressures, probably bordering on the extrapolated boundary between the low pressure granulite and intermediate pressure granulite fields of Green and Ringwood (1967b). The latter is inferred from the absence of garnets in granulites and the presence of coronas surrounding olivine in contact with plagioclase in mafic igneous rocks which cooled at the peak of metamorphism (M 1).

Both the biotite and hornblende from sample K58 (Table 43) are rich in titanium and hornblende has a high (Na + K) content which is compatible with their occurrence in a rock of the granulite facies. The biotite has the highest iron content of all the biotites analyzed from mafic and pelitic rocks in the area.

#### 7.2.6 Retrograde Metamorphism in the Ultramafic rocks of Tantalite Valley

Where the ultramafic rocks are highly sheared as is the case for the north-east plug Map (1) they may be completely metamorphosed to greenschists or greenstones composed of chlorite and calcic amphibole (mostly tremolite-actinolite). The

shear zones probably acted as channelways for introducing water into the rocks.

The olivine in the ultramafic rocks is generally altered to serpentine and the presence of this mineral implies that  $X_{\text{CO}_2}$  must have been less than 10 mole% in the fluid phase during metamorphism (Winkler 1974). Haematite and secondary magnetite were also formed by retrogressive metamorphism and their co-existence implies that  $P_{\text{O}_2}$  was high (Miyashiro 1973, p 51 - 52). Alteration of plagioclase to saussurite, cancrinite and prehnite is ubiquitous. The presence of the retrograde assemblages actinolite + hornblende + chlorite + cummingtonite and forsterite + serpentine + talc implies the transition between low and medium stage metamorphism (Winkler 1974).

#### 7.2.7 Discussion of the Status of Charnockites in the Namaqua Metamorphic Complex

The name charnockite is used for rocks falling within the "granite" field of the QAPF triangle (Tobi 1971) and is one of a series of genetically related rocks from acidic to ultrabasic in composition (eg mangerite, farsundite, enderbite etc) called the charnockite suite or charnockite series. Members of the charnockite

series are characterized by the presence of hypersthene and/or almandine, hornblende, quartz, and perthitic and/or antiperthitic feldspar. The rocks have a greasy appearance and owe their greenish-grey colour to the presence of grey-blue quartz and blue-green feldspar (Pichamuthu 1953). They are commonly associated with rocks found in the granulite facies particularly granulites, anorthosites and migmatites and are restricted to well-defined areas of Precambrian shields (Tobi 1971). Charnockites form under conditions of the granulite facies of metamorphism, ie high temperatures and  $P_{\text{total}} \gg P_{\text{H}_2\text{O}}$ .

It is apparent from a review of the literature that there are three ways in which charnokite<sup>c</sup> can be formed. As summarized by De Waard (1969) these are:

- (a) Anatexis of rocks of granitic bulk composition with crystallization of the melt under granulite facies metamorphic conditions. Von Platen (1965) proposed that there is a reaction between biotite and quartz, giving rise to a potash feldspar component + hypersthene + water, forming with quartz and plagioclase in an anatectic melt under reducing conditions.
- (b) Granulite facies metamorphism of rocks of

granitic bulk composition.

- (c) By crystallization of magmas of granitic bulk composition produced by mechanisms other than anatexis (eg differentiation) under granulite facies conditions.

It is my opinion that the term charnockite should only be used for orthopyroxene-bearing rocks of granitic bulk composition and for rocks similar in all respects to charnockites from the type-area in India.

Jackson (1976), while acknowledging Tobi's definition of a charnockite, uses the term in a general sense for all orthopyroxene-bearing plutonic rocks regardless of their age or origin so that it is hardly surprising that he attributes a widespread distribution to these rocks in the Namaqua Metamorphic Complex. In fact, the only true charnockites which occur in the entire region are found in an area between Upington and Kakamas (Von Backström 1964; Geringer 1973), and just to the north of Klein Karas (Blignault 1977). All the other examples quoted by Jackson (1976) including Tantalite Valley and Nouzees are either granulites or mafic igneous rocks which have been intruded and cooled during or after metamorphism. Work on the co-existing pyroxenes found in the fresh mafic and ultramafic

rocks of the Tantalite Valley Complex, indicate intrusion temperatures greater than  $1\ 000^{\circ}\text{C}$  which is about  $200^{\circ}\text{C}$  higher than the value given by Hlewins (1975) for various charnockites and granulites. In the Kumkum area, the Kumkum gabbro-norite has been called a granulite and a mafic charnockite by Toogood (1976) and Jackson (1976) respectively. Although co-existing pyroxenes in this rock-type give temperatures of around  $800^{\circ}\text{C}$  (Wood and Banno 1973) the evidence points to an intrusion temperature of about  $1\ 200^{\circ}\text{C}$ . Firstly this rock-type is texturally, compositionally and mineralogically almost identical to the Tantalite Valley gabbro-norite which was intruded near  $1\ 200^{\circ}\text{C}$ . Exsolution features in pyroxenes and sulphides from the Kumkum gabbro-norite, clouding in feldspars, reaction relationships between olivine and plagioclase and opaque mineral-orthopyroxene symplectites indicate cooling from magmatic temperatures. Both the Kumkum and Tantalite Valley gabbro-norites are orthocumulates which may have a well developed igneous lamination defined by plagioclase laths. The presence of sapphirine-bearing metapelites from the Kumkum area indicates that temperatures near  $1\ 000^{\circ}\text{C}$  were achieved locally, probably close to contacts with the Kumkum gabbro-norite (the only mafic intrusive in the area apart from minor plugs of Eselruh



gabbronorite and layers of amphibolite) since such high temperatures are usually not encountered in most regional metamorphic terrains (Winkler 1974). Clifford (1974) found that African members of the 'charnockite-granulite' series plot mainly in the "high alumina basalt" field in an alkali-silica diagram of Kuno (1968) and that in an AFM diagram they show a trend similar to that observed for calc-alkali and alkali igneous suites. In contrast the mafic and ultramafic rocks from the Tantalite Valley and Kumkum areas plot in the 'tholeiite series' field in an alkali-silica diagram (Figs 69 and 72) and show a tholeiitic or iron-enrichment trend on an AFM diagram (Figs 67 and 71). Reid and Moore (pers comm) have investigated the Nouzees mafic complex near Pofadder and describe rocks which bear a striking resemblance to the metagabbro and gabbronorite of Tantalite Valley. Further the mafic rocks from Nouzees all plot in the 'tholeiite series' field in an alkali-silica diagram of Kuno (1968) and define an iron-enrichment trend on an AFM diagram. The Stolzenfels norite or enderbite formation of Toogood (1976) is seen in the Kumkum area to be a hybrid rock occurring as it does along the contacts of the Kumkum gabbronorite with the Inequigranular granite gneiss and grading by degrees into unaltered gabbronorite away from

the contact. It is postulated here that the intrusion of the Kumkum gabbro-norite into the granite gneiss caused partial melting in the latter and led to the formation of a series of contaminated, recrystallized mafic rocks, termed here the Contaminated gabbro-norite.

In general, though mafic intrusives commonly occur associated with rocks of high-stage metamorphism in the Namaqualand Metamorphic Complex, (Jackson 1976) these should not be referred to as charnockites but should be dealt with as igneous rocks of continental origin. The intrusion of the mafic rocks could however, have contributed to high-grade metamorphism in the region mainly as a result of increasing the prevailing geothermal gradient but also by dehydrating the country rocks.

#### 7.2.8 Summary of Metamorphism in the Tantalite Valley and Kumkum Areas

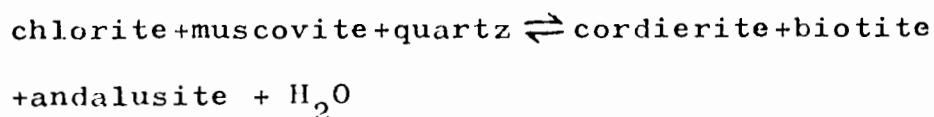
Along and north of the axis of the Tantalite Valley Lineament the rocks have been affected by two main regional metamorphic events. The first of these (M 1) was characterized by pressures between 5 and 6 kb with  $P_{\text{total}} > P_{\text{H}_2\text{O}}$  and temperatures between 750 and 860°C. Locally temperatures may have reached as high as 1 000°C near mafic intrusives.

During this event garnet-and cordierite-bearing gneisses were formed as restites after the removal of an anatectic melt. Muscovite and quartz became unstable and were replaced by sillimanite and melt. These melt-producing reactions led to the formation of migmatites. During M 1 the Kumkum gabbro-norite and Eselruh gabbro-norite were emplaced and cooled.

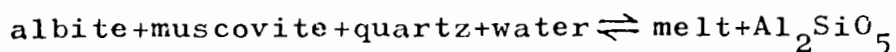
M 2 was a retrograde event which occurred at a pressure from 4 to 5,5 kb,  $P_{\text{total}} > P_{\text{H}_2\text{O}}$  and temperatures increased northwards from the axis of the Tantalite Valley shear zone into the Kumkum area, ie from 580 to 760°C. Cordierite formed from garnet during M 2 and a second generation of migmatites was produced probably due to the local entry of water into the rocks. Amphibolites were also probably formed as a result of this introduction of water. During or slightly after M 2 the Tantalite Valley metagabbro was emplaced into the core of the Tantalite Valley Lineament and produced aureole temperatures of about 575 to 820°C in the country rocks, at a pressure of about 5,5 kb. Small degrees of partial melting took place.

The region occupied by rocks of the Namaqua Metamorphic Complex to the south of the axis of the Tantalite Valley Lineament has not been studied in any detail in this work. This domain is characterized

by the minerals cordierite, andalusite, sillimanite, garnet (not co-existing with cordierite), diopside and muscovite. Beukes (1973) has suggested that the reaction:



marks the lower temperature limit for metamorphism in the area. Hirschberg and Winkler (1968) found that this reaction occurs at 500 to 550°C over a pressure range of 0,5 to 4 kb. The reaction:



proposed by Beukes (1973) as an upper temperature limit occurs at a temperature of 650°C over a wide range of pressures above 3 kb according to the experimental work of Storre and Karotke (1971). From the widespread distribution of andalusite in this area pressures could not have been much greater than about 4 kb using the  $\text{Al}_2\text{SiO}_5$  triple point of Holdaway (1971). Thus metamorphism in the area south of the Tantalite Valley shear zone axis occurred at temperatures between about 550 and 650°C and pressures of between 3 and 4 kb. In this region mafic rocks are rare and granitic rocks are absent. Migmatites only make their appearance in rocks in which sillimanite is found near to the axis of the

### Tantalite Valley Lineament.

Both Moore (1975) and Toogood (1976) noted the abrupt northward change from low-medium to high-stage metamorphism (Winkler 1974) on crossing the axis of the Tantalite Valley Lineament. Toogood (1976) attributed this phenomenon to the northern block having been more deeply eroded than the southern block. Beukes (1973) defined the "cordierite-in" and "sillimanite-in" isogrades in the southern block, both of which are approximately parallel to the trend of the Tantalite Valley Lineament. The grade of metamorphism decreases towards the contact of the Namaqua and Richtersveld Tectonic Provinces, which lie far to the south-west of the Tantalite Valley Complex. Beukes (1973) also showed that the metamorphism in the area is characteristic of the Abukuma low P/high T facies series which, according to Miyashiro (1973), means that the geothermal gradient was high: between 25 and 50°C/km.

Low-grade retrogressive metamorphism is only significantly developed along zones of shearing, particularly the Tantalite Valley shear zone, where some local partial melting occurred probably as a result of the introduction of water. The overall poor development of retrogressive effects can

probably be ascribed to the already dehydrated nature of the rocks and to a sharp decrease in heat flow in the area, probably as a result of rapid uplift.

#### 7.2.9 Regional Variations in High-Stage P T Conditions in the Namaqua Metamorphic Complex in South West Africa

An examination of mineral assemblages reported by various workers from metamorphic rocks from the Namaqua Metamorphic Complex in southern South West Africa indicates that at the highest grades of metamorphism realized, P and T conditions were almost constant throughout the entire region.

The estimates of pressure (8 kb) reported by Toogood (1976) from the Onseepkans area are almost certainly too high, because he based his conclusion on the erroneous assumption that garnet, and not garnet plus cordierite, was stable during the highest grade of metamorphism in the area. Results from the present study area, which overlaps that of Toogood (1976), indicates that P was between 5 and 6 kb and T between 750 and 860°C.

In the area between Ai-ais and Klein Karas, Blignault (1977) has estimated that the peak of

metamorphism occurred at 6 kb and  $740^{\circ}\text{C}$ . The temperature is probably too low, because he does not take into account the presence of charnockite sensu stricto in the zone of highest-grade metamorphism. From Hewins (1975), it is likely that temperatures required for the formation of these charnockites must have been close to  $800^{\circ}\text{C}$ .

In the Aus area, Jackson (1976) has demonstrated that granulite facies metamorphism occurred at  $P_t = 5 - 7 \text{ kb}$  and  $780 \text{ to } 860^{\circ}\text{C}$ .

