

**THE GEOLOGY OF THE MT. MARE AREA,
PIETERSBURG GREENSTONE BELT,
SOUTH AFRICA.**

by

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ABSTRACT

This thesis is a study of the geological evolution of Archaean supracrustal rocks from the Mt. Mare area, Pietersburg greenstone belt, South Africa. Combined field, microstructural and geochemical data indicate the area is a poly-deformed (D1-D2), poly-metamorphosed (M1-M2) tectono-stratigraphic terrain comprised of older simatic rocks and BIF and younger detrital sediments.

Circa 3.4 Ga simatic rocks and BIF horizons constitute the tectonically dismembered remnants of a single, 3.5–10 km thick, intrusive-extrusive igneous complex. The extrusive rocks were derived from deeper level intrusive rocks by low pressure crystal fractionation and/or partial melting. This was concurrent with regional-scale, high temperature (upper amphibolite to lower greenschist facies), low pressure (<3 kb), retrograde, open-system metasomatism (M1). Metasomatic addition of Mg to the simatic pile and loss of SiO₂ and Fe, may have mimicked primary igneous differentiation trends. Syndeformational magmato-metasomatic interaction was responsible for the development of spinifex-textured amphiboles in cross-cutting fibrous extension veins or dykes.

Subsequent development of a northwest verging, foreland propagating, imbricate fan fault system (D2) lead to the tectonic stacking of emergent thrust sheets of the igneous complex and syn-D2 alluvial 'piggy-back' basins in a 'foreland basin' environment. Early D2, near surface faults were characterized by syn-sedimentary deformation, fluidization and a gradation from tectonic to sedimentary processes. In the Snymansdrift Shear Zone the gradation is marked by a distinctive tectono-sedimentary melange formed by progressive, brittle-ductile deformation of BIF units. Analysis of clast orientation data in the melange indicates the kinematic vorticity number, $W_k = 0.76-0.8$ and that the deformation path involved both pure and simple shear. Thus, continued D2 deformation lead to the sedimentary basins becoming steepened and flattened to form shear zones.

Syn-D2 dynamothermal metamorphism (M2) which occurred in the brittle–semi-brittle field, caused development of chlorite-quartz and sericite-quartz phyllonites by the progressive homogenization of the textural, chemical and metamorphic differences between precursor lithologies. Syn-D2 chloritoid and clinozoisite porphyroblast growth suggest greenschist facies metamorphic conditions (350–500°C, 3 kb *maximum*) were attained during M2.

The D2 faults were characterized by stick-slip fault movement, closely associated with episodic hydraulic fracturing and fluid release from deeper crustal levels to a near surface environment. This tectono-metasomatic activity produced a distinctive fuchsite–carbonate–quartz gneiss in two of the shear zones. One such fluid flow event was mineralizing and precipitated quartz–tourmaline–arsenopyrite–pyrite–gold veins which were subsequently deformed and are only preserved in low strain lenses.

CHAPTER 1

INTRODUCTION

1.1 INTRODUCTION

This thesis is a study of the stratigraphy, structure and geochemistry of Archaean supracrustal rocks in the Mt. Mare area, Pietersburg Greenstone Belt, located in the northern half of the Kaapvaal Craton, South Africa (Fig. 1.1). The belt is one of six major greenstone belts found on the craton and it lies subparallel to, and 50 to 100km south, of the Limpopo Mobile Belt which separates the Kaapvaal and Zimbabwean Cratons. The Pietersburg area has not been extensively studied in the past and hence it provides an ideal opportunity to investigate problems in Archaean greenstone belt evolution without being prejudiced by preconcieved ideas.

1.2 GEOLOGICAL CHARACTERISTICS OF ARCHAEOAN ROCKS

Archaean rocks are exposed in at least small areas on all of the continents and they comprise 3.8 to 2.5 Ga crustal provinces that are roughly equidimensional in plan view and vary in size from less than 0.1 up to 2.6 million km² in area (Condie 1981).

Three rock associations occur in Archaean provinces which in order of decreasing abundance are; the granite-greenstone association, the high grade metamorphic association and the cratonic basin association (Condie 1981). The granite-greenstone association is characterized by supracrustal successions comprising mafic-ultramafic igneous rocks and subsidiary sediments engulfed by various granitic rocks. This association dominates in Archaean provinces in North America, southern Africa, and Australia. The high grade metamorphic association, which dominates Archaean provinces in central and northern Africa, Antarctica, Greenland and the Soviet Union is characterized by gneiss-migmatite-granulite complexes, layered intrusions and highly metamorpho ed supracrustal remnants. The cratonic basin a sociation has mainly been described in the Kaapvaal craton in South Africa where the 2.7-3.0 Ga Witwatersrand Basin and 3.0 Ga

Pongola Supergroup are found. Both successions are composed dominantly of quartzites, shales and carbonates with smaller amounts of volcanic rock (Condie 1981).

Boundaries of Archaean crustal provinces fall in to one or a combination of three categories; unconformities with younger sedimentary terrains, rapid increases in metamorphic grade, and tectonic contacts. Increases in metamorphic grade and tectonic contacts characterize boundaries with Archaean-Proterozoic mobile belts.

1.3 CURRENT PROBLEMS IN GRANITE-GREENSTONE TERRAINS

Archaean rocks have become increasingly well documented in the past 25-30 years and their study is of particular importance because they provide constraints on the origin and early development of the Earth's lithosphere and by inference the core, mantle, hydrosphere and biosphere. A central question in Archaean geology is to what extent were crustal processes in the Archaean different to those that operate at the present time? Were the processes themselves fundamentally different or is it that they were similar, but perhaps modified in terms of the rate they operated at or their distribution in the Earth's structure?

Until recently, most greenstone belts were thought of as extremely thick (up to 20-30 km) stratigraphic sections of mafic-ultramafic lavas with lesser sediments and felsic volcanics, all metamorphosed to greenschist and/or amphibolite facies and engulfed in a sea of intruded granitoid plutons. Their overall shape and tectonic style were thought to be due to density inversions and diapirism, the inferred result of an unstable configuration of sialic crust overlain by a dense simatic layer. Hence vertical tectonic processes were thought to dominate Archaean crustal evolution. However, these early models were proposed before the plate tectonic revolution had had an impact on the interpretation of crustal development, and by that time they had become well entrenched in the literature. This has led many writers to infer that crustal processes in the Archaean were different to the present time. This apparent difference was accentuated because despite inter-cratonic variation there is a very generalized similarity in the stratigraphy, structure, metamorphism, mineralization, associated granite and geotectonic setting of greenstone belts around the world. Hence they could all be explained by one all encompassing model despite the fact that they varied in age by up to one billion years (de Wit and Ashwal 1987).

The vertical tectonic models were greatly influenced by the "Barberton Model" of Anhaeusser (1973, 1975, 1986) based on studies in the Barberton Greenstone Belt, South Africa as a part of the Upper Mantle Project (Viljoen and Viljoen 1969a). Two of the basic assumptions of this model are that the preserved 22km of stratigraphic thickness is tectonically intact and that the majority of the simatic rocks are extrusive. Ramsey (1963), Williams and Furnell (1979), de Wit (1982, 1983), de Wit *et al.* (1982, 1987a,b), Paris (1984), Lowe *et al.* (1985), Lamb (1986), have all presented field evidence that shows these basic assumptions require re-evaluation; a significant part of the simatic sequence could in fact be intrusive and there is structural repetition throughout the entire apparent stratigraphy. These observations clearly have profound implications for studies of crustal evolution in the Archaean.

There is now growing evidence that greenstone belts record extremely complex geological histories that can in fact be explained within the framework of modern theories of plate tectonics. For instance, some greenstone belts have now been interpreted as ophiolites (Helmstadt and Padgham 1986, Taylor 1986, Kusky 1986, de Wit *et al.* 1987a), and the role of thrust and nappe tectonics is becoming increasingly well documented (eg Coward 1976, de Wit 1982, Lamb 1984, 1987, Paris 1984, Spray 1985, 1986). In order to establish whether or not such interpretations are valid it is necessary to address the following problems which are central to any study of greenstone belt evolution; Firstly, what processes resulted in the development of the simatic and related felsic rocks in the belt? Secondly, what processes affected the associated sediments? Thirdly, what are the geochemical and structural relationships between greenstone belts and the surrounding granitoid terrains? Finally, given answers to the previous problems, what tectonic environments are found in granite-greenstone terrains?

It is evident that these problems are interrelated and in many cases interdependent suggesting that one of the reasons for the complexity that clearly does exist in greenstone belts is that more than one geological process was operating at any one time and place in their evolution. It is probably because of this overlap that the geological evidence concerning timing relationships in greenstone belt history is so often incomplete or ambiguous and this in turn is responsible for the conflicting interpretations of their formation.

1.4 PREVIOUS WORK IN THE PIETERSBURG GREENSTONE BELT

As a result of the discovery of gold on the farm Eersteling 17KS (Fig 1.2) in 1871 (Willemse 1938), the south west portion of the Pietersburg belt received early attention in the geological literature. However, little work has been done in the area since the late 1970's and the report by SACS (1980) is the only work to have formally classified the rocks in the belt as a whole. The rocks were named the Pietersburg Group and this consists of six formations (Table 1.1, Fig 1.2). SACS (1980) recognised a basal sequence of predominantly ultramafic volcanic rocks interbedded with various sialic lavas and banded iron formations (Mothiba and Ysterberg Formations) which were overlain by mafic lavas and various schists (Eersteling, Zandriverspoort and Vrischgewaagd Formations). This sequence was unconformably overlain by the coarse detrital sediments of the Uitkyk Formation and the entire Group was intruded by several different types of granitoid pluton.

Gotz (1887) was the first to study rocks from the Pietersburg belt. He described the petrographic features of rocks from the Mt. Mare and Eersteling areas but was unaware of their field relationships. He recorded the presence of actinolite, chlorite, quartz-tourmaline schists, ottrelite (manganese bearing chloritoid) and andalusite schists as well as phyllites and calico rock (colloquial term for iron formation). Gotz concluded that the rocks had,

"...undergone a more or less intense change in mineralogical compositionbrought about not merely by the chemical activity of circulating waters but also by pressure due to orogenic forces".

Molengraaf (1904) differed from Gotz's interpretation, noting the presence of a variety of schists in the Mt. Mare area which were,

"...greatly modified in their structure by contact metamorphism produced by granitic intrusions. In the contact zone the schists show an immense variety of types characterised by a large number of contact minerals such as staurolite, andalusite, ottrelite, garnet and corundum etc."

Hall (1909) produced the first geological cross sections of the Pietersburg belt in the Mt. Mare area. He described a stratigraphic section some 4450 ft. thick which was characterised by a basal sequence of various chloritic schists, sheared quartzites and conglomerates overlain by banded iron formation and collectively called the Lower Mt. Mare

series. This was overlain by the Middle Mt. Mare series, a sequence of chloritic schists and serpentinites capped by a second iron formation unit. The Upper Mt. Mare series, the uppermost exposed rocks, consisted of the andalusite and ottrelite schists (metasediments) overlain by coarse conglomerates, shales and finally a third iron formation unit. Hall noted that pervasive tourmalinisation had occurred in many of the lithologies.

The abundance of schistose rocks and the occurrence of identical iron formation units in the stratigraphy led Hall to consider the possibility of stratigraphic duplication by *isoclinal folding or thrusting*. The uniform dip of all the beds in the area, their variable thickness which was not repeated, plus the lack of any other field evidence led Hall to reject structural repetition in favour of a normal cyclical succession. He also favoured an intrusive contact between the surrounding granites and the greenstone belt, noting that the granites (which were also tourmaline bearing) became sheared themselves towards some of their contacts with the greenstones, but that the latter were also cut by undeformed granitic veins. The intrusive nature led Hall to propose a contact metamorphic origin for the mineral assemblages seen in the metasediments although the field evidence was equivocal. Hall also noted the similarities of the rock types in the Mt. Mare area, especially the conglomerates, with those in the Barberton Mountain Land and concluded they both belonged to the Swaziland System (Archaean).

Willemse (1938) produced the first regional map of the Pietersburg area concentrating on the south western portion of it. He divided the rock types seen in the greenstone belt into two groups, para-rocks (metamorphosed sediments) and ortho-rocks (metamorphosed igneous rocks). The para-rocks consisted of banded iron formation and detrital arenaceous sediments some of which had been metamorphosed and contained andalusite and ottrelite. Willemse noted the presence of,

"...conglomerates of an uncertain origin on the southern slopes of the Mt. Mare area",

which he assigned to the para-rocks. The ortho-rocks comprised schistose and massive varieties of ultrabasic rocks, amygdaloidal rocks and serpentinites. The iron formation and ortho-rocks were considered to be contemporaneous and these were unconformably overlain by the detrital sediments. Willemse, like Hall, interpreted the stratigraphy as being one of simple layer cake stratigraphy with no stratigraphic repetition. He also showed through his mapping that there had been considerable (approx 10km) strike-slip movement along a major fault that had offset the greenstone belt in a sinistral sense. De Wit (1985, 1987) called this fault the Willemse Shear Zone and it is marked by a poorly exposed schist zone

up to 1km wide. The faulting postulated a period of intrusive activity which resulted in the development of the granites.

Most of Willemse's (1938) paper deals with the gold deposits in the area. In his introductory statements Willemse makes the following important observations;

"During the final manifestations of igneous activity associated with the intrusion of thegranite, fissures formed in the Swaziland system and in the granite. These were filled predominantly with barren vein quartz, but occasionally they were reopened and sulphides, gold and other minerals were deposited in them. The metallization also extended in to fracture zones in the schist so that ore minerals need not be associated with large quantities of quartz".

Willemse makes it very clear he believed the mineralization was related to granitic activity but there was a strong structural control to the deposits overall, many of them being associated with fault zones. It is clear also from his observations that there was also repeated fluid flow through the faults and fractures.

Grobler (1972) remapped Willemse's field area and reported astonishing similarities with the Barberton Greenstone Belt which lead him to apply the 'Barberton Model' to the Pietersburg area. This analogy was quoted in subsequent reconnaissance work in the area (Muff and Saager 1979, SACS 1980, Viljoen *et al.* 1982 and Tankard *et al.* 1982). Grobler was the first writer to assign formal stratigraphic names to the units in the Pietersburg Greenstone Belt. He reported the presence of a lower ultramafic to mafic extrusive komatiitic unit (the Landsberghoek Formation) capped by an exhalative banded iron formation, which he correlated directly with the Lower Ultramafic Unit capped by the Middle Marker of the Barberton Model (Anhaeusser 1973, 1983, 1986). This unit was overlain by a series of mafic to felsic volcanics (the Eersteling Formation), analagous to the mafic-to-felsic unit of the Barberton Model. The Landsberghoek Formation (subsequently renamed the Muthiwa Formation by SACS 1980, Table 1.1), was unconformably overlain by the coarse arenaceous sediments of the Uitkyk Formation. Although the relationship of these sediments to the Eersteling Formation was unclear, they were apparently the youngest rocks in the belt and analagous to the Moodies Group of the Barberton Model. The group was interpreted as having been deposited in a synclinal trough.

Grobler inferred overturned isoclinal folding was responsible for the repetition of iron formation and detrital sedimentary units in the Mt. Mare area and that this folding

postdated an earlier deformation event that predated the Uitkyk Formation. There is little structural evidence (bedding cleavage relationships, lineations and younging direction) in Grobler's thesis to support these conclusions. Grobler also believed that two types of metamorphism had affected the belt. Firstly, he noted that the simatic rocks contained retrograde mineral assemblages suggestive of amphibolite to greenschist facies regional metamorphism, and secondly, that granite-related contact metamorphism was responsible for the andalusite and chloritoid-bearing mineral assemblages found in some of the Uitkyk sediments.

This early work in the Pietersburg area is fascinating in that it shows that many of the central arguments about granite-greenstone terrains appearing in the literature today are long-standing and that questions which were first being asked up to 100 years ago are still being asked today. Although the general lack of research in the Pietersburg area means that many of these questions have not been adequately resolved in that time, it is clear that previous workers generally believed the following: first, that the majority of the simatic rocks were extrusive; secondly, that the metamorphic features observed were due to regional metamorphism and/or contact effects related to granitoid intrusion; thirdly, there was no fault related stratigraphic repetition; fourthly, the mineralization was caused by granitic fluids migrating along structural pathways; and finally, that similarities exist between the Pietersburg and the better studied Barberton greenstone belts. These early conclusions act as hypotheses which are tested in this thesis. The wide applicability of the arguments involved in these hypotheses suggests that any possible answers derived from this study may have implications for greenstone belt studies in general.

1.5 BACKGROUND AND OBJECTIVES OF THIS STUDY

As part of an ongoing gold exploration programme in the Pietersburg Greenstone Belt, Anglo-American Corporation (AAC) of South Africa contracted Dr. M.J. de Wit of the University of the Witwatersrand to undertake regional reconnaissance mapping of the south-western portion of the belt. This work (de Wit 1985) suggested the existence of previously unrecognised structural complexities in the Mt. Mare area due to the possible presence of thrust related stratigraphic repetition. However, he was not able to demonstrate conclusively that thrusts existed in the area. De Wit also suggested there was an early phase of metamorphism which predated the deposition of the sediments and intrusion of the granites, and that this may have resulted in an early phase of gold mobility. Deposits

formed during this event may have subsequently been reworked and possibly upgraded during later deformation, igneous activity and sedimentation. Hence de Wit believed that there was evidence for polyphase gold mineralization in the Mt. Mare area and that it could have occurred over a considerable time span (up to 800 million years).

The present study was initiated to follow up on de Wit's work in the Mt. Mare area with the following overall strategy in mind: Can we deduce, at least in part, the geological evolution of the area by tracing the history of gold mineralization in the belt? Within this overall framework the following specific questions were asked: first, when was the gold first introduced and by what mechanism? Secondly, has the gold been affected by later deformation, metamorphism or igneous activity and if so, were previous gold concentrations upgraded or downgraded by these processes? Finally, has there been any subsequent introduction of "new" gold in to the system?

In order to tackle these problems it was clear that an integrated approach to the study needed to be taken. Virtually all aspects of the evolution of the belt would need to be considered (to varying degrees) to try and unravel the history of gold mineralization. At the same time it was thought that this approach may be one way to overcome the problems of interpreting greenstone belts that arise from the interaction of geological processes in time and space (chapter 1.3). The basis for such an integrated approach is the careful documentation of field relationships.

1.6 PROJECT FRAMEWORK

A total of ten months were spent in the field. The majority of the time was spent remapping selected areas studied on a reconnaissance basis by de Wit (1985) and extending his work to the northern parts of the Mt. Mare area. Fig 1.2 shows the location of the Mt. Mare area and Fig 1.3 (see map pocket) is a 1:50,000 compilation map of the geology showing the areas mapped by the writer as well as by de Wit. The areas mapped focused on major fault zones, namely the Kuschke, Hollandsdrift and Snymansdrift Shear Zones (Fig 1.3), which had been identified as important targets for gold exploration. AAC were (and are still) actively exploring for gold in several areas within the Snymansdrift Shear Zone and mapping there was primarily aimed at fitting the local prospect geology in to the overall geology of the fault zone to see if any patterns emerged.

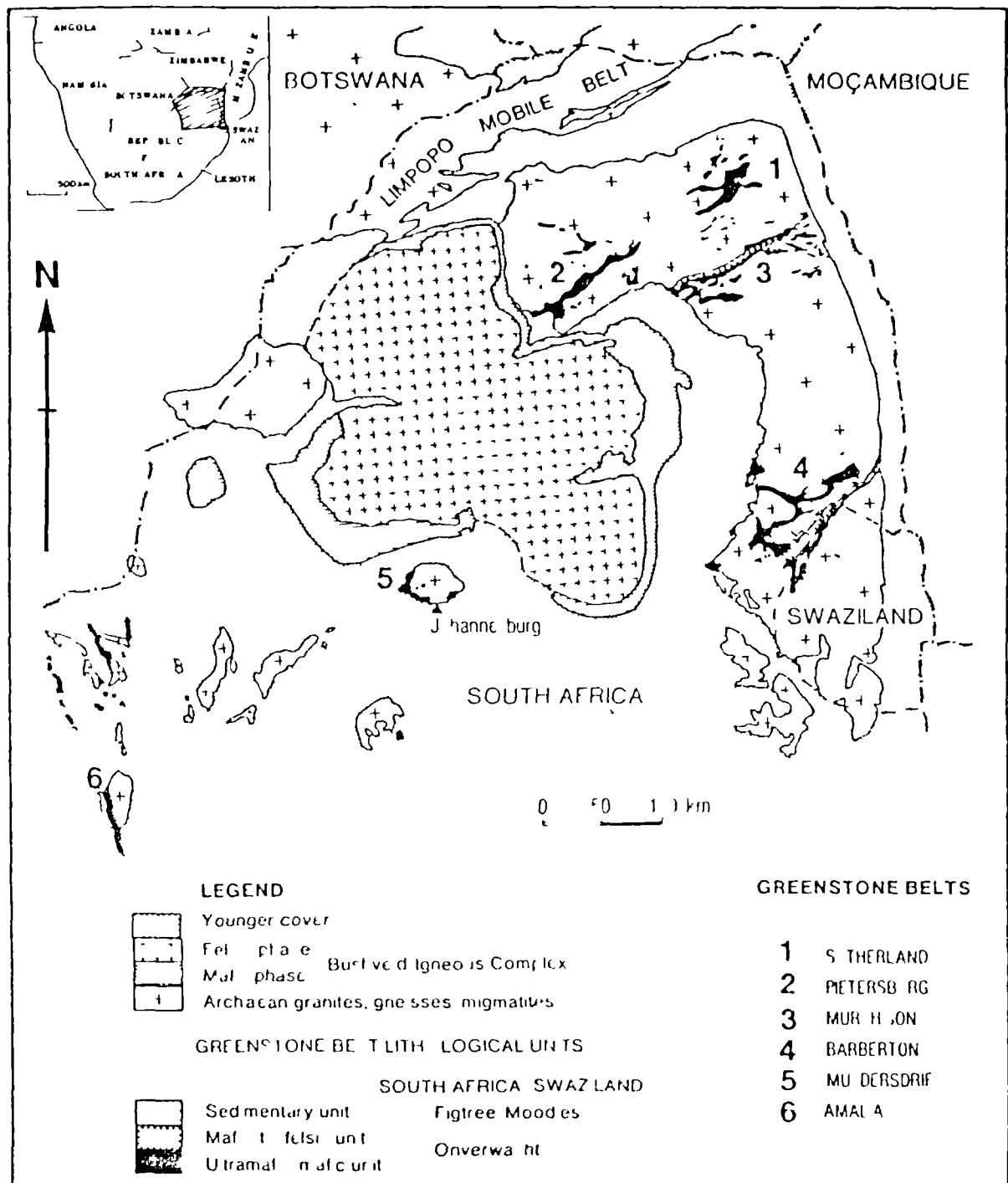
Three of the prospects, Snymansdrift #3 Hill, Snymansdrift #4 and #5 Hills and Zandrivier (Fig 1.3), were mapped at a scale of 1:1000 on grids previously surveyed in by AAC. These prospects all contain old gold workings and AAC's main exploration philosophy in the area was to reinvestigate these deposits at depth. The prospects were then linked by 1:5000 scale mapping using aerial photographs as base plans (Fig. 1.3 and Fig. 5.1). The Hollandsdrift and Kuschke Shear Zones as well as the intervening sediments were also mapped at 1:5000 scale on aerial photographs (Fig. 1.3 and Fig. 2.1). The aim of this mapping was to establish the regional setting of the Snymansdrift Shear Zone in order to elucidate the geological history of the Mt. Mare area as a whole and to identify possible target areas for further exploration. The geological detail was transferred from the aerial photographs onto a 1:5000 photographic enlargement of the 1:50,000 topographic government survey map using standard procedures (Allum 1978). The 1:1000 maps were transferred to this map using an optical pantograph. The detailed nature of the mapping allows confidence limits of between 1 and 5 metres to be placed on the accuracy of localities mentioned in the text and marked on the maps themselves. A combination of contact mapping and mapping between contacts was employed to ensure that every outcrop in the mapping area was visited.

All the major lithologies; 'greenstones', sediments and fault rocks were sampled for follow-up whole rock geochemistry. Particular care was taken in the field to avoid samples containing macroscopic veins and visible signs of weathering. Where this was not possible the samples were taken back to camp and trimmed using a diamond saw. Oriented hand specimens were taken for structural petrographic work. The surface sampling and mapping was backed up by detailed core logging (scale 1:100) and sampling of 1000 metres of diamond drill core from the Snymansdrift #3 Hill and Zandrivier prospects.

The samples collected were analysed for their whole rock geochemistry using the ICPAES technique. Ordinary and polished thin sections were prepared for use in transmitted and reflected light and scanning electron microscope petrographic studies. Mineral identification and analysis was undertaken by XRD and by electron microprobe. The methods used and precision and accuracy limits for the ICPAES technique employed are given in Appendix 1.

1.7 STRUCTURE OF THE THESIS

This chapter has outlined current research and some of the problems that exist in the study of granite-greenstone terrains and also summarised previous work in the Pietersburg Greenstone Belt. The next chapter briefly describes the geological characteristics of the two major rock sequences in the study area, the metamorphosed simatic rocks and Uitkyk Formation sediments. Chapter 3 deals in more detail with the petrography, metamorphism and geochemistry of the simatic rocks. A detailed field and microstructural study of the Snymansdrift Shear Zone and its associated gold mineralization is given in chapters 4 to 7. Chapter 8 describes other shear zones in the area and their relationship with the surrounding granites. These observations are combined with the earlier descriptions to provide an integrated model for the evolution of the Mt. Mare area. The thesis closes in chapter 9 with a series of conclusions and suggestions for further work.



**Figure 1.1 Location of the Pietersburg Greenstone Belt
(from Franey 1987).**

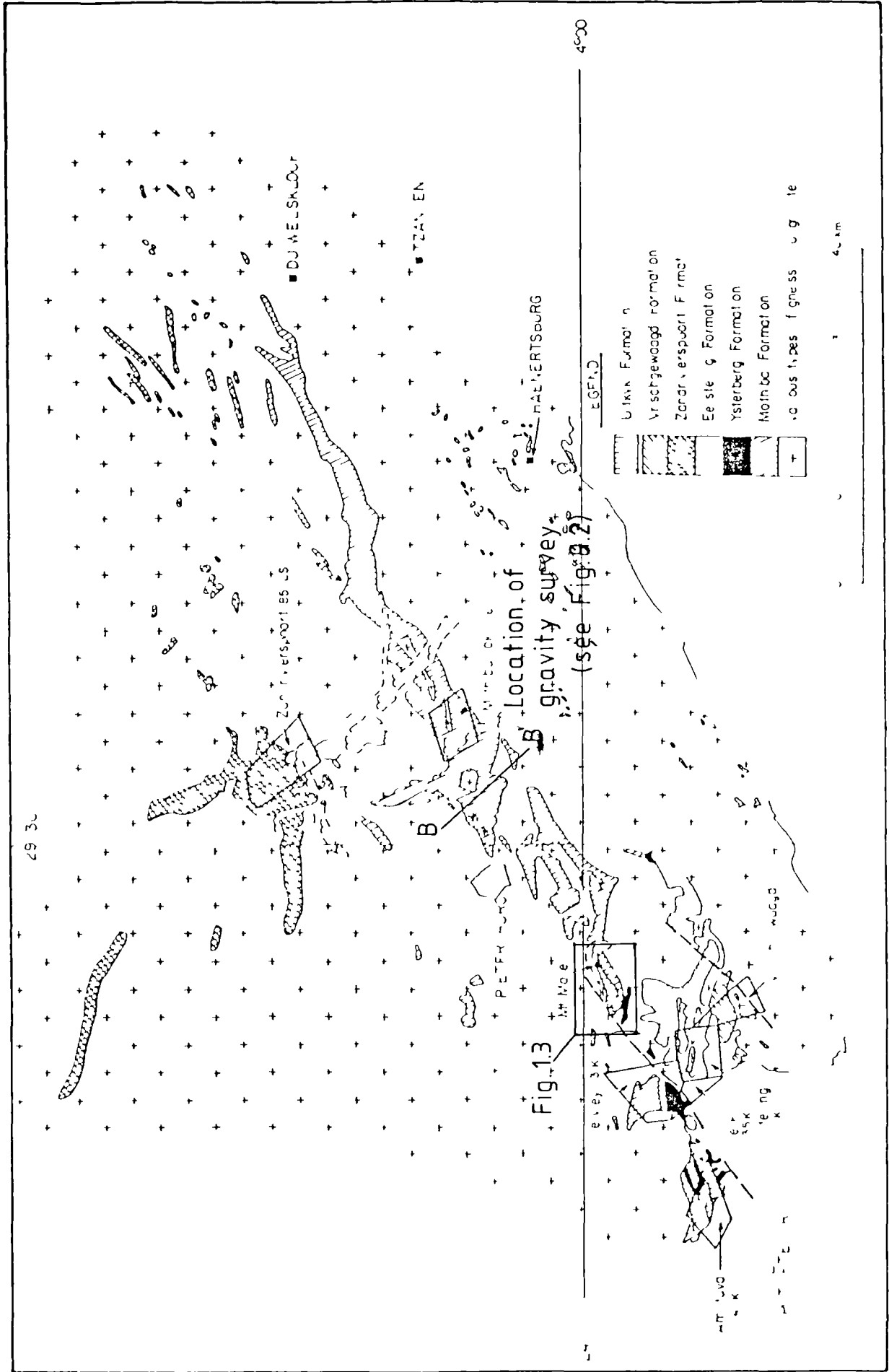


Figure 1.2 Geology and localities of the Pietersburg Greenstone Belt (from SACS 1980).

**Table 1.1 LITHOSTRATIGRAPHIC SUBDIVISION OF THE PIETERSBURG GROUP
(from SACS 1980)**

Formation	Lithology	Max. Thickness (m)
Uitkyk	Quartzites, conglomerates, quartz-mica schists, shale, banded ironstones.	1<500
Vriscgewaagd	Chloritic and quartzitic schists.	500-1,1000
Zandriverspoort	Mafic metalava with interlayered magnetite-quartzite.	500
Eersteling	Mafic metalava with subordinate ultramafics, quartz-feldspar porphyry and sialic lavas.	10,000
Ysterberg	Sialic lavas and tuff, shale, interbedded quartzitic schist, banded ironstone/ chert, quartzite.	1,400
Mothiba	Ultramafics with minor amphibolite, quartzitic schist, quartz-feldspar porphyry, banded ironstone.	7,500
	Various types of granite and gneiss intrusive into all the formations.	

CHAPTER 2

STRATIGRAPHY, BASIC STRATIGRAPHIC AND STRUCTURAL RELATIONSHIPS AND FIELD OBSERVATIONS OF LITHOLOGIES

2.1 INTRODUCTION

The combined field, microstructural and geochemical data presented in this study will show that the Mt. Mare area constitutes a poly-deformed (D1-D2) and poly-metamorphosed (M1-M2) supracrustal tectonostratigraphic complex. In this chapter a basic stratigraphic and structural framework is established for the field area based on the mapping carried out in this project and important differences to previous frameworks are pointed out. Field observations of the main lithologies are also described. The geology of the Mt. Mare area is shown summarised at 1:50 000 scale in Fig. 1.3 and the main localities mentioned in the text are shown on a transparent overlay. Also shown on this overlay are the areas mapped in detail, the main lithological units and high strain zones.

2.2 STRATIGRAPHY AND AGE RELATIONSHIPS IN THE MT MARE AREA

The study area is underlain by three major lithological sequences (Fig. 1.3) whose tectono-stratigraphic relationships are shown in Fig. 2.1. First, is a sequence comprised of intrusive and extrusive igneous rocks and banded iron formation (BIF), hereafter jointly referred to as the 'greenstones'. Secondly, is a sequence of variably deformed siltstones, sandstones, conglomerates and breccias which form part of the Uitkyk Formation as defined by SACS (1980); and thirdly a sequence of granitoid rocks. These three sequences are bounded by faults, fault-related melanges, unconformities or intrusive contacts and thus they can be considered as discrete tectono stratigraphic units. Minor basic intrusives crosscut all lithologies and geological contacts.

The main unit all dip southeast in the east of the study area but dip to the southwest in the west of the study area (Fig. 1.3). This change in strike is primarily related to a major change in strike of the Uitkyk Formation sediments above the Snymansdrift Shear Zone at locality A (Fig. 1.3). Above this sedimentary unit the stratigraphy to the southeast and

south west is different (Fig. 2.1).

Quantitative ages of the three main lithological sequences are very poorly constrained and only qualitative statements can be made. The major timing relationship established in the study area is a well defined unconformity that exists between the greenstones and overlying Uitkyk Formation sediments. Fig. 2.2 is a geological map of the central west portion of the Mt. Mare area (Fig. 1.3) which shows that the predominantly NE-SW strike of the near vertical Kuschke BIF is clearly truncated by the NW-SE strike and southwesterly dip of the sediments. At locality A (Fig. 2.2) the contact is marked by irregular, basal, breccia-conglomerate layers of BIF clasts in the sediments, which are derived directly from the underlying BIF units. There is no foliation visible parallel to the contact which in fact is cut at a high angle by a locally well defined cleavage. A similar relationship is seen at the eastern extremity of the Kuschke BIF except that the sediments dip to the east (Fig. 1.3). Thus in common with previous writers (Chapter 1.4) it is concluded the Uitkyk Formation is *younger* than the greenstones. The age of the greenstones themselves is poorly constrained and only de Wit (1985), from unpublished lead isotope data from J. Barton, gives an age for them of 3.4 Ga.

Granitoid rocks in the Mt. Mare area constitute both the structurally highest and lowest units in the stratigraphy (Fig. 2.1). However, they appear to belong to the main plutonic complexes which surround and generally intrude the Pietersburg belt (Fig. 1.2) and which have been dated at 2.6 - 2.7 Ga (SACS 1980, Brandl pers-comm 1987). Therefore, although the contact between the structurally lowest granite and overlying unit of simatic rocks is now sheared (Fig. 1.3), the granite does not constitute a 'basement' to the greenstone belt as a whole.

Combined, these relationships indicate that the stratigraphy of the Mt. Mare area is comprised of several intercalated, variably aged tectono-stratigraphic units including 'repeats' of younger and older rocks that are bounded by faults (Figs. 1.3, 2.1). This raises directly the question of stratigraphic repetition (Chapter 1.4).

2.3 EVIDENCE FOR POLYPHASE DEFORMATION AND STRATIGRAPHIC REPETITION BY TIMING

2.3.1 Evidence for an early deformation event (D₁)

As well as indicating that the Uitkyk Formation is younger than the greenstones, the non-conformity between the two units also indicates the existence of a deformation event which

predated the sediments which only affected the older greenstones. Such a discontinuity also indicates the BIF (and associated simatic rocks) must have been at least locally steeply dipping prior to the deposition of the detrital sediments. To date field evidence has been found for only one early deformation event and as it is the earliest yet recorded in the belt, it is termed the 'D1' event.

At locality B (Fig. 2.2) a distinct shallow dipping (15° SE), gneissose carbonate-fuchsite-quartz tectonite lithology is found as an isolated low strain lens in a post-D1 shear zone. This rock type is reminiscent of similar units seen by the writer in the Barberton Greenstone Belt, where they have been interpreted as early extensional shear zones (de Wit 1982, 1986). At locality B (Fig. 2.2) the sense of shear associated with the tectonite is not known and the lithology is interpreted simply as evidence for a pre-Uitkyk shear zone, possibly related to D1.

Supportive evidence for an early deformation event has been found in the Rietvley area by de Wit (1987 - Fig. 1.2). There, on the basis of structural data and younging directions, de Wit suggested the simatic rocks and associated BIF may have been overturned prior to deposition of the Uitkyk sediments. A fold structure in that area, although related to the main tectonic event that deformed the Uitkyk Formation, may be a downward facing antiformal syncline. Hence at least some parts of the simatic stratigraphy were probably overturned as well as steeply dipping prior to the deposition of the cover rocks.

