THE GEOLOGY OF THE AREA EAST OF POFADDER WITH EMPHASIS ON SHEARING ASSOCIATED WITH THE POFADDER LINEAMENT, NORTH WEST CAPE

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ABSTRACT

To the east of Pofadder in Bushmanland, N.W. Cape, an area of some 3600 km^2 has been geologically mapped. Emphasis has been placed on the role of shearing in the deformation history of the area.

The study area occurs within the Namaqualand Metamorphic Complex and consists of gneisses, schists and quartzites. These rocks are overlain to the south by the Dwyka Formation of the Karoo Sequence. The rocks of the Namaqualand Metamorphic Complex have undergone polyphase deformation and have been mapped according to their fabric and composition. The Bushmanland Group consists of supracrustal rocks and the following lithologies have been mapped: quartzite (granular and glassy varieties), muscovite-quartz schist, calc-silicate gneiss, quartzo-feldspathic rocks (quartzo-feldspathic gneiss, leucogneiss and leptite), para-amphibolite gneiss and pelitic gneiss. Pre-tectonic intrusive rocks consist of the Nouzees Gabbronorite Suite, a mafic granulite, amphibolite and alkali-feldspar granitic gneisses. Syn to post-tectonic intrusive rocks include a granodiorite-tonalite suite, diorite, pegmatite and vein quartz. Dolerite is post-tectonic, most likely of Karoo age and occurs as sills and remnant hillocks. Non-diamondiferous kimberlite pipes occur in the west and their emplacement appears to have been structurally controlled, being situated along the Nouzees shear zone.

The study of metamorphism has been considered in terms of metamorphic petrology, from which a petrogenetic grid has been constructed, and also in terms of geothermometry and geobarometry. The geothermometry was carried out using Fe^{2+}/Mg partitioning for various mineral pairs, and by using calcium partitioning between ortho and clinopyroxene. The geothermometers give results of approximately 580 to 770°C, which agrees reasonably well with those from the petrogenetic grid of 625 to 675°C. These geothermometers also indicate that the exsolution in pyroxene record higher temperatures (830 - 990°C), which could represent the crystallisation of the mafic rocks. A geobarometer using Fe^{2+}/Mg partitioning between garnet and cordierite indicates that pressures of approximately 5,25 kb existed, which agrees reasonably well with that of 4,5 kb determined from the petrogenetic grid.

In the structural analysis, the structures and fabrics have been classified into five stages of deformation. These stages are based on age relationships to the regional foliation (S2), east-northeast - trending F3b folds (which fold the S2 foliation), mylonitic foliation (S4) and lastly (brittle) fractures and faults (S5). These stages of deformation describe the sequence of deformation and an important aspect of this study has been to group these stages into three phases of deformation, each phase having a different controlling process. The first phase, namely the "fabric-development phase" involves two stages of folding (F1 and F2), the folds being tight to isoclinal and defined by the oldest recognisable surface, namely bedding or its transposed equivalent S0/1. P-T conditions were at, or just prior to, those of peak metamorphism during the F2 folding, and it is during this stage (D2) that the regional foliation (S2) and a conspicuous mineral lineation (L2) developed, the foliation being axial-planar to the F2 folds. The second phase of deformation, namely the "fabric-folding phase" involves two generations of

folding (F3a and F3b). These folds are of the flexural slip type as a result of their folding of the S2 surface which facilitated the slip. The F3b folds are common, and plunge gently to the east-northeast with their axes being co-linear to the L2 lineation as a result of a strongly developed L2 anisotropism in the rock. Little is known about the older F3a folds as they were not identified in the field, but they are structurally necessary to yield the dome-and-basin fold interference pattern observed in the area, and have been reported to exist east of the present study area (R.W.Harris, pers.comm., 1984). The trend of these F3a folds appears to be northerly to northwesterly and it is thought that the folds are associated with west-southwest-directed thrusting. The thrusts are mylonitic and have been folded by the F3b folds. The final phase of deformation, namely the "shearing and wrenching phase" followed on from the previous phase in which shearing initiated along the northern limbs of the F3b synforms. In this phase, however, shearing intensified and formed continuous shear zones which controlled the mode of deformation. The shearing occurred in two directions, with steeply dipping shear planes and a predominant component of strike-slip movement. The shears form an east-southeast set of dextral shears, typified by the Pofadder Lineament, and an east-northeast set of sinistral shears, as typified by the Putsberg shear. The two sets of shears intersect at approximately 30° and there is evidence that they form a conjugate shear pair which probably developed in response to a north-south direction of maximum stress. Shearing is most intense in the dextral set of shears, and most notably within the Pofadder Lineament. There is evidence that the lineament extends from Lüderitz to Vanwyksvlei in the southeast, a distance of approximately 800 km. The displacement across the shear appears to be 50 km which is significantly less than the displacement of 85 km that was determined by Toogood (1976). An important conclusion regarding this phase of deformation is that the movement associated with the shearing continued over an extended period of time, and as conditions became brittle, the zone of movement became a wrench zone. The wrench zone follows the mylonite belt of the lineament, and Riedel fractures, which are marked by prominent veins of quartz, transect the mylonitic foliation. The wrench action has effected the Karoo Sequence rocks, and seismic evidence indicates that movement is still occurring along the lineament.

Finally, mylonitic fabrics have been studied under the microscope. The classification of mylonitic fabrics and also the factors that control their development are discussed.

CONTENTS

I. INTRODUCTION..1
A. Regional Setting..1
B. Previous Work around Pofadder...................................3
C. Present Investigation...4
D. Acknowledgements..5

II. LITHOLOGY..6
A. Introduction..6
B. Bushmanland Group...6
1. Pelitic Gneiss..6
2. Para-amphibolite Gneiss...7
3. Quartzo-feldspathic Rocks.......................................8
4. Calc-silicate Gneiss...12
5. Biotite Schist...14
6. Muscovite-quartz Schist..15
7. Quartzite..16
C. Pre-tectonic Intrusive Rocks...................................18
1. Mafic Granulite..18
2. Olivine Gabbronorite...20
3. Amphibolite..22
4. Granitic Gneiss..24
D. Syn to Post-tectonic Intrusive Rocks...........................26
1. Late Granitoids and Diorites...................................26
2. Pegmatite and Vein Quartz......................................29
3. Dolerite...29
4. Kimberlite...30
E. Cover Rocks and Superficial Deposits...........................31
1. Karoo Sequence...31
2. Calcrete...31
3. Aeolian Sand...31
4. Alluvium and Sand Pans...32

III. METAMORPHISM...33
A. Introduction...33
B. Diagnostic Mineral Parageneses.................................33
1. Low-grade Rocks..33
2. Medium-grade Rocks...34
3. High-grade Rocks...35
C. Mineral Parageneses and Reactions..............................35
1. Pelitic Rocks..35
2. Calc-silicate Rocks..37
3. Felsic Rocks...38
4. Mafic Rocks..38
D. The Petrogenetic Grid..41

E. Geothermometry and Geobarometry....................43
1. Introduction....................43
2. Iron-Magnesium Distribution Coefficients....................44
3. Results....................44
F. Conclusions....................46

IV. STRUCTURE....................49
A. Introduction....................49
B. Theory on Shears and Their Brittle-state Analogues....................51
1. Wrench Zones....................51
2. Brittle to Ductile-state Transition....................53
3. Shear Zones....................54
C. Orientation of Fold Axes....................55
D. Regional Structure....................57
1. Description of Structures....................57
2. The Pofadder Lineament....................60
3. Structural Areas....................62
E. The Deformation Mode....................65
1. Introduction....................65
2. Classification of Structures....................65
3. Deformation Processes....................66
F. Ductile Structure Analysis....................67
1. Introduction....................67
2. Gordonia Sub-province....................70
(a) Introduction....................70
(b) Lucas Vlei Vlakte Region....................70
(c) Longziek Vlei Synform Region....................75
(d) Longziek Vlei Interference Region....................76
(e) Bank Vlei Region....................77
(f) T'Oubep Region....................77
(g) Conclusions....................78
3. The Pofadder Lineament....................79
(a) Introduction....................79
(b) Listric Shears....................80
(c) Fabric within the Pofadder Lineament....................80
(d) Conjugate Shearing....................84
(e) Discussion....................84
4. Bushmanland Sub-province....................86
(a) Nouzees Region....................87
(b) Houmoed Region....................87
(c) Putsberg Shear....................87
(d) Vraa-Weer Region....................89
(e) Conclusions....................91
G. Fracture Analysis....................91
H. Structural Synthesis....................94
1. Fabric-development Phase (D1 and D2)....................95
2. Fabric-folding Phase (D3a and D3b)....................96
3. Shearing and Wrenching Phase (D4 and D5)....................96

V. MYLONITE TEXTURES..99
A. Introduction..99
B. Classification of Mylonite Rock..99
C. Factors Controlling the Development of Mylonitic Fabrics.....101

VI. CONCLUSIONS..106

VII. REFERENCES..110

APPENDICES

Appendix I: Microprobe analyses of clino and orthopyroxene mineral pairs in the Nouzees Gabbronorite Suite (specimen HM 7) and a mafic granulite (specimen HM 261).................118
Appendix II: Microprobe analyses of coexisting cordierite, garnet and biotite in pelitic gneiss for specimen HM 25........119
Appendix III: Microprobe analysis of garnet in calc-silicate gneiss for specimen HM 196....................................120

PLATES................................121

GEOLOGICAL MAP
(Annexure)

Geological Map of the area east of Pofadder

PREFACE

Cataclastic Rocks

These are rocks in which there has been a brittle fragmentation of mineral grains with rotation of grain fragments accompanied by frictional grain boundary sliding and dilatancy (Sibson,1977).

Crystallisation and Recrystallisation

Crystallisation refers to the formation of a new mineral that is different from any other pre-existing minerals. Recrystallisation is the crystallisation of existing mineral phases, forming more crystals of that phase.

Deformation Stage and Phase

The term "deformation stage" refers specifically to the sequence in which structures and fabrics developed. In contrast, the term "deformation phase" refers to one or more consecutive stages of deformation which are controlled by a common process of deformation. Whilst the former term is useful for the purpose of description, the latter is of greater tectonic significance.

Fold Tightness

The tightness of a fold is defined in terms of the following interlimb angles (Ramsay, 1967, p.349):

gentle to open	180° - 70°
close	70° - 30°
tight to isoclinal	30° - 0°

Granulite

The term granulite is restricted to metamorphic rocks with mineral assemblages that are diagnostic of the regional hypersthene zone. Since the concept of metamorphic "grade" is used in this thesis and not metamorphic "facies", the term "granulite high grade" is used to define the anhydrous field of high grade metamorphism. This is the same as Winkler's (1967) "granolite high grade".

Leptite

Fine-grained rocks consisting chiefly of quartz and feldspar (Spry, 1969, p.294) with a granular texture.

Linear Orientation

Linear features are recorded in terms of their trend followed by their plunge, the two measurements being separated by a slash. For example, a lineation that trends due west and which has a plunge of 50° is recorded as 270/50.

Lithologic Symbols

The various lithologies that have been mapped are symbolised by a capital letter which defines the erathem followed commonly by two letters. The first letter designates the compositional nature of the lithology, and in some cases this letter can be self-sufficient as for example in quartzite. The second letter after the erathem designates the fabric of the rock. In most cases the two letters after the erathem adequately describe the lithology, but in rare cases, a third is required to designate any unusual feature, for example epidotization. The symbols used are listed in Table 1.

Locality Reference

Locations on the geological map are specified by reference to co-ordinates of latitude followed by longitude, separated by a slash. The co-ordinates are given in degrees, minutes and seconds, the last mentioned being approximated to quarters of a minute unless greater accuracy is required. For example, a reference such as the following could be given (29°15'45"S/19°40'00"E).

Mylonitic Rocks

Mylonites are rocks in which the minerals that are least resistant to deformation, namely the soft minerals, "have undergone intense intracrystalline plastic deformation" as a result of simple shear, causing the rock to be "foliated, usually with a strong inosculating L-S shape fabric (fluxion structure)" (Sibson,1977) with a cohesive matrix. Blastomylonites are rocks in which grain growth during syntectonic recrystallisation predominated resulting in a strain-free texture (Sibson, op.cit., Table 1).

Planar Attitudes

Of the two strike directions that are possible for any measurement of a planar attitude, only the direction for which the dip is to the right is recorded. The strike direction is followed by the dip measurement, separated by a slash. For example, a surface having an east-west strike and which dips at 50° to the north is recorded as 270/50.

Porphyroclast

A relatively large fragment of a crystal, mineral grain, or aggregate of crystals or grains in a "mylonitic or cataclastic rock. Porphyroclasts are not produced by neomineralisation" (crystallisation) "or recrystallisation (as opposed to porphyroclasts)" (Higgins,1971), but undergo grain-size reduction, usually due to brittle cataclasis, though sometimes by recovery.

Table 1: Lithologic symbols used for the Geological Map

Erathem	Compositional Features	Fabric Features	Additional Features
Q Quarternary	a alkali-felds-	a aggregates or	e epidotization
K Cretaceous	par granite	clusters	m marble
J Jurassic	b biotite-rich	b streaky bands	o ortho-
C Carboniferous	c calc-silicate	c conglomerate	amphibolite
N Namibian	d dolerite	f float	p clino-
M Mogolian	Dwyka Fm.	g granular	pyroxenite
	diorite	gravel	
	f feldspathic	l layering	
	g granodiorite	(gneissose)	
	h hornblende-	m megacrystic	
	rich	p porphyritic	
	i ironstone	pegmatitic	
	k kimberlite	s schist	
	k-feldspar-	shale	
	rich	t tillite	
	l leptite	v vein-like	
	m muscovite-		
	rich		
	n gabbronorite		
	o orthopyroxene		
	-bearing		
	p pelitic		
	q quartzite		

Example: Mcl = calc silicate gneiss of Mogolian age

Table 2: Mineral abbreviations used

Ac	actinolite-tremolite	Di	diopside	Or	orthoclase
Alm	almandine	Ep	epidote	Pi	piemontite
Am	amphibole	Fs	ferrosalite	Pg	plagioclase
And	andradite	Ga	garnet	Py	pyrite
Aug	augite	Gd	gedrite	Qz	quartz
Ap	apatite	Hb	hornblende	Sc	scapolite
At	anthophyllite	Hy	hypersthene	Sil	sillimanite
Bi	biotite	Kf	K-feldspar	Sp	sphene
Cc	calcite	Ma	magnetite	St	staurolite
Cd	cordierite	Mc	microcline	Tr	tremolite
Ch	chlorite	Mu	muscovite	Zc	zircon
Cpx	clinopyroxene	Ol	olivine	Zo	zoisite
Cz	clinozoisite	Opx	orthopyroxene		

Chapter I

INTRODUCTION

A. Regional Setting

An area of Bushmanland in the North West Cape, occurring to the east of Pofadder and bounded by longitudes 19°30' and 20°15'E and latitudes 29°00' and 29°30'S has been geologically mapped. Topographic maps covering this area are numbered 2919 BA,BB,BC, BD; 2920 AA and AC. The study area, as shown in Fig.1, covers some 3600 km^2 and is well served by gravel roads from Kenhardt and Kliprand besides a tar road which connects Pofadder to Springbok and Kakamas.

The area mapped occurs within the Namaqualand Metamorphic Complex and towards the south these rocks are overlain by rocks of the Karoo Sequence. The morphology over the entire area is extremely flat and has been described by Paizes (1975) as conforming to a post-Karoo erosional surface, with gravel or float of the Dwyka Formation covering extensive areas in the south. Outcrop of the Namaqualand Metamorphic Complex is therefore obscured to a large extent in the south and this is made worse by the presence of a well-developed cover of calcrete. The calcrete is most common at the interface between the gneisses and the Dwyka Formation, as well as in topographic depressions such as drainage courses. Sand is widespread throughout and in the northwest permanent sand dunes have developed, typified by the red, well-rounded, Kalahari-type sand. Rocks of the Namaqualand Metamorphic Complex that outcrop most frequently are those that resist weathering the most, such as quartzite and calc-silicate gneiss. Fortunately, strike lines seen on aerial photographs have aided structural interpretation of the area considerably. Rock outcrop is greatest in the northwest of the area along the Mattheusgat Mountains which mark a well exposed part of the Pofadder Lineament. The lineament which is thought to extend from Lüderitz (S.W.A./Namibia, passes through the present study area but cannot be followed further southeast due to cover of the Dwyka Formation. The lineament is considered, however, to extend as far to the southeast as Vanwyksvlei since the overlying Dwyka Formation appears to have been effected (Joubert,1974c). If this extension is correct, then the lineament has a length of approximately 800 km.

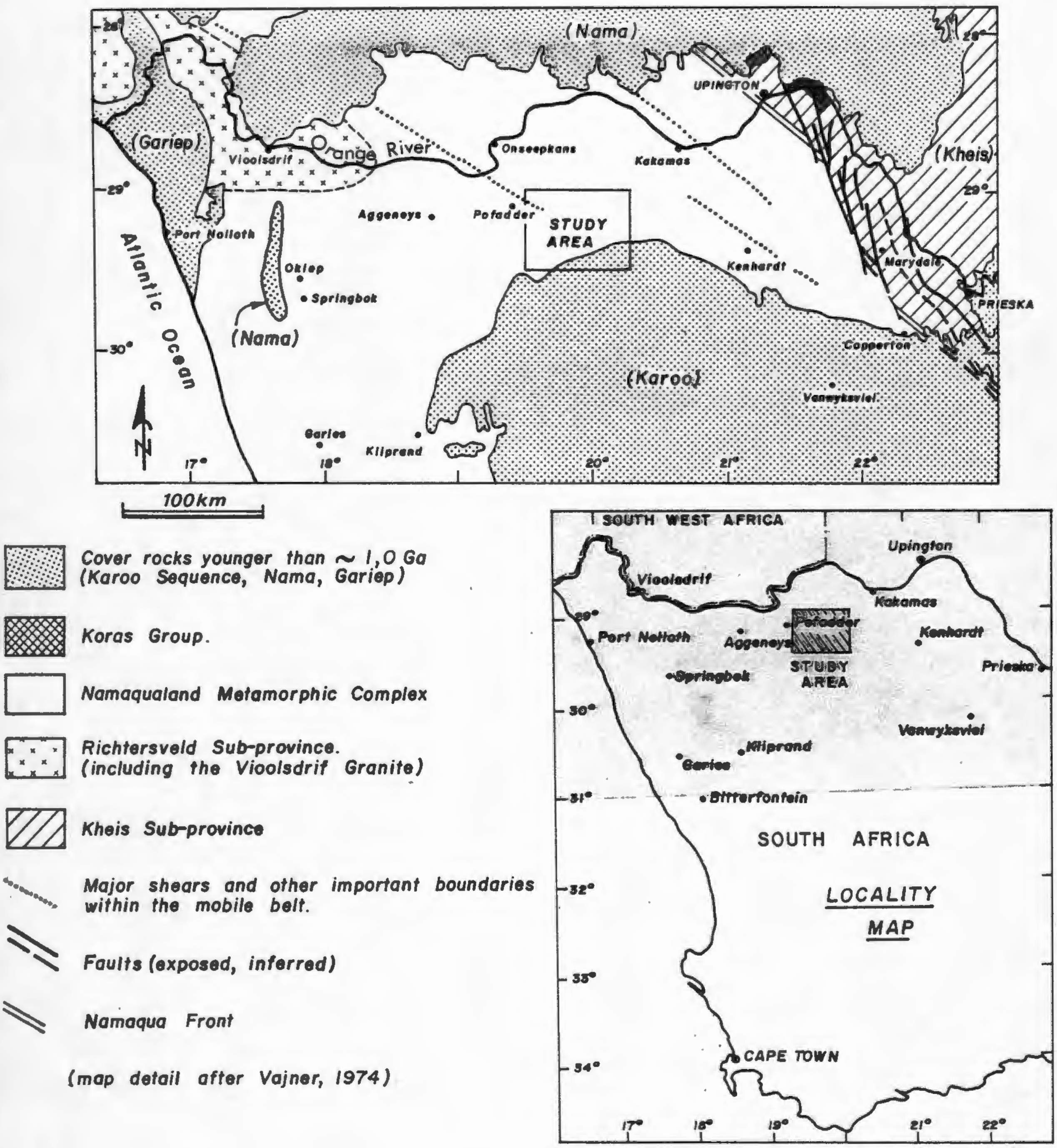

Fig.1: The location of the study area showing the regional geology

B. Previous Work around Pofadder

Early in this century, isolated mineral deposits were discovered in Bushmanland, for which descriptions appeared in the literature. Pegmatite is very common in Bushmanland and Gevers (1936) has described mineralisation in these rocks. Much of the pioneer studies were undertaken by Coetzee (1941,1958) who has described sillimanite-cordierite rocks, especially in the Pella vicinity, and manganiferous iron ore, haematite, barite, and sillimanite on Gams. De Jager and Von Backström (1961) have described sillimanite deposits near Pofadder and De Jager (1963) gives a review of sillimanite reserves. Within the present study area, prospecting for sillimanite, which is hosted in biotite schist, was carried out by Mr R.G. Niemöller on the farm Steenkampsvlei. Mr Niemöller has also prospected for amazonite and gadolinite in pegmatite on the farm Nouzees. Black tourmaline (shorl), garnet, and rare occurrences of purple tourmaline (possibly rubellite) and green tourmaline (possibly verdelite) can also be found in the prospect pits. An isolated occurrence of amazonite has also been prospected by Mr Niemöller on the farm Adjoining Geelvloer where pegmatite intrudes calc-silicate gneiss. Exploration work has been carried out on pegmatites to the north of the study area in the vicinity of the farm Coboop by Esso Minerals of S.A. Limited in the late 1970's, but the exploration was terminated after a few years. Several kimberlite pipes have been prospected by Mr. Niemöller to the southeast of Pofadder, of which several pipes occur on the farm Nouzees in the present study area. The pipes are not diamondiferous and have been described by Frick (1974) as being of the basaltic or autolithic type. These pipes appear to be associated with deep-seated structures (ibid.) and mapping shows them to occur in close proximity, and also within, the Nouzees shear.

Bushmanland has been the centre of attraction for base-metal exploration in recent years and during the period from 1971 to 1973 stratiform lead-zinc deposits with minor amounts of copper and silver were discovered at three localities: Two of these localities, namely on the farm Zuurwater near Aggeneys and at Gams, occur approximately 60 km west of Pofadder while the third, which occurs on the farm Rozynen Bosch occurs approximately 40 km northwest of Kenhardt. From a geographical aspect the area between Aggeneys and Kenhardt is a likely target area for further base metal exploration and several companies have carried out exploration in the area. A copper deposit was found by Johannesburg Consolidated Investment Company (J.C.I.) in 1974 during follow-up ground work of aeromagnetic anomalies, but the deposit is considered to be of sub-economic value by J.C.I. in present circumstances (Campbell and Mason, 1979). The copper mineralisation is intimately associated with thin siliceous horizons within biotite-sillimanite schist, and consists of chalcopyrite, occasional sphalerite and rare galena (ibid.). Present mapping has shown that the deposit is located within the Putsberg shear zone and therefore the mineralisation most likely originated as a result of mobilised fluids within the shear, especially if movement along the shear continued during the brittle stages of deformation. Exploration pegs can still be seen in the field and they follow the shear. Gold Fields of S.A. Ltd. (G.F.S.A.) and Gencor are at present both involved in base-metal exploration in the area, and the latter company appears to have found a mineralised body on the farm Adjoining Geelvloer.

Base-metal exploration in the last decade has revealed that the geology is very complex, typified by polyphase deformation and as a result research units at various South African Universities have shown great interest, and their research has been largely subsidised by interested mining houses. Joubert pioneered much of the research work in Namaqualand (1971a,b) and Bushmanland (1973; 1974a,b,c; 1975). Other studies in Namaqualand, Bushmanland and southern S.W.A./Namibia that have concentrated on the Namaqualand Metamorphic Complex have been prolific since Joubert's work, and in the vicinity of the present study area, the following work has been done: Beukes (1973), Blignault (1974), Blignault et al.(1974), Toogood (1974,1976), Paizes (1975), Kröner and Blignault (1976), Moore (1976,1981), Moore (1977,1980,1983), Lipson (1978,1979), Colliston (1979), Reid et al. (1979), Maclaren (1983), Odling (1983) and Albat (1983a,b). As a result of this and ongoing research, a reasonably co-ordinated picture of the geology is developing which should go hand-in-hand with future mineral exploration.

C. Present Investigation

The main aim of this study has been to describe the role of shearing in the deformation of the area studied, and if possible to determine whether the Pofadder Lineament terminates within the area studied or if it continues further southeast, possibly as far as Vanwyksvlei. During field work this required distinction between sheared and unsheared rock, which in some places is very subtle. Mylonitic textures were studied in thin section and metamorphic P-T conditions have also been determined.

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Chapter II

LITHOLOGY

A. Introduction

As the rocks of the Namaqualand Metamorphic Complex have undergone several phases of deformation and tectonic duplication is almost certainly responsible for the repetition of similar lithologies and lithological assemblages in the area, a stratigraphic approach in the description of the various lithologies would lead to confusion. In the descriptions of the rocks of the area, therefore, classification is according to the fabric and the names are prefixed with the most common mineral(s) in the assemblage.

Rocks that are intrusive have been classified according to their composition and also according to their fabric, when it is developed. The I.U.G.S. classification scheme for plutonic rocks (Streckeisen, 1973) has been used for their compositional classification.

The Dwyka Formation of the Karoo Sequence occurs as extensive sheets of scree in the south. Distinction has been made as to whether or not it occurs in situ, as a coarse float or as a fine gravel. Shale beds in the Dwyka Formation have also been distinguished on the map.

B. Bushmanland Group

The Bushmanland Group includes the pre-tectonic rocks for which there is no evidence of an intrusive origin. Most of these rocks are without doubt of sedimentary origin, but the origin of some of the quartzo-feldspathic gneisses is debatable. These rocks all have a metamorphic imprint and they possess a reasonably well developed fabric.

1. Pelitic Gneiss (Mpl)

Outcrops of the pelitic gneisses were found in proximity to the large ortho-amphibolite body on the farms Nouzees and Vaalputs where they are interbanded with quartzites, and as pods of pelitic gneiss within para-amphibolite gneiss on the farm Steenkampsvlei. In Table 3 the locations for samples of pelitic gneisses and their estimated mineral modes are listed.

The rocks are fine to medium grained, a dark bluish colour and are strongly foliated. In places sillimanite needles impart a mineral lineation. In thin section the rock has an inequigranular seriate texture with quartz, labradorite (An 55 by Michel-Lévy method), cordierite and microcline defining a granoblastic interlobate texture. Lepidoblastic biotite and sillimanite, the latter occurring commonly as mats of fibrolite having a preferred alignment and concentrated in bands, define a foliation and a weak lineation. Almandine garnet occurs commonly as sieve-textured porphyroblasts.

Table 3: Estimated mineral modal percentages for pelitic gneisses

Locality: HM 25 29°16'10"S/ 19°34'10"E
HM 31 29°14'25"S/ 19°30'00"E
HM 33 29°15'30"S/ 19°34'45"E
HM108 29°22'55"S/ 19°32'25"E

	Qz	Pg	Cd	Sil	Ga	Bi	Kf	Ch	Mu	Ma
HM 25	20	1	50	12	9	6				2
HM 31	45		41	7	2	1	2			2
HM 33	1	45	40	5				4	1	4
HM 108				94		1			4	1

2. Para-amphibolite Gneiss (Mhl)

Para-amphibolite gneiss differs from ortho-amphibolite as it is commonly strongly banded and the mineral bands, which possibly represent original layering in the paragneiss, are boudinaged in many places. Recrystallised quartz defines much of the banding and also a weak mineral lineation. The rock is commonly associated with calc-silicate and quartzo-feldspathic gneisses, and biotite schist. On the farm Steenkamplvlei para-amphibolite gneiss contains pods of biotite schist which in turn hosts pelitic gneiss and pockets of sillimanite. Prospect pits for the sillimanite have been dug on this farm near the Pofadder-Kliprand road. In the vicinity of these pits, to the east of the road, an F3 fold closure is developed in the para-amphibolite gneiss and garnet porphyroblasts in the hinge zone (specimen HM 109) are well developed.

In thin section the rock is characterised by:

(a) A fine to medium-grained inequigranular, seriate and granoblastic texture with interlobate grain margins.
(b) Mineral concentrations of quartz, hornblende and in places diopside, which define laminae.
(c) A mineral lineation defined by elongate quartz grains which have commonly annealed into sub-grains forming linear aggregates.
(d) Idioblastic garnet or magnetite.

Mineral modes for the para-amphibolites are listed in Table 4. Of particular interest is specimen HM 179, collected from the Adjoining Geelvloer shear (29°20'30"S/ 20°08'30"E) and which occurs as pods within biotite-rich quartzo-feldspathic gneiss. The rock is magnesium, and to a lesser extent, alumina-rich since it contains gedrite and cordierite but lacks sillimanite. Pyrite is finely disseminated throughout the rock. The rock is of unknown origin, but on the basis that it is comprised predominantly of plagioclase, amphibole and quartz it has been included with the para-amphibolite gneiss. The rock is metasomatic, however, as the plagioclase is albite (having a refractive index less than quartz) and no K-feldspar is present. The rock therefore appears to be enriched in sodium and depleted in calcium, possibly as a result of shearing.

Table 4: Estimated mineral modal percentages for para-amphibolite gneisses

	Pg	Hb	Qz	Ga	Di	Cd	Sp	Ma	Zr	Bi	Ep	Cz	Ch	Ap	Kf
HM 39	20	15	61					2	1		1				
HM 53	25	10	40				1	4	5		14			1	
HM 83	20	20	56					4							
HM 109	47	35	7	4				7							
HM 129	65	18	10		1		3	3							
HM 130	60	20	9				1	1	1	1	1	2	1		3
HM 133	43	30	25				1							1	
HM 208	1	99													
HM 179+	26	38*	25			8				2		(Py=1)			

+ Mg-rich rock of uncertain origin * Gedrite

3.Quartzo-feldspathic Rocks

Quartzo-feldspathic Gneiss (Mfl)
Megacrystic variety (Mfm)
Leucogneiss (Mkl)
Megacrystic variety (Mkm)
Leptite (Mll)

Three varieties of quartzo-feldspathic rocks were identified in the field, namely quartzo-feldspathic gneiss, leucogneiss and leptite. The first variety is a medium to coarse-grained rock of pinkish colour on fresh surfaces, but which weathers to brown. The type area for this variety is on the northern slopes of the Mattheusgat Mountains and also on the north-eastern side of the Nouzees shear zone to the north of the Kakamas road. On the farm Gemsbok Vlakte (at locality 29°04'15"S/ 19°31'00"E) the rock is intruded by a granitic gneiss and has a very weak foliation, resembling a granitic gneiss itself. The leucogneiss variety is medium grained and a very pale whitish colour; the type area being on the farm Lucas Vlei Vlakte where it occurs as a paragneiss interbanded with lenses of para-amphibolite in places. Here the rock is intruded by a red granitic gneiss in the extreme north of the study area and the rock is also brecciated, silicified, strongly epidotised and is intruded by vein quartz in places along a postulated low angle, probably northward-dipping fault. The rock is sheared in numerous areas and although it has the same mineralogy, it commonly weathers to a greyish colour. The rock is megacrystic along the northern slopes of the Mattheusgat Mountains, but further east along strike it weathers to a blackish colour and contains bands of rounded quartz and feldspar pebbles of up to two cm in diameter. These are clearly pebble bands defining bedding and the pebbles within them are aligned at an angle to the bands defining a transecting foliation. The leptite variety is fine grained, pink coloured, has a moderately developed foliation and is characterised by the presence of "quartz-eyes". The type area for the leptite is on the farm Adjoining Geelvloer, where it occurs adjacent to a major shear zone and is associated with para-amphibolite and calc-silicate gneisses.

The estimated mineral modes for the quartzo-feldspathic rocks are shown in Table 5. The quartzo-feldspathic gneiss variety is medium to coarse grained and in thin section has an inequigranular, seriate and granoblastic texture with interlobate grain boundaries. In places the rock appears to change from a seriate-grained rock to a porphyroclastic rock as a result of shearing (HM 87) whereby the quartz grains undergo grain-size reduction more rapidly than the feldspars. The K-feldspars in these rocks consist of microcline and commonly perthite and orthoclase as well. Plagioclase is estimated to be An 32 (Michel-Lēvy method) in HM 92. This specimen is from the farm Gemsbok Vlakte (locality 29°04'15"S/ 19°31'00"E), where the gneiss has only a very weak foliation and has a coarsely crystalline texture. A whole-rock analysis taken from the same outcrop for the Namaqualand Floor Rock Project, National Cooperative Scientific Programme, (M.K. Watkeys, pers.comm., 1983), sample NF 38, has a CIPW-Norm as follows:

Quartz	: Albite	: Anorthite
39.60	29.52	30.88

This plots on the cotectic line of the quartz-albite-orthoclase diagram (Winkler, 1976, Fig.18-2) very near the the minimum melt composition (ibid.). The rock at this outcrop is therefore possibly of igneous origin and is more red in colour than most other quartzo-feldspathic gneisses with the exception of the rocks in the same vicinity, south of the Pofadder-Kakamas road. On this basis, the latter rocks may also be of igneous origin.

The leucogneiss consists predominantly of quartz, K-feldspar and plagioclase in varying proportions with lesser amounts of opaque mineral, and minor biotite. The plagioclase ranges in An content from at least 30 to 46 (Michel-Lévy method) and is therefore andesine. The K-feldspar is commonly perthite or antiperthite and less commonly microcline. The garnet appears to be almandine rich on the basis of its red to brownish colour. The epidotized variety (HM 263) has probably formed in response to brittle deformation and according to Deer <u>et al</u>. (1966), its formation is favoured by low temperature and shear stress. The epidote appears to have replaced feldspar and the excess calcium liberated by the reaction may have resulted in the carbonate (calcite) seen in thin section.

Table 5: Estimated mineral modal percentages for quartzo-feldspathic rocks

	Qz	Kf	Pg	Ga	Ma	Zr	Sp	Hb	Ep	Ap	Cc	Bi	Mu	Ch	
HM 92	40	47	10		1							1	1		
HM193	55		34		3							5		3	
HM 61	50	24	10		5		1			1		4	5		+
HM 62	45	4	42				3	5		1					QFGn +
HM 72	37	50	7		1				1	1		3			+
HM 87	30	40	20									10			+
HM 85	40	35	20										5		
HM101	15	68	15		1							1			
HM102	20	70	2		2			2				4			L e
HM120	35	5	54	2	1							3			u c o
HM124	30	45	25												
HM170	40	15	42		2							1			G n
HM188	20	50	20		2		1	5	1		1				e i
HM206	35		62						2				1		s s
HM236	40	1	55	4											
HM265	35	53	10		1	1									

(continued)

(Table 5 continued)

	Qz	Kf	Pg	Ga	Ma	Zr	Sp	Hb	Ep	Ap	Cc	Bi	Mu	Ch	
HM 11	25	44	20		3		1	4		1		2			Sheared Leuco Gneiss
HM 13	15	75	3		2	1				1		2		1	
HM 20	40	25	24							1		8		2	
HM 36	45	44	2		1			2	1	1		4			
HM 58	50	28	70						1			9	1	1	
HM 59	55	42							1					2	
HM 76	25	20	50		1							3		1	
HM119	38	2	60												
HM125	50	33	10						2			5			
HM175	48	20	15		3	2					1		7	4	
HM248	50	40	5		1							3		1	
HM263					5				82		3				*
HM182	35	48	10		2	1	1	2		1					Leptite
HM183	56		38		2		3				1				
HM191	50		34	6	2		2			1	(Di=5)				

+ Megacrystic rock
* Epidote-rich rock
QFGn Quartzo-feldspathic Gneiss

Texturally the leucogneiss is fine to medium grained, inequigranular, seriate and granoblastic with interlobate grains of feldspar and quartz, with only minor phyllosilicates which define the foliation. The feldspar, and less commonly the quartz, form mineral concentrations in places.

Where the leucogneiss is sheared, chlorite has formed in response to retrograde metamorphism, and the inequigranular grain size becomes bi-modal because with strain the quartz recrystallises more rapidly than the feldspar. The porphyroclasts commonly consist of microcline or perthite and less commonly orthoclase. The perthite has, in a number of cases, changed in lattice structure to microcline within the strained margins and this may reflect a continuous series of microcline obliquities from monoclinic to triclinic (Deer et al., 1966). These rocks have a foliation that interbraids between the porphyroclasts and which is defined by the concentration and alignment of chlorite and biotite within laminae of recrystallised quartz.

The leptite is very-fine to fine grained and in thin section shows an inequigranular, seriate, granoblastic texture with interlobate grain boundaries. The foliation is defined by mineral concentrations and by aggregates of quartz grains, the latter conforming at least in shape, to flattened detrital grains.