Beukes (1973), Jackson (1976) and Blignault (1977) have shown that metamorphism of the gneisses of the Namaqua Metamorphic Complex in southern South West Africa is typical of the low P/high T (Abukuma) facies series, with the corollary that geothermal gradients were high. Most workers also agree that the peak of metamorphism in the region was achieved early on and that this was followed by retrograde metamorphism, (Jackson 1976; Toogood 1976; Blignault 1977). Metamorphic isograds are either north-west trending, as in the Warmbad and Ai-ais area, (Beukes 1973; Blignault 1977), or strike approximately east-west, as in the Aus area (Jackson 1976), with the grade of metamorphism increasing in a north-easterly or northerly direction towards a high-grade core zone after which it

decreases again (Blignault 1977). Although a decrease in grade has not as yet been shown for rocks lying north of the Kumkum area, this may be found to be the case with further work. In the Warmbad and Ai-ais areas the isograds are parallel to the trend of the Tantalite Valley Lineament and its north-westerly extension, the Kanabeam Lineament (Beukes 1973; Blignault 1977). In the Warmbad area the Tantalite Valley Lineament forms the boundary between rocks containing K-feldspar and sillimanite north of the lineament and rocks containing muscovite and quartz to the south of the lineament (Moore 1975; Toogood 1976) whereas in the Ai-ais area the K-feldspar + sillimanite + isograd is only reached some 20 km north-east of the Kanabeam Lineament which does not define any isograd in itself (Blignault 1977). In the Aus area, the Kuckaus Lineament, the probable north-westerly extension of the Kanabeam lineament, cuts completely across the metamorphic isograds without displacing them. (Jackson 1976). It appears that both horizontal and vertical displacement were negligible here, whereas in the Onseepkans area, Toogood (1976) has estimated a minimum horizontal displacement of 85 km and attributes differences in grade across the shear zone to significant vertical movements having taken place along the latter with erosion



of the uplifted block exposing deeper levels of metamorphism.

## CHAPTER VIII

## GEOCHEMISTRY

"It is a capital error to theorize before one has data".

Scandal in Bohemia - Sir Arthur Conan-Doyle.

### 8.1 Chemical Alteration

#### 8.1.1 General Statement

By chemical alteration is meant any process such as weathering, metasomatism, hydrothermal alteration, which changes the chemical composition of the rock after its formation (Reid 1977).

Chemical alteration is certain to have taken place in those mafic intrusive rocks which have been uralitized, retrogressively metamorphosed or show other evidence that metasomatism has occurred. The effects of sub-aerial weathering in the form of desert varnish generally can be neglected if the weathered surface is removed prior to crushing of the rock.

The author did not collect all the rocks for which analyses are presented in this report (see Appendix 5) so that the normal sample selection criteria such as freshness of samples, absence of contamina-

tion and shearing, representability of samples, cannot be said to have been always applied.

In order to assess the degree of alteration of the sample analysed it is useful to consider such variables as  $H_2O^+$  (water lost at  $1\ 000^\circ C$ ) and  $CO_2$ , or a combination of these in the form of loss on ignition (ie L.O.I. = total volatile loss at  $1\ 000^\circ C$ ), the  $Fe_2O_3/FeO$  ratio and the presence or absence of corundum in the C.I.P.W. norm.

In the following sections major oxide concentrations are given in weight percentages, referred to simply as "%".

#### 8.1.2 Volatile Content

Fresh gabbroic rocks of Karroo age from Rhodesia were found by Hossain (1977) to have  $H_2O^+$  values ranging from 0,29 to 0,81% ( $n = 8$ ). This can be taken to represent the primary water content of gabbroic rocks of tholeiitic composition which have been unaffected by metamorphism. Any gabbroic rock containing  $H_2O^+$  greatly in excess of about 0,80% can then be said to be altered. Applying this criterion to the mafic and ultramafic rocks of the Kumkum-Tantalite Valley areas it can be seen that the Tantalite Valley gabbro-norite (av.  $H_2O^+$  content =  $0,75 \pm 0,06\%$ ), the Kumkum gabbro-norite

(av.  $H_2O^+$  content =  $0,62 \pm 0,16\%$ ) and the Contaminated gabbronorite (av.  $H_2O^+$  content =  $0,52 \pm 0,22\%$ ) can be regarded as virtually unaltered. The mottled metagabbro and metagabbro which are essentially hornblende + plagioclase rocks contain on average almost the same amounts of  $H_2O^+$  ie  $1,82 \pm 0,06\%$  ( $n = 4$ ) and  $1,87 \pm 0,92\%$  ( $n = 8$ ) respectively. Samples of the mafic-ultramafic group from Tantalite Valley have a wide range of  $H_2O^+$  values from  $1,93 - 9,07\%$  (av =  $6,11 \pm 2,96\%$ ,  $n = 5$ ) and with respect to  $H_2O^+$  content can be said to be the most altered of all the mafic rock-types. This is to be expected in rocks which have had water introduced into them along shear zones, resulting in serpentinitization, saussuritization, chloritization and uralitization. The Eselruh gabbronorites which show many of these alteration features in thin section have the highest  $H_2O^+$  contents of all the mafic rocks in the Kumkum area excluding amphibolites.

$H_2O^+$  is plotted against D.I. (the differentiation index of Thornton and Tuttle (1960) (see Appendix 4) for the mafic and ultramafic rocks of Tantalite Valley and Kumkum (Fig 66 ). Generally there is a tendency for the rocks with lower D.I. values to have higher  $H_2O^+$  contents. This is because rocks which contain large amounts of olivine and/or

pyroxene are highly susceptible to alteration since these minerals are unstable under high  $P_{H_2O}$  and low temperature conditions. Further, the more fractured nature of the ultramafic rocks compared with the gabbroic rocks probably allowed freer access of water. Tantalite Valley metagabbro sample TV277 contains a relatively large amount of chlorite which explains why this sample has an anomalously high water content (Fig 66 ). The Kumkum gabbro norite contains more magmatic hornblende and biotite than Tantalite Valley gabbro norite which is reflected in the higher  $H_2O^+$  values of Kumkum gabbro norite samples compared with those of Tantalite Valley gabbro norite specimens. This implies that the Kumkum gabbro norite magma was the more hydrous of the two.

The mafic and ultramafic rocks from Kumkum and Tantalite Valley have  $CO_2$  contents just below 0,2% on average. This is higher than the average  $CO_2$  content of 0,1% obtained by Hossain (1977) from fresh Karroo gabbros from Rhodesia. This implies that either the mafic and ultramafic rocks of the Kumkum - Tantalite Valley area were derived from more  $CO_2$ -rich magmas or they were altered. The Tantalite Valley gabbro norite which has the water content of a fresh rock has surprisingly high  $CO_2$

values (av 0,33  $\pm$  0,12%, n = 5). This is ascribed to calcite veining which is commonly observed on the surface of the rock in the field. Even when no veining is present the CO<sub>2</sub> content is still high (ie 0,18%) so that it is likely that the CO<sub>2</sub> content of the magma was relatively high. Plots of CO<sub>2</sub> versus D.I. for the Tantalite Valley and Kumkum mafic rocks are given in Fig ( 66 ). The Tantalite Valley samples show a large spread in values with the average centred about 0,18% CO<sub>2</sub>. The samples with values higher than this have calcite in their modes or show calcite veining. The Kumkum rocks show a general increase of CO<sub>2</sub> with decreasing D.I., a similar trend to that noted for H<sub>2</sub>O<sup>+</sup>. Thus CO<sub>2</sub> content of the rock increases with increasing alteration.

### 8.1.3 Bulk Fe<sub>2</sub>O<sub>3</sub> / FeO ratios

It is important to have an approximate value for the initial Fe<sub>2</sub>O<sub>3</sub> / FeO ratio in a sample so that the degree of oxidation can be gauged compared with similar rocks and because the magnitude of this ratio affects the C.I.P.W. norm (Le Maitre 1976). Although it is not possible to exactly determine the original Fe<sub>2</sub>O<sub>3</sub> / FeO ratio of a rock due to analytical difficulties eg partial oxidation of FeO to Fe<sub>2</sub>O<sub>3</sub> and post-consolidation oxidation

processes, it can be constrained by determining the FeO content of a sample using standard wet chemical methods (eg Shapiro and Brannock 1956).

FeO and  $\text{Fe}_2\text{O}_3$  values are available for all of the rocks analysed from the present study area so that it is possible to plot  $\text{Fe}_2\text{O}_3 / \text{FeO}$  versus D.I. for the mafic intrusives of the Tantalite Valley and Kumkum areas (Fig 66). It can be seen that most of the rocks have an  $\text{Fe}_2\text{O}_3 / \text{FeO}$  ratio of about 0,2. Further, the  $\text{Fe}_2\text{O}_3 / \text{FeO}$  ratio of basaltic rocks is commonly adjusted to 0,2 before calculating the C.I.P.W. norm, following the approach adopted by Melson and Thompson (1973). On the strength of these two facts, C.I.P.W. norms for the mafic and ultramafic rocks of the Kumkum - Tantalite Valley area were calculated assuming  $\text{Fe}_2\text{O}_3 / \text{FeO} = 0,2$ . Fig ( 66 ) shows that some of the Tantalite Valley ultramafic rocks have unusually high  $\text{Fe}_2\text{O}_3 / \text{FeO}$  ratios. The same samples also have high  $\text{H}_2\text{O}^+$  values. The introduction of large quantities of water into these rocks resulted in advanced serpentinization of olivine with the production of abundant secondary magnetite. From Fig ( 66 ) it can also be seen that Kumkum gabbro-norite sample K65 has a higher than normal  $\text{Fe}_2\text{O}_3 / \text{FeO}$  ratio. The reason for this is not clear since the rock has  $< 0.5\%$  modal ilmenite and is fresh except for some

alteration of pyroxene along cracks to hornblende.

#### 8.1.4 Normative Corundum

Normative corundum is calculated in the C.I.P.W. norm when the  $\text{Al}_2\text{O}_3$  content of a rock exceeds that of the sum of  $\text{Na}_2\text{O}$ ,  $\text{K}_2\text{O}$  and  $\text{CaO}$  (Shand 1952, p 40). Removal of alkalis and/or calcium from a rock during alteration may be indicated by normative corundum.

Normative corundum is only found in the C.I.P.W. norms of four of the many samples of mafic and ultramafic rocks examined from the Kumkum and Tantalite Valley areas. Significantly it is restricted to samples of ultramafic rocks of Tantalite Valley (ie T7, TV45, TV117 and GBW376 given in Table (44) which have anomalously high loss on ignition or  $\text{H}_2\text{O}^+$  and  $\text{Fe}_2\text{O}_3 / \text{FeO}$  ratios and have the lowest concentrations of alkalis and calcium of all the mafic rocks. Further, since  $\text{Na}_2\text{O}$ ,  $\text{K}_2\text{O}$  and  $\text{CaO}$  are generally very mobile during alteration (Pearce 1976) it is likely that the removal of these elements by migrating fluids could have given rise to modified bulk compositions for which normative corundum would be calculated in the C.I.P.W. norm.



#### 8.1.5 Discussion

It has been postulated in section ( 4.7 ) that Si, P, K and Na have been introduced metasomatically from granitic rocks into the mottled metagabbro. Except for this and the loss of alkalis and calcium noted for four ultramafic rock samples it is clear that the main alteration which the rest of the samples have undergone involves addition of water and  $\text{CO}_2$  and the oxidation of iron, processes which do not change the relative amounts of the major elements in the rock. Studies of uralitization of gabbroic rocks have been discussed in section (7.2.1). It was concluded that this process is essentially isochemical and only involves introduction of water into the rock. Smith (1968, 1969) showed that the bulk composition of volcanics containing up to 10 - 12%  $\text{H}_2\text{O}^+$ , when calculated volatile free is comparable to that of fresh, unaltered rocks from the same unit. On the strength of this all major element analyses of rocks from the Kumkum - Tantalite Valley area have been recalculated to 100% volatile free for representation on variation diagrams assuming as a first approximation that only water has been added to the rock during alteration.

## 8.2 Major Element Composition of the mafic and ultramafic rocks of the Tantalite Valley Complex

Discussion is based on 43 bulk analyses of mafic rocks from the Tantalite Valley Complex and immediate vicinity. These include 5 of mottled metagabbro, 12 of metagabbro, 9 of gabbro-norite, 2 of the gabbro-norite satellite body and 15 of the ultramafic rocks. These are probably representative of the range of compositions found in the Complex as a whole. When major element variations are discussed the values used are those obtained by recalculating the analysis to 100%, volatile free. C.I.P.W. norms were calculated from the volatile free analyses assuming  $\text{Fe}_2\text{O}_3 / \text{FeO} = 0,2$  (see section 8.1.3).

### 8.2.1 C.I.P.W. norm

The metagabbro and mottled metagabbro are characteristically oversaturated rocks as is reflected by quartz in their C.I.P.W. norms. Textural evidence such as the presence of ocelli, points to the fact that some  $\text{SiO}_2$  was metasomatically introduced into the mottled metagabbro from neighbouring granitic rocks. In gabbro-norite an equal number of samples contain normative quartz as do those which contain normative olivine. In ultramafic rocks only sample TV30 is quartz norma-

tive; all the rest are olivine normative. Sample TV30 is essentially a tremolite-actinolite-cummingtonite rock. It is likely that uralitization of the original pyroxenes resulted in the formation of quartz which then concentrated in parts of the rock.

All of the samples of mafic and ultramafic rocks of Tantalite Valley contain normative hypersthene. In addition most of the fresh rocks have hypersthene in their modes. From the absence of rocks of intermediate composition with corundum in their norms (Cawthorn et al 1976) and the ubiquitous presence of hypersthene normative rocks it is concluded that the mafic and ultramafic rocks of Tantalite Valley are tholeiitic rather than calc-alkaline.