2.3.2 Evidence for stratigraphic repetition by thrusting (D2)

Younging directions deduced from cross bedding structures, show that above the unconformity at locality A (Fig. 2.2) the Uitkyk Formation sediments young to the southwest. However, in an area to the north of this locality not mapped previously, a further unit of Uitkyk Formation sediments has been found by the writer which are seen to strike NE-SW, dip under the simatic rocks and young *towards* them. The two units are separated by a high strain zone. Hence the Uitkyk Formation underlies, as well as overlies, rocks that are older than itself, indicating that there must be stratigraphic repetition in the Mt. Mare area. In all cases cleavage in the sediments is steeper than bedding (Fig. 2.2) signifying that all exposures are the right way up and structurally upward facing. These relationships show that isoclinal folding (as inferred by Grobler 1972) can not be responsible for the repetition seen and hence the northern most, high strain boundary of the central zone of simatic rocks, where it is in contact with the sediments, must be a thrust-related feature. The central package of simatic rocks is therefore made up of at least one thrust sheet or nappe and is allochthonous.

Figure 1.3 shows that the northern-most sediments are also underlain by another unit

of simatic rocks, whereas the package of sediments above the unconformity at Locality A (Fig. 2.7) youngs toward, and dips under, a further unit of simatic rocks (containing the Hollandsdrift BIF). This apparent cyclical stratigraphy of alternating layers of greenstone and sediment was first noted by Hall (1909). Assuming at present that all the packages of simatic rocks seen in Fig. 1.3 were formed during the same igneous event, the structural evidence presented here suggests that all of the units of simatic and related rocks may be thrust sheets and that thrust and nappe tectonics has played an important part in the evolution of the Mt. Mare area. This is the second (and main) phase of deformation documented in the study area and it is termed 'D2'. Mapping in this project was concentrated on the geological relationships in three of the D2 high strain zones; the Snyman'sdrift, Kuschke and Hollandsdrift shear zones (Fig. 1.3). Away from these zones only basic observations were made on two of the main tectono-stratigraphic units; the greenstones and the Uitkyk Formation.

2.4 FIELD OBSERVATIONS OF THE GREENSTONES

2.4.1 Introduction

Grobler (1972) and SACS (1980) are the only publications reporting the relative positions of the different types of simatic rock within the south-western portion of the Pietersburg belt (Fig. 1.2). They both concluded that in this area there was a basal sequence of ultramafic volcanic rocks, the Mothiba Formation, conformably overlain by pyroclastic rocks and chemical exhalative sediments of the Ysterberg Formation. The overlying Eersteling Formation, a series of more basaltic volcanic rocks, was thought to be conformable with these sediments. In total, SACS (1980) believed that these formations made up a preserved stratigraphic thickness of 19 km within which there was assumed to be little or no stratigraphic repetition (Table 1.1). Although noting that the simatic and associated rocks had been deformed prior to the deposition of the Uitkyk Formation, previous studies did not record the existence of a pre-Uitkyk period of metamorphism.

The simatic and related rocks in the belt are found as discrete litho-tectonic packages, some of which are clearly allochthonous (see above) and contain a significant amount of intrusive rocks. Two of these packages have been studied in detail and they are found immediately above and below the Kuschke Shear Zone (Fig. 3.1). Most of the different *individual* litho-tectonic packages in the belt as a whole contain a variety of simatic lithologies, and all combinations of pillow lavas and breccia, BIF, net gabbro and gneiss, mafic gabbro and serpentinites seem to be present. Such observations strongly suggest that the preserved stratigraphic sequence in the Pietersburg belt is

probably the tectonically dismembered remnants of what was originally a single intrusive-extrusive igneous complex.

Although further work in the area is required, a working estimate of the minimum thickness of simatic lithologies has been obtained by measurement of the width of different litho-tectonic packages identified in this study and those of de Wit (1987) and Franey (1987). These estimates are shown in Table 2.1. In total, the preserved thicknesses of simatic rocks may represent an original stratigraphy that was probably less than 10 km thick.

2.4.2 Banded Iron Formation (BIF)

Description: Steeply dipping BIF units are found throughout the south-west of the Pietersburg Greenstone Belt, intercalated with both mafic and ultramafic rocks (Fig. 1.2, Table 1.1), and two distinct types are seen. In the Mt. Mare and Rietvley areas the BIF is exclusively oxide facies and consists of alternating fine beds and laminations of microcrystalline (20 to 50 microns), white to grey quartz (chert) and magnetite. A minor percentage (<5%) of goethite/limonite probably represents weathered ankerite (Plate 2.1a). In AAC drill core from the Eersteling area the BIF units are silicate oxide facies and magnetite-grunerite-chlorite assemblages are common. In both types of BIF the thickness of layering is variable and ranges from microbands less than 1mm in width, up to beds a maximum of 4cm wide. Close to later shear zones the BIF units have been altered to black and white "zebra" cherts or been completely silicified.

Although poor outcrop and subsequent deformation have obscured many of the original relationships between the BIF units and surrounding simatic rocks, there are many localities within the south-western part of the greenstone belt where such relationships are preserved. In the central part of the Mt. Mare area it is evident that the BIF units are intruded by the simatic rocks. This relationship is best observed at locality C (Fig. 2.2) (Plate 2.1b). This type of igneous cross-cutting relationship is common throughout the central area and therefore the rather disjointed nature of the BIF units seen in Fig. 3.1 is probably best explained if they represent a train of megaxenoliths. The intrusive relationship also indicates the BIF is the oldest lithology in the Mt. Mare area.

Some of the silicate facies BIF units from the Eersteling area have also been intruded by the simatic rocks and Plate 2.1c shows a well developed epidote-garnet quartz-rich contact zone in the BIF, related to a simatic intrusive with a well developed chilled margin. The chert layers in these sediments have recrystallised in essentially a static fashion and triple point junctions are ubiquitous (Plate 2.1d).

De Wit (1987) reports similar intrusive relationships between the simatic rocks and

BIF in the Rietvley area, but notes in addition, that the BIF units are also intercalated with extrusive volcanic breccias and epiclastic sediments. Some of the intrusives are particularly large sills.

Interpretation: BIF and associated chert units in greenstone belts have commonly been interpreted in terms of 'seafloor' hydrothermal activity and chemical exhalative processes (Viljoen and Viljoen 1969a, Anhaeusser 1973, Fripp 1976, Boyle 1976). However, other workers such as Lowe and Knauth (1977), de Wit *et al.* (1982), Lowe *et al.* (1985) and Paris *et al.* (1985) in the Barberton Mountain Land, Boultter *et al.* (1987) in the Pilbara Block, Australia and Dimroth (1977) in Canada, have all documented BIF units and cherts that were originally interpreted as chemical exhalative sediments, but were shown to be silicified and ferruginized sediments and simatic rocks. The reinterpretations were based on the presence of remnant sedimentary structures and igneous textures in the rocks. In the Pietersburg belt the evidence for the origin of the BIF is equivocal. Nonetheless, their abundance indicates that considerable hydrothermal activity, which resulted in the precipitation of significant quantities of iron and silica, must have operated early in the history of the greenstone belt. Given that the BIF is intruded by the simatic rocks at Mt. Mare but is intercalated with volcanics in the Rietvley area, it is most likely the formation of the BIF overlapped in time with the evolution of the simatic rocks as a whole and that the iron and silica were leached from the surrounding simatic pile.

2.4.3 Intrusive Rocks

Three types of intrusive rock can be recognised in the central Mt. Mare area (Fig. 3.1). Although poor outcrop obscures the contacts between these lithologies, the general outcrop pattern and lack of intervening structural fabric suggests the contacts are not tectonic and that the intrusives may represent some sort of igneous complex. No completely primary igneous mineralogies were observed in the field and it is evident that all the intrusive lithologies seen have been metamorphosed and they are thus strictly metamorphic rocks.

Volumetrically the most important intrusive rock is a purple to grey, medium to coarse grained, massive serpentinite found in the south western portion of the area (Fig. 3.1). Surface weathering has rendered it difficult to distinguish textural variation in the serpentinites but on fresh surfaces probably pseudomorphs of olivine up to 1 mm in diameter can be seen. Thus these serpentinites probably represent magnetitised serpentinites. The serpentinites are cross cut by numerous anastomosing veinlets of fibrous asbestos up to 2cm wide with alteration halos up to 5cm wide around them. The veins can be up to several metres long and are commonly sub-parallel to each other, giving the rock a layered appearance. There are numerous small veins cutting the rock and these are cross cut by the asbestos

veins. Similar serpentinites occur to the north of the Mt. Mare area where they are found both in tectonic contact with, and as xenoliths within, an underlying granite gneiss (Fig. 3.1). Here the rocks are also quite layered but there are no asbestos veins (Plate 2.1e).

In the Rietvley area (Fig. 1.2) there is an old asbestos mine in serpentinites similar to those in the Mt. Mare area. Here, asbestos veins have developed to a greater extent and constitute the main part of the rock. The asbestos is also found as slickenslide-like features along shear planes (Plate 2.1f). The development of the asbestos fibres may be analogous to the crack-seal mechanism of fibrous extension vein and shear vein formation of Ramsey (1980). The presence of serpentine as well as asbestos, magnetite and ankerite veins indicates that the serpentinites have been considerably affected by metasomatic processes.

The second type of intrusive rock is found in the eastern portion of the Mt. Mare area (Fig. 3.1). It consists of nodular to tabular to fairly massive bodies of dark green to grey amphibolite with minor associated serpentinite. The amphibolites are medium to coarse grained rocks which are locally carbonated and contain albite, epidote and amphibole as well as occasional ocelli. They appear to represent metagabbros. In AAC drill core from the Waterval prospect (Fig. 1.2) there are good examples of cumulate layering in metagabbros, with occasional possible basal graded layering of feldspars (Plate 2.2a and b). There is also evidence for complex magmatic processes within the gabbroic sequence and Plate 2.2c shows one of the coarse feldspar layers that actively invaded and brecciated an earlier phase of finer grained gabbro. These intrusive rocks were mapped as volcanics by earlier workers such as Grobler (1972) and SACS (1980), and assigned to the Eersteling Formation.

The final intrusive rock type in the Mt. Mare area is found close to the BIF xenoliths in the northwest portion of the area (Fig. 3.1). It is a blue green, porphyritic rock containing coarse crystals of randomly oriented skeletal amphibole crystals set in a fine-grained, variably carbonated, chlorite talc matrix. The rocks often have a brown colour (weathered carbonate) close to the xenoliths. In the Rietvley area similar intrusive rocks grade into coarse, oriented zones of spinifex-like fibrous amphibole crystals set in a talc-chlorite matrix. These rocks are texturally similar to the Geluk-type basaltic komatiites of the Barberton Greenstone Belt (Viljoen and Viljoen 1969b). However, the amphibole crystals are found in well defined zones up to 75cm wide which have a preferred orientation perpendicular to the edges of these zones. The zones are also texturally symmetrical and often have deformation features such as kink bands in them. At locality A (Fig. 3.1) the zones have orientations that differ by as much as 90° in one outcrop. These observations are not consistent with simple layering of crystals in volcanic flows. Rather they are better explained as cross-cutting zones of spinifex-like rocks. It is not clear from the field relationships if these textures are igneous in origin or are a result of later metamorphism which has affected the lithologies.

2.4.4 Extrusive Rocks

In the Mt. Mare area, to the north of the intrusive rocks, is a fault bounded package of metamorphosed extrusive rocks (Fig. 3.1, Fig. 2.2). It consists of a gradational sequence of pillow lavas, hyaloclastites and pillow breccias, as well as mafic and felsic lavas and occasional sills and dykes.

The commonest extrusive rocks are non-vesicular pillow lavas. They can be picked out by brown to cream coloured, concentrically zoned chilled margins and are up to 1.5m by 75cm in size, although generally smaller (Plate 2.3a). They lie in a fine grained tuffaceous matrix. Some of the larger pillows form a "figure-of-eight" shape indicating they may have been budding (splitting into two smaller, more equidimensional pillows). Draping of pillows over one another is not common and so unequivocal younging directions were not observed. Although no three dimensional outcrops of the pillows were seen, the elongate nature of some of them suggests they may represent lava tubes as well as pillows. In the Lersteling area there are good outcrops of metabasaltic pillow lavas in stream sections and possible edges of pillows were observed in core samples from the Girlie prospect. No extrusive ultramafic lavas (*komatiites sensu stricto*) were observed.

A well developed sequence of pillow breccias also occurs in the area which consists predominantly of mafic, often carbonated, clasts up to 20cm³ in size. Clasts with both entirely chilled margins and angular, fragmented margins are present (Plate 2.3b) and they represent broken pillows and small globules and fragments of chilled lava, all of which lie in a predominantly hyaloclastic matrix. Individual fragments have deformed both in a brittle fashion by *in situ* brecciation and in a ductile manner within the unit, indicative of a more explosive effusion of magma than the pillow lavas. The breccias also contain grey cherty clasts which may represent remnant metasediments. Similar but more extensive breccias, plus related lapilli tuffs and accretionary lapilli tuffs, were briefly described from the Rietvley area by de Wit (1987). De Wit also identified palagonite in the units showing that the breccias, despite their age, still retain a very low grade of metamorphism.

A distinctive feature of the extrusive rocks is the presence of more felsic material. The pillows contain abundant ocelli, 1-30cm in diameter (Plate 2.3c) with smaller ocelli spalling off from the larger felsic masses into the mafic magma (Plate 2.3d). The pillow breccias are also invaded by felsic magma showing similar textures (Plate 2.3e). A small number of aegirite dykes are seen cross cutting the extrusives as well as occasional possible slump structures (Plate 2.3f). This suggests that the felsic material may have intruded into, or extruded onto, emplaced tuff as well as consolidated rocks. It also implies that the fault bounded unit of volcanic lithologies is the right way up and dips under the intrusives. Fig 2.2

Ocelli (and related structures) have been variously explained as being the product of liquid immiscibility (Ferguson and Currie 1972, Gelinas *et al.* 1976, 1977, Coltorti *et al.* 1987), alteration (Hughes 1977, Philpotts 1977), and magma contamination (Frost 1985). Although the ultimate origin of the felsic material is unclear in the study area, the field relationships described are best interpreted as being indicative of co-magmatic felsic and mafic magmas which were immiscible. Such immiscibility could result in the observed spalling-off texture.

2.5 THE UITKYK FORMATION

2.5.1 Introduction

In the Mt. Mare area the Uitkyk Formation consists of several sequences of coarse detrital sediments that are intercalated with discrete units of simatic rocks (Fig. 1.3). There have been no published studies of the sedimentology of these rocks and little is known about their exact age, metamorphic grade or structural position in the greenstone belt stratigraphy, apart from the fact that they appear to be younger than the greenstones themselves. In this section, the sedimentology, interpreted depositional environment and provenance of the Uitkyk Formation are outlined.

2.5.2 Sedimentology

A section along traverse line A-A' (Fig. 2.2) was measured across the thickest exposed sequence of sediments, found in the central-west of the study area. The section (Fig. 2.3) is truncated by a high strain zone, the Hollandsdrift Shear Zone; hence the original *total* thickness of the Uitkyk Formation is not known. In the measured sedimentary section, visible internal strain is low and primary sedimentary structures are well preserved, so the existing stratigraphic thickness of 1 km may be close to the original, albeit a minimum, pre-deformation thickness.

The Uitkyk Formation, as seen in the Mt. Mare area, consists of a basal thin band of homogeneous green shale overlain by a sequence of coarsening upward cycles of detrital arenaceous and rudaceous sediments (Fig. 2.3). Successive cycles contain relatively greater volumes of conglomerate and thus the sequence is one of overall coarsening upwards. Three main lithofacies, designated A (coarse grained pebbly sandstones and constituting 60 % of the section), B (clast supported conglomerate, 20% of the section) and C (matrix supported conglomerate, 10% of the section), dominate the sequence, all of which interdigitate rapidly along strike towards the top of each cycle. The conglomerates become more abundant

towards the top of each cycle until they form a virtuously continuous layer, with only minor sand tone intercalations, which cap the top of each cycle. The rudites suddenly die out and are overlain by lithofacies A marking the sharp base of the next cycle. Table 2.2 summarises the sedimentology of the individual lithofacies along section line A-A'. The central-west package of the Uitkyk Formation sediments is the only one to show lithofacies other than A because elsewhere there is insufficient outcrop and or structural truncation of the sequence has occurred.

A qualitative study of the types of clast in the Uitkyk Formation has been undertaken to place constraints on the provenance of the detritus. Table 2.3 lists all of the clasts observed in this study together with a list of ore minerals found by Muff and Stager (1979) at the Mt. Robert deposit on the farm Amatava, southwest of the Mt. Mare area (Fig. 1.2). All the conglomerates are polymictic and virtually all the different clasts can be seen within individual beds.

2.5.3 Interpretation of depositional environment and provenance

The facies described in Table 2.2 represent immature sediments that are similar to those described from alluvial environments; rapidly interdigitating channel, sheet flood and debris flow deposits (Rust and Koster 1985, Lamb 1984, DeCelles 1986). For these reasons the Uitkyk Formation is interpreted as having formed as an alluvial fan or proximal braided alluvial plain. The overall lack of channelled deposits can be attributed to rapid lateral migration within a wide river system. Vertical aggradation was probably the dominant sedimentary process. The 'flashy' nature of many of the debris flow deposits, which would have required a considerable flow velocity and sedimentary load, and the cyclical nature of upward coarsening sediments, are features characteristic of an allogenic influence on sedimentation with tectonic activity inferred to increase towards the top of each cycle (Rust and Koster 1985). The increasing volume of conglomerates towards the top of the exposed formation, and hence overall coarsening upwards of the sequence, suggests that the alluvial environment was prograding and that the tectonic front that supplied the detritus may have been migrating towards its depositional basin. It is not possible to state on the basis of the sedimentological study alone if the sediments are related to an extensional, thrust related, oblique-slip or strike slip tectonic regime. Table 2.3 indicates the detritus was shed from an emergent provenance of predominantly greenstone belt rocks. Only late in the depositional history was a granitoid source tapped. Clasts of intraformational sandstone and mudstone attest to local within basin reworking of the sediments.

2.5.4 Post depositional characteristics

As depicted in the block diagram, the Uitkyk Formation is situated between

affected by metasomatism and metamorphism, and these processes were most effective in and around shear zones. Here however, only the effects of these processes away from such zones are considered.

Metasomatism

Two distinct metasomatic fluids seem to have affected the sediments. Along the unconformity to the north of locality A (Fig. 2.2) the sandstones are locally brecciated and invaded by quartz-tourmaline veins (Plate 2.5c). A more passive infiltration of similar fluids can be seen to have imparted a grey colour to the sandstones along most of this part of the unconformity.

Higher up in the sedimentary package the matrix to many of the conglomerates and debris flows is clearly sideritic, being picked out by its dark brown colour due to weathering (Plate 2.5d). Sideritic alteration fronts that cross-cut bedding in the conglomerates are also seen locally. This siderite is a post-deposition, hydrothermal alteration effect related to the passage of iron-magnesium carbonate bearing fluids through the sediments. These conglomerates also contain siderite nodules (Table 2.2, Plate 2.5e) that were formed prior to their incorporation in the sediments. Their presence suggests that hydrothermal fluid flow of carbonates may have operated over a prolonged period of time and so could have overlapped in time with tectonism and sedimentation. The presence of siderite in the conglomerates only, suggests that porosity/permeability contrasts between the conglomerates and sandstones controlled the fluid flow in these rocks.

Both the quartz-tourmaline and siderite alteration zones are cut by the cleavage seen in the sediments. This observation allows the timing of the metasomatism to be constrained between deposition of the sediments and the onset of the cleavage-forming deformation event.

Metamorphism

Away from later deformation zones there were no visible signs in the field or in thin section of authigenic metamorphic minerals apart from the minor development of micas along cleavage planes (Plate 2.5f). There is some clast alignment in zones of more intense cleavage development where the neomineralised micas show features indicative of incongruent pressure solution such as aligned fibrous quartz and mica overgrowths on quartz clasts (Rutter 1983). It would seem therefore that the overall metamorphic grade of the majority of the Uitkyk Formation in the Mt. Mare area is probably no more than lower greenschist facies.

In the Rietvley area (Fig. 1.2) de Wit (1987) reported that the sediments contained microscopic evidence of thermal metamorphism (triple point junctions in quartzites) close to the contact with granitic rocks. Combined with macroscale cross-cutting relationships of the granites to the sediment, this evidence led de Wit (1987) to suggest that at least some of the granites surrounding the greenstone belt had intruded into the Uitkyk Formation. SACS (1980) also documented intrusive relationships between the surrounding granites and sediments. The granites were reported to be between 2.3 and 2.6 Ga which places the only age constraint known on the Uitkyk Formation, that it is older than 2.6 Ga.

Table 2.3 shows that contained within the Uitkyk Formation there are clasts of simatic rocks that are clearly at a higher grade of metamorphism (greenschist to amphibolite metamorphic mineral assemblages) than the sediments in which they are found. Throughout the entire Mt. Mare area in situ exposures of the simatic rocks were all found to be at the same higher grade than the sediments. This simple field evidence suggests therefore that, as proposed by de Wit (1985), there was a period of greenschist to amphibolite facies metamorphism, apparently confined to the simatic rocks, that predates the deposition of the sediments and hence *also* the intrusion of some of the granites into the belt.

2.6 DISCUSSION

The following conclusions can be drawn from the basic stratigraphic and structural relationships outlined in this chapter.

First, many features of the simatic and related rocks of the southwest Pietersburg belt suggest there are serious problems with earlier interpretations and that the previously accepted stratigraphy requires revision. Field relationships suggest that a significant proportion of the simatic rocks may be intrusive. Specifically, it appears that in the Mt. Mare area, the Mothiba Formation of SACS (1980) *intrudes* the Ysterberg Formation which SACS interpreted as being younger than the Mothiba Formation. The BIF units are therefore the oldest rocks in the Mt. Mare area. Furthermore, the presence of previously undocumented thrusts and isoclinal folding (chapter 2.3, de Wit 1985, 1987), and possible shear zones within the Eerteling Formation (Franeý 1987) suggest that the preserved lithologies probably constitute an original stratigraphic sequence that was considerably thinner than SACS' (1980) estimate of 19 km. SACS' (1980) figure appears to be based simply on the current exposed thicknesses of lithologies.

The conclusions alone seriously undermine the previous interpretations of the stratigraphy and therefore the evolution of the Pietersburg greenstone belt.

Clearly much more fieldwork is required in the belt before a new stratigraphy of the simatic rocks can be established with confidence.

Secondly, the relationships described have allowed most of the major timing relationships in the Mt. Mare area to be deduced. It seems that the earliest geological event recorded in the study area is the evolution of the simatic and related rocks. These lithologies then underwent at least one phase of deformation and metamorphism prior to the deposition of the Uitkyk Formation. The detrital sediments in this younger formation probably represent an alluvial fan or small alluvial plain whose detritus was shed predominantly from a greenstone provenance during active tectonism before 2.6 Ga. The sediments were infiltrated by quartz-tourmaline and siderite bearing fluids soon after deposition. Age relationships, bedding/cleavage relationships and younging directions indicate that both the sediments and simatic rocks were probably involved in a phase of thrust-related deformation that resulted in stratigraphic repetition under generally low-grade conditions of metamorphism. The resulting complex then seems to have been intruded, firstly by granitic rocks and secondly by basic sills and dykes (Fig. 1.3).

This outline of the geological evolution does not however really suggest that any of the events overlapped in time and space (chapter 1.3). In order to obtain supportive evidence for such overlap as well as the overall geologic history of the study area, it is necessary to investigate in more detail both the nature of the simatic rocks and the relationship of the sediments with inferred fault zones.

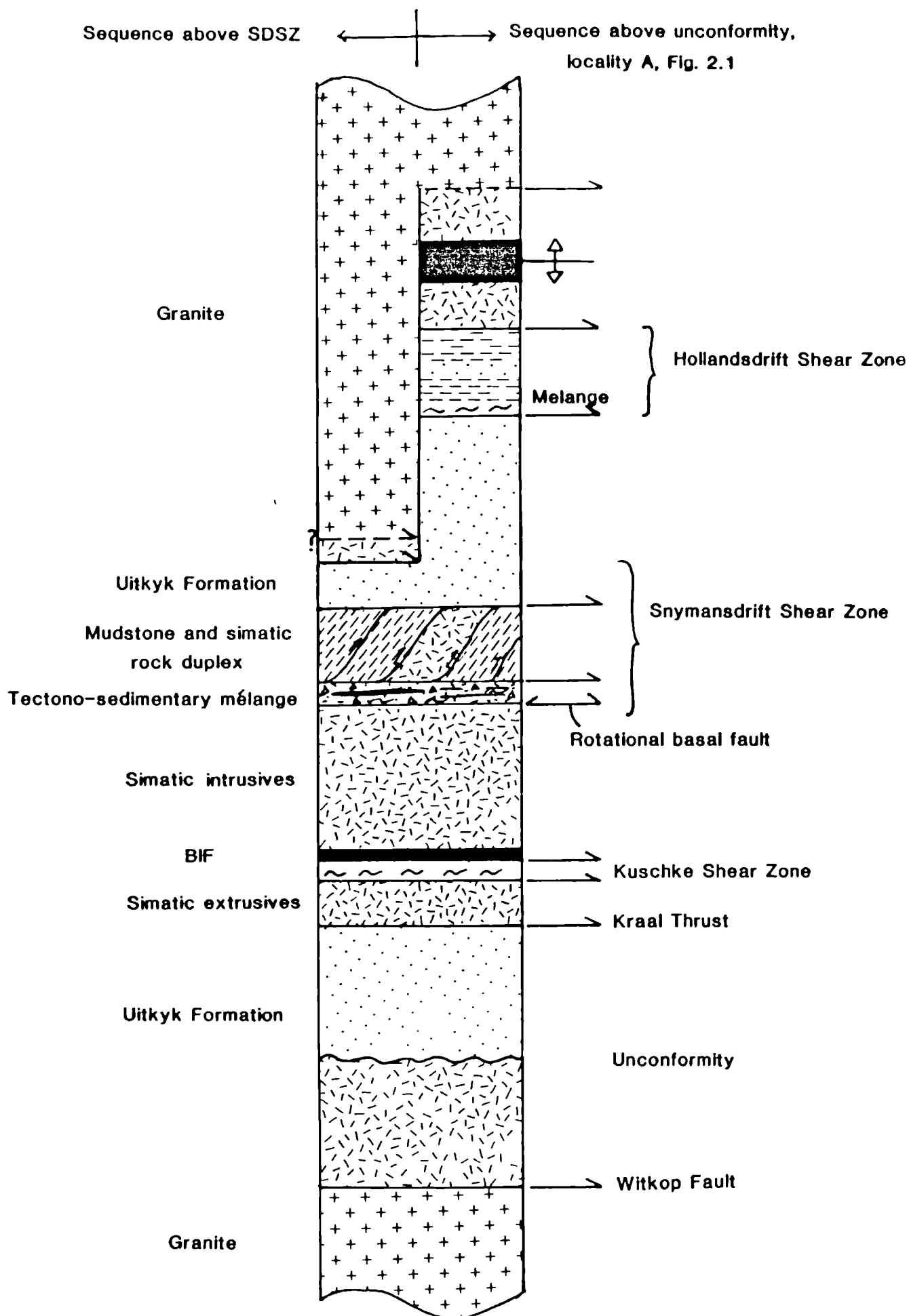
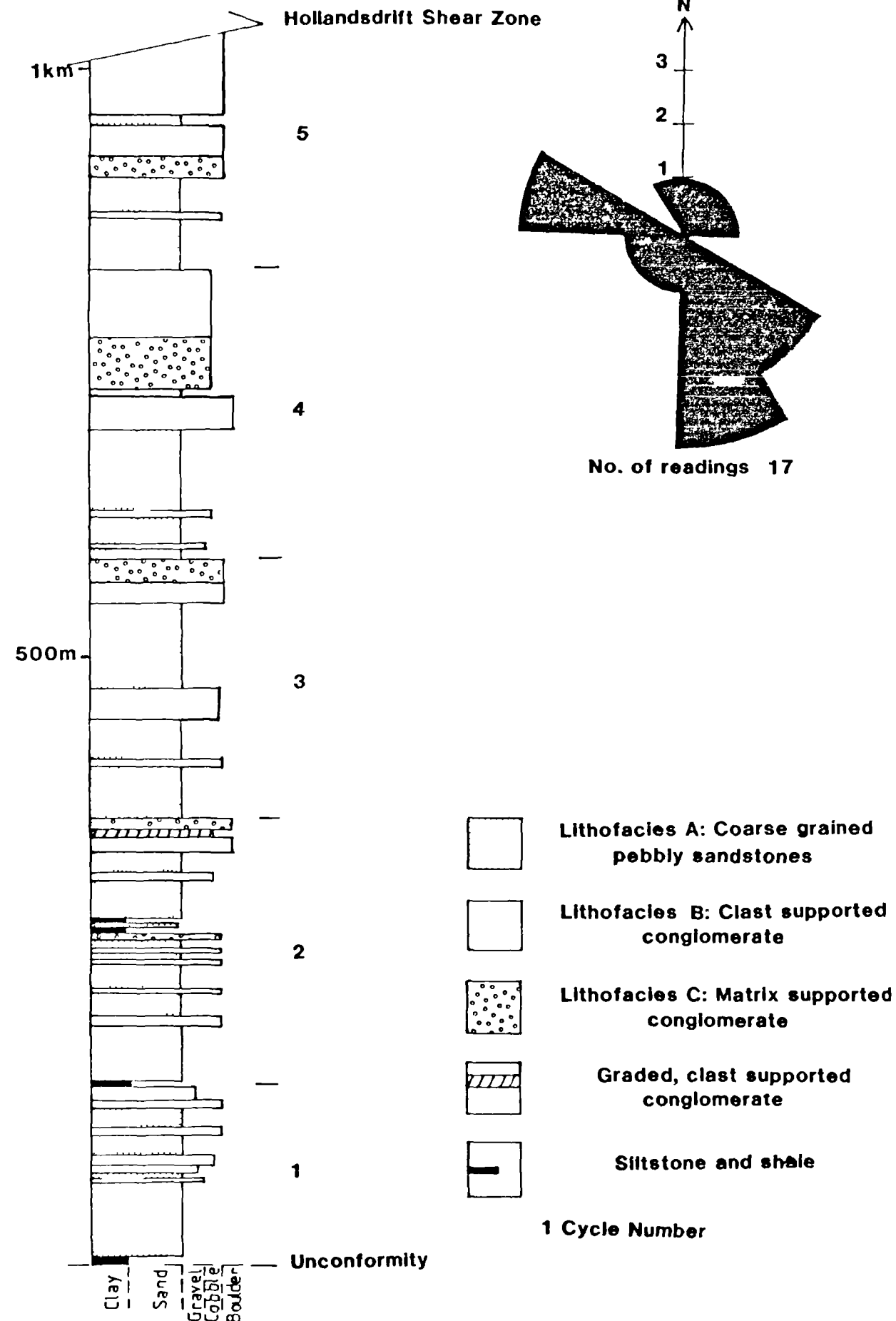


Fig. 2.1 Proposed new tectono-stratigraphy for the Mt. Mare area

Vertical scale 1: 25 000

Measured Section along
traverse line A-A', Fig. 2.1



**Figure 2.3 Sedimentological data from the Uitkyk Formation
in the Mt. Mare area.**

TABLE 2.1 SUGGESTED REVISED STRATIGRAPHY FOR THE SIMATIC AND ASSOCIATED LITHOLOGIES IN THE SOUTHWEST PIETERSBURG GREENSTONE BELT

	Lithologies	Minimum Thickness
Extrusives	Ocellular pillow lavas, breccias, tuffs, minor felsic volcanics and B F, (formerly the Ysterberg and part of the Eersteling Formations).	0.5 - 1 km
Intrusives	Serpentinites (metapendotites and metadunites), metagabbros and amphibole spinifex-textured rocks, (formerly the Mothiba Formation).	1 - 1.5 km
Mixed Intrusives & Extrusives	Pillow lavas, metabasites, layered metagabbros (formerly the Eersteling Formation).	1.5 - 5 km
Total:		3 - 7 km

TABLE 2.2 SEDIMENTOLOGY OF THE UTRKYK FORMATION

Sedimentary Unit	Description	Interpretation
Lithofacies A	Poorly sorted, granular and pebbly, coarse grained, polymict sandstones with basal conglomerate ags. Occasional pebbly cobble grade conglomerate beds.	Immature sediments deposited in high energy environment.
	Bedding ranges from 30cm to 2m, large length to width ratios, occasional erosional bases. Trough and festoon cross-bedding (plate 2.4a) ubiquitous throughout lithofacies; sets less than 50cm wide co-sets up to 2m wide (plate 2.4b). Occasional bar-like features (plate 2.4c). Grading rare but one example of both normal and reverse grading seen (plate 2.4d)	Sandstones deposited as migrating bars in a sheet braided river system during turbulent flow. Comparable to non-repetitive bar types of modern braided streams (Smith 1971, 1974) and of the Proterozoic Mbaia Group (Andrews - Speed 1986)
Lithofacies B	Poorly sorted, clast supported, pebbly and cobble grade, polymict conglomerates. Become increasingly common and rapidly interdigitate with Lithofacies A towards the top of each cycle.	Proximal, gravel-bed, braided stream deposits. 'Fashy' similar lithofacies found in the Cretaceous Kootenai Formation of Montana (De Celles 1986) and Triassic Jurassic Newark Basin, USA (Arguden and Rodolfo 1986).
	Beds - generally have erosional bases and channels into underlying sandstones, vary from thin sheets (1.5m thick and over 20m wide) to moreenticular beds, massive but crude horizontal stratification seen where coarse sandstone intercalations are present (plate 2.1e). No cross stratification or grading. Casts - moderate to good rounding, poor to moderate sphericity.	
Lithofacies C	Poorly sorted, matrix supported, pebble and cobble grade polymict conglomerates. Matrix of coarse sandstone. Also become increasingly common towards top of each cycle. Occur as sheet like bodies up to 50m wide, little evidence of erosional bases and are internally structureless (plate 2.4f). Clast size and shapes heterogeneous and boundary grade clasts are common. Base of cycle 4 (Fig. 2.4) marked by a distinctive bed of siderite nodules.	Superficially similar to debris flows (Walker 1984, Rust and Koster 1984). But lack of argillaceous matrix suggests they represent non-turbulent grain-flow deposits (Lowe 1976) in which clast support is provided by dispersive pressure between grains. In such flows, movement of clasts and their rounding or alignment is inhibited. Friable clasts thus tend not to break up as evidenced by cobble grade clasts of lithofacies C (Table 2.3)
Other Lithofacies	Two minor lithofacies make up 10% of observed section. Lithofacies B and C commonly overlain by thin 10-15cm wide layers of horizontally laminated green to brown siltstone, often marked by overlying flows to form mudflake conglomerates.	Deposited in periods of waning flow after rudite deposition in high energy environment.
	Occasional lenticular beds up to 2m thick of well sorted and graded clast supported conglomerates. Clasts are well rounded and have moderate to good sphericity (plate 2.5a)	'Resedimented' conglomerates, coarse grained turbidite like flows (Walker 1984).
Lithofacies Outside Section A	Horizontally stratified matrix to clast supported conglomerates found to the north of Mt. Mare (Fig. 2.2). Beds generally less than 50cm wide (plate 2.5b). Grading and cross bedding poorly developed.	Such deposits are common in several different sedimentary environments. Interpretation of origin usually based on association with more diagnostic facies (Walker 1984).

**TABLE 2.3 TYPES OF CLAST AND ORE MINERALS
FOUND IN THE UITKYK FORMATION**

Clasts - This Study	Ore Minerals (Muff and Saager 1979)
Fuchsite very angular fragment up to boulder grade in size Banded Iron Formation including jasperite; folded clasts are common Various cherts - grey, green, white, laminated Vein quartz Shale fragments - intraformational Sandstone intraformational Metabasalts Ocelli bearing metabasalts Spinifex-textured amphibolite Mafic schists Siderite nodules Granitoid - only found towards top of section	Pyrite Rutile Leucoxene Chromite Zircon Chalcopyrite Gold - inclusions in pyrite Ilmenite Molybdenite Magnetite Carbonaceous matter
INTERPRETATIONS BASED ON CLAST STUDY	
All clast types have been observed in situ in their original lithologies in the study area indicating a provenance of greenstone belt rocks. Only late in the depositional history was a granitoid source tapped.	
Fuchsite, mafic schist and BIF clasts all carry pre-depositional tectonic fabrics and structures. Thus there was at least one pre-depositional deformation event. Metabasalt, amphibole textured spinifex clasts indicate a pre-Uitkyk period of metamorphism. Chert and BIF indicate pre-Uitkyk metasomatism.	

Plate 2.1 Banded Iron Formation and serpentinites – field and thin section observations.