4. Calc-silicate Gneiss

Calc-silicate Gneiss (Mcl)
Calc-silicate Conglomerate (Mcc)

Calc-silicate gneisses are commonly associated with quartzo-feldspathic rocks, amphibolites and quartzites. They are thought to represent part of a paragneiss sequence and occur on the farms Vaalkop and Graafwater. These rocks are typified by mineral banding on a mm to cm scale and by the positive relief of quartz-rich bands on the weathered surfaces. The quartz bands have a vein-like appearance, but are sufficiently continuous and conformable to the foliation to be regarded as graded bedding, accentuated by metamorphic processes. In places cross-bedding is preserved, as shown in Plate 1. The rocks are pale green in colour, weathering to a grey colour. They vary in composition and appear to grade into either quartzo-feldspathic rocks or into a granular quartzite in places. Less commonly the rock contains lenses of marble or pockets rich in garnet, scapolite or diopside. In one locality (29°12'00"S/ 20°06'00"- 20°10'00"E) on the farm Brul Kolk, the rock is conglomeratic with rounded clasts of quartzite and overlies a granular quartzite.

The mineral modes for the calc-silicate gneisses are listed in Table 6, which shows that these rocks are composed predominantly of quartz and plagioclase, while sphene is a common accessory mineral. The other calcic mineral(s) are usually hornblende or an amphibole of the actinolite - tremolite series and andraditic garnet (see microprobe analysis for HM 196, Appendix III). Calcic minerals that are less common are clinopyroxene (diopside or rarely ferrosalite), scapolite and calcite. K-feldspar is a common mineral comprising microcline, perthite or orthoclase and rarely antiperthite (HM 73). Accessory minerals are apatite, zircon, ore and muscovite.

Texturally, the calc-silicate gneisses are fine to medium grained, inequigranular, seriate and granoblastic with polygonal grain boundaries in places. Their lamination results from mineral concentrations of epidote and garnet, actinolite-tremolite, diopside, plagioclase, opaque minerals or quartz, the last-mentioned being most common.

Table 6: Estimated mineral modal percentages for calc-silicate gneisses

	Qz	Pg	Hb	Ga	Kf	Di	Sc	Sp	Cc	Ap	Zr	Ma	Mu	Ch	Ep
HM 30	45	45	7					1		1	1				
HM 34	40	25	3			20		1	1	1	4	5			
HM 78	40	28	5		15			3				2		7	
HM151	20	30	8		35			4	2	1					
HM162	2		15		5	20	55	1	1						1
HM167	25	50	5		10			1		1	2			5	1
HM196	20	44	2	8++	5	1		2	3			5			10
HM203	35	40	7		15			2			1				
HM212	34	60	4					2							
HM216	50	20	18	5	1			1	1			2			2
HM247	60	15	2			15		3	1	1		3			
HM135		1	95*			4									
HM144	15	25	5*		44			1	2			3			5
HM153	60		39*					1							
HM168		15	71*			7						1			(Zo=6)
HM224	50	15	4*		2			1	1				1		26
HM 29		65		15		10			1						9x
HM 32		20		30	18	30			2						
HM 44	1	40		21		30**		5		1	2				
HM152				92					5			2	1		
HM156	9	52		8	15	10	4		2						
HM189	34	50		8		5		1	1					1	

(continued)

(Table 6 continued)

	Qz	Pg	Hb	Ga	Kf	Di	Sc	Sp	Cc	Ap	Zr	Ma	Mu	Ch	Ep
HM 73	10	76			2	9		1				2			
HM 79	35	30			11			1	2	1		2	8		10xx
HM 80	9	5			3	25+		2		1					55
HM155	6	84				5+		2				2			1
HM190	25	15			40	18		2							

Symbols:
- * Actinolite-tremolite
- ** Hedenburgite?
- x Zoisite associated with Epidote
- xx Piemontite associated with Epidote
- \+ Ferrosalite
- ++ Andraditic garnet (see Appendix III)

5. Biotite Schist (Mbs)

Biotite schist is not common and typically occurs as minor lenses within quartzo-feldspathic rocks, as on the farm Pofadder East or as pods within para-amphibolite rocks, as on the farm Steenkampsvlei. Biotite schist also crops out as lenses in quartzite within the Mattheusgat Mountains, where it is migmatitic having quartzo-feldspathic veins and biotite selvedges. In places minor neosomes appear to intrude the sheared quartzite in the form of stringers parallel to the foliation. The mineral modes for the biotite schist examined in thin section are given in Table 7.

Table 7: Estimated mineral modes for biotite schist

	Qz	Bi	Mu	Pg	Ma	Sil	Ch	
HM 65	40	45	10	4	1	*		
HM 106	76	7	1	6	4	6		
HM 86	21	20	20	35	4			Migmatites
HM 88	59	20	20		1			Migmatites

* Sillimanite noted in field but not in thin section

The biotite schist is medium grained and has a strongly developed foliation in most areas. Where sillimanite is present, the rocks are commonly lineated due to the acicular habit of this mineral. Interspersed between the mica concentrations are equigranular, granoblastic, semi-polygonal assemblages of quartz and plagioclase, which form vein-like aggregates in places with selvedges of mica around them, and they resemble neosomes (HM 86).

6. Muscovite-quartz Schist (Mms)

Muscovite-quartz schist occurs within the Pofadder Lineament in the Mattheusgat Mountains. The rocks are invariably sheared and they appear to be structurally duplicated with mylonitic quartzite. Minor magnetite-rich laminae can be seen to define trough cross-bedding in places (Plate 2). On weathered surfaces the rocks are dark grey in colour, presumably due to the liberation of iron from magnetite. The mineral modes of the muscovite-quartz schists examined in thin section are listed in Table 8.

Table 8: Estimated mineral modal percentages for muscovite-quartz schists

	Qz	Mu	Ma	Pg	Kf	Ep	Pi	Cz	Ch
HM 24	95	4	1						
HM 60	87	7	3	2		1			
HM 63	91	6*	2				1		
HM 89	59	40	1						
HM 105	87	10	1	1				1	
HM 112	89	8	2			1			
HM 140	15	35	1	5	42				2

* Rose muscovite

The schists commonly contain more than 80 per cent quartz and have a mylonitic fabric, but are clearly different from mylonitic quartzite as muscovite imparts a schistose fabric. A further difference is that opaque mineral concentrations define bedding in places.

The rock is fine to medium grained and has a strong foliation defined by the recrystallisation of quartz into laminae. These laminae are also rich in muscovite which shows a strong alignment, and are most prominent in highly strained areas such as around the margins of porphyroclasts. The recrystallised quartz is very fine grained, granoblastic and relatively free of strain. Detail on the process of mylonitisation is given in Chapter V. The muscovite is relatively undeformed, cuts across quartz grains and is commonly of a rose-pink colour. It appears to have formed from retrogression during the shearing episode of deformation. In places, however, there is an older greenish-coloured muscovite with exsolution of opaque minerals along its cleavage (HM 89).

7. Quartzite

Glassy Quartzite	(Mq)
Granular Quartzite	(Mqg)
Iron Formation	(Mi)

Two types of quartzite were identified in the field; a glassy, milky coloured quartzite and a granular-textured detrital quartzite. The former is a medium-grained semi-monomineralic quartzite with minor amounts of opaque minerals and micas, in places being a micaceous quartzite. The granular quartzite is fine grained and grades into calc-silicate gneiss due to the development of lime-rich lamellae within the rock.

The glassy quartzite is typically associated with quartzo-feldspathic gneiss and commonly outcrops along shear zones forming excellent marker bands. In places it is of a dark blue-grey colour due to very fine-grained opaque minerals, which are in places (rare) sufficiently concentrated to form pods of massive haematite. Ironstone is also associated with the granular quartzite, but in this case tends to be banded.

The estimated mineral modes for the quartzites and ironstones studied in thin section are listed in Table 9. In thin section the glassy quartzite is medium grained, ranging from being inequigranular, seriate to equigranular and has a granoblastic interlobate texture. The rock commonly lacks a foliation, though in the micaceous glassy quartzite it is weakly developed with sillimanite needles also contributing to the foliation in some instances. Finer grained plagioclase, K-feldspar, chlorite after biotite, sillimanite and quartz are concentrated along quartz grain boundaries and in particular at grain junctions. This points toward a possible detrital origin for at least some of the glassy quartzites. In some specimens the rocks are feldspathic (HM 54, 93) and appear to grade into quartzo-feldspathic gneiss.

Where sheared, the glassy quartzite has a bimodal grain size comprising strained quartz porphyroclasts and bands of recrystallised quartz concentrated with mica defining a mylonitic foliation. Fibrolite is seen to occur as inclusions in quartz and also to cut across quartz grain boundaries indicating that quartz recrystallisation probably post-dates the fibrolite.

The granular quartzites are fine to medium grained, inequigranular, seriate and granoblastic with interlobate to polygonal grain margins. In places (HM 154) aggregates of quartz grains in approximate optical continuity exist and these are surrounded by a sheath of muscovite, indicating that the quartz grains could have been of detrital origin. As for the glassy quartzite, these rocks have a significant feldspar content in places and would appear to grade into quartzo-feldspathic gneiss. Sample HM 122 is metasomatised, and is comprised of quartz, epidote and minor amounts of opaque minerals only. The opaque mineral is thought to have resulted from the liberation of iron in mica and has itself been included into the recrystallised quartz. In places the opaque mineral still defines a relict foliation. There are also fluid inclusions in the rock forming "trails" within the quartz. The rock appears to have had all the alkali minerals flushed out.

Table 9:Estimated mineral modal percentages for quartzite and banded ironstone

Glassy Quartzite

	Qz	Mu	Bi	Ma	Zr	Kf	Pg	Sil	Ga	Cc	Ch	Ep	
HM 54	71	4	1	1		20		3					"glassy"
HM246	90	3		2				4				1	
HM 51	85		1			1	4	3		1	5		micaceous
HM 93	55	2	8			30	5						in
HM242	90	1	1	1				6	1				specimen
HM 18	95	3		1							1		mylonitic
HM 19	82	4		14									in
HM 35	93	1	1	2			2	1					specimen

Banded Ironstone

	Qz	Mu	Bi	Ma	Zr	Kf	Pg	Sil	Ga	Cc	Ch	Ep
HM114	74	4	1	18			2				1	
HM199	83		1	10	2		3				1	

Granular Quartzite

	Qz	Mu	Bi	Ma	Zr	Kf	Pg	Sil	Ep	Di	Hb	Ap	Sp
HM 95	85			1			10		2	2			
HM122*	79			1					20				
HM154	91	4	1			3		1					
HM161	80	2	3	1	1	3			6x		2	1	1
HM225	72	2				15			10		1		
HM256	66	1			1	30			2				

* Metasomatic rock x Zoisite with lesser epidote

In the banded ironstone, grains of magnetite and quartz have concentrated into alternate laminae which define a highly contorted surface that is probably bedding. The grains within these laminae are, however, aligned parallel to a foliation which transects the contorted mineral laminae. The intersection of the two surfaces results in an intersection lineation. Study in reflected light reveals that the opaque mineral in the banded and massive ironstones is magnetite which has partially been replaced by haematite. This has resulted in magnetite exsolving haematite as fine lamallae forming a challaise texture. In more extensively replaced grains the haematite lamellae have widened until the whole grain is replaced. The replacement is probably a near-surface weathering phenomena. Veins of secondary goethite also occur.

C. Pre-tectonic Intrusive Rocks

Pre-tectonic intrusive rocks have a pseudo-igneous texture in most cases, with a metamorphic imprint on them. Their intrusive origin is not always evident from field relations, usually because of poor exposure. Rock types included here are mafic granulites, ortho-amphibolites, cluster amphibolites, olivine gabbronorites, and granitic gneisses.

1. Mafic Granulite (Mo)

Mafic granulite (HM 261) occurs on the farm Gemsbok Vlakte at locality 29°01'40"S/ 19°35'30"E and has intruded leucogneiss. The granulite is interbanded with zones of ortho-amphibolite on a 5 to 50 cm scale. The amphibolite appears to be either a retrograde equivalent of the granulite or else to represent zones of weakness which had access to water during prograde metamorphism.

The estimated mineral mode of a mafic granulite is given in Table 10. The rock is a fine-grained, two-pyroxene, hornblende, plagioclase (An 71 by Michel-Lévy method) rock with a porphyroblastic texture. The porphyroblasts consist of a core of clinopyroxene (augite ?) with exsolution of two minerals, one being possibly a spinel and the other possibly an orthopyroxene, the latter occurring in very fine lamellae and at an angle to the former exsolved mineral. The clinopyroxene cores are thought to be of igneous origin and are surrounded by a partial or complete corona of hypersthene. In some porphyroblasts the core is sub-ophitic enclosing plagioclase, as shown in Fig.2.

Table 10: Estimated mineral modal percentages for mafic granulite

	Pg	Hb	Hy	Cpx
HM261	45	35	15	5

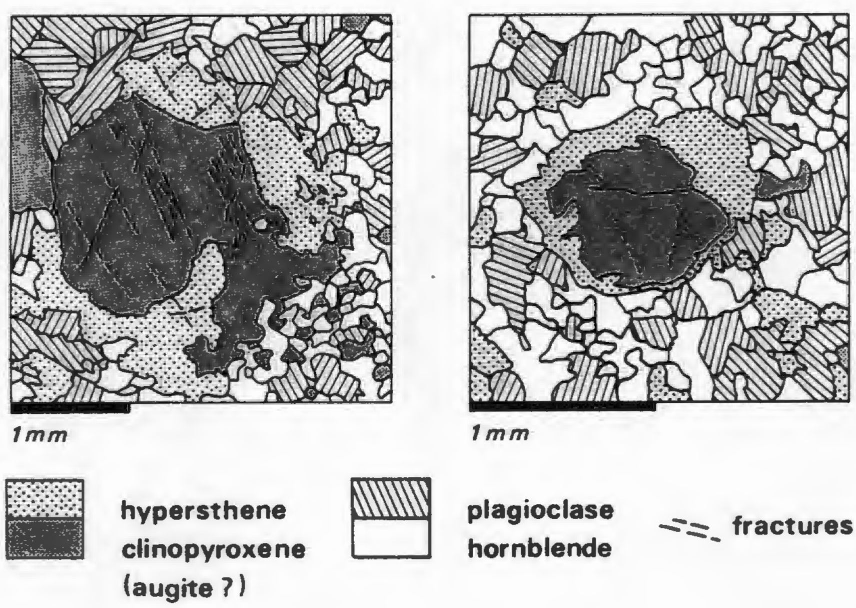


Fig.2: Corona textures in mafic granulite (HM 261). Partial or complete coronas of hypersthene surround grains of clinopyroxene which are in places sub-ophitic.

The coronas indicate slow cooling, from either an igneous or a metamorphic temperature, enabling calcium diffusion to occur, depleting the rim in this element. The matrix is equigranular, granoblastic and in many places polygonal with 120° triple junctions. Hornblende is a brown colour whereas it is a green colour in the surrounding amphibolite rocks. The rock has a metamorphic texture (D. Waters, pers.comm., 1983) and is a true granulite.

2. Olivine Gabbronorite (Mn)

The type area for olivine gabbronorite is within the Samoep synform west of the Nouzees shear zone. The only other occurrence of this lithology is on the farm Graaf-Water (29°17'25"S/ 20°04'30"E). The rock is typically associated with ortho-amphibolite which lacks the lamellae or banding of the para-amphibolite. There is generally no exposure around these rocks as they usually crop out as hillocks surrounded by sand and little is known about their contact relationships.

Table 11: Estimated mineral modal percentages for olivine gabbronorite

	Pg	Cpx	Opx	Ol	Am	Bi	Sp?	Ma	
HM 1	75	6	4	8	3	1	1	2	
HM 2	40	48	2		4	1		5	
HM 3	46	35	5	7	4	1	1	1	
HM 4	53	35	4	2	1	2	1	2	
HM 5	55	30	5	5	2	1	1	1	
HM 6	60	15	12	3	4	2	1	3	Nouzees Gabbronorite Suite
HM 7	50	35	4	2	2	2	1	4	
HM 8	45	40	3		6	2		4	
HM 9	45	45	3	1	1	2		3	
HM 10	55	34	3	1	1	2		4	
HM200	55	24	10		4	2	1	4	Graaf-Water

In Table 11 the estimated mineral modes for this lithology are listed and, in accordance with the I.U.G.S. Sub-commision (Streckeisen, 1973), these rocks are classified as gabbronorites since they contain both clino and orthopyroxene. The name is prefixed with olivine as the mineral comprises up to eight per cent of the rock in some specimens. The rock is medium grained and in thin section is seen to be inequigranular comprising decussate laths of plagioclase with a sub-ophitic texture. The plagioclase grains range from at least An 60 to 75 (by Michel-Lévy method) and many are zoned. Petrographically, the followiing points are of importance:

(a) Inclusions of olivine occur within plagioclase and vice versa in the same thin section (HM 5) indicating contemporaneous crystallisation of the two minerals.

(b) Coronas around olivine grains consist of an inner rim of orthopyroxene (in some olivine grains the whole grain is altered to orthopyroxene with a skeletal opaque mineral) and an outer rim of symplectite. The symplectite consists of a very fine-grained intergrowth of a green mineral having very high relief (possibly hercynite) and a green needle-like mineral of lower relief with inclined extinction (possibly an amphibole). The outer rim is usually only partially developed, typically forming between pyroxene and plagioclase. The corona texture is virtually identical to those described by Albat (1983b) in the Kliprand area of Namaqualand. In this case the coronas have an inner shell of orthopyroxene which is seen to replace the olivine, an intermediate shell of inclusion-free green amphibole and an outer shell of amphibole-spinel symplectite. The coronas typicallly occur between an olivine-plagioclase interface, as in the present study. Similar corona textures have also been described by Mason (1967) and by Whitney and McLelland (1973) all of whom postulate that a sub-solidus mechanism involving diffusion of ions via a fluid phase took place in a closed system, with water being the only additive. Albat (op.cit.) reached the same conclusion, postulating that primary olivine and plagioclase with water were the only reactants involved, producing orthopyroxene, amphibole and spinel. It is proposed that the same applies to the corona textures studied here since they have the same features.

(c) Exsolution has occurred in both clino and orthopyroxene:

i) Augite has exsolved pigeonite as fine lamellae, the lamellae commonly having been initiated in augite along cracks which may have acted as nucleating sites. In some augite crystals a different type of exsolution was noted, formed by very fine blebs of two phases, possibly an opaque mineral and an orthopyroxene, the two phases being orientated at a slight angle to each other.

ii) Hypersthene crystals have very fine exsolution lamellae, which are absent near the grain margins in many instances indicating that calcium in the marginal parts of the crystal was able to diffuse away.

(d) Vermicular myrmekites involving both ortho and clinopyroxene are also common.

(e) Double coronas around an opaque mineral (spinel ?) are also present, consisting of a relatively thick rim of biotite which is in turn rimmed by a brown amphibole.

(f) Recrystallisation of the pyroxene has occurred, forming subgrains which are still in approximate optical continuity.

(g) In fine-grained dykes of this rock type (HM 12) the sub-ophitic texture is still present, but most of the ferromagnesian minerals have been altered to amphibole. Possible relict grains of olivine occur as "spongy" aggregates of an opaque mineral.

It is clear that the reactions mentioned above (b-g) require diffusion to occur over a long period of time and hence the rocks must have cooled slowly after emplacement or metamorphism.

3. Amphibolite

Ortho-amphibolite (Mhlo)
Cluster amphibolite (Mha)

These amphibolites consist predominantly of plagioclase and hornblende, they lack the well-defined quartz lamellae so characteristic of the para-amphibolites and have only a weakly developed foliation, defined by hornblende.

In outcrop, the ortho-amphibolite is found in association with pre-tectonic mafic intrusives and is considered on the basis of this association to be the retrograded equivalent. The ortho-amphibolites crop out in two regions; they are associated with the Nouzees Gabbronorite Suite and also occur on the farm Gemsbok Vlakte where they are interbanded with mafic granulite on a 5 to 50 cm scale.

The cluster amphibolites form hillocks of similar outcrop style to the olivine gabbronorites, and as for those rocks, no contact relationships with the country rock were observed. Clusters of hornblende, surrounded by leucocratic plagioclase-rich haloes (Plate 3) are characteristic of this rock type. It must be remembered that they were also observed in the calc-silicate gneiss, but in the latter case they are not common and the host rock is much more strongly foliated.

For the purpose of description the rocks are divided into three groups, namely; the ortho-amphibolites associated with the Nouzees Gabbronorite Suite, those associated with the mafic granulite and, the cluster amphibolites. The estimated mineral modes for these rocks are listed in Table 12.

The ortho-amphibolite associated with the Nouzees Gabbronorite Suite is medium grained, inequigranular, seriate and granoblastic with interlobate to semi-polygonal grain boundaries. Hornblende has recrystallised into bands of somewhat coarser grain size with numerous 120° triple junctions, micas have altered to chlorite and to epidote in places, and plagioclase is commonly altered to clinozoisite.

The ortho-amphibolite associated with the mafic granulite on the farm Gemsbok Vlakte is fine to medium grained, equigranular and granoblastic with interlobate grain boundaries. In zones of high strain the hornblende is strongly aligned resulting in an interbraided shear fabric. The composition of the plagioclase is estimated to be An 70 by the Michel-Lévy method. In large areas of the thin section (HM 260) hornblende grains show approximate optical continuity and they may possibly represent coarse original grains of pyroxene which have subsequently recrystallised into sub-grains and retrograded to hornblende.

The cluster amphibolite is characterised by aggregates of medium-grained hornblende, in places surrounded by leucocratic haloes rich in plagioclase and almost devoid of amphibole and opaque minerals. The haloes are visible in hand specimen (Plate 3), and the clusters define a lineation in some places. In thin section the rock is fine to medium grained and inequigranular with porphyroblasts of hornblende in a matrix of plagioclase with minor amounts of finer grained hornblende, quartz and micas. In places the hornblende clusters have the following features which indicate that they could originally have been coarse-grained pyroxene:

(a)Weakly developed sets of inclusions inclined at right angles (HM 186) possibly conforming to a relict pyroxene cleavage.
(b)An intergrowth with pyroxene (HM264).
(c)A porphyroblastic sieve texture with inclusions of plagioclase (HM 98). These porphyroblasts have subsequently recrystallised into sub-grains which are still in approximate optical continuity.

A further step in recrystallisation is seen by the presence of fine-grained prismatic hornblende within the hornblende clusters and it appears to define a mineral lineation. The plagioclase An content appears to range from at least 40 to 75 (Michel-Lévy method) and some grains are zoned. The micas are commonly altered to epidote and chlorite.

Table 12: Estimated mineral modal percentages for ortho, and cluster amphibolites

	Pg	Hb	Qz	Sp	Di	Ap	Ma	Mu	Bi	Ch	Cz	Ep		
HM 27	49	40		1		3	2	1		1	2	1	*	Ortho-amphibolite
HM255	45	49		2			1		3				*	
HM260	55	44					1						+	
HM 23	63	20	5	2		1	1		8					Cluster-amphibolite
HM 28	48	50					2							
HM 94	37	45	5				4			2	5	2		
HM 98	25	73						1			1			
HM134	43	50									7			
HM186	54	40		1				1		1	1	2		
HM264	30	54			15							1		

* Associated with Nouzees Gabbronorite Suite
\+ Associated with Mafic Granulite on the farm Gemsbok Vlakte

4. Granitic Gneiss

Red Alkali-feldspar Granitic Gneiss	(Mal)
Megacrystic Alkali-feldspar Granitic Gneiss	(Mam)
Streaky Gneiss	(Mab)

The granitic gneisses have a moderately to strongly developed foliation and pre-date the shear zones. There are two main types of granitic gneiss; firstly a red, commonly porphyritic, alkali-feldspar granitic gneiss and secondly, a megacrystic, commonly biotite-bearing, alkali-feldspar granitic gneiss which appears to be transformed into a streaky gneiss in zones of high strain.

The type area for the "red" variety is on the farm Nouzees (29°13'20"S/ 19°32'45"E) where the rock is characterised by K-feldspar phenocrysts up to 1.5 cm in diameter. They are readily weathered out leaving the rock with a pitted surface. Also, in strongly weathered areas the rocks foliation is accentuated and the rock resembles a quartzo-feldspathic gneiss. Lensoid bodies of leucocratic porphyroblastic rock occur as possible xenoliths in the rock. The porphyroblasts consist of coarse microcline grains up to 2 cm in diameter.

The type area for the megacrystic variety is to the south of the east-west-striking shear zone in the northern part of the farm Pofadder East. These rocks are well exposed along the Pofadder-Kakamas road cutting in the extreme west of the field area and xenoliths (for which the internal and external foliations are parallel) can be observed. In the north of the farm Nouzees, in the extreme west of the field area, the rock type is exposed adjacent to the Nouzees shear zone. Here the megacrysts have been deformed into augen. It is possible that this rock with its well developed augen represents an intermediate in the transformation from the megacrystic variety to the streaky gneiss. The latter type does not commonly crop out, but appears to be associated with shear zones and is generally migmatitic (29°02'00"S/ 19°44'30"E).

The estimated mineral modes for the granitic gneisses studied in thin section are listed in Table 13. The quartz, alkali-feldspar and plagioclase mineral proportions for these rocks are also shown (Fig.3), from which they are classified as alkali-feldspar granitic gneisses (Streckeisen, 1973). In thin section the red alkali-feldspar granitic gneiss is of medium grain, inequigranular, seriate and granoblastic with interlobate to polygonal (rare) grain boundaries. In sheared rock (HM 116, 258) quartz has recrystallised along grain boundaries forming anastomosing zones of recrystallised quartz. Phenocrysts are composed of microcline, plagioclase or quartz. In thin section the porphyroblast-rich lens mentioned earlier (p.24), which in the field resembles a xenolith, could on petrographic grounds equally well be a megacrystic granite. The rock has a granoblastic texture and is composed of quartz with lesser amounts of plagioclase, some muscovite and interstitial microcline in a matrix with coarse-grained porphyroblasts of microcline which enclose quartz and plagioclase. The plagioclase is estimated to be An 10 by the Michel-Lévy method.

Table 13: Estimated mineral modal percentages for granitic gneisses

	Qz	Pg	Kf	Hb	Bi	Mu	Ch	Ep	Ap	Sp	Zr	Ma		
HM116	20	13	65			2							+	Red Alkali-Feldspar Granitic Gneiss
HM210*	35	12	50		1							2		
HM258	31		65					1				3	++	
HM 17	50	3	45				1					1	x	
HM104	45	7	43		2	1						2	x	
HM 91	25	10	58		3		1	1		1		1		Megacrystic Alkali-Feldspar Granitic Gneiss
HM202	35	10	45		5		1		1	1	1	1		
HM259	20	20	43	2	10		1		1	1	1	1		
HM 21	36	5	50	2	5					1		1	+	
HM250	30	5	55	8						1		1	++	
HM257	25	15	51		8		1						+	Streaky Gneiss

x Porphyritic
+ Protomylonite
++ Ultramylonite
* Not red in colour

In the megacrystic alkali-feldspar granitic gneiss the rock is inequigranular, with coarse-grained microcline or perthite. In some instances (HM 21) the megacrysts show a gradual transition from perthite in the core to microcline at the margins which appears to be in response to strain. Also, in some areas the megacrysts have formed sub-grains, some being perthite, although most are microcline (HM 202). Quartz has recrystallised extensively, especially around the megacrysts and, as in the red alkali-feldspar granitic gneisses, forms anastomosing recrystallised bands with abundant mica.

In thin section the streaky gneiss has the same fabric as the megacrystic variety, but quartz veins are also developed corresponding to the neosomes seen on outcrop scale and which at that scale can be observed to have biotite selvedges about them.

D. Syn to Post-tectonic Intrusive Rocks

1. Late Granitoids and Diorites

Granodiorite-tonalite	(Mg)
Pyroxene-bearing Diorite	(Md)
Fine-grained Quartz-plagioclase Porphyry	(Mdp)

The late granitoids and diorites vary considerably in their field appearance, but seldom have any well developed foliation. They are thus relatively undeformed and in places intrude shear zones on a mesoscopic scale. The following varieties were observed:

(a) Granodiorite-Tonalite

i) A biotite-rich, megacrystic granitoid intruding ortho-amphibolite on the farm Nouzees. In places it occurs in the form of pegmatitic veins one or two metres wide that are rich in biotite forming coarse "books" one to two cm across. Also on the farm Nouzees a biotite-rich granitoid (HM 252) has intruded quartzite and has abundant xenoliths, typically one to four cm across.

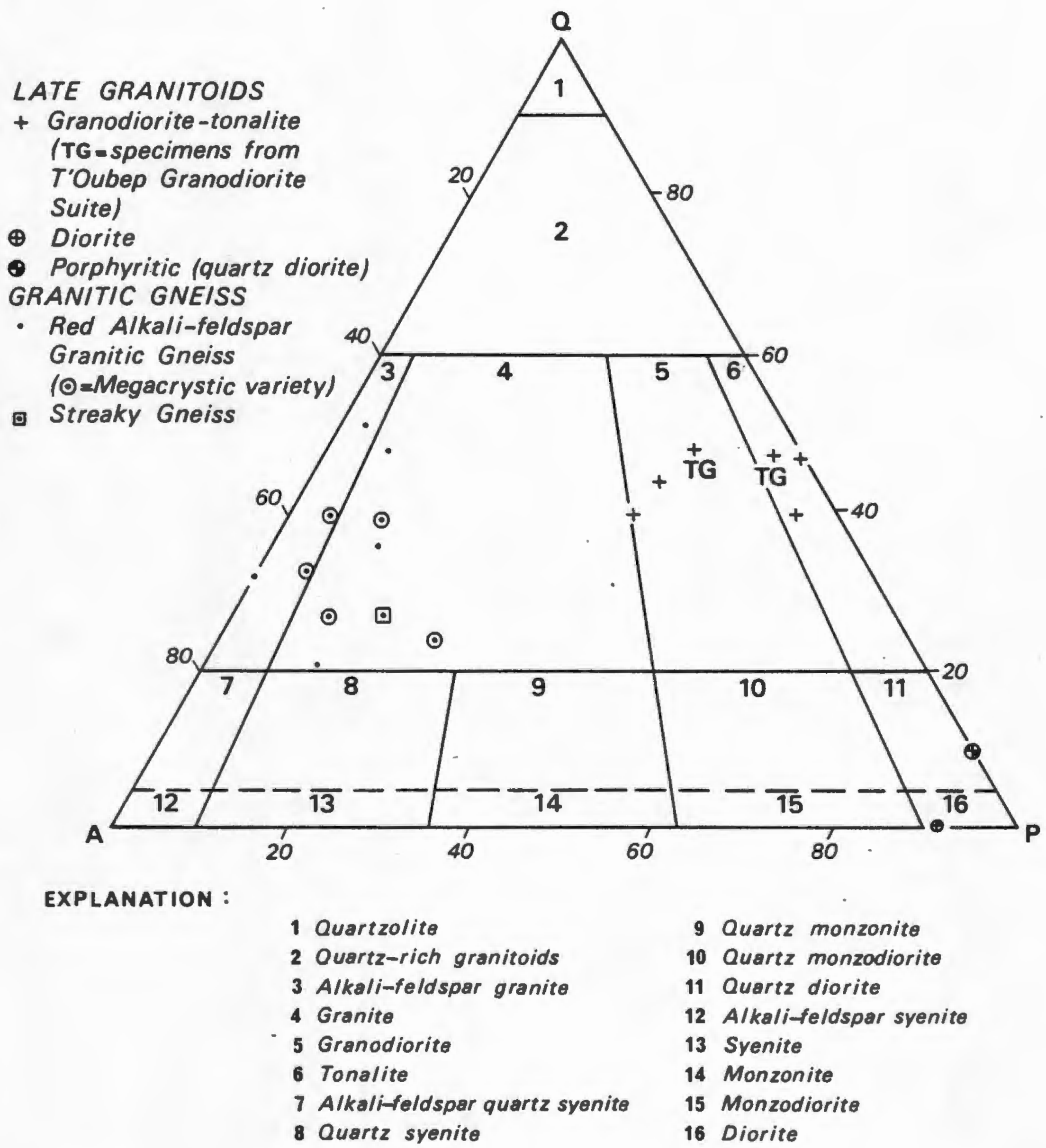
ii) A leucocratic plagioclase-quartz rock, typically lacking a foliation (HM 126) and which has intruded a minor shear on the farm Lucas Vlei Vlakte (19°00'45"S/ 19°42'40"E).

iii) A leucocratic, black weathering, medium to coarse-grained hornblende-bearing granitoid. The T'Oubep Granodiorite Suite is comprised of this rock type, forming a large body approximately 10 km across from north to south. The body occurs on the farm T'Oubep, straddling the eastern margin of the field area (HM 205,207) and is strongly jointed in places, cropping out as isolated hillocks. Calc-silicate rock xenoliths are common in the rock, on both a mesoscopic and a megascopic scale. As is the case for the host rock, these xenoliths are also black weathering, but from field work they appear to be common only in the south of the intrusive body.

(b) Pyroxene-bearing Diorite - On the farm Adjoining Geelvloer rocks similar to the T'Oubep Granodiorite Suite crop out along the northern side of an east-southeast-striking shear zone, but they contain clinopyroxene in addition to the hornblende and lack quartz (HM 174).

(c) Fine-grained Quartz-plagioclase Porphyry - A micro-granitoid with megacrysts of plagioclase and of quartz in a very fine-grained feldspathic matrix (HM 235) occurs in the south of the farm Drooge Grond, near the eastern boundary of the field area.

Quartz, alkali-feldspar and plagioclase mineral proportions of the late granitoids and diorites have been plotted (Fig. 3) and used to classify the rocks (Streckeisen, 1973). The granodiorite-tonalite rocks range in composition from granodiorite to tonalite, and in contrast, the pyroxene-diorite and the fine-grained quartz-plagioclase porphyry plot in the plagioclase-rich corner of the diagram. The porphyry is a quartz-diorite following this classification scheme.



EXPLANATION :

1 *Quartzolite*
2 *Quartz-rich granitoids*
3 *Alkali-feldspar granite*
4 *Granite*
5 *Granodiorite*
6 *Tonalite*
7 *Alkali-feldspar quartz syenite*
8 *Quartz syenite*
9 *Quartz monzonite*
10 *Quartz monzodiorite*
11 *Quartz diorite*
12 *Alkali-feldspar syenite*
13 *Syenite*
14 *Monzonite*
15 *Monzodiorite*
16 *Diorite*

Fig.3: Quartz - alkali-feldspar - plagioclase diagram for classifying the granitic rocks studied (after Streckeisen, 1973).

Table 14: Estimated mineral modal percentages for late granitoids and diorites

	Qz	Pg	Kf	Hb	Cpx	Bi	Ch	Ep	Ap	Cc	Sp	Zr	Ma	
HM 15	40	35	15	1		3	1	1*	1		1		2	
HM 16	40	45				10	1	1	1			1	1	
HM126	40	56	4											Grano-
HM205**	45	38	10	4					1		1	1		diorite/
HM207**	45	47	3	3					1		1			Tonalite
HM252	36	35	20			7		1					1	
HM174		73	7	15	2			2*		1				x
HM235	10	86				3	1							+

* clinozoisite (+- Ep)
**T'Oubep Granodiorite Suite
x Pyroxene-bearing Diorite Gneiss
\+ Fine-grained Quartz-plagioclase Porphyry

The estimated mineral modes for the late granitoids and diorites are listed in Table 14. In thin section the granodiorite-tonalite granitoids are medium grained (the T'Oubep granitoids are medium to coarse grained) and have an inequigranular, seriate, granoblastic texture with interlobate grain boundaries, being polygonal in places. The grain boundary shape indicates that these rocks bear some metamorphic imprint. Myrmekitic texture was observed in thin section for HM 16, 205, 207 and 252, possibly indicating a replacement mechanism, as proposed by Becke (1908):

$$\underset{\text{orthoclase}}{KAlSi_3O_8} + Na \longrightarrow \underset{\text{albite}}{NaAlSi_3O_8} + K \qquad (1)$$

$$\underset{\text{orthoclase}}{2KAlSi_3O_8} + Ca \longrightarrow \underset{\text{anorthite}}{CaAl_2Si_2O_8} + 4SiO_2 + 2K \qquad (2)$$

The silica would precipitate as vermicular quartz, and the albite and anorthite would mix to give, for example, oligoclase (Phillips,1980).

The pyroxene-bearing diorite (HM 174) is composed predominantly of plagioclase (An 41 by Michel-Lévy method) with lesser amounts of hornblende and K-feldspar. Clinopyroxene, epidote, carbonate and clinozoisite occur in minor amounts. The rock is medium to coarse grained with euhedral plagioclase and clinopyroxene grains, the latter being replaced by a pale blue-green coloured hornblende in a patch-work mosaic.

The quartz-plagioclase porphyry is a very fine-grained hypabyssal rock with plagioclase and quartz phenocrysts occurring in a feldspathic matrix. The quartz phenocrysts are rimmed by biotite which has partially been replaced by chlorite.