#### 8.2.2 Iron-enrichment

The degree of iron-enrichment of igneous rocks is generally shown by plotting liquid compositions on an AFM diagram. For rocks from the Tantalite Valley Complex such a diagram is of limited value because of the presence of cumulate rocks such as the gabbro-norite, the satellite body and ultramafic rocks. However Clifford (1974) used such a diagram to show that African charnockites

and granulites define a so-called "calc-alkaline trend". When the mafic rocks of Tantalite Valley are plotted on an AFM diagram they define an iron-enrichment or tholeiitic trend (Fig 67 ). The ultramafic rocks have predominantly cumulus olivine and orthopyroxene present and tend to plot towards the Mg-corner of Fig ( 67 ) whereas the gabbronorites and some of the metagabbros are displaced towards the  $(\text{Na}_2\text{O} + \text{K}_2\text{O})$  ie ALK-corner of this diagram because plagioclase is a major cumulus phase in these rocks. The mottled metagabbro samples are relatively more Fe-rich than the rest of the mafic rocks and have a relatively high alkali content. The iron-enrichment tendency of a number of different magma series is shown in Fig ( 68 ) for comparison.

### 8.2.3 Total alkalis-silica variation

In a  $(\text{Na}_2\text{O} + \text{K}_2\text{O})$  versus  $\text{SiO}_2$  diagram (Fig 69 ) after Kuno (1968) the mafic and ultramafic rocks of the Tantalite Valley Complex all plot well within the "tholeiitic series" field. Rocks which are enriched in cumulus plagioclase such as the gabbronorite and rocks which have had alkalis added to them during metasomatism such as the mottled metagabbro are displaced towards the

"high-alumina basalt" field.

#### 8.2.4 Major Element Abundances

Barnes (1978) has devised a discriminant analysis diagram using the canonical variables  $\text{SiO}_2$ , FeO total, MgO and NiO for distinguishing between the four main types of serpentinites found in nature based on several hundred bulk analyses from the literature. She called the four groups (a) layered complexes (b) Alpine/mantle fertile (c) Alpine/mantle differentiated and (d) Alpine/mantle depleted. Although her diagram shows a large area of overlap for the four classes, Tantalite Valley ultramafic rock samples AM88A, AM89B, T7 and T10, which satisfy the requirements for representation on this diagram (ie they approximate serpentinite composition), all plot clearly and unambiguously in the field of serpentinites from layered complexes which indicates that they probably have a tholeiitic affinity.

The metagabbro and gabbro-norite have average major element compositions similar to those given for high- $\text{Al}_2\text{O}_3$  olivine tholeiites (abyssal tholeiites) dredged from along mid-ocean ridges (Engel et al. 1965; Green and Ringwood 1967a). Compare values in Tables ( 45 ) and ( 46 ). Compared with the

average olivine gabbro and gabbro of Nockolds (1954, p 1 020) the average Tantalite Valley metagabbro and gabbro norite are lower in  $\text{TiO}_2$ , FeO total, MnO, MgO,  $\text{Na}_2\text{O}$  and  $\text{K}_2\text{O}$  and higher in  $\text{SiO}_2$ ,  $\text{Al}_2\text{O}_3$  and CaO (see Tables 45 and 46 ). The low  $\text{K}_2\text{O}$  content (ie  $< 0,5\%$ ) and  $\text{TiO}_2$  content (av.  $\sim 0,25\%$ ) of the unaltered gabbroic and ultramafic rocks is also a feature of similar rocks from the Bushveld Igneous Complex, and the Skaergaard and Rhum intrusions (Wager and Brown 1968). There is also a remarkable similarity in compositions of the Tantalite Valley gabbro norite and metagabbro as can be seen by comparing the average compositions of these rocks (see Table 45 ). This indicates that the metagabbro has formed by iso-chemical metamorphism of a rock similar in composition to the gabbro norite.

#### 8.2.5 Major oxide - D.I. Variation

If the major oxides of the mafic rocks of the Tantalite Valley Complex are plotted against D.I. as in Fig ( 70 ) a number of features become apparent:

1. Although the ultramafic rocks and the gabbroic rocks (metagabbro and gabbro norite) often define an approximate trend there is normally a hiatus along this trend which separates the

fields of ultramafic and gabbroic rocks.

2. There is generally a large degree of scatter for points along the ultramafic-gabbroic rock trend.
3. The mottled metagabbro and the satellite body rocks plot in their own distinct fields.

For  $\text{SiO}_2$  versus D.I. (Fig 70 ) the ultramafic rocks define two trends. This is a reflection of the two main ultramafic rock types found in the Complex ie troctolites and pyroxenitic rocks. Varying the amount of olivine in troctolite and the amount of orthopyroxene in norite will give rise to these trends. This conclusion is based on the control line which can be established between olivine and orthopyroxene for ultramafic rocks in Fig ( 79 ).

The gabbro-norite and metagabbro form a distinct cluster due to their similar  $\text{SiO}_2$  contents and range in D.I.

For MgO versus D.I. (Fig 70 ) the ultramafic rocks define a single trend controlled by the amount of olivine and orthopyroxene in the rock. (See Fig 79 ). Sample T10 is high in MgO because it is enriched in cumulus olivine, having the highest olivine content (  $\sim 83\%$  ) of all the ultramafic

rock samples.

The trend for metagabbros and gabbro-norites in the MgO versus D.I. plot (Fig 70) is controlled by the amount of orthopyroxene or its metamorphosed equivalent present (see Fig 79).

The mottled metagabbro is characterized by constant MgO values and in this respect is distinct from the other gabbroic rocks.

In  $\text{TiO}_2$  / D.I. space (Fig 70) the ultramafic and gabbroic rocks define an overall shallow positive trend. In the ultramafic rocks  $\text{TiO}_2$  is found as a minor component in pyroxenes and phlogopite whereas in the gabbros Ti-rich minerals such as ilmenite and sphene are common. The mottled metagabbro has the highest  $\text{TiO}_2$  content of all the mafic rocks. Since Ti is immobile, this is clearly a primary feature of the rock.

The  $\text{Al}_2\text{O}_3$ , CaO and  $\text{Na}_2\text{O}$  versus D.I. trends (Fig 70) for the ultramafic and gabbroic rocks are essentially similar and represent progressive enrichment in plagioclase as compositions become less mafic. The satellite body gabbroids are depleted in plagioclase compared to metagabbros and gabbro-norites, a feature which is clearly seen in Fig ( 70 ).

In the  $\text{K}_2\text{O}$  versus D.I. plot in Fig ( 70 ) most of



the ultramafic and gabbroic rocks plot along a shallow positive trend, which is controlled mainly by the amount of plagioclase and hornblende in the rock. Samples which are enriched in  $K_2O$  define a similar trend but with a steeper gradient. The presence of relatively large amounts of primary phlogopite in ultramafic rock samples T10 and T11, metasomatism of ultramafic rock sample T6 (see section 8.4.2.2) or possibly contamination of metagabbro samples TV265 and TV277 and satellite body sample AM123A are responsible for enriching samples in  $K_2O$ . The mottled metagabbro shows a wide scatter in alkali content which would be expected if Na and K were mobile during metasomatism (see section 4.7).

The ultramafic-gabbroic rock trends on the  $FeO^*$  and  $MnO$  versus D.I. diagrams (Fig 70 ) are essentially the same and suggest a control by mafic minerals. Some samples (eg T36, TV277, TV205A, JWVB25) are enriched in both  $FeO^*$  and  $MnO$  with respect to the average trends but it is not clear which mafic mineral or minerals <sup>or are</sup> is responsible for this since most of the samples named have high plagioclase contents and have relatively low contents of mafic minerals. The enrichment of  $FeO^*$  in addition to  $MgO$  in ultramafic rock sample T10

must be due to olivine accumulation in the absence of phases like pyroxene.

### 8.3 Major Element Composition of the mafic rocks of the Kumkum area and their relationship to the mafic-ultramafic rocks of Tantalite Valley

The following discussion is based on 17 analyses of gabbroic rocks from the Kumkum area made up of 7 samples of Kumkum gabbro, a sample of the Einsiedler gabbro (Beukes 1973), 5 samples of Contaminated gabbro, 3 samples of Eselruh gabbro and a sample of gabbro from Verloorkoppe (Beukes 1973). These rocks have been plotted on the same diagrams as were used in discussing the Tantalite Valley rocks and identical conventions were followed. One granulite sample is included in the plots for purposes of comparison.

#### 8.3.1 C.I.P.W. norm

All of the samples are hypersthene normative and are hence of tholeiitic affinity. Only the Contaminated gabbro has normative quartz and this can be used to distinguish it from the other mafic rock types which are olivine normative. None contain normative corundum.

### 8.3.2 Iron-enrichment

On an AFM diagram (Fig 71 ) the gabbros of the Kumkum area show an iron-enrichment trend with the olivine-pyroxene rich rocks, the Eselruh gabbronorite in particular, plotting nearer the Mg-corner than more plagioclase-rich rocks. There is a clear separation of the recrystallized Contaminated gabbronorite from the unreconstituted Kumkum and Eselruh gabbronorites. The Verloorkoppe gabbronorite and Kumkum granulite both plot together with the Contaminated gabbronorite while the Einsiedler gabbronorite plots together with the Kumkum gabbronorite.

### 8.3.3 Total alkalis-silica variation

All of the mafic rocks from the Kumkum area plot within the "tholeiite series" field in a total alkali-silica variation diagram (Fig 72 ) and in this respect resemble the Tantalite Valley mafic rocks.

### 8.3.4 Major element abundances

The average Kumkum gabbronorite composition closely resembles that of high- $\text{Al}_2\text{O}_3$  olivine tholeiites (Engel et al 1965, Green and Ringwood

1967a). (Compare the analyses in Table 45 and 46 ).

#### 8.3.5 Major oxide-D.I. Variation

In order to compare the mafic rocks from the Kumkum and Tantalite Valley areas, separate Major oxide/D.I. plots were done for the Kumkum samples and the fields in which the Tantalite Valley mafic rocks occur were superimposed. From the plots it is clear that not enough samples were analysed because it cannot be shown whether regular trends exist or not. From Fig ( 73 ) it can be seen that:

1. There is a wide scatter of points.
2. Contaminated gabbonorite samples fall in or near the field defined by mottled metagabbro samples from Tantalite Valley in all of the plots, which provides support for the hybrid origin hypothesis proposed for the Contaminated gabbonorite in section ( 3.3.5.2). Both rock-types are characterized by higher  $\text{TiO}_2$  contents than the rest of the mafic rocks.
3. The Verloorkoppe gabbonorite and the Einsiedler gabbonorite samples plot in a similar way to Contaminated gabbonorite and Kumkum gabbonorite samples respectively.
4. The Kumkum gabbonorite has higher  $\text{FeO}^*$ ,  $\text{MnO}$ ,

$\text{TiO}_2$ ,  $\text{MgO}$  and about the same amounts of  $\text{SiO}_2$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{CaO}$ ,  $\text{Na}_2\text{O}$ ,  $\text{K}_2\text{O}$  and  $\text{P}_2\text{O}_5$  compared with Tantalite Valley gabbronorites and metagabbros.

5. Compared with the Kumkum gabbronorite the Eselruh gabbronorite has higher  $\text{MgO}$ ,  $\text{K}_2\text{O}$  and  $\text{MnO}$ , about the same  $\text{SiO}_2$  and  $\text{FeO}^*$  and lower  $\text{TiO}_2$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{Na}_2\text{O}$ ,  $\text{P}_2\text{O}_5$  and possibly  $\text{CaO}$ . Eselruh gabbronorite sample K59A has about 70% modal clinopyroxene and secondary hornblende, and 2% modal secondary calcite which are responsible for the high  $\text{CaO}$  content of this sample. Large amounts of phlogopite (10% in K60) and secondary hornblende (K59A, K7) are responsible for the Eselruh gabbronorite having higher  $\text{K}_2\text{O}$  content than the Kumkum gabbronorite. It is shown in section (8.5.4) that the compositions of the Kumkum gabbronorite and Eselruh gabbro-norite are controlled mainly by the minerals plagioclase, olivine and orthopyroxene.

#### 8.4 Comparative Trace Element Geochemistry

In this section a comparison is made between the concentrations of trace elements in Kumkum and Tantalite Valley mafic rocks and with other rocks of similar composition in general. All of the trace elements are plotted against D.I. with either arithmetic scales or, where the range in concentra-

tions is great, with log-normal scales. In the diagrams for the Kumkum area, the fields occupied by Tantalite Valley samples have been superimposed for comparison. Average trace element compositions for each rock-type are given together with standard deviations (Table 45 ) to express the uncertainties of these values.

For the mineralogy observed in the mafic rocks from the Kumkum - Tantalite Valley area, the trace elements V, Cr, Ni, Co, Sr, Sc, Zn and Cu are "compatible elements" ie they are elements which tend to concentrate in the phases crystallizing from the magma, whereas the trace elements Rb, Ba, Zr, Nb and Y are "incompatible elements" ie they are elements which are strongly partitioned into the liquid fraction of the basic magma. In the following section trace elements are discussed under the headings compatible and incompatible elements with the elements showing similar behaviour eg Zr - Nb-Y grouped together under these headings.

#### 8.4.1 Compatible Elements

##### 8.4.1.1 Ni - Co

Nickel characteristically shows a strong depletion with fractionation because it is strongly

partitioned into olivine ( $K_D = 12$ , Dale and Henderson 1972) and is partitioned into orthopyroxene ( $K_D = 5$ , Le Roex and Reid 1978) and clinopyroxene ( $K_D = 2$ , Häkli and Wright 1967). It also shows an affinity for sulphur forming discrete crystals of pentlandite (Taylor 1965) or as is found in the Kumkum - Tantalite Valley area, it may go into pyrrhotite, later to be exsolved as pentlandite.