- 2.1a. Oxide facies (quartz-magnetite) BIF. Note possible cross-bedding (arrowed) but also evidence for tectonism. GR X 2 658 000, Y -33 750.
- 2.1b. Truncation of BIF layers by metaperidotites. A chilled margin some 15-20cm wide is present. GR X 2 658 080, Y -33 420.
- 2.1c. Thin section showing garnet-epidote-quartz assemblage in metasediments next to a chilled margin of an amphibole-bearing intrusive. (Sample DDH GR3, 124.2m, Eersteling prospect). PPL. Field of view – 2.2cm.
- 2.1d. Thin section of quartz layer in silicate facies BIF showing well-developed triple-point junctions suggestive of static (thermal) recrystallisation. Silicate layer is composed of grunerite-chlorite and magnetite. Sample DDH GR3, 124.3m. Field of view – 8.5mm. Cross polars (XPL).
- 2.1e. Field appearance of serpentinites in the north Mt. Mare area close to a tectonic contact with granitic rocks (Fig. 1.3).
- 2.1f. Asbestos slickensides along an early (D1?) shear zone in serpentinites from the Rietvley area.

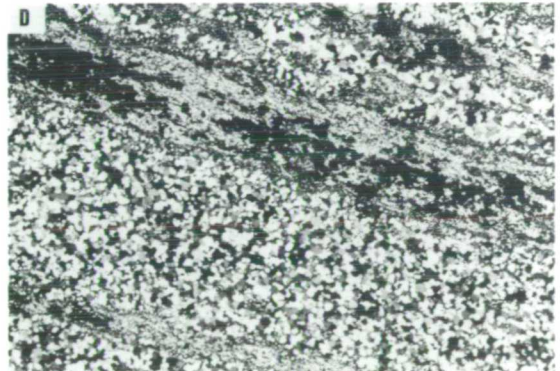
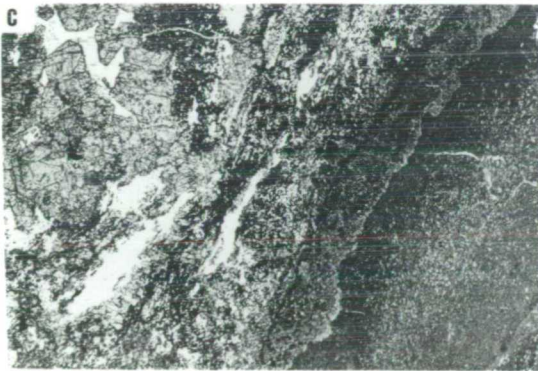
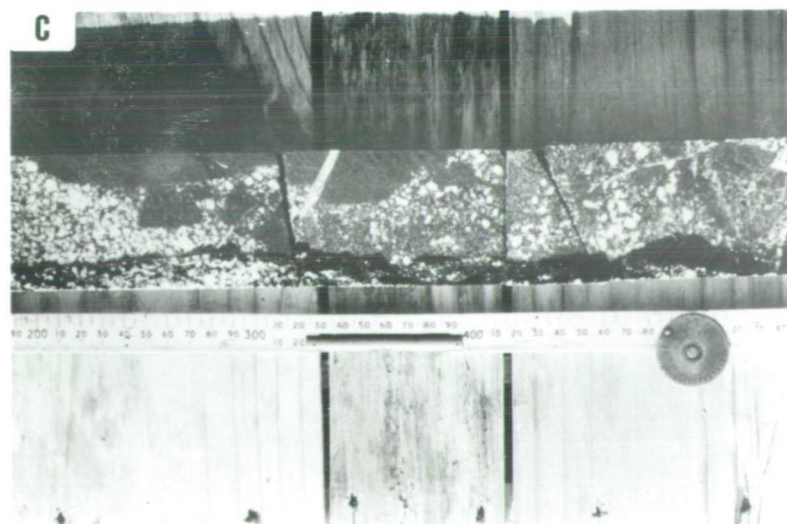
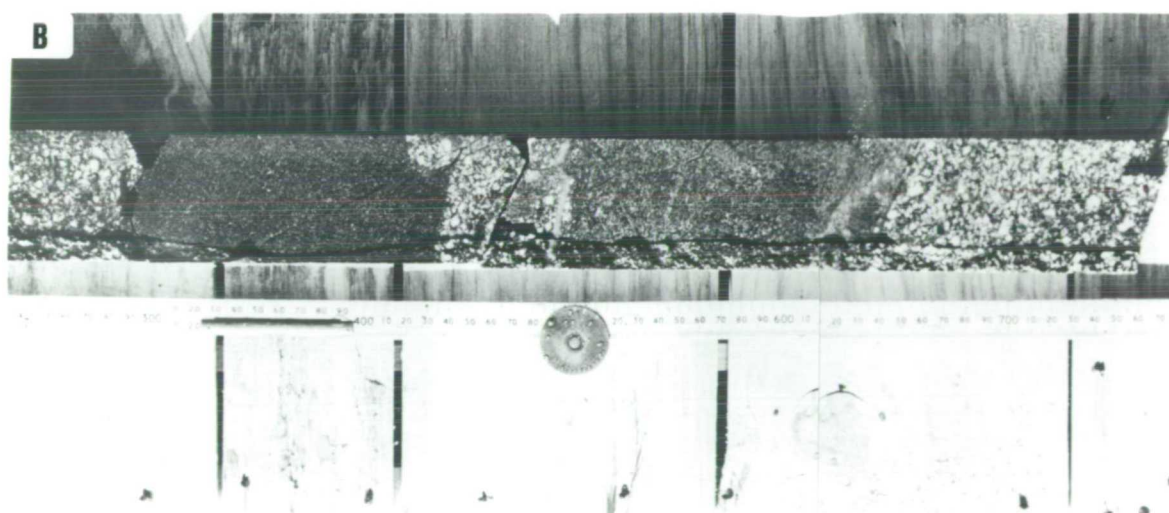


Plate 2.2 Metagabbros from the Waterval prospect - field observations.

- 2.2a. Cumulate layering with original igneous mineralogy of pyroxene and calcic plagioclase pseudomorphed by hornblende and albite.**
- 2.2b. Layered metagabbro suggesting that multiple magma injections (into a chamber?) occurred.**
- 2.2c. Evidence for forceful injection of new magma into the gabbroic sequence. Angular fragments of fine grained gabbro in a matrix of younger, coarse grained, feldspar-rich gabbro.**



**Plate 2.3 Extrusive rocks from north of the Kuschke Shear Zone
- field observations.**

- 2.3a. Pillow lavas lying in a tuffaceous matrix. Note the elongate and figure-of-eight shapes suggestive of lava tubes and budding. The rims of the pillows are picked out by concentric weathering of ankeritic alteration.
GR X 2 657 860, Y - 33 100.**
- 2.3b. Pillow breccia showing rounded and angular fragments and *in situ* brecciation leading to development of hyaloclasite.
GR X 2 657 -850, Y -33 250.**
- 2.3c. Individual pillow lava with felsic ocelli in its centre (arrowed). GR 2 657 840, Y -33 100.**
- 2.3d. Immiscibility texture with felsic ocelli spalling-off a larger felsic mass at the contact with a mafic magma. Slight D2 flattening is responsible for the elongate nature of these ocelli. GR 2 657 900, Y -33 100.**
- 2.3e. Ocelli spalling-off (arrowed) in the pillow breccias. Such textures are difficult to explain by alteration and are probably primary igneous textures.
GR X 2 657 850, Y -33 275.**
- 2.3f. Possible slump structure. Felsic sill or lava with chilled margin (arrowed) lying on mafic tuff. This suggests the sequence youngs northwards (upwards in the photograph). GR X 2657 800, Y -33 390.**



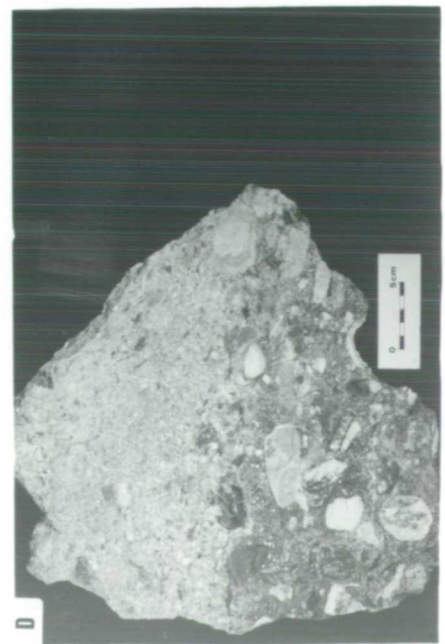
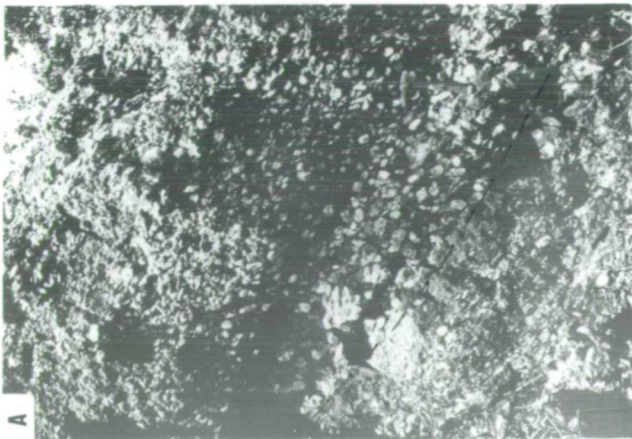
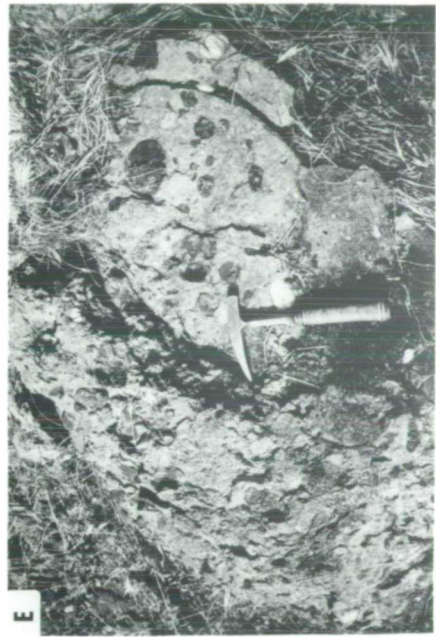
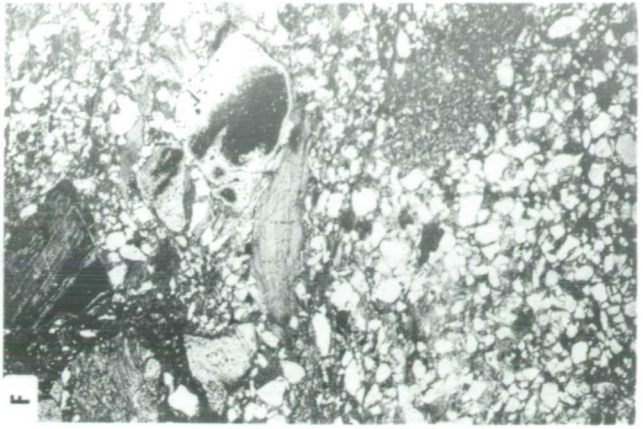
Plate 2.4. Sedimentary structures in Lithofacies A, B and C.

- 2.4a. Large scale trough cross-bedding marked by conglomerate and heavy mineral lags (Lithofacies A).
GR X 2 658 800, Y -32 050.**
- 2.4b. Large scale co-sets of festoon trough cross-bedding (Lithofacies A). GR X 2 658 700, Y -32 000.**
- 2.4c. Sand bar-like feature isolated by cross-bedding (Lithofacies A). GR X 2 658 800, Y -31 500.**
- 2.4d. Example of reverse and normal grading in successive beds of Lithofacies A. GR X 2 659 300, Y -33 100.**
- 2.4e. Clast-supported conglomerate with sandstone intercalations. Note the large fuchsite clast (arrowed). Clasts of simatic rocks metamorphosed to greenschist/amphibolite facies are also present (Lithofacies B). GR X 2 658 750, Y -31 650.**
- 2.4f. Matrix-supported conglomerate. Note the ankeritic matrix, now weathered to a dark brown colour, and the lack of sorting. GR X 2 658 850, Y-31 700.**



**Plate 2.5. Sedimentary structures in other lithofacies and...
post-depositional features.**

- 2.5a. Well sorted and graded, clast-supported "resedimented" conglomerate. Photo shows top of one bed and base of overlying bed. GR X 2 658 600, Y -32 100.**
- 2.5b. Horizontally stratified conglomerate lying structurally below older simatic rocks. GR X 2 657 700, Y -32 400.**
- 2.5c. Tourmaline hosted sandstone breccia. Note the rounding of some of the breccia fragments. Such features are common along the part of the unconformity. GR X 2 657 900, Y - 32 100.**
- 2.5d. Sideritic matrix clearly confined to the lower conglomeratic part of a single sedimentary bed. GR X 2 659 200, Y -33 000.**
- 2.5e. Large siderite nodules in lithofacies A. Locally these nodules can make up over 75% of single beds. GR 2 658 850, Y -32 100.**
- 2.5f. Thin section of Lithofacies A showing large chert and fuchsite clasts. Note the poorly developed cleavage marked by neomineralised micas suggesting that the sediments are at a low grade of metamorphism. Sample CG1, GR X 2 659 200, Y -33 150. Long side of photograph = 1.7cm, plane polarised light (PPL).**



CHAPTER 3

THE PETROGRAPHY AND GEOCHEMISTRY OF THE SIMATIC

ROCKS AND THE NATURE OF THE M1 EVENT

3.1 INTRODUCTION

Field observations of the greenstones described in the previous chapter indicated that the igneous rocks are all metamorphosed. It was established in chapter 2.5.4 that the metamorphic event responsible occurred early in the history of the area *prior* to the deposition of the Uitkyk Formation and therefore prior to the intrusion of the granitoid rocks into the Pietersburg belt. Because it is the earliest metamorphic event so far recognised in the greenstone belt it is called the 'M1' event.

Although the majority of the simatic lithologies in the worlds greenstone belts contain mineral assemblages similar to those observed as part of the M1 event in the study area, there have been few recent published detailed studies of their metamorphic petrology. Thus it is concluded that little is known about the type of metamorphism responsible for those assemblages and therefore any associated elemental mobility. It is of fundamental importance to try and understand this metamorphism because interpretations of whole rock geochemical data from Archaean simatic rocks commonly assume that major elements such as SiO_2 , MgO , Fe_2O_3 , and Al_2O_3 , were immobile during such post-igneous processes. The first step in understanding the metamorphism and to see if such an assumption is valid, is a petrographic study.

3.2 PETROGRAPHY

3.2.1 Extrusive Rocks

Both the ocellular-bearing pillow lavas and the pillow breccias in the Mt. Mare area contain well developed alteration assemblages which have completely overprinted any primary igneous minerals. They are made up of a fine grained (<250 micron) intergrowth of chlorite and quartz which has been pervasively carbonated and euhedral fine grained

rhombs of ankerite are ubiquitous. The ocelli consist of a fine grained intergrowth of quartz and albite (Plate 3.1a). In the Rietvley area similar extrusives contain palagonite replacing original glass shards and are extensively chloritised (de Wit 1987). Overall, the extrusive rocks in the Mt. Mare and Rietvley areas are at relatively low grades of metamorphism and the mineral assemblages present represent low to mid-greenschist facies conditions. The absence of actinolite in the e lithologies suggests that temperatures did not exceed 350°C (Evarts and Schiffman 1983).

In the Eersteling area, reconnaissance logging of AAC drill core revealed that the monotonous sequence of massive fine grained metabasalts seen are extensively fractured and cut by quartz veins containing varying amounts of carbonate, epidote and chlorite. In thin section, no primary igneous mineralogies were observed but igneous textures such as sub-ophitic intergrowth of actinolitic amphibole and albite are preserved. There is a significant degree of replacement of the amphibole to chlorite. Quartz, with a notably ragged morphology, and epidote are common accessory minerals. Overall, the metabasalts at Eersteling appear to be at a slightly higher metamorphic grade than at Mt. Mare and Rietvley. Mid- to upper-greenschist and epidote-amphibolite facies mineral assemblages are ubiquitous, although primary igneous textures are generally well preserved throughout.

3.2.2 Serpentinities

Although possible pseudomorphs of olivine were visible in some hand specimens, no original mineralogy or even textures were observed in thin section. The rocks constitute an unoriented felted mass of fine grained serpentine, ubiquitous chromite and minor magnetite and probably represent metaperidotites and possibly metadunites (de Wit 1985). Kenyon (1983) reported that some of the peridotites were extensively carbonated to produce magnesite-talc-chlorite assemblages. The presence of asbestos veins and quartz-magnetite-siderite veins in these rocks (chapter 2.4) is consistent with the release of considerable iron, silica and possibly calcium during the serpentinisation process as suggested from theoretical and experimental studies (Janecky and Seyfried 1986). The temperature at which this hydrothermal alteration occurred is difficult to determine but it was below 500°C (Krauskopf 1979, Kimbal and Gerlach 1986, Janecky and Seyfried 1986).

3.2.3 Metagabbros

The metagabbros all contain the following mineral assemblage (in varying proportions), amphibole-feldspar-epidote-chlorite-sphene-quartz + \ - carbonate. However, in each of the samples studied, the amphibole is always being *replaced* by chlorite and sometimes epidote. Therefore, although the above mineral assemblage is common, it does not reflect an equilibrium assemblage, rather it indicates widespread disequilibrium. Like

the extrusive rocks, igneous textures such as sub-ophitic intergrowths are well preserved.

1. Feldspar, epidote, chlorite, sphene

In most of the metagabbros the feldspars have been completely albitised or sausseritised and only in the layered metagabbros are occasional kernels of more calcic plagioclase preserved. Representative microprobe analyses are given in Appendix 2. The feldspars in the layered metagabbros from the Waterval prospect (Fig. 3.1) also show evidence of remnant igneous zoning, marked by secondary mineral inclusions suggesting that the textures seen are pseudomorphs of original igneous textures (Plate 3.1b). The albitisation of feldspars attests to mobility of calcium and sodium in the gabbros. Further evidence for calcium metasomatism comes from the presence of epidote in many of the samples, occurring as small euhedral laths and larger knots of more equidimensional euhedra. Later sausseritization of the feldspars is common.

Chlorite (up to 5-10%) is found in all the gabbros as fine grained felted masses in the groundmass and as an alteration phase of the amphiboles. These features suggest that the majority of the chlorite represents a retrogressive metamorphic overprint on an earlier higher temperature mineral assemblage. Occasionally the gabbros have been entirely retrogressed to chlorite-carbonate-albite-quartz assemblages. Also present in some of the gabbros are subhedral to euhedral crystals of sphene up to 0.5 x 0.4 mm in size. However their morphology suggests they may be a pseudomorphic replacement of ilmenite (Plate 3.1c, Stern and Elthon 1979). The crystals are often rimmed by magnetite and pyrite.

2. Amphibole

The metagabbros all contain amphiboles which are texturally and compositionally variable. Samples GC, G3, G8 and G8a (Fig. 3.1) illustrate the range seen. Representative microprobe analyses that reflect both within-crystal and within-sample variations are given in Appendix 2 and shown diagrammatically in Fig. 3.2. This compositional data shows that there is considerable variation in Si and Al(IV) within individual samples and commonly within individual minerals. In contrast the Mg/Mg+Fe ratios are relatively constant suggesting that they may reflect the bulk rock composition. Whole rock geochemical data for these samples are given in Table 3.1.

Description: Sample GC is one of the least hydrated gabbros studied (Table 3.1) and comes from the layered gabbros in the Waterval area. The amphiboles, especially from the coarse grained layers, are subhedral to euhedral, up to 0.5 cm in size and are intergrown with a significant percentage of free quartz (Plate 3.1b), a feature commonly seen in the alteration of pyroxene to amphibole (Brodie pers. comm. 1987). Although no relict pyroxenes were

observed it is thought that most of the amphiboles probably represent completely altered pyroxenes rather than late igneous amphiboles. The amphiboles are predominantly magnesio hornblendes (Fig. 3.2).

Samples G3, G8a and G8 show a progressive, three-fold, related textural and chemical sequence as follows;

A). The amphiboles range in composition from relatively iron-rich ferro-hornblendes with small rims of ferro-actinolite around them (Sample G3, Fig. 3.2), through to actinolitic hornblende and actinolite (Sample G8a, Fig. 3.2) and finally to the most magnesium rich amphiboles in sample G8 (actinolite, actinolitic hornblende, magnesio-hornblende, Fig. 3.2). This sequence reflects an increase in the Mg/Mg+Fe ratios of the amphiboles.

B). The chemical change is directly associated with a boudinage-fibrous overgrowth phenomenon which results in textural modification of the amphiboles. In sample G3 the uralitic rims of ferro-actinolite consist of fibrous overgrowths up to 50 microns long associated with micro-boudinaging of the ferro-hornblendes (Plate 3.2a). In sample G8a the fibrous overgrowths are better developed and are invariably 100-200 microns long. They attach fragments of amphibole that have been boudinaged and also grow into feldspar crystals (Plate 3.2b and c). In sample G8 the degree of amphibole deformation is greater still and this again is reflected in the length of the fibres which are commonly over 500 microns long (Plate 3.2d and e). Some of the overgrowths seen are quite complex and suggest that there may have been a local rotational component to the deformation that caused the fracturing of the amphiboles.

C). Changes A and B are both accompanied by an increase in the L.O.I. values in the whole rock geochemistry data (Table 3.1) from 2.01% in sample G3 to 4.12% in G8a and 8.60% in G8. These samples contain virtually no carbonate and thus this change reflects an increase in the degree of hydration of the samples. Significant changes in the major element geochemistry between these samples are also seen (Table 3.1). MgO increases from 3.85% to 25.66% between samples G3 and G8, plus there is an accompanying decrease in the Fe_2O_3 content from 16.48% to 10.25% and decrease in SiO_2 from 50.59% to 36.95%. Furthermore, the reaction hornblende to actinolite is an alumina-producing reaction and this is reflected in the marked decrease in alumina content of amphiboles in all samples from sample G3 to G8 (Table 3.1).

Interpretation: The boudinage-fibrous overgrowth features are post igneous metamorphic textures similar to those observed by Misch (1969) in naturally deformed amphiboles and Rutter *et al.* (1985) in their experimental work on the syntectonic hydration of basalt. The

fibres have preferentially developed on dilatant interfaces between boudins as well as epitaxial overgrowths across amphibole-feldspar grain boundaries. Because many of them actually join individual boudins they are best interpreted as having grown concomitantly with the boudinage and hence are syn-deformational features, a mechanism outlined in Fig. 3.3. Because the development of the fibrous textures is directly associated with changes in amphibole composition (Fig. 3.2), it is likely that incongruent pressure solution, where water-assisted diffusive mass transfer aids grain boundary diffusion (Rutter *et al.* 1985), was an important process in their development. Cataclasis (and the associated dilatancy) of the amphiboles probably aided the related fluid flow in these rocks.

The chemical and textural features described here represent what is actually a continuous spectrum of features that come from closely related rocks (Fig. 3.1) which look virtually identical in the field. Hence it is inferred that the samples with the best developed fibrous, and most magnesian, amphiboles (G8) represent deformed, metamorphosed and hydrated versions of the least modified rocks (G3), and they may have originally had similar compositions. Because no truly unaltered samples of gabbro were found, it is difficult to quantify (by mass balance procedures) what gross chemical changes may have accompanied this modification. However, because the amphiboles are the main magnesium bearing mineral in the metagabbros, their chemical change probably did involve the quantitative addition of magnesium to the rocks and concomitant removal of silica, iron and possibly alumina. The observed chemical variations suggest that significant major element mobility accompanied the hydration of the gabbros.

It is evident that the overall change seen in the amphiboles from hornblende to actinolite rich assemblages is a retrograde metamorphic feature and probably occurred at temperatures between 600°C and 350°C. Raase (1974) suggested that the TiO_2 content of metamorphic amphiboles was a function of temperature, the two increasing together. A decrease in the TiO_2 content of the amphiboles in this study is seen (Appendix 2) which is consistent with Raase's (1974) conclusions and the interpretation that the sequence seen represents decreasing temperature and is therefore a retrograde effect. Similar conclusions for secondary amphiboles were reached by Evarts and Schiffman (1983). The retrograde event clearly postdates the alteration from igneous to metamorphic assemblages (reflected in sample GC) but is earlier than a lower temperature event responsible for the development of chlorite-epidote bearing assemblages in the rocks. If the albite-hornblende-quartz assemblage seen in sample GC does represent a metamorphosed calcic plagioclase-pyroxene +/- olivine igneous assemblage then the metamorphic event responsible was also retrograde in nature and probably occurred at temperatures between 750° and 600° C (Kimbal and Gerlach 1986, Stakes and Vanho 1986).

Although estimates of the temperatures involved in the metamorphism can be

inferred from the mineral assemblages seen, they do not provide strict pressure constraints (Gerlach and Kimbal 1986). However, the Al (vi) occupancy content of all the secondary amphiboles (Appendix 2) may indicate low lithostatic pressures, 3Kb or less, or fluid pressures significantly less than lithostatic (Evarts and Schiffman 1983). All analyses plot in the low pressure field of Raase (1974) (Fig. 3.4).

In summary, the textural and chemical variations seen in many of the metagabbros reflect a wide range of temperature conditions but all occurring under conditions of low pressure (<3Kb). Early, high temperature (750°C-600°C) pseudomorphic replacement of igneous mineral assemblages was subsequently overprinted by lower temperature events (600°C-350°C and possibly lower), during which significant element mobility and textural modification occurred. This retrograde sequence of events was accompanied by the hydration of the rocks.

3.2.4 Spinifex-textured rocks

1. Introduction

The term 'spinifex-texture' has long been used in igneous petrology to describe skeletal crystal forms in mafic and ultramafic igneous rocks, textures believed to be common in Archaean simatic rocks (Nesbitt 1973, review by Donaldson 1982). Although 'spinifex' was originally used for describing skeletal olivine or pyroxene crystals of various shapes and sizes, the term has commonly been misused in the Archaean-related literature. Many writers have assumed that such delicate textures were originally igneous in origin and have simply been pseudomorphed; thus olivine had been isochorically replaced by serpentine-talc and pyroxene by amphibole eg Viljoen and Viljoen (1969b), Nesbitt (1971), Viljoen *et al.* (1983). This assumption of simple replacement is now well entrenched in the literature to such an extent that the terms 'olivine and pyroxene spinifex' are often applied directly with little or no mention of the alteration assemblages which are clearly present. Such misuse of the term often gives the impression that Archaean simatic rocks are not metamorphosed.

Having recognised the above problem, the term spinifex is still a useful descriptive term for skeletal-textured rocks. In this study the term is retained but its use is extended to minerals other than olivine or pyroxene, being used in conjunction with the *observed* mineralogy. No *a priori* assumption concerning the possible original mineralogy is made.

The main reason a distinction between primary and possible secondary mineralogies should be made is that processes other than igneous crystallization have been proposed for the origin of *some* spinifex textures. Evans and Tromsdorff (1973) attribute talc/serpentine spinifex textures in ophiolites to hydrothermal contact metamorphism and de Wit *et al.*

(1987a) show good examples of both serpentine and amphibole spinifex overgrowing what they interpret to be original igneous textures suggesting both types may be the product of metamorphism. De Wit *et al.* (1987a) also note that spinifex-textures ascribed to igneous processes have been described in both intrusive and extrusive mafic/ultramafic rocks indicating that the presence of such textures alone can not be used to prove an extrusive origin for the host lithology. A similar conclusion was reached by Donaldson (1982) on the basis of experimental work on the growth of such textures.

The amphibole spinifex-textured rocks described from the Rietvley area (chapter 2.4) have been studied in order to establish their origin; the three likeliest possibilities are that they represent either pseudomorphed pyroxenes, igneous amphiboles or metamorphic amphiboles. The spinifex-textured rocks occur as oriented fibrous amphiboles confined to texturally symmetrical vein-like structures. On both sides the veins are surrounded by randomly oriented blocky and more massive amphiboles. Both vein and host rock are considered here. None of samples studied contained any evidence of pyroxene growth, either in crystal morphology or as remnant kernels in the amphiboles, nor was there any large degree of free quartz as was seen in the layered metagabbros. Representative microprobe analyses of the amphiboles are given in Appendix 2 and plotted in Fig. 3.2. This data shows that there is far less variation in the mineral chemistry compared to the amphiboles in the metagabbros, suggesting the crystals, which are all actinolite, are not chemically zoned to a significant degree and that both vein and host rock amphiboles have similar compositions.

2. Textural description and interpretation - spinifex-textured amphiboles

The following four textural relationships are considered to be the most important:

A. Description: The fibrous amphibole crystals are narrow (2mm or less) but very elongate crystals with aspect ratios of up to 1:100 being common, which show a strong degree of preferred orientation (Plate 3.3). The crystals lie in a fine grained matrix of chlorite-talc-serpentine and very small (10 microns wide) but elongate (200 microns or more), variably oriented amphibole fibres. In cross-section the large crystals are arranged in bundles of between 6 and 10 which show good optical continuity and amphibole morphology. In elongate sections (Plate 3.3 and 3.4a), individual fibrous crystals within bundles commonly diverge from that bundle and then merge to become an integral part of another bundle. Some of the divergent crystals appear to have been broken off their parent bundle (Plate 3.4b). This slight 'anastomosing' habit is responsible for the variations seen in the overall preferred orientation of the crystals and also results in the many cross-cutting and truncation relationships seen between them (Plates 3.3 and 3.4).

In detail, the individual fibrous crystals are themselves made up of aggregates of extremely narrow and elongate asbestiform fibres (10 microns or less wide, aspect ratios $\gg 1:100$). Although most of these fine fibres are arranged sub-parallel to the long axis of the main individual crystal, some of the crystals are made up of fibres showing complex rotational geometries (Plate 3.4c).

Interpretation: The optical continuity between individual fibrous crystals within bundles and connection of bundles by one or more such crystals, suggests the bundles represent larger megacrysts which themselves are commonly linked in three dimensions. Thus all of the crystals are inferred to have grown approximately synchronously. Their strong preferred orientation suggests the crystals formed under conditions of stress. The extreme fibrous nature of the amphibole crystals in general suggests they may not be true 'double chain' amphiboles but belong to another member of the biopyribole family, pyribole asbestos (Veblen *et al.* 1977). Although HVTEM studies are required to confirm this interpretation, such crystals are made up of a combination of mainly double and triple chain silicates, rather than the single and double chains alone that are seen in pyroxenes and true amphiboles respectively. This structural configuration allows easy fibrous growth.

B. *Description:* The most notable feature about the fibrous crystals are the deformation textures associated with them, textures which *are not seen* in the surrounding matrix. Virtually all the crystals show undulose extinction and are bent, kinked and fractured. Boudinaging sub-perpendicular to the long axis of the crystals is common, with the gaps between crystal pieces being filled mainly by matrix minerals. Some boudins are commonly attached by sheaves of fine amphibole fibres in an identical fashion to that described in the metagabbros (chapter 3.2.3, Plate 3.2 – Plate 3.4b). The boudinage of the oriented crystals indicates significant extension sub-parallel to the crystals' length. An estimate based on restoration across a single thin section suggests that a minimum of 10% extension can be attributed to the boudinage. SEM observations reveal that boudin ends, fractures and also the long edges of the main crystals are areas of patchy chlorite alteration and intense micro- and crypto-scale cataclasis (Plates 3.5a and b).

Interpretation: The deformation textures are confined to the amphibole crystals and are not seen in the matrix. Such features indicate the deformation affected the *amphiboles* and therefore occurred *after* amphibole formation. If the amphiboles were originally igneous pyroxenes then the deformation must have postdated the pseudomorphic metamorphic event. Specifically, the undulose extinction, indicative of straining of the *amphibole* crystal lattice, would not be expected to be inherited from any parent pyroxene crystal.

The boudinage-fibrous overgrowth phenomenon is identical to that observed in the metagabbros (chapter 3.2.3) and is interpreted similarly as indicating co-existent boudinage

and fibrous growth and therefore extension sub-parallel to the length of the fibrous crystals (Misch 1969). Most of the brittle deformation structures are associated with chlorite alteration suggesting that the majority (but not all) of the cataclasis occurred at slightly lower temperatures and/or more hydrous conditions than that responsible for the development of the main amphibole crystals. Because the matrix minerals are generally undeformed and fill in fractures and spaces between broken crystals, it appears the fibrous crystals are behaving as rigid objects in a more ductile matrix.

C. Description: In Plate 3.6a, some unusual chaotic areas are seen lying either as lobes along side the main amphibole crystal or as more linear zones oriented at a high angle to the crystal axis. Such zones are the 'norm' at crystal-matrix boundaries. Plate 3.6b shows one such zone in more detail and it consists of a micro-agglomeration of randomly oriented amphibole clasts with some chlorite. It is evident that part of the zone is derived from the edge of the main fibre whereas others have interlocking crystal faces. Plate 3.6c shows a zone oriented at a higher angle to the main fibrous crystal. The zone has apparently been separated from its parent crystal by matrix minerals. In this micrograph however the crush zone itself has a curvilinear form suggestive of folding. The limb of the postulated fold in the matrix is a long (>200micron), undeformed asbestiform fibre which at least locally appears to overprint the matrix minerals. However, as the fibre is traced around the hinge of the fold it becomes broken up and forms an integral part of the chaotic zone. Similar fibres are common in thin section and sometimes have a preferred orientation, lying at angles between 20° and 40° to the long axis of the main fibrous crystals.

Interpretation: The lobate and folded morphology of these structures are both suggestive of relative movement of fibrous crystal and matrix, sub-parallel to the long axis of the crystals. Initial cataclasis, aided by alteration, followed by sliding of the crystal past the matrix and accompanied by concomitant grain growth of new amphiboles could produce such textures by scraping off of fine grained amphibole fragments and piling them up into crush zones during fluid movement. Limbs of some of the folded crush zones are made up of long, undeformed asbestiform fibres. Their extreme relative length makes it unlikely they were also ripped off the main crystal and rotated through the matrix. Rather they probably grew during the folding. Such long fibres and interlocking features are common in the matrix. These textures indicate cataclasis and sliding was followed by, or was locally concurrent with, further amphibole growth, thus indicating P-T-x conditions fluctuated rapidly during their formation. A similar origin for microbreccia zones alongside hornblendes undergoing syntectonic retrogression in proto-mylonitic gneisses was proposed by Obee (1985).

D. Description: Deformation textures also characterize the junctions of diverging/merging fibrous crystals where they meet parent bundles. Plate 3.5c, d and e show examples of these junctions and it is evident the fibrous crystals are often joined to the main bundle and are

often in optical continuity with each other (see also Plate 3.4). Commonly however, the structural lattices of the divergent/mergent crystal and parent bundle are mismatched *across* fractures. These fractures may die out along their length (Plate 3.5e) or cut individual crystals giving the appearance of the divergent crystal having been 'ripped-off' from the main bundle (Plate 3.5c). Other fibrous crystals are attached to crystals in the main bundle by the boudinage-fibrous amphibole growth structure.

Interpretation: Although much of the boudinage and other brittle structures have a deformational origin, the timing of the growth of the main fibrous crystals with respect to deformation is more difficult to establish. The structure of the junctions between crystals of different orientations is thought to favour an origin of syndeformational growth for all the amphiboles.

The structural mismatches across fractures and broken fibrous crystals seen at the crystal junctions are suggestive of individual crystals being broken off or rotated away from the main bundle. Although some crystals can be rotated back into their original position others cannot, despite being clearly part of a parent bundle. These fibres also cross-cut boudinaged fibres in other bundles eg Plate 3.5c. It is inferred from this feature that the mismatched crystals and broken crystals grew *after* fracturing and rotation. The cross-cutting relationships and connecting nature of the bundles therefore suggest simultaneous fracturing, rotation and growth was occurring throughout the vein. Matrix minerals, which essentially constitute a hydrothermal mineral assemblage, formed in the resultant dilatant spaces. Similar textures and interpretations of syndeformational growth across fractures and lattice mismatches were reported by Bierman (1977) in elongate (but non-fibrous) tectonically deformed, regionally metamorphosed schists, the so-called 'garbenscheifer texture'. Veblen *et al.* (1977) show, in a HVTEM study, similar lattice mismatches in pyribole asbestos and they infer such features result from syndeformational crystal growth.