2. Pegmatite and Vein Quartz

Vein Quartz (Nqv)
Pegmatite (Ngp)

Pegmatite is common throughout the area mapped. It occurs as narrow (<10cm) veins along shear zones and also along zones of brittle fracture. On the farm Nouzees a large body of pegmatite crops out and has been prospected for amazonite and also gadolinite (P. Joubert, pers.comm., 1983). Other minerals associated with the pegmatite are black tourmaline (shorl), purple tourmaline (very rare, possibly rubellite), green tourmaline (possibly verdelite) and garnet.

Pegmatite scree has been observed throughout the southern part of the field area, extending from the Putsberg shear zone in the south-west to the shear zone on the farm Adjoining Geelvloer in the south-east. Much pegmatite scree was also observed north of the T'Oubep Granodiorite Suite.

Vein quartz was noted to cap pegmatite in areas on the farm Nouzees but more commonly crops out along fractures, for example those which extend south-east from the Mattheusgat Mountains.

3. Dolerite (Jd)

Dolerites are post-tectonic and in places exhibit onion-skin weathering patterns. The rock commonly has a similar appearance to the olivine gabbronorite which is also unfoliated, but is somewhat less melanocratic. Dolerite outcrops in four areas, namely:

(a) On either side of the Pofadder-Kakamas road near the north of the field area, cropping out as hillocks which probably conform to a relict sill. In the extreme north the dolerite can be seen to have intruded leucogneiss at "Tafelkop" as a sill, truncating a subvertical foliation in the rock.
(b) At the junction of the farms De Neus, Sandgat and Mattheusgat, dolerite intrudes as a sill, once again cutting the foliation of the host rock. Small hillocks crop out between this locality and the previous one at "Tafelkop", possibly suggesting the presence of an even more extensive relict sill.
(c) In the extreme north of the farm Valsvlei dolerite also crops out, but here no contact relationships were seen.
(d) In the extreme east of the field area, on either side of the Drooge Grond - Brul Kolk farm boundary the presence of dolerite dykes was recorded.

In Table 15, the estimated mineral modes for dolerites that were studied in thin section are listed. Dyke rocks have also been studied, but they are too fine grained for their mineral modes to be estimated.

Table 15: Estimated mineral modai percentages for dolerite

	Cpx	Pg	Ol	Ch	Ma
Generalised estimate (HM:55,67,81,82,111, 123,127,128,173,262)	48	40	4	2	6
HM 223 (olivine-rich)	40	35	21	1	2

The majority of the specimens are medium grained and sub-ophitic with decussate plagioclase laths. The plagioclase is labradorite (An 57-71, by Michel-Lévy method). The pyroxene is predominantly augite although orthopyroxene (pigeonite ?) does occur in places. The olivine-rich specimen comes from the eastern-most locality, (d) above, and although it is finer grained, it has the same texture.

In thin section the dolerites differ markedly from the olivine gabbronorites as they lack coronas and do not show any signs of metamorphism.

4. Kimberlite (Kk)

Frick (1974) describes the occurrence of ten kimberlites which occur to the southeast of Pofadder, six of which occur in the present study area on the farm Nouzees. The assumption by Frick (op.cit.) that the northwest-southeast arrangement of kimberlites is structurally controlled has been substantiated by the present mapping as they occur in close proximity to the Nouzees shear zone.

The pipes are not easily observed in the field and were located by prospectors (ibid.) with the use of infrared scanning and aeromagnetic techniques. In the field a correlation between the extent of prospect pits and topographic depressions where calcrete is better developed could be made and on this basis the possibility of two further pipes, also located along the Nouzees shear zone was noted.

The pipes described by Frick (op.cit.) were discovered by Mr. R.G. Niemöller and prospected by him, but without success. The prospect dumps indicate that the pipes are garnetiferous and rich in an opaque mineral, and fragments of dolerite in the dumps are also fairly common. Frick (op.cit.) has classified the kimberlites as basaltic kimberlites following the nomenclature of Wagner (1914) and as autolithic kimberlites following the nomenclature of Rabhkin et al. (1962).

E. Cover Rocks and Superficial Deposits

1. Karoo Sequence

Dwyka Gravel (Qdg)
Dwyka Float (Qdf)
Dwyka Shale (Cds)
Dwyka Tillite (Cdt)

The lower-most formation of the Karoo Sequence, the Dwyka Formation, occurs in the south of the area and consists of angular fragments set in a finer grained matrix of silty material. In most areas where the rock occurs it forms float of round boulders up to 20 cm across, but in the east there are also extensive areas of Dwyka gravel, especially on the farm Loogkolkjes and environs. Shale beds were also mapped in two localities, one on the farm Gras Kopjes in the southwest, and another on the farm Loogkolkjes in the southeast where the shale contains pockets of gypsum. The float increases in thickness southward and was noted to be in situ only in the very southwest of the area.

Small areas of in situ Dwyka crop out amongst the float in many areas and are too small to be mapped. In one such area, to the southeast of the homestead "Nuwedam" in the south of the farm Sandgat (southeast of the quartzite hills, on the northern side of the road) the rock is intensely fractured. The locality coincides with the trace of the Putsberg shear zone, strongly suggesting that movement of the shear continued at least during the Carboniferous.

2. Calcrete

The calcrete is thickly developed between the Karoo Sequence and the underlying gneisses, with Dwyka float scattered throughout the calcrete. Calcrete also occurs in proximity to the calc-silicate gneiss and also within topographic depressions along drainage courses.

3. Aeolian Sand

The sand is red, well rounded windblown material, similar to the Kalahari type. Sand accumulations have formed, and still are forming, in a region that extends from the farm Gemsbok Vlakte in the northwest of the field area to the farm Valsvlei further southeast. In the northwest of the occurrence, sand dunes have formed trending approximately normal to the general direction of accumulation, namely in a south-southwest - north-northeast direction. Elsewhere, the sand does not form dunes, but occurs either as a flat-lying cover over extensive regions or against the slopes of hills on the side of approach of the prevailing wind.

4. Alluvium and Sand Pans

Alluvial sand and silt is encountered in most of the dry water courses and also in topographic depressions from which there is no outward drainage. When it rains, water stagnates in these depressions and a thin veneer of silt and clay accumulates. This results in the development of pans, commonly accompanied by dust-bowl conditions. As a result of their water-holding ability due to the clay content, these pans are normally surrounded by a ring of vegetation.

Chapter III

METAMORPHISM

A. Introduction

An attempt to determine the P-T conditions of metamorphism has been seriously hampered by the sparse outcrop in the area. Furthermore, pelitic rocks, which are the most informative metamorphic rocks to study are the least common lithology and only four specimens were collected. The approach towards studying metamorphism has been as follows:

1) To define diagnostic criteria which are to be used in determining the metamorphic grade of mineral parageneses.
2) To subdivide the lithologies into compositional groups and for each group to determine the metamorphic grade(s) from the diagnostic mineral parageneses. Mineral reactions that appear to have occurred are also described.
3) To construct a petrogenetic grid in order to constrain the P-T conditions of metamorphism.
4) To use selected geothermometers and geobarometers, and to compare their results with those from the petrogenetic grid.

B. Diagnostic Mineral Parageneses

For each metamorphic grade the following criteria have been used to define diagnostic mineral parageneses:

1. Low-grade Rocks

Chlorite in the presence of muscovite and quartz is used to define low grade metamorphism because, at the start of medium grade chlorite disappears by either of the following reactions (Winkler, 1976, pp.76-77):

$$\{Ch + Mu + Qz\} = \{Cd + Bi + Al_2SiO_5 + H_2O\} \qquad (3)$$

$$\{Ch + Mu\} = \{St + Bi + Qz + H_2O\} \qquad (4)$$

In mafic rocks the presence of chlorite in the paragenesis

$$Pg + Hb + Cz/Ep + Qz + Ch + Ga$$

is diagnostic of low grade (ibid.,p.170) and garnet will not be present unless the pressure is sufficiently high (Fig.6). The presence of actinolite-tremolite instead of hornblende indicates the lower temperature range of low grade, namely less than about 500°C for medium to high pressures (ibid.,p.171). In this region hornblende is unstable and is converted to actinolite/tremolite, chlorite and clinozoisite or epidote (ibid.,p.170).

2. Medium-grade Rocks

In mafic rocks hornblende is stable from the higher temperature part of low grade up to high grade (Winkler, 1976, p.167) and mineral parageneses within the "amphibolite facies" (hornblende + plagioclase) are not diagnostic of metamorphic grade. The absence of chlorite is assumed to indicate that the metamorphic grade exceeds low grade (ibid.,p.76,168). A further distinction between low grade and higher grades is that for low grade mafic rocks the An content of plagioclase must be on the lower side of the An 5 - 17 gap. Note, however, that the gap commonly occurs at 20-40°C lower than the low to medium grade transition (ibid.,p.166).

The beginning of anatectic melting in muscovite-bearing gneiss is taken to represent the start of high-grade metamorphism, provided that H_2O pressure is greater than about 3,5 kb, as given by the reactions:

$$\{Mu + Qz + Pg(albite) + H_2O\} = \{liquid + Al_2SiO_5\} \qquad (5)$$

$$\{Kf + Qz + Pg(albite) + H_2O\} = \{liquid\} \qquad (6)$$

Note that Winkler (op.cit.,p.85) shows that the presence of muscovite in reaction (6) is not necessary. When H_2O pressure is less than about 3,5 kb the reaction:

$$\{Mu + Qz\} = \{Kf + Al_2SiO_5 + H_2O\} \qquad (7)$$

occurs and plagioclase is not a necessary reactant (ibid.,p.85). If sillimanite is included in the mineral paragenesis for which muscovite is stable in the presence of quartz, the P-T conditions are restricted to a relatively small field of P-T space, marking the upper range of medium grade.

3. High-grade Rocks

In pelitic rocks, cordierite is broken down according to the reaction

$$Cd = \{Alm + Sil + Qz\} \qquad (8)$$

(Winkler, 1976, p.229) and if biotite is also present, the reaction

$$\{Cd + Bi\} = \{Alm + Kf + H_2O\} \qquad (9)$$

will take place together with the former as a coupled reaction at the same conditions (Currie, 1971; Winkler, op.cit., p.229). For these rocks K-feldspar is an index mineral and muscovite cannot coexist with quartz (Winkler, op.cit.,p.82). This is the [cordierite-almandine]-high grade metamorphic mineral paragenesis which is as follows (ibid.,Fig.14.11b):

Cd + Alm + Sil/Bi + Kf + Qz

C. Mineral Parageneses and Reactions

The lithologies are subdivided into four compositional groups, namely pelitic, calc-silicate, felsic and mafic rocks. Mineral reactions are described for each group of lithologies and the diagnostic mineral parageneses are shown in Table 16.

1. Pelitic Rocks

In pelitic gneiss the following aspects are of importance:

(a) Two generations of sillimanite are present (HM 25) with the older sillimanite defining a foliation due to its fibrous habit and mat-like development. The younger sillimanite occurs as coarse sub-idioblastic aggregates and is intergrown with quartz (HM 33) in places, indicating reaction (5) to have occurred. It appears that neither reaction (7) nor (10) occurred in place of (5) since K-feldspar and biotite were not formed as reaction products.

(b) Of the specimens studied in thin section HM 31 is of [cordierite-almandine]-high grade showing the following mineral paragenesis:

Cd + Ga + Sil + Mc + Qz.

Specimen HM 25 is also stable at high grade, but is not diagnostic as it lacks K-feldspar. Further evidence that HM 31, and possibly HM 25, reached high grade is that they lack muscovite in the presence of quartz and that in HM 25 veins of recrystallised quartz can be seen under the microscope indicating that partial melting probably occurred.

(c)Biotite was not observed to be in contact with garnet in HM 31, and from reaction (9), (see also Figs.4 and 6) this probably indicates that garnet formed from the biotite and that its formation was controlled by diffusion processes.

It is therefore likely that high grade metamorphic conditions occurred in places with the development of sillimanite from quartz, muscovite, plagioclase and water and that limited partial melting occurred. An interesting point is that the high grade pelitic rocks occur in the immediate vicinity of the Nouzees Gabbronorite Suite and it is possible that their high grade metamorphism is the result of contact metamorphism from intrusion of the gabbronorite. Such a metamorphism, if it occurred, must pre-date the peak of metamorphism since the gabbronorite itself was subjected to peak metamorphism. With this in mind, it is reasonable that the pelitic gneiss has a regional metamorphic fabric since peak metamorphism would have overprinted the contact metamorphism. The garnet in the rock is, however, idioblastic, possibly reflecting the original texture. The mineral paragenesis for the pelitic gneiss (HM 31) is listed in Table 16.

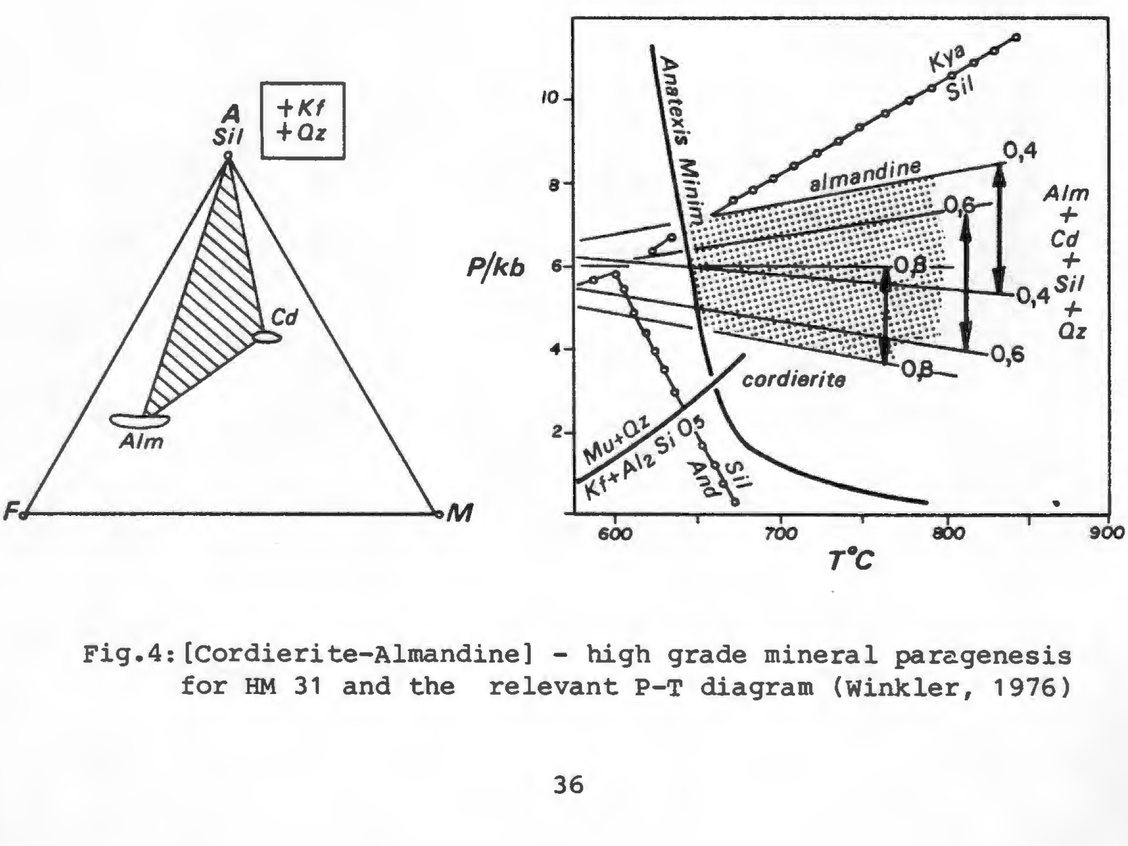


Fig.4: [Cordierite-Almandine] - high grade mineral paragenesis for HM 31 and the relevant P-T diagram (Winkler, 1976)

2. Calc-silicate Rocks

In calc-silicate gneiss there is evidence for both prograde and retrograde metamorphism and the reactions observed will be discussed in the order that they are assumed to have occurred:

(a) The rock is thought to originally have been comprised of plagioclase, garnet, diopside, hornblende, carbonate, a mica and quartz. Some diopside is seen to enclose hornblende within porphyroblasts (HM 34), indicating that at least some diopside post-dates the hornblende, although they appear to have been in equilibrium.

(b) During prograde metamorphism, recrystallisation predominated. In particular, specimen HM 44 collected from the extreme southwest of the area is comprised of, inter-alia, polygonal garnet and diopside. The two minerals are intergrown in places forming a sieve texture and also contain plagioclase inclusions. During the prograde event scapolite could have formed from a reaction such as the following (Deer et al., 1966, p.388):

$$\underset{\text{anorthite}}{3\ CaAl_2Si_2O_8} + \underset{\text{calcite}}{CaCO_3} \longrightarrow \underset{\text{meionite}}{Ca_4Al_6Si_6O_{24}CO_3}$$

(c) During retrograde metamorphism the amphibolite is considered to have undergone at least two stages of reaction, of which the second is possibly associated with shearing. In the first stage the original hornblende is considered to have remained unchanged, but clinopyroxene probably underwent retrograde metamorphism to hornblende, as seen in the thin section HM 247. In thin sections for HM 151,168,216 and 247 hornblende occurs in aggragates of medium to coarse grain size forming clusters which are surrounded by leucocratic haloes rich in plagioclase and deficient in opaque minerals. It is thought that the clusters formed after coarse-grained pyroxenes. In the second stage of retrograde metamorphism, concerning amphibole, the original hornblende recrystallised into bands consisting of the amphiboles of the tremolite-actinolite series and has a somewhat coarser grain size (HM 135,153). The hornblende clusters have also crystallised, in part into amphiboles of the actinolite-tremolite series (HM 135) as prismatic grains defining a mineral lineation. Other retrograde metamorphic reactions that have been noted, especially in the sheared rocks, are as follows:

i) Epidote after clinopyroxene and hornblende (HM 80, 162).
ii) Clinozoisite after plagioclase, and in places a mortar of epidote has formed between the two minerals (HM 29).
iii) Chlorite after a mica (biotite?), commonly defining a mylonitic foliation (HM 78,167).

The common mineral parageneses for the calc-silicate gneiss are shown in Fig.5 and Table 16. From a microprobe analysis (Appendix III) the garnet in the calc-silicate gneiss (HM 196) is highly calcic, being enriched in the andradite molecule as it contains 31,79 weight per cent CaO and 25,01 weight per cent total iron.

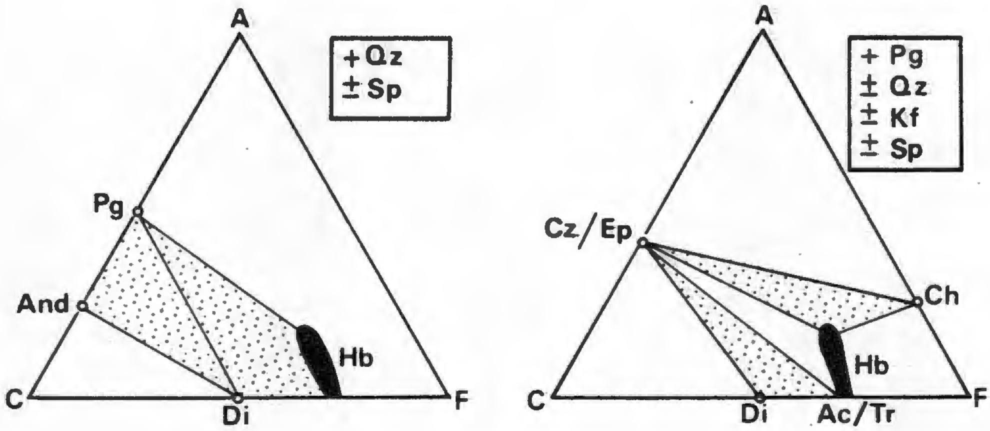


Fig.5: Ternary diagrams for a) medium grade and b) low grade mineral parageneses of calc-silicate gneiss. Diagrams after Turner, 1981, (Figs. 10-3a and 9-18a) with modifications.

3. Felsic Rocks

The felsic rocks include the lithologies quartzo-feldspathic gneiss, muscovite-quartz schist and quartzite. In both the quartzo-feldspathic gneiss and the muscovite-quartz schist two generations of muscovite exist. The older, in contrast to the younger generation is coarser grained, does not cut across quartz grain boundaries and is noticeably deformed. This generation, being stable in the presence of quartz and plagioclase indicates that peak (?) metamorphism was medium grade and did not enter high grade. The younger generation is associated with epidote minerals and chlorite; and is diagnostic of low grade metamorphism overprinting the above-mentioned medium grade mineral paragenesis. In the quartzite only one generation of muscovite is apparent and it occurs in medium and low grade mineral parageneses, as described for the previous-mentioned lithologies. The mineral parageneses for these felsic lithologies are shown in Table 16.

4. Mafic Rocks

The mafic rocks consist of para, ortho and cluster amphibolites, biotite gneiss, mafic granulite and olivine gabbronorite. The ortho and cluster amphibolites and the olivine gabbronorite are, however, omitted since their parageneses do not contribute to a better understanding of metamorphic grade, but merely define the "amphibolite facies". The mafic rocks reflect medium to high grade metamorphic conditions, overprinted by low grade conditions. In terms of the main metamorphic event the para-amphibolites record medium grade metamorphism, the biotite schist records the medium to high grade transition and the mafic granulite records granulite high grade. The rocks shall therefore be discussed in this order, reflecting equilibrium at progressively higher temperatures.

In the para-amphibolite, growth of hornblende is thought to have predominated during prograde conditions, resulting in the inclusion of grains of epidote (HM 53), feldspar and quartz (HM 130) in hornblende crystals which are in optical continuity. It is interesting to note that in the cluster amphibolite, hornblende is thought to have started replacing coarse-grained pyroxene during prograde metamorphism whereas in calc-silicate gneiss pyroxene was stable during these conditions.

In the biotite schist, partial melting resulted in the development of quartzo-feldspathic neosomes with selvedges of biotite. However, reaction (5) probably did not run to completion since muscovite, which appears to be primary, is still present, coexisting with quartz and plagioclase. This seems to indicate that the P-T conditions of equilibrium coincided with the medium to high grade transition. The mineral paragenesis for this lithology,

Bi + Mu + Qz +- Pg +- Sil

indicates that upper medium grade metamorphic conditions prevailed (See Section B.2.).

In the mafic granulites clinopyroxene is surrounded by coronas of hypersthene which are discontinuous in places. The rock texture is porphyroblastic with clinopyroxene grains ranging from porphyroblasts to sub-ophitic grains (see Fig.2). The texture is metamorphic and the coronas could be due to calcium migration into the clinopyroxene as a result of slow cooling. There are two types of exsolution within the clinopyroxene cores, namely a greenish mineral with high relief (spinel,possibly hercynite) and possibly orthopyroxene which is very finely exsolved. The exsolved phases are poorly developed though and hence the rock is expected to serve as a meaningful geothermometer.

It is uncertain whether the coronas formed as a result of regional or contact metamorphism. Since the rock is a mafic intrusive it is plausible that the coronas formed by contact metamorphism resulting from successive pulses of intrusion (see Section F.). The mineral paragenesis for the rock (HM 261) is

Hb + Hy + Cpx + Pg.

The local development of this granulite high grade metamorphism is attributed to the fact that the mafic body would have had considerably less water available than its surrounding paragneisses. The rock is intercalated with narrow bands of amphibolite, which may represent possible zones of weakness which had access to water during prograde metamorphism. Alternatively, the amphibolite bands could have formed as a result of retrograde metamorphism at a later stage but this does not explain why the granulite high grade rocks are developed only locally and rather implies an almost complete retrogression on a regional scale, an idea which seems rather unlikely.

Table 16: Diagnostic mineral parageneses

	LOW-GRADE low tempera- ture range			MEDIUM-GRADE		HIGH-GRADE	
	Ac/Tr +Cz/Ep +Ch	Mu+Ch +Qz	Pg+Hb +Qz+Ch +Ep/Cz	Pg+Hb +Qz±ZO (no Ch)	Mu+Qz+ Sil±Pg	Cd+Alm +Sil+Kf +Qz	Hy- bear- ing
				....	....	->	
Pelitic Gneiss						Qz Mc Alm Cd Sil	
Calc-silicate Gneiss	Pg Qz Ac/Tr Di Cz/Ep Ch			Pg Qz Hb Ga Di Sp			
Quartzo-feldspathic Rocks		Ch Mu Pg Qz ±Ep		Pg Qz Ga Di Hb ±Mu			
Muscovite-quartz Schist		Ch Mu Ep Cz ±Pi					
Quartzite		Qz Mu ±Ep/Zo ±Ch			Qz Mu Bi Sil Alm		
Para-amphibolite			Pg Qz Hb Ch Cz	Pg Qz Hb Zo/Ep Sp ±Ga			
Biotite Schist					Pg Qz Mu Bi Sil		
Mafic Granulite							Hy Hb Cpx Pg

D. The Petrogenetic Grid

In order to place some constraints on the P-T conditions of peak metamorphism, the positions of reaction curves in P-T space are now considered so as to develop a petrogenetic grid, (Fig.6) using data from Winkler (1976). As a first constraint, it is assumed that P-T was completely within the sillimanite stability field as no kyanite or andalusite was observed in any of the thin sections studied. Secondly, two reaction curves, namely:

$$\left\{St + Mu + Qz\right\} = \left\{Bi + Al_2SiO_5 + H_2O\right\} \qquad (10)$$

and

$$\left\{1Tr + 3Cc + 2Qz\right\} = \left\{5Di + 3CO_2 + 1H_2O\right\} \qquad (11)$$

occur in proximity to each other and both serve as constraints on the same side of the P-T space concerned. In regard to reaction (10) no staurolite has been observed in the pelitic rocks whereas quartz, sillimanite and less commonly biotite do occur indicating that the mineral paragenesis stabilised to the right side of the reaction curve. Similarly, with regard to reaction (11), no tremolite has been observed in the calc-silicate gneisses whereas diopside, quartz and less commonly calcite occur indicating that this reaction stabilised also to the right side of the reaction curve. Thus both reactions (10) and (11) define the lower temperature limit in P-T space with the former reaction acting as a tighter constraint of the two (Fig.6). Migmatites (see Section C.4.) are present in some areas indicating that P-T conditions must have, in places at least, crossed the melting curves (5) or (6). Finally, the [cordierite + almandine] stability field cannot be precisely defined (see Figs.4 and 6) because no whole-rock analyses were done and hence the bulk Fe/(Fe + Mg) ratio is unknown. As an estimate the most iron-rich stability field (Winkler, 1976) where Fe/(Fe + Mg) = 0.8 is used as a lower pressure limit. This ratio of 0.8 is also the value obtained by Moore (1977) for the pelites of the Namiesberg (analysis VGH1) which is located approximately 37 km west of the present study area.

In the light of all the above constraints and estimates for peak metamorphism (see Fig.6), temperature was probably in the range of 625 to 675°C and pressure probably approximated 4,5 kb.

The P-T conditions for low grade metamorphism could not be precisely established, but appear to have been within the low-temperature range of low grade metamorphism (greenschist facies). The presence of actinolite-tremolite in some of the calc-silicate gneiss specimens indicates temperatures to have been less than about 500°C for pressures greater than approximately 4 kb. The absence of garnet in low grade mafic rocks,however, indicates pressure to have been lower than 4 kb (see Fig.6). As a compromise, and without any further available data, it is tentatively suggested that for low grade metamorphism, temperature was less than approximately 500°C and that pressure was less than approximately 4 kb.

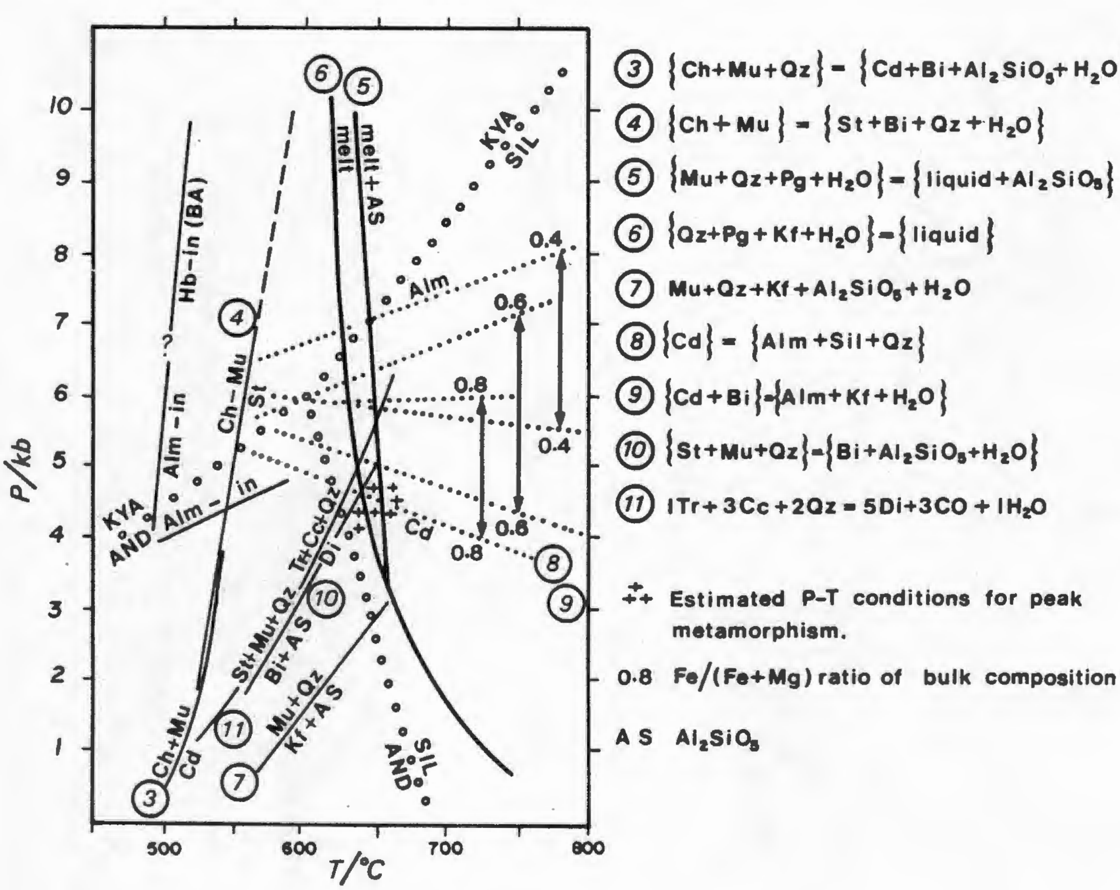


Fig.6: Petrogenetic grid for the pelitic rocks, calcareous rocks (reaction 11) and mafic rocks (BA) (after Winkler, 1976)

E. Geothermometry and Geobarometry

1. Introduction

For the purpose of geothermometry and geobarometry, specimens of pelitic gneiss (HM 25), calc-silicate gneiss (HM 196), olivine gabbronorite (HM 7) and mafic granulite (HM 261) were analysed with a Cambridge Microscan 5 microprobe. The routine operating conditions for the microprobe have been described by Albat (1983b). The analyses are listed in Appendices I to III together with their collection locality. The sparse outcrop of lithologies that are suitable for geothermometry and geobarometry, and especially those which have not undergone retrograde metamorphism, renders it impossible to determine a regional distribution pattern of equilibrium P-T conditions for the whole study area. The specimen localities for the lithologies analysed are shown in Fig.7. For the purpose of this thesis the results from geothermometry and geobarometry are not considered to be statistically meaningful, but rather to indicate the approximate P-T conditions and are compared against the results obtained from metamorphic petrology.

Four different types of geothermometers have been applied in this study, namely:

a) Fe^{2+}/Mg partitioning between garnet and cordierite (Currie, 1971; Thompson, 1976; Holdaway and Lee, 1977)
b) Fe^{2+}/Mg partitioning between garnet and biotite (Ferry and Spear, 1978)
c) Fe^{2+}/Mg partitioning between clino and orthopyroxene (Wood and Banno, 1973; Wells, 1977)
d) Calcium partitioning between coexisting pyroxenes (Lindsley, 1983)

It was intended to use a fifth type involving Fe^{2+}/Mg partitioning between garnet and clinopyroxene for the calc-silicate specimen, HM 196, but the microprobe analysis of the garnet (Appendix III) reveals that it is highly calcic, being enriched in the andradite molecule, containing 31,79 weight per cent CaO and only 0,12 weight per cent MgO. The high CaO content would seriously effect the exchange equilibria between Fe^{2+} and MnO, and hence this geothermometer could not be used. The Fe^{2+}/Mg partitioning between garnet and cordierite mentioned above as a geothermometer has also been used as a geobarometer (Albat, 1983a) and is used in this study. The principles of equilibrium thermodynamics and its application to the above-mentioned geothermometers and geobarometers is discussed in detail by the authors cited above and shall not be repeated here.

2. Iron-Magnesium Distribution Coefficients

A serious drawback in the microprobe technique of analysis is that of the two valencies of iron, only total iron is detected, recorded as Fe^{2+}. As a consequence of this, minerals that have complex structure in which it is not clear how to calculate the quantity of Fe^{2+} from total iron, are calibrated according to the (total Fe)/Mg distribution coefficient. The garnet-biotite (Ferry and Spear, 1978) and the garnet-cordierite (Thompson, 1976; Holdaway and Lee, 1977) distribution coefficients are therefore calibrated on this basis and it follows that even though Ryburn et al.(1975) have devised a method of calculating the Fe^{2+} content in garnet from total iron on a charge basis, total iron should still be used since otherwise the distribution coefficients would need recalibrating.

Pyroxene has been calibrated on an Fe^{2+}/Mg basis (Wood and Banno, 1973; Wells, 1977) as the mineral has a relatively simple stoichiometry and in order to use the geothermometers of Wood and Banno (op.cit.); and Wells (op.cit.) Fe^{2+} has been calculated from total iron on a charge basis (Ryburn et al., op.cit.). The method of calculation (ibid.) involves molecular proportions based on 12 oxygens, whereby:

$$Fe^{3+} = 4 - 2Si - 2Ti - Al - Cr + Na + K$$

which allows Fe^{2+} to be determined, since

$$Fe^{2+} = Fe(total) - Fe^{3+}.$$

Allocation of the cations to their structural sites is done according to Wood and Banno (1973) and Powell (1978). The tetrahedral site is filled with aluminium and if there is a deficiency of aluminium then it is filled with ferric iron. Sodium, calcium and manganese are placed on the M2 site, while chromium, titanium and any remaining aluminium and ferric iron are placed on the M1 site. The distribution of Fe^{2+} and Mg between M1 and M2 sites is assumed to be random, so that (Wood and Banno, 1973, equ.25):

$$(Mg/Mg + Fe^{2+})M1 = (Mg/Mg + Fe^{2+})M2 = (Mg/Mg + Fe^{2+})pyrox.$$

3. Results

The iron-magnesium distribution coefficient (Kd) and the results obtained for the various geothermometers and geobarometers are listed in Table 17. The results for each lithology are now discussed in turn.

(a) Pelitic Gneiss - The equilibrium temperature given by the garnet-cordierite geothermometer (Currie, 1971) is more than 100°C higher than for any other geothermometer used for the pelitic gneiss. This geothermometer can be ignored since the method is not particularly sensitive for garnets which have an Fe/Fe + Mg ratio that exceeds 0,7 (Currie, 1971). The two garnets probed, HM 25(i) and HM 25(ii) have ratios of 0,85 and 0,87

Table 17: Results of geothermometry and geobarometry

P	METHOD	SAMPLE	Kd	T/°C	Er	P/kb
E						
L	Garnet-Cordierite					
I	Holdaway & Lee	HM 25 (i)	,1174	[631]4,5kb	50	
T	(1977)	HM 25(ii)	,0966	[584]4,5kb	50	
I	Thompson	HM 25 (i)	8,5147	[646]4,5kb	50	
C	(1976)	HM 25(ii)	10,3509	[591]4,5kb	50	
	Albat	HM 25 (i)	1,2601			[4,52-5,58]632°C#
G	(1983a)	HM 25(ii)	1,3639			[5,41-5,98]554°C#
N	Currie	HM 25 (i)	8,5139	795	50	
E	(1971)	HM 25(ii)	10,3509	846	50	
I	Garnet-Biotite					
S	Ferry & Spear	HM 25 (i)	,2344	[682]4,5kb	50	
S	(1978)	HM 25(ii)	,1997	[618]4,5kb	50	
-						
G	Two Pyroxene					
A	Wells	HM 7 (i)	,1576	991	70	
B	(1977)	HM 7(ii)	,1468	963	70	
B		HM 7(ii)*	,1409	771	70	
R	Wood and Banno	HM 7 (i)	,1576	985	60	
O	(1973)	HM 7(ii)	,1468	953	60	
N		HM 7(ii)*	,1409	745	60	
O	Lindsley	HM 7 (i)	(cpx)	[990]5,0kb	50	
R	(1983)	HM 7 (i)	(opx)	[700]5,0kb	100	
I		HM 7(ii)	(cpx)	[890]5,0kb	50	
T		HM 7(ii)	(opx)	[1100]5,0kb	100	
E		HM 7(ii)*	(opx)	[>1100]5,0kb	-	
-						
G	Two Pyroxene					
R	Wells	HM261 (i)	,2180	966	70	
A	(1977)	HM261(ii)	,1374	865	70	
N	Wood and Banno	HM261 (i)	,2180	901	60	
U	(1973)	HM261(ii)	,1374	828	60	
L	Lindsley	HM261 (i)	(cpx)	[880]5,0kb	50	
I	(1983)	HM261 (i)	(opx)	[950]5,0kb	100	
T		HM261(ii)	(cpx)	[620]5,0kb	50	
E		HM261(ii)	(opx)	[750]5,0kb	100	

* orthopyroxene rim
(all other grains comprised cores)
\# second approximation of pressure, the first having been at 4,5 kb in the Holdaway and Lee (1977) method,
Er error in °C

respectively, being predominantly almandine. There is good agreement with the results obtained using the other geothermometers for garnet-cordierite and garnet-biotite mineral pairs, spanning the temperature range of 584 to 682°C. The garnet-cordierite geobarometer (Albat, 1983a) indicates equilibrium at a pressure somewhere between the extremes 4,52 kb (P H_2O = 0) and 5,98 kb (P H_2O = P total). The P-T conditions of 4,5 to 6,0 kb and 584 to 682°C are in reasonable agreement with those of 4,5 kb and 650°C obtained from the petrographic study.