A trend for Ni versus D.I. is shown in Fig ( 75 ) for mafic and ultramafic rocks from Tantalite Valley. There is a clear separation between ultramafic rocks with 700 to 2 000 ppm Ni and metagabbro and gabbro-norite with 400 to 60 ppm Ni. This is a reflection of the greater abundance of olivine, pyroxenes and sulphide minerals in the ultramafic rocks compared to the metagabbro and gabbro-norite. Ultramafic rock sample T10 has 83% olivine and 2.5% pentlandite and ultramafic rock sample T11 has 0,5% pentlandite and 75% orthopyroxene, which explains why these samples plot off the average trend. Troctolite sample TV157A is apparently depleted in nickel. The only phase in this rock in which nickel could be significant is olivine, so that it is likely that serpentinization of this mineral might be responsible for low Ni. Fluctuations about the Ni-trend line

for the metagabbros and gabbro-norites are probably caused by variations in the modal percentage of such minerals as olivine, sulphides and pyroxenes. Metagabbro sample TV126 has 88% modal plagioclase which accounts for the low Ni-content of this sample.

For a similar range in D.I. values the Kumkum mafic rocks have higher Ni values than Tantalite Valley mafic rocks.

Cobalt is partitioned into olivine ( $K_D = 4.4$ , Dale and Henderson 1972) and into orthopyroxene ( $K_D = 2.1$ , Onuma et al 1968) and sulphide minerals. Co is not accepted to the same extent as Ni in these minerals because of the larger size of the  $\text{Co}^{2+}$  ion. Co, therefore, like Ni, decreases with fractionation but to a lesser extent than the latter (Taylor 1965). The Co versus D.I. trend is similar to the Ni versus D.I. trend (Fig 75 ) for the ultramafic and mafic rocks of Tantalite Valley. Once more there is a separation between the ultramafic rocks with 90 - 170 ppm Co and gabbro-norites and metagabbros with 20 - 80 ppm Co. The presence of 83% olivine and 5% sulphide minerals in ultramafic rock sample T10 is responsible for the high Co concentration in this sample. The presence of 88% modal plagioclase in metagabbro sample TV126 ex-



plains the low Co concentration in this sample.

Co like Ni appears to be enriched in the mafic rocks of Kumkum compared to those of Tantalite Valley for samples having a similar range of D.I. values.

#### 8.4.1.2 Cr - V

Cr and V concentrate in minerals which readily accommodate trivalent cations such as pyroxenes and some opaque minerals (eg magnetite and chromite) and are excluded from those that cannot, such as olivine and ilmenite (Taylor 1965). In a Cr versus D.I. plot for Tantalite Valley samples (Fig 75), it can be seen that there is a distinct separation of ultramafic rocks, with about 2 000 to 5 000 ppm Cr, from gabbro-norites and metagabbros which have about 800 to 100 ppm Cr. This is because chromite is a common cumulus phase in the ultramafic rocks whereas in the gabbros Cr is mostly found as a minor element in silicates. Certain samples are exceptional such as ultramafic specimen T10, the low Cr content of which can be ascribed to the large amount of olivine present. Ultramafic sample TV117 is highly altered. The Cr deficiency in this rock was caused by loss of  $\text{Cr}_2\text{O}_3$  from chromite during the alteration of the latter to

ferritchromite. The low Cr concentration of metagabbro sample TV265 is puzzling. This rock has < 0.5% opaque minerals and consists essentially of plagioclase and calcic amphibole in equal proportions. It is possible that uralitization of the pyroxenes liberated Cr but this seems unlikely since the other metagabbro samples have Cr values lying along the average trend in Fig ( 75 ). The fluctuations along the visually estimated trend line for Cr concentrations is due to variations in the amount of Cr-bearing cumulus phases such as spinels and pyroxenes which are present. The range of Cr-values is essentially the same for Kumkum and Tantalite Valley rocks with similar D.I. values.

Fig ( 74 ) shows V plotted against D.I. for Tantalite Valley samples . Two trends are produced for the ultramafic rocks; one with a negative slope, the other with a positive slope. The latter is the result of V increasing with increasing pyroxene content in norites containing little or no olivine while the former is produced by the amount of olivine decreasing relative to plagioclase in peridotites and troctolites.

The metagabbro and gabbro-norite samples on the whole have a narrow range of V concentrations from 75 to 100 ppm and higher values than this are due

to the rock being enriched in pyroxenes (or their altered equivalents) and/or opaque minerals relative to olivine and/or plagioclase while lower values are a reflection of the presence of large amounts of plagioclase (eg TV126). As can be seen in the V versus D.I. plot (Fig 74 ) the primary V content of the mottled metagabbro was higher than that found in the gabbro norite and metagabbro.

The gabbro norites from the Kumkum area have higher concentrations of V than gabbro norites with about the same D.I. range from Tantalite Valley (Fig 74 ).

#### 8.4.1.3 Cu - Zn

Most copper in igneous rocks is found in sulphide phases but it may also occur in ferromagnesian and feldspathic constituents with a preference for the former. Its concentration increases in successive residual melts until it is high enough for sulphide to precipitate, after which the remaining melt is depleted in Cu (Floyd 1968).

In the Tantalite Valley mafic rocks copper shows a wide scatter in values when plotted against D.I. (Fig 74 ). There is a general positive correlation, with Cu increasing from the ultramafic rocks to the gabbroic rocks. The scatter of points is due to the presence or absence of chalcopyrite and the

fact that copper is especially mobile during alteration. Sulphide minerals were mobilized in part from metagabbros and emplaced into hornfels as veins.

There is less scatter when Cu is plotted against D.I. for the mafic rocks from the Kumkum area (Fig 74 ). Further the various types of gabbro-norite separate out. This is because only sulphide-free samples were analysed.

Although Zn is a chalcophile element, the small amount of sulphur in silicate magmas tends to combine with the abundant Fe and minor Cu and Mo, so that Zn substitutes for  $Mg^{2+}$  and  $Fe^{2+}$  in various ferromagnesian minerals such as pyroxenes or amphiboles in the case of gabbroic rocks (Gurney and Ahrens 1973). From the plot of Zn versus D.I. for Tantalite Valley mafic rocks (Fig 74 ) it can be seen that there is a well-defined trend from ultramafic rocks (av Zn = 76 ppm) to gabbro-norite and metagabbro (both have av Zn = 45 ppm). The range of Zn concentrations from ultramafic rocks to gabbros to anorthositic gabbros of Tantalite Valley is compatible with the values quoted by Gurney and Ahrens (1973) ranging from pyroxenite (80 ppm) to norite/gabbro (39 ppm) to anorthosite (10 ppm) from the Bushveld Igneous Complex, indicating that

the amount of pyroxene present is an important controlling factor of the Zn concentration in the rock. Le Roex (1979) has found that Zn may be partitioned into olivine crystallizing from basaltic magmas ( $K_D = 1.8$ ). That olivine (or sulphide minerals) are not responsible for the trend in Fig ( 74 ) can be seen from peridotite sample T10 which has the lowest Zn content of all the ultramafic rocks in spite of containing 83% olivine and 5% sulphide minerals. The presence of large amounts of biotite in a sample (eg TV265) in addition to pyroxenes may be responsible for Zn contents higher than the average trend. Relatively high Zn content distinguishes the mottled metagabbro from the gabbro and metagabbros.

In the plot of Zn versus D.I. (Fig 74 ) it can be seen that the Kumkum mafic rocks have higher Zn values than mafic samples from Tantalite Valley having a similar range in D.I. values.

#### 8.4.1.4 Sr

Of the major minerals found in gabbroic rocks only plagioclase partitions Sr to any extent (ie  $K_D = 1.8$ , Philpotts and Schnetzler 1970). Of the minor minerals only apatite, sphene and K-feldspar may contain appreciable Sr (Taylor 1965). In basic tholeiitic intrusions such as the Skaergaard

and Stillwater, plagioclase shows increasing Sr and Sr/Ca ratios with increasing fractionation (Berlin and Henderson 1968).

On the Sr versus D.I. plot for the Tantalite Valley samples (Fig 76) there is a clear separation between the various mafic and ultramafic rock types. This is controlled by the amount of plagioclase which is found in each rock-type eg in the ultramafic rocks plagioclase is present as a rare intercumulus phase whereas in the gabbroic rocks it is the main cumulus mineral present. Further metagabbro sample TV126 which has the highest plagioclase content of all the samples (ie 88%) also has the highest Sr-content. Potassium-feldspar is absent from these rocks and apatite and sphene occur in trace amounts so that the effect of these minerals on Sr can be ignored.

The gabbroic rocks from the Kumkum area have Sr contents intermediate between those of the Tantalite Valley ultramafic and gabbroic rocks for D.I. values covering the range of Tantalite Valley gabbroic rocks and mottled metagabbro (Fig 76). The reason for this is that the Kumkum gabbros have less plagioclase than their Tantalite Valley counterparts. The Tantalite Valley gabbro-norite contains on average 63% plagioclase ( $n = 34$ ) compared with

the Kumkum gabbro-norite, which has an average plagioclase content of 46% ( $n = 23$ ). The anomalously high Sr content of Eselruh gabbro-norite sample K59A (Fig 76 ) may be due to this sample containing 2% secondary calcite which may have large amounts of Sr substituting for Ca in its crystal lattice.

#### 8.4.2 Incompatible elements

##### 8.4.2.1 Zr - Nb - Y

These elements are immobile during alteration (Erlank and Kable 1976) and are generally incompatible with respect to all the minerals found in the mafic rocks in the Kumkum - Tantalite Valley area. Zr, Nb and Y are depleted in the ultramafic rocks, metagabbros and gabbro-norites of Tantalite Valley compared with average values of these elements found in tholeiites. The range of Zr concentrations for Tantalite Valley mafic rocks can be seen in the Zr versus D.I. plot (Fig 77). Most of the samples of ultramafic rock, metagabbro and gabbro-norite have Zr concentrations in the range of 10 to 20 ppm. The average Zr content of tholeiites is 112 ppm (Winchester and Floyd 1977). The gabbros from the Kumkum area show a wide scatter of Zr concentrations in Fig ( 77 ). The unrecrystallized mafic rocks, namely the Kumkum and Eselruh gabbro-

norite generally have higher Zr concentrations than mafic rocks of Tantalite Valley with similar values of D.I.

Most of the samples of mafic and ultramafic rocks from Tantalite Valley and the Kumkum area have Nb values below the detection limit of 3 to 4 ppm. The average Nb content of tholeiites is 16 ppm (Winchester and Floyd 1977).

The ultramafic rocks, the metagabbro and gabbro-norite from Tantalite Valley show a narrow range of Y concentrations from 5 to 10 ppm (Fig 77 ). The enrichment of Y above the average value for metagabbro sample TV277 is due to the presence of apatite. The average Y content of tholeiites is 27 ppm (Winchester and Floyd 1977). The mafic rocks from the Kumkum area show a wide scatter of Y values (Fig 77 ) but it can be seen that Y is more abundant in the Eselruh and Kumkum gabbro-norite compared with mafic rocks with similar D.I. values from Tantalite Valley.

An important conclusion which can be reached from the previous discussion is that the source area in the upper mantle from which the mafic and ultramafic rocks of Tantalite Valley and the Kumkum area were derived was depleted in Nb, Y and Zr. Further, the source area of the Kumkum rocks was not the same



as that for the Tantalite Valley rocks being less depleted in Y and Zr.

#### 8.4.2.2 Rb

Rb is very similar in ionic size and chemical character to K and shows a close association with it in rocks (Taylor 1965). Both K and Rb are incompatible with respect to all the minerals found in the ultramafic and mafic rocks from the Kumkum - Tantalite Valley area except phlogopite and biotite. Most of the ultramafic rocks, gabbro-norites and metagabbros of Tantalite Valley have Rb values below the detection limit of about 4 ppm (Fig 76 ) which is in agreement with the value of between 0 and 3 ppm quoted by Erlank et al (1968) for gabbros, norites, anorthosites, pyroxenites and hartzburgites from the Bushveld Igneous Complex. The average Rb content of island arc basaltic tholeiites is 5 ppm (Jakes<sup>v</sup> and White 1972) so that the Tantalite Valley mafic rocks on the whole are depleted in Rb. The mafic rocks from the Kumkum area have higher values of Rb than mafic rocks from Tantalite Valley with similar D.I. values (Fig 76 ). Generally, the Kumkum samples have more biotite or phlogopite than Tantalite Valley samples, which accounts for this

enrichment. The presence of phlogopite or biotite also explains why Rb is enriched in some samples from Tantalite Valley (eg T10, TV265 and TV277).

In ultramafic rock sample T6 from Tantalite Valley, Rb is greatly enriched relative to K. Where a rock is a product of extreme fractionation such as pegmatite, Rb may be concentrated relative to K so that country rock abutting such an intrusion may become enriched in Rb relative to K (Taylor 1965). It is possible that the high Rb content of T6 can be explained in this way since a number of large pegmatite bodies cut the ultramafic rocks. The Rb enrichment effect dies out over the 2 m vertical distance which separates T6 from T7 in the same borehole.

The range of K/Rb ratios for Tantalite Valley and Kumkum mafic rocks can be seen in Fig ( 76 ) where K/Rb is plotted against D.I. For the Tantalite Valley samples, the different rock-types generally separate out into distinctive fields. The gabbro-norites have the highest K/Rb ratios with a range from 400 to about 750 which overlaps that of the metagabbros (range from 200 to 500 ppm). This range is similar to that of 500 - 1 000 quoted by Erlank et al (1968) for most norite and gabbro samples examined from the Bushveld Igneous Complex.

The low K/Rb ratios observed for the Tantalite Valley ultramafic rocks is in agreement with the low K/Rb ratios (200 - 400) given by Erlank et al (1968) for most pyroxenite samples examined from the Bushveld Complex. The Kumkum mafic rocks show a large scatter of K/Rb values (Fig 76 ) but the range of these values is similar to that of the Tantalite Valley metagabbro.