The strong preferred orientation of crystals and their vein-like geometry in the field indicates the amphibole spinifex rocks are best interpreted as some form of extension veins that grew under upper greenschist to lower amphibolite facies conditions. They are somewhat analagous to fibrous asbestos veins and quartz veins that form at lower temperatures although they do not appear to have formed by the crack-seal mechanism of Ramsey (1980) because inclusion bands and trails of wall rock material are not present. Rather, the spinifex veins represent zones of brittle-ductile extension which were sites of mineral deposition and therefore fluid flow.

3. Amphibolites

Description: The amphibole crystals in this lithology exhibit a notably different morphology to the fibrous variety. The crystals, which are actinolitic in composition (Fig. 3.2), are unoriented, blocky and prismatic and can be up to 5cm long and about 0.5mm² in cross-section. The crystals are not made up of asbestiform fibres and thus they are probably 'true' double-chain silicate amphiboles. No petrographic evidence of former pyroxenes was observed. The matrix to the crystals comprises variably oriented, dendritic sheaves up to several cm² in size, of fine grained asbestiform amphibole intergrown with lesser amounts of chlorite and talc/serpentine. Some undulose extinction of the sheaves is seen. Textural variation in the larger amphiboles is shown in Plate 3.7.

Like the oriented amphiboles, deformation textures such as undulose extinction, kinking and fracturing are ubiquitous in the large crystals but are generally not seen in the matrix. Bifurcation of single amphibole crystals into two separate crystals across a fracture is seen (Plate 3.7b). These are similar to the lattice mismatch structures in the oriented amphiboles. The two most frequently observed textures are stellate growth (Plate 3.7c), and also clusters of crystals which show complex truncation relationships between the individual component amphiboles (Plate 3.7d).

Interpretation: The stellate and 'crystal cluster' textures are best explained by the simultaneous random nucleation throughout the lithology of actinolite crystals and their subsequent growth. From each nucleation point, radial (stellate) growth of several crystals or individual crystal growth could occur. As these crystals grew they would mutually interfere and so producing the observed complex truncation relationships.

The deformation textures are predominantly confined to the amphiboles suggesting the *amphiboles* deformed as brittle objects in a more ductile matrix. There was apparently no preferred orientation to this deformation. The bifurcation structure is thought to be indicative of syn-deformational growth of the amphiboles as in the fibrous amphiboles - the amphibole grew, was fractured and then continued growing with a structural mismatch across the fracture.

4. Discussion

The petrographic features described above and interpretations thereof, point to the crystallization and growth of the *amphiboles* rather than invoking their being pseudomorphed pyroxenes. An alternative viable explanation for the observed textures can be put forward and, although the pre-existence of pyroxenes can not be totally ruled out, any that were present must have been altered long before the majority of the textures seen

developed.

Many of the textures are suggestive of a syndeformational growth origin for the crystals. However, the deformation may be related to either igneous processes, where the amphiboles represent magmatic amphiboles, or metamorphic/metasomatic processes where the amphiboles represent a secondary crystal growth phenomenon, or even a combination of both. The characteristic of brittle-ductile behaviour between the amphiboles and matrix minerals is thought at present to favour a combined magmatic-metasomatic related deformation origin as described below and shown schematically in Fig. 3.5.

The unoriented actinolite amphiboles are envisaged as having formed in a partially crystallized magma, or 'crystal mush' ("an aggregate of solid crystals lubricated by compressed water vapour" - Krauskopf 1979 p. 342). Deformation of newly formed, more rigid crystals in such a hydrous, more ductile mixture would result in new surfaces becoming available for continued amphibole growth (Fig. 3.5a). The dendritic sheaves of amphibole in the matrix are suggestive of rapid crystallization and overall the crystal mush may have been quenched in its last stages by interaction with external fluids and/or exsolution of volatiles from the mush itself.

The crystal mush was subjected to an external stress field (of unknown orientation) which lead to the development of zones of extension and growth of oriented amphibole crystals essentially as hydrous magmatic veins or dykes but which had access to hydrothermal fluids. As with other proposed fibrous growth mechanisms eg. Phillips (1973), Ramsey and Huber (1983, 1987), the growth of fibres must reflect a delicate balance between rate of extension and rate of chemical supply to the zone (Misch 1969). If the rates are equal, then the amphiboles would grow by diffusion of material into the dilatant space as the walls of the vein separated (Fig. 3.5b). An analogy for this growth would be the separation of two microscope slides under water. If the rate of extension exceeded the rate of chemical supply then it might be expected brittle deformation (boudinage etc.) would occur, probably at lower temperatures and/or under higher fluid pressures (resulting in the observed chlorite alteration) (Fig. 3.5c). The textural features described do suggest that the veins underwent rapid fluctuations in both temperature and fluid pressure. Thus, both types of amphibole-bearing lithology came into contact with hydrous fluids during or in the late stages of, their formation. These fluids would have aided the development of the hydrothermal mineral assemblage seen in the matrix. The temperature of vein formation was probably between 600° and 350°C under conditions of low pressure (<3Kb - Fig. 3.4).

Although the evidence presented here points to a magmatic-metasomatic origin for the *amphibole spinifex* textures seen in this study, the presence of gradations in structure between many of the members of the bi-pyribole family has an important implication for

further petrological work on them. Namely, that the simple optical similarity of pyroxene and amphibole spinifex is not evidence for pseudomorphing. Indeed, the similar basic structural building blocks involved in the biopyribole minerals could conceivably lead to virtually identical textures that result from totally different geological processes. The writer believes that spinifex texture can result from both igneous and metamorphic processes and that detailed petrographic and crystallographic work is necessary to distinguish between them.

3.3 DISCUSSION OF PETROGRAPHIC DATA AND THE NATURE OF THE M1 EVENT

The most notable petrographic feature of the M1 metamorphic event is that all of the simatic rocks have been hydrated to varying degrees and locally carbonated, whilst in most cases, the rocks have preserved their original igneous textures with only slight modification by metamorphism. Thus the M1 event was essentially a metasomatic, low pressure-high temperature event.

On the basis of the petrographic observations, four distinct metamorphic facies may be distinguished in the simatic rocks and these are summarised in Table 3.2. The divisions employed have been defined by what appear to be equilibrium assemblages. The common disequilibrium mineral assemblage, actinolite-chlorite-epidote-albite-quartz, may represent a transitional upper greenschist facies. In the field, some degree of overlap between these different facies is visible within individual litho-tectonic packages of simatic rocks and, furthermore, a strong lithological control is apparent on both the grade and style of metamorphism. Overall, the facies seen define a *retrogressive* sequence of increasing grade, but decreasing intensity, of metamorphism from the extrusive basalts to intrusive gabbros. There was also some associated hydrothermal contact metamorphism close to intruded BIF units (chapter 2.4). Grobler (1972) also documented that the simatic rocks contained retrograde mineral assemblages. The serpentinization and asbestos veins seen in the metaperidotites are thought to be a late stage, lower temperature part of the M1 metasomatism.

If it is accepted for the moment that the intrusive rocks underlie the extrusive rocks in the postulated simatic igneous complex (chapter 2.4), then the observed zonation of metamorphic mineral assemblages could have been produced in response to a thermal gradient imposed from below, and the entry and convective circulation of cooler fluids from above. This would occur if fluid circulation overlapped in time with, or closely followed, the formation of the simatic rocks. Given the estimated preserved thicknesses of simatic

rocks (Table 2.1) this interpretation implies very steep vertical thermal gradients existed in the simatic pile during its metamorphism, possibly as great as 100°C/km.

Three lines of evidence suggest that fluid circulation did overlap with the formation of the simatic complex and that the fluid was most likely to have been Archaean seawater:

1. The presence of pillow lavas implies there was instantaneous fluid-magma interaction. Given this submarine environment of extrusion, it would seem unavoidable that fluid/rock interaction on a larger scale would occur (Spooner and Fyfe 1973).

2. The interpretation of the formation of the spinifex textures implies the magmas probably had access to external fluids.

3. The presence of BIF units intruded by the simatic rocks and intercalated with volcanics (chapter 2.4) also infers the presence of an early metasomatic event. It seems likely that the iron and silica in the BIF units were leached from the simatic rocks as an integral part of the M1 episode. The petrographic and chemical variations recorded in the metagabbros suggested both these elements were released during hydration. Fluid circulation deep into the simatic pile may have possibly been aided by D1 shear zones.

A review of the literature reveals that the textural and metamorphic patterns described in the simatic rocks are in fact remarkably similar to those associated with both oceanic ridge hydrothermal systems and Phanerozoic ophiolites. A comparison of features associated with regional metamorphism, ocean floor metamorphism and features observed in the study area is shown in Table 3.2 and they are interpreted here in a similar fashion (Spooner and Fyfe 1973, Hart 1976a and b, Coish 1977, Stern and Elthon 1979, Evarts and Schiffman 1983, Alt *et al.* 1986, Stakes and Vanho 1986). The strong similarities indicate that the Pietersburg greenstone belt affected by ocean floor metamorphism. Petrographic and stable isotope evidence for early "ocean floor metamorphism", including both hydration and carbonation of the simatic rocks, has been reported in other Archaean greenstone belts by Cloete (1986), Hoffman *et al.* (1986), Barley and Groves (1987), Golding *et al.* (1987) and de Wit *et al.* (1987a).

In such an open hydrologic environment as is envisaged for the Pietersburg area, vast quantities of seawater would have passed through the simatic complex and approximated a "single pass" hydrothermal system where fluid flow rates would generally have tended to exceed chemical reaction rates and hence would cause disequilibrium mass transfer (Fyfe *et al.* 1978, p.127). Furthermore, it would be *expected* that large scale open-system metasomatism and widespread pseudomorphic replacements would occur and hence that the mineral compositions in the lithologies are likely to reflect fluid chemistry at least a much

as rock chemistry (Evarts and Schiffman 1983). In order to quantify mass balance changes in such an environment a considerable amount of chemical data would be needed. Such data accumulation has not been attempted in this study. However, there is sufficient evidence to indicate considerable major element mobility occurred during the hydration and carbonation of the simatic pile. This evidence is also summarized in Table 3.2 and is interpreted as indicating that there was a gross loss of silica, iron and calcium from, and a gross gain of magnesium and possibly sodium to, the simatic pile during its metasomatism. These features should be reflected in the whole rock geochemistry of these rocks.

3.4 WHOLE ROCK GEOCHEMISTRY

3.4.1 Introduction

In order to obtain supporting whole rock data for the postulated elemental mobility, 31 samples of simatic rocks and 1 felsic sample (large ocelli) were selected and analysed by ICPAES. Details of the analytical procedure and the precision and accuracy of this method are given in Appendix 1. The samples were chosen with the aim of covering the complete range of rock types (both intrusive and extrusive) seen in this study. Their location is shown in Fig. 3.1 and the results presented in Table 3.1. The data is presented and used in its 'raw' form. Recalculation to anhydrous values has not been carried out. The results are presented in order of increasing L.O.I. values, some of which are quite large (up to 18%) reflecting the considerable hydration and/or carbonation of the rocks.

Table 3.1 demonstrates that there is a large degree of chemical variation in the simatic rocks which can be interpreted in terms of three models (Smith and Erlank 1982, de Wit *et al.* 1987a);

- a) the primary igneous chemistry controlled the composition and proportion of hydrated phases (hence there was no elemental mobility).
- b) the hydrated phases control the observed chemistry, or alternatively,
- c) the present composition reflects some intermediate situation between these two extremes.

The approach adopted here is to first characterise the relevant igneous processes by considering the variation in abundance of elements considered least mobile under a wide range of metamorphic conditions, such as TiO_2 , Zr and P_2O_5 (Pearce and Cann 1973, Stern

and Elthon 1979), and compare their behaviour with elements that may have been mobilised. Then, using the two end member models above, hypotheses are set up that predict the major element behaviour in each case and these are tested using binary variation diagrams. Molecular proportion ratio plots as employed by Beswick (1982) as a test for element mobility have not been employed because Butler (1986), Rollinson and Roberts (1986) have shown that such ratioing is statistically biased because of the constant sum problem (Chayes 1960) in geochemical analyses.

3.4.2 Igneous processes

1. Incompatible/immobile element variation

For all samples, TiO_2 , Zr and P_2O_5 show positive correlations with each other (Fig. 3.6a and b). This coherency is not affected by rock type or metamorphic grade and is consistent with the postulated relative immobility of these elements. Furthermore, in low pressure mafic-ultramafic magmatic systems Ti and Zr are considered to be incompatible elements and thus are expected to vary systematically with igneous processes such as fractional crystallization and partial melting (Nesbitt and Sun 1976; Coish 1977; Stern and Elthon 1979; Ludden and Gelinas 1982; de Wit *et al.* 1987a). The positive covariance of TiO_2 with Zr (Fig. 3.6b) is therefore also consistent with the field and petrographic data, indicating that the simatic rocks sampled are co-magmatic and evolved as a single intrusive-extrusive igneous complex (chapter 2.4). Ni and Cr values decrease exponentially with increasing Zr content (Fig. 3.6c and d) and this is consistent with early fractionation of olivine and chromite from the mafic liquid (Coish 1977). The rocks containing high chromium and nickel values are the serpentinites (Table 3.1) and hence the data is consistent with the derivation of the gabbros, and then the basalts, from the peridotites by fractional crystallization or partial melting (Cox *et al.* 1983). Combined with the field and petrographic data this suggests that the sequence basalt-gabbro-serpentinite also represents increasing depth in the igneous complex.

2. Major element variation

If the rock suites have not been altered by post-igneous processes, then given the relative immobility and incompatibility of Zr, their whole rock chemistry should exhibit regular trends (straight or exponential curves) of changing element concentration with changing Zr content (Coish 1977). Variation diagrams of Zr against some of the major elements are shown in Fig. 3.7 and it is clear that no linear relationships are present. Fig. 3.7e (MgO vs Zr) exhibits an exponential curve consistent with the early fractionation of MgO rich phases and partitioning of Zr into the residual liquid. However, the other major elements exhibit trends against Zr that are difficult to explain in terms of igneous processes.

suggesting many of the major elements may have been mobilised to varying degrees during post-igneous metamorphism.

If this is correct, it should then follow that there would be very little coherence in the behaviour between major elements. In mafic-ultramafic magmatic systems major element variation diagrams with MgO as the abscissa are commonly used to interpret chemical diversity because it is presumed that solids crystallizing from the magma collectively contain more MgO than the liquid. The correct use and limitations of this type of diagram are reviewed by Cox *et al.* (1983). Here, such diagrams are used to test hypotheses that predict the chemical behaviour in the simatic complex where a crystallization order of olivine-pyroxene-calcic plagioclase is expected. As fractional crystallization or partial melting proceeded, a decrease in MgO content of the remaining liquid should be accompanied by;

- 1) an increase in the SiO_2 and Al_2O_3 content;
- 2) an initial increase in CaO followed by a decrease as plagioclase crystallizes;
- 3) an exponential increase in the incompatible elements such as TiO_2 ;
- 4) a probable slight increase in Fe_2O_3 .

The related variation diagrams are shown in Fig. 3.8 and they show a significant degree of coherence does exist in the major element chemistry and that the trends exhibited are reasonably consistent with those predicted above.

Many writers, have suggested that because of this consistency between predicted igneous behaviour and observed chemical trends, that the related major elements have remained relatively immobile during metamorphism. Such an interpretation assumes that the primary chemistry did directly control the composition and proportion of hydrated phases that formed during metamorphism, and therefore recalculation to anhydrous values is valid. However, in this case such an interpretation is not consistent with either the major element/Zr data (Fig. 3.7) nor the field and petrographic data. In order to investigate this apparent paradox it is first necessary to see if the data set can be interpreted in a different way.

3.4.3 Secondary metamorphism

An alternative approach to analysing the data is to look at element mobility as a function of volatile content.

Hart (1973), and Coish (1977), plotted element concentrations of metamorphosed ocean floor basalts and ophiolitic rocks against their water contents to determine the

direction of movement of elements during low to medium grade metamorphism. They interpreted strong positive correlations of an element with water content as evidence for the addition of that element to the rock suite concerned, and negative correlations as evidence for the element being leached from those rocks. It is implicit in these interpretations that the degree of hydration is *proportional* to the degree of addition or subtraction of an element to or from the rock concerned. Staudigel and Hart (1983) showed in an experiment on seawater/basalt interaction that SiO_2 was leached from the rock during its hydration. A comparison of the chemical composition of the basalt before and after the experiment showed that when the hydrated values were recalculated on a volatile free basis, the SiO_2 value was virtually identical to its original value even though the rock had lost silica. Two important implications arise from this; first, the degree of hydration was proportional to the degree of silica removal and hence the use of the volatile plots may be valid, and secondly, any elements that were added to the basalt would have been *enhanced* over their original values by the recalculation. It should be evident therefore that recalculation of chemical data is a dangerous procedure *until* elemental immobility has been proved.

In a similar fashion to the MgO variation diagrams, hypotheses can be set up that relate elemental variation to metasomatic processes. On the basis of the field and petrographic data presented above the major elements should exhibit the following relationships when compared to increasing L.O.I. values;

- 1) SiO_2 , CaO, Fe_2O_3 and Al_2O_3 should exhibit negative trends because the evidence discussed suggests these elements were released during the hydration of the simatic rocks;
- 2) MgO on the other hand, should exhibit a positive trend if it has been metasomatically added during hydration;
- 3) Samples showing evidence for extensive ankeritic carbonation would be expected to show positive trends for Fe_2O_3 and CaO against increasing L.O.I.;
- 4) There should be no linear trends of the immobile elements TiO_2 and Zr;
- 5) Na_2O should exhibit a positive trend, especially in the basalts and gabbros, due to the albitisation.

The corresponding variation diagrams are shown in Fig. 3.9 and some interesting relationships are apparent. The plots discriminate very effectively a subgroup of 5 samples that petrographic work identified as being significantly carbonated. These have been distinguished on the diagrams and, as predicted, increasing carb nate content is associated with increasing iron and calcium (Fig. 3.9a and b) indicating calcium and iron may have been metasomatically added to these rocks. These samples are clearly altered and, in common with previous write s, it is recommended that carbonated samples should not be used in data sets to be interpreted in terms of igneous processes. Furthermore, in future, L.O.I. values in published analyses should distinguish between carbonate and water content. The 'success' of these plots in ident fy ng t e car onated sampl s s gests th y may be f

use in alteration studies.

If the carbonated samples are ignored, the remaining data set reveals strong negative correlations between L.O.I. (which is inferred to represent the degree of hydration of the samples) and SiO_2 , CaO and Al_2O_3 (Fig. 3.9a, b and c) and a strong positive correlation between L.O.I. and MgO (Fig. 3.9d). No simple relationships exist between L.O.I. and Fe_2O_3 , Na_2O and TiO_2 , (Fig. 3.9e, f and g). Following a similar line of reasoning as for the carbonated samples (and that followed by Hart 1973, and Coish 1977), these relationships are interpreted as indicating that SiO_2 and CaO have been quantitatively leached from, and MgO has been added to, the simatic pile during its hydration. The basalts and gabbros are quite enriched in Na_2O , and although there is no definite trend, is consistent with albitisation of more calcic plagioclase. These observations are consistent with the combined field and petrographic data. Fe_2O_3 does not clearly exhibit the expected trend.

3.4.4 Discussion

It would seem from the above interpretations that the whole rock geochemical data is equivocal and that there is evidence to support both end igneous and metasomatic member models presented in chapter 3.4.1. However, two features of the binary diagrams are considered significant. First, they do not distinguish the carbonated samples to the same degree as the L.O.I./major element diagrams and they could easily be incorporated as 'unaltered samples' in some of the trends. Secondly, the major elements show a higher degree of correlation with L.O.I. than they do with the traditional indices of crystal fractionation (Figs. 3.6 and 3.7). Such features, when combined with the field and petrographic evidence suggest that alteration has played an important part in modifying the chemistry of the simatic rocks in the belt.

However, the use of the volatile plots is in itself fraught with difficulties because if the degree of hydration is directly controlled by the igneous chemistry, then it would be expected that the most magnesium (olivine) rich rocks may be the most susceptible to alteration; a circular argument! It is pertinent then to ask why, given the strong field and petrographic evidence for metasomatism, is the data set not more unequivocal? One of the reasons may be that the metasomatism is cleverly disguised.

If, for instance silica is leached from the rocks and magnesium is added to them, possibly in proportion to each other, then the overall effect would be to simply enhance the expected igneous differentiation trend of magnesium decrease coupled with silica increase. Therefore, although MgO does reflect a sympathetic relationship with SiO_2 (and Al_2O_3) in these rocks (Fig. 3.8), the possibility exists that these trends have been enhanced or even imposed by metasomatism and that the values of these oxides are not the original igneous

values. If this is the case, then the interpretation that, if binary variation diagrams can be interpreted in terms of igneous processes then the elements concerned have remained essentially immobile, is not valid. The reason it is not valid is that there is a hidden assumption in the interpretation and one that is inherent in the use of ratioing techniques to detect alteration (Beswick, 1982, Barnes, 1985 and Butler, 1986). That assumption is;

metamorphism acts to randomise the chemical components of the system.

If SiO_2 removal and MgO addition has occurred in tandem, then it is evident that metamorphism did not randomise the chemical components but rather, by the unique nature of the metasomatism, produced "pseudo-igneous" differentiation trends (Hart and de Wit 1984). Experimental data shows this type of metasomatic behaviour may indeed occur.

Bischoff and Dickson (1975), Mottl and Holland (1978), Bischoff and Seyfried (1978), Hajash and Chandler (1981), Reed (1983), and Janecky and Seyfried (1986) have all carried out experimental seawater/basalt interactions and shown that at moderate to high temperatures (200–500°C) over a wide range of water/rock ratios, magnesium is indeed precipitated out of the seawater and silica leached from the basalt. This process occurs even at low degrees of hydration (1 to 6% H_2O added). In some cases eg Reed (1983), the basalt was chemically transformed by the precipitation of magnesium rich silicates to a peridotite composition.

At lower temperatures (<150°C) such patterns are often reversed and there is the potential for some very complex chemical patterns to develop. Fig. 3.10 summarizes the experimental data for magnesium and silica. Seyfried (1987) in a review of seawater/rock interactions shows that the behaviour of other elements such as calcium and iron is more complex.

The alteration model is also supported by metamorphic petrology and geochemical studies of Phanerozoic ocean floor rocks and ophiolites (Hart 1973, 1976a and b, Coish 1977, Stern and Elthon 1979, Evarts and Schiffman 1983, Alt *et al.* 1986). Stakes and Vanho (1986) have shown that ocean floor metamorphism of gabbros is polyphase and very complex because of later hydrothermal activity that occurs as the water/rock system cools. Such a feature was recorded in gabbros in this study (chapter 3.2.3). There is also a growing body of stable isotope evidence for the early pervasive and widespread simatic rock seawater interaction in greenstone belts eg Beatty and Taylor (1982), de Wit *et al.* (1987a), Golding *et al.* (1987), Barley and Groves (1987).

In summary, it is likely that both magmatic differentiation and mobilisation during metasomatism have produced the chemical variations demonstrated in the rocks. However,

the field petrographic and geochemical data are more consistent with a model involving the metasomatic alteration of the simatic rocks that has resulted in significant major element mobility on a regional scale.

3.5 SUMMARY AND DISCUSSION

The field, petrographic and geochemical data suggest that the simatic rocks in the belt evolved as one intrusive-extrusive complex in which a low pressure differentiation sequence of peridotite-gabbro-basalt occurred. Gabbroic magma chambers may have played an important role in this igneous evolution. The simatic pile underwent a high temperature-low pressure metasomatic event that extensively hydrated (and locally carbonated) the pile during or soon after its formation and resulted in significant mobilisation of at least Mg, SiO₂, Ca, and Fe. The metamorphic mineral assemblages define a sequence of increasing metamorphic grade but decreasing intensity from the basalts to the gabbros. The metaperidotites were extensively serpentinised. This alteration is very similar to that seen in Phanerozoic ocean floor rocks and ophiolites. It is not possible at this stage to say how thick the original igneous sequence was although an estimate based on preserved sequence thicknesses suggests the complex was between 3.5 and 10 km thick. These interpretations are shown schematically in Fig. 3.11.

A further important observation is that although many of the amphiboles seen are secondary, the spinifex-textured amphiboles are not pseudomorphed pyroxenes. Such an interpretation implies that either the parental magmas were quite hydrous and/or they had access to external fluids.

Archaean greenstone belts are commonly assumed to contain remnants of ancient oceanic crust (see Windley 1976, 1984, and Condie 1981 for overviews) and therefore the mafic and ultramafic units and their associated cherts have been implicitly equated to Phanerozoic ophiolite complexes (de Wit *et al.* 1987a). Helmstadt and Padgham (1986), Kusky (1986), Harper (1986) and de Wit *et al.* (1987a) have all interpreted some greenstone belts as being ophiolite complexes that are direct analogues of Phanerozoic oceanic crust. The Pietersburg belt, although not conforming to the Penrose conference definition of an ophiolite (Anonymous 1972), does contain a Steinmann Trinity of hydrothermally altered pillow lavas (or spilites), cherts and serpentinites. Layered gabbros are also present. Therefore, the 'Steinmann Trinity' is a useful working hypothesis that can be used in further studies in the area. A key field issue would be to investigate the relationship between intrusive and extrusive rocks within individual structural packages to see if sheeted dykes are present. If present, they would provide the strongest evidence for some form of oceanic

spreading having occurred in the Archaean. Given that the belt is tectonically dismembered one problem in recognising possible ophiolite sequences is that parts of the 'Penrose' stratigraphy could easily be missing.

Although there is insufficient evidence at present to infer the Pietersburg greenstone belt is an ophiolite, it does appear to have been affected by ocean floor metamorphism. If such hydrothermal activity affected other greenstone belts then there is an important implication. If magnesium metasomatism of the simatic rocks has occurred, then it is difficult to state with confidence the original magnesium content of the magmas involved. Estimates of the magnesium content are vital in the evaluation of the temperature of the Archaean upper mantle and degrees of partial melting required to produce the parental liquids to the simatic rocks (Bickle 1982, Elthon 1986). Such parameters have been used in models of Archaean Plate tectonics (Arndt 1983). If magnesium metasomatism has affected the lithologies then it may be that these parameters have been overestimated in the past. It is interesting to note that the parental liquids to the recently reported relatively fresh komatiites from the Belingwe area (Nisbet *et al.* 1987) were inferred to contain 17 to 20% MgO, significantly less than the 25 to 30% estimated from altered komatiites from the same belt (Nisbet *et al.* 1977).

A further implication if ocean floor metamorphism has affected greenstone belts concerns the concentration of gold. Saager *et al.* (1982) analysed 15 samples of undeformed BIF from the south-west Pietersburg belt for their Au content and found that they contained an average of 191 ppb Au (range 3.7 to 667 ppb). They proposed that this gold had been leached from the simatic rocks during early seafloor hydrothermal activity along with the iron and silica. This is the earliest phase of gold mineralisation seen in the Pietersburg belt but as yet no economic concentrations have been found. If the fluid involved in the alteration of the simatic rock is inferred to be Archean seawater it is likely the fluid was rich in Cl⁻ based ligands. Thermodynamic considerations (Henley 1985, Neall 1987) suggest that Cl⁻ ligands are not suitable for forming economic concentrations of *gold-only* deposits. Indeed many auriferous BIF units have been interpreted to have had their gold introduced epigenetically (Phillips *et al.* 1984). This suggests concepts of syngensis may not be useful in exploration programmes for gold. However, such early gold concentrations may be amenable to upgrading (or even downgrading) during subsequent hydrothermal activity and/or deformation metamorphism.

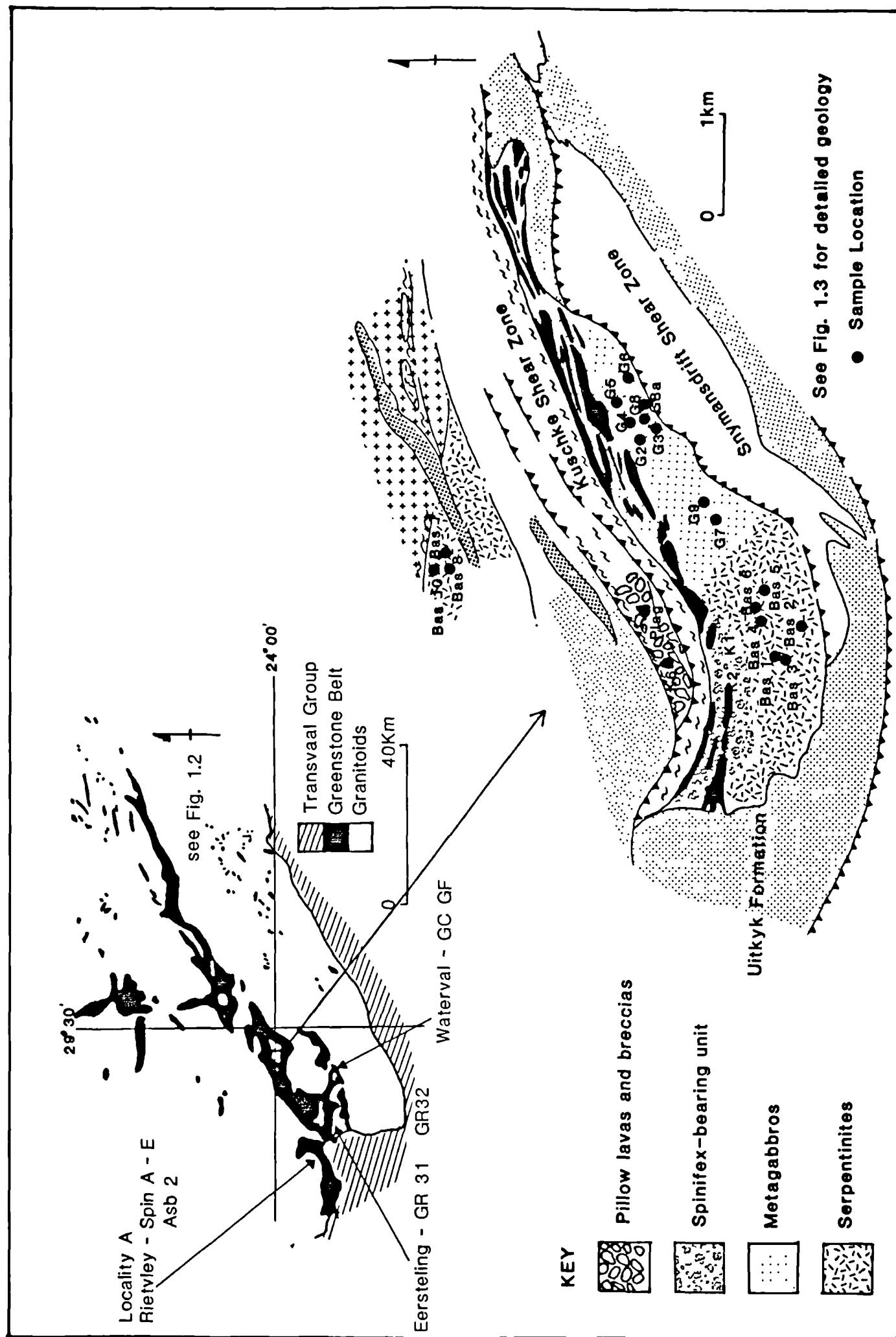


Fig. 3.1 Geology of the Simatic Rocks and Sample Locality Plan

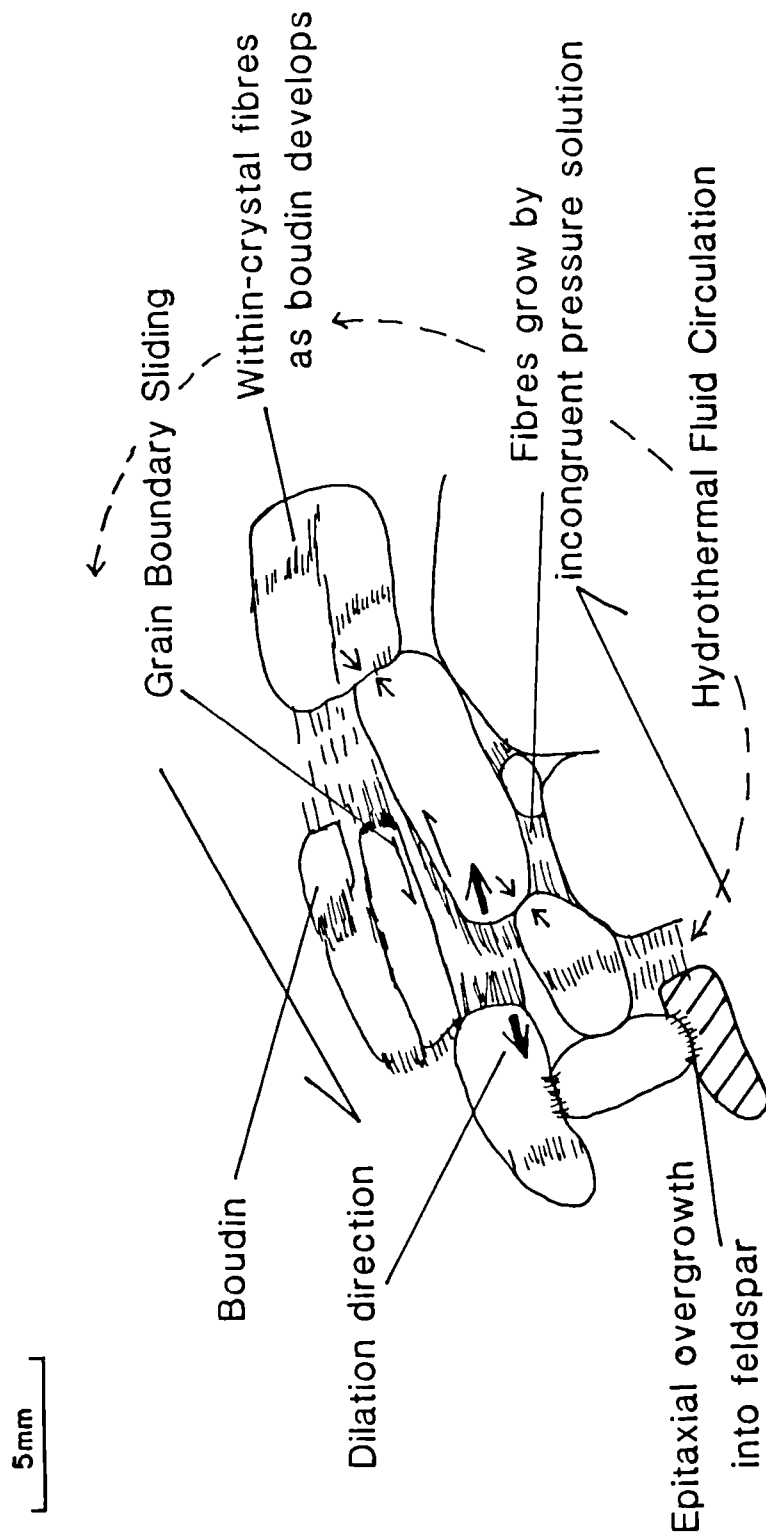


Fig. 3.3 Mechanism for Fibrous Growth
(modified after Rutter et al. 1985).