(b) Olivine Gabbronorite - The olivine gabbronorite appears to reflect an igneous crystallisation temperature overprinted by sub-solidus re-equilibration. The igneous temperature is revealed from analyses of pyroxene cores, but could be affected by fine exsolution. The temperature obtained from iron-magnesium equilibria ranges from 953 to 991°C. The geothermometer based on calcium partitioning (Lindsley, 1983) gives temperatures of 890 to 990°C for clinopyroxene, which are similar to the above-mentioned results, but the temperatures for orthopyroxene are sporadic, possibly as a result of the low calcium content of the mineral, and also as a result of the close spacing of isotherms in the model, and are therefore ignored. The orthopyroxene rim, based on iron-magnesium partitioning with clinopyroxene gives a temperature of 745 to 771°C using the Wood and Banno (1973), and Wells (1977) methods, probably reflecting sub-solidus re-equilibrium at lower temperatures.

(c) Mafic Granulite - Temperatures obtained from Fe^{2+}/Mg partitioning in pyroxenes for the mafic granulite range from 828 to 966°C, and it is noted that in HM261(ii) using the calcium partitioning geothermometer (Lindsley, 1983) temperatures are considerably lower, namely 620°C for clinopyroxene and 750°C for orthopyroxene. This may reflect sub-solidus calcium migration, possibly during cooling, as indicated by the presence of a finely exsolved phase in the clinopyroxene.

F. Conclusions

It is proposed that for the igneous rocks two temperatures are recorded; an igneous temperature and a medium to high grade metamorphic temperature, the latter showing good correlation with the (peak ?) temperature of metamorphism recorded in the pelitic gneiss. These temperatures as well as a pressure obtained from the pelitic gneiss, are listed in Table 18 and their distribution is shown in Fig.7. The results agree reasonably well with those obtained from metamorphic petrology, namely that for prograde metamorphism, temperature was in the range of 625 to 675°C and that pressure was approximately 4,5 kb.

An interesting aspect is that Beukes (1973), Blignault et al. (1974) and Toogood (1976) have reported that higher metamorphic grades exist to the north of the Pofadder Lineament, with high grade and granulite high-grade mineral parageneses to the north and medium grade mineral parageneses to the south. The structural implications of this are dealt with in Chapter IV, but it is important to note that the difference in metamorphic grade across the lineament cannot be accounted for in terms of shearing as this was predominantly of strike-slip movement, and that contrary to expectation, the dip-slip component seems to be north-down. There are two possible explanations why the metamorphism is of higher grade to the north of the lineament that do not invoke shearing. Firstly, west-southwest-directed thrusting has been identified in the present study area, and since it pre-dates the shearing, the Pofadder Lineament may represent an older thrust with crustal uplift to the north, which was reactivated by strike-slip shearing. Alternatively, the difference in metamorphic grade could be accounted for in terms of contact metamorphism related to mafic intrusive bodies which have been shown by Moore (1981,Fig.1) to be concentrated within and to the immediate north of the Pofadder Lineament. The relatively high temperature and low pressure metamorphism in the area, namely 650°C and 4,5 kb, does favour the presence of a magmatic heat input. It is possible that mafic magmatism was localised along the northward-dipping Pofadder Lineament (M.K.Watkeys, pers.comm., 1983) and hence the outcrop surface north of the lineament would have had a greater heat input than to the south. It is noteworthy that the specimen of high grade pelitic gneiss collected south of the lineament, in proximity to the Nouzees Gabbronorite Suite (HM 31), indicates that the distribution of high grade metamorphic rocks could be closely related to mafic magmatism, and need not be restricted to the north of the lineament, at least on a local scale.

Finally, shearing which post-dates the thrusting and the mafic magmatism resulted in retrograde metamorphism, with temperatures probably having been less than 500°C and pressures less than 4 kb.

Table 18: Temperature ranges recorded from geothermometry which are postulated to represent crystallisation and sub-solidus re-equilibration due to prograde metamorphism. Pressure is also recorded from the pelitic gneiss. (Note that +- indicates a range and not an error.)

Lithology	Crystallisation Temperature (°C)	Sub-solidus (Metamorphism ?) Temperature (°C)	Pressure (Kb)
Pelitic Gneiss		635 +- 55	5,25 +- 1
Olivine Gabbronorite	970 +- 20	755 +- 15	
Mafic Granulite	900 +- 70	685 +- 65	

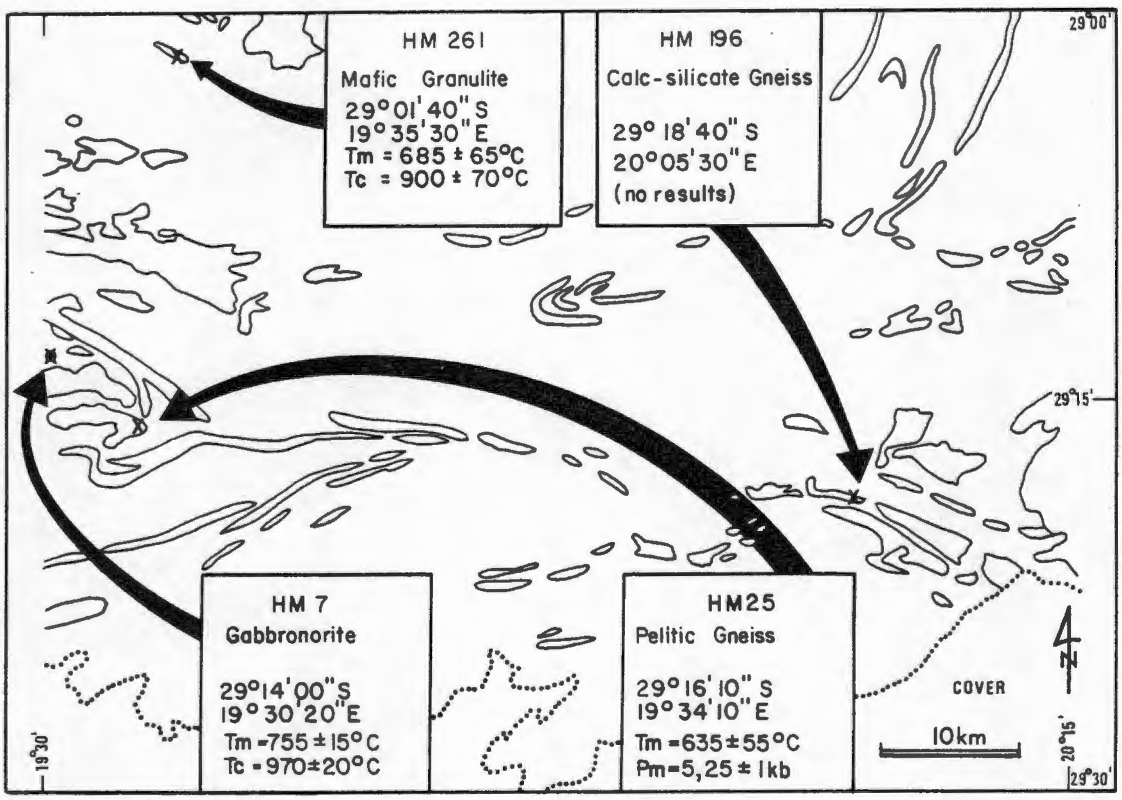


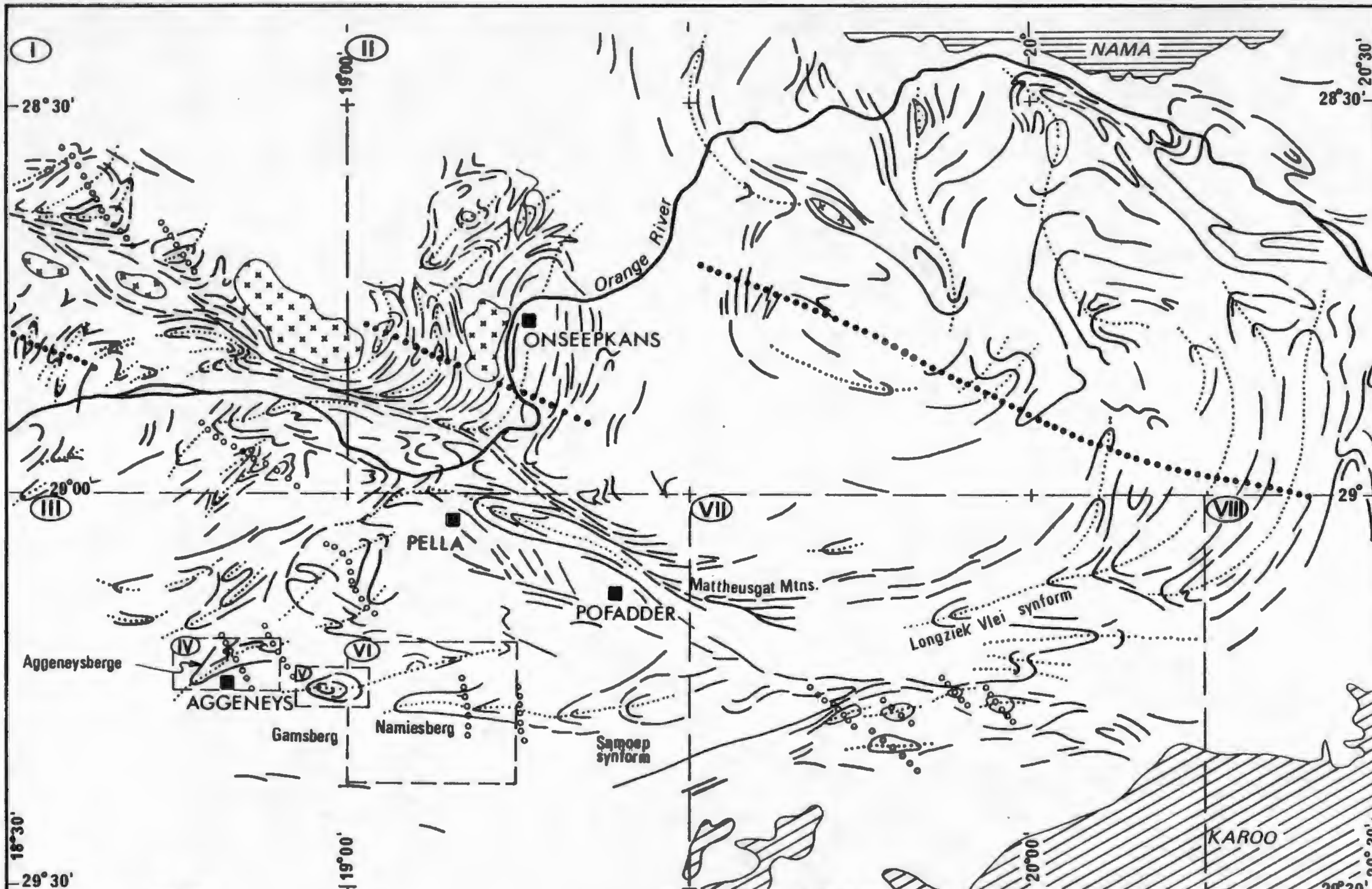
Fig.7: Locality map of specimens that were analysed by electron microprobe. Lithology of each specimen and their locality co-ordinates are listed. The general outcrop pattern is shown. Average metamorphic temperature (Tm), metamorphic pressure (Pm) and crystallisation temperature (Tc) obtained from Table 18 are also listed for each specimen.

Chapter IV

STRUCTURE

A. Introduction

The main structural trends for a part of the Namaqualand Metamorphic Complex are shown in Fig.8; an area that includes the study area and much of the Pofadder Lineament, bounded by longitudes 18°30'and 20°30'E and latitudes 28°30'and 29°30'S. The Pofadder Lineament is a major dextral shear, having been traced from Lüderitz (SWA/Namabia) to beyond Pofadder (Blignault 1974). Toogood (1976) has calculated that a minimum displacement of 85 km has taken place along the lineament. From this study the displacement is considered to be somewhat lower, approximating 50 km. Most of the area shown in Fig.8 has been mapped (Joubert,1973; Beukes,1973; Paizes,1975; Toogood,1976; Moore,1977; Lipson,1978 and Odling,1983) and mapping by R.W Harris in the east is in progress. As a result of all this work the regional geology has already been described in detail and the purpose of this thesis is to concentrate on the aspect of shearing in the study area and to describe the pattern of dextral and sinistral shears which seem to form a conjugate shear pair. The development of a mylonitic fabric has also been studied and is dealt with in Chapter V.



KEY

- Mafic Intrusives
- E.R.T.S. trends of the Namaqualand Metamorphic Complex
- Axial trace of F_{3a} fold ?
- Axial trace of F_{3b} fold
- Axial trace of F_4 fold.

AREAS MAPPED

I Beukes (1973)
II Toogood (1976) (north of river)
III Joubert (1972)
IV Lipson (1978)
V Odling (1983)
VI Moore (1977)
VII Maclaren (present study)
VIII Harris (in preparation)

Fig. 8: **LANDSAT Strike-line map for the environs of the study area and the Pofadder Lineament**

B. Theory on Shears and Their Brittle-state Analogues

Following Ramsay (1980) the localisation of intense deformation into narrow sub-parallel sided zones in the crust has been loosely termed shear zones by geologists. Shears appear to be the dominant deformation mode whereby large masses of physically rather homogeneous rock can change shape under medium to high grades of metamorphism. In many regions, (ductile) shears appear to be deep-level counterparts of faults, thrusts or wrench zones which occur in the brittle mode of deformation.

The term shear has been used for ductile shears, ductile-brittle shears and brittle shears by Ramsay (op.cit.), but in this thesis the term will be reserved for the ductile type only with the brittle state analogues being referred to in the usual manner as thrusts, wrench zones, normal faults and reverse faults as these are more descriptive. In the study area the shears typically have steeply dipping shear planes and a predominant strike-slip sense of movement, which is analogous to wrench zones in the brittle state. It is therefore necessary to describe briefly the main features associated with wrench zones, how they grade into shears with increasing ductility, and in more detail the features of classic shears.

There is much literature that deals with strain theory, based mainly on an analysis of the following aspects:

(1)The variation in the orientation of foliation as a result of simple shear in a previously unstrained rock (Ramsay and Graham, 1970; Ramsay and Allison, 1979; Ramsay, 1980; and Simpson, 1981).
(2)The variation in the orientation of pre-existing bands such as dykes (Escher et al., 1975).
(3)The shapes of deformed ellipsoidal particles (Dunnet, 1969; Dunnet and Siddans, 1971).

None of these methods could be utilised in the study area and hence they shall not be discussed further.

1. Wrench Zones

Wrench zones are high-angle strike-slip faults of great extent (tens of kilometres) and their fault planes generally have dips that exceed 70°and are commonly vertical as a result of horizontal shear couples in the earth's crust (Wilcox et al.,1973). They are generally considered to be deep-seated structures, having any of the following characteristics (ibid.):

(a)En echelon folds - En echelon folds form early in the deformation process whilst conditions are ductile. This is followed by a combination of ductile and brittle processes until finally only fracturing and faulting occur.

(b)<u>Conjugate strike-slip fractures</u> - The conjugate pair is comprised of a synthetic (Riedel) and an antithetic (conjugate Riedel) pair of strike-slip fractures.

(c)<u>Main wrench fracture (P-fracture)</u> - P-fractures are symmetrical to the Riedel fractures with respect to the general direction of movement of the wrench zone.

(d)<u>Normal faults or tension joints (T-fractures)</u> - T-fractures are orientated parallel to the axis of maximum stress.

By examining the development of wrench zones on all scales from the microscopic to sub-continental (Tchalenko, 1970), and also by studying the development of fractures resulting from simple shear according to the Coulomb failure criterion (Tchalenko and Ambraseys, 1970), the following important features emerge (Fig.9):

(a)With increasing deformation the resistance to shearing also increases and just prior to peak strength or at peak strength, failure initiates at inclinations of $\emptyset/2$ (for Riedel fractures) and $90°- \emptyset/2$ (for conjugate Riedel fractures) with respect to the general direction of movement (d-direction). The peak angle of shearing resistance is $\emptyset$. The Riedels generally form at about 12° to the d-direction and hence a realistic value for $\emptyset$ is about 24°. The apex of the angle between the Riedels and the d-direction points against the relative direction of movement. Since the Riedels are at a low angle to the d-direction they accommodate most of the displacement whereas the conjugate Riedels, at a high angle, tend to deform passively and also to take on a sigmoidal shape. In some cases tension fractures are formed in place of, or in addition to, the Riedels at 45° to the d-direction.

(b)With continued deformation the resistance to shearing decreases and hence deformation tends toward direct shearing conditions. P-fractures develop, interconnecting the Riedels along a semi-continuous line inclined approximately symmetrical to the Riedels (at $-\emptyset/2$ from the d-direction).

(c)Lastly, the resistance to shear attains a stable value, and either one or several principal displacement fractures (D) develop and are orientated along the d-direction.

In profile some wrench zones have a "flower" structure (Harding and Lowell, 1979) expressed as an upward spreading fault zone whose elements commonly have reverse separations. In the case of the San Andreas fault zone (Sylvester and Smith, 1976), crustal shortening as well as a more predominant shearing has occurred in a zone separated from relatively undeformed rocks on either side by steep faults. These faults branch upward and flatten abruptly into low-angle, outward directed, thrusts. It is hypothesised (<u>ibid.</u>) that the confining pressure resulted in uplift and local outward directed thrusting. Field observations (<u>ibid.</u>) reveal that basement in the wrench zone is fractured and sheared pervasively indicating that the mechanism of basement deformation was significantly ductile.

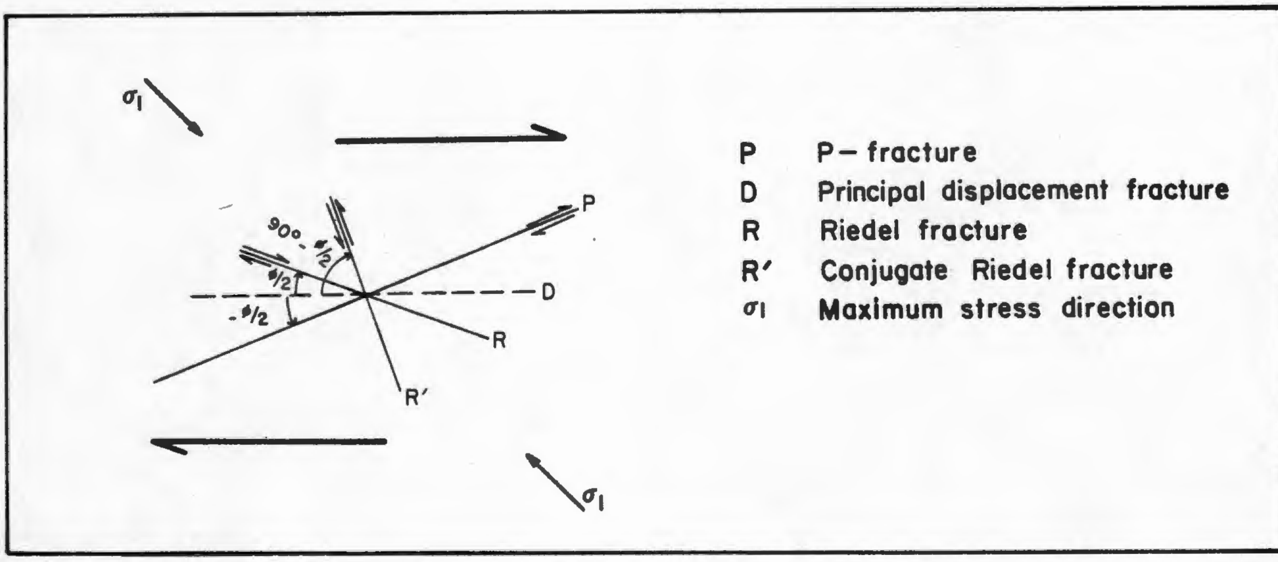


Fig.9: Mechanistic terminology for (brittle) simple shear structures according to the Coulomb failure criterion (Tchalenko and Ambraseys, 1970, Fig.2b).

2. Brittle to Ductile-state Transition

Little is known about the transition from a wrench zone to a shear with depth, or vice versa. Ramsay (1980) has observed that where shears pass upward into wrench zones they commonly exist as shears until they have passed the basement-cover unconformity. At higher levels the rheological contrast in the cover rock becomes increasingly apparent resulting in flexural-slip folding. Any shortening which is not taken up by folding is accommodated by low-angle thrusts that finally pass into strata-guided "glide-sheets" which ramp upwards.

In a contrasting manner, shears have also been postulated to steepen upwards with a reverse sense of movement and hence are analogous to thrusts at depth. Such a situation has been proposed for the Limpopo Mobile Belt (Coward,1980), but in contrast to thrusts, to have a predominant strike-slip sense of movement. The thrust component can result in antiformal bulges developing on the overthrust side (Coward,1983) due to either compression or to displacement of the hanging wall over a series of underlying flats and ramps.

3. Shears Zones

Shear zones have already been defined. An important concept regarding their ductile state of deformation is strain softening. According to White et al. (1980) shears develop when the hardening capacity, namely the ability of the rock to be stressed, has been exceeded. Their development marks the onset of a strain softening process (the reduction of stress at constant strain), which continues until a critical volume of mylonite is developed which wholly accommodates the imposed strain rate that the rock cannot accommodate by bulk deformation alone.

Ideally, the constraints for shears are that they have plane strain, simple shear, parallel margins and that the strain profile must be constant along their length (Ramsay and Allison, 1979). In reality however, shears tend to coalesce, enclosing lozenge shaped blocks of less deformed rock within which there is a tectonic fabric usually at a high angle to the shear (Coward,1976). In the Limpopo Mobile Belt the foliation changes orientation around such blocks, but the lineation maintains a constant trend. Similar observations have been made in the Grenville Province (Baer, 1977), in the Alps (Ramsay and Allison, 1979; Simpson, 1981) and in the Woodroffe "Thrust" which is a ductile mylonite zone in Australia (Bell,1978). There appear to be two reasons for the development of these lozenge-shaped blocks:

(a) The sigmoidal form and hence the coalescence of shears is favoured by a high rate of strain relative to the rate of propagation (Coward,1976).

(b) Conjugate shears may effectively enclose lozenge shaped blocks of relatively-less deformed material (Ramsay and Allison, 1979). One criterion that can be used to determine whether the shear sets are a conjugate pair is that the fabric should be continuous at the areas of intersection (Simpson ,1981). It has been postulated that for a brittle-state of deformation one set of a conjugate pair of fractures may appear to post-date the other set in one locality and not in another, due to a mechanism of alternating criss-cross faulting (Freund,1974). Such a mechanism is less likely for ductile deformation, but should be borne in mind. It should not be assumed a priori that one set post-dates the other.

Theory regarding the termination of shears is very pertinent to the present study, in particular with regard to the Pofadder Lineament. Strain and displacement variations for the ends of shears are mathematically very complex (Ramsay,1980) and either involve non-plane strain models or models which assume plane strain and a change in boundary conditions (Coward, 1976; Ramsay and Allison, 1979; Ramsay, 1980). Ramsay and Allison (op.cit.) have suggested that for plane strain the terminal displacement must be spread over an increasingly wider area until the strains are so low that no visible structure forms. If in this model the boundary conditions are constrained, then there will be a tendency for the tips of shears to bend and propagate; in a clockwise manner for dextral shears and in an anticlockwise manner for sinistral shears (Ramsay, 1980, Fig.18). The foliation may be unequally developed at the ends of the shear, being most strongly developed on the advancing side. In the non-plain strain model, constrictive and flattening deformations are set up on either side of the shear tip and all displacement takes place in the Y direction, normal to the shear (Ramsay and Allison, 1979, Fig.14). By analogy, for brittle deformation, Chinnery (1966) concluded from his analysis of secondary faults that displacement falls off quite rapidly near the ends of a master fault and also that the stress is not relieved at the ends to the same extent as over the length of the fault. This accentuates certain types of secondary faults, the most common being the "splay" or "horsetail" type which has also been described by Anderson (1951), McKinstry (1953), Freund (1974) and Casey (1980). It is important that these "splays" are commonly asymmetric and tend to extent themselves along curved paths, as in the San Andreas Fault Region (Chinnery, 1966). Evidence has therefore been given by Ramsay (1980) and Chinnery (op.cit.) that curvature at the ends of zones of movement is common to both shears and master faults. Simpson (1981) drew attention to the fact that these curved "splays" are developed in the Rosas Granodiorite which is associated with the Maggia Nappe, Switzerland; and not in the more ductile Maggia Nappe. Simpson (op.cit.) concluded that the "splays" tend to form in ductile-brittle rocks as opposed to ductile rocks. Thus, with decreasing depth (or at a later stage in deformation at the same depth) a truly ductile shear could transform into a wrench zone and where the ductility is sufficiently reduced, "splays" or other curved structures will develop.

C. Orientation of Fold Axes

The orientation of fold axes and the manner of folding is of importance in any structural analysis. In this study consideration is given to theory regarding (1) the angular relationships between folds and the bulk strain ellipsoid since this is of significance to folds in areas of high strain such as shears and (2) angular relationships of superimposed folding that can result in certain types of fold interference patterns.

(1) The bulk strain ellipsoid - It is generally agreed that for passive folding, where rheologic layering is not important, that fold axes tend to show a strong frequence maximum centred about the maximum principal axis of

the bulk strain ellipsoid. This is seen as a result of progressive rotation of linear elements toward the stretching direction (Borradail,1972; Sanderson,1973; Escher and Watterson,1974; Bell,1978; Cobbold and Quinquis,1980). Cobbold and Quinquis (op.cit.) have modelled the development of sheath folds for passive folding. At high shear strains most deflections become sheath like and highly asymmetrical.

Where, however, there is a strong rheologic layering, active folding occurs and the fold axial orientation is controlled by the orientation of the two-dimensional strain in the competent layers, with fold axes forming parallel to the maximum extension direction (Ramsay,1980).

Another aspect of folding worth mention is the development of a mylonitic fabric due to shearing that can become folded in the same event. If shearing is initiated in zones of weakness, the deformation must be inherently progressive; the regions of weakness are initially mylonitized and then the rock adjacent to them is affected, and so on. Hence, anastomosing early formed mylonite which is not parallel to the overall XY plane of the incremental strain ellipsoid, undergoes rotation during subsequent shearing. Determination of the sense of movement from the fold asymmetry requires care. Ramsay et al. (1983) have shown that during the initial stages of deformation, when shear strain is low, folds develop an asymmetry that is consistent with the applied shear. However, when the enveloping surface of the folds rotates into the extension field of the applied deformation the sense of asymmetry changes.

(2) Angular Relationships of Folds - Full details of fold interference patterns are given by Ramsay (1967) and, by Thiessen and Means (1980). For the purpose of this thesis only the angular relationships of folds that result in either Type 1 (dome-and-basin) or Type 3 structures developing will be discussed. This is done because both Type 1 and Type 3 structures are thought to exist in the present study area and hence an understanding of the angular relationships of the folds that cause them is required. In Type 1 the "a-direction" of the second generation folds must be contained within the axial plane of the first generation folds (Ramsay,op.cit.; Thiessen and Means,op.cit.). Since the "a-direction" is also contained within the axial plane of the second generation folds, it follows that the "a-direction" marks the intersection of the axial planes for both folds. Thus the folds must be commutative in order for Type 1 patterns to form (O'Driscoll,1964). From modelling (ibid.) the "a-direction" should, ideally, be vertical for the interference pattern to be developed on a horizontal plane, and where it does deviate from the vertical, en echelon chains of domes and basins are produced. The angle of the second fold axis with respect to the first is not specified, except that it cannot be parallel (Ramsay,op.cit.).

Type 3 fold interference patterns require that the "a-direction" of the second folds must be at a high angle to the axial plane of the first folds and that the trend of the second fold axis must be close to that of the first fold axis (ibid.). The two-dimensional pattern produced on outcrop surface does not show closed outcrop shapes because the periodic undulations of the first fold hinges are not well developed, but show continuously diverging or converging forms that are folded (ibid.).

D. Regional Structure

1. Description of Structures

The structural sequences which have been proposed by various workers in Bushmanland and their suggested correlation are shown in Table 19. The area under consideration is shown in Fig.8. The earliest known deformation stage (D1) is controversial, but has been reported by Joubert (1971a,b, 1974a) to have occurred in both Namaqualand and in the Pofadder area. The earliest known S-surface is possibly the original bedding (S0) or else a transposed bedding surface resulting from early deformation (D1?), and is thus termed S0/1. According to Joubert (1971b), this surface is deformed by F1 folds as intrafolial folds with sharp hinges, that predate the main stage of deformation (D2). From a study in the Aggeneysberge, Lipson (1978) has shown that the F1 folds are necessary in order to account for the duplicated succession that is found there.

The second stage (D2) was the main deformation and is distinctive, having been recognised by everybody who has mapped in the area. D2 serves as a structural benchmark to which other events can be related. The most characteristic feature of this deformation is the development of a regional fabric (S2) that is manifest as a gneissose mineral banding. Small to large scale, tight to isoclinal folds with rounded hinge zones formed and they commonly have an axial-planar foliation. It is only possible to distinguish F1 from F2 folds where the former occurs in the hinge zone of the latter since the regional foliation (S2) usually parallels the limbs and cuts the hinge zone in both generations. A conspicuous lineation is expressed by the orientation of mullion structures, and according to Joubert (1971b), also by the orientation of elongate minerals and mineral aggregates. Joubert (op.cit.) also states that it was during, or just after, the second stage of deformation that the highest grade of metamorphism was reached, whilst retrogressive metamorphism followed during the subsequent stages.

In the third stage of deformation (D3), large open folds (F3) developed which fold the regional foliation (S2) and which do not commonly have any axial-planar foliation. Two sets of folds, both having the above characteristics, are considered to be F3 folds, namely F3a and F3b. These two sets have also been identified by Lipson (1978) in the Aggeneysberge and also by Moore (1977) in the Namiesberg. Age relationships between these two sets of folds can only be determined where one is seen to deform the other (Lipson,op.cit.). F3b folds are much more common than the F3a folds and generally have east-northeast-striking axial surfaces, a gentle plunge in that direction and a monoclinal style. This set is commonly developed on a megascopic scale and is of the flexural-slip fold type. In the present study area only one set was observed and it has been correlated with the younger F3b set on account of its east-northeast trend. The F3a folds cannot, however, be omitted as it has been found that they are structurally necessary, as will be shown in Sections F.2.b., F.2.c. and F.4.d. These older folds are necessary in order that when they are cross-folded, a dome-and-basin fold interference pattern is produced. F3b synforms are common and along their steeper dipping

Table 19: A possible correlation of structures for the area shown in Fig. 8, as mapped by various workers

Present correlation	This study	Joubert (1971) (1974)	Beukes (1973)	Toogood (1976)	Moore (1977)	Lipson (1978)	Harris (pers. comm.)
DI Isoclinal, intrafolial folds seen in places to be deformed around F2 fold axes. FI folds pre-date the S2 foliation.	FI	FI SI		SI	FI	FI	
D2 Development of the regional foliation (S2) and tight ENE-trending folds (F2) for which the foliation is axial planar. L2 mineral lineations are common.	F2 S2 L2	F2 S2 L2	FI	S2 L2	F2 L2	F2 S2 L2	F2 S2 L2
D3 Meso to megascopic folding (F3) of the regional foliation (S2). F3a and F3b folds are only distinguishable where F3a occurs in an F3b closure. F3b folds generally trend ENE and cause fold interference patterns. F3a folding associated with thrusting.	F3a S3a L3a F3b* S3b	.... F3* S3 L3	.... F2* L2	... ? ... F3 S3 L3 F5?	F3 .... F4 S4~	F3a S3a F3b	F3 S3 .... F4*
D4 WNW to W-striking dextral shears and ENE to NE-striking sinistral shears, possibly being a conjugate pair. F4 drag folds associated with the shearing.	F4 S4~ L4	F4 S4~	S2~ F3*	F4 S4 L4 .... F6*	F5 S5~	F4*	F5 S5~ L5
D5 NW-SE - striking faults and Riedel fractures. Jointing developed.	S5					S5	

* interference fold pattern
~ mylonitic or sheared foliation
S foliation surface
L lineation
F fold

northern limbs, zones of refoliation have developed which with further deformation have become shears (Joubert,1971b). Mapping, and the plotting of structural data on stereonets has shown that within the folded surface (S2) of these folds, the L2 mineral lineation (or mineral aggregates) almost invariably parallels the axial orientation of the F3b folds. This parallelism probably reflects co-linear F3b folding about the L2 mineral lineations since the lineations would have formed a strong linear anisotropism in the rocks. It must be noted that in the structural analysis to follow, distinction will be made between the F3a and F3b fold generations, but that in a less specific sense, the D3 deformation stage will include both D3a and D3b.

As already mentioned, shearing (in its early stage of development) is related to F3b folding, and hence the fourth stage of deformation (D4), in which shearing controlled the mode of deformation, probably followed on from the D3 stage. The main direction of shearing is west-northwest with a dextral displacement and a steep dip to the northeast, as characterised by the Pofadder Lineament. A sinistral set of east-northeast striking shears is also developed, characterised by a sub-vertical dip. Both sets of shears are associated with retrograde metamorphism and evidence, which is to be discussed in the structural analysis, suggests that the shears form a conjugate pair.

Folds that deform the large scale F3b folds are monoclinal and appear to be associated with the D4 shearing stage of deformation, and hence are given the age of F4. The Longziek Vlei synform in the study area has, for example (see Fig.10), been "drag-folded" with the sense of drag being in sympathy with the dextral movement along the Pofadder Lineament. In the area mapped by Toogood (1976) along the lineament, to the northwest of the present study area, F4 folds (F6 according to Toogood,op.cit.) occur parallel to the lineament and appear to represent drag folds. The folds are crescent shaped and have been classified by Toogood (op.cit.) as the H-type (Type 2 interference folds) after Ramsay (1967). It is questionable whether they represent Type 2 as movement along the lineament was predominantly strike-slip whereas a steep "a-direction" of shear would be required to form these interference patterns. It is more likely that these crescent shaped interference patterns represent drag-folding of already existing (possible dome-and-basin) interference patterns that could have formed due to the superimposition of F3b folds on an older fold generation. Evidence to support this is that both Beukes (1973) and Toogood (1976) have reported that there is a disparity of the structure on either side of the lineament, with F3b fold axes being at a higher angle on the northern side and that fold interference structures are also developed on that side. If the interference structures were to have formed during shearing, then they would be expected to have formed on both sides of the lineament.

In the last deformation stage (D5), fractures and faults developed with a general northwest strike. These structures tend to branch out of the Pofadder Lineament and in places cut across it (Joubert,1974a). Prominent quartz veins follow the faults in places.

2. The Pofadder Lineament

The Pofadder Lineament (Joubert,1974c) is probably the largest discrete shear in southern Africa and is best exposed between Warmbad in S.W.A./Namibia and Pofadder in South Africa. It has also been termed the Tantalite Valley Megaskuifskeurzone, (Beukes,1973), the Pilgrim Lineament (Toogood,1974), and the Tantalite Valley Mylonite Belt (Blignault et al.,1974). The lineament attains a maximum width of 7 km at Tantalite Valley, S.W.A./Namibia (Moore,1976). Its northwesterly extension is poorly exposed, but Blignault et al. (1974), and Jackson (1976) have shown that it extends as far as Lüderitz. The lineament thus has a length in excess of 450 km and Joubert (1974c) has indicated that it probably extends southeast of Pofadder as far as Vanwyksvlei, which means that the total length may be up to 800 km. There is not much sign of shearing to the southeast of Pofadder, but the following evidence does, however, indicate that the lineament probably continues as far to the southeast as Vanwyksvlei:

(a) A 60 km linear zone of superficial deposits lying on Karoo Sequence rocks is shown to the northeast of Vanwyksvlei on the official 1:1 000 000 Geological Map of South Africa, 1970 edition (Joubert 1974c). This zone is parallel to the west-northwest shearing in Namaqualand and lies almost in line with the extension of the Pofadder Lineament (ibid.).