#### 8.4.3 General

As in the major element versus D.I. plots (Fig 73 ) the Contaminated gabbro-norite plots in or near the field defined by mottled metagabbro samples of Tantalite Valley in the trace element versus D.I. plots (Figs 74-77). This adds support to the hybrid origin hypothesis postulated for the Contaminated gabbro-norite. Contamination by granitic fluids may explain the higher Y and Zr contents of the Contaminated gabbro-norite and mottled metagabbro compared with the other gabbroids.

Compared with other tholeiites, the gabbros and ultramafic rocks of Tantalite Valley and to a lesser extent those of the Kumkum area are depleted in Zr, Nb, Y and Rb. This implies that the source area for these rocks in the mantle was also depleted.

The presence of cumulus minerals in these rocks indicates that some of the depletion in incompatible elements observed may be a function of dilution

The different mafic rock-types of Tantalite Valley separate out into fields in V, Co, Sr, K/Rb, Y and Zr versus D.I. diagrams (Figs 74-77). This tendency was also noted in major element versus D.I. plots (Fig 70 ). It should thus be possible to identify unknown samples in plots where separation is achieved.

Compared with the Tantalite Valley gabbro-norite and metagabbro, the Kumkum and Eselruh gabbro-norites have higher concentrations of V, Co, Ni, Zn, Rb, Y Zr (and  $\text{FeO}^*$ , MnO, MgO,  $\text{TiO}_2$ ) and lower concentrations of Sr. The Kumkum and Eselruh gabbro-norites are more hydrous (ie they contain more primary hornblende and biotite) and are more mafic (ie they contain less plagioclase) than the unaltered gabbro-norites of Tantalite Valley. It is concluded that the Kumkum rocks are not less differentiated than those from Tantalite Valley because they have higher concentrations of incompatible trace elements, but were derived from a less depleted source area in the upper mantle.

In view of the resemblance of the Tantalite Valley Complex to Uralian-Alaskan zoned mafic-ultramafic bodies (Moore 1975) and the association of Pt with these rocks some samples of ultramafic rock from Tantalite Valley were analysed for Pt by Johannesburg Consolidated Investment (Pty) (Ltd). All of the values obtained were below the detection limit although what this was could not be ascertained.

## 8.5 Petrogenesis of the Mafic and Ultramafic Rocks of the Tantalite Valley Complex

### 8.5.1 Mafic-Ultramafic Rock Association

Moore (1975) pointed out the similarities which exist between the Tantalite Valley Complex and the so-called "concentrically zoned dunite-peridotite association" (see Wyllie 1967, p 4) typified by the occurrences in Alaska and in the Urals. The features in common are:

1. Multiple Intrusion (Irvine 1967). The Tantalite Valley Complex comprises several different intrusive rock-types which were emplaced at various times.
2. Ultramafic rocks have been emplaced into gabbros (Taylor 1967). In the Uralian-Alaska complexes the ultramafic rocks show no genetic

relationship with the gabbros.

On major element and trace element versus D.I. plots for rocks from the Tantalite Valley Complex, there is usually a distinct break in the trend, going from ultramafic to gabbroic rocks. (In section 8.5.2 this is taken as evidence that some cumulation process has given rise to the two rock-types). The ultramafic cores of these zoned complexes are characteristically younger than the gabbros which enclose them, a feature which distinguishes this association from that of the classical layered complexes such as the Bushveld Igneous Complex where most of the ultramafic rocks, underlie or are older than the gabbros;

3. The bulk of the ultramafic rocks have been formed by crystal accumulation in situ.  
Layering is commonly developed in the ultramafic rocks (Fig 4 ) but is comparatively rare in the gabbros (Irvine 1967).
4. They all occur in orogenic belts (Taylor 1967).
5. These bodies are all small and pipe-shaped and either have ultramafic rocks predominating in volume over gabbroic rocks as is found in Alaska, or they have gabbroic rocks present in larger volumes than ultramafic rocks as is the case in the Uralian occurrences (Taylor 1967) and the Tantalite Valley Complex.

6. The gabbroic rocks have tholeiitic affinity and characteristically contain modal hypersthene (Irvine 1967, Findlay 1969).
7. The ultramafic rocks were intruded after the main periods of folding and metamorphism had affected the orogenic belt (Taylor 1967).
8. The contact between ultramafic rocks and enclosing gabbros is sharp although in the case of the Uralian-Alaskan occurrences the ultramafic bodies have commonly saussuritized and uralitized the gabbros along contacts (Taylor 1967). Whether this has occurred at Tantalite Valley is uncertain because the metagabbro acquired its present mineralogy before the emplacement of the ultramafic rocks (Moore 1975).

However, there are a number of important differences between rocks of the Tantalite Valley Complex and those from zoned complexes from the type areas in Alaska and the Urals. These are:

1. The Tantalite Valley Complex is Precambrian in age whereas the zoned complexes found in the Urals and Alaska are respectively Palaeozoic (Devonian) and early to middle Cretaceous (Taylor 1967).
2. The ultramafic rocks of the Tantalite Valley

Complex are not zoned, a feature which is typical of ultramafic rocks from the Uralian-Alaskan zoned Complexes (Taylor 1967).

3. Mineralogy-Clinopyroxene, primary hornblende and abundant primary magnetite are common in the gabbros and ultramafic rocks of the Uralian-Alaskan zoned complexes (Taylor 1967; Moore 1973; ) but are scarce to absent in comparable rock-types from the Tantalite Valley Complex. Clinopyroxenes from ultramafic rocks from Uralian-Alaskan zoned complexes have high  $\text{Al}_2\text{O}_3$  contents ( $\sim 4 - 7$  weight%) (Taylor 1967) compared with the 1,75 - 3,50 weight% concentration found in clinopyroxenes from gabbro-norite and ultramafic rocks from the Tantalite Valley Complex (Table 43). Orthopyroxene and plagioclase are absent from the ultramafic rocks of the Uralian-Alaskan zoned complexes (Irvine 1967; Taylor 1967) but may be quite abundant (especially orthopyroxene) in the ultramafic rocks of the Tantalite Valley Complex. Olivine is more magnesian (up to  $\text{Fo}_{95}$ ) (Taylor 1967) in ultramafic rocks from zoned complexes compared with those from the Tantalite Valley Complex (Table 43) while dunites which are a common constituent of the cores of the zoned complexes are not represented



in the ultramafic rocks of the Tantalite Valley Complex. Olivine gabbros are rare in the zoned complexes but are a common constituent of the gabbros into which the ultramafic rocks have been intruded in the Tantalite Valley Complex.

4. Ultramafic rocks of Uralian-Alaskan Complexes may be intruded by pegmatitic hornblende-plagioclase rocks (Irvine 1967) whereas in the Tantalite Valley Complex a similar rock-type is found only in the metagabbro where it is a residual component of the latter. Further, pegmatitic hornblende-plagioclase rocks from Alaska are characterized by anorthitic plagioclase (Irvine 1967) in contrast to the andesine plagioclase found in similar rocks forming part of the metagabbro of Tantalite Valley.
5. Chemistry - Ultramafic rocks from zoned complexes are silica undersaturated, which is reflected in the presence of such minerals as nepheline and leucite in C.I.P.W. norms, and have alkali-basalt affinities (Irvine 1967) compared with ultramafic rocks from the Tantalite Valley Complex which mostly have normative olivine and pyroxenes, and have tholeiitic affinities. Platinum may be found associated with the ultramafic rocks of zoned complexes (Taylor 1967) but is absent from similar rock-

types from the Tantalite Valley Complex.

The Tantalite Valley Complex cannot be classified as belonging to the "Alpine-type peridotite-serpentinite association" (see Wyllie 1967, p 4) because of such features as the absence in the former of serpentinites, highly magnesian olivine, podiform chromite deposits, complicated structural relations between gabbroic and ultramafic rocks and structures and textures characteristic of high-grade metamorphic rocks (Thayer 1967).

The Tantalite Valley Complex resembles the Permo-Triassic Yakuno Complex in Japan in that the latter consists of metagabbros which have been intruded by minor volumes of granitic rocks and later by ultramafic rocks (Igi 1973). The Yakuno Complex was however emplaced into a high pressure metamorphic terrane and the mafic rock components have an alkali basalt to high-alumina basalt affinity (Igi 1973). There is also some similarity between the Tantalite Valley Complex and the Mecklenburg Complex of North Carolina (Hermes 1968, 1970), because each has a regionally metamorphosed metagabbro intruded by a post-metamorphic gabbro. However, on a total alkalis versus silica diagram (see Kuno 1968) the Mecklenburg gabbros plot in the "high alumina basalt series" field compared to the

Tantalite Valley Complex metagabbro and gabbro-norite which fall in the field of tholeiites. Further, the Mecklenburg gabbros contain more clinopyroxene and primary hydrous phases, such as biotite and hornblende, compared with those from Tantalite Valley.

From the previous discussion it can be seen that the Tantalite Valley Complex, although it has some features in common with Uralian-Alaskan zoned complexes, cannot be classified definitively into any one of the ultramafic-mafic associations as defined by Thayer (see Wyllie 1967, p 4). It was shown in section ( 8.2) that the mafic and ultramafic rocks of Tantalite Valley have tholeiitic affinities and that some of the ultramafic rocks in particular are equivalent chemically to "serpentinites" from large layered complexes. In the Uralian-Alaskan complexes the ultramafic and gabbroic rocks are not even of the same magma type and at least two periods of partial melting of the upper mantle at different depths are usually invoked to explain their association (Taylor 1967). In the following section the relationship between the gabbros and ultramafic rocks of Tantalite Valley are discussed.

### 8.5.2 Relationship between Gabbroic and Ultramafic Rocks of the Tantalite Valley Complex

The fact that both the gabbros and ultramafic rocks of the Complex are tholeiitic suggests that they may be related by fractional crystallization. In order to assess the rôle of cumulus phases in controlling the compositions of mafic rocks of the Tantalite Valley Complex a number of plots were prepared for which major elements are represented, recalculated volatile free, together with mineral analyses of plagioclase, olivine, clino- and ortho-pyroxene. In the  $Al_2O_3$ , CaO and  $Na_2O$  versus MgO diagrams (Fig 78 ) the rôle of plagioclase can be evaluated. On all three diagrams there is a tendency for gabbroic and ultramafic samples to separate out with little or no overlap between the two clusters of points. The trend defined by ultramafic rocks also has a much shallower slope than for that of gabbroic rocks. In the  $Al_2O_3$  versus MgO diagram (Fig 78 ) the control of gabbroic rock compositions by the amount of plagioclase and pyroxene present in the rock is evident. In the CaO versus MgO diagram (Fig 78 ) clinopyroxene is seen as not being significant in controlling the overall composition of the gabbros, but rather that orthopyroxene, plagioclase and

olivine are involved. The  $\text{CaO}$  versus  $\text{Al}_2\text{O}_3$  diagram (Fig 79 ) emphasizes the unimportance of clinopyroxene in influencing bulk rock compositions.

Plots of  $\text{MgO}$  versus  $\text{SiO}_2$  (Fig 79 ),  $\text{FeO}^*$  versus  $\text{MgO}$  (Fig 78),  $\text{FeO}^*/\text{MgO}$  versus  $\text{MgO}$  (Fig 78 ) and  $\text{TiO}_2$  versus  $\text{MgO}$  (Fig 78 ) were carried out in order to test the above conclusions reached concerning mafic minerals. In the  $\text{MgO}$  versus  $\text{SiO}_2$  diagram, (Fig 79 ) points representing ultramafic rock samples define a control line between olivine and orthopyroxene compositions whereas points representing gabbros lie along a line which can be drawn between plagioclase and orthopyroxene compositions, although there is some displacement from this trend as a result of the minor influence of clinopyroxene. Ultramafic sample TV117 lies within the triangle defined by olivine, orthopyroxene and plagioclase compositions which is an indication of its probable mineralogy before metamorphism to a tremolite-actinolite-chlorite rock. In the  $\text{FeO}^*$  versus  $\text{MgO}$  diagram (Fig 78 ) the composition of the mafic and ultramafic rocks can be seen to be influenced by plagioclase, orthopyroxene and olivine. In the  $\text{FeO}^*/\text{MgO}$  versus  $\text{MgO}$  and  $\text{TiO}_2$  versus  $\text{MgO}$  diagrams (Fig 78 ) there is a definite separation of gabbros (metagabbro and gabbro-norite), ultramafic rocks and

mottled metagabbro of the Tantalite Valley Complex into separate fields. Plagioclase-rich metagabbro sample TV126 is anomalous in that it is depleted in MgO and enriched in FeO\*, compared with other gabbro samples (Fig 70 ). The constant FeO\*/MgO ratio of the ultramafic rocks in Fig ( 78 ) suggests that these rocks were all derived from the same source if a partial melting model with equilibrium melting is assumed.

A diagram in which  $\text{MgO}/\text{Al}_2\text{O}_3$  is plotted against  $\text{SiO}_2/\text{Al}_2\text{O}_3$  (Pearce 1976) for the gabbro-norites and ultramafic rocks is given as Fig ( 81 ). The quantities MgO,  $\text{Al}_2\text{O}_3$  and  $\text{SiO}_2$  were converted into molecular proportions from the values obtained when the bulk analysis was recalculated volatile free. For the ultramafic rocks three possible accumulation trends are revealed, namely: olivine accumulation, orthopyroxene accumulation and a combination of these. The feasibility of the first two are shown by the presence of samples containing large proportions of olivine (ie T10) and orthopyroxene (ie T11) at the limits of these trends. Serpentinization of olivine causes a change in slope of the olivine line ie Mg:Si changes from 1:1 to 3:2. The line which can be drawn bisecting the olivine and orthopyroxene accumulation trends is therefore taken to represent the case where both

olivine and orthopyroxene were accumulated. These findings are in agreement with conclusions reached above concerning the minerals which influenced ultramafic rock compositions. The gabbronorites in comparison with the ultramafic rocks have undergone little or no fractional crystallization as is seen by the limited spread of points in Fig ( 81 ). The compositions of gabbronorite and metagabbro are almost identical when average values are compared (Table 45 ) so that the metamorphism of the parent rock to form metagabbro was probably isochemical. Thus the metagabbro samples can be omitted from Fig ( 81 ) without any major loss of information. The relationship between the gabbronorite and the ultramafic rocks in Fig ( 81 ) is not clear but they could be related by fractional crystallization.