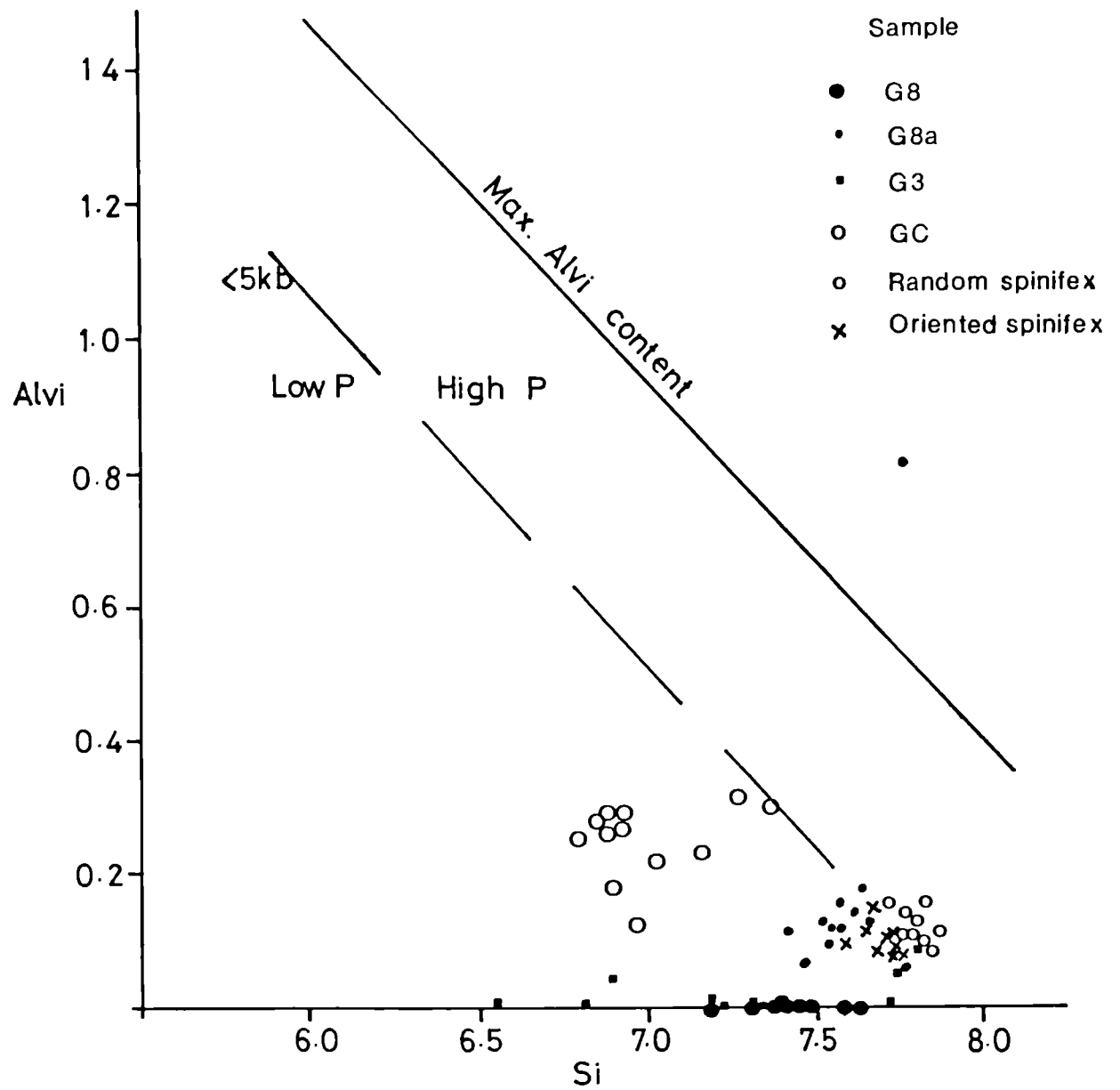


Fig. 3.4 Si vs Alvi occupancy for amphiboles
(after Raase 1974)
data from Tables 3.3 and 3.5

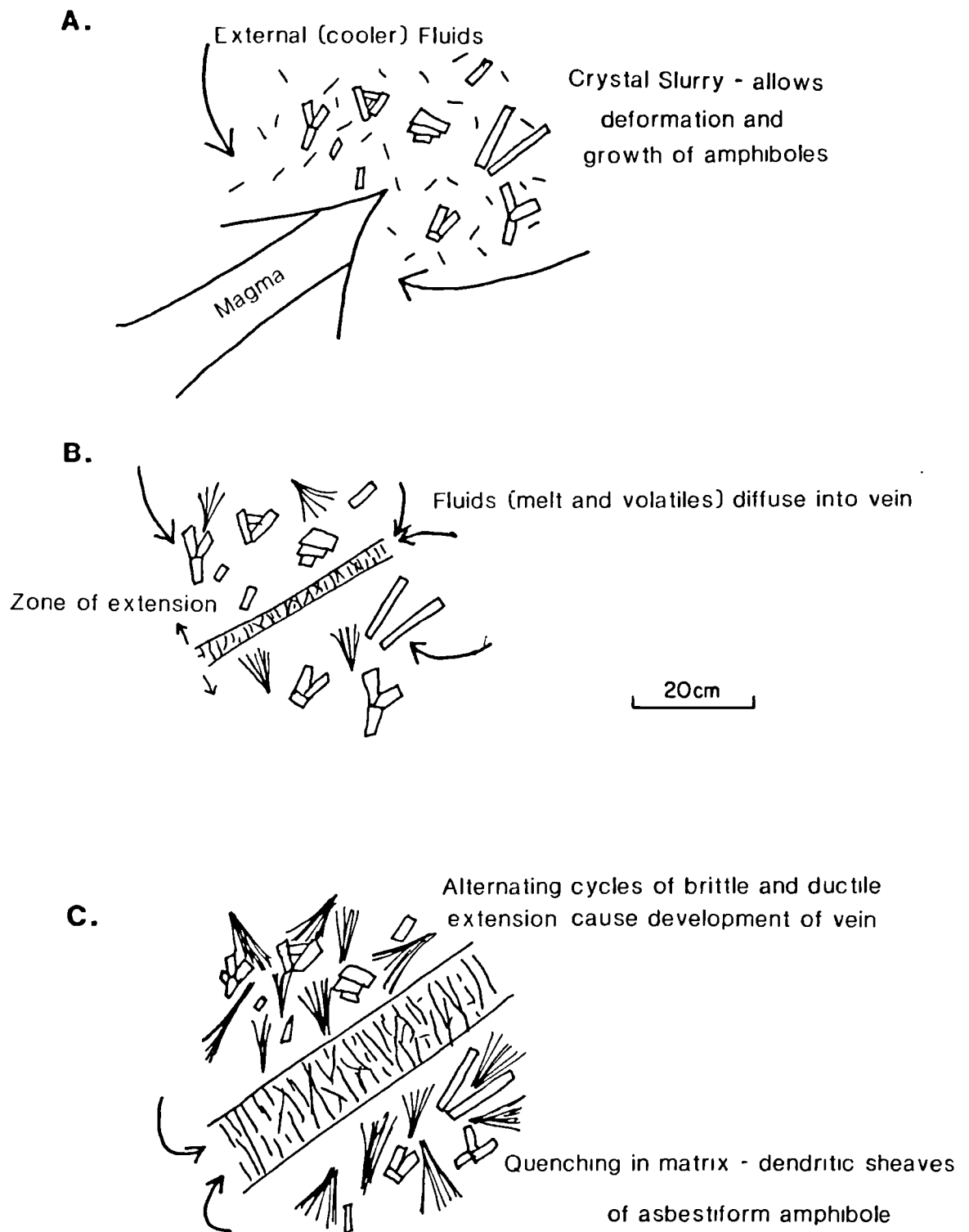


Fig. 3.5 Formation of amphibole spinifex textures

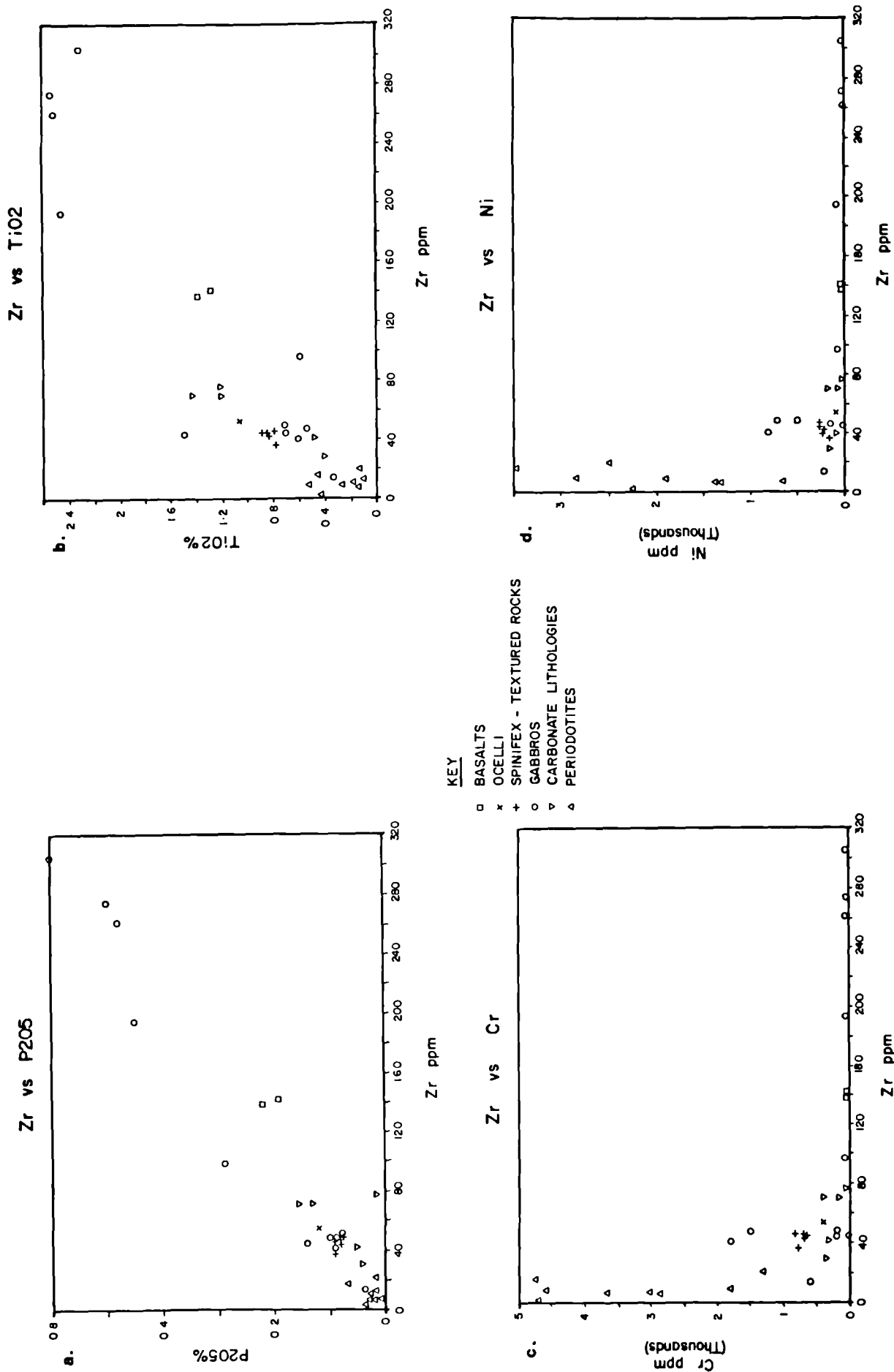
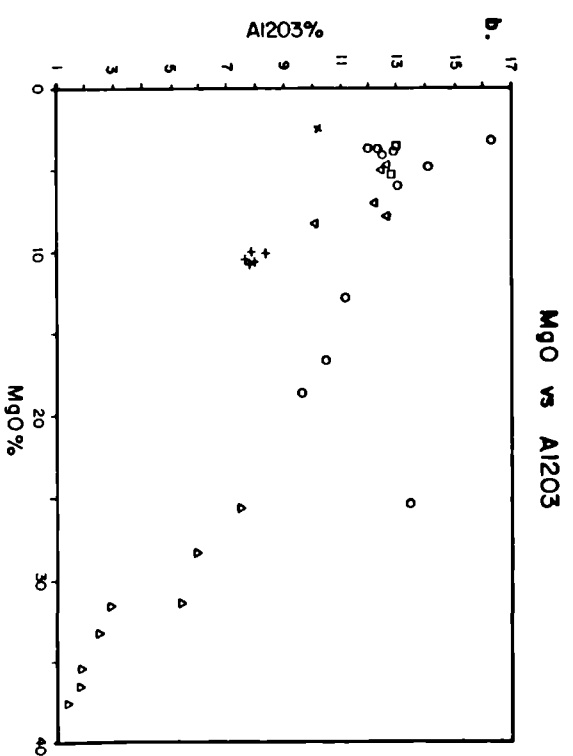
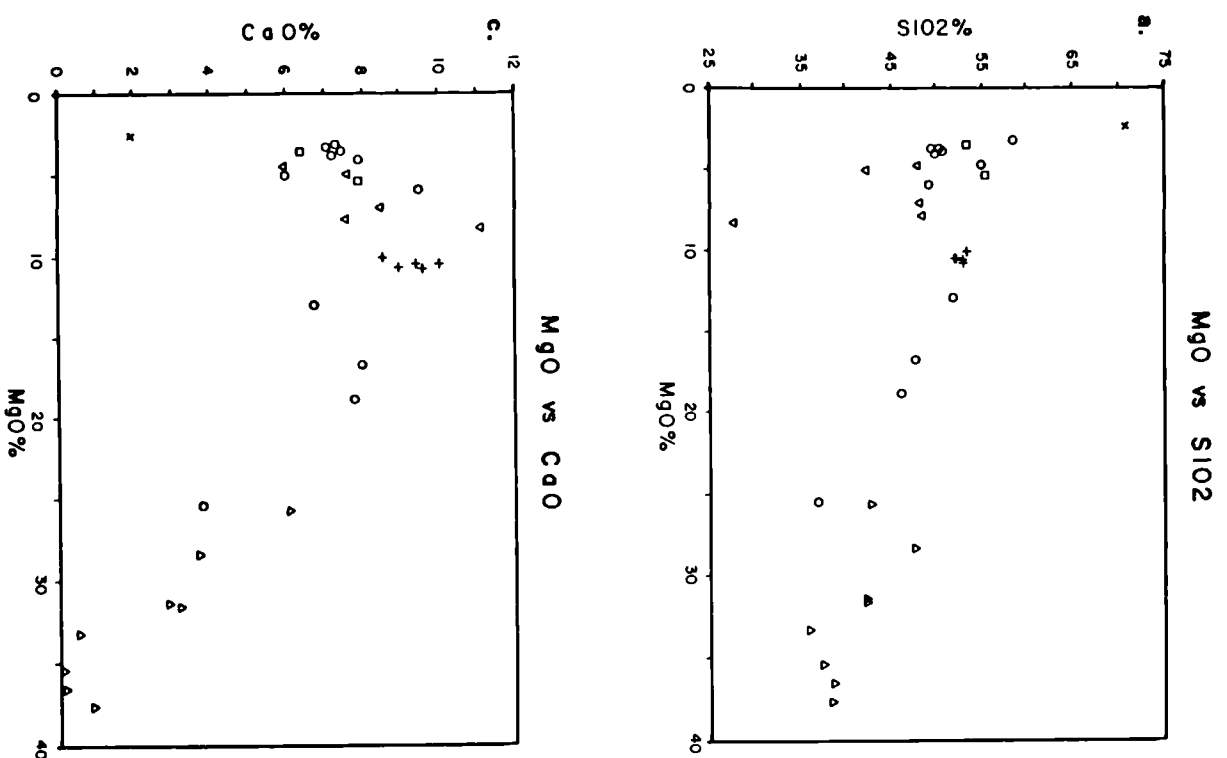


Fig. 3.6 Binary variation diagrams of Incompatible/Immobile elements



KEY

- BASALTS
- x OCELLI
- + SPINFEX - TEXTURED ROCKS
- GABBROS
- ▽ CARBONATE LITHOLOGIES
- △ PERIODOTITES

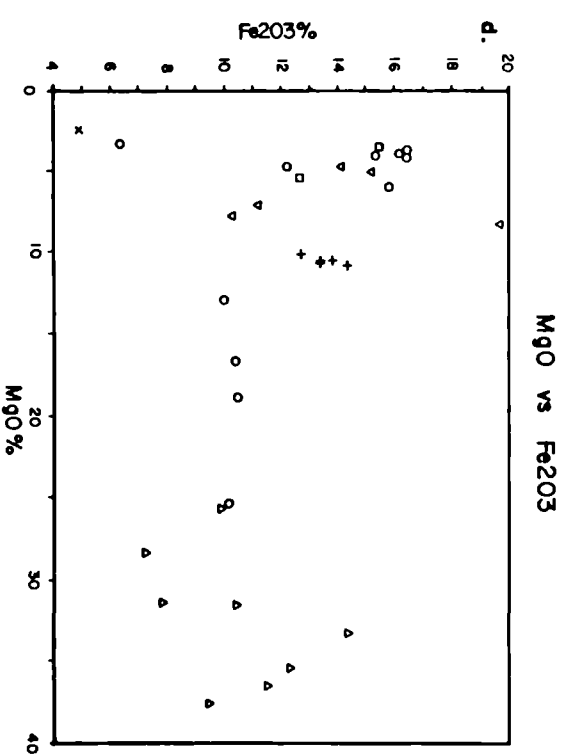


Fig. 3.8 Major element variation diagrams

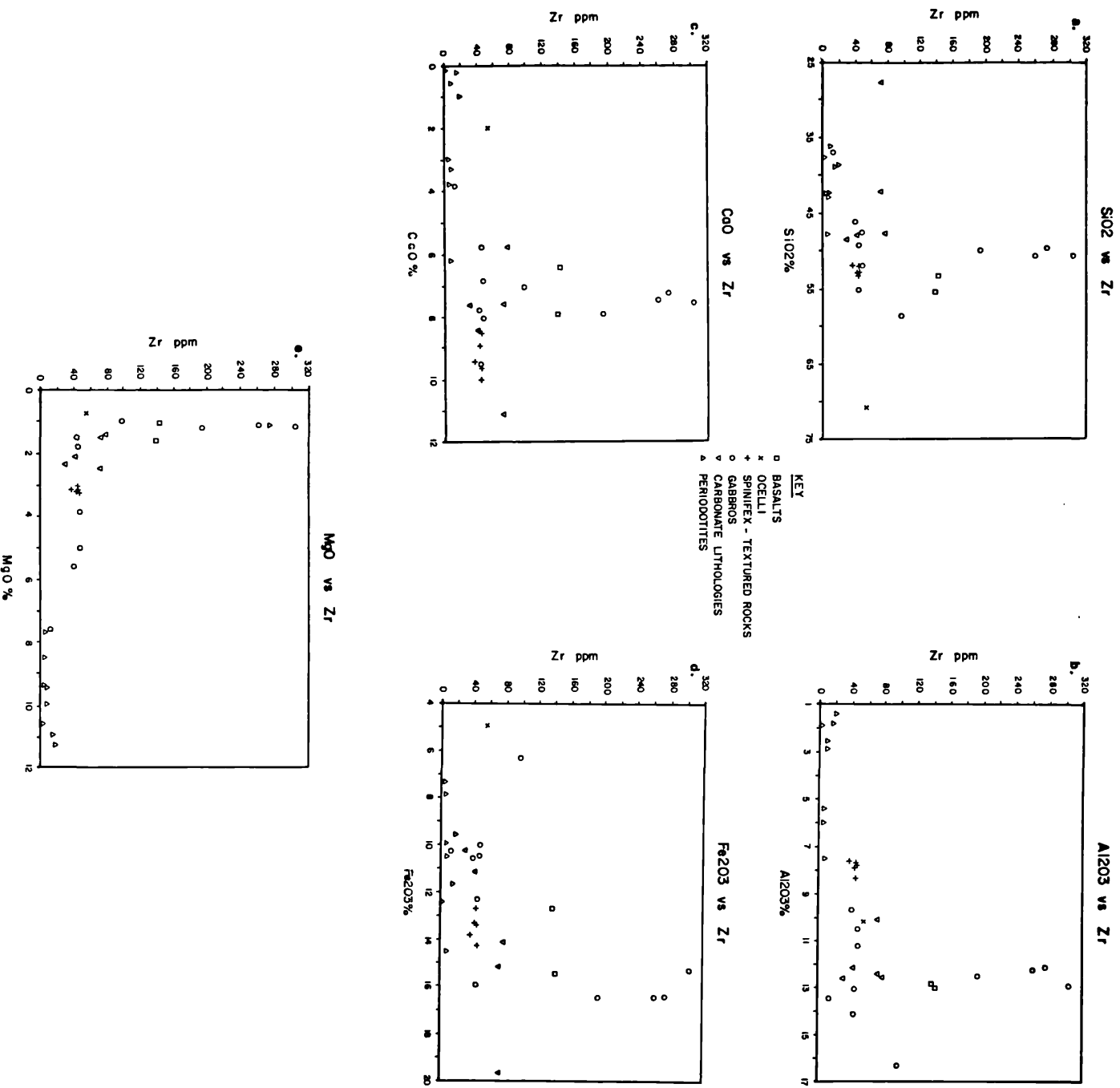
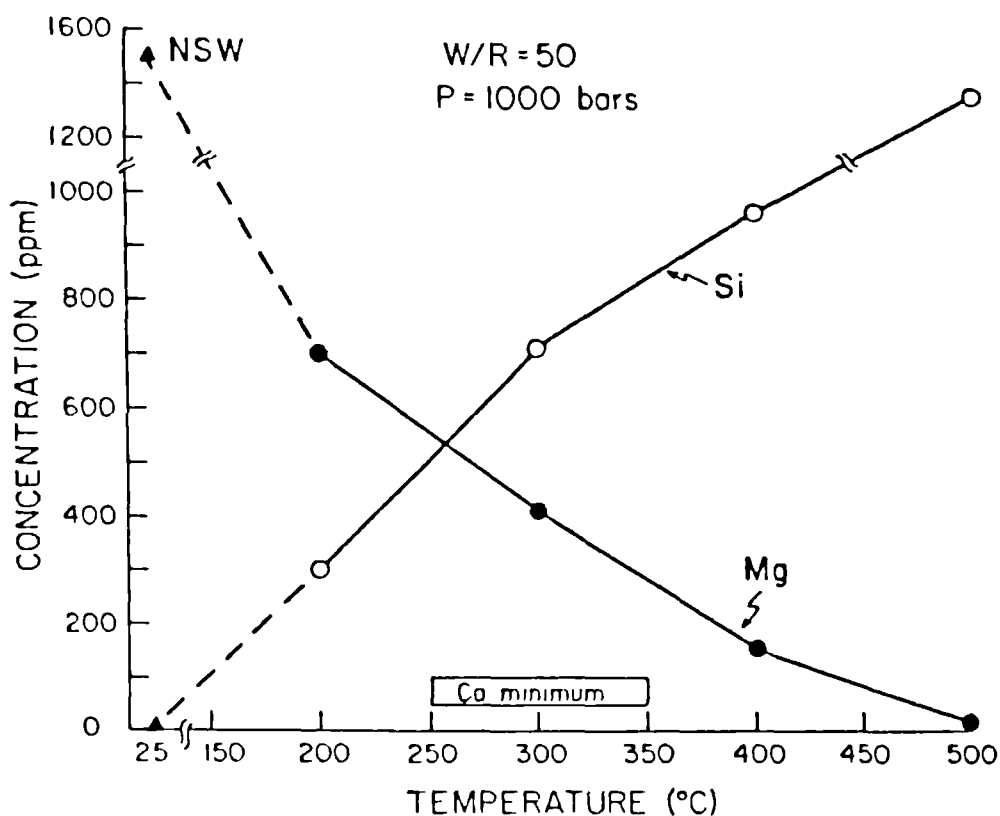


Fig. 3.7 Zirconium/Major element variation diagrams

Fig. 3.10 Summary of Seawater-Basalt Interaction**(from de Wit et al. 1987a)**

The fluid composition of Si and Mg in seawater/basalt experiments plotted as a function of temperature. The data is adopted from Hajash and Chandler (1981) for a water/rock ratio of 50 and a pressure of 1000 bar. The plot indicates the Mg content of seawater decreases with temperature probably because of deposition of Mg (OH) rich phase such as chlorite, saponite, actinolite-tremolite and serpentine. At elevated temperatures Si is leached from the rock and subsequently precipitated at low temperature. Unlike Si and Mg which vary linearly with temperature, the solubility of Ca in seawater describes a parabolic curve with a minimum in the 250–350°C temperature range.

TABLE 3.2 SUMMARY OF METAMORPHIC PETROLOGY AND EVIDENCE FOR OCEAN FLOOR METAMORPHISM IN THE PIETERSBURG GREENSTONE BELT

Metamorphic Petrology

Facies	Low to mid-greenschist facies
Mineral assemblage	Palagonatised glass, chl, ep, ab, sphene, quartz, carbonate
Inferred max temp	350 °C
Lithology	Extrusives and some gabbros
Facies	Lower amphibolite facies
Mineral assemblage	Fibrous actinolite and actinolite, ab, sphene, quartz, carbonate, talc
Inferred max temp	650–350°C
Lithology	Metagabbros and some extrusives. Minor textural obliteration new metamorphic growth
Facies	Upper amphibolite facies
Mineral assemblage	Blocky euhedral magnesio-hornblendes and ferro hornblendes, calcic plagioclase, sphene, quartz. Textural preservation.
Inferred max temp	750–600°C
Lithology	Layered metagabbros
Facies	Serpentinite
Mineral assemblage	Chrysotile, antigorite lizardite, talc, magnetite, carbonate
Inferred max temp	<500°C
Lithology	Peridotites and dunites. Mixed textural preservation.

Evidence for Metasomatism

Element	Fe
Evidence	BIF units, increasing Mg/Mg+Fe ratios in amphiboles, serpentinization and associated magnetite-siderite veins, carbonation of associated extrusives.
Inferred gain/loss from simatic complex	Loss
Element	Ca
Evidence	Serpentinization, ubiquitous epidote in gabbros and veins throughout complex, albitisation, carbonation of extrusives.
Inferred gain/loss from simatic complex	Loss
Element	Si
Evidence	BIF units, serpentinization and associated quartz veins, ubiquitous quartz veining throughout complex
Inferred gain/loss from simatic complex	Loss
Element	Na
Evidence	Widespread albitisation
Inferred gain/loss from simatic complex	Gain

Summary of features reported from studies of ocean floor metamorphism found in the Pietersburg greenstone belt – from references mentioned in text

1. Effects are restricted to greenstones and occurred early in the history of the simatic rocks concurrent or soon after extrusion/formation
2. Rapid lateral and vertical variations in intensity of metamorphism going from greenschist to amphibolite facies at between 2 and 3km depth. Grade and intensity increase downwards
3. Maximum grade reached in metagabbros, which are classified as amphibolites, but widespread textural preservation
4. The textures indicate a sequential series of metamorphic reactions occurring during decreasing temperature. Thus there is no prograde event *sensu stricto* and the greenschist amphibolite transition with depth is a retrograde boundary
5. Ubiquitous development of metamorphic minerals up to amphibolite grade without development of schistosity and textures indicative of widespread disequilibrium.
6. Ragged and feathery grain boundaries developed during essentially pseudomorphic replacement thus allowing textural replacement
7. Calcic plagioclase replaced by albite, then clinozoisite, epidote and chlorite. Often myrmekitic and granophyric intergrowths of replacement minerals
8. Replacement of ilmenite by sphene
9. Serpentinization of ultramafics

These features as a group are not found in regional, dynamothermal metamorphic schist terrains

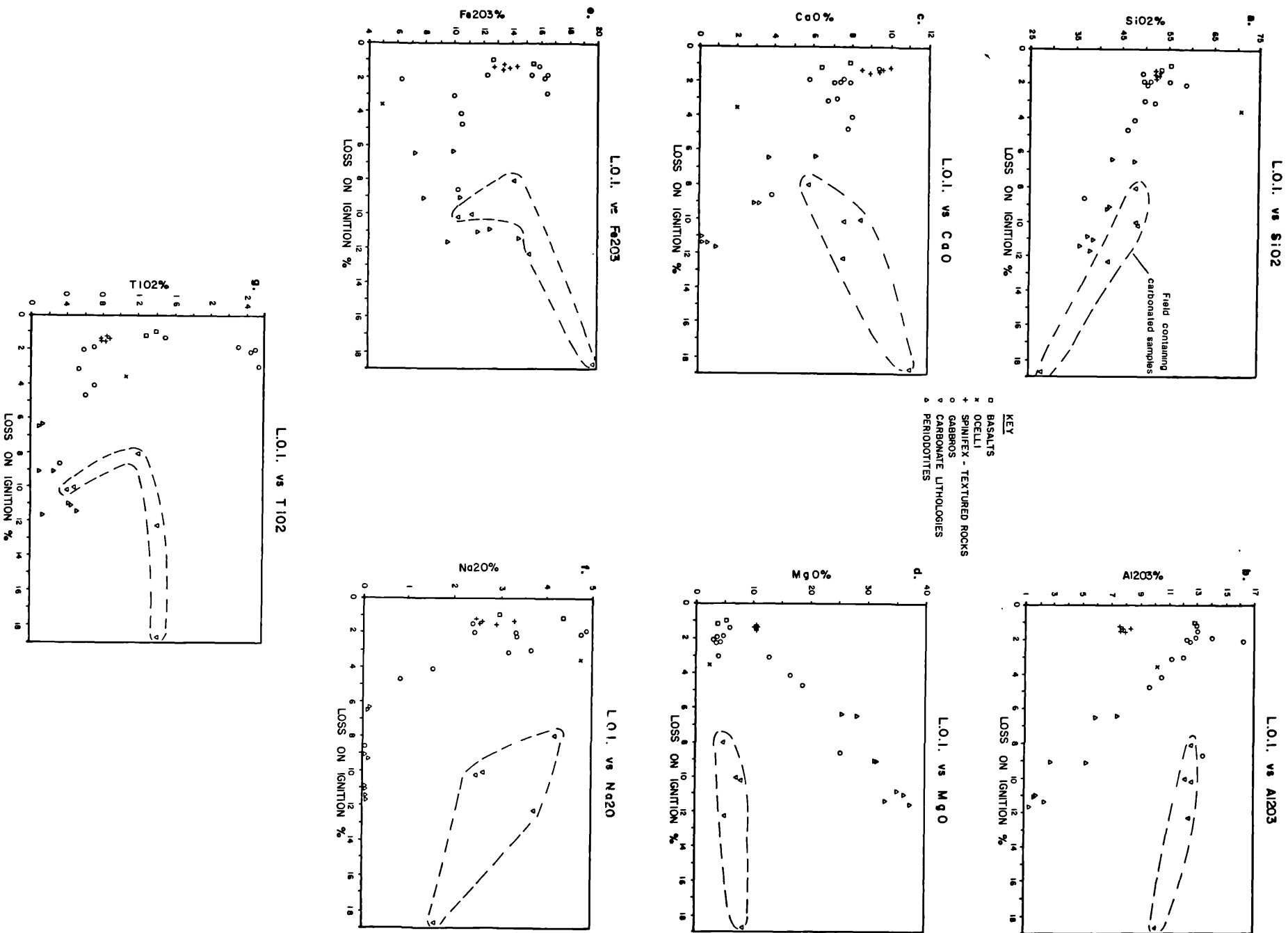


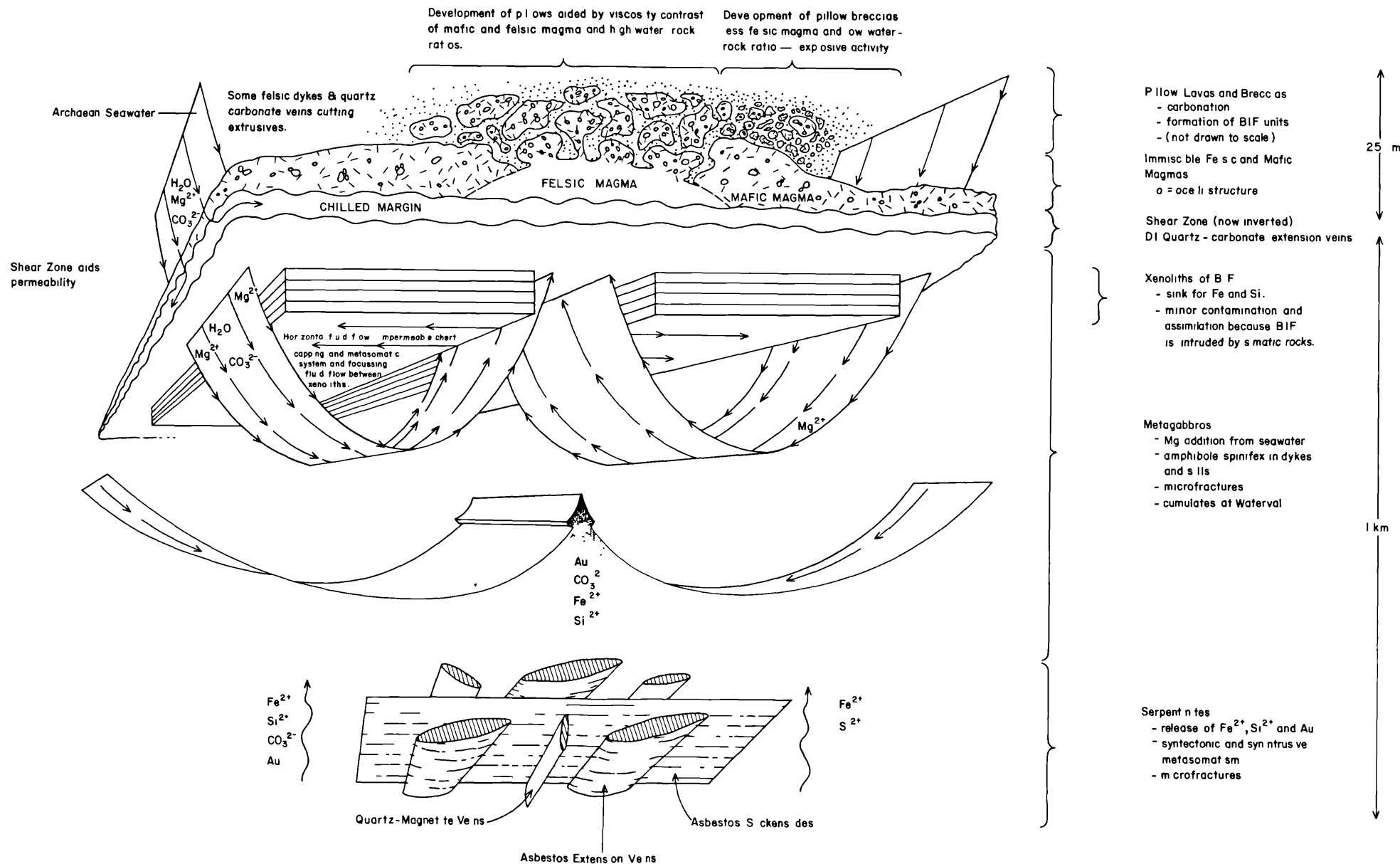
Fig. 3.9 Variation diagrams for L.O.I. vs major elements

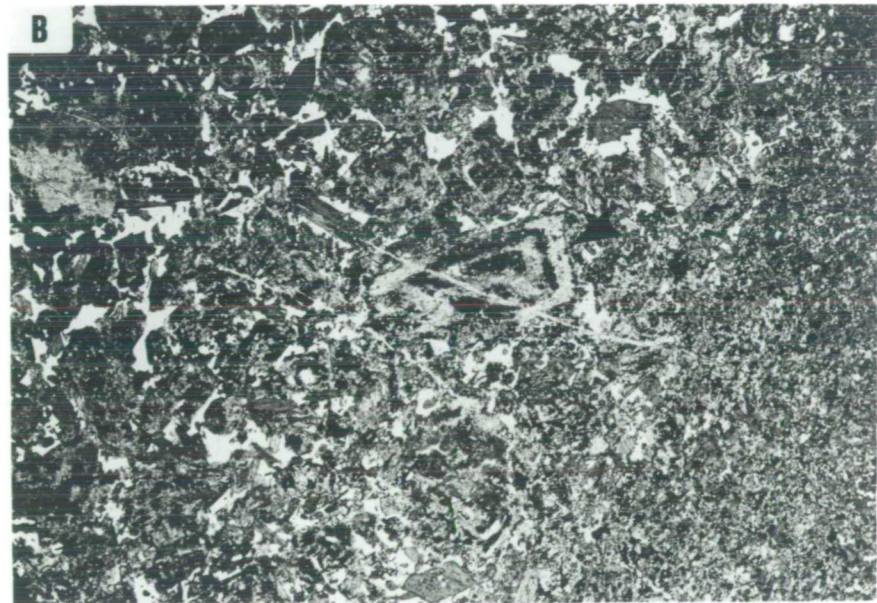
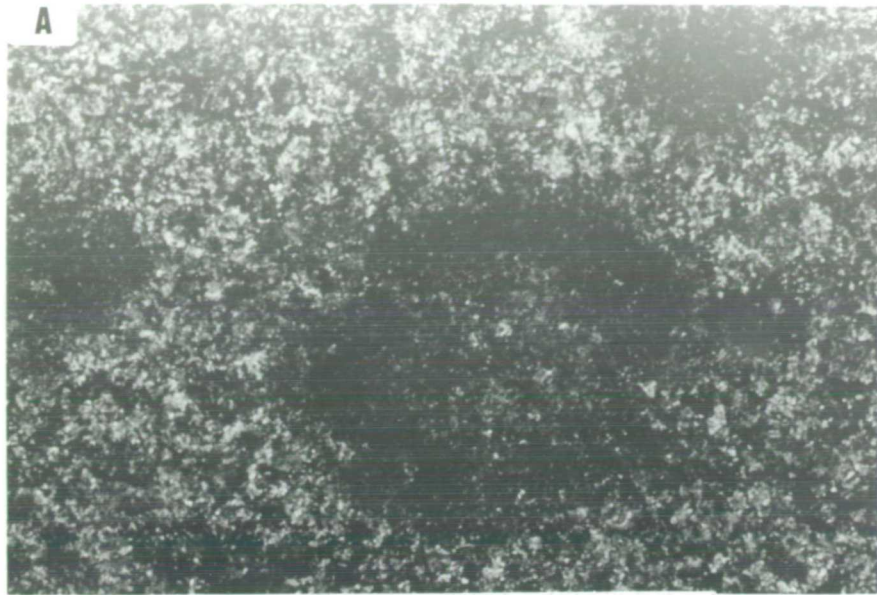
Plate 3.1 Thin section observations of simatic rocks.