(b) Many large mafic and ultramafic bodies as well as numerous small mafic intrusives have been emplaced along the Pofadder Lineament. Mafic bodies still appear for some distance along the southeastly extrapolation of the lineament (Joubert,op.cit.).

(c) The epicentres of earth tremors that occurred during 1979 and 1980 (Fernández, 1983a, 1983b) show that seismic activity still exists along the southeasterly extrapolation of the Pofadder Lineament (Fig.22).

An important feature in regard to the Pofadder Lineament is that higher grades of metamorphism are encountered to the north of it, with high grade and granulite high-grade metamorphic mineral parageneses occurring to the north whilst medium grade mineral parageneses occur to the south (Beukes,1973; Blignault et al.,1974; Toogood,1976). From a study of mylonitic lineations within the lineament, however, Toogood (op.cit., Fig.52, and p.132) has indicated that the shearing was predominantly strike-slip, and hence the higher grade of metamorphism to the north cannot be accounted for in terms of shearing. In this regard Toogood (op.cit.) proposed that crustal uplift occurred to the north of the lineament in only the final stages of movement, post-dating the strike-slip shearing. In the present study area there is no evidence for this, and it seems more likely that the uplift was the result of west-southwest - directed thrusting which has been identified in the area and which pre-dates the shearing. It is proposed that the lineament could represent an older thrust which accounts for the difference in metamorphic grade from north to south, and that it was reactivated by strike-slip shearing. The main problem with this idea is that it is unlikely for a thrust to be laterally contunuous over the entire length of the lineament. An alternative explanation of the higher metamorphic grade to the north is that contact metamorphism may have been caused by the intrusion of mafic bodies which occur predominantly within and to the north of the lineament. This concept has already been discussed in Chapter III (Section F) and need not be repeated here.

A second important feature regarding the lineament is that the structural pattern is reported to differ on either side (Beukes,1973; Blignault,1974; Toogood,1976). The structural pattern developed to the north is that of a Type 2 fold interference pattern (Toogood,op.cit.), and it is developed on a megascopic scale. In the immediate south of the lineament this interference pattern is not recognised and Toogood (op.cit.) postulates its apparent absence as being due to:

(a)Lower P-T conditions to the south of the Pofadder Lineament.
(b)The Vioolsdrif Granite to the south of the Pofadder Lineament, which probably acted as a rigid body during shearing and hence created a pressure shadow that shielded a large area south of the lineament and east of the granite during this deformation.

Toogood (1976) has calculated the displacement on the Pofadder Lineament to be as follows:

(a)100 km provided that the strain involved 100 per cent simple shear
(b)90 km if a 20 per cent component of pure shear to the simple shear existed
(c)85 km if the pure shear component was as large as 40 per cent.

The calculations appear to be somewhat unreliable because of the following assumptions:

(a)The F3b folds (classified as F5 by Toogood,op.cit.) were initiated at the start of shearing
(b)The axial planes of these F3b folds:
  i)are parallel to the b-axis of shearing for the Pofadder Lineament and plunge steeply northwards
  ii)develop at 45° to the a-axis of shearing for the Pofadder Lineament. The a-axis is approximately horizontal, as deduced from displacement patterns.

Using assumption b(ii) above, Toogood (op.cit.) was able to apply the Ramsay and Graham (1970) model for strain analysis since the model considers the rotation of a material line which is initially at 45° to the shear plane. A main shortcoming of the approach, however, is that because not many direct measurements of the axial surfaces were possible in the field, Toogood (op.cit.) resorted to using foliation measurements from a traverse across the lineamemt which avoided hinge zones. Such an approach obviously must assume that the foliation parallels the axial surfaces and since the area has been so complexly folded the assumption seems too simplistic. The estimate of an 85 km displacement is, however, not unrealistic for a shear zone of approximately 800 km in length, but it will be demonstated (Section F.3.e.) that the displacement was probably somewhat lower, approximating 50 km. This estimate is done on the basis of correlating a synform in the vicinity of Pella on the west of the lineament with the Longziek Vlei synform to the east.

3. Structural Areas

The basic structural features of the area are shown in Fig.10. A hierarchical system of analysis is used which allows structural areas to be compared and described at all scales. From largest to smallest scale the hierarchy constitutes structural provinces, sub-provinces and regions. In this context, the term Namaqua Province constitutes the rocks collectively referred to as the Namaqua Metamorphic Complex (Kröner and Blignault, 1976). The province thus wholly includes the present study area and is therefore not useful in this study. The Pofadder Lineament is considered to separate the province into the Gordonia Sub-province to the northeast and the Bushmanland Sub-province to the southwest, the latter having lower metamorphic grade (P.Joubert, pers.comm., 1984). Next in the hierarchy, the sub-provinces are subdivided into regions. The various regions within any given sub-province are considered to be of similar metamorphic grade, but to have deformed as separate structural units. Wherever possible, boundaries to the regions are chosen such that they occur along shear zones, since if they are considered to have deformed as separate units, then any strain difference would be localised into shears. As an additional criterion, the structural regions are considered to be divided along contrasts in lithological associations wherever possible. The regions are not intended to define a homogeneous fabric as would be required by the use of "domains" (Turner and Weiss, 1963), but rather to reflect a geometric pattern, such that when a fabric element is plotted onto a stereonet, the pattern is descriptive of the type of structure concerned. The various regions are shown in Fig.11. Finally, it is stressed that the present study area is a critical area of Bushmanland in which two sub-provinces and several structural regions exist. The choice of these sub-provinces and regions has been made such that they are compatible with the regional divisions of the Namaqualand Metamorphic Complex (Joubert,1984; Stowe,1984).

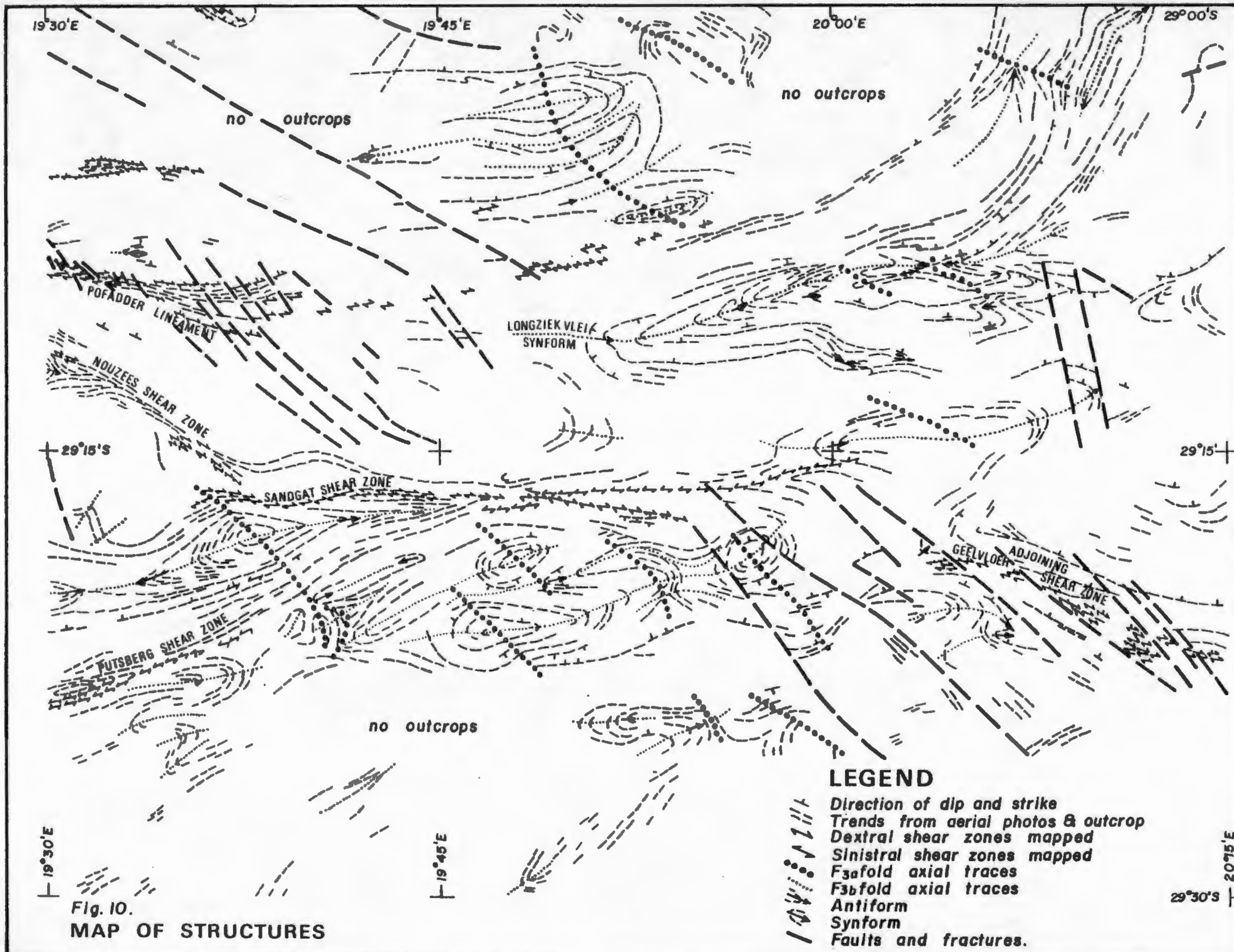


Fig. 10.
MAP OF STRUCTURES

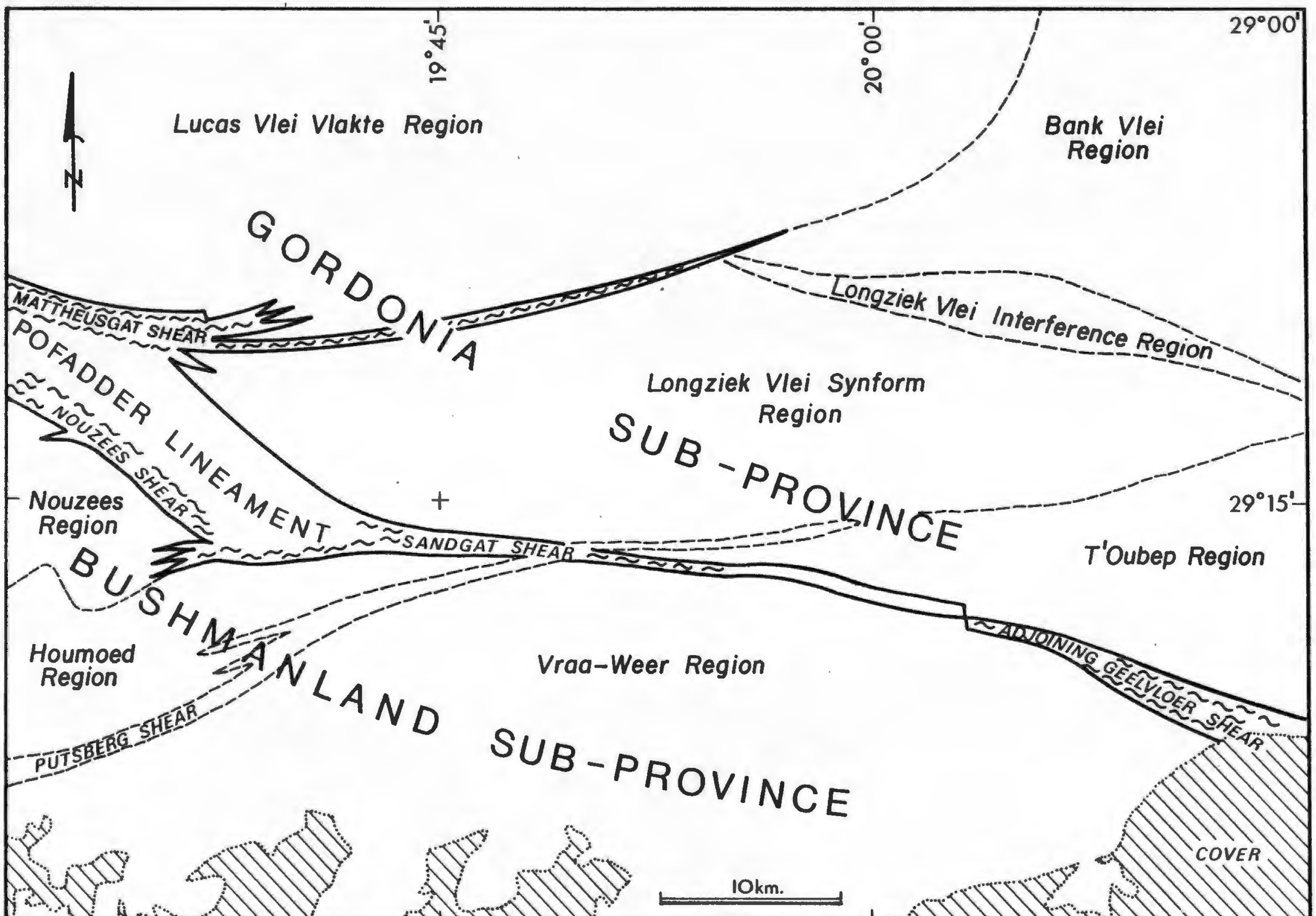


Fig.11 : MAP OF STRUCTURAL AREAS USED IN THE STRUCTURAL ANALYSIS

E. The Deformation Mode

1. Introduction

The regional structure and the proposed stages of deformation have already been described. In this section the various criteria that were used for the classification of the structures are recorded, and significant aspects of the structures are discussed with regard to the process of deformation.

2. Classification of Structures

Following the description of the regional structure (Section D.1.), the classification of structures according to their order of development relied on three main aspects. Firstly, does the foliation in the structure correspond to the regional foliation (S2) or to a surface of refoliation (S4) due to shearing? (Note that the S3 foliation is very uncommon, occurring only in some F3b folds within the hinge zones.). Secondly, does the development of that structure pre, syn, or post-date the above-determined foliation? Thirdly, is the deformation ductile or brittle? In terms of the above-mentioned aspects two benchmarks can be used to classify the structure, namely the age of the foliation (D2 or D4) and the brittle fractures or faults (D5) which the ductile structures must pre-date. It follows that:

(a)If the foliation is S2 in age, then

i)D1 structures pre-date the foliation
ii)D2 structures are synchronous with the foliation
iii)D3 (a and b) and D4 structures post-date the foliation, and hence the foliation is deformed in these structures, but that they pre-date the brittle structures (D5).

In order to distinguish F1 folds from F2 folds (see Section D.1.) for which the latter, and commonly the former fold generations have limbs that are parallel to the foliation and hinge zones that are cut by it (in the case of F2 folds the foliation must be axial-planar), the F1 fold must be observed to be deformed around the hinge zone of the F2 fold.

F3a, F3b and F4 folds all deform the S2 foliation and since F3b and F4 (and possibly F3a) folds are usually monoclinal, it is difficult to distinguish them. The only practical way of distinguishing them is to assume that the east-northeast trending folds (for which the S2 foliation is deformed) are of the F3b generation. As a consequence, F3a folds can only be identified if they deform the S2 foliation and are seen to have been deformed around the hinge of an F3b fold. Similarly, an F4 fold can only be identified if it is seen to be younger than the F3b folds by having deformed the latter about its hinge zone. The F4 folds are typically associated with drag-folding along shear zones and tend to be more gentle than the F3b folds.

(b)If the foliation is of S4 age, then

i)D1, D2 and D3 (a and b) structures pre-date the S4 surface and can only be distinguished if the S2 surface has been preserved, by using the criteria mentioned above in case (a).

ii)D4 structures post-date the F3b folds, are synchronous with the development of shears (S4) and pre-date the brittle structures.

The identification of S4 on the basis of whether or not it is a refoliated surface by being mylonitic, and thus distinguishing it from S2, is not ideal. (Note that S3 is very uncommon, occurring as an axial-planar foliation in the hinge zones of some F3b folds.) The problem lies in when to distinguish between a non-mylonitic and a mylonitic foliation since there is in many cases a continuous spectrum between the two types of foliation. In general, however, it is safe to say that the refoliated areas or shears have a flaggy appearance in the field, a steep dip and a well-developed mineral or mineral aggregate lineation (L4).

3. Deformation Processes

It is necessary to record some basic criteria that are indicative of the processes of deformation, and the number of processes required for the total deformation. The most important aspects to consider are:

(a)The development of the regional foliation (S2) is considered to be a separate process from that of D3 in which the S2 foliation acted as a plane of weakness facilitating slip along the foliation, and in particular flexural-slip folding.

(b)The F3b folds generally have a gentle plunge to the east-northeast, steeper dipping limbs to the north of synforms, and these limbs are commonly refoliated into shears. The significance of this is that at some stage, movement on S2 surfaces must have changed from having a large component of dip-slip (since the folds have gentle plunge) with a small component of sinistral strike-slip to one having a large component of dextral strike-slip. This change would take place when flexural-slip folding gave way to dextral shearing.

(c)The Samoep synform which is an F3b fold, is truncated by the Nouzees shear which substantiates point (b), in that shearing post-dates F3b folding, even though the shearing may have followed on from the F3b folding.

In the light of the above it is postulated that at least three different deformation processes operated during the deformation history, although there need not necessarily have been a change in the stress field. Thus, as a working model, the structural synthesis of the present study area which follows is based on the folowing hypothesis:

(a)The first phase of deformation involved tight to isoclinal folding of the S0/1 surface, probably forming two generations of folds, namely F1 and F2, of which it was during the second that an axial-planar foliation (S2) developed, roughly coincident with the peak of metamorphism.

(b)The second phase of deformation utilised the S2 foliation, involving flexural-slip folding. Two generations of F3 folds probably occurred, namely F3a and F3b. Little is known about the first set (F3a) but it does appear to be structurally necessary, for reasons which are explained in Sections F.2.b., F.2.c. and F.4.d. F3b folds are common and tend to be large-scale, east-northeast trending, shallow plunging monoclinal folds.

(c)As a possible continuation of the F3b folding, the northern limbs of the F3b synforms became refoliated in many cases, developing into shears. Once shearing intensified, however, the process probably changed from one of flexural-slip folding to one of shearing and drag folding. The D4 stage of deformation was therefore probably controlled by the process of shearing and is regarded as being different from D3, although it probably did follow on from it. This difference is illustrated by the fact that the F3b Samoep synform is truncated by the Nouzees shear.

(d)It is possible that the same stress field which resulted in the shearing (D4) may have resulted in the brittle deformation during (D5), but this is not certain and will be considered in the light of the structural synthesis for the present study area.

F. Ductile Structure Analysis

1. Introduction

The ductile structures discussed here, are considered to have behaved differently in each structural area, whereas the brittle structures, which are discussed later are considered to have behaved the same over the whole study area. In the structural analysis a map showing the locality of each structural measurement was compiled and then the boundaries to each structural region or shear zone were drawn in to subdivide the data. Thus for each structural area (shear zones or regions) the relevant data can be used to plot stereonets that are area defined. The type of data used comprises.

(a)The D2 fabric elements. This includes the S2 foliation, L2 mineral lineations and F2 fold axes, of which the last has not been plotted since too few readings could be taken. For convenience the mean east-northeast - trending fold axis (F3b) deforming the S2 surfaces is also shown with these D2 elements.

(b)D3 structures, namely, F3b fold axes that were measured in the field. Note that no F3a folds were observed, but they are presumed to exist.

(c)S4 fabric elements which include a mylonitic foliation (S4) and a mylonitic mineral lineation (L4).

To set the data out in the above-mentioned manner, stereonets have been constructed to represent data for D2, D3 and D4 fabric elements separately. To be most informative, these stereonets are arranged into matrices such that D2, D3 and D4 fabric elements appear in different rows and that stereonets for different structural areas appear in different columns. In this arrangement it is easy to compare the same fabric elements for different structural areas or to compare different fabric elements for the same structural area.

As a consequence of this rigorous approach the data has been greatly subdivided, with the result that the sample size for any stereonet is significantly reduced. This has had two consequences, firstly, a given fabric will not be represented unless there are more than five readings and secondly, the structure will be treated qualitatively. The stereonets have not been contoured because according to Stauffer (1966,p.490) more than 400 points are required for samples from weak concentrations, otherwise the contour pattern will be erratic and may reflect a sampling variation. It follows that unless a sample population is greater than 400 an empirical study of symmetry and grouping is the best way of determining preferred orientations and hence this approach will be adopted here.

Prior to embarking on the structural analysis of ductile deformation for each structural area, it is beneficial to have an overview of the area as a whole. In this regard Fig.12 shows the L2 and L4 mineral lineation trends and their angles of plunge. The following points are noteworthy:

(a)Both L2 and L4 mineral lineations have a general northeast trend and are sub-parallel to the F3b fold axes.
(b)The Mattheusgat shear (see also Fig.11) has shallow-plunging L4 mineral lineations whilst the Nouzees, Sandgat and Adjoining Geelvloer shears have steeper plunging L4 lineations.

It is therefore likely that there is a relationship between L2 and L4 mineral lineations which accounts for their similar orientation. Also, within the Pofadder Lineament, the Mattheusat shear is considered to be a zone of predominantly strike-slip movement, if the lineations are assumed to parallel the "a-direction" of shearing, and that there is a greater component of dip-slip along the Nouzees, Sandgat and Adjoining Geelvloer shears.

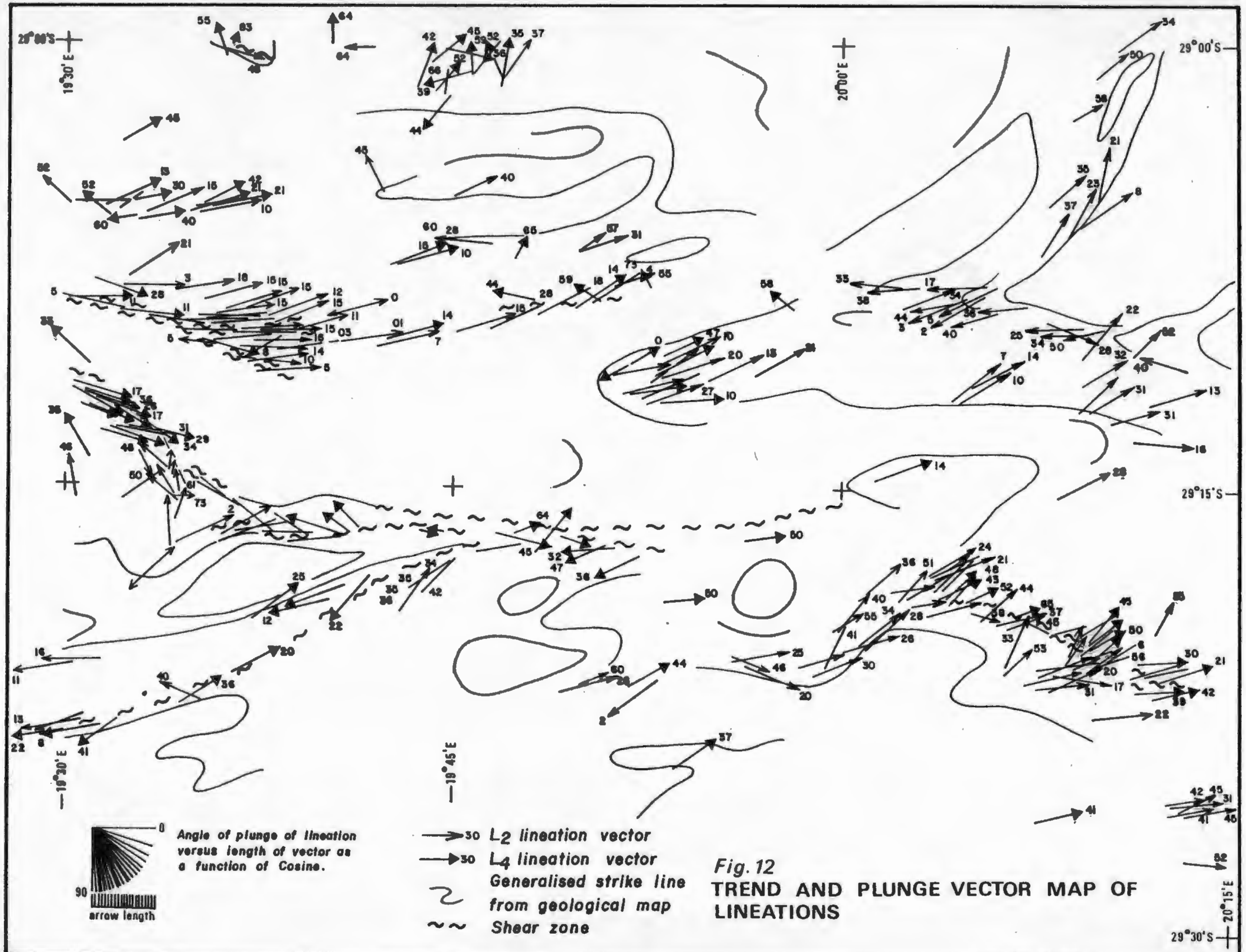


Fig. 12
TREND AND PLUNGE VECTOR MAP OF LINEATIONS

2. Gordonia Sub-province

(a) Introduction - The Gordonia Sub-province is separated from the Bushmanland Sub-province in the south by the Pofadder Lineament. The reason for separating the two sub-provinces is that the lineament is a regional feature. The lineament is thought to be up to 800 km in length, with a significant displacement that is possibly as much as 85 km (Toogood,1976), although a more conservative estimation is 50 km (see Section F.3.e.). Rocks north of the lineament (the Gordonia Sub-province) have been reported to be of higher metamorphic grade than to the south as a result of uplift, but the reason for this is controversial and has been dealt with in Section D.2. and Chapter III (Section F). It is stressed that there is not any apparent difference in the lithologies of the Gordonia Sub-province as compared to those of the Bushmanland Sub-province, but that they have merely been displaced by the lineament, by a movement that was predominantly strike-slip.

(b) Lucas Vlei Vlakte Region - The Lucas Vlei Vlakte Region is separated from the Longziek Vlei Synform Region by a listric shear zone that splays off from the Pofadder Lineament, following the northern limb of the synform. In contrast to the major F3b folding in the Longziek Vlei Synform Region, this region is characterised by fold interference patterns which can clearly be seen from strike lines taken from aerial photographs (Fig.10). In order for these interference patterns, which approximate dome-and-basin structures (Type 1 after Ramsay,1967), to have developed it is necessary that:

i) both fold generations must post-date the development of S2, if S2 is assumed to define the interference patterns.

ii) in order to have a dome-and-basin interference pattern developed on the horizontal plane (plan view of the map) it is necessary that the "a-direction" of movement (ibid.) be steeply inclined. This negates F4 folding as being the cross-folding event since F4 folds are associated with strike-slip shearing which would have a shallow plunging "a-direction".

The two fold generations therefore occurred between the F2 and F4 fold generations, of which one is correlated with the large-scale east-northeast-trending F3b folds. On a mesoscopic scale these interference fold patterns are not common, having been observed in one locality only (29°01' 31"S/19°43'39"E). The outcrop (Fig.13, Plate 6) is comprised of megacrystic alkali-feldspar granitic gneiss. Within the rock are one to two cm wide feldspar-rich bands which form enclosed structures of oval shape in plan view with steeply dipping surfaces. The megacrysts appear to be prolate in shape and to define a planar foliation, considered to be the XY plane of the bulk strain ellipsoid. The X-direction is considered to be the direction of elongation of the megacrysts. The mineral banding could either be the earliest S0/1 surface or the S2 surface formed during peak metamorphism. The structure appears to have resulted from the interference of two fold generations, and then the interference pattern itself seems to have been folded. Concerning the two fold events that resulted in the interference pattern it is noteworthy that:

i) The younger fold generation would have an axial trace sub-parallel to the megacrysts since the fabric they define has not been folded to form the interference pattern.

ii) The younger fold generation must be of F3b age and not F4 since the latter folds have a shallow plunging "a-direction" of movement (as a consequence of their association with strike-slip shearing) and thus cannot result in dome-and-basin (Type 1) structures being exposed on a sub-horizontal plane (Ramsay, 1967) as is the case here.

It is therefore hypothesised that:

i) If the mineral banding is S0/1, then the banding would most likely have been subject to F2 folding and to have been cross-folded by either F3a or F3b folds.

ii) If the mineral banding is S2 then the two fold generations must post-date the F2 fold event, involving both F3a and F3b folds.

The former case does not seem likely because, if the surface was S0/1 then an S2 foliation should be present, axial-planar to the older fold (F2) axis, and this is not the case. It is therefore concluded that the structure is the result of two F3 fold events, and that the F3b folds trend approximately east-northeast, with the F3a fold axes being at some angle to them.

Stereonets for D2, D3 and D4 fabrics are shown in Fig.14. In the field the majority of the foliations are of the S2 generation, and when plotted on a stereonet (Fig.14, D2 fabric) their distribution defines an F3b fold axis plunging at 10° towards 078°. This direction is sub-parallel to the L2 mineral lineations and to many of the F3b fold axes that were measured in the field (Fig.14, D3 fabrics). This probably indicates that the L2 mineral lineations resulted in a strong linear anisotropism in the rock which induced the subsequent F3b folding to be co-linear.

With respect to the D4 structures, a dextral set of shears with strike varying from northwest to west, and a sinistral set with a northeast strike have developed, although the latter set is not as strongly developed as the former. The dextral set is observed in the following four localities:

i) On the farm Gemsbok Vlakte (locality 29°00'S/19°36'E) a northwest-striking shear is exposed for a distance of about 500 m and it is comprised of calc-silicate gneiss, leucogneiss and blastomylonite with a core of mylonite and quartz veins. The leucogneiss is isoclinally folded (F3b?) with Z-asymmetric bands of feldspar and biotite (Plate 4), with garnet porphyroblasts in places. The orientation of the biotite flakes defines the mylonitic foliation which is axial-planar to the folds.

ii) A possible extension of the above-mentioned shear occurs on the farm Lucas Vlei Vlakte (locality 29°01'15"S/19°43' 00"E). The rock is a mylonitic leucogneiss (HM 125) with a northwest strike and the mylonite is refoliated into narrow bands of up to half a metre in width with a north-northeast strike, commonly defined by blastomylonite. Deflection of the older fabric into the refoliated bands is consistent with sinistral

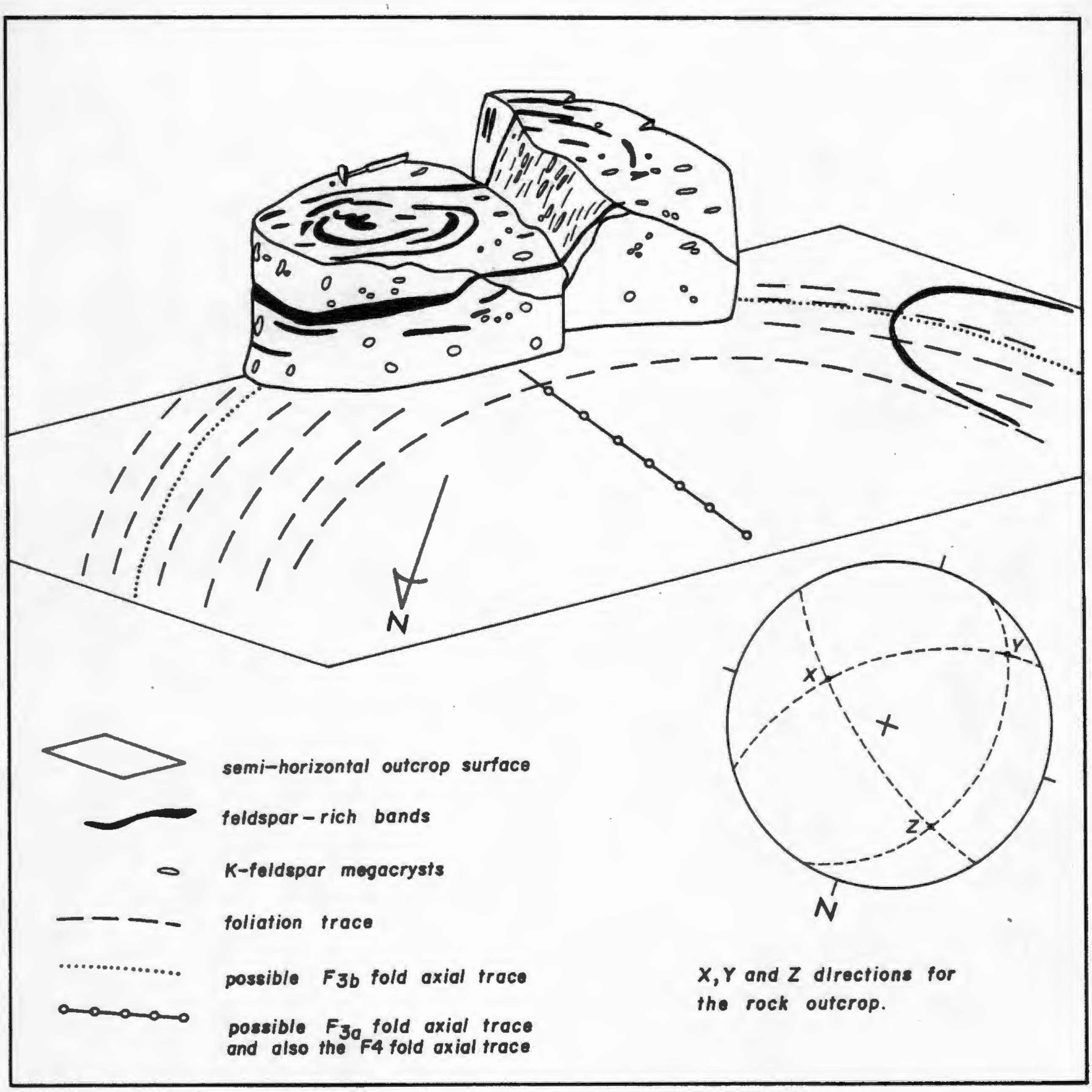


Fig.13: Sketch of outcrop features (Plate 6) in megacrystic alkali-feldspar granitic gneiss, showing evidence of possible interference folding. X,Y and Z strain axes determined from megacrystic shape (locality 29°01'31"S/19°43'39"E).

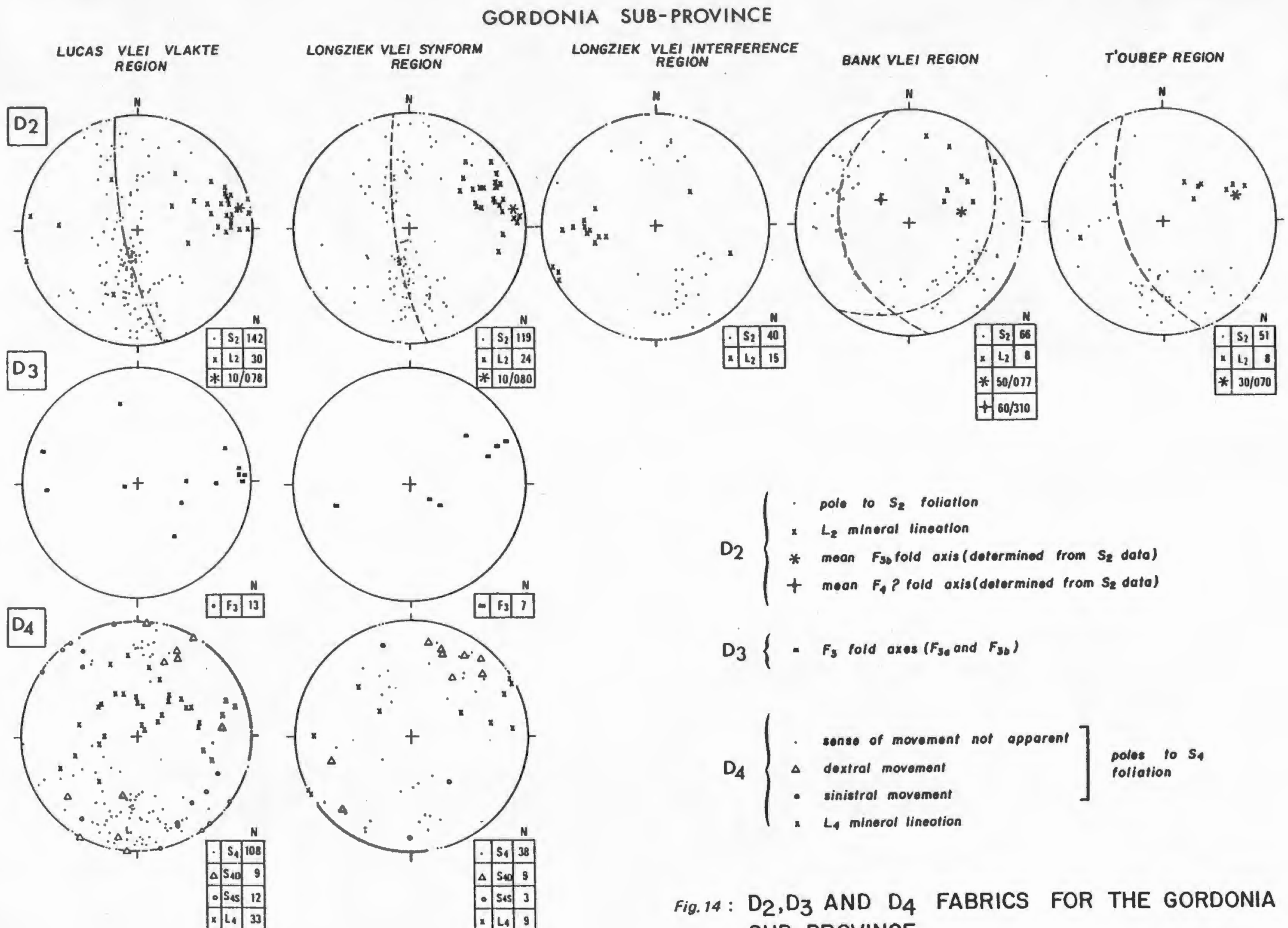


Fig. 14 : D_2, D_3 AND D_4 FABRICS FOR THE GORDONIA SUB-PROVINCE.

movement and indicates that sinistral shearing post-dated the dextral shearing, at least at this outcrop.

iii) On the farm Pofadder East (locality 29°05'45"S/19°30'00" - 19°35'30"E) a band of strongly mylonitised quartzite crops out which can be followed further west into the main part of the Pofadder Lineament (Joubert,1973).

iv) Extending from the farm Mattheusgat to Lucas Vlei Vlakte (locality 29°07'30"S/19°42'00" – 19°48'00"E) is an east-west - striking shear comprised of mylonitic quartzite and muscovite-quartz schist.