Although it cannot be conclusively proved that the ultramafic and gabbroic rocks are related, it is tentatively postulated here that they were derived from a common parent by a crystal fractionation process. This would explain why there is commonly a hiatus in the trends observed between gabbroic and ultramafic rocks in plots concerning major and trace elements. Unfortunately, no chilled contacts were observed in the rocks making up the Complex so that this model cannot be quantitatively tested by

calculating least squares approximations to the major element composition of each proposed derivative rock-type using the method of Bryan et al (1969) because the composition of the parent is unknown.

### 8.5.3 Origin of the Tantalite Valley Complex

With respect to the origin of the Tantalite Valley Complex, the mechanisms proposed by Igi (1973) and <sup>V</sup>Misar (1971) for similar complexes elsewhere are of relevance. Both models involve the separation of gabbroic magma from ultramafic cumulates by a process of fractional crystallization in a magma chamber deep in the earth's crust and a fracture zone penetrating as deeply as the upper mantle is seen as providing the channelway for ascending magmas. In the case of the origin of the Tantalite Valley Complex, a combination of the hypotheses of Igi (1973) and <sup>V</sup>Misar (1971) can explain most of the features observed.

It is proposed here that partial melting of depleted upper mantle material gave rise to liquids of high  $Al_2O_3$  olivine tholeiite composition which rose along a deep fault zone, the Tantalite Valley Lineament, into a magma chamber situated at a depth of between 20 and 30 km in the crust



(Misar<sup>✓</sup> 1971). Some of the magma may have reached the surface or close to it where it was intruded or extruded (ie the proto-mottled metagabbro).

Partial remobilization of granite gneiss led to the development of mottled metagabbro from the proto-mottled metagabbro by metasomatism. Meanwhile, under the influence of fractional crystallization in the magma chamber, olivine and orthopyroxene cumulates settled out under gravity, sinking to the floor of the chamber leaving behind more iron-rich liquids containing phenocrysts of plagioclase and some olivine. These magmas then rose upward until they crystallized as gabbro at a depth of about 20 km in the crust, which is equivalent to pressures of about 5 kb derived from mineral assemblages in hornfelses which were formed along the contacts of the country rock and the gabbro.

The emplacement of the gabbro also led to partial melting in adjacent rocks giving rise to the local development of tonalites and granodiorites. The gabbro underwent slight differentiation as it cooled as is seen by the presence of (a) layering near the contacts with country rocks, (b) quartz-bearing and quartz-free varieties and (c) residual pegmatitic fractions. A period of regional metamorphism followed which converted the proto-mottled

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metagabbro and gabbro into essentially hornblende plus plagioclase-bearing rocks, ie metagabbros. This process was essentially isochemical. At some later time a further period of partial melting in the upper mantle led to a new injection of magma into the magma chamber which then underwent fractional crystallization as before to yield orthopyroxene-olivine cumulates and more iron-rich liquids enriched in cumulus plagioclase and some olivine. This liquid then rose along the fault zone to be intruded eventually into zones of weakness in metagabbro as sheets and plugs of gabbro-norite. Minor differentiation of olivine gabbro-norite then gave rise to quartz-bearing gabbro-norites. Finally, possibly as a result of tectonic influences, the residual ultramafic crystal mush in the magma chamber rose along the fracture zone and found its way into zones of weakness in the solidified gabbroic rocks where it crystallized fully. The emplacement of these crystal mushes would occur at temperatures closely approaching that of basaltic magmas (Wyllie 1967, p 405). The gabbro-norite satellite body rocks, which are chemically and mineralogically intermediate between gabbroic and ultramafic rocks and are enriched in  $K_2O$ ,  $P_2O_5$  and the incompatible trace elements, most likely represent a contaminated variety of ultramafic rock.

It is noteworthy that in the study area, ultramafic rocks are only found along the axis of the Tantalite Valley Lineament even though mafic rocks are fairly abundant north and south of this line (Jackson 1976). This is probably due to the centre of the Shear Zone providing the major channelway for ultramafic rocks tapped at depths reaching the surface.

#### 8.5.4 The Origin of the Kumkum and Eselruh gabbronorite

The trend noted by Jackson (1976) towards mafic rocks enriched in  $\text{SiO}_2$  and  $\text{K}_2\text{O}$  on going northward from the axis of the Tantalite Valley Lineament may be a function of increasing crustal contamination. In the region of the axis of the shear zone, magmas would rise quickly and undergo relatively little contamination but further away from the shear zone, magmas would find it more difficult to rise, and during their slower ascent, would be more liable to incorporate  $\text{K}_2\text{O}$  and  $\text{SiO}_2$  and incompatible elements from the surrounding rocks. This hypothesis is however considered unlikely to explain the chemistry of the Kumkum and Eselruh gabbronorites since these rocks show no evidence of crustal contamination such as xenoliths. The Kumkum and Eselruh gabbronorites are more mafic than similar

rock-types from the Tantalite Valley area (see section 8.4.3). They cannot be said to be less differentiated equivalents of the Tantalite Valley rocks since they contain higher concentrations of incompatible elements. The chemical differences between Kumkum and Eselruh gabbronorites and those from Tantalite Valley are here ascribed to lateral inhomogeneity in the source region in the upper mantle. The magma from which the Kumkum gabbronorite crystallized was also more volatile-rich than the magmas from which the Tantalite Valley gabbronorite crystallized. This is seen by the common occurrence of magmatic hornblende and biotite in the Kumkum gabbronorite compared with the rarity or absence of these minerals in the Tantalite Valley gabbronorite. The age relationship of the Eselruh gabbronorite to the Kumkum gabbronorite is uncertain, since no contact relationships exist, although by analogy with Tantalite Valley where the ultramafic rocks are younger than the gabbros, it is suspected that the Eselruh gabbronorite which is more mafic than the Kumkum gabbronorite, is younger than the latter. These gabbronorites, like those of Tantalite Valley, are tholeiitic and have Ni-Cu sulphides present in variable amounts.

Control line diagrams for the mafic rocks of the Kumkum area, using  $\text{Al}_2\text{O}_3$  and CaO against MgO, and

including the compositions of the essential minerals (Fig 80 ), shows that points representing the Kumkum and Eselruh gabbronorite samples form a trend line between plagioclase and orthopyroxene/olivine. The deviation of sample K59A from this control line in the CaO versus MgO diagram (Fig 80 ) is due to the presence in this sample of 70% clinopyroxene and secondary hornblende and 2 modal% secondary calcite. As in the case of the Tantalite Valley mafic rocks, clinopyroxene generally does not influence the composition of these rocks to any extent, with the exception of Eselruh gabbronorite.

#### 8.5.5 Magma Genesis and Source Characteristics

It is generally accepted that the upper mantle is peridotitic (Green and Ringwood 1967a; O'Hara 1968) and, when partially melted as a result of an increase in heat flux or a decrease of pressure, can give rise to liquids of basaltic or picritic composition (Hess 1960; Eaton and Murata 1960; Hurley et al 1962; Yoder and Tilley 1962; Green and Ringwood 1967a). According to Arndt (1977) at least 8% (by volume) partial melting is required before magma segregates from residual refractory crystals which probably have a composition similar to peridotite or dunite (Green and Ringwood 1967a). From this point, the magma behaves independently of the crystal residuum, and fractionates by cooling and

crystal settling on its way to the surface. The chemical composition of the magma before it starts fractionating is a function of the depth at which magma segregation rather than the depth at which partial melting occurs (see Yoder and Tilly 1962; Kushiro 1968). Quartz-normative tholeiitic liquids can only be generated by partial melting of parental peridotite at depths of 0 - 15 km (ie  $< 5$  kb) (Boyd et al 1964). Below this depth magmas will always be olivine-normative. Segregation of a 20 - 25% partial melt derived from parental peridotite at depths of between 15 and 35 km (ie 5 to 12 kb) will yield a high  $Al_2O_3$ -olivine tholeiite magma (Green and Ringwood 1967a). At depths between 35 and 70 km basaltic liquids segregating will vary in composition from alkali olivine basalt to olivine-rich tholeiite, depending on the percentage of partial melting. The above assumes dry melting and no influence by volatiles such as  $H_2O$  and  $CO_2$ .

In the Kumkum-Tantalite Valley area the mafic and ultramafic rocks are essentially tholeiitic in character, but have high  $Al_2O_3$ . Tholeiite basalt is, according to Wedepohl (1969), the commonest member of the basalt series and is the most abundant volcanic rock to have been extruded throughout geological time. The generation of this basalt-type requires a large amount (20 - 30%) of partial

melting (Gast 1968). The metagabbro and gabbro-norite of Tantalite Valley and the Kumkum gabbro-norite have average compositions similar to those given for high- $\text{Al}_2\text{O}_3$  olivine tholeiites dredged from along mid-ocean ridges (see sections 8.2.4 and 8.3.4). According to the experimental work of Green and Ringwood (1967a), rocks with these compositions can be derived by 20 - 25% partial melting of an assumed parental peridotite (or pyrolite) (62% olivine, 19% orthopyroxene, 11% clinopyroxene, 4% plagioclase, 4% minor phases (phlogopite, spinel, ), at 1 250°C followed by magma segregation at a depth of 15 - 35 km (ie 5 to about 12 kb). Crystal fractionation of this liquid at lower pressures under closed system conditions will lead to enrichment in  $\text{Al}_2\text{O}_3$  due to removal of olivines and pyroxenes and to the production of quartz-tholeiites. Separation of both olivine and low-alumina enstatite, joined at lower temperatures by sub-calcic clinopyroxene from this liquid will occur and accumulate as troctolites and peridotitic layered complexes (Green and Ringwood 1967a, Fig 13).

The occurrence of troctolites and peridotitic rocks associated with gabbro-norites from Kumkum and Tantalite Valley lends support to this hypothesis except that they apparently do not occur as the cumulate fraction, but rather as late-stage intrusives.



Further, textures in cumulate rocks from the study area indicate that olivine and orthopyroxene were important liquidus phases in these rocks. From the experimental work of Green and Ringwood (1967a) on olivine-tholeiite compositions, this indicates that crystallization of these rocks took place at pressures less than 13.5 kb and probably less than 9 kb since plagioclase accumulation is important in the gabbro-norites. Orthopyroxenes from the mafic rocks of the Kumkum-Tantalite Valley area have lower  $\text{Al}_2\text{O}_3$  contents than those from olivine-tholeiites crystallized at 9 kb (compare the  $\text{Al}_2\text{O}_3$  values of orthopyroxene given in Table 43 with those in Table 9, Green and Ringwood 1967a). This implies that the orthopyroxenes from the study area formed under pressures lower than 9 kb. Independent evidence from the Tantalite Valley aureole indicates that the prevailing pressure at the time of emplacement of the metagabbro was around 5.5 kb which serves as a minimum estimate.

In the ultramafic rocks of the Uralian-Alaskan zoned complexes magmatic hornblende is abundant and orthopyroxene is absent, suggesting that the water content of these magmas

was much

higher than that for the magmas from which the Kumkum-Tantalite Valley mafic rocks were derived.

Yoder and Tilley (1962) showed that clinopyroxene and amphibole in rocks of basaltic composition have a wide range of stability at less than 2 kb  $P_{H_2O}$  compared with the same system at higher pressures. Further, Eggler and Burnham (1973) demonstrated that andesitic magmas do not crystallize magnetite at water pressures greater than 2 kb. Thus the presence of abundant magnetite, clinopyroxene and hornblende in ultramafic rocks from Uralian-Alaskan complexes implies that water pressure was below 2 kb.

Compared with other tholeiitic rocks, those from the Kumkum-Tantalite Valley area are depleted in incompatible elements which is taken to indicate that the source area is depleted (see section 8.4.3). This implies that the source area has undergone an earlier partial melting episode. Magmatism was probably a function of tectonism along the Tantalite Valley Lineament.

High  $Al_2O_3$ -olivine tholeiites are characteristically found along zones of tension eg the mid-oceanic ridges so that the occurrence of this rock-type in an area transected by a major fault, the Tantalite Valley Lineament, is predictable.

#### 8.6 Geochemistry of Amphibolites from Tantalite Valley and the Kumkum area

Beukes (1973, p 132) has called the amphibolites which occur within his Umeis Formation (ie Grey Gneiss sequence), para-amphibolites, based on their concordant and gradational contacts with the host rocks. He believes that the amphibolites (including epidosite) represent the metamorphic products of marls which were intercalated within a greywacke succession. On the other hand, Toogood (1976, p 27) working to the east of the present study area, described amphibolites showing cross-cutting relationships with foliation and with other amphibolites as is found in dyke and sill complexes, features which argue convincingly for the presence of ortho-amphibolites in the area. In the present study area, field relationships such as have been described above were observed indicating that both ortho- and para-amphibolites are probably present. In addition, a third variety, here called metasomatic hornblendite, was distinguished.