- 3.1a.** Ocelli (dark spheroids) made up of a fine-grained intergrowth of quartz and feldspar lying in a heavily carbonated matrix that also contains chlorite and quartz. Field of view - 8mm (XPL). Sample K6.
GR X 2 657 840, Y -33 100.
- 3.1b.** Margin of coarse and fine grained metagabbro from the Waterval prospect, showing a feldspar with remnant zoning (arrowed) lying in a matrix of euhedral amphiboles and a large percentage of free quartz. Field of view - 2.4cm (PPL). Sample DDH WV2, 75.2m.
- 3.1c.** Spene showing trellis-like habit suggestive of ilmenite (Stern and Elthon 1979). Similar crystals have been found. Field of view - 4mm (PPL). Sample G3 (see Fig. 3.1 for location).

Fig. 3-11 **SYNOPTIC DIAGRAM OF THE Mt. MARE INTRUSIVE-EXTRUSIVE SIMATIC TECTONO-COMPLEX**

- VIEWED UPWARDS FROM MANTLE
- VERTICAL SCALE APPROXIMATELY 1:5000





TAB E 3.1
WHOLE ROCK GEOCHEMISTRY OF SIMATIC ROCKS

	GR3(1)	GR3(2)	SPIN A	SPIN B	SPIN C	GF	SPIN D	SPIN E	GC	G9	G2	G3	G4	G5	G6	G8a
S 02	55.48	53.37	52.06	53.31	52.87	49.27	51.94	52.84	58.67	50.73	55.18	50.59	49.99	49.67	51.98	47.62
A 2.3	12.83	8.31	7.65	8.31	7.78	13.04	7.63	7.93	16.50	12.90	14.10	12.33	12.50	12.10	11.19	10.52
Fe203	12.67	15.44	13.40	12.68	14.30	15.91	13.82	13.38	6.55	15.37	12.28	16.48	16.44	16.41	10.02	10.47
MnO	5.37	3.58	10.52	10.16	10.73	5.98	10.51	10.59	3.30	3.96	4.87	3.85	4.09	3.81	12.80	16.68
CaO	7.93	6.43	10.01	8.52	9.62	9.50	9.43	8.93	7.08	7.58	5.80	7.47	7.93	7.23	6.82	8.00
Mg2O	2.96	4.37	2.46	3.29	2.59	2.42	2.53	2.88	4.78	2.45	4.87	3.36	3.37	3.68	3.18	1.55
K2O	1.02	0.53	0.12	0.14	0.17	0.03	0.17	0.17	1.13	1.73	0.04	0.34	0.22	0.40	0.04	0.15
PT05	0.22	0.19	0.09	0.09	0.08	0.14	0.09	0.08	0.29	0.60	0.08	0.48	0.45	0.50	0.10	0.09
TiO2	1.39	1.29	0.86	0.89	0.79	1.50	0.79	0.84	0.59	2.32	0.71	2.51	2.46	2.54	0.53	0.72
MnO	0.15	0.23	0.21	0.17	0.19	0.17	0.18	0.19	0.08	0.19	0.12	0.20	0.19	0.20	0.15	0.16
H2O-	0.19	0.27	0.17	0.05	0.16	0.07	0.12	0.11	0.02	0.43	0.20	0.28	0.22	0.61	0.30	0.18
LOI	0.98	1.19	1.27	1.37	1.39	1.41	1.50	1.57	2.09	1.90	1.92	2.01	2.13	3.02	3.12	4.12
TOTAL	101.19	99.89	98.82	98.98	100.67	99.46	98.71	99.51	100.70	100.16	100.17	99.90	99.99	100.15	100.34	100.26
Ba	218	182	34	40	50	180	47	54	468	443	23	198	145	300	22	109
Be	1	1	0	0	1	1	0	1	2	2	0	2	3	2	1	0
Cr	47	38	687	629	812	182	775	681	71	56	15	53	70	56	191	1495
Cu	195	200	133	16	81	75	28	59	65	55	4	80	88	104	51	57
La	6	18	4	6	14	11	0	6	54	38	0	34	37	42	16	8
Ni	37	29	217	177	216	142	175	187	70	18	21	11	92	6	512	699
Sr	176	320	94	39	64	500	67	48	978	198	152	153	168	162	145	699
V	313	203	225	224	230	328	214	234	132	288	312	328	350	338	160	189
Zn	104	159	104	103	117	124	109	114	67	123	55	140	146	135	79	83
Zr	138	142	45	45	46	44	37	43	97	304	45	261	194	274	48	48
Co	61	66	66	67	73	93	58	80	76	37	40	41	67	36	82	59
A 203/TiO	9.23	10.08	8.90	9.34	9.85	8.69	9.66	9.44	27.93	5.56	19.86	4.91	5.08	4.76	21.11	14.61
CaO/TiO2	5.71	4.98	11.64	9.57	12.18	6.33	11.94	10.63	12.00	3.27	8.17	2.98	3.22	2.85	12.87	11.11
CaO A 203	0.62	0.49	1.31	1.03	1.24	0.73	1.24	1.13	0.43	0.59	0.41	0.61	0.63	0.60	0.61	0.76
MgO/TiO2	3.86	2.78	12.23	11.42	13.58	3.99	13.30	12.61	5.59	1.71	6.86	1.53	1.66	1.50	24.34	23.17
Zr/TiO2	99.28	110.08	52.33	50.56	58.23	29.33	46.84	51.19	164.41	131.03	63.38	103.98	78.86	107.87	90.57	68.67
Fe203/MiO	2.36	4.31	1.27	1.25	1.33	2.66	1.31	1.26	1.92	3.88	2.52	4.28	4.02	4.31	0.78	0.63
MnO	0.16	0.17	0.12	0.19	0.16	0.09	0.14	0.18	0.17	0.14	0.26	0.17	0.53	1.43	1.43	1.07
H2O-	0.23	0.25	0.08	0.11	0.64	0.36	0.29	0.15	0.26	0.18	0.32	0.18	0.63	0.28	0.52	0.06
LOI	4.73	6.37	6.50	8.05	8.60	9.08	9.08	10.04	10.21	10.85	11.05	11.68	11.40	12.29	18.72	3.56
TOTAL	99.80	99.47	100.16	98.89	99.61	99.91	100.51	100.38	100.38	89.29	101.31	100.39	99.66	100.69	99.45	100.03
Ba	123	13	11	35	6	8	23	15	13	47	34	10	66	74	22	19
Be	0	0	0	1	0	0	0	0	0	0	1	0	0	0	1	1
Cr	1777	3035	2871	50	596	3678	1820	308	348	4718	4749	1323	4586	155	382	376
Cu	65	28	16	162	0	7	40	66	33	40	65	3	54	138	22	322
La	0	0	3	0	0	0	11	10	6	0	8	6	0	1	1	14
Ni	806	664	1324	218	229	1383	1902	127	149	2261	3499	2507	2847	71	176	97
Sr	57	21	18	216	10	14	18	102	70	10	11	9	16	172	161	69
V	150	150	78	169	189	94	91	251	210	83	98	39	111	317	252	180
Zn	103	53	57	107	44	56	68	78	83	80	102	30	105	106	143	48
Zr	41	8	7	77	13	7	10	42	39	3	17	21	10	71	71	54
Co	66	77	76	42	53	83	114	57	56	156	188	97	175	46	61	58
Al2O3/TiO	15.87	53.79	59.80	10.39	40.73	49.36	10.78	25.29	31.45	4.36	3.98	10.00	4.77	8.66	7.03	9.51
CaO/TiO2	12.77	44.07	37.70	4.78	11.67	26.91	12.11	17.67	18.98	0.32	0.43	6.93	1.00	5.29	7.78	1.86
CaO Al2O3	0.80	0.82	0.63	0.46	0.29	0.55	1.12	0.70	0.60	0.07	0.11	0.69	0.21	0.61	1.10	0.20
MgO/TiO2	39.74	183.28	283.00	3.93	76.70	285.45	116.93	14.77	19.80	80.39	79.28	268.21	62.66	3.54	5.78	2.33
Zr/TiO2	67.21	57.14	70.00	63.64	39.39	63.64	37.04	87.50	75.00	6.82	36.96	150.00	18.67	49.65	49.65	50.47
Fe203/MiO	0.56	0.39	0.26	2.97	0.40	0.25	0.33	1.58	1.29	0.35	0.32	0.26	0.44	3.00	2.38	1.98

Sample description and locality - see also Fig. 3.1

GR3 (1) and (2)
SPIN A and E
SPIN B, C and D
GF and GC
G2 to G8 G9
BAS1 to 8 Bas10
ASB 2
K1 and K2
PLAG
Large felsic ocelli from extrusive lithologies

Metabasalts from the Eersteling area
Fibrous amphibole "spinifex-textured" rocks
Random amphibole "spinifex-textured" rocks
Fine and coarse grained melagabbro from Waterfall prospect
Melagabbros from Mt. Maria area
See pentinite from the Mt. Maria area
Carbonated metaintrusives from central Mt. Maria area close to Ulf-xenoliths
Pillow lava from north of Kuschie shear zones

Plate 3.2 Amphibole textures in metagabbros. All photographs XPL. Sample locations on Fig. 3.1.

- 3.2a. Ferrohornblende, with small 'uralitic' rims of ferro-actinolite marked by small fibrous overgrowths (arrowed). Note fracturing of main crystal (sample G3). Field of view - 6mm.**
- 3.2b. Actinolitic hornblende to actinolitic amphibole showing boudinage with boudins joined by fibres (Sample G8a). Field of view - 5mm.**
- 3.2c. Fibres of actinolitic-hornblende and actinolite amphibole growing into feldspar crystals (Sample G8). Field of view 5mm.**
- 3.2d and e. Extreme fibrous morphology of the most magnesian actinolites, indicating considerable metamorphic modification of original igneous textures. Field of view 5mm.**

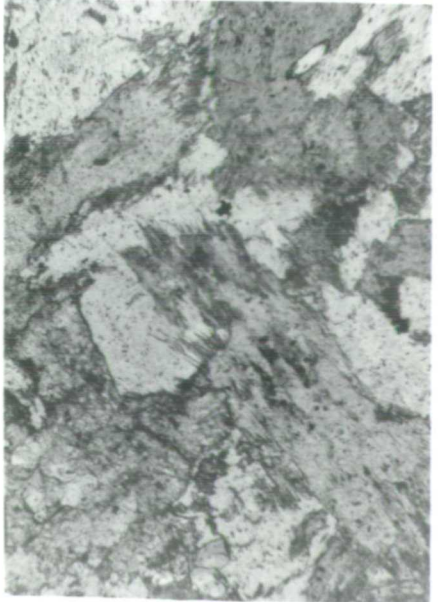
A



B



C



D



E



Plate 3.3 Oriented fibrous amphibole extension vein.
Sample Spin A, see Fig. 3.1 for location.

Thin section of "spinifex-textured" amphiboles showing preferred orientation and a kink-band structure. The fibres occur in texturally symmetrical units up to 75cm wide with the fibres oriented perpendicular to walls of the units. They are best explained as extension veins. The fibrous crystals occur in bundles with individual fibres commonly linking separate bundles giving the impression of two preferred orientation directions. This anastomosing habit occurs in and out of the plane of the thin section and this is responsible for the majority of the smaller amphibole crystals, which are therefore actually longer than observed. The fibres lie in a matrix of talc-chlorite and minor serpentine(?) and carbonate. The vein is cut by a later epidote-chlorite vein. PPL.

**Plate 3.4 Textural variation in fibrous amphiboles.
Sample Spin A.**

- 3.4a. A junction between the two apparent preferred orientations of amphiboles. It is evident that individual fibres diverge and merge, locally truncating fibres in the different bundles, to become integral parts of separate bundles. Field of view - 2.4cm. PPL.**
- 3.4b. Detail of one boudinaged crystal indicating significant extension along the crystals length and therefore sub-parallel to the walls of the unit containing them. Brittle deformation of other fibres is also present. Note how the crystals are joined by smaller and narrower asbestiform amphibole and/or chlorite (arrowed cf. plate 3.5). Field of view - 2.4cm. PPL.**
- 3.4c. Small fragment of amphibole crystal. It is not clear if this fragment is a crystal growing out of the plane of the thin section. The fragment comprises an aggregate of extremely narrow, curved asbestiform amphiboles suggesting that complex rotational components of deformation were operative during their growth. Field of view - 1.6cm. XP.**

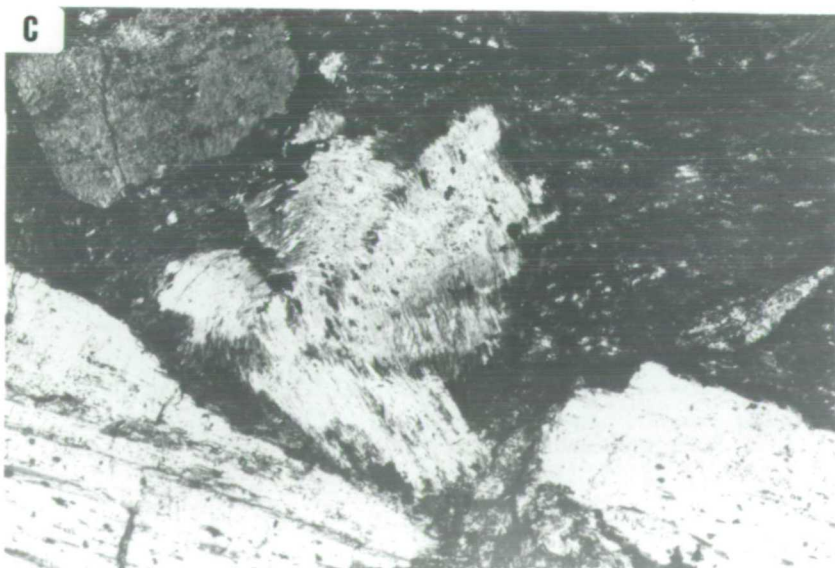
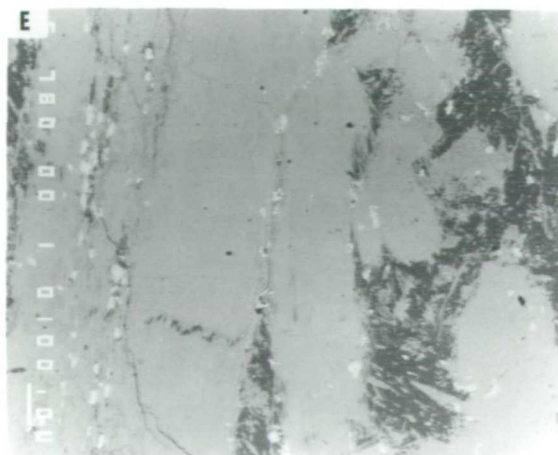
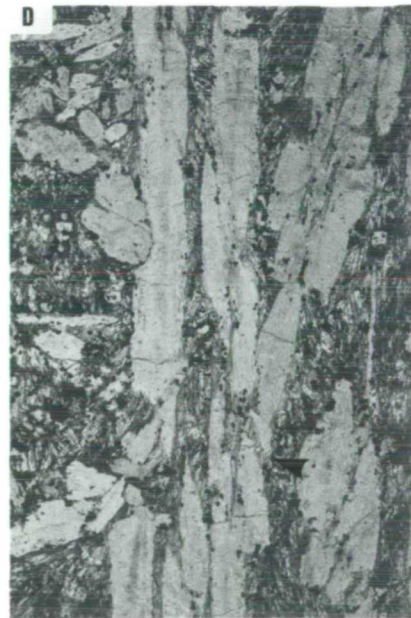
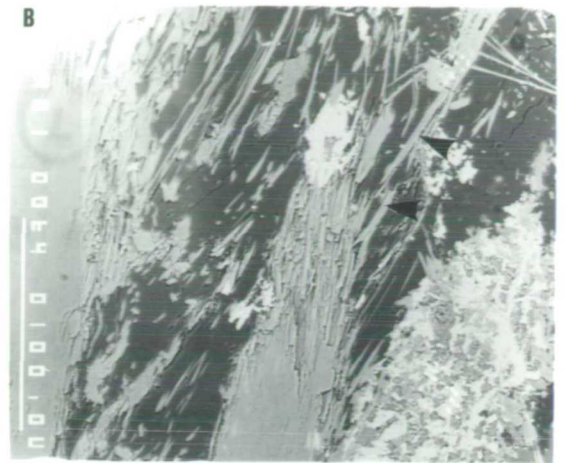
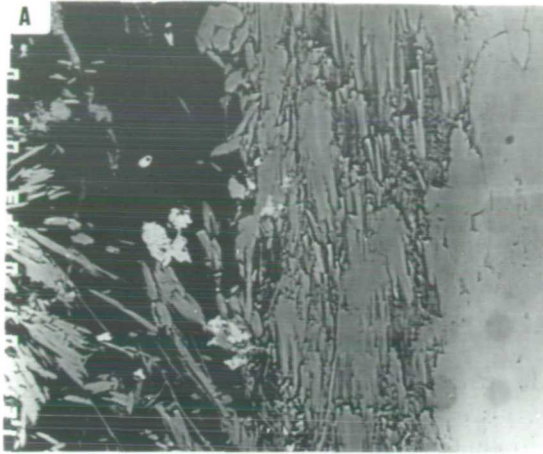
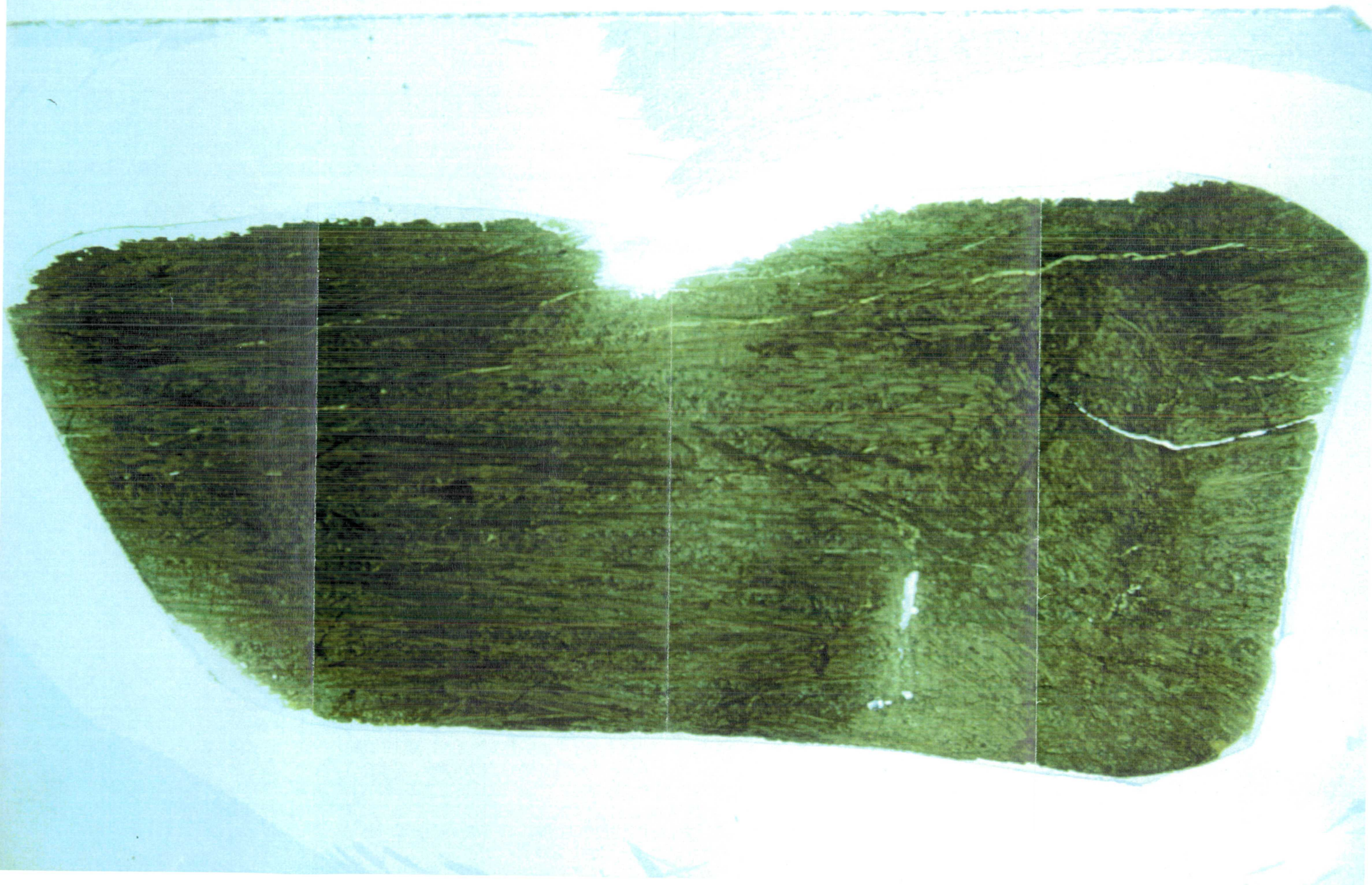


Plate 3.5 Evidence for retrogression of fibrous crystal ends and edges and also for their overall deformation-related growth. Sample Spin A.

- 3.5a. Edge of fibrous amphibole crystal showing retrogression to chlorite (patchy alteration) that has penetrated along the amphibole chains and aided/caused cataclasis of the fibre. SEM micrograph.**
- 3.5b. End of small amphibole boudin showing retrogression but also note the fibres with extreme aspect ratios in the matrix, some of which merge with the boudin end (arrowed). SEM micrograph.**
- 3.5c. Junction of diverging fibres. Note how the single crystal has apparently been ripped-off the left-hand main bundle and a fine fracture at the contact is seen (arrowed). The 'ripped-off' crystal cross-cuts a second bundle to the right suggesting that the fibres continued growing after the deformation. Field of view - 11mm. PPL.**
- 3.5d. A further example of a crystal junction. Note that the inclined crystal actually merges with the straight crystal and there is a boudin-fibrous overgrowth feature close to the junction (arrowed). The inclined crystal can not be rotated back in to parallelism with the straight crystal. Field of view - 9mm. PPL.**
- 3.5e. SEM micrograph of fibre junction showing fracture between the two differently oriented fibres. Note the fracture does not cut the entire crystal.**





**Plate 3.6 Matrix textures in fibrous spinifex-textured rocks.
Evidence for differential movement of fibre and matrix.
Sample Spin A.**

- 3.6a. SEM micrograph showing edge of a fibrous amphibole crystal with small chaotic zone oriented at varying angles to it, all lying in a matrix of talc-chlorite and possibly serpentine.**
- 3.6b. A zone shown in greater detail. It consists of a microbreccia of amphibole and some chlorite derived from the edge of the main crystal.**
- 3.6c. A further zone appears here to be attached to the main fibre and also folded, suggesting relative movement of crystal and matrix has occurred. However the limb of the fold in the matrix is a thin (<10um) fibre of amphibole that passes out of the photograph. Similar fibres (commonly with a preferred orientation between individual main fibres) can be seen to overprint the matrix as well.**



Plate 3.7 Textures in random amphibole spinifex bearing rocks.
Sample Spin D, Rietvley area (Fig. 3.1).

- 3.7a. Large overview of spinifex showing elongate crystals. Many of the smaller crystals are elongate out of the plane of the thin section. The section is extra thick and masks a fine grained matrix of dendritic sheaves of asbestiform amphibole, talc and chlorite. Field of view - 6.1cm. PPL.
- 3.7b. Lattice mismatch across a fracture similar to that seen in the fibrous amphibole spinifex-textured rocks. Note the bifurcation of one crystal into two across this fracture. Deformation textures are common in most of the amphiboles. Field of view 8mm. PPL.
- 3.7c. Stellate growth indicating nucleation of amphiboles from a single point and subsequent radial growth outwards. Field of view - 9mm. PPL.
- 3.7d. Complex truncation textures arising from the growth of amphiboles from different nucleation points and their subsequent interference. Field of view - 5mm PPL.



CHAPTER 4

THE GEOLOGY OF THE SNYMANSDRIFT SHEAR ZONE:

LITHOLOGIES AND CONTACTS BETWEEN LITHOLOGIES

4.1 INTRODUCTION

Field mapping in this project was concentrated on the Snymansdrift Shear Zone (SDSZ), a high strain zone that marks the southern, uppermost, boundary of the allochthonous unit of simatic rocks in the central Mt. Mare area. The zone is host to several important old gold workings (Fig. 1.3). Fig. 4.1 is a 1:5000 scale geological map of the SDSZ on which the lithologies and structural data have been differentiated.

The SDSZ developed during the second main deformation event so far recorded in the Mt. Mare area. This event is termed 'D2' and affected both the simatic (and related) rocks and the younger Uitkyk Formation sediments (chapter 2.3). The zone trends NE-SW, dips moderately to steeply (55-80°) SE, has a down-dip plunging lineation and can be traced for a distance of 6 km along strike from Snymansdrift No. 2 Hill in the southwest to the Greenwald (GW) area in the northeast (Fig. 4.1). The intensity of deformation increases along strike from the southwest, where the SDSZ dies out completely, towards the central area (Zandrivier), before decreasing again towards the northeast. Therefore *virtually the full lateral extent of the zone is exposed*.

Although the northern, lowermost boundary of the zone is well defined, its southern, upper boundary, is not. De Wit (1985) placed the upper boundary where a distinctive and unusual unit of tectonic breccia passed conformably into an overlying sequence of shales and sandstones. However, the sedimentary rocks themselves have been considerably deformed and are cut by narrower high strain zones. Therefore, for the purposes of this study, the upper boundary of the SDSZ has been extended out to include the entire sedimentary package, upto isolated outcrops of sheared breccio-conglomerate that lie immediately above the sedimentary units and which form the postulated extension of the Hollandsdrift Shear Zone (Fig. 4.1). Thus the SDSZ can be considered as a sequence of tectonites and deformed sediments bounded by two major strain discontinuities.

In this chapter field (and some microstructural) characteristics of the surface

lithologies associated with the zone and the contacts between them are described. This is followed in chapter 5 by a more detailed microstructural study of the operative deformation mechanisms and metamorphism present. In chapter 6, the geometrics of the overall structural geology (second order folds and faults) are discussed and the results of a vorticity analysis to determine the 'ratio' of pure and simple shear within the zone presented. The information from all three chapters is then integrated into a model for the structural evolution of the SDSZ. The fault rock terminology used in this chapter is summarised in Appendix 3.

The SDSZ consists of a basal chlorite-quartz-carbonate phyllonite layer, successively overlain by a tectono-sedimentary melange, tectonically interleaved packages of deformed chloritic mudstones, sandstones and simatic rocks and finally well-bedded sandstones and conglomerates of Lithofacies A of the Uitkyk Formation (section line B-B' Fig. 4.1). Contacts between the sedimentary lithologies are variably strained but have all been seen to be gradational at some point along strike. Thus the sequence is essentially conformable but has been locally structurally thickened. Early formed syn-sedimentary deformation structures have been strongly overprinted in most places by fabrics produced by continued D2 compression.

4.2 THE TECTONO-STRATIGRAPHY OF THE SDSZ

Fig. 4.1 illustrates the tectono-stratigraphy of the SDSZ. It will be shown that this final stratigraphy was established during progressive D2 deformation but first several important features must be highlighted.

1. The stratigraphy is variable along strike both in thickness and in composition. At both the NE and SE extremities of the SDSZ, basal fault contacts, marked by chlorite-quartz phyllonites, are overlain by deformed breccio-conglomerates, poorly bedded mudstones and finally lithofacies A of the Uitkyk Formation (Fig. 4.1). Structures visible in this sequence suggest it is conformable.

In the centre of the SDSZ the stratigraphy is thicker (Fig. 4.1) and the basal chlorite-quartz phyllonite layer is successively overlain by a tectono-sedimentary melange which passes conformably into a sequence of deformed chloritic mudstones and sandstones. These sediments are tectonically interleaved with variably deformed 'sheets' of simatic rocks. In places the mudstones/sandstones pass conformably into Lithofacies A of the Uitkyk Formation. It is critical to emphasize here that the *contacts* between the lithological units have been observed to be either sedimentary (in the case of sediment-sediment contacts) or

are marked by early D2 high strain zones interpreted as syn-sedimentary faults *at some point* along their strike; either at the edges or in areas of preserved low strain within the more deformed parts of the SDSZ. In many places however, these original relationships have been obscured by subsequent strain and fabric development and therefore *taken in isolation* certain sections across the SDSZ may not reveal the initial relationships.

2. Two further problems were encountered in the field mapping. First, field and microstructural observations indicate different lithologies were texturally and chemically homogenised during progressive D2 deformation and metamorphism. Thus in the central part of the SDSZ phyllonitic units have uncertain protoliths and could be the product of deformation and metamorphism of either simatic or sedimentary lithologies. Secondly, in some places weathering has resulted in most chlorite-rich lithologies becoming red-brown in colour and ferruginous, further compounding protolith recognition. In most cases however, along strike from such outcrops, features such as remnant clasts or remnant igneous textures were observed and thus the origin of the units could be determined. However further detailed surface mapping and reconciliation with drill core logging may show some units to have been misidentified in this study. This is unlikely to significantly change the overall interpretation of the SDSZ.

4.3 CHLORITE-QUARTZ-CARBONATE-TALC PHYLLONITES

This unit defines the base of the SDSZ and is 50 to 150m thick. It is derived from the progressive deformation of footwall simatic rocks (medium grained metagabbros) and an increase in strain is indicated by the development of a penetrative foliation. Only occasional remnant textures such as flattened ocelli and remnant lenses (2-3m) of amphibole spinifex textured rocks are preserved in the unit. Small lenses of highly deformed BIF are also present. The development of the foliation, which trends approx. NE-SW and dips moderately to steeply (55-80°) SE, indicates an increase in strain and suggests the basal contact of the SDSZ is tectonic along its entire length. The rocks are predominantly chloritic and hence they are best termed phyllonites (Appendix 3).

4.4 TECTONO-SEDIMENTARY MELANGE

4.4.1 Introduction

This unit is the main lithology in the SDSZ and is marked by a coarse, locally chaotic, chert breccia in a pervasively sheared matrix which is up to 150m thick. It was first defined as a tectono-sedimentary melange by de Wit (1985) who suggested it formed during

concurrent shearing and sedimentation early in D2.

The term "Melange" in the sense of a mixed fragmental rock was first introduced into the geological literature by Greenly (1917) and since then its definition has been a contentious issue (see Raymond (1984) and Rast and Horton (1989) for reviews). For the purpose of this study the definition of Silver and Beutner (1980) has been adopted which, although perhaps a rather broad definition, is descriptive and avoids the use of specific terms that have been argued over in the past. They define melange as;

"...a mappable (at 1:25000 or smaller scale), internally fragmented and mixed rock body containing a variety of blocks, commonly in a pervasively deformed matrix".

Rast and Horton (1989) use the term melange in a more restrictive sense for a *tectonically deformed* body of internally fragmented and mixed rock so as to distinguish it from an olistostrome which is used for a rock body in which both sedimentation and tectonic deformation were involved in the fragmentation and mixing. However the term olistostrome often has a genetic connotation assigned to it (implicitly or explicitly) of having formed by submarine slumping and/or sliding. There is no evidence for such processes in the SDSZ and therefore the preferred name for the unit studied is (with a genetic precursor to the descriptive noun) tectono-sedimentary melange.

One of the key issues in the study of melanges concerns the relative timing of stratal disruption of the sedimentary units/blocks and fabric development, especially where the disruption is inferred to have occurred whilst the sediments were still partially consolidated ('soft sediment deformation'). Bosworth (1989) has shown in a Taconic melange that on the basis of cross-cutting fabric relationships, fabric development and stratal disruption probably occurred well after lithification. However others have shown that at least some stratal disruption occurred whilst the sediments were clearly unconsolidated (see papers in Rast and Horton 1989). Generally such structures are quite chaotic and are not associated with much fabric development. However studies of sediments in both ancient and modern accretionary prisms (eg Moore 1986) have shown strong fabrics can develop during consolidation and dewatering, thus both soft sediment structures and fabric development can be synchronous at least in the early stages of deformation. It is important to emphasize that the role of crystal plasticity is minor in the fabric development, cataclasis and pressure solution predominate. Thus it is important to determine the processes involved in the fabric development, their overprinting relationships and spatial relationships of the various structures. In this chapter structures are described which indicate synchronous tectonism and sedimentation were occurring early in D2 and lead to widespread soft sediment stratal disruption textures. These structures however have been variably overprinted by continued fabric development as D2 progressed. The effects of this later deformation are emphasized

in chapter 5.

The distinctive feature about the SDSZ melange is that it is heterogeneous both along and across strike in terms of its thickness, clast size and shape, clast:matrix ratios and internal structures. A gross overall fining upwards and along strike away from the centre of the unit can be seen, but even then there are local coarser zones. The following descriptions are based on type section B-B' (Fig. 4.1, Fig. 4.2) although it must be emphasized that, because of the heterogeneity, the features seen are not found throughout the melange. The main departures from the sequence seen in the type section are also described. The term 'breccia' is used here in its simplest descriptive form as a sequence of angular fragments lying in a finer grained matrix and no genetic connotation is assigned to it.

4.4.2 Lowermost Melange

The contact between the underlying phyllonites and the melange is gradational, occasionally marked by a thin (<75cm wide) discontinuous layer of BIF. The phyllonites grade rapidly upwards (over 1 to 5m) into a spectacular 20-30m thick zone of breccia (Fig. 4.2). The clasts in the breccia are exclusively white chert and laminated white and brown to red chert, similar to the BIF units found in the simatic rocks, and also chert clasts seen in the Uitkyk Formation (chapter 2.5). The majority of the clasts are clearly rectangular to rhombohedral in shape (plate 4.1) but have extremely variable sizes, being up to 40cm long, 1 to 5cm wide but generally less than 1 to 2cm thick. Brittle break up of the clasts is ubiquitous. The matrix to the breccia is pervasively sheared and chloritic (phyllonitic), although locally it can be shaley in appearance. It carries the same tectonic fabric as the underlying phyllonites. It is a red to brown colour in most outcrops due to a significant percentage (generally 5 to 10%) of weathered iron-rich carbonate. The clast to matrix ratio in this part of the melange is high and it is predominantly clast supported. Small 1-2m lenses of chlorite-talc-carbonate phyllonites (schistose mafic-ultramafic material) which grade out into the melange zone are also present.

The breccia clasts show a clear preferred geometric orientation which defines a lineation that for the most part plunges down-dip of the matrix fabric (plate 4.1). Together, the fabric and the lineation define the strain ellipsoid (Ramsey and Huber 1983). Despite their elongate nature, there is no microstructural evidence for crystal plastic deformation in the clasts and thus their present shape is predominantly an original one, not greatly modified by plastic strains. Therefore, the preferred orientation of the clasts is not a stretching lineation but a simple shape or geometric orientation lineation.

There are areas of higher strain within the breccia that do show greater degrees of ductile strain. These contain clasts with small tails of fine grained quartz at their ends, often

preferentially developed at clast impingement points (plate 4.2a). They appear to be pressure solution phenomena. In places, there seem to be 'domains' up to 2 to 3 metres long by 1 metre wide that behaved in a ductile manner. In these domains longer clasts are slightly warped and all clasts show a greater degree of tail development (plate 4.2b). This ductile deformation is an overprint on the clasts in the breccia and appears to affect individual clasts at first, then larger domains which locally coalesce. Overall, brittle and diffusive mass transfer processes were the dominant deformation mechanisms in the lowermost melange and hence the majority of the breccias can be termed c-type mylonites (Appendix 3).