The above-mentioned shears are sub-parallel to the Pofadder Lineament and are considered to be of the same age.

In the Mattheusgat Mountains, just north of the Pofadder Lineament (locality 29°08'15"S/19°30'00"E) quartzo-feldspathic gneiss was noted to unconformably overlie megacrystic alkali-feldspar granitic gneiss with a knife-sharp contact. The overlying quartzo-feldspathic gneiss is very fine grained up to five cm away from the contact. The contact could be the result of any of the following processes:

i) An intrusive contact due to the granitic gneiss intruding the quartzo-feldspathic gneiss

ii) A discrete surface of movement

iii) Flexural slip between two rheologically different rock types as a result of folding.

The last seems most probable since slightly further east, also to the north of the lineament, the quartzo-feldspathic gneiss is folded.

In the stereonet for D4 fabrics (Fig.14) the foliation (S4) has a wide scatter, but appears to show some concentration sub-parallel to the northerly-dipping S2 foliation. The L4 mineral lineation has a broad range in orientation varying from west-southwest to east-northeast as a result variation in plunge. This broad range may be partly caused by plotting the L4 mineral lineations for both dextral and sinistral shears without distinction. This is unavoidable, however, since in most cases the sense of movement is not apparent.

In shears it is not uncommon to see the mylonitic fabric folded. This is illustrated in Fig.15 with the development of highly asymmetric S-folds within the dextral shear in the north of the farm Pofadder East. It is noted, however, that the fold has an S and not a Z-asymmetry despite the fact that the shear, being part of the Pofadder Lineament, has dextral displacement. Another S-asymmetric mylonitic fold from the same shear can be seen in Plate 5. Why these folds have S and not Z-asymmetry is not certain but as has been pointed out in Section C.1., the asymmetry could have changed from Z to S for high shear strain, provided that the enveloping surface of the folds had rotated into the extension field of the applied strain (Ramsay et al., 1983).

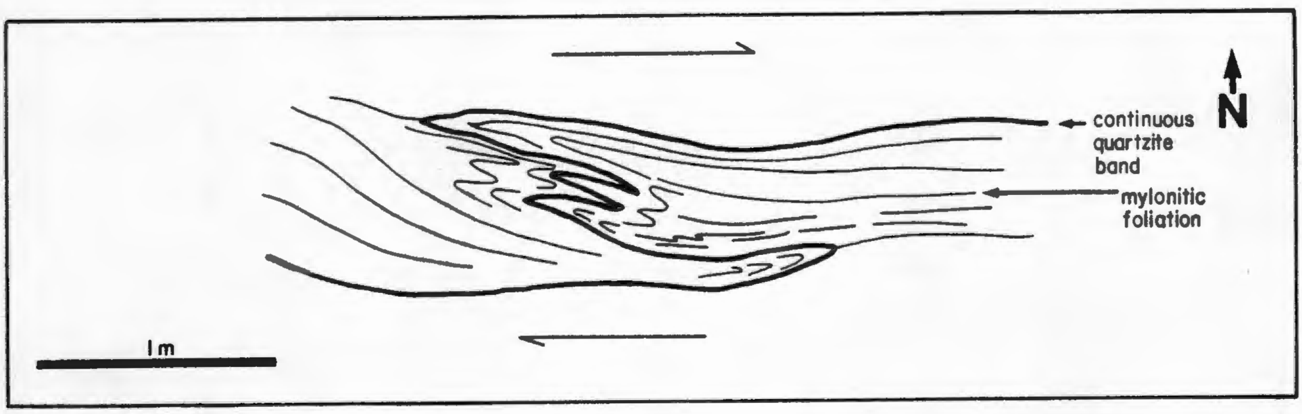


Fig.15: Field sketch of a strongly asymmetric S-fold of mylonitic quartzite in a dextral shear in the north of the farm Pofadder East. (locality: 29°05'40"S/19°32'15"E).

(c) Longziek Vlei Synform Region - This region has a mean F3b fold axis that plunges at 10° toward 080° (Fig.14). When viewed down plunge the synform has an interlimb angle of approximately 75° and can thus be described as an open to close fold after Ramsay (1967,p.349).

On the farm Brul-Kolk, in the very south of this region (locality 29°12'10"S/20°08'00"E) a fold interference pattern was observed in the field (see Geological Map). The structure is elongate in an east-northeast direction and is defined by alternate bands of granular quartzite and quartzo-feldspathic gneiss, with a foliation passing through the closures. The closures, probably having an axial-planar foliation, are either of F2 or F3b age. It is likely that they are of F3b age on the basis of their east-northeast trend, as this is the common trend for F3b folds. The closures are therefore considered to be of F3b age, and an interesting point regarding the interference pattern, if this conclusion is correct, is that the other generation of cross-folding can therefore not be F4 in age since a steep "a-direction" of movement is required for the second fold (Ramsay,1967). This "other"direction of folding must therefore be either F2 or F3a in age. The important point is that F3b is the last fold event of the cross-folding, and also the most tight.

Another important aspect of this region is that in the closure of the Longziek Vlei synform, strike lines taken from aerial photographs (Fig.10) reveal that an older fold generation is present, rotated about the closure. The age of this fold generation could not be determined from field work, but on the basis that the final fold pattern resembles the Type 3 interference pattern (Ramsay, 1967) the older folds should have an axial orientation sub-parallel to the F3b folds and could thus be of the F2 generation, as in general F2 and F3b fold axes do have sub-parallel trends. As in the other region already described, the mean F3b fold axial orientation and the minor F3b fold axes measured directly in the field, parallel the L2 mineral lineations. The synform region has been sheared by both dextral and sinistral shear sets, but this has occurred on a small scale without the development of major shears. The dextral sets appear to have moderate to steep dips and northwesterly strikes, and the sinistral sets also have moderate to steep dips but east-northeast strikes (Fig.14, D4 fabrics). The L4 mineral lineations on mylonitic surfaces trend sub-parallel to the F3b fold axes in most cases (Fig.14) and in the actual closure of the fold (Fig.12) the lineations appear to parallel the axial trend, plunging gently to the east-northeast. It is therefore assumed that the L4 mineral lineations are co-linear to the F3b fold axes, and since the mylonitic surface (S4?) in which these lineations are developed defines the closure of the fold, the hypothesised age relationship of the L4 lineations post-dating the F3b folds appears to be negated, at least at this locality. Unfortunately the age relationship cannot be resolved by studying the lineations as they are co-linear to the fold axis. It is suggested that either (i) the mylonitic surface and lineations developed prior to F3b folding, being older than the mylonites within the Pofadder Lineament which post-date the F3b folding, or (ii) that the mylonite post-dates the F3b folding, having formed as a result of movement along the folded surface parallel to the fold axis. The first hypothesis is the most likely as there are kinematic problems associated with the second, and it indicates that movement along the moderate to gently dipping mylonite surface occurred as thrusting, possibly associated with the F3a generation of folding.

(d) <u>Longziek Vlei Interference Region</u> - This region is a linear zone in which L2 mineral lineations are reversed in trend, trending west to west-southwest (Fig.14). Outcrops are parallel to the zone of reversed lineation, striking in a west-northwest direction (Figs.11 and 12) and strike lines from aerial photographs indicate that this is a zone of fold interference (see Geological Map). The axial directions of the two sets of folds required to produce the fold interference pattern are not clear. However, by comparing the D2 data for the Longziek Vlei Synform Region with that in this region, it appears that the foliation surfaces and the lineations have been rotated about a gently plunging axis. This axis plunges either to the north-northwest or south-southeast, and is probably responsible for the fold interference pattern. Since this trend crosses the Longziek Vlei synform, the other fold generation is presumably the east-northeast - trending F3b set of folds.

(e) Bank Vlei Region - The northern boundary of the Bank Vlei Region is marked by the northern limb of the Longziek Vlei synform, which is folded by an F4 fold with a northwest trending axial trace. The strike of the limb is rotated from a south-southwest direction in the northeast of the study area to west-southwest nearer the Pofadder Lineament. This rotation is consistent with dextral drag associated with the shears and hence the fold is considered to be an F4 (drag) fold. The attitudes of S2 foliations (Fig.14) indicate that the F4 fold axis plunges steeply northwest. Outcrop defining the northern limb of the Longziek Vlei synform alternates in dip-direction in a manner that appears to reflect the fold interference pattern shown in Fig.10 (see also Geological Map). This would indicate that, as for the Longziek Vlei Interference Region, at least two stages of folding are required.

To the immediate south of the farm road which crosses the northern limb of the Longziek Vlei synform, on the farm Drooge Grond (locality 29°05'00"S/20°09'30"E), a band of ortho-amphibolite shows signs of considerable flexural slip. The rock is pervaded with minor quartzo-feldspathic veins parallel to its strike and in places these veins display S-shaped folds indicative of sinistral movement.

In the south of the farm Drooge Grond, on the northern side of the Kakamas road (locality 29°09'S/20°13'E), strongly foliated calc-silicate gneiss has been folded and forms west-southwest - closing synforms which are in mutual contact. There is much pegmatite in the area and it is postulated that the rocks have been sheared, destroying the intervening antiforms. This shearing could possibly be related to the Putsberg shear in terms of both the west-southwest orientation of the folds and their location.

(f) T'Oubep Region - When the location of the T'Oubep Granodiorite Suite is considered in terms of the overall shear zone geometry, it is striking that the intrusive body is situated in the eastern quadrant of intersection for the dextral Adjoining Geelvloer shear (which is considered to be part of the Pofadder Lineament) and the extension of the sinistral Putsberg shear. This quadrant has the acute angle formed by the intersecting shears as its apex, an angle of approximately 30°, and since it hosts the syn to post-tectonic granodiorite body there is strong evidence that a system of conjugate shearing existed in which the east-west - directed quadrants were under an overall tensional stress. For the T'Oubep Granodiorite Suite, S2 attitudes depict the common east-northeast F3b fold axial trend, and have a plunge of 30° towards 070°. This trend is sub-parallel to the L2 mineral lineations, as for the other regions. If the hypothesis that the T'Oubep Granodiorite Suite intruded under a tensional stress system, and that the same stress system was responsible for both the D2 fabrics and the D4 conjugate shearing is correct, it follows that the deformation stages D2 to D4 were all controlled by one stress field.

(g) Conclusions - In the Gordonia Sub-province the L2 lineations plunge gently to the east-northeast in most cases. The lineation is considered to have resulted in a strong linear anisotropism in the rock and as a consequence F3b folding deformed the S2 foliation about axes which are co-linear to the L2 lineation. F3b folding is of a large scale in the Longziek Vlei Synform Region, and is of monoclinal style, resulting in fold interference patterns in isolated areas. Within the closure of the synform there is fold interference which is possibly of Type 3, resulting from superimposed folding of F3b on an older fold, probably of the F2 generation in a semi co-axial manner. In contrast to Type 3, Type 1 interference folding is noted to occur in the Longziek Vlei Interference Region, in the south of the Longziek Vlei Synform Region on the farm Brul Kolk, and in the Lucas Vlei Vlakte Region. In the first case, the zone of interference crosses the Longziek Vlei synform which is of the F3b generation and in the last two cases the interference structures are elongated in an east-northeast direction, strongly suggesting that they too are associated with F3b folding. It is therefore assumed that one of the fold generations involved in the development of the Type 1 fold interference patterns is F3b. Furthermore, since the "a-direction" of the cross-folds must be steeply inclined (Ramsay,1967) it is concluded that the cross-folds were not of the F4 generation, but F3b. Thus, the cross-folded set of folds must pre-date F3b folding, and yet if the structures are assumed to be defined by the regional foliation (S2) then the F3a folds are necessitated, being of intermediate age between F2 and F3b. It is postulated that when F3b folds were superimposed on F3a folds, Type 1 fold interference structures resulted, and that when they were superimposed on F2 folds, then Type 3 intereference fold patterns developed.

Two generations of mylonitic fabrics are considered to be present in the area, namely a moderate to gently-dipping fabric that is possibly associated with thrusting and which pre-dates the F3b Longziek Vlei synform, and a steeply-dipping post-F3b fabric associated with shearing. Mineral lineations within the mylonite that formed as a result of thrusting (L3a?), which are indistinguishable from the L4 lineations, have not been rotated as a result of the F3b folding since they are co-linear to these fold axes. The lineations plunge gently to the east-northeast and the direction of thrusting was probably parallel to the lineations, but in their "up-plunge" direction, namely west-southwest. The younger mylonitic fabric, which is associated with the shear zones, formed during the D4 deformation stage and probably followed on from the F3b folding as the northern limbs of F3b synforms have been refoliated in many places to form shear zones. Two directions of shearing developed during the D4 stage of deformation, namely west-northwest - striking dextral shears and east-northeast - striking sinistral shears. These shears are considered to be a conjugate pair, and their intersection forms quadrants of tension and compression as a result of the sense of movement for the intersecting shears. This is substantiated by the presence of the syn to post-tectonic T'Oubep Granodiorite Suite which occurs between the extension of the Putsberg shear which is sinistral and the Adjoining Geelvloer shear which is dextral. It is hypothesised that the same stress field which resulted in the conjugate shearing must have existed since at least the D2 stage of deformation because the D2 fabrics are developed in the T'Oubep Granodiorite Suite.

Finally, F4 folding associated with the east-southeast trending dextral shears is thought to have occurred. This is demonstrated by the rotation of F3b fold axial surfaces as they approach the Pofadder Lineament, as for example in the Bank Vlei Region. It is thought that the rotation developed contemporaneously with a tightening of the F3b folds as a result of increasing shear strain toward the shear zone since shearing is considered to have followed on from F3b folding.

3. The Pofadder Lineament

(a) Introduction - The main difficulty in studying the Pofadder Lineament is that within the present study area outcrop is poor. As a consequence, only the lithologies that are relatively resistant to weathering are exposed and within the lineament these rocks occur as isolated shears within the larger shear belt. The Pofadder Lineament is regarded to comprise the following shears (Fig.11):

i) Mattheusgat shear - The Mattheusgat shear splays off from the Pofadder Lineament in an easterly direction and is best developed within the Mattheusgat Mountains. The shear can be followed along the northern limb of the Longziek Vlei synform.

ii) Nouzees shear - The Nouzees shear is aligned parallel to the main trend of the Pofadder Lineament and occurs about 5 km to the southwest of the Mattheusgat shear, separated by an extensive sand plain. Within the plain, sporadic outcrop of granodiorite - tonalite rock occurs, which possibly represnts a late-stage intrusive into the Pofadder Lineament.

iii) Sandgat shear - The Sandgat shear was observed to initiate within the Houmoed Mountains to the east of the Houmoed farm boundary. It initiates as 1-3 m wide bands of mylonite in quartzite which are separated by quartzo-feldspathic gneiss. The bands converge eastwards with a concomitant intensification of shearing. In line with the extension of the Nouzees shear the mylonite bands rotate slightly to the south and then continue eastwards. This rotation is seen as the effect of dextral displacement along the Nouzees shear, but nowhere is it disrupted by the shear and the two shears are considered to merge. The shear continues eastward and strike-lines taken from aerial photographs indicate it to transect the Putsberg shear and thus to post-date it although this seems questionable (see Section F.4.c.). The shear can be followed as far east as the domal structure on the farm Haartebeest Vlei and it is postulated that the shear rotates back into parallelism with the main trend of the Pofadder Lineament to continue as the Adjoining Geelvloer shear.

iv) Adjoining Geelvloer shear - The Adjoining Geelvloer shear is of similar strike as the main part of the Pofadder Lineament, but as a consequence of the more eastward trend of the Sandgat shear, it is developed slightly to the northeast of the extrapolated extension of the main part of the lineament.

(b) Listric Shears - Shearing within the Mattheusgat Mountains rotates in strike direction from the main east-southeast direction to an easterly strike, and diverges into listric shears to the east. The listric zones are not clearly visible on an outcrop scale, but are clearly seen on aerial photographs. The eastward swing of the listric shears is shown in Plate 7. In the field they are mylonitic and their main function is regarded as one of rotating the zones of movement associated with the northern limbs of the F3b synforms into east-southeast - striking shears. Alternatively, they could provide a means of dissipating the shear strain over a progressively wider area to the east. The latter is not favoured, however, as the listric zones appear to be associated with the older F3b fold structures and are therefore probably related to the early stages of shearing rather than the final dissipation of shearing.

In a gorge that exposes a north-south section of the eastern tip of the Mattheusgat Mountains (locality 29°09'30" - 29°10'00"S/19°39'00"E) a listric shear is seen in section on a large scale (Fig.16). The foliation dips to the north and is steepest within the listric zone, becoming gentle to sub-horizontal away from the zone. This section has a sense of drag associated with it that possibly indicates a north-down component of vertical movement. This sense of movement is questionable as it indicates that tension existed across the shear in a north-south direction, which is not in agreement with the regional pattern of shearing in which compression is considered to be north-south, as will be determined in Section G. The north-down component of movement does, however, appear to be substantiated in terms of the mylonitic lineations (L4), as is discussed in Section F.3.c.(iv).

On the northern side of the southern-most valley within the Mattheusgat Mountains (locality 29°10'15"S/19°36'00"E) the listric shears can be observed along strike. They are characterised by ribbon-like bands of mylonite approximately 200-400 m long and 3-5 m wide (Plate 7). These bands appear to be interbraided on a larger scale which is consistent with the development of lozenge-shaped pods in the cross-sectional view (Fig.16).

Rocks situated between the listric shears are considerably less deformed than those within them and bedding is even preserved in places. In one locality, just north of the listric shear mentioned above, overturned trough cross-bedding is preserved in the west of the gorge (Plate 2).

(c) Fabric within the Pofadder Lineament - The fabrics and structures for the D2, D3 and D4 deformation stages are shown for each shear within the Pofadder Lineament in Fig.17 and the following points emerge:

i) Not many measurements could be taken of the D2 and D3 fabrics and structures within shears as they have been largely overprinted and deformed by the D4 stage of shearing. Most D2 and D3 data was collected from the Mattheusgat shear, and the data (Fig.17) shows that the S2 foliation has been folded about an F3b fold axis which has a mean plunge of 12° towards 084°. This orientation is similar to that of the minor F3b fold axes that were measured directly in the field, and is parallel to the L2 mineral lineations. The orientation and the above-mentioned parallelism between F3b fold axes and L2 mineral lineations is the same as

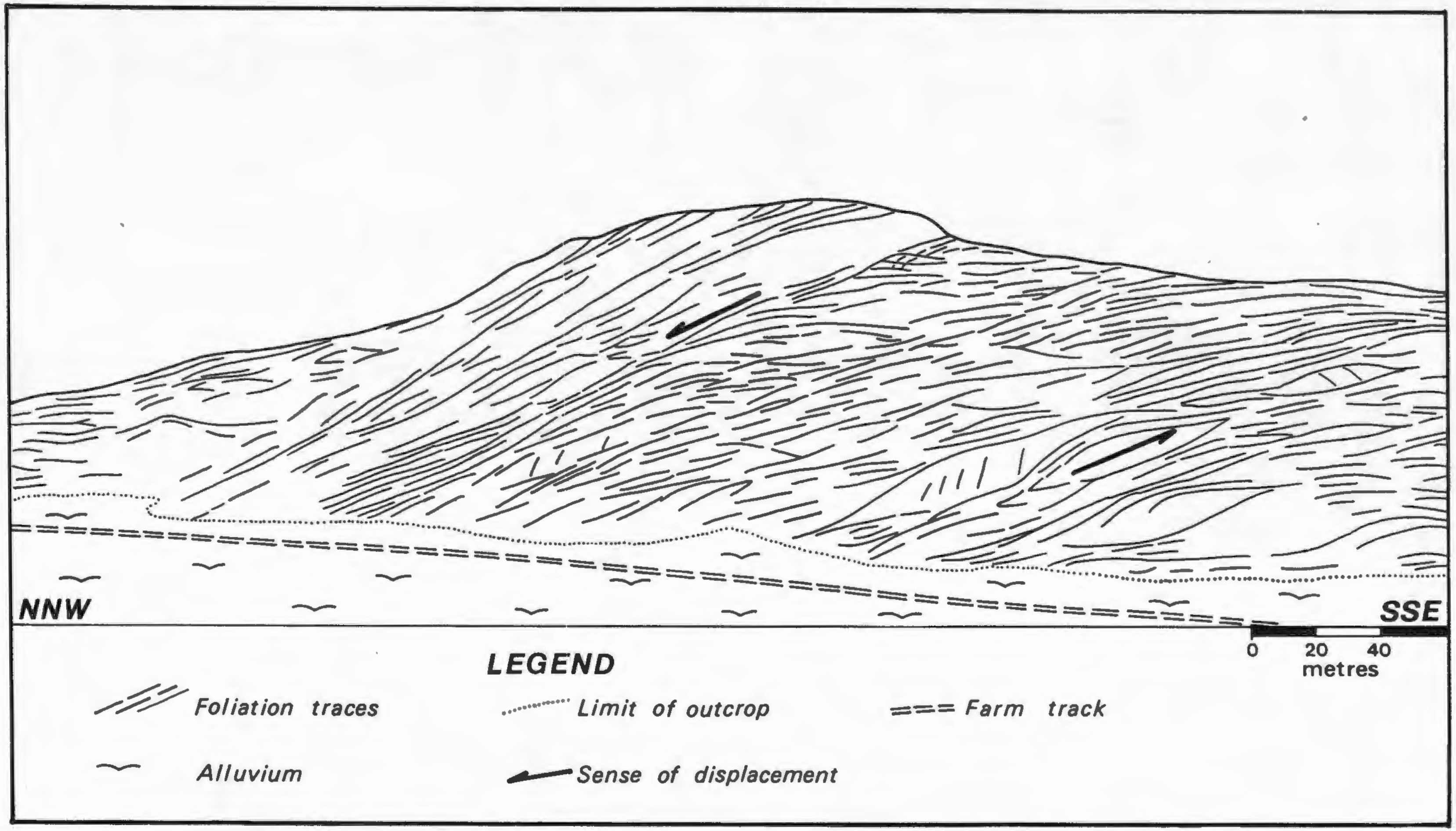


Fig.16: NNW - SSE cross-sectional view of an east-west striking listric shear in the east of the Mattheusgat Mountains. Detail traced from a photograph.
Lithology: muscovite-quartz schist
Locality : 29°10'00"S/19°39'00"E

for the Gordonia Sub-province. This strongly indicates that firstly, the D2 and D3 fabrics and structures were of similar orientation throughout the study area prior to shearing, and secondly, that these fabrics and structures have not undergone any significant rotation during shearing since they have the same orientation both within and outside the shears.

ii) The attitude of the S2 surface appears to have had a pronounced effect on the shearing. Within the Mattheusgat shear, S4 mylonitic foliation appears to have utilised the northward-dipping S2 foliation in its development. This is seen as being the result of slip along the existing S2 surfaces during shearing. Furthermore, L4 mineral lineations are in many cases sub-parallel to the older L2 lineations and F3b fold axes. This parallelism of linear structures and fabrics probably indicates that the older L2 lineation is markedly anisotropic and that it had a significant effect on subsequent deformation.

It is interesting to note that in the case of the Nouzees shear the S4 mylonite foliation has utilised the southward-dipping S2 foliation in its development rather than the northward-dipping S2 surfaces, and that lineations plunge to the southeast. This attitude of S4 and orientation of L4 seems to be unique to the Nouzees shear, and it is thought to reflect a predominance of southward-dipping S2 surfaces prior to shearing. Alternatively, and there is no evidence for this, the D4 fabrics could have been overturned to the south at a subsequent stage to their development. The former hypothesis is preferred.

iii) It is noted that of the mylonitic S4 foliations that have been plotted for the Mattheusgat shear, those for which the sense of shearing is apparent in the field plot away from the mean, as determined by visual inspection. This is seen as the result of S4 foliations developing along S2 surfaces, and where S2 was at a high angle to the shear plane, the mylonitic foliation developed along an S2 surface that was undergoing rotation. The rotation indicates the sense of shear.

iv) The attitude of the foliation and the orientation of both folds and lineations in the Mattheusgat and Adjoining Geelvloer shears are the same as those recorded by Toogood (1976,Fig. 52c) for the main part of the Pofadder Lineament to the northwest of the present study area. In most cases the S4 foliation dips to the north and L4 lineations plunge gently to the east-northeast. Since the shearing is of a dextral sense, an overall north-down dip-slip component of shearing is indicated.

It is apparent from the first three points that S4 fabrics tend to form parallel to the S2 fabrics, probably as a result of slip along the former surface and this implies that refoliation does not commonly transect older surfaces, but rather forms parallel to suitably orientated surfaces wherever possible. The development of a transposed foliation has, however, been observed in the field (locality 29°09'10"S/19°48'30"E). This process involves the progressive folding of an older muscovite-rich foliation in the lithology muscovite-quartz schist. The folds become overturned to the north as the shear is approached, and finally the folds become isoclinal within the shear and form a transposed surface.

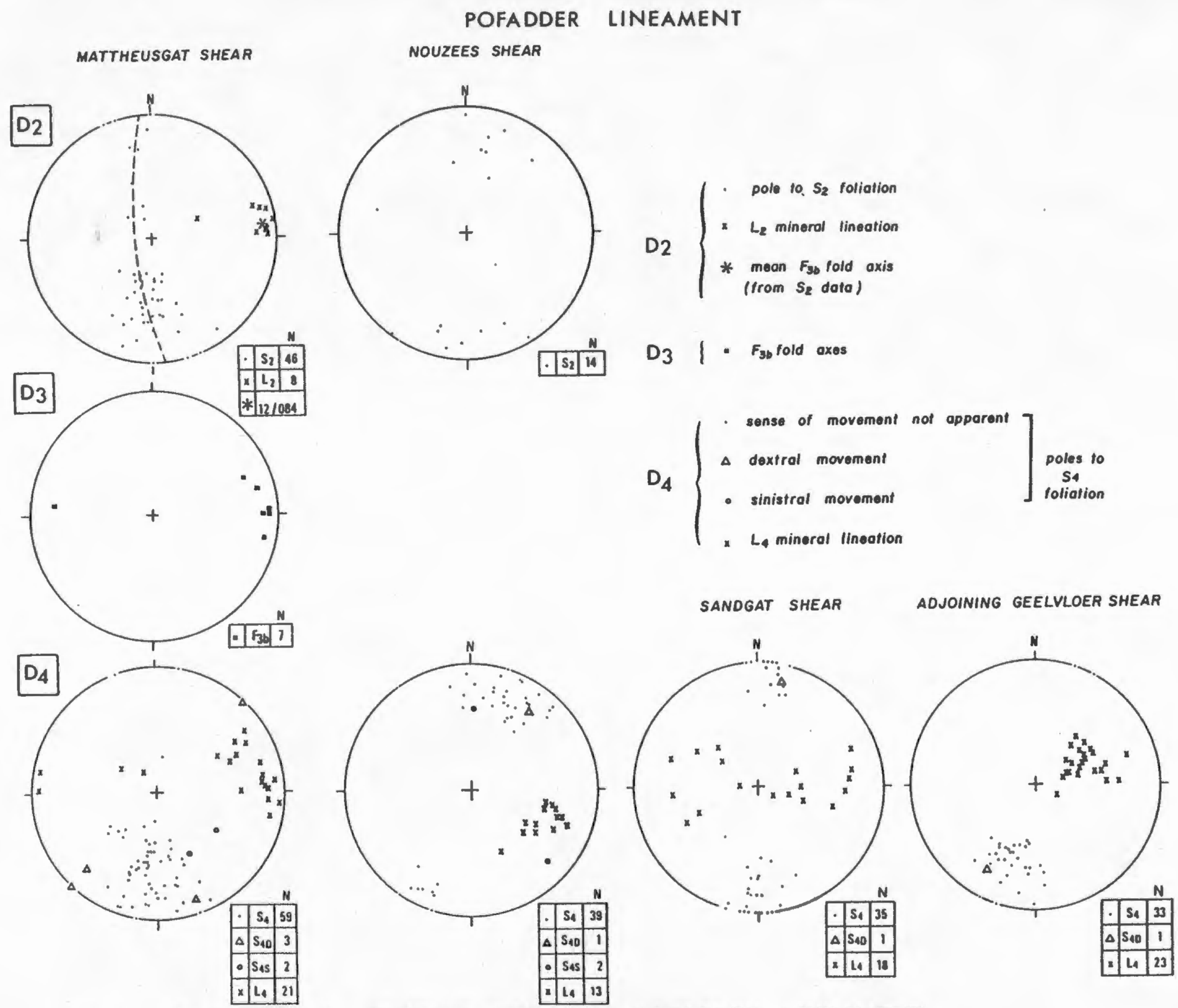


Fig. 17 D_2, D_3, AND D_4 FABRICS FOR THE POFADDER LINEAMENT

(d) Conjugate Shearing - It has been mentioned (Section F.2.f.) that the Adjoining Geelvloer shear, which is considered to be part of the Pofadder Lineament, forms a conjugate shear pair with the sinistral east-northeast - striking Putsberg shear. To the immediate north of the northern margin for the Nouzees shear, where dextral shearing is not as pervasive as within the shear, further evidence of a conjugate shear system can be obtained by studying the foliation pattern in the vicinity of intersecting shears on a mesoscopic scale (Fig.18). The foliation has formed in a geometric pattern that is compatible with both sets of shears, which have opposite senses of displacement. In Fig.18b, however, the sinistral shear appears to post-date the dextral shear, and since it includes the same features as in Fig.18c, though on a larger scale, caution is required if the latter is to be regarded as evidence of contemporaneous shearing. The most convincing evidence of conjugate shearing is that on a regional scale (Fig.20) neither the Pofadder Lineament (the Sandgat shear in particular) nor the Putsberg shear appear to displace one another at their intersection, and yet from west to east, the Pofadder Lineament rotates from a southeast to an east-southeast orientation in sympathy with the sinistral movement along the Putsberg shear, and the Putsberg shear rotates from northeast to east-northeast in sympathy with the dextral movement of the Pofadder Lineament. Thus, although neither shear is displaced at their intersection, each shear is seen to rotate the other which suggests that the two were contemporaneous, namely a conjugate pair. The lineations in the Putsberg shear plunge gently in either the east-northeast or the west-southwest directions (Fig.19), indicating that the movement had a major component of strike-slip, as is the case for the Pofadder Lineament.

(e) Discussion - One perplexing aspect about the lineament is to decide whether the vertical component of movement is north-up or vice versa. In the section on metamorphism it was noted that granulite high-grade metamorphic rocks occur to the north of the lineament, which substantiates the conclusion reached by various workers (Beukes,1973; Blignault,1974; Toogood,1976) that the mineral parageneses are stable for higher grades of metamorphism to the north. This metamorphic evidence, together with the asymmetry of the gentle plunging F3b folds indicates a north-up sense of movement. However, the L4 mineral lineations associated with the lineament plunge to the east or east-northeast at about 15° (Fig.12). This orientation together with the knowledge that movement of the lineament was dextral along steeply northward dipping S4 surfaces indicates that the vertical component of movement was relatively small, but that contrary to the other evidence, was north-down. Furthermore the cross-sectional view of the listric shear (Fig.16) substantiates this conclusion. It is accepted that the sense of movement was north-down since the evidence for this seems more reliable than the metamorphic evidence. Why then does the north have higher grade mineral parageneses? Toogood (1976) proposed that crustal uplift occurred only in the final stages of movement, post-dating the development of the L4 mineral lineations which he refers to as being "sub-horizontal". The possibility of thrusting towards the west-southwest (see Sections F.2.c. and F.2.g.), which is probably associated with F3a folding, indicates that crustal uplift could have occurred at an earlier stage in the deformation than proposed by Toogood,

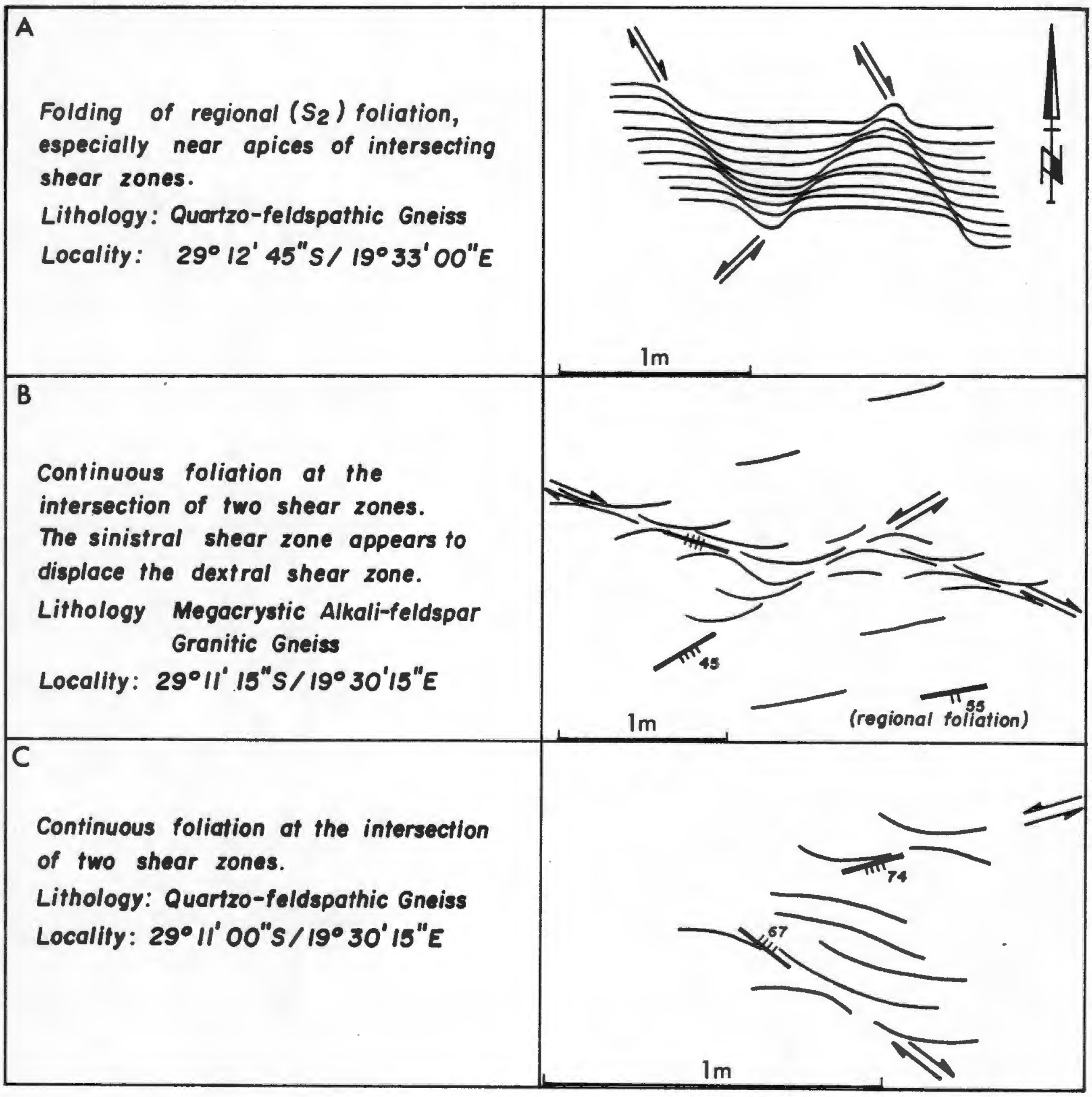


Fig.18: Foliation traces in possible conjugate shear pairs. In some cases, as in B, the sinistral set appears to post-date the dextral set.

pre-dating the development of the Pofadder Lineament. An alternative idea is that the higher grade mineral parageneses north of the lineament may reflect metamorphism that is not directly accounted for in terms of uplift on one side of the lineament. For example, in the present study the distribution of high grade metamorphic rocks could possibly reflect the occurrence of mafic intrusive rocks rather than a vertical component of movement associated with the lineament. In this regard high grade metamorphic rocks have been located both north and south of the lineament. The mafic granulite occurs to the north, on the farm Gemsbok Vlakte and is considered to be of intrusive origin (note that the rock has a metamorphic texture and does appear to be a true granulite with metamorphic orthopyroxene). High grade pelitic gneisses occur in close proximity to the Nouzees Gabbronorite Suite south of the lineament. In this light (see Chapter III) it is possible that high grade rocks could be associated with mafic intrusive aureoles (M.K.Watkeys, pers.comm., 1984), especially since most of the mafic bodies occur just north of or within the lineament.