In this study only a limited number of samples of amphibolite have been chemically analysed, so that this section serves mainly as a starting point for further more detailed investigations. The usual chemical methods to evaluate whether amphibolites are igneous or sedimentary in origin depend on examining a large suite of samples so as to define a

trend which controls the distribution of major and trace elements. (Evans and Leake 1960; Leake 1964). Since only six amphibolite samples from Tantalite Valley and the Kumkum area were analysed for major and trace elements, the approach used here rather has had to depend largely on methods for distinguishing the origin of amphibolites from single samples.

Shaw and Kudo (1965) have devised a method for distinguishing para- and ortho-amphibolites using empirically derived discriminant functions based on the content and distribution of major and minor elements. Although these authors stress that their apparently successful method cannot necessarily be applied to rocks of other areas and of other ages with expectation of like results, it is nevertheless interesting to compare values obtained for amphibolites from the present study area. If the discriminant value  $X_3$  ie:

$$X_3 = 7,07 \log \text{TiO}_2 + 1,91 \log \text{Al}_2\text{O}_3 - 3,29 \log \text{Fe}_2\text{O}_3 + 8,48 \log \text{FeO} + 2,97 \log \text{MnO} + 4,81 \log \text{MgO} + 7,80 \log \text{CaO} + 3,92 \log \text{P}_2\text{O}_5 + 0,15 \log \text{CO}_2 - 15,0$$

is calculated for major and minor elements of the amphibolites, then positive values are obtained, indicating that the amphibolites may be of igneous origin.

Evans and Leake (1960) and Leake (1964) have shown that Cr, Ni and Co are the most useful trace elements

in distinguishing between sedimentary and igneous trends in amphibolites. This is because in the differentiation of igneous rocks, a decrease in Niggli mg is accompanied by a decrease in these trace elements, whereas mixtures of pelite and limestone, or dolomite, show an increase in Cr, Ni and Co with a decrease in the Niggli mg value, since limestone and dolomite are poorer in these trace elements compared with pelites. It is assumed that during metamorphism, individual rock layers acted essentially as closed systems and that metasomatism did not occur to a significant degree. Orville (1969) states that the presence of assemblages containing muscovite or an aluminosilicate within, or in the immediate surroundings of amphibolite layers would rule out the possibility that metasomatism has played a significant part in the formation of the amphibolite. This statement is valid for most of the amphibolites from the study area, which occur as layers in the Grey Gneiss sequence, a unit which is characterized by the presence of aluminosilicates or muscovite. However, the Metasomatic hornblendite, as its name implies, can be seen in the field to have formed along the contacts of pegmatites and ultramafic rock, mainly as a result of the metasomatic introduction of  $K_2O$  from pegmatite into the latter, as

is seen by the development of abundant phlogopite in the reconstituted ultramafic rock.

Based on the theory outlined by Evans and Leake (1960) and Leake (1964), Van de Kamp (1969) devised plots of Cr versus 100 Niggli mg and Ni versus 100 Niggli mg to distinguish between ortho- and para-amphibolites, which are reproduced here as Figs 83 and 82 respectively. Also included on these plots are the average of eight ortho-amphibolite analyses (NAMORTHO) and one para-amphibolite analysis from the Namiesberg (D86) (Moore 1977) (Table 44), and the average of rocks from the present study area which correspond to amphibolites mineralogically and are indubitably of igneous origin. These include the metagabbro proper, satellite body metagabbro and mottled metagabbro. Metabasite hornfels analyses are also represented but only on the Cr versus 100 Niggli mg diagram, since nickel sulphides may occur in these samples. Most of the samples represented on Figs ( 82 ) and ( 83 ) define a trend which is opposite from that defined by the shale-carbonate rocks. Only amphibolite samples AM76 and AM113 plot within the field of shale-carbonate rocks in both Figs ( 82 ) and ( 83 ). Metabasite hornfels sample TV13 98 plots just outside the shale-carbonate field in Fig ( 83 ) but lies well away from the trend defined by rocks

of igneous origin, suggesting that it may also have a sedimentary origin. Para-amphibolite sample D86 from the Namiesberg plots in or near the field of shales and carbonate rocks in both Figs ( 82 ) and ( 83 ) as expected. Amphibolite sample K<sup>40</sup> from the Kumkum area is probably of sedimentary origin since it plots in the shale-carbonate field in the Cr versus 100 Niggli mg diagram. (Fig 83 ). No conclusion as to the origin of this sample can be reached in the Ni versus 100 Niggli mg diagram (Fig 82 ) since this specimen contains Cu-Ni sulphides which have probably been introduced from other mafic rocks in the area since sulphides occur only along joints in this sample. The Metasomatic hornblendite sample AM157 is characterized by high Cr and Ni contents, which reflects its origin from an ultramafic parent.

Although Rb and Sr are mobile trace elements, Rb/Sr ratios may be useful indicators of rock origin as inferred from the data of Faure and Hurly (1963). Basic igneous rocks generally have ratios well below 0,15 and the ratios in shales are about 0,50 or higher. Para-amphibolites have an average Rb/Sr ratio of 0,11 with a range of values from 0,01 to 0,58 (Van de Kamp 1968). For the amphibolites from the present study area, only samples AM113 and AM157

have Rb/Sr ratios which suggest an undoubtedly sedimentary origin. For sample AM113 this conclusion is considered as valid but for AM157 the result is regarded as meaningless since large amounts of K and thus Rb were introduced into this rock during metasomatism, which can be seen by comparing the chemical data for this rock with that of the average ultramafic rock from Tantalite Valley (see Table 44 and Table 45).

The high Sr contents of samples AM76 and K40 makes an igneous genesis unlikely since rocks of basaltic composition contain on average 465 ppm Sr (Turekian and Wedepohl 1961). The plagioclase in sample AM76 is highly calcic (ie  $An_{95}$ ) and since Sr is known to substitute for Ca and since in a rock containing mainly hornblende and plagioclase, the latter will partition most of the Sr present (Tauson 1965), the high Sr content of the rock is to be expected. The high Zr content of AM76 compared with mafic igneous rocks of the Tantalite Valley-Kumkum area favours a sedimentary origin for this sample (see Table 44 and Table 45).

The concentration of copper in the amphibolites (Table 44) shows a wide range, from values below the detection limit ( $< 1,5$  ppm) to 738 ppm. This is in agreement with the findings of Floyd (1968)



who states that the Cu content in ortho-amphibolites varies from about 5 to 500 ppm with an average value of 107 ppm. The generally low concentrations of Cu in the amphibolites examined here may be due to the relatively high mobility of this element during metamorphism (Floyd 1968). According to Van der Kamp (1968), Cu, Y, Zn and V are of little use in distinguishing the parentage of amphibolites.

The ACF diagram (Fig 90) allows a further check on the origin of the amphibolites in question. Samples AM76 and K94 plot in the field corresponding to marls with 35 to 65 percent carbonate while all other samples (AM77, AM113, K40 ) fall in the field occupied by basaltic rocks. This does not however rule out a sedimentary origin for AM113 and K40 since this field overlaps that of calcic marls. Sample AM157 plots unequivocally in the field of ultrabasic rocks.

From the above, it is concluded that AM76, AM113 and K40 are likely to be of sedimentary origin while the rest of the samples, including AM157, have probably been derived from igneous rocks. The metasomatic origin of AM157 must, however, be borne in mind when considering its possible parentage.

Since the minor and trace elements  $\text{TiO}_2$ ,  $\text{P}_2\text{O}_5$ , Nb,

Zr and Y are considered as being relatively immobile during metamorphism, Winchester and Floyd (1976) devised a number of plots using these elements to determine the magma-type from which rocks which are now represented as ortho-amphibolites have been derived. Included on these diagrams are the fields occupied by alkaline and tholeiitic rocks. In addition to including ortho-amphibolites from the study area, rocks equivalent to ortho-amphibolites such as the metagabbro proper, metagabbro from the satellite body and mottled metagabbro as well as the two para-amphibolite samples AM76 and AM113 are included in the diagrams for comparison. From the plots of  $\text{TiO}_2$  versus Y/Nb (Fig 84 ),  $\text{P}_2\text{O}_5$  versus Zr (Fig 85 ),  $\text{TiO}_2$  versus  $\text{Zr}/\text{P}_2\text{O}_5$  (Fig 86 ) and Nb/Y versus  $\text{Zr}/\text{P}_2\text{O}_5$  (Fig 87 ) it can be seen that these samples generally plot in the tholeiitic field. The anomalous behaviour of satellite body metagabbro sample AM125 in the  $\text{P}_2\text{O}_5$  versus Zr plot (Fig 85 ) and of the Metasomatic hornblendite sample AM157 in the Nb/Y versus  $\text{Zr}/\text{P}_2\text{O}_5$  plot (Fig 87 ) can probably be ascribed to the fact that the original chemistry of both of these rocks has been altered by contamination and metasomatism respectively.

### 8.6.1 The occurrence of garnets in amphibolites

Amphibolites from terrains metamorphosed under medium to relatively high pressures are characterized by the presence of garnet, whereas in low pressure metamorphic terrains, eg the present study area, garnet is rarely found in amphibolites (see Beukes 1973; Toogood 1976). Leake (1972) showed that Niggli mg , Oxidation Ratio (O.R.) (see Table 48 ) and MnO content of the rock in percent are the main variables controlling the occurrence of garnet in amphibolites in Connemara. He considered that it was possible to use the presence or absence of garnet in amphibolite as a geothermometer if rocks of similar composition were used. He devised the garnet formation parameter,  $g = \text{Niggli } \underline{\text{mg}} \times \frac{\text{O.R.}}{\text{MnO\%}}$  where  $g$  diminishes as the rock composition becomes more favourable for garnet formation. Or, stated in another way, "as the metamorphism becomes more favourable for garnet formation, the maximum size of this number will increase in garnetiferous amphibolites". (Leake 1972, p 659). Garnet is not found in amphibolites analysed here above a value for  $g$  of 0,756 and probably not above a value for  $g$  of 0,372 (Table 48 ) which indicates that the pressure of metamorphism was below that at which the Green Beds of the

Dalradian of S.W. Scotland were metamorphosed. Garnetiferous amphibolites from the latter area have an average value for  $g$  of 1.31 and this is found in rocks which have been postulated as having been metamorphosed at pressures of about 7 kb. (Leake 1972).

Although no garnetiferous amphibolites have been analysed from the present study area, such a rock has been analysed from the Namiesberg, an area in the Namaqua Metamorphic Complex which was metamorphosed under similar conditions to that found in the Tantalite Valley-Kumkum area (Moore 1977). The value of  $g$  obtained from this sample (ie A21) was 0,08 which is approximately the same as the average value (ie 0,12) calculated by Leake (1972) for garnetiferous amphibolites from Connemara in Ireland, where metamorphism took place at around 5 kb (Leake 1972). This is not an unrealistic estimate for the pressure at which metamorphism took place in the Kumkum-Tantalite Valley area (see section 5.4 ).

#### 8.7 Geochemistry and origin of Granitic Rocks from the Kumkum-Tantalite Valley area

The term 'granitic rocks' encompasses all coarse-grained quartzo-feldspathic rocks of

igneous origin (Middlemost 1969), a group which can be further divided into alkali granite, granite sensu stricto, granodiorite and quartz-diorite (incorporating trondhjemite and tonalite) on the basis of the relative amounts of sodic and potassic feldspars present (Streckeisen 1967).

Whole rock analyses and C.I.P.W. norms of granitic rock samples from the Kumkum-Tantalite Valley area are given in Tables ( 44 ) and ( 47 ). Like most granitic rocks, these are peraluminous (Shand 1952). Since most of the  $K_2O$  and  $Na_2O$  present in granitic rocks is contained in feldspars, some workers (eg Harpum 1963) have used the  $K_2O/Na_2O$  ratios of granitic rocks to supplement the classification based on the relative amounts of feldspars present. This is done on a diagram (Fig 88 ) where  $K_2O$  is plotted against  $Na_2O$ , where both quantities are in weight % and calculated volatile free, and includes the fields occupied by granite, adamellite, granodiorite and tonalite. In Fig ( 88 ) Inequigranular granite gneiss from the Kumkum area (ie K49 and GBW744), Warmbad granite (ie GBW458 and 608) and sample TV299 from east of the Tantalite Valley Complex plot in the granite field while samples TV305 (tonalite sensu stricto) and TV306 (Inequigranular granite gneiss) from west of the Tantalite Valley Complex, plot in

the tonalite field. On the basis of modal analysis, sample TV299 was termed a granodiorite using the nomenclature of Streckeisen (1967).

Phillips (1972) has shown that where perthitic or antiperthitic feldspars are present in a granitic rock, the name given to such a rock, based on the recorded feldspar percentages present (Streckeisen 1967) may be grossly incorrect. The presence of abundant antiperthite in sample TV299 is probably the main contributing factor towards the disagreement between names given on the basis of the chemistry and the modal analysis.

Granitic rocks are also classified according to:

1. whether they were intruded before, during or after an orogeny, termed pre-, syn-, or post-tectonic (Marmo 1967).
2. the depth at which they were emplaced. The adjectives Catazonal, Mesozonal and Epizonal are used to qualify granitic rocks emplaced into successively lower levels in the crust (Buddington 1959).
3. those which have originated from anatexis of rocks of sedimentary origin (S-type granitoids) and those which have been derived directly from pre-existing rocks of igneous origin (I-type granitoids) (White and Chappell 1977). S-types

have metasedimentary xenoliths whereas I-types contain mafic hornblende-rich xenoliths. S-types are peraluminous and have low Na/K ratios compared to I-types where total Na, K and Ca are high relative to Al and Na/K is high.