At locality A (Fig. 4.1) conjugate fibrous quartz-carbonate extension veins are present within which the fibres are straight and plunge parallel to the clast lineation. These veins contain inclusion trails of wall rock and therefore appear to have formed syntectonically with respect to D2 by the 'crackseal' mechanism of Ramsey (1980). The straightness of the fibres suggests the incremental strain ellipsoid did not significantly change orientation during D2 (Ramsey and Huber 1983).

4.4.3 Large BIF blocks in the melange

The lowermost melange is 'capped' by a disrupted BIF unit of variable thickness (Fig. 4.2). It consists of a series of large blocks lying at approximately the same horizon but totally within the breccia (Fig. 4.1). The largest blocks are up to 150m long but none are more than two metres thick at most (plate 4.2c and d). The BIF blocks are therefore large clasts in the melange zone and the lamination in them, as well as their overall shape, is aligned sub-parallel to the fabric in the matrix. These relationships are evident on Fig 4.1 and plate 4.2c and d.

Although the BIF blocks themselves are not folded, *internal*, or intrafolial, similar folding of their laminations is common (plate 4.3). Note that both the quartz and magnetite layers display ductile behaviour, the folds being characterised by thinning of the limbs and thickening of the hinges. Thin sections show that both the quartz and magnetite remain as fine grained (<10-20 microns) intergrowths in the relatively higher strained limbs and lower strained hinges of the folds (plate 4.5d). Small phyllosilicates mark many of the grain boundaries. There is no evidence of dynamic or static recrystallisation textures or any sign of preferred crystallographic orientation of the quartz. Such features suggest that grain boundary sliding probably accommodated by diffusion (DAGBS - Appendix 3) was the dominant deformation mechanism within the blocks.

The fold axes of all the folds seen are plunging sub parallel to the lineations in the surrounding breccia and their axial planes are parallel to the fabric in the melange matrix.

and the original laminations in the BIF (Fig. 4.1). When viewed down plunge, there is no apparent systematic vergence of the folds and S and Z folds seem to be of equal predominance. Within the blocks, M and W style similar folds are seen (plate 4.3a) as well as occasional brittle-ductile deformation styles such as tectonic imbrication of chert layers and irregular white chert breccias (both within a siderite with subordinate magnetite matrix – plate 4.3 b and c). In these cases the chert has clearly behaved in a more brittle fashion than it did in the folds.

At the edges of many of the larger BIF blocks remnant fold hinges can still be seen where the laminations are truncated by the matrix that surrounds the blocks (plate 4.3d). Some of these show in a most spectacular fashion that the clasts in the melange surrounding the BIF blocks are directly derived by the tectonic delamination and consequent milling of chert layers in brittle-ductile folds at the block edges. Plate 4.4a shows a remnant fold hinge, delineated by white chert boudins, that are breaking down into a predominantly sideritic matrix. Note that the boudins are not folded, rather the structures represent boudinaged folds. Plate 4.4c shows a more irregular breakdown and it is clear that the chert clasts are "spalling-off" from the blocks. The magnetite-bearing laminations in the BIF are being reworked in a ductile manner into the matrix of the melange during the fragmentation. This process appears to be concurrent with a chemical change of the magnetite to siderite. The fluids which caused the carbonation have also penetrated 2–10cm into the BIF units. The horizon of BIF blocks within the melange therefore marks a zone of significant cataclasis, stratal disruption and dilatancy and the breccias formed directly from the blocks are cataclasites. It is not clear if the brecciation was continuous or occurred in pulses.

Along the bases, and less frequently along the tops of the BIF blocks, simple boudinage of the external laminations is seen. Investigation of 3D outcrops reveals that the boudinage is predominantly irregular style, chocolate-tablet boudinage which produces a dessication crack-like appearance on bedding planes. The boudinage is responsible for producing many of the larger slabs or clasts of chert that were seen in the underlying breccia zone (plate 4.5a, b and c). This is consistent with the earlier interpretation that the clast size and shape in the lowermost melange are not due to elongation by plastic deformation. The break up of the BIF has not been aided by any pre-existing joints.

The stratal disruption of the BIF blocks results in a supply of cataclasites to the matrix that are initially oriented at varying angles to the matrix. If this process did produce all the clasts in the lowermost melange, it can be inferred that the c-type mylonites represent more deformed versions of the cataclasites. Therefore, once formed, the clasts undergo rigid body rotation through the matrix (ductile flow) during progressive deformation to give the preferred orientation observed in the c-type mylonites.

4.4.4 Uppermost melange

Immediately above the disrupted BIF layer a breccia, of similar style to that of the lowermost melange again dominates the tectonic-stratigraphy (Fig. 4.1) but there is a greater degree of *in situ* brittle deformation of the clasts within the matrix. This has resulted in significant grain size reduction. Towards the top of the unit, which is a maximum of 75m thick (Fig. 4.2), the clast to matrix ratio decreases considerably and the melange becomes increasingly matrix supported. The average size of clast decreases upwards and their degree of sphericity increases, although the angularity of most of them is still high (plate 4.6a and b). There is also a significant decrease in the degree of preferred orientation of the clasts although the matrix, which becomes distinctly shaley in appearance, still has a moderately developed fabric (plate 4.6a and b). The matrix also displays possible signs of deformation. Plate 4.6c and d show small 1-2m diameter 'eye' structures in the matrix (interpreted as sheath folds) as well as isoclinal folds, all of which are axial planar to D2. Similar structures have been reported by Higgins *et al.* (1989) in melanges from the Southern Appalachians. Locally there are good examples of grading in the unit, some of which may be turbiditic (plate 4.7a and b). These structures are best exposed at localities B, C and D (Fig 4.1). In places remnant fold hinges of BIF can be seen and these are also breaking down *in situ* (plate 4.6e).

A small percentage (<5%) of clasts other than white chert fragments are present in the upper melange, such as grey chert and deformed sandstone carrying a pre-existing fabric. Although the provenance of these clasts is outside the main melange zone, their precursor lithologies have been observed further up sequence and thus they are intraformational (chapter 4.9.1). The distinctive siderite nodules seen in the Uitkyk Formation are also present in this part of the melange (plate 4.6f, *cf.* plate 2.5e), and are up to 10cm by 5cm in size. There are also layers and lenses up to 5cm thick of preferential siderite development in the melange where the carbonate has strongly overprinted the matrix.

Overall, the uppermost parts of the melange are best classified as poorly sorted, granule-sized, sedimentary conglomerates with poorly developed bedding. At localities C, E and F (Fig. 4.1) the melange unit grades directly, *without recognisable tectonic break*, into the cross bedded sandstones and conglomerates of lithofacies A of the Uitkyk Formation (Table 2.2). The sediments young upward in the section, to the south and southeast. In the centre of the SDSZ the melange grades up into a thick sequence of chloritic mudstones as well as sandstones (eg. locality A, Fig. 4.1). These contacts are now cross-cut by the regional D2 fabric (plate 4.7c). Hence the upper contact of the melange is, in many places, *sedimentary*.

4.4.5 Discussion

The features described in this unit conform to Silver and Beutner's (1980) definition of melange. Furthermore, there are no recognisable tectonic breaks between the lowermost phyllonitic unit of the SDSZ (which is in tectonic contact with the underlying simatic rocks) and the well-bedded sandstones and conglomerates of the Uitkyk Formation. The melange therefore marks the gradation *from* purely tectonic *to* purely sedimentary processes and is best termed a tectono-sedimentary melange. This relationship is evident both along strike to the NE and SW of the centre of the melange as well as across strike (Fig. 4.1).

Most of the small chert clasts seen in the melange are boudins formed by the *progressive* tectonic delamination of the large BIF blocks. Because the axial planes and plunges of the parent folds *within* the BIF blocks are parallel to the foliation and clast orientation respectively in the surrounding matrix, the disruption, and melange development as a whole was synkinematic with respect to D2. Both the folding and other internal structures are best explained as the result of layer parallel shearing of the BIF blocks *within* the melange. This resulted in a decrease in the size of the blocks and their virtual destruction, due to a progressive "sheathing-off" of their external laminations (Fig. 4.3a). The irregular chocolate-tablet boudinage structures associated with this sheathing-off suggest the strain path was, at least locally, not one of plane strain but of flattening strain. Therefore the regular rectangular to rhombohedral shape of most of the clasts in the breccias is strongly controlled by a combination of the original thickness of quartz/magnetite laminations in the BIF units as well as the strain history of the rocks.

A key factor in the development of the boudins is the change in style of deformation of the chert layers from ductile behaviour in the centres of the BIF blocks to being brittle at the edges. The initially fine grain size of the quartz and magnetite, coupled with the presence of fine grain-boundary phyllosilicates is interpreted to have been an important control on the ductile behaviour and probably allowed DAGBS to operate early in the deformation history without the need for previous grain size reduction. Therefore, it would seem that the deformation mechanism in the chert layers changed from being a grain size sensitive process in the block centres (if DAGBS was operative), to a pressure sensitive process (cataclasis) at the edges. This implies there was a *relative* local increase in strain rate associated with the cataclasis.

The cause of this change in operative deformation mechanism is difficult to determine but it is probably related to one or a combination of the following factors;

- 1) In certain progressive deformation histories early formed folds may be subsequently extended as they rotate, the so called 'field of boudinaged folds' (Fig. 4.3b - Ramsey and Huber 1983, p. 224).

2) In order to maintain strain compatibility in zones of shearing, the edges of rigid bodies within such zones must experience an increase in shear strain rate, the so called 'stress riser effect' (Sibson 1980).

3) The magnetite layers in the centres of the blocks BIF continue to behave in a ductile fashion at the edges and are reworked into the matrix of the melange. This reworking was accompanied by a chemical change of the magnetite to siderite. Such a transformation is unlikely to have been isochoric and a volume increase probably occurred. This would have caused the stresses in the iron-rich layer to rapidly increase, so aiding the fracturing of the enclosing chert layers (Fig. 4.3c). If the volume increase was an important factor in the delamination process then the associated metamorphic transformation may have been an important strain softening process in the development of the melange.

The field and textural relationships of the *actual* brecciation of the BIF, and shown specifically in plate 4.4a and b, are suggestive of fluidization phenomena (Reynolds 1954). The slurry-like appearance of the breccias (plate 4.4b) is suggestive of entrainment, whilst abrasion and cataclasis of the constituent particles is ubiquitous; the overall fining upwards of the clasts and direct passage into alluvial sediments indicate elutriation was probably a major process in the formation of the uppermost melange. Overall these features are considered to represent the "expanded bed" stage of a fluidized system (Reynolds 1954). Murphy (1984) has described fluidized breccias from Ireland in which cleavage development and folding was synchronous, or cyclically alternated, with brecciation.

The abundant carbonation associated with the fluidization points to the influx of an aqueous fluid phase, saturated in dissolved CO₂ gases, into the melange. Such a fluid is likely to have migrated from depth (metamorphic or magmatic origin) and hence the melange represents a zone of significant fluid release into a near surface environment. Cooling of the fluid during its ascent, coupled with chemical interaction with magnetite in the BIF blocks, would have caused phase separation leading to volume expansion under high fluid pressures, concomitant brecciation and fluidization as well as mineral precipitation. The carbonate-bearing fluids would also have passed upwards into the overlying sediments, mixing with meteoric waters. This is thought to be responsible for the pre-cleavage siderite infiltration seen in the Utkyk Formation (chapter 2.5, plate 2.5d).

Although there is a great deal of internal deformation within the melange, contacts between lithologies were preserved as essentially planar to curvilinear features. This suggests that overall, the deformation in the melange was controlled by its boundary conditions and thus it may have behaved as a large scale ductile shear zone (Ramsey and Huber 1983). Because the lateral terminations of the zone, especially in the west, have been identified (chapter 4.1 Fig 4.3) then the alignment of the gradation into the surrounding

reflects a decrease in deformation and displacement such as is seen towards fault tips (Coward and Kim 1981, Ramsey and Huber 1987). The sense of shear in this zone is not immediately clear in the field but is discussed in chapter 6.

The melange is marked by a lithological sequence which passes from the footwall upwards through phyllonites, c-type mylonites, cataclasites and sediments (Fig. 4.2). An important question concerning this distribution is; does it reflect either

- 1) the tectonic juxtaposition of rocks from different crustal levels, or
- 2) a 'snapshot' of an actual spatial distribution during deformation (Appendix 3).

This question is tackled here in two parts:

First, the overall gross fining upwards of the clasts from the BIF blocks to the overlying sediments indicates that the layer parallel shear and associated cataclasis of the blocks was concomitant with sedimentation. Clasts were being actively derived both from the blocks at the base of a sedimentary pile and added to the top by alluvial processes (chapter 2.5). This is best explained if sedimentation was occurring *above* a propagating fault plane.

Secondly, below the BIF blocks a progressive sequence of fault rock development of cataclasite-c-type mylonite was inferred. During *progressive* layer parallel shearing of the BIF blocks it would be expected that early formed cataclasites would be deformed to form c-type mylonites *whilst* brecciation around the edges of the remaining blocks was still occurring. However, this does imply the c-type mylonites formed synchronously (at the base of a sedimentary pile) with sedimentation (at the top of the pile). The stratigraphic distance over which the transition occurs may be unrealistically small (Fig. 4.2) and therefore much of the fabric in the lowermost melange may be a later D2 overprint. The sedimentary contacts at the top of the melange are cross cut by the D2 fabric and attest to post-lithification flattening (*sensu lato*). However, synchronous c-type mylonite formation and sedimentation can not be ruled out on the basis of the field evidence. A qualitative investigation of the degree of strain seen in overlying lithologies may help estimate the degree of post-lithification deformation.

The matrix to the lowermost melange rocks is phyllonitic. It is derived from three main sources; 1) chlorite from the simatic rocks, 2) fine grained brittle and ductile deformed quartz from the original clasts, and 3) siderite derived from reworking of magnetite laminae in BIF units and associated introduction of carbonate fluids. In the upper melange (where sedimentary processes) have an influence, although detritus was probably being introduced from outside the main zone of BIF disruption, a significant part of the matrix may have been derived from similar phyllonitic material. Under continued hydration, carbonation

and mechanical reworking, the phyllonites would break down to form sand, silt and clay grade material. This could be reworked by near surface processes and progressively mixed with true sedimentary detritus.

In summary, it is inferred that the melange probably represents an original BIF layer or layers intercollated with simatic rocks that was subjected to shearing, initially in a near surface environment. The unique style of the tectonic disruption of the BIF, and the type of fault rock produced, was strongly controlled by its original lithological properties, namely the fine grain size and thickness of the quartz/magnetite laminations, as well as its strain history. The structures are best explained by the propagation of a syn-sedimentary fault at the base of a sedimentary pile.

4.5 STRATALLY-DISRUPTED, CHLORITOID-BEARING, CHLORITIC MUDSTONES AND SANDSTONES

Description: In the central parts of the SDSZ the melange is overlain by a series of predominantly very cleaved/schistose chloritic mudstones (shales and siltstones) and sandstones that have weathered to a red to brown colour in most outcrops. The weathering is most pronounced in parts of the unit that are pervasively carbonated. The sediments contain isolated clasts and small breccio-conglomerate lenses of angular chert and deformed sandstone that are demonstrably intraformational. Chloritoid porphyroblasts are locally well developed.

The lithology is up to 300m thick but has been intensely deformed as evidenced by the fabric development and ptymatically folded quartz veins (axial planar to D2 fabric). This deformation, the lithology's generally fine grain size and weathering, make it difficult to determine the original thickness of the lithology and also to distinguish between primary sedimentary lamination and tectonic fabric. In many outcrops two fabrics can be seen, differing by up to 25° in orientation, but it is generally not clear if they represent bedding and cleavage, an anastomosing tectonic fabric, or two separate tectonic fabrics. In other outcrops one or other of these fabrics is better developed. In order to overcome this problem of fabric origin, the fabric that was sub parallel to the regional fabric in other lithologies was recorded as the tectonic fabric whereas the second fabric was recorded as a 'secondary fabric' (Fig. 4.1). Where later deformation is particularly intense it is difficult to distinguish this lithology from possible phyllonites derived from the simatic rocks.

At locality G (Fig. 4.1) good exposures of the lithology in an area of qualitatively we s al de a e t the t o fabric pre nt a e the p duct of a a m s' g

dark green mudstone matrix that encloses elongate ellipsoidal bodies of pale green sandstone. The ellipsoids plunge steeply to the southeast, subparallel to the clast lineation seen in the melange. Plate 4.8a and b show, respectively, sections normal to and parallel to the long axes of the ellipsoids. They show a sandstone–mudstone interface and indicate that the ellipsoids are sandstone ‘blebs’ which have been progressively separated from their parent body to lie in an anastomosing matrix of the dark green chloritic mudstone. The boundaries between the lithologies are sharp to diffuse and locally there appears to have been complete ‘mixing’ of the two (plate 4.8a). In thin section, individual grains at the edges of the sandstone bodies become incorporated into the mudstone matrix and locally this is accompanied by grain breakage of the clasts (plate 4.8c). These textures are overprinted by pressure solution seams (plate 4.8a).

Along strike towards the centre of the SDSZ from locality G, fabric intensity increases and most original textures become unrecognizable. Although the metamorphism of this lithology is discussed in detail in chapter 5, as a first order field observation it appears that as fabric intensity within the unit increases, so does the degree of chloritoid development. At locality H (Fig. 4.1), the chloritoid may make up over 70% of the lithology and completely obliterates earlier textures. This is an unusual rock type and is the ‘ottrelite schist’ of Hall (1909) and Willemse (1938).

Interpretation: The textures at locality G are interpreted as evidence for stratal disruption of partially consolidated sandstone–mudstone couplets during deformation of the sedimentary package. The parallelism between the lineation defined by the ellipsoids and also the clasts in the melange suggest the deformation was an early D2 feature.

Similar ‘soft sediment’ deformation structures have been reported from Phanerozoic and Recent melange terrains and accretionary prisms by Cowan (1985), Lucas and Moore (1986), Knipe (1986), Lundberg and Moore (1986) and Agar (1988). The variables controlling the operative deformation mechanisms during such deformation have been shown by these writers to be complexly interrelated and include such factors as porosity, permeability, compaction rates and strain rates. Such inter–relation causes deformation conditions to change rapidly during the burial and tectonic history of the sediments. Much of the deformation is usually inferred to have occurred under conditions of high pore fluid pressure (see Moore 1986 for overview). The textures observed in this study are best explained as follows (taken partly from Lucas and Moore (1986).

Undeformed, weakly consolidated sandstone–mudstone units were subjected to layer parallel shear. The progressive tectonic mixing of the lithologies probably proceeded by non-destructive particulate (cataclastic) flow aided by excess pore fluid pressure and viscosity contrasts between the sandstone and mudstones. These processes would result in

initial "spalling-off" of sandstone blebs into the mudstones. Continued shearing and burial of the sediments would cause dewatering and hence an increase in the effective confining stress. This in turn would lead to incipient cataclasis, grain size reduction and strain hardening, causing the localisation of deformation along discrete shear zones. These zones propagated along original sandstone-mudstone interfaces so continuing to dismember the sandstone bodies. Hence, the dark green matrix represents an anastomosing cataclastic shear zone array in which material is made up of deformed sandstone grains and original mudstone. Such 'lithological mixing' has therefore resulted in the *transposition* of primary bedding in much of this lithology (plate 4.5d) and indicates the two fabrics seen represent an anastomosing D2 tectonic fabric.

This textural interpretation is consistent with the interpretation of the underlying tectono-sedimentary melange as indicating concurrent deformation and sedimentation early in D2. However, in the centre of the SDSZ *these sediments have been considerably affected by post-lithification D2 deformation and metamorphism.*

4.6 WELL BEDDED SANDSTONES AND CONGLOMERATES; LITHOFACIES A OF THE UITKYK FORMATION

Description: At localities C, E, and F (Fig. 4.1) the melange grades, *without recognisable tectonic break*, into Lithofacies A of the Uitkyk Formation (Table 2.2). Elsewhere within the SDSZ the chloritic mudstones also grade, at least locally, into Lithofacies A. This lithofacies is traceable along the entire length of the SDSZ and consistently youngs upwards towards the southeast. As such the sediments appear to 'drape' over and around the central and more deformed parts of it. At SD # 3 Hill at the southwestern extremity of the SDSZ, the melange and chloritic mudstones thin dramatically and grade along strike into a thin (1-5m wide), chloritic mudstone unit that lies between two sandstone layers (Fig. 4.1). At locality L (Fig. 4.1) these layers exhibit some important structures.

Plate 4.9a shows the hinge zone of a small syncline at locality L which is axial planar to D2 and plunges at 26° to the SW. Several generations of fibrous extension veins and a small bedding-parallel slickenside layer can be seen in this area, within which the fibres plunge sub-parallel to the fold axis. In the upper right of the photograph is an unusual 'flame-like' structure shown in more detail in plate 4.9b. The structure is macroscopically ductile and appears to be the result of the 'injection' of the mudstone into the sandstone approximately along the axial plane of the fold. The 'flame' appears to truncate earlier fibrous quartz veins.

The mudstone also penetrates at least several metres into the sandstone along discrete thin (mm scale) zones that isolate ellipsoidal bodies of sandstone to define a mesoscopic D2 fabric (plate 4.9c). At the actual contact of the two lithologies, small 'blebs' of sandstone can be seen to be 'spalling-off' into the mudstone and mixing of the two lithologies has occurred. Plate 4.9d shows a larger sandstone body lying in the mudstone unit at the same locality. This body is still just attached to the underlying sandstone unit from which it appears to have been 'ripped-off'. The edges of this body also show the 'spalling-off' texture. These textures are cross cut by small faults and quartz veins. D2 cleavage is not well developed at this locality but is evident elsewhere in the SD # 3 area (Fig. 4.1).

In the hinge zone of the flame structure, the mudstone contains small clasts of white chert. Throughout the mudstone layer at SD # 3 Hill, there are larger clasts of white chert up to 50cm³ in size which were undergoing *in situ* distributive cataclasis (plate 4.9e and f). The cataclasis is associated with the penetration of the mudstone matrix along the fractures in the clasts. The resultant smaller particles were subsequently strung out and underwent ductile flow in the mudstone. It is this process that supplied the smaller clasts seen in the fold hinge.

Interpretation: The flame structure, lithological mixing of sandstone and mudstone, and the disaggregation of large chert clasts are all interpreted as evidence for deformation and stratal disruption of partially consolidated sediments that occurred early in D2. The cataclasis of chert clasts and associated penetration of mudstone into fractures is suggestive of hydraulic fracturing and infers the deformation occurred under conditions of high fluid pressure in the mudstone unit. Similar textures and interpretations have been reported in melanges from New England, USA by Bailey *et al.* (1989). High fluid pressures would aid the injection of mudstone into the sandstone during folding of the 'soft' sediments, so producing the flame structure. The presence of fibrous veins also attests to (at least transitorily) high fluid pressures during part of the deformation. The soft sediment deformation was generally distributive and was probably the result of particulate (cataclastic) flow. It was overprinted by localized cataclasis producing the later cross-cutting faults, and also by later cleavage development.

4.7 OTHER LITHOLOGIES

4.7.1 Simatic rocks

There are several elliptical lenses of chlorite-talc schist found within the SDSZ (Fig. 4.1) as well as tabular sheets of relatively undeformed simatic rock (with schistose

boundaries) intercalated with the chloritic mudstone-sandstone sequence. The lenses are a blue-green colour in fresh exposures but are often weathered to a brown colour due to pervasive carbonation. The lenses are all very highly deformed and are either bounded by, or found within, narrow (2-5m wide) D2 high strain zones within the main SDSZ. At locality M, (Fig. 4.1) within one such high strain zone, the chlorite-talc schist is strongly crenulated and the crenulations define a rodding lineation. At locality N deformed elongate ocelli were seen within one of the lenses and which plunge down dip parallel to the D2 lineation elsewhere in the SDSZ. Strain analysis by de Wit (1985) on the ocelli indicated minimum D2 strains of 111% extension, 31% extension and 63% shortening in the X,Y and Z planes respectively of the finite strain ellipsoid. This corresponds to an ellipsoid ratio of 5.8:3.6:1. Thin section work (this study) indicates there is little or no variation in grain size between undeformed ocelli (which were initially fine grained; 20-50 microns - plate 3.1a) and deformed ocelli. DAGBS is inferred to be the dominant deformation mechanism.

The tabular sheets of simatic rock are difficult to identify at surface (chapter 4.2) but at least one major one is present and is clearly visible in drill core from the Zandrivier prospect (DDH ZR1 (see Fig. 5.1), ZR2 and ZR3). Original textures are well preserved although at the edges of the unit (and locally within it) fabric development obliterates such features. The contacts between these simatic rocks and surrounding sediments are commonly marked by variously coloured chert bars and/or narrow zones of chert breccia, both interpreted as high strain zones (see chapter 4.8 below).

The lenses and sheets are interpreted as remnants of simatic rocks that are probably ellipsoidal in 3D. They could represent either large blocks of simatic rocks caught up in the original sedimentary package and which were subsequently strained, or, fault-bounded sheets or horses that were tectonically introduced into the zone during, preferentially along pre-existing lithological contacts, during later deformation. Their association with high strain zones (chapter 4.8.2) and the extreme deformation seen in them favours the latter interpretation.

4.7.2 Late basic intrusives

At localities O and P (Fig. 4.1) two small exposures of basic intrusives are seen. They were not mappable at 1:5000 scale. These are late stage pyroxenites that intruded either as sills sub-parallel to the fabric or as more irregular bodies that cross-cut the foliation and are essentially undeformed. An important feature of these units is that they have been retrogressed and uraltised rims of actinolite around augite phenocrysts are ubiquitous.

4.8 CONTACTS BETWEEN LITHOLOGIES

Apart from the basal phyllonite contact with the underlying simatic rocks, it has been emphasised in previous sections that the contacts between all the major lithologies above the tectono-sedimentary melange, were observed to be gradational and sedimentary *at some point* along their strike length. Contacts between lithologies are in fact variable along strike and three different types have been identified: sedimentary contacts, late tectonic contacts and early syn-sedimentary fault contacts. These different types have been distinguished in Fig. 4.1 and a gradation between them was observed in the field.

4.8.1 Sedimentary contacts

These contacts have already been described above. They are all cross-cut at a low angle by the D2 cleavage. The lithologies were probably first affected by D2 whilst not fully lithified (chapter 4.6 and 4.7) but continued to be progressively deformed resulting in D2 cleavage development.

4.8.2 Tectonic contacts

Post-lithification deformation results in an increasing tectonic overprint on the original sedimentary contacts, most of which can be traced along strike into tectonic contacts. A good example of this gradation is seen between the contact of the melange and the overlying chloritic mudstones. At locality A (Fig. 4.1) the contact is sedimentary and cut by the D2 cleavage (plate 4.7c). Along strike to the NE, the original contact becomes transposed into parallelism with the tectonic fabric and is marked by phyllosilicate-rich, c-type mylonites and locally mylonites (plate 4.10a). These lithologies contain a well developed lineation formed by the pressure solution and possibly dynamic recrystallisation of clasts in the sediments. At locality Q (Fig. 4.1) the contact sharply truncates a one metre wide quartz vein in the footwall and is therefore tectonic. The contact now represents a high strain zone that nucleated along a pre-existing major lithological contact. Tectonised sedimentary contacts are best developed in the central SDSZ.

Locally the high strain zone between localities A and Q becomes very 'cherty' in appearance and passes into variably coloured chert layers. Similar cherts are common along the high strain zones identified within the main SDSZ and they have been distinguished in Fig. 4.1. The layers, or bars, are between 50cm and 8m thick, although 1-2m is the normal thickness, and consist of massive grey, green and white chert, laminated grey and white 'zebra' cherts (plate 4.10b) and laminated green and white cherts. Gradations between all the different types were seen. The bars also surround *all* the isolated lenses of highly deformed simatic rock seen in the melange (chapter 4.7).

Although it is difficult to prove movement along many of the chert bars they do contain textures that suggest they all represent silicified tectonites such as silicified breccias (plate 4.10c) and remnant rotated clasts (plate 4.10d). Different coloured cherts appear to represent different silicification episodes (eg plate 4.10e) and at least three distinct, cross-cutting silicification events (represented by the white, grey and green cherts) have been counted. Thus the cherts also record a complex polyphase history of fluid flow, which was focussed along these high strain zones. Some of the breccias seen in the chert bars contain well rounded clasts and resemble 'pseudo-conglomerates' (plate 4.10f) reported by Paris *et al.* (1985) who inferred they are formed under high fluid pressures.

The cherts are cut by at least two distinct generations of quartz veining. Some of these veins are ptlygmatically folded axial planar to D2 whereas others are undeformed fibrous extension veins. The fibres in these latter veins, which have well defined median lines and inclusion trails, plunge more or less down dip, parallel to the lineation seen in the melange (plate 4.10a). This suggests that the stretching direction (X) of the strain ellipsoid did not vary significantly during D2 and that straight dip slip movement along the contacts predominated (Ramsey and Huber 1983). Late tectonic stylolites sub-parallel to the D2 fabric cross-cut most of the features seen in the cherts (plate 4.10f). Fragments of the cherts and recrystallised sediments are found as clasts in the upper melange (section 4.4.4).

Overall, the deformation seen along such contacts indicates considerable late D2 deformation has occurred within the SDSZ but that much of it was localised along pre-existing sedimentary contacts, thus preserving original sedimentary structures and soft sediment deformation textures within the main parts of the lithologies and domains of lower strain along those contacts. The deformation was concentrated in the centre of the SDSZ and the overprint dies out laterally, especially to the west towards SD # 3 Hill. It is thought that the variation in strain seen along strike represents a variation in the amount of movement along the contact.

4.8.3 Syn sedimentary faults

Description: In all cases, the chert bars seen in the SDSZ (interpreted as tectonic contacts) become tectonically disrupted and brecciated themselves along strike to form linear breccia ones generally less than 1.5m wide. These are also high strain zones often characterised by the strong preferred alignment of the clasts in the breccia and are best exposed on the Zandrivier prospect, SD # 4 and 5 Hills and SD # 2 Hill (localities R, S and T respectively - Fig. 4.1).

The breccias are characterised by very heterogeneous clast shapes and sizes. The clasts are predominantly angular to subangular and vary in shape from very fine grained

angular fragments to tabular clasts, up to 50cm x 10cm x 10cm in size, and more equidimensional clasts up to 30cm in radius. Larger blocks up to 1.5 m³ are also seen. Plate 4.11a shows the breccia zone from locality R (Fig. 4.1). The breccia lies at the contact between underlying mudstones and overlying sandstones, and sub-parallel to bedding in the latter. Plate 4.11b is a close up of 4.11a and shows the penetration of a sandstone veinlet along a fracture in one of the larger clasts. Disaggregation of other chert clasts into the surrounding sediment is also seen and is similar to that described in section 4.6. At locality S (Fig. 4.1) draping of the overlying sandstone over the clasts and penetration of sandstone apophyses into the zone are common. Stratal disruption textures, including the complete disaggregation of sandstone (plate 4.11c), are also seen. At this same locality, bedding in the overlying sandstone is being systematically rotated into the breccia zone which as a result is transgressive to bedding at a low angle (10–20°). The breccia forms the northern limb of a D2 syncline. Along strike to the southwest, towards SD #3 Hill, the breccias die out and this same contact becomes purely sedimentary (locality E, Fig. 4.1).

The chert breccias have also locally undergone varying degrees of subsequent ductile deformation. The overprint is again strongest in the centre of the SDSZ (and weakest at the southwestern, along-strike termination of the zone) and has led to the development of foliated cataclasites (plate 4.11d and e) and (subsequently) c-type mylonites (plate 4.11f). The strain ellipsoid defined by the resultant clast alignment and tectonic fabric is parallel to that seen elsewhere in the SDSZ and is also related to D2. Strong pressure solution cleavages, often picked out by later tourmaline and/or graphitic alteration, are well developed within one metre of such zones (plate 4.12a). All but one of the old gold workings in the SDSZ are located along these breccia zones.

Interpretation: The features described above indicate the breccia zones are fault horizons that have had a prolonged deformation history. The sandstone injection structure and evidence for stratal disruption suggest that a large degree of cataclasis of the chert in the zones occurred whilst the surrounding sediments were partially consolidated and therefore probably occurred under conditions of excess pore fluid pressure. The breccia zones are interpreted as early D2 syn-sedimentary growth faults that initially propagated along lithological contacts in partially consolidated sediments. Because the breccias can locally be traced into sedimentary contacts it appears that displacement (but probably not direction of movement) varies along the fault zone. In the centre of the SDSZ, continued D2 tectonism lead to the progressive ductile overprinting of these breccias to form more mylonitic rocks.

Because the chert bars from which these breccias are derived are themselves D2 tectonites (chapter 4.8.2), then the breccia zones represent faults that have had more than one phase of movement along them. Furthermore, as some of the chert bars contain silicified breccias it follows that there were at least two phases of cataclasis in the fault

zones. Therefore the deformation could have been of a 'stick-slip' nature and related to earthquake activity (Sibson 1986, 1987).

4.8.4 Discussion

The gradation between sedimentary, tectonic and syn-sedimentary contacts can be explained in terms of a cyclical model of a propagating, stick-slip growth fault during progressive D2 deformation (Fig. 4.4). Early sedimentary contacts progressively become tectonically overprinted to become c-type mylonites which locally become silicified to form chert bars. Some early cataclasis also occurred. As the fault zone propagated, further cataclasis caused brecciation of the chert in unlithified sediment along strike (and probably up dip in three dimensions). This interpretation implies that at least some of the ductile deformation seen in the SDSZ did overlap in time with soft-sediment deformation *on the scale of the shear zone*. The ductile deformation, which is concentrated in the centre of the shear zone, does overprint earlier soft sediment structures but also passes along strike, over a distance of 3 km, into soft sediment deformation at its edges. Hence the current lithological distribution could represent close to a 'snapshot' of a developing shear zone (chapter 4.4.5).

4.9 SUMMARY

Observations from surface exposures in the SDSZ have lead to the following conclusions;

1). All structures within the high strain zone are related to the D2 deformation event but the strain distribution within it is very heterogeneous. A complex progressive history of early D2 syn-sedimentary deformation being progressively overprinted by later D2 fabric development has been revealed.

2). The lowermost contact of the zone is well defined and is a phyllonite (tectonite) whereas the upper contact is less well defined and sedimentary. Therefore, the lithological sequence of phyllonite melange-chloritic mudstones stratally disrupted mudstones and sandstones-lithofacies A of the Uitikyk Formation, is essentially *conformable*.

3). The gradation from tectonic to sedimentary processes is marked by a melange zone produced by the progressive tectonic disruption of BIF units whose original lithological properties BIF had a strong control on the development of tectonites and overall deformation style. A progressive fault rock development sequence of cataclasite-c-type mylonite

4). The sediments which overlie the melange contain structures which suggest considerable early D2 deformation occurred whilst the sediments were partially consolidated. However, there is a variably developed and locally intense late D2 overprint of ductile deformation on earlier formed D2 structures but it is concentrated in the centre of the SDSZ. However, the soft-sediment brecciation of the chert-bar tectonites suggests that at least some ductile deformation was occurring in the centre of the SDSZ mid- to late D2 whilst soft sedimentary processes were occurring at its edges as the whole zone developed and propagated in three dimensions.

5). The original contacts between the sedimentary lithologies in the SDSZ strongly controlled the propagation of later faults as deformation of the sequence continued. Actual decollement along some contacts seems to have occurred. A complete gradation between sedimentary contacts, tectonic contacts and syn-sedimentary faults is seen. This gradation can be explained in terms of a model of a propagating, stick slip growth fault along which displacement varies. Tectonic introduction of lenses of simatic rock into the sequence along the tectonic contacts may have also occurred.

6). The SDSZ is a zone of preferential fluid flow and deformation generally proceeded under conditions of high fluid pressures. Pervasive carbonation within the melange, silicification to form chert bars and local tourmalinisation all suggest considerable quantities of fluid have passed through the zone.

7). Although the timing of tectonism with respect to metamorphism is not clear in the field, it appears later D2 deformation may have been concurrent with prograde metamorphism of the sediments and was concentrated in the centre of the zone. Because the majority of the sediments in the Uitkyk Formation do not contain any porphyroblasts (chapter 2.5), the sediments in the SDSZ are at a higher metamorphic grade than those outside it. Therefore it would seem that the SDSZ is a zone of preferential heat flow.