Finally, the estimation of displacement across the lineament will be discussed. The southern limb of a synform to the east of Pella is defined by a prominent quartzite horizon (Joubert, 1973) which is seen to cross the lineament, extending into the Mattheusgat Mountains within the present study area, having been mapped as intercalated quartzite and muscovite-quartz schist. Since muscovite-quartz schist also defines the closure of the Longziek Vlei synform, and bearing in mind that lithologies must rotate in sympathy with the dextral movement of the lineament, this synform is correlated with that near Pella. It is thought to be totally fortuitous that the Longziek Vlei synform is in approximate alignment with the Samoep synform to the west of the lineament, and indeed this synform cannot possibly be the equivalent of the Longziek Vlei synform, as this would indicate that there is no displacement across the lineament. It has been mentioned that Toogood (1976) has calculated the displacement to have been 85 km, provided that the pure shear component associated with the simple shear was as high as 40 per cent, and that the dispacement could have been further if the pure shear component had been less. It would seem that a displacement of 85 km is excessive, since if the synform east of Pella is correlated with the Longziek Vlei synform, and if it is assumed that the orientation of the fold axial trace was east-northeast, prior to shearing, then a displacement within the range of 45 to 60 km is indicated, approximating 50 km. It is stressed that this is a tentative estimate.

4. Bushmanland Sub-province

The Bushmanland Sub-province lies to the southwest of the Pofadder Lineament and is considered to have been up-lifted relative to the northern side of the lineament during D4, for reasons already mentioned in the previous section.

The Bushmanland Sub-province is comprised predominantly of quartzite and calc-silicate gneiss, the former commonly having a mylonitic fabric. There is sparse outcrop in this region with large areas being obscured by pegmatite scree or calcrete, and in the south there is considerable float of the Karoo

Sequence rocks.

The most characteristic feature of this sub-province is the presence of closed structures which are clearly visible on aerial photographs (see Fig.10). Foliation measurements in the environs of these closed structures were difficult to obtain on account of the sparse outcrop, but in most instances it could be ascertained whether the closed structures be domes or basins. In general there is a predominance of domes, the basins possibly being concealed by superficial deposits or cover. Field mapping has also shown that these structures are commonly defined by a mylonitic fabric at their margins where the foliation is steepest. Closed structures are especially well developed inbetween convergent shears, as in the case of the Putsberg and Sandgat shears (Fig.10; Geological Map). The sub-province has been sub-divided into three regions, namely the Nouzees, Houmoed and the Vraa-Weer Regions. The first differs from the other two as it consists predominantly of the Nouzees Gabbronorite Suite and thus has deformed as a competent body without the development of a strong foliation. Both the Houmoed and the Vraa-Weer Regions are characterised by paragneisses and quartzite which define dome-and-basin structures, and they are separated by the Putsberg shear.

In regard to the stereonets for the Bushmanland Sub-province (Fig.19), the regions are commented on as follows:

(a) Nouzees Region - The gabbronorite has undergone retrograde metamorphism to amphibolite, the latter commonly possessing a weakly developed fabric. Most of the structural readings for the region were measured from the supracrustal rocks in the area as their foliation is strongly developed. The D2 fabrics show more scatter than for the other regions in the sub-province, this possibly being a consequence of the competent nature of the gabbronorite body. Measurements of the F3b fold axes were obtained from the supracrustal rocks, and their orientation is typically east-northeast. The D4 fabrics show considerable scatter but in general the sinistral shears strike to the northeast and the dextral shears strike to the east.

(b) Houmoed Region - Paizes (1975,p.160) ascribed the shearing in the Houmoed Mountains as being the result of flexural slip associated with the "Houmoed Antiform" whereas in the light of the present mapping it is considered as a dome, located at the convergence of two shears. From Fig.19 the S2 data reflects the direction of tightest curvature of the dome to be east-northeast, as shown in Fig.10.

(c) Putsberg Shear - The Putsberg shear is defined by mylonitic quartzite along most of its length, and lineations are defined by stringers of recrystallised quartz. Stereonets for this shear (Fig.19) indicate that both S2 and S4 foliations strike east-northeast with steep dips and that the mylonite lineation L4 is gently plunging to sub-horizontal trending parallel to the shear. In the extreme west of the shear a para-amphibolite band occurs and the strike lines taken from aerial photographs (Fig.10) indicate a closed structure to be developed within it, revealed as a basin by mapping. Within

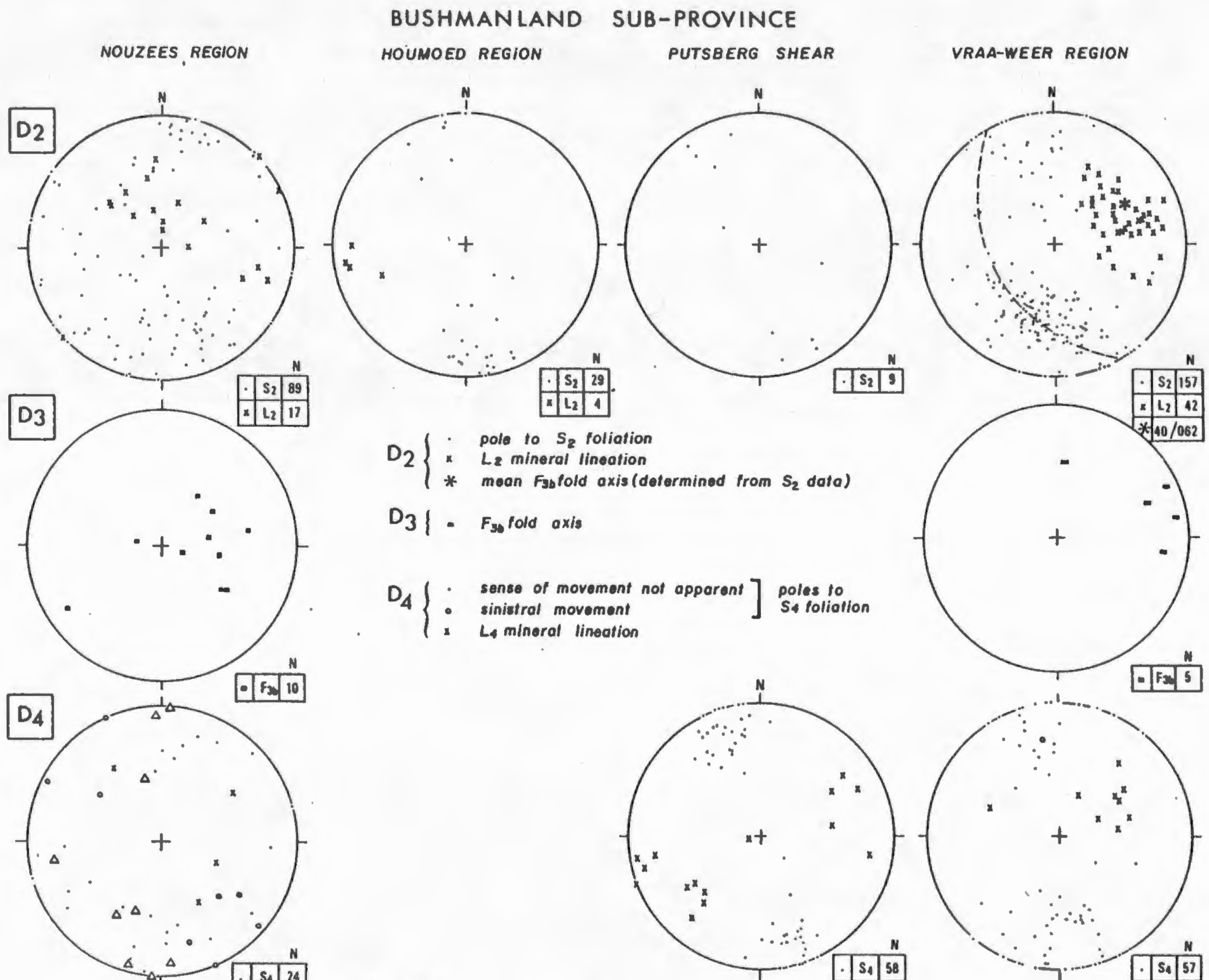


Fig. 19:

D2,D3 AND D4 FABRICS FOR THE BUSHMANLAND SUB-PROVINCE the para-amphibolite are numerous quartz veins parallel to the shearing and lenses of biotite schist which in turn hosts pods of sillimanite. A specimen from such a pod (HM 108), shows in thin section that the sillimanite fibres define a foliation which has been folded at least twice, one generation being kink folds, and the other which appears to post-date the former, being gentle, rounded folds. The sillimanite forms contorted bands on a millimetre scale, the banding existing as white and brown-coloured sillimanite, the boundaries of which are cut by the foliation. These bands could represent either an original bedding or a secondary alteration, but the latter is not likely since both the brown and white sillimanite are distinct from a coarser grained secondary sillimanite which also exists. This rock is therefore considered to show a contorted bedding and two successive stages of folding.

In the vicinity of the homestead "Nuwedam" in the south of the farm Sandgat, the Putsberg and Sandgat shears intersect as shown in Fig.20. In two localities (Fig.20, localities A, B) the Putsberg shear appears to post-date the Sandgat shear but the field evidence is not conclusive. On a larger scale this evidence is apparently contradicted by inspection of the strike lines taken from aerial photographs (Fig.20: location map). The age relationship of the two shears therefore remains unresolved and the possibility of them being contemporaneous still exists although locally one may deform the other and vice versa.

(d) Vraa-Weer Region - The Vraa-Weer Region has very sparse outcrop, most of which is confined to a band of quartzite and calc-silicate gneiss. The band is up to 2 km in apparent width and is probably structurally duplicated in many parts. This band occurs south of a series of closed-structures and is gently folded in a shape that defines the margins of these structures. The band extends from the farm Willems-Opdam and can be followed eastward, finally rotating to the southeast upon entering the Adjoining Geelvloer shear.

The stereonets for the Vraa-Weer Region (Fig.19) show that the S2 surface is folded about an F3b fold axis that plunges at 40° towards 062°. As in the other regions this axial orientation and the axes of minor F3b folds parallel the L2 mineral lineations indicating that folding was controlled by the L2 linear anisotropism in the rocks. The linear anisotropism and the presence of the S2 foliation are considered to have also controlled the D4 stage of deformation as the L4 mylonitic lineations have formed parallel to the linear structures and fabrics, and the S4 mylonitic foliation has formed parallel to the S2 foliation.

The D2 fabric (S2 and L2) which defines the fold interference pattern in the Vraa-Weer Region appears to have been folded about an east-northeast-trending fold axis (Fig.19) which is considered to be the axis of an F3b fold. The reason why the direction of the older cross-folded set of folds is not apparent in Fig.19 is probably a consequence of two factors: firstly, these folds are thought to be more gentle than the east-northeast - trending F3b folds judging from the interference pattern shape, and secondly, the data sets are not large enough to reflect the complete pattern.

The F3b, east-northeast-trending folds must be the younger of a set of cross-folds to produce the dome-and-basin interference pattern (see Section C.2.) as the younger set must have a steeply inclined "a-direction" of

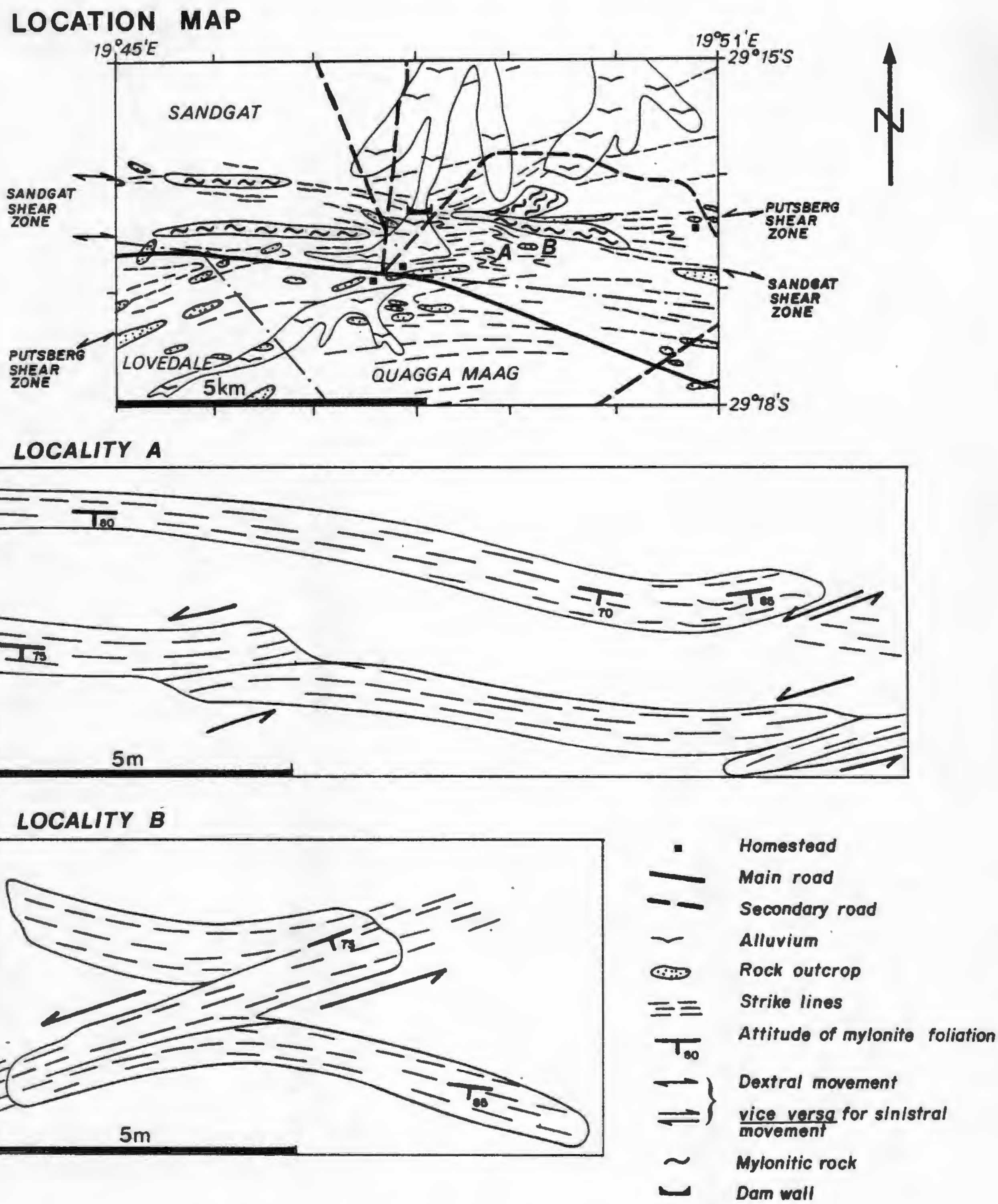


Fig.20: Field sketches for two localities (A and B) in which the Putsberg shear appears to post-date the Sandgat shear. On a larger scale (see location map) the opposite is indicated.

movement, a condition which is not met by the F4 folds as they are associated with strike-slip deformation. In contrast, the F3b folds have gently plunging axes, and assuming that the "a-direction" is contained in the axial surface at a high angle to the axis (preferably at right angles) this condition is met. Accepting that the east-northeast-trending fold axes are the youngest of a pair of cross-folds, and assuming that the dome-and-basin structures are defined by the regional fabric (S2), then the older fold set must be of an intermediate age between F2 and F3b. For this reason, and as for the Gordonia Sub-province, the older set of folds which have been cross-folded is postulated to be of F3a age, their trend being uncertain.

(e) Conclusions - For the Bushmanland Sub-province as a whole, it is concluded that:

(a) F3a and F3b folds, the latter being of east-northeast trend, resulted in dome-and-basin fold interference patterns.
(b) The Putsberg shear is characterised by a predominant component of strike-slip movement and is considered to be a conjugate shear to the dextral shears of the Pofadder Lineament.
(c) The presence of dome-and-basin structures which are of D3 in age is thought to have controlled the D4 shearing to some extent as they probably behaved as relatively stable "augen" It is thought that the domes or basins tended to control the pattern of shearing, such that the conjugate shear pairs converge on either side of them.

G. Fracture Analysis

In the field it was possible to distinguish between fractures and (ductile) shears, as the former occur as sharp breaks in the rock, without any associated drag or refoliation. In general no displacement was observed along the fractures, but on a megascopic scale, faults which strike northwest show a limited dextral displacement.

Brittle deformation is assumed to have been the same over the whole study area and hence all the fracture data is represented on a single stereonet (Fig.21) for which there is an accompanying Rose diagram showing strike-directions of the fractures. It is thought that the Pofadder Lineament acted as a major wrench zone in its final stages of movement for the following reasons:

(1) The fractures have dips that predominantly exceed 70° and therefore intersect along a steep to vertical axis indicating that sub-horizontal shear couples were operative in the area, which is characteristic of wrench zones (Moody and Hill, 1956; Wilcox et al.,1973).

(2)The strike directions (Fig.21) for the fractures are consistent with the theory of Riedel fractures (R), conjugate Riedel fractures (R'), P-fractures (P), and Principal Displacement fractures (D) for wrench zones (Tchalenko and Ambraseys, 1970).

(3)The fractures have only limited strike-slip displacement which is typical of wrench zones (Wilcox <u>et al.</u>,1973).

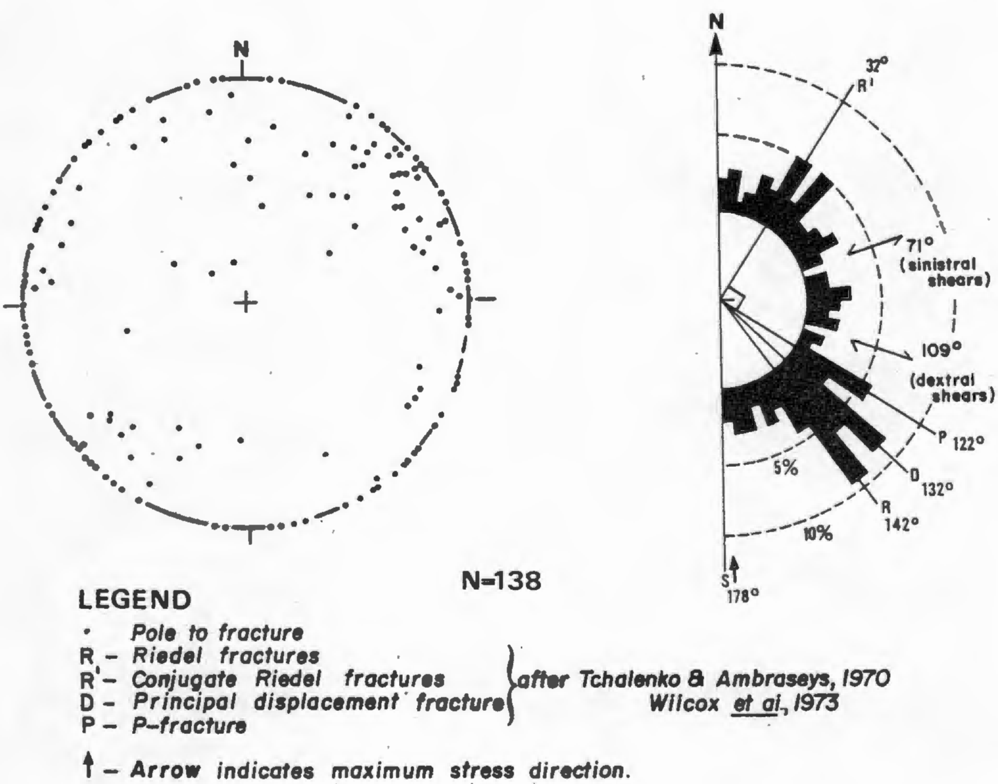


Fig.21: Fracture measurements and their correlation with wrench-zone fracture directions. This correlation is independant of the mean directions of shearing for the area which are also shown.

The theory (Section B.1.) of fracture development in wrench zones using the Coulomb failure criterion has been applied independently of the known directions of shearing that operated during the ductile stage of deformation. Firstly, R, D and P fracture directions (see Figs.9 and 21) are at an acute angle to each other with D being the bisector, parallel to the general southeast direction of movement. Secondly, the angle between P and R' is a right angle. Secondary fractures resulting from stress redistribution after the development of the wrench system probably account for the numerous other strike directions and are obviously too complicated to unravel as there are too many directions of older movement involved, namely P, D, R and R'.

The following aspects are of importance concerning the wrench-zone fracture directions:

(1)The direction of maximum stress was orientated at 178°, (north-south), since the maximum stress direction is the acute bisector of the Riedel and conjugate Riedel fractures according to Wilcox et al.(1973).
(2)The principal displacement fracture (D) is orientated at 132°, thus it has a southeast strike.
(3)By drawing a line from the Mattheusgat shear in the northwest to the Adjoining Geelvloer shear in the southeast the local trend of the Pofadder Lineament in the study area is 109°. The mean trend of the Putsberg shear is 71°. The bisector of the obtuse angle between the shears is the maximum shortening direction, provided that the shears form a conjugate pair (Ramsay and Allison, 1979). In this case the direction would be 180°, namely north-south, which can be regarded as the same as that for the wrench zone (178°), and hence the shearing and wrenching probably developed under the same stress field.

Wilcox et al.(1973) have shown experimentally that deformation in shears is initially ductile and that with progressive deformation the shears narrow to become effectively faults. Such a process of deformation could have operated in the study area, with the movement that was associated with the dextral shears continuing into a time when brittle conditions prevailed. This is substantiated by the following observations:

(1)Throughout deformation, movement was essentially horizontal as demonstrated by:

a)the shallow plunging mineral lineations in shears.
b)the sub-vertical attitude of the fracture surfaces.

(2)The fracture pattern is symmetrical with respect to the mean directions of shearing (see Fig.21).
(3)The Riedel fractures have a dextral displacement and they are localised along the Pofadder Lineament, this being clearly shown in Fig.10.
(4)Epicentres of earth tremors in southern Africa (Fernández, 1983a,b) show that there is a concentration of seismic activity along the southeasterly extension of the Pofadder Lineament (Fig.22). These are shallow-level tremors (eg. 10 km) and are considered to indicate that movement along the shear is still active.

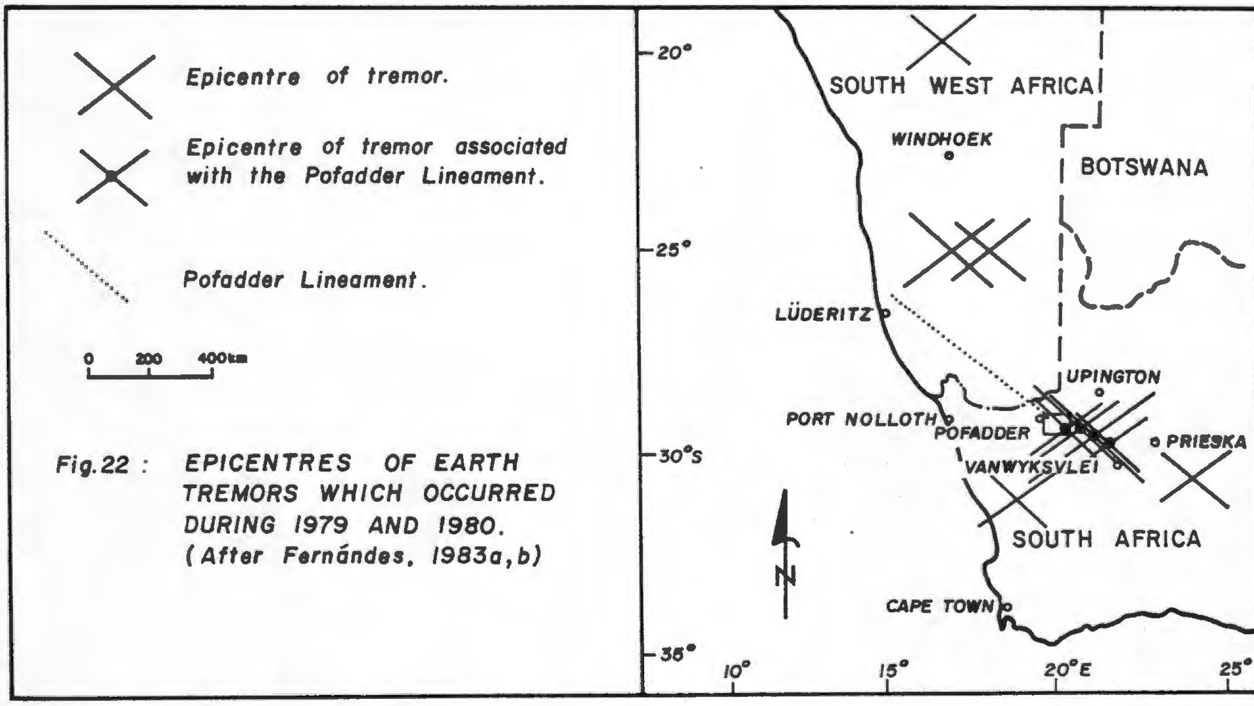


Fig.22 : EPICENTRES OF EARTH TREMORS WHICH OCCURRED DURING 1979 AND 1980. (After Fernándes, 1983a,b)

H. Structural Synthesis

In the preceding sections the structure has been described in detail for ductile deformation in terms of structural sub-provinces, regions and shear zones; and for brittle deformation in terms of a fracture analysis for the whole area. The deformation history is to be discussed for the study area as a whole, with special reference to any difference that could have existed from one region to another. A bias is given toward the deformation processes since the deformation sequence (stages D1 to D5) and the structures have already been described. In this regard the deformation stages are grouped into three phases of deformation, each phase being characterised by a different set of processes. The first phase involves folding of the S0/1 surface, resulting in two sets of folds, namely F1 folds and F2 folds, the latter deforming the former and having an axial-planar foliation. This foliation (S2) is the regional foliation, and probably marks the most intense stage of deformation in this first phase of deformation, being coincident with, or just prior to, the peak of metamorphism. The second phase of deformation (not stage but phase) involves folding of the S2 foliation into the F3a and F3b fold generations, the foliation facilitating flexural slip folding to occur. The F3a folds are shallow plunging and appear to have northwest or southeast trending fold axes. Little is known about these F3a folds, but they appear to be associated with west-southwest-directed thrusting. The thrust surfaces are shallow dipping mylonitic surfaces and have been folded by the younger F3b folds, as for example in the closure of the Longziek Vlei synform. The F3b folds have resulted in Type 1 fold interference patterns where they are superimposed on F3a folds and Type 3 interference patterns where they are superimposed on F2 folds. The third phase of deformation, which probably followed on from the last, but in which shearing controlled the mode of deformation, involved the D4 shearing stage and the D5 wrenching stage. The above-mentioned phases of deformation are to be discussed in more detail.

1. Fabric-development Phase (D1 and D2)

The oldest surface recognisable in the study area is S0/1 which is either the original bedding in the paragneisses or else a transposed equivalent of the bedding. This surface has been overprinted by the main fabric and is only recognisable as an isoclinal rootless folded surface (F1 folds). The same surface (S0/1) which defines the F1 folds also defines the F2 folds and it is likely that the tight F2 folding occurred within the same folding process as the F1 folds, but as a late-stage development with a concomitant development of an axial-planar foliation (S2). It does not seem justifiable to separate the two fold generations (even though F2 folds are seen to post-date F1 folds) into separate phases of deformation as they could have developed within a single deformation process. In this regard Williams (1983) has shown that within a single event of shearing, perturbations in the rock strata can result in early folds developing which in turn become refolded as the shearing continues. It is considered that deformation became most intense during the second stage of this phase, namely during D2, and that P-T conditions were at, or just prior to, those of peak metamorphism, with the result that the regional foliation (S2) developed. A mineral lineation (L2) is developed parallel to the F2 fold axes and has resulted in a strong linear anisotropism in the rocks. It is postulated that until this stage of the deformation history the whole study area underwent similar deformation and hence the various structural areas had not yet begun to deform differently.

2. Fabric-folding Phase (D3a and D3b)

The start of the D3 deformation stage is marked by a change in the deformation process since the S2 surface, which is a penetrative fabric, facilitated flexural-slip folding. Evidence regarding the D3a stage is slender, but as concluded for the Gordonia and Bushmanland Sub-provinces, the development of dome-and-basin structures (Type 1 interference pattern, after Ramsay,1967) requires the presence of gently plunging F3a folds which must be at an angle to the east-northeast-trending F3b folds. The orientation of the former folds is not certain, but in the Longziek Vlei Interference Region (see Section F.2.d.) stereonets for the D2 fabrics (Fig.14) indicate that the S2 foliation surfaces and the L2 lineations have been rotated about shallow plunging north-northwest (or south-southeast) trending fold axes which are thought to be of the F3a fold generation, as mentioned in Section F.2.g. Thus, the F3a folds probably trend north-northwest (or south-southeast). This direction is in agreement with the fold interference patterns shown in Fig.10 in which the F3a folds are indicated to have northwest (or southeast) trending axial traces. Since these folds are shallow plunging, their axial traces will approximate the direction of their fold axes and hence it can be accepted that the folds trend to the northwest (or southeast). Little is known about the kinematics of this fold event, but as mentioned in Sections F.2.c. and F.2.g., these folds may be associated with thrusting in a west-southwest direction, pre-dating the F3b folding. The F3a folding and associated thrusting is more common to the east of the present study area (R.W.Harris, pers.comm., 1984) and hence it seems that these structures are better preserved away from the Pofadder Lineament where they have been overprinted to a lesser extent.

Post-dating the F3a folds, large-scale, open F3b folds with east-northeast-trending axes developed with a gentle plunge. These folds have had two major efects. Firstly, their gentle plunge necessitates that the "a-direction" of movement was steeply inclined and thus dome-and-basin fold interference patterns resulted from the refolding of the F3a folds. Secondly, refoliation occurred along the northern more steeply dipping limbs of the synforms, and as deformation continued the refoliated limbs developed into shear zones. For example, the northern limb of the Longziek Vlei synform is sheared. The association of shearing with the east-northeast-trending F3b folds indicates that the shearing, which only became pronounced during the D4 deformation stage, started during the D3 stage. As a corollary, D4 is thought to have followed on from D3 without any necessary change in the stress field.

3. Shearing and Wrenching Phase (D4 and D5)

It is postulated that D4 followed on from the D3 stage of deformation when shearing intensified to such an extent that it controlled the mode of deformation. Shear zones no longer developed along the northern limbs of F3b synforms, but formed continuous zones commonly truncating the synforms, as seen by the Nouzees shear which truncates the Samoep synform in the west of the present study area. Such truncation does not necessarily imply a change in the stress field between the D3 and D4 deformation stages, and there is no evidence to suggest such a change. In terms of the sub-division of the area into sub-provinces and regions, it is postulated that it was late in the D3 stage of deformation that shearing developed and that the various regions started to deform differently from this stage onward, being separated from one another by shears. The regions are considered to have formed in two groups, namely the Gordonia and Bushmanland Sub-provinces when the shearing reached its maximum development, separated by the Pofadder Lineament. The lineament has a dextral sense of movement, an east-southeast strike, a moderate to steep northward-dip and a small north-down component of dip-slip. The lineament comprises the Mattheusgat, Nouzees, Sandgat and Adjoining Geelvloer shears, which occur as isolated shears due to the lack of exposure. The dip-slip has in all cases, except for the Nouzees shear, a normal sense of movement in cross-section indicating that a tensional stress existed across the shears. In the Nouzees shear the opposite applies.

Sinistral shearing is not as widespread as the dextral shearing, and is localised predominantly along the Putsberg shear which has an east-northeast strike and a sub-vertical dip. The shear is predominantly of strike-slip displacement, with possibly a small north-down component of dip-slip.

Convergent shears, commonly having opposite senses of movement, tend to define the tapered margins of dome-and-basin structures. These dome-and-basin structures are considered to pre-date the shears, probably having formed as a result of superimposion of the F3b folds on F3a structures, and hence are considered to have acted as "augen" during shearing with the mylonite interbraiding around them. Regarding the intersection of the Sandgat and Putsberg shears, the following observations appear to substantiate that they form a conjugate pair:

a) Neither shear appears to displace the other in the vicinity of intersection.
b) The effect of contemporaneous shearing is seen by the rotation of each shear as a result of the other. Towards the intersection, the Pofadder Lineament rotates from an east-southeast strike to a more easterly strike. This rotation is seen to be the result of the Putsberg shear. Similarly, the Putsberg shear rotates from an east-northeast to a more easterly strike as a consequence of the Pofadder Lineament.
c) The T'Oubep granodiorite body has intruded into the eastern quadrant of the intersection. Tension would have existed in this quadrant provided that the two shears were contemporaneous.
d) To the immediate north of the northern margin of the Nouzees shear, where dextral shearing is not as pervasive as within the shear, small-scale dextral and sinistral shears have continuous foliation traces at their intersections, indicating that the shear sets probably form a conjugate pair.

F4 folds have northwesterly trending axial traces, and appear to be the result of "drag-folding" of F3b folds, most probably due to dextral shearing. Since shearing is considered to have followed on from the F3b folding, and possibly within the same stress field, it is likely that the F4 drag-folding occurred in the final stages of F3b folding. F4 folds have a gently plunging "a-direction" of movement as the shearing was predominantly of a strike-slip sense. The folding would therefore be best exhibited in surfaces at a high angle to the "a-direction", namely in steeply dipping surfaces such as the northern limb of the Longziek Vlei synform.

The most important aspect of this study has been to draw conclusions regarding (a) the extension of the Pofadder Lineament to the southeast of the present study area, and (b) the ductile-brittle conditions of the lineament. The lineament can be traced through the study area, but it is covered by rocks of the Karoo Sequence in the extreme southeast and hence it is uncertain how far the (ductile) shearing continues in this direction. In terms of the period of time for which shearing occurred it is most likely that movement along the lineament continued over an extended period of time and was even active when the rocks became brittle. Movement during these brittle conditions resulted in a wrench zone developing, straddling the Pofadder Lineament, and this is seen by the presence of a well-developed set of Riedel fractures. These fractures are clearly seen in the field as they are commonly filled with quartz (see Geological Map or Fig.10). It is postulated that the development of the wrench zone is a continuation of the shear deformation and not a reactivation of the shears, since both the shears and the wrench zone appear to have been subjected to stress fields of similar orientation, which possibly indicates that a single stress field accounted for them both. The extent of the wrench zone is not certain, but it appears to have affected the Karoo Sequence rocks, and the presence of a 60 km linear zone of superficial deposits lying on the Karoo rocks which can be traced as far southeast as Vanwyksvlei (Joubert,1974c) is thought to reflect a depression in the rocks due to the wrenching. Thus the wrench action is considered to continue at least as far to the southeast as Vanwyksvlei. The total displacement across the Pofadder Lineament, which is mostly accounted for by ductile shearing and only to a limited extent by wrenching, is considered to be approximately 50 km. In terms of the period of time during which movement was active, it is safe to say that it was still active during the Carboniferous as rocks of the Karoo Sequence have been affected. Seismic activity still exists along the lineament and hence it is probable that movement is still continuing. It is a well-known fact that movement along many wrench zones was initially ductile (Wilcox et al., 1973) and the Pofadder Lineament seems to be no exception.

Chapter V

MYLONITE TEXTURES

A. Introduction

Since shearing has been the most important topic dealt with, the recognition of sheared rock and in particular of mylonitised rock, both in the field and under the microscope, has been vital to this study. Classification criteria on a microscopic scale will be discussed and the effect of the mineral mode of the sheared rock is also introduced since it is found that a mylonitic fabric tends to be best developed in rocks that have the least amount of hard minerals. The term "hard" is used in the same sense as by White et al.(1980,p.181), referring to minerals in a rock that are relatively resistant to recovery and recrystallisation, tending to behave in a semi-brittle manner by means of rigid body rotation and fragmentation. In contrast the term "soft" is used for minerals that deform in a crystal-plastic (ductile) manner, as for example quartz.