Beukes (1973), using the criteria listed by Marmo (1967) and Buddington (1959) for the recognition of the environment of formation of granitic rocks classified the Inequigranular granite gneiss (Eendoorn granite) as a syntectonic granite of the Catazone, and the younger Warmbad granite as a post-tectonic granite of the Mesozone. The so-called tonalites and granodiorites associated with the Tantalite Valley Complex can be described as late syn-tectonic to post-tectonic granitic rocks of the Mesozone using some of the criteria listed by Buddington (1959) since they: (1) display discordant and concordant contact relationships with the country rocks (2) they are mainly foliated where they are in contact with other rocks, otherwise they are massive (3) they are not related in any way to volcanic rocks or their metamorphosed equivalents and (4) they may be associated with pegmatites. All of these granitic rocks are peraluminous which indicates that they are S-types. In addition, Na/K ratios are high in, Inequigranular

granite gneiss samples from the Kumkum area, (which may contain metapelite restite minerals like sillimanite, cordierite and garnet), Warmbad granite samples and a sample of "granodiorite" associated with the Tantalite Valley Complex (Fig 88 ).

However, an Inequigranular granite gneiss sample and tonalite sample from the eastern end of the Complex have low Na/K ratios and these rocks may contain xenoliths of mottled metagabbro in outcrop. No garnet, cordierite or sillimanite were recorded in the Inequigranular granite gneiss from Tantalite Valley. The Tantalite Valley Inequigranular granite gneiss is found only along, or north of, the axis of the Tantalite Valley Lineament, the domain in which partial melting has occurred, which favours the interpretation that this rock-type, in spite of possessing certain characteristics of I-type granites, formed as a result of anatexis during regional metamorphism. Magma mixing of Tantalite Valley Inequigranular granite gneiss with mottled metagabbro may have produced the unusual chemistry of the former. That mottled metagabbro and Inequigranular granite gneiss could mix intimately is seen in outcrop and in the textures displayed in the mottled metagabbro. (See section 4.7 ). The chemistry of the tonalite sample TV305 may be explained in a similar way, since these rocks are



probably the products of crustal anatexis as is seen from their association with metagabbro - country rock contacts. It is unlikely that microcline megacrysts found in the Inequigranular granite gneiss were formed by blastesis (see Beukes 1973) since the original porphyritic texture of the rock is apparent where the rock is undeformed.

Winkler and Von Platen (1961) have shown that granites are formed mainly from melts produced by anatexis of gneisses. On a Q, Ab, Or diagram (Fig 89 ) the granitic rocks from the study area plot mainly along the border of the contour line representing 86% of all granites (represented by over a thousand samples) (Winkler and Von Platen 1961). To a first approximation this field represents anatectic granites so that it is possible that the granitic rocks from the Kumkum-Tantalite Valley area formed by this mechanism. This conclusion is partly supported by the work of Wyllie et al (1976) who state that a batholith, a term which can correctly be applied to the Inequigranular granite gneiss because of its large outcrop area, is produced from crustal rocks 'as a normal consequence of regional metamorphism, with the formation of  $H_2O$  - undersaturated granite liquid and mobilized migmatites'. Further, the geothermal gradient in

the area was high during metamorphism (see section 7.2.9) which would be conducive to the production of granitic rocks by anatexis. The most convincing evidence for anatexis is the presence of garnet, and more rarely cordierite and sillimanite, in the Inequigranular granite gneiss from the Kumkum area. The association of this rock-type with garnet-cordierite-gneisses throughout the area may be significant. Green (1976) showed that it is possible to produce S-type granites by partial melting of pelitic rocks to give garnet, cordierite residues at  $800 \pm 20^{\circ}\text{C}$ . Assuming that biotite is still stable, garnet-bearing granites will be produced at pressures greater than 5 kb and cordierite-bearing granites at pressures less than 5 kb. Cordierite in the Inequigranular granite gneiss occurs solely as rims around garnet which means that the rock was emplaced before M 2 (see section 5.4 ) and probably during M 1 when temperatures were of the order of  $800^{\circ}\text{C}$  and the pressure was greater than 5 kb. Pelitic gneiss also occurs associated with Inequigranular granite gneiss at Tantalite Valley but not to the same extent as in the Kumkum area. The Inequigranular granite gneiss is so strikingly similar throughout the entire area that it is unlikely that it had a different origin in the Tantalite Valley area. What does

seem likely is that the Inequigranular granite gneiss may have been <sup>partially</sup> remobilized during the emplacement of the mottled metagabbro since the granite gneiss forms an envelope around the mottled metagabbro and often contains xenoliths of this rock-type. This would explain why Inequigranular granite gneiss appears to be emplaced during M 2 in the Tantalite Valley area compared with during M 1 in the Kumkum area.

Aplites which appear to be genetically related to the Inequigranular granite gneiss in the Kumkum area, since they always intrude the latter, contain garnets which are spessartine-rich, and are interpreted as primary (Hsü 1968). Only almandine-rich garnets are found in the granite gneiss.

#### 8.8 Geochronology

Geochronological studies using U - Pb and Rb - Sr methods have shown that the Namaqua Metamorphic Complex is characterized by ages ranging from about 900 to about 1 300 Ma (Nicolaysen 1962; Nicolaysen and Burger 1965; Burger and Coertze 1973; Clifford et al 1975a). This distinguishes it from the Richtersveld Province in the west, where ages of around 1 900 Ma have been obtained (Reid 1977) and the Kheis tectonic province to the east, where ages of between 2 500 and 2 900 Ma have been

recorded (Vajner 1974).

Ages determined for rocks which can be correlated with those from the Kumkum - Tantalite Valley area are given in Table ( 50 ). It can be seen that the ages are characteristic of the Namaqua Metamorphic Complex. The age of the Tantalite Valley Complex and the Kumkum gabbro-norite can be obtained approximately since these units intrude the Inequigranular granite gneiss, dated at 1 218 Ma (using the  $Pb^{207}/Pb^{206}$  age for the Beenbreek Granite, Table 50, Toogood 1976) and are themselves intruded by discordant, young pegmatites dated at 988 Ma (using the age of the Lithium pegmatite of Tantalite Valley, Table 50, Nicolaysen 1962.) The Orangefall biotite gneiss (Grey Gneiss) which is representative of the pre-tectonic gneisses, gives ages of between 1 142 to 1 720 Ma (Table  $Pb^{207}/Pb^{206}$  ages,  $n = 2$ , Toogood 1976). Kröner (1974, 1975) had drawn attention to other localities in the Namaqua Metamorphic Complex where ages of between 1 600 and 1 800 Ma have been recorded. These anomalous ages may reflect an imprint of the Richtersveld province igneous activity, tectonism and low-grade metamorphism on some of the rocks of the Namaqua Metamorphic Complex. The predominantly 1 000 Ma ages may mark the onset of closed system

behaviour associated with late tectonic uplift (Kröner and Blignault 1976), or the main high grade metamorphic event (M 1). However, Blignault (1977) believes that the main metamorphism of rocks of the Namaqua Metamorphic Complex commenced at least at about 1 900 Ma since he correlates rocks of the Vioolsdrif igneous complex (1 900 - 2 000 Ma) with the Grey Gneiss sequence found in the Namaqua Metamorphic Complex.

## CHAPTER IX

## Summary and Conclusions

"It is an old maxim of mine that when you have excluded the impossible, whatever remains, however improbable must be the truth".

The Beryl Coronet - Sir Arthur Conan-Doyle.

An original metasedimentary sequence has undergone polyphase metamorphism and deformation and has been intruded by a large variety of mafic, ultramafic and granitic rocks. A prominent right-lateral shear zone, the Tantalite Valley Lineament, has produced mylonites along its axis, and since all the ultramafic rocks occur along or close to this line, it is probable that it facilitated the tapping of basic to ultrabasic magmas from the upper mantle. The shear-zone axis also defines the "muscovite out" isograd in the area, with high-grade metamorphic rocks occurring to the north and medium-grade metamorphic rocks being found to the south of the shear zone. This distribution is probably due to greater erosion of the northern block compared with the southern block as a result of greater vertical displacement of the former and is not a primary feature. The Inequigranular granite gneiss formed

as a result of partial melting of pelitic rocks leaving garnet-cordierite-sillimanite restites (the Black Gneiss). Both the Black Gneiss and Inequigranular granite gneiss are restricted to areas north of, or along, the axis of the Tantalite Valley Lineament. Migmatites are found over the entire area but are best-developed in the domain in which the Black Gneiss and Inequigranular granite gneiss are found. The Black Gneiss parageneses formed during M 1, which was characterized by pressures of 5 to 6 kb and  $P_{\text{total}} > P_{\text{H}_2\text{O}}$  and temperatures between 750 and 860°C. During this event, the Inequigranular granite gneiss and the Kumkum and Eselruh gabbro-norites were emplaced, in that order. Partial remobilization of the Inequigranular granite gneiss occurred where it was intruded by the gabbro-norites giving rise to a hybrid rock-type, the Contaminated gabbro-norite. These intrusions cooled to the ambient high-grade metamorphic conditions. Partial melting, with the production of migmatites and concordant pegmatites was common during this event. The Tantalite Valley Shear Zone was probably initiated during the retrograde metamorphic event, M 2, as a result of high strains becoming localized within the approximately planar region of the Tantalite Valley Lineament (Moore 1975). M 2 was characterized by pressures

from 4 to 5.5 kb,  $P_{\text{total}} > P_{\text{H}_2\text{O}}$  and temperatures from 580°C along the axis of the Lineament to 760°C north of it. For the region south of the axis of the Lineament, temperatures were probably between 550 and 650°C.

Pressure was constant over the entire area. The presence of cordierite rims on garnet and hydration reactions such as uraltization in such rocks as granulites are ascribed to this event. Where water was available, partial melting could give rise to migmatites and conformable pegmatites.

In the Tantalite Valley area, the proto-mottled metagabbro was intruded along the axis of the Lineament and together with the enclosing gneisses developed a penetrative schistosity. Its emplacement may have led to <sup>partial</sup> remobilization of the Inequigranular granite gneiss which occurs as an envelope around the mottled metagabbro. The proto-mottled metagabbro was then converted into mottled metagabbro as a result of metasomatism by granitic liquids. The proto-metagabbro, probably a gabbro-norite, was intruded late-tectonically into comparatively dry rocks. This gave rise to a narrow contact aureole and local partial melting resulting in the emplacement of tonalite and granodiorite along metagabbro-country rock contacts. Cordierite-anthophyllite and



garnet-cordierite-sillimanite hornfels were formed as restites. Temperatures in the aureole were in the range from 575 to 820°C and pressure was constant at about 5,5 kb with  $P_{\text{total}} > P_{\text{H}_2\text{O}}$ . Pyroxene geothermometry on the gabbro-norite indicates intrusion temperatures near 1 000°C. Therefore in general the emplacement temperature of the metagabbro is unlikely to have been much less. The proto-mottled metagabbro and proto-metagabbro then underwent extensive hydration metamorphism, during which they were converted completely to metagabbros. The gabbro-norite was then intruded into the metagabbro at temperatures of approximately 1 200°C as plugs, sheets and dykes and cooled slowly giving rise to olivine-plagioclase coronas, clouding in plagioclase and opaque mineral-orthopyroxene symplectites. Following this mafic-ultramafic rocks were intruded as plugs into the Complex and close to it in the enclosing gneisses at about 1 200°C and like the gabbro-norite underwent slow cooling. Retrograde mineral reactions in the hornfels, eg staurolite breakdown, indicate that a drop in pressure occurred, probably as a result of rapid uplift. During this stage the shear zone was reactivated which resulted in the production of a weak schistosity in the hornfels and the outer margin of the Tantalite Valley Complex. Local partial melting of the gneisses

led to the emplacement of large discordant, mineralized pegmatites into low-pressure areas around the Complex (Moore 1975). Shearing accompanied by introduction of water in the mafic rocks produced low-grade metamorphic assemblages on a small scale and was responsible for the development of such minerals as prehnite and chlorite in pelitic hornfels.

Generally metamorphism in the area is typical of the low-pressure facies series which is characterized by high geothermal gradients.

Although the Tantalite Valley Complex has many features in common with the Zoned Complexes of Alaska (see Wyllie 1967), the most striking of which is multiple intrusion, with the ultramafic rocks following the gabbroic rocks in order of emplacement, most of the characteristics point to it being a variety of Layered Complex. Partial melting of peridotite in the upper mantle gave rise to basaltic liquids which moved upward into a magma chamber situated at a depth of between 20 and 30 km. Here differentiation occurred. Ultramafic cumulates sank to the floor of the chamber leaving iron-enriched basic liquids in which phenocrysts of plagioclase and olivine were suspended. Successive tapping of this magma chamber then gave rise to the emplacement of the intrusives making up the Complex

in the sequence observed.

The mafic and ultramafic rocks from the Kumkum - Tantalite Valley area have a tholeiitic affinity. Comparisons of major and trace element concentrations in mafic rocks from the Tantalite Valley and Kumkum areas rules out the possibility that the differences in chemistry of these suites of rocks can be explained by crystal fractionation but should rather be ascribed to lateral inhomogeneity in the upper mantle. Concentrations of incompatible elements in the mafic rocks indicates that the source area was depleted. This is possibly the result of some earlier melting episode having taken place. The gabbroic rocks have compositions similar to high  $Al_2O_3$  - olivine - tholeiites which are found along the mid-oceanic ridges. The association of these rocks with zones of tension can explain the presence of similar rocks in the study area which is dominated by a major fault, the Tantalite Valley Lineament.

The hypothesis that the mafic and ultramafic rocks are members of the charnockite suite as proposed by Toogood (1976) and Jackson (1976) is rejected because of the high intrusive temperatures and atypical mineralogy and texture of these rocks compared with charnockite from the type-area in India.

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