B. Classification of Mylonite Rock

Until fairly recently the universal view was that mylonite fabrics developed by a process of brittle deformation and that recrystallisation textures commonly found in these fabrics were post-tectonic (Higgins,1971). More recent studies, however, (Bell and Etheridge, 1973; White,1973; Lister et al., 1977) have shown recrystallisation to be syn-tectonic. This led to a clear distinction between cataclasites and mylonites, simplifying the terminology as shown in Table 20.

In the field sheared rock is commonly of a finer grain, flaggy in outcrop and has a steep to sub-vertical dip. The foliation surface commonly contains string-like aggregates of quartz and augen-shaped porphyroclasts which are commonly feldspar. Both the string-like aggregates and the porphyroclasts define a strong mineral lineation. In regard to the classification scheme (Table 20), the shear zones are all characterised by foliated and cohesive textures, with the proportion of matrix commonly exceeding 10 per cent. The rocks therefore typically form the mylonite series and in some cases even blastomylonites.

Table 20: Classification of fault rocks (Sibson,1977)

			RANDOM-FABRIC	FOLIATED	PROPORTION OF MATRIX
INCOHESIVE			FAULT BRECCIA (visible frags. > 30% of rock mass)	?	
INCOHESIVE			FAULT GOUGE (visible frags. < 30% of rock mass)	?	
COHESIVE	NATURE OF MATRIX	Glass/ devitrified glass	PSEUDOTACHYLYTE	?	
COHESIVE	NATURE OF MATRIX	Tectonic reduction in grain size dominates grain growth by recrystallisation & neomineralisation	CRUSH BRECCIA FINE CRUSH BRECCIA MICROBRECCIA	(fragments > 0.5 cm) (0.1 cm < frags. < 0.5 cm) (fragments < 0.1 cm)	0-10%
COHESIVE	NATURE OF MATRIX	Tectonic reduction in grain size dominates grain growth by recrystallisation & neomineralisation	PROTOCATACLASITE (Cataclasite Series)	PROTOMYLONITE (Mylonite Series)	10-50%
COHESIVE	NATURE OF MATRIX	Tectonic reduction in grain size dominates grain growth by recrystallisation & neomineralisation	CATACLASITE (Cataclasite Series)	MYLONITE (Mylonite Series; Phyllonite Varieties)	50-90%
COHESIVE	NATURE OF MATRIX	Tectonic reduction in grain size dominates grain growth by recrystallisation & neomineralisation	ULTRACATACLASITE (Cataclasite Series)	ULTRAMYLONITE (Mylonite Series; Phyllonite Varieties)	90-100%
COHESIVE	NATURE OF MATRIX	Grain growth pronounced	?	BLASTOMYLONITE	

C. Factors Controlling the Development of Mylonitic Fabrics

The classification scheme for mylonitic rocks (Table 20) indicates a basic trend in textural development from a protomylonite, through mylonite to ultramylonite in direct response to grain-size reduction. In more detail, however, the grain-size reduction depends on the following factors:

1) The role of pore fluid and associated, commonly retrograde, hydration reactions (White, 1976, p.82; Beach, 1980; Vernon et al., 1983, p.142)
2) The mineralogy of the rock being sheared (White, 1973, 1976, 1977; Vernon, 1974, 1976; Lister and Price, 1978; Dixon and Williams, 1983; Vernon et al., op.cit.)
3) Grain size (White, 1976, 1977; Beach, op.cit.; Vernon et al., op.cit.)
4) Strain rate versus recovery and recrystallisation (Vernon, 1976; White, 1976, 1977; Lister and Price, op.cit.; Beach, op.cit.)

The above-mentioned factors have received much attention in the literature and hence do not need to be repeated here. A qualitative approach is adopted for the study of mylonitic fabric development as many of the mylonitic specimens have a high proportion of hard minerals which causes heterogeneity of strain within the matrix (Lister and Price, op.cit.). It follows that the measurement of quartz c-axis fabrics is not warranted in this study and it is therefore omitted. The following points, varying from those observed on a mesoscopic scale to those on a microscopic scale, are noted to be of importance regarding the development of mylonitic fabrics:

1) Mapping of shear zones has revealed that lithology has a marked effect on the development of mylonitic fabrics. Lithologies in which there is a high proportion of hard minerals such as feldspar, garnet, amphibole or pyroxene, tend to have a limited bulk-ductility with deformation being controlled by the hard minerals. In contrast, lithologies with a high proportion of soft minerals such as quartz and mica, tend to deform in a predominantly plastic manner.
2) On a microscopic scale, quartz and mica behave in a very ductile manner and show considerably better alignment than the hard minerals.
3) Strain is concentrated in the matrix of the rock, which is most commonly comprised of quartz and mica. Strain seems most intense closest to porphyroclasts, on the sides facing the foliation.
4) Where quartz has recrystallised, it commonly forms fine-grained laminae that define the foliation and which are concentrated with mica. It is uncertain whether the mica has recrystallised, but it most likely has as this would account for its concentration within the laminae.
5) In coarse-grained megacrystic rocks such as the megacrystic granitic gneiss (HM 21) the mylonitic foliation anastomoses about the coarse megacrysts which have been flattened, resulting in an augen-gneiss texture.
6) Quartz is very sensitive to strain, the first noticeable effect being the development of dislocations which are marked by bands of undulose extinction. The dislocations form elongate sub-grains.

In order to illustrate the effect that an increasing proportion of hard minerals in the sheared rock has on the resultant mylonitic texture, a concept that has not been adequately described in the literature, a schematic model has been devised using selected examples of mylonitic rocks in illustration. The following points have been considered in order to devise the model:

1)In monomineralic rocks the distinction between brittle (cataclastic) and ductile (mylonitic) deformation during shearing is a simple matter. In polymineralic rocks the accepted way of distinguishing the two is by examining the behaviour of only the matrix (White et al., 1980). This distinction, however, gives no consideration to the bulk ductility of the rock and hence there is a need to show how the development of the mylonite depends on the proportion of the soft minerals in the rock.

2)The relationship between temperature and strain rate is important. At high strain rates and low temperatures the dislocation density in the minerals will be high, resulting in small sub-grains developing due to recovery (White,1977). The dislocations inhibit grain growth and hence recrystallisation results in grain refinement (White,1976). At relatively high temperatures and low strain rates, cross-slip and climb of dislocations enable the rate of recovery to keep pace with the development of dislocations and thus work-hardening is overcome (Vernon,1976). Watterson (1979) has recently pointed out that in many high-temperature shear zones the rocks become very deformed, but do not show a marked grain refinement.

3)The temperature of recrystallisation of a mineral has been found to decrease with decreasing grain size and increasing strain (Spry, 1969, Figs.36,37). It follows that during shearing, with temperatures being below that at which grain growth predominates, an increase in shear strain and a concomitant decrease in grain size will occur. This results in the required temperature of recrystallisation for a given mineral being lowered with progressive deformation and hence the rate of recrystallisation will increase. Thus, the slope of the boundary separating the field of grain growth from grain refinement must have a negative slope (Fig.23), decreasing in temperature with increasing matrix proportion (and strain).

4)The proportion of recrystallised grains increases with strain until a "steady state" microstructure is attained (White et al., 1980). For quartz mylonite, as strain increases the coarse-grained quartzite is converted into a protomylonite, then a mylonite and finally an ultramylonite (ibid.).

5)White (1976,1977) has shown that for a grain size less than 100 µm the deformation mechanism changes from dislocation creep to Coble creep which occurs more rapidly and can render the rock effectively superplastic.

Taking the above-mentioned points into account Fig.23 shows a model in which mylonitic textures are shown as a function of temperature and matrix proportion. By assuming that the matrix proportion increases with progressive shearing, except in the case of high-temperature deformation where grain growth dominates, the model enables mylonitic fabrics to be regarded in terms of deformation versus temperature.

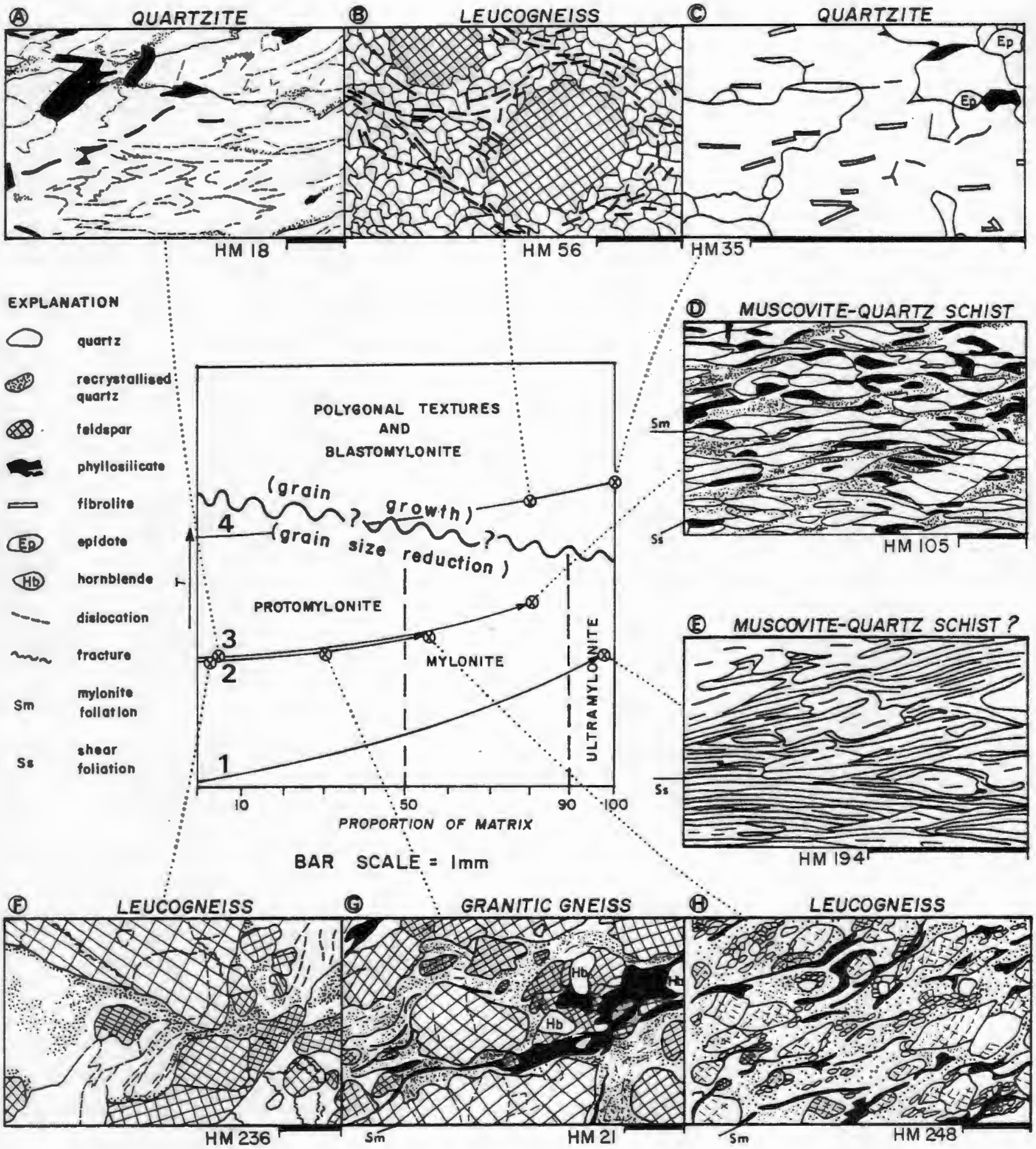


Fig.23:Selected textures (A-H) from various lithologies showing a progressive development of mylonitic and polygonal textures. The textures are shown as a function of matrix proportion (which increases with progressive shear strain, recovery, and recrystallisation) and temperature.

The following points require explanation in order to show how the model accounts for the development of the selected textures (A-H) shown in Fig.23:

1)All sheared rocks are assumed to have begun their shear history with no matrix - that is, no grain-size refinement had yet started as a consequence of deformation, even though soft minerals existed in the rock.

2)Quartz-rich rocks are the most sensitive to strain. In quartz, low angle sub-grain boundaries develop which are highlighted by bands of undulose extinction (Fig.23a).

3)Soft minerals in the rock tend to recrystallise into "matrix" grains and hence the proportion of soft minerals in the rock prior to shearing strongly influences the proportion of matrix that will develop in a mylonite. It follows that quartz-rich rocks which undergo grain-size reduction, soon develop abundant matrix and hence usually have either a mylonite or ultramylonite texture (Fig.23d and 23e), but rarely protomylonite. Note that curves 2 and 3 are separated since the former depicts a deformation "path" for the quartzo-feldspathic rocks having an abundant hard mineral content, whilst the latter depicts a deformation path for quartz-rich rocks. In contrast to the rapid development of mylonite and ultramylonite textures for the quartz-rich rocks, the quartzo-feldspathic rocks, rich in hard feldspar, rarely leave the protomylonite field even though the strain for these rocks is probably comparable to that for the quartz-rich rocks.

4)The texture shown in Fig.23e has been included since it is mylonitic as it is cohesive, even though the sample comes from a fault (Riedel fracture) that is thought to post-date the mylonite fabric of the Pofadder Lineament. It would appear to have deformed at a lower temperature than for curve 3, as shown by curve 1.

5)Lastly it is postulated that temperature was high enough in some cases for the deformation path to enter the field of grain growth, as shown by curve 4. The texture shown in Fig.23b is of a blastomylonitic leucogneiss which is polygonal, and some coarse grains (porphyroblasts?) appear to show no sign of strain. The rock texture appears to have overprinted an older foliation defined by biotite which wraps around the porphyroblasts. The sample comes from area 29°01'20"S/ 19°43'10"E in which sinistral shears are seen to refoliate a dextral mylonite belt into narrow bands which are less than half a metre in width. It is therefore thought that the biotite foliation could be associated with the dextral shearing and that during subsequent sinistral shearing grain growth kept pace with the strain rate, resulting in the develoment of a blastomylonite.

In Fig.23c a polygonal textured rock from a mylonitised quartzite is shown. The rock is associated with sheared ortho-amphibolite on the farm Nouzees (locality 29°14'10"S/19°31'30"E), and yet no sign of strain can be noticed, indicating that the rate of recrystallisation kept pace with, and possibly exceeded, the strain rate.

According to White (1977) and White <u>et al.</u> (1980) mylonitic rocks tend to deform in a superplastic manner when their grain size has been reduced to less than 100 μm. In terms of the above-mentioned model, mylonitic rocks of the finest grain size will tend to plot within the ultramylonite field since grain-size reduction is accompanied by an increase in the proportion of matrix. As a consequence it is deduced that superplasticity is most likely to occur in rocks that are sheared at relatively low temperatures and which have a high proportion of soft minerals.

Finally, the model illustrates that in situations where the lithology is the only variable, such as on a mesoscopic scale; it is the rocks which have the highest proportion of soft minerals that will show the most pronounced mylonitic texture. As a consequence, cognizance must be taken of the differing degree of textural development in sheared rocks, and when mapping shear zones, the lithologies that have a relatively high proportion of hard minerals should not be underestimated in their degree of strain.

An interesting aspect is that blastomylonitic rocks occur in the northwest of the area (for example HM 56, Fig.23b) on the farms Gemsbok Vlakte and Lucas Vlei Vlakte. Since mafic granulite occurs on the farm Gemsbok Vlakte (see Chapter II) this area was most likely one of relatively high temperature during the peak of metamorphism, and possibly also an area of higher temperature during the subsequent period of shearing. Thus the development of blastomylonites in the northwest is in good accord with the metamorphic evidence that temperatures were higher there than elsewhere during deformation.

Chapter VI

CONCLUSIONS

The present study has had an emphasis on the deformational history of the field area, and in particular the role of shearing.

Few conclusions have been made regarding the stratigraphy of the area. A series of paragneisses have been mapped which constitute the Bushmanland Group, and they consist of quartzite, muscovite-quartz schist, biotite schist, calc-silicate gneiss, several varieties of quartzo-feldspathic rocks, para-amphibolite and pelitic gneiss. The basement upon which these rocks were deposited has not been identified. In one locality a sharp contact between an alkali-feldspar granitic gneiss and an overlying quartzo-feldspathic gneiss was observed, but the contact appears to be structural. No contacts with pre-tectonic gabbronorite rocks were observed and the pre-tectonic mafic granulite clearly intrudes the Bushmanland Group rocks. It is interesting to note that there is a difference in the mineral mode between the pre-tectonic granitic gneiss and the syn to post-tectonic granitoids and diorites. On a quartz - alkali-feldspar - plagioclase projection scheme (Streckeisen, 1973) the former rocks plot as alkali-feldspar granitic gneisses whilst the latter plot as granodiorite-tonalite and diorite rocks. A final conclusion regarding lithology is that the post-tectonic kimberlite pipes have utilised deep-seated structures during their emplacement, as seen by their localisation along the Nouzees shear.

From the study of metamorphism, several geothermometers and one geobarometer were used besides a study of the metamorphic petrography. It is found that the P-T conditions could be more accurately derived on the basis of the petrography, namely the petrogenetic grid, than from the geothermometers and geobarometer. It is concluded that temperatures of 625 - 675°C and pressures of approximately 4,5 kb existed during the peak of metamorphism, and that these conditions straddled the boundary between medium and high grade metamorphism. The geothermometry does, however, show that in the gabbronorite and mafic granulite the pyroxene exsolution records a higher temperature of equilibrium. This temperature is within the range of 830 - 990°C (see Table 18), possibly representing a sub-solidus temperature, and it is assumed to approximate the crystallisation temperature. The peak of metamorphism is regarded to have occurred during, or just after, the development of the regional foliation (S2) and that during subsequent deformation retrogression occurred. The P-T conditions for the retrograde metamorphism could not be precisely established, but appear to have been within the low-temperature range of low-grade metamorphism (greenschist facies).

In the structural synthesis, five stages of deformation have been recognised, namely D1 to D5. These stages of deformation represent the sequence in which the fabrics and structures formed, and in order for them to be of tectonic significance, they have been grouped into three phases of deformation, each phase having a different controlling process. A detailed description of these phases has been given in the structural synthesis

(Chapter IV) and only the main conclusions will be discussed. The first phase of deformation involved intense folding of the oldest recognisable surface, namely bedding or its transposed equivalent (S0/1). Two sets of folds were formed in this phase (F1 and F2) and there is no reason why a single deformation cannot produce two stages of folding. To substantiate this, Williams (1983) has shown that if early perturbations existed in the strata they could have caused early folds to develop, which with continued deformation become refolded. This phase of deformation is thought to have culminated in the development of an axial-planar foliation (S2) and a well-developed mineral lineation (L2) at, or just prior to, the peak of metamorphism. An important result of this deformation phase is that the L2 lineation caused a marked linear anisotropism to develop in the rock which controlled subsequent deformation. The L2 lineations generally plunge gently to the east-northeast.

The second phase of deformation is considered to have involved folding that was not as intense as that of the first phase, and these folds formed by means of flexural-slip along the S2 surface. Two generations of folds formed during this phase of deformation, namely F3a and F3b. The F3b folds are common and have a gentle plunge towards the east-northeast. The F3a folds were not identified in the field, but they are structurally necessary in order to yield the dome-and-basin fold interference pattern in the area. There are two reasons why the F3a folds are necessary. Firstly, the regional foliation (S2) defines the fold interference pattern and hence these folds (and the superimposed folds) must post-date both the S2 foliation and the F2 folds. Secondly, these folds must be cross-folded by the F3b folds as the latter have gently plunging fold axes and hence a steeply inclined "a-direction" of movement, a condition which is required (Ramsay,1967) if superimposed folding is to result in a dome-and-basin pattern. In regard to the shape of the dome-and-basin pattern the F3a folds are considered to have a northwest trend. Little is known about the kinematics of this fold event, but the folds could be associated with thrusting in a west-southwest direction, pre-dating the F3b folding.

During the late stages of F3b folding, shearing developed along the northern limbs of the synforms, and the third phase of deformation is considered to have followed, being characterised by shearing. The two deformation phases are, however, regarded as being separate since it is postulated that flexural-slip folding controlled the deformation of the former phase whereas the shearing, which increased in intensity, controlled the deformation during the latter phase. It is thought that when the shearing intensified, the locally developed shears which occurred along the northern limbs of the F3b synforms merged by means of listric shears into an east-southeast strike and formed continuous shears in that direction. These shears have a dextral sense of movement and are characterised by the Pofadder Lineament which is of megascopic scale, extending from Lüderitz to the present study area, and probably as far to the southeast as Vanwyksvlei. The length of this shear is therefore considered to be 800 km. The displacement across the lineament is not considered to be as high as 85 km, as calculated by Toogood (1976), but rather to be within the range of 45 to 60 km. This estimate is made by correlating the synform near Pella, which occurs to the southwest of the lineament (Joubert,1973) with the Longziek Vlei synform to the northeast.

On the basis that the mylonitic lineations plunge gently to the east-northeast, and that the lineament has a dextral sense of displacement, the lineament is considered to have had a small north-down component of movement. It has been reported, however, that the rocks are of higher grade to the north of the lineament and that this is presumably the result of crustal uplift on that side (Toogood,1976). It is suggested that since numerous mafic intrusives are present both within and to the north of the lineament, that the higher metamorphic grade to the north is not the result of crustal uplift associated with the Pofadder Lineament, but possibly the effect of contact metamorphism on a local scale. Within the study area, high grade metamorphic rocks occur locally on both sides of the lineament, namely the mafic granulite to the north on the farm Gemsbok Vlakte and the pelitic gneisses to the south in close proximity to the Nouzees Gabbronorite Suite. Since these high grade metamorphic rocks occur on either side of the lineament and also because they seem to be associated with mafic magmatism, the concept of contact metamorphism does seem to be favourable, especially since the metamorphism was of relatively high temperature and low pressure (approximately 650°C and 4,5 kb). Alternatively, the higher metamorphic grade to the north could be the result of west-southwest-directed thrusting which has been shown to pre-date the lineament. In this context, the lineament, being a zone of predominantly strike-slip movement, would be required to have reactivated a large thrust in order to explain the presence of the higher metamorphic grade to its north. This idea is not favoured though as thrusts are not likely to be laterally continuous over the entire length of the Pofadder Lineament.

The Mattheusgat Mountains represent the southeastern-most part in which the effects of the Pofadder Lineament are well exposed. Extending from these mountains to the southeast of the present study area, exposure along the extension of the lineament is poor and it is exposed only as individual shears, namely the Nouzees, Sandgat and Adjoining Geelvloer shears. The Sandgat shear and a sinistral east-northeast - striking shear, namely the Putsberg shear, intersect, and evidence suggests that they form a conjugate shear pair. In this regard several important conclusions have been drawn. Firstly, the east and west directed quadrants between the intersecting shears are expected to have been areas of overall tension and this is substantiated by the presence of the syn to post-tectonic T'Oubep Granodiorite Suite being located within the eastern quadrant. Secondly, since the S2 foliation and the L2 mineral lineation are present in the granodiorite, and if it is assumed that the rock was emplaced during the conjugate shearing, it can be concluded that the stress system which resulted in the conjugate shearing (D4) probably also accounted for the D2 and D3 stages of deformation. It is postulated that folding associated with the Pofadder Lineament, which is the most intense shear in the area, occurred in two ways. As the lineament is approached, the increasing shear strain resulted in a tightening of the large-scale F3b folds as well as drag folding about northwest-trending F4 fold axes. Since the F4 folds are the result of strike-slip movement, they are best developed in steeply dipping surfaces and their trend is therefore expected to be highly variable.

An important feature in regard to the three deformation phases is that the parallelism between the L2 mineral lineations, mylonitic lineations in thrust surfaces (L3a?), F3b fold axes and the L4 mylonitic lineations in shear zones is a widespread phenomenon throughout the present study area. To recap, L2 lineations which formed just prior to, or at, the peak of metamorphism resulted in a strong linear anisotropism developing in the rocks and it is thought that this resulted in subsequent linear structures and fabrics developing parallel to the L2 lineations. From a kinematic point of view, the parallelism suggests that a single stress field existed during the ductile deformation.

From the fracture analysis it has been shown that the stress system responsible for the fracturing was of a north-south direction, and it thus had the same direction as that which resulted in the conjugate shearing. The argument that the stress field remained the same from the D2 to the D4 stage of deformation may now be extended to include D5. Thus although little is known about the earliest stage of deformation (D1), all the remaining stages seem to be characterised by a single stress field. The fracture analysis indicates that the movement which occurred during the ductile (shearing) deformation continued into a time when brittle deformation prevailed, and that a wrench zone developed. Prominent quartz veins mark the presence of large-scale Riedel fractures which are localised along the lineament and which transect the mylonite foliation. The movement was active during the Carboniferous as the fractures have affected the rocks of the Dwyka Formation, and seismic evidence shows that movement is still active along the lineament.

Finally, the study of the development of mylonitic fabrics indicates that the final fabric is highly dependant on the mineral mode of the rock. For a given set of P-T conditions a lithology that is comprised predominantly of soft minerals such as quartz and mica will have a higher bulk ductility than a lithology with few soft minerals, and this will favour the development of a mylonite as opposed to a cataclasite. Thus the formation of a mylonite depends not only on the P-T conditions of shearing, but also on the lithology. Blastomylonites are considered to form instead of mylonites when the temperature is elevated to the extent that recrystallisation keeps pace with the strain rate, enabling grain growth to dominate over the process of grain-size reduction.

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APPENDICES

Appendix I: Microprobe analyses of clino and orthopyroxene mineral pairs in the Nouzees Gabbronorite Suite (specimen HM7) and a mafic granulite (specimen HM261). Ferric iron determined on a cation charge basis (Ryburn et al., 1975). Allocation of cations to structural sites is done according to Wood and Banno (1973), and Powell (1978).

Specimen	HM 7	HM 7	HM 7	HM 7	HM 7	HM261	HM261	HM261	HM261
Mineral Pair	(i)	(i)	(ii)	(ii)	(ii) rim	(i)	(i)	(ii)	(ii)
Mineral	cpx	opx	cpx	opx	opx	cpx	opx	cpx	opx
SiO_2	50,50	54,08	53,36	52,87	53,71	52,94	51,73	52,60	51,60
TiO_2	0,23	-	0,19	0,76	-	0,04	-	0,11	0,04
Al_2O_3	2,48	1,76	1,90	1,40	1,41	1,26	0,76	1,21	0,75
Cr_2O_3	-	-	-	-	-	-	-	0,04	-
FeO	6,15	16,56	6,53	17,80	17,02	9,56	26,45	9,74	27,97
MnO	0,17	0,39	0,22	0,46	0,45	0,29	0,68	0,26	0,62
MgO	16,08	27,00	15,81	25,51	26,27	14,23	19,32	13,44	18,47
CaO	21,71	0,36	21,79	0,93	1,45	21,02	0,83	22,70	0,62
Na_2O	0,37	-	0,56	-	-	0,27	-	0,26	-
K_2O	-	-	-	-	-	-	-	-	-
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TOTAL	99,69	100,15	100,36	99,73	100,31	99,61	99,77	100,36	100,07
- - - ATOMIC PROPORTIONS BASED ON 12 OXYGENS - - -									
Tetrahedral sites									
Si	1,937	1,953	1,958	1,938	1,949	1,980	1,973	1,966	1,975
Al	0,063	0,047	0,042	0,060	0,051	0,020	0,027	0,034	0,025
				(0,002)**					
M1 sites									
Al	0,045	0,028	0,040	-	0,009	0,036	0,007	0,019	0,009
Cr	-	-	-	-	-	-	-	0,001	-
Ti	0,006	-	0,005	0,021	-	0,001	-	0,003	0,001
Fe^{3+}	0,038	0,019	0,032	0,020	0,042	0,002	0,020	0,027	0,014
Fe^{2+}	0,134	0,237	0,150	0,262	0,238	0,262	0,417	0,256	0,444
Mg	0,777	0,716	0,773	0,697	0,711	0,699	0,556	0,694	0,532
M2 sites									
Fe^{2+}	0,018	0,244	0,018	0,262	0,237	0,035	0,407	0,021	0,437
Mg	0,107	0,737	0,092	0,696	0,710	0,094	0,542	0,055	0,552
Na	0,026	-	0,040	-	-	0,020	-	0,019	-
Ca	0,858	0,014	0,857	0,037	0,056	0,842	0,034	0,909	0,025
Mn	0,005	0,012	0,007	0,014	0,014	0,009	0,022	0,008	0,020
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SUM	4,014	4,007	4,014	4,009	4,017	4,000	4,005	4,012	4,004

Analyst R. O. Moore, 1983

Specimen locality: HM 7 29°14'00"S/19°30'20"

HM 261 29°01'40"S/19°35'30"E

** Ferric Iron used to fill the tetrahedral site

Appendix II: Microprobe analyses of coexisting cordierite, garnet and biotite in pelitic gneiss for specimen HM 25.

	AREA ONE			AREA TWO		
	Cordierite	Garnet	Biotite	Cordierite	Garnet	Biotite
SiO_2	48,40	37,15	35,01	48,94	37,54	35,24
TiO_2	-	-	2,59	-	0,06	1,74
Al_2O_3	33,60	21,79	19,82	32,40	20,51	19,77
Cr_2O_3	-	-	-	-	-	-
FeO	9,45	34,51	21,93	9,40	34,71	21,11
MnO	0,18	2,34	-	0,17	2,56	0,04
MgO	7,48	3,21	8,70	7,88	2,81	8,56
CaO	-	1,63	0,05	0,05	1,40	0,09
Na_2O	0,12	-	0,22	0,12	-	0,28
K_2O	-	-	8,58	-	-	8,69
TOTAL	99,23	100,63	96,90	98,96	99,59	95,52

	- - ATOMIC PROPORTIONS BASED ON SELECTED NUMBER OF OXYGENS - -					
OXYGEN	18	12	24	18	12	24
Si	4,961	2,967	5,753	5,031	3,037	5,857
Ti	-	-	0,320	-	0,004	0,217
Al	4,059	2,051	3,839	3,926	1,956	3,873
Cr	-	-	-	-	-	-
Fe (total)	0,810	2,305	3,014	0,808	2,349	2,934
Mn	0,016	0,158	-	0,015	0,175	0,006
Mg	1,143	0,382	2,131	1,207	0,339	2,120
Ca	-	0,139	0,009	0,006	0,121	0,016
Na	0,024	-	0,070	0,024	-	0,090
K	-	-	1,799	-	-	1,843
SUM	11,013	8,002	16,935	11,017	7,981	16,956

Analyst: R. O. Moore, 1983

Sample locality: 29°16'10"S/19°34'10"E

Appendix III: Microprobe analysis of garnet in calc-silicate gneiss for specimen HM 196. The average of two analyses is shown and atomic proportions are based on 12 oxygens.

OXIDES		ATOMIC PROPORTIONS	
SiO_2	35,69	Si	3,164
TiO_2	0,47	Ti	0,031
Al_2O_3	3,88	Al	0,405
Cr_2O_3	0,04	Cr	0,003
FeO	25,01	Fe (total)	1,854
MnO	1,42	Mn	0,107
MgO	0,12	Mg	0,016
CaO	31,79	Ca	3,020
Na_2O	-	Na	-
K_2O	-	K	-
	-----		-----
TOTAL	98,42	SUM	8,600

Analyst: R. O. Moore, 1983

Sample locality: 29°18'40"S/20°05'30"S

PLATE 1: An overturned bedding surface in calc-silicate gneiss along the northern margin of the Adjoining Geelvloer shear (29°18'S/20°04'E).

PLATE 2: Overturned trough cross-bedding in quartz-muscovite schist within the Pofadder Lineament (29°09'45"S/19°39'00"E).

PLATE 3: Hornblende-mineral aggregates with plagioclase-rich haloes in cluster amphibolite on the farm Brul Kolk (29°11'30"S/20°09'00"E).

PLATE 4: Tight Z-asymmetric F3b folds defined by biotite and feldspar mineral bands. View is eastward (29°00'S/19°36'E).

PLATE 5: Northerly view of an S-asymmetric F4 fold in mylonite on the farm Pofadder East (29°05'40"S/19°32'50"E).

PLATE 6: A possible F3a/F3b fold interference pattern in megacrystic alkali-feldspar granitic gneiss on the farm Lucas Vlei Vlakte (29°01'30"S/19°43'39"E).

PLATE 7: Northeasterly aerial view of the eastern part of the Mattheusgat Mountains. Note the ribbon-like mylonite bands along strike as well as the rotation and divergence of these bands to the east.

Mkm = megacrystic leucogneiss
Mfl = quartzo-feldspathic gneiss
Mms = muscovite-quartz schist
Mq = glassy quartzite
Jd = Karoo dolerite


GEOLOGICAL MAP
OF THE AREA
EAST OF POFADDER
QUARTERNARY
Alluvial Sand
Deposits in Pans
Aeolian Sand
Calcrete
Sand and superficial cover
Gravel of Dwyka Fm
Float of Dwyka Fm
CRETACEOUS
Kimberlite Pipe (Kkp possible Kimberlite Pipe)
JURASSIC
Karoo Dolerite
CARBONIFEROUS
Dwyka Formation
Shale
Tillite
NAMIBIAN
Vein Quartz
Pegmatite
MOKOLIAN
POST TO SYN-TECTONIC
Fine-grained Quartz-plagioclase Porphyry
Pyroxene-bearing Diorite
Granodiorite-tonalite
PRE-TECTONIC
BUSHMANLAND GROUP (NOT IN STRATIGRAPHIC ORDER)
Glassy Quartzite
Granular Quartzite
Iron Formation
Muscovite-quartz Schist
Biotite Schist
Calc-silicate Gneiss
Marble
Pyroxenite
Calc-silicate Conglomerate
Quartzo-feldspathic Gneiss
Megacrystic Quartzo-feldspathic Gneiss
Leptite
Leucogneiss
Megacrystic Leucogneiss
Para-amphibolite Gneiss
Pelitic Gneiss
NOT IN INTRUSIVE ORDER
Red Alkali-feldspar Granite Gneiss (porphyritic in places)
Megacrystic Alkali-feldspar Granite Gneiss
Streaky Gneiss
Olivine Gabbronorite
Ortho-amphibolite
Cluster Amphibolite
Mafic Granulite and Ortho-amphibolite undifferentiated
SCALE 1:100 000
0 1 2 3 4 5 6 7 8 9 10km
KEY
Strike and dip of F_2 and F_3 fold axial planes
Trend and plunge of
A) (a) tight to isoclinal folds
(b) close folds
(c) gentle to open folds
B) (a) F_1 folds
(b) F_2 folds
(c) F_3 folds
Trend and plunge for (a) mineral and (b) intersection lineations
Strike and dip of bedding (inclined and vertical)
Strike and dip of S_2 foliation (inclined and vertical)
Strike and dip of S_4 foliation (inclined and vertical) with sense of movement shown if ascertained
Strike and dip of fractures (inclined and vertical)
Lithological contact or extent of outcrop
Extent of sporadic outcrop
Lines of strike taken from aerial photographs
Rocks possessing a mylonite fabric
Fault trace
Trace of listric shear zone
S-asymmetric fold
Z-asymmetric fold
Symmetrical fold
Trigonometric beacon
Dry water-course
Farm boundary
Main road
Secondary road
Farm road
Power-line
Homestead
LOCALITY MAP
Garies
Kliprand
Bitterfontein
REPUBLIC OF SOUTH AFRICA
Cape Town
PRU
pЄ
Compiled by A H Maclaren. Precambrian Research Unit for Geological Survey
Drawn by R M E Kovats 1984
KONKOONSIES
NONGCAIP
GEMSBOK VLAKTE
KABAS
MATTHEUS-GAT
POFADDER EAST
MATTHEUSGAT MTNS
VAALPUTS
NOUZEES
DE NEUS
SAND-GAT
LUCAS VLEI
LUCAS VLEI VLAKTE
FALS-VLEI
BLAD GROND SOUTH
ARRIBEES
BANK-VLEI
DROOGE GROND
VOGELSTRUIS LEEGTE
BRUL KOLK
LONGZIEK-VLEI
HARTEBEEST-VLEI
GRAAF-WATER
T'OUBEP
HOUMOED MTNS
HOUMOED
KLEIN HOUMOED
LOVEDALE
DUSSA MAAG
GANNA-POORT
VAN TITTENS VILLE
PUTS-BERG
STEENKAMPS-VLEI
VOLMOED
WILLEMS-OPDAM
VAALKOP
GEELVLOER
SAND KOLKJES
GRAS KOPJES
TIMAS-DOS
KEMPENS KRAAL
KORT DRAAI
SWANEPOELS PUTS
VRAA WEER
GRAPPIES
T'CAIMOEPS LAAGTE
T'CAIMOEP
LOOS KOLKJES
GEEL VLOER
To Pofadder
To Kakamas
To Bladgrond
To Kliprand
19°30'
19°45'
20°00'
20°15'
29°00'
29°15'
29°30'