The gradation from purely tectonic to purely sedimentary processes seen in the tectono-sedimentary melange is best explained as the result of a fault zone which early in D2 propagated along the base of a sedimentary pile that was being thickened by active sedimentation at the top of the pile. This implies that the sediments and the original sedimentation were at least in part allochthonous or semi-allochthonous (Ricci Lucci 1985). The SDSZ may therefore be analogous to a deformed piggy-back basin (Ori and Friend 1984). A corollary of this interpretation is that it is likely the sediments, and hence SDSZ as a whole, were at a shallower dip early in D2 than is seen now. They were therefore *significantly rotated, steepened and compressed as D2 progressed*. It is now necessary to investigate the effects of the later D2 flattening overprint and metamorphism on the lithologies in the more deformed parts of the SDSZ.

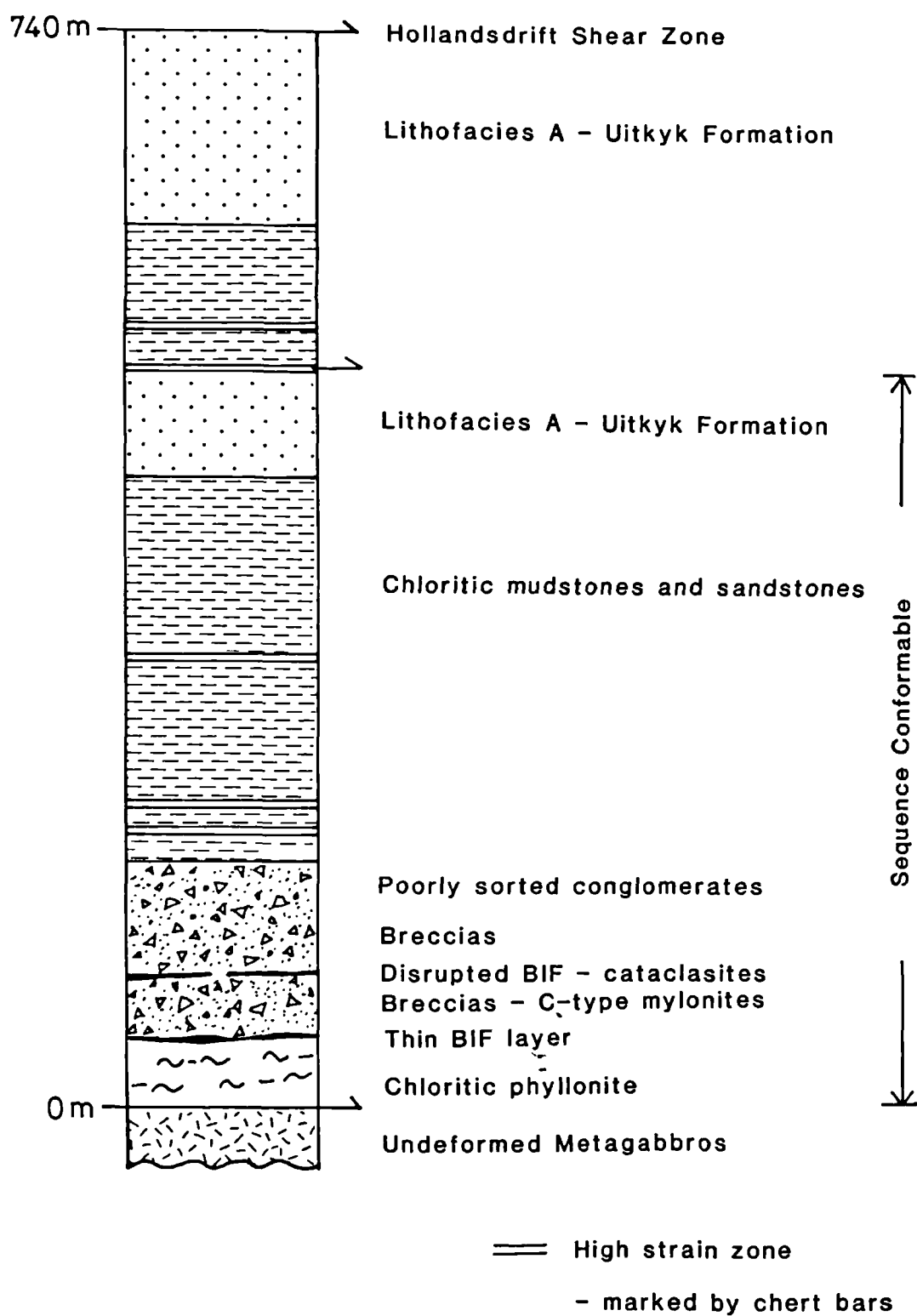
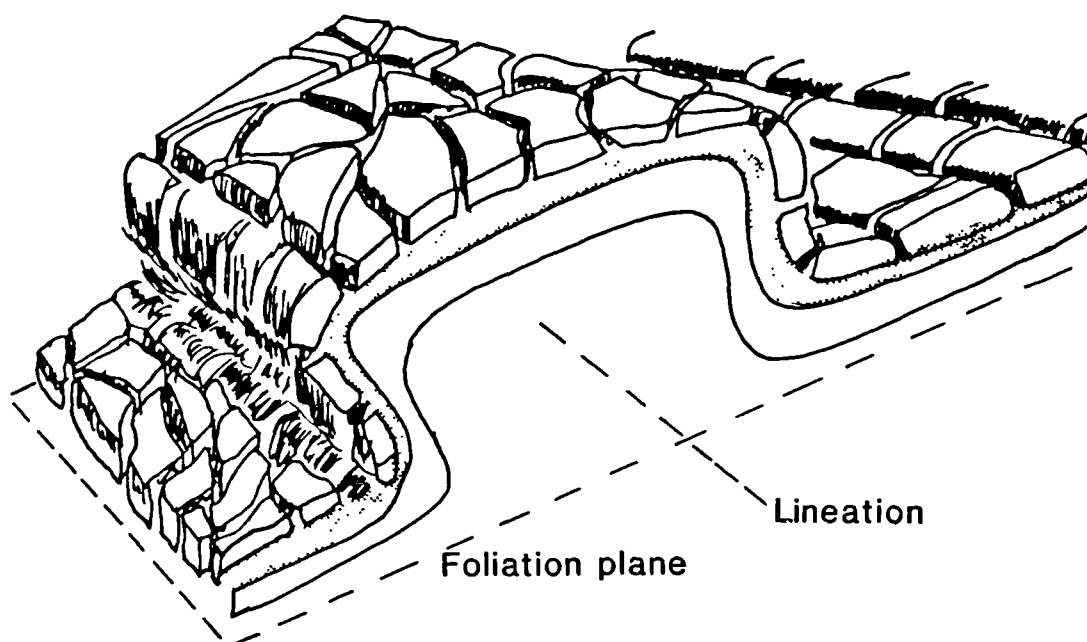
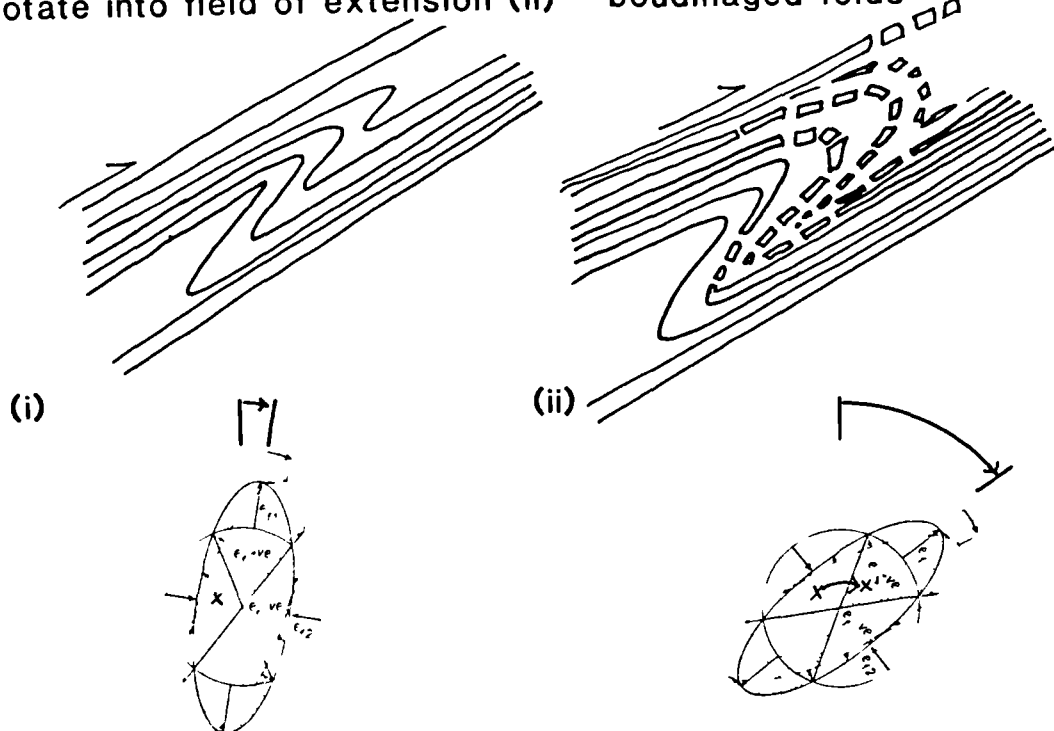


Fig. 4.2 Measured section along traverse line B-B' Fig. 4.1

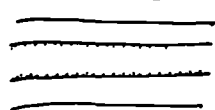
A. Irregular chocolate boudinage and progressive sheathing-off of chert layers



B. Progressive layer-parallel shear of BIF – early formed folds (i) rotate into field of extension (ii) – boudinaged folds



C. Non-isochoric conversion of magnetite to siderite causes hydraulic fracturing



Mag → Sid

$$\delta_s - p = t$$

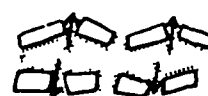
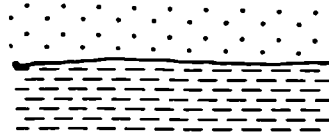


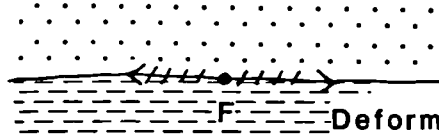
Fig. 4.3 Deformation of BIF blocks in melange

1. Sedimentary contact



2. Fault development along contact and propagation sideways

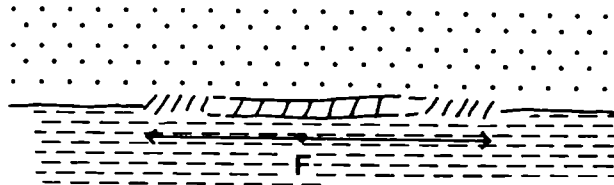
F - Fault



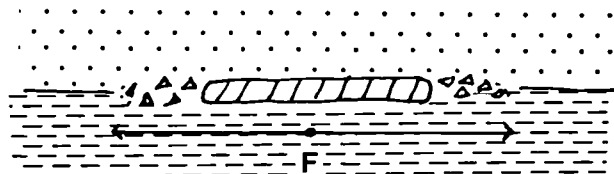
Deformed sedimentary contact

3. Continued deformation leads to protomylonite development

and local mylonites - progressively silicified to form chert bars

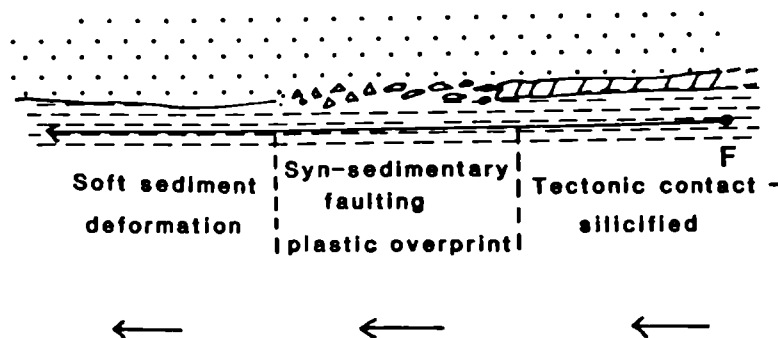


4. Subsequent cataclasis causes brecciation of chert as fault propagates into soft sediment at fault extremities



5. Further deformation causes clast reorientation and plastic overprint

← 3km along strike →



These boundaries migrate outwards as deformation progresses

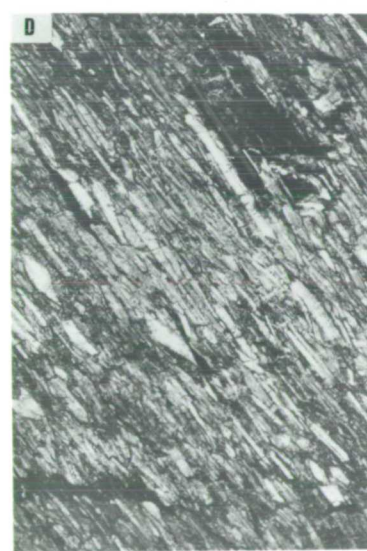
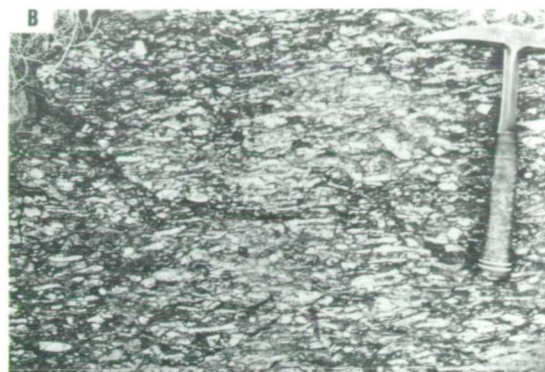
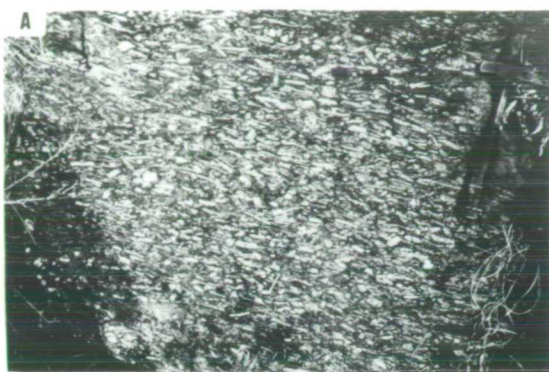
Fig. 4.4 Model to explain observed gradation between lithological contacts

Plate 4.1 Textures of breccias at base of melange

4.1a and b. Two different examples of the breccia in the YZ plane of the strain ellipsoid, which in this case is a horizontal plane with South to the top of the photographs. Note the clast supported framework, regular rectangular to rhombohedral shape of the majority of the clasts, and their strong preferred geometric orientation sub-parallel to the phyllonitic matrix fabric.
GR X +2 658 150, Y -35 000.

4.1c and d. These photographs show an example (and close up) of the XZ plane of the strain ellipsoid from the same outcrop as plate 4.1b, with South now to the left of the photographs. There is a stronger preferred orientation in this plane than the YZ plane and also the clast supported framework is developed to a greater degree. Note also that there are a greater number of very long clasts (up to 40cm long) but the majority of them still have well defined sharp edges with little evidence of recrystallised tails; indeed, brittle deformation of the clasts is prevalent. GR X +2 658 150, Y -35 000.

4.1e and f. Two more views of breccia in the YZ and XZ plane of the strain ellipsoid respectively. Note in 4.1 e the breccia has a more irregular appearance and resembles a sedimentary breccio-conglomerate. There is still a strong preferred orientation in the XZ plane in 4.1f.
GR X + 2 658 000, Y -35 320.



**Plate 4 2 Evidence for plastic deformation of clasts and domains,
and nature of large BIF blocks in the melange.**

- 4.2a. Clast with recrystallised tail in the YZ plane of the strain ellipsoid. The degree of clast packing is now much greater than in areas where recrystallisation is not seen in this plane (plate 4.1a and b) indicating that more intense flattening across the fabric has occurred here. Note that pressure solution, especially at clast impingement points (arrowed), may be an important deformation mechanism in the melange. GR X +2 657 890, Y -35 370.
- 4.2b. Plate 4.2a (arrowed) comes from a larger domain, shown here, that has been deformed in a more ductile manner. A 40cm long clast in the top left hand corner (also arrowed) shows evidence of slight folding or warping. GR X + 2 657 890, Y -35 370.
- 4.2c. Large BIF block (arrowed) lying totally within the melange; note the underlying breccia. This particular block is 40cm wide and nearly 20m long, but blocks up to 150m long and 2m wide are also present. The blocks, and the laminations within them, are also aligned sub-parallel to the fabric in the rest of the melange. GR X +2 657 750, Y -35 600.
- 4.2d. A smaller BIF block lying in a finer grained, shaley part of the melange. Such blocks are not common and a continuous range in block sizes is not seen; instead, a bimodal distribution of large blocks and matrix clasts is present. GR X +2 657 700, Y -35 690.

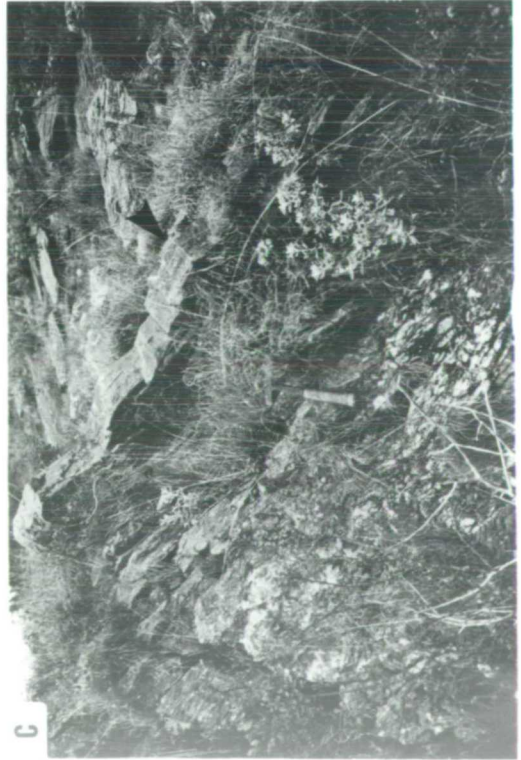


Plate 4 3 Internal deformation of BIF blocks in melange.

- 4.3a. M/W style fold within a single 1 metre wide, 150m long BIF block, taken in the YZ plane of the strain ellipsoid. The fold plunges parallel to the X direction (into the page). Note that both the quartz and magnetite are behaving in a ductile manner. GR +2 658 220, Y -34 970.
- 4.3b. Tectonic imbrication (or mini duplex) of chert slabs in a siderite/magnetite matrix suggesting considerable layer parallel shear occurred within the BIF blocks. It is not clear if the slabs have been imbricated individually or as groups. The sample is a loose block and the *in situ* vergence direction can not be ascertained. GR X + 2 658 300, Y -34 900.
- 4.3c. Internal breccia of white chert clasts derived from the brittle break-up of the quartz laminations. GR X + 2 657 980, Y -35 310.
- 4.3d. Truncation of fold laminations by matrix of melange. However both fold and fabric seem to be of a D2 origin. GR X +2 658 175, Y -35 010.

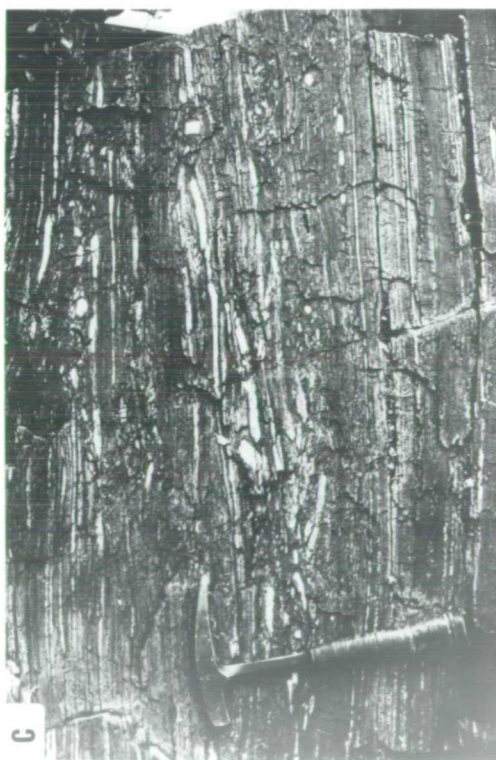
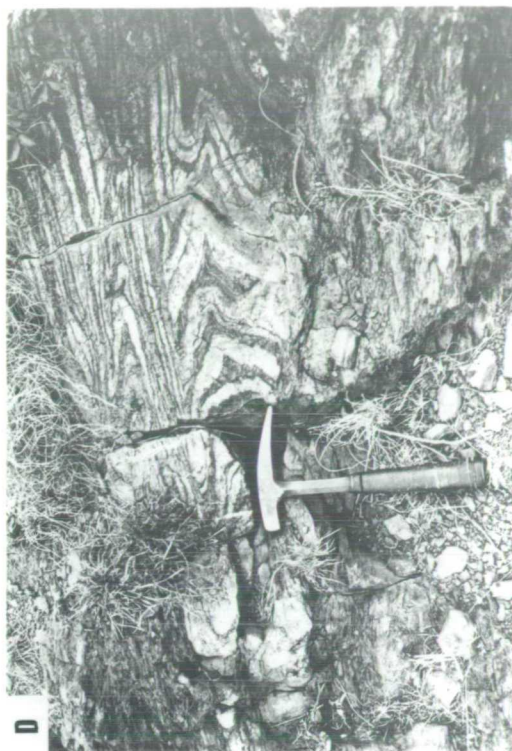
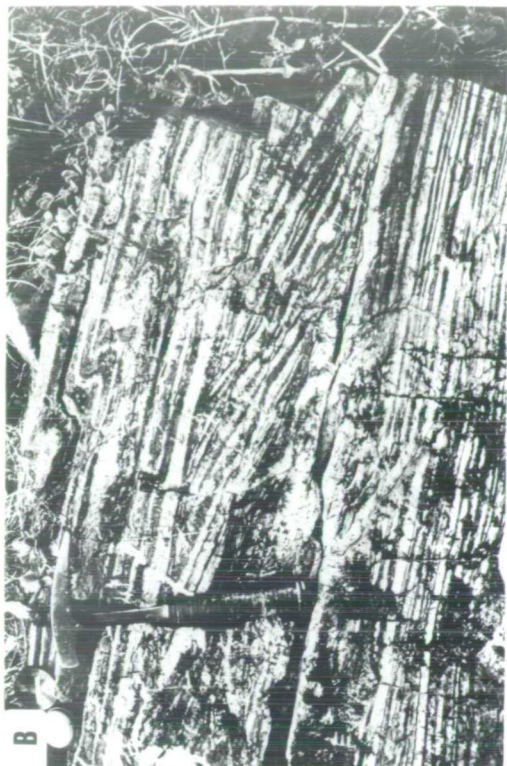


Plate 4.4 Brittle-ductile deformation at edges of BIF blocks
 (i) - 'spalling-off' process.

- 4.4a. Remnant fold hinge in BIF clearly breaking down into the matrix breccia. Layer-parallel extension leads to boudinage and spalling-off of the chert laminations to produce the regular rectangular to rhombohedral clasts seen in the surrounding breccia. Note that the sideritic matrix is behaving in a ductile manner. The fold and brecciation are syn-kinematic with respect to D2. GR +2 658 350, Y -34 850.
- 4.4b. A more irregular breakdown of the BIF blocks. The matrix and breccia, derived from the block, has an irregular, slurry-like, appearance with clast break-up and remnant fold hinges still visible (arrowed). The textures in this plate are thought to represent the processes that produced all the ~~clasts~~ in the lowermost melange by the tectonic milling of the blocks during progressive D2 deformation.
 GR X +2 658 350, Y -34 850.





Plate 4.5 Brittle-ductile deformation of BIF blocks in melange (ii).

- 4.5a. YZ plane of strain ellipsoid. Boudinage of top, outermost layers of a large BIF block indicating that a flattening strain has been an important part of the strain history of the unit. It is thought that the majority of the boudinage is chocolate-tablet like in 3D. GR X +2 657 710, Y -35 660.
- 4.5b. A BIF block with boudinaged upper layers grading into the overlying breccia. GR X +2 657 230, Y -35 590.
- 4.5c. A further example of boudinage, this time at the base of a slab, showing how large slabs are spalling-off from their parental blocks and undergoing concomitant brittle deformation. GR X +2 658 230, Y -34 950.
- 4.5d. Micrographs (in XP) of the hinge (left) and limb (right) of a D2 fold, internal to a BIF block. The grain size, which is very fine, is identical in each and there is no evidence of a preferred crystallographic or shape orientation in the quartz or magnetite layers. Long axis of photograph 1.5cm

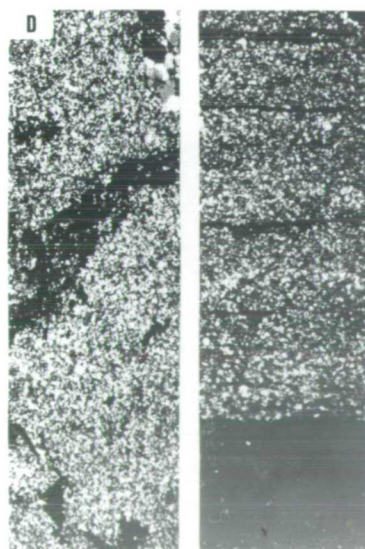


Plate 4.6. Textures in the uppermost melange (i) – evidence for deformation.

- 4.6a and b. Textures showing the marked decrease in clast size and degree of preferred orientation as well as an increase in the percentage of matrix, resulting in a matrix-supported framework. These units are described as poorly sorted conglomerates. GR X +2 658 250, Y -35 030.
- 4.6c. An 'eye' structure visible in the matrix of uppermost melange. It is not clear if the lamination visible represents a true tectonic fabric or is in part a sedimentary lamination. GR X +2 658 220, Y -35 080.
- 4.6d. Isoclinal fold in matrix. All folds seen in the uppermost melange are axial planar to D2. GR X +2 658 210, Y -35 120.
- 4.6e. A small dislocated fold hinge (axial planar to D2) undergoing active deformation in the matrix. Note the extreme granulation of the outer layers (arrowed) and also a possible axial planar stylolite. GR X +2 658 205, Y -35 120.
- 4.6f. Siderite nodules (arrowed) identical to those found in the Uitkyk Formation (chapter 2.5, plate 2.5e). Note the quartz veins that cut only the nodules, suggesting that the veining occurred prior to their incorporation in this part of the melange. GR X +2 658 205, Y -35 120.

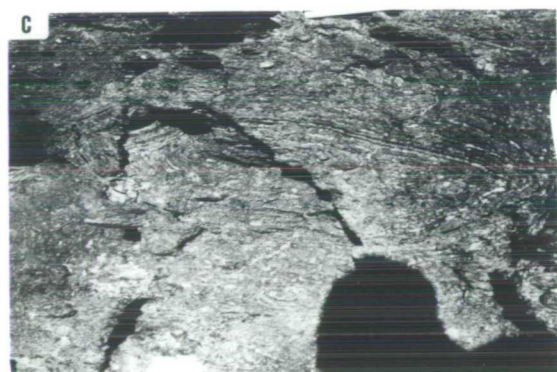


Plate 4.7 Textures in the uppermost melange (ii) – evidence for sedimentation.

- 4.7a. Example of grading from locality C (Fig. 4.1).
GR X +2 657 310, Y -36 260.
- 4.7b. Grading (possibly turbiditic) from locality B (Fig. 4.1).
Note the truncation of the horizontal lamination by an overlying unit (arrowed). A sharp to diffuse silicification front is also present. GR X +2 658 950, Y -34 090.
- 4.7c. Upper contact of melange and overlying chloritic shales and siltstones from locality D (Fig. 4.1). This shows the gradational, sedimentary nature of the contact now cross-cut by the D2 fabric (marked by pen). The shales have also been considerably flattened. GR X +2 658 250, Y -35 080.



Plate 4.8 Evidence for stratal disruption in chloritoid-bearing sandstones and mudstones.

- 4.8a. XZ plane of strain ellipsoid. Sample showing interface between sandstone and mudstone units showing 'blebs' of sandstone spalling-off into the finer grained material and suggestive of soft-sediment deformation. Mixing of the lithologies is occurring (arrowed). Dark, later stage pressure solution seams cross cut these structures. Sample SD 72, locality G, Fig. 4.1. GR X +2 658 780, Y -35 700.
- 4.8b. XY plane of the strain ellipsoid, showing elongate nature of the sandstone bodies enveloped by anastomosing chloritic mudstone zones that act as discrete cataclastic shear zone arrays that progressively dismember the sandstone. The 'spotting' seen is caused by small chloritoid crystals. The white angular fragments are vein quartz clasts. Sample SD 72, locality G, Fig. 4.1. GR X +2 658 780, Y -35 700.
- 4.8c. Optical micrograph showing cataclasis of sand grains and spalling off into the bounding clay zones (arrowed). Late chloritoid laths overprinting the original textures are also present. Sample SD 72, locality G, Fig. 4.1. GR X +2 658 780, Y -35 700.
- 4.8d. Highly deformed original stratal disruption texture. This anastomosing appearance results in the appearance of two fabrics in outcrop, neither of which represents true bedding which has been transposed. GR X +2 658 800, Y -35 730.

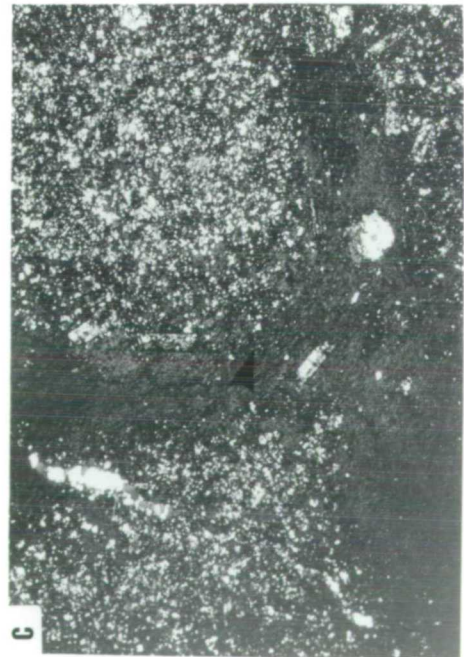


Plate 4.9 Evidence for soft sediment deformation at the base of Lithofacies A, Uitikyk Formation at SD #3 Hill.

Locality L, Fig. 4.1. GR X +2 658 025, Y -33 210.

- 4.9a. View of a small, upright D2 syncline, gently plunging down the hammer handle. The exposure shows an irregular contact of a lower sandstone unit (upper unit in photograph) overlain by a dark green mudstone. Note the fibrous extension veins and slickenslides plunging parallel to the fold axis.
- 4.9b. An unusual 'flame' structure seen in the top right of plate 4.9a. The mudstone unit is apparently injecting into the sandstone unit, axial planar to the D2 fold. The contact between the two lithologies is sharp to diffuse. Note the flame truncates an earlier fibrous extension vein but is cut by a later vein indicating high fluid pressures existed at least transitorily during the soft sediment formation. Note also the presence of chert clasts in the mudstone. Some late (conjugate) D2 faults locally offset the veins.
- 4.9c. Contact of sandstone and mudstone taken just below the flame structure (position shown in plate 4.9a). The mudstone is apparently injecting into the sandstone along discrete planes for up to 1 metre from the contact isolating elliptical bodies of sandstone. Spalling-off of these bodies into the mudstone is occurring (arrowed), further evidence for stratal disruption.
- 4.9d. Large sandstone body lying in the mudstone layer but still attached to the lower sandstone layer in the left of the photograph. The spalling-off texture is well developed around the edges of the body, which has apparently been ripped off the sandstone layer during soft sediment deformation.
- 4.9e and f. Photograph taken only several metres from the flame structure (Fig. 4.3) showing large chert clasts that underwent cataclastic disaggregation in the mudstone. This process supplied the small chert clasts seen in the flame structure and indicates that the cataclasis also occurred whilst the sediments were weakly consolidated.



Plate 4 10 Textures seen along tectonic contacts.

- 4.10a Fibrous extension veins with fibres plunging sub-parallel to the wall rock stretching lineation.
GR X +2 657 890, Y -35 530.
- 4.10b. Laminated grey and white chert bar along a tectonized sedimentary contact. GR X +2 658 500, Y -35 400. Such chert bars are common along tectonic contacts.
- 4.10c. White chert breccia lying in a grey chert matrix.
GR +2 659 000, Y -33 110.
- 4.10d. A remnant rotated clast in a green chert with the D2 fabric wrapped around it. GR X +2 659 110, Y -33 120.
- 4.10e. The chert bars also record a complex history of fluid flow. Here, a green chert (light grey in photograph) is invading and replacing an earlier grey chert. This grey chert is host to the white chert breccia (5.10c above).
GR+2 659 040, Y -34 160.
- 4.10f. Rounded fragments of green chert in a fine grained white quartz matrix. the breccia is cut by dilational quartz veins and both are truncated by a D2 tectonic stylolite.
GR X +2 658 500, Y -35 410.

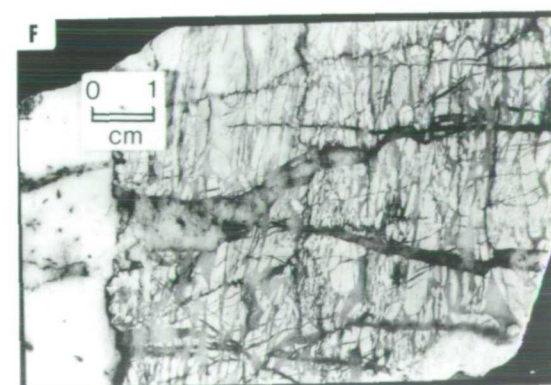
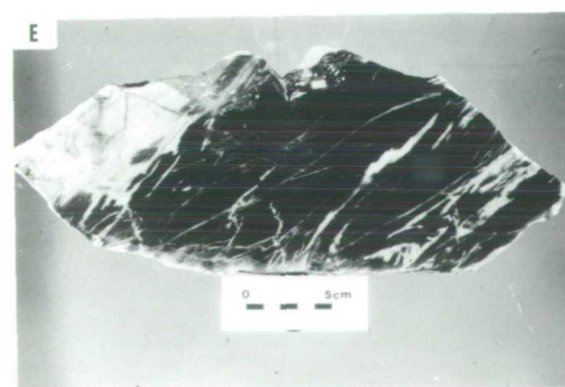
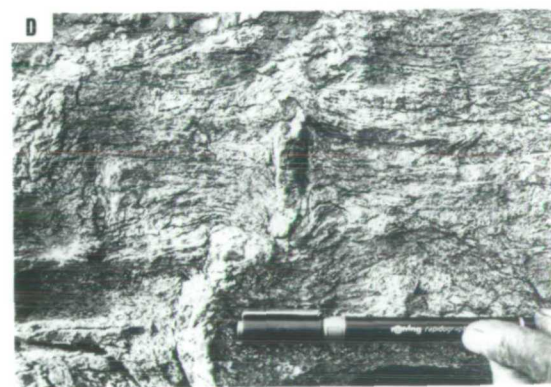


Plate 4.11 Textures along syn-sedimentary faults.

- 4.11a and b. Breccia zone from Zandrivier prospect, locality R (Fig. 4.1) showing evidence for cataclasis in soft sediment. Arrowed in b is a sandstone veinlet that is penetrating into a large chert clast aiding its fragmentation. Disaggregation of the clasts is also evident and is similar to that seen at SD #3 Hill (plate 4.9e and f). The sandstone is draped over and around the clasts and hence appears to have been in a unconsolidated state during brecciation. The breccia zone is bedding parallel (along hammer handle) and occurs at the boundary with an underlying chloritic siltstone layer. This locality is host to a major gold working that exploited a quartz vein along the bedding-parallel fault. The geology of this locality is discussed in more detail in chapter 7. GR +2 658 000, Y -35 760.
- 4.11c. Example of stratal disruption from within a soft-sediment fault zone. Note the strong fabric. GR X +2 659000, Y -34 140.
- 4.11d. Breccia showing a strong, preferred geometric clast orientation fabric.
- 4.11e. A slabbed sample (from the outcrop shown in 4.11d) showing a section close to a circular section of the strain ellipsoid. The matrix to this sample is cemented rock flour and the breccias are best described as foliated cataclasites. GR X +2 659 040, Y -33 050.
- 4.11f. Strongly foliated breccia. Note anastomosing fabric around clasts and associated rounding of clasts and development of pressure solution tails. The strain ellipsoid defined by this fabric is the regional D2 ellipsoid. GR X +2 659 050, Y -35 060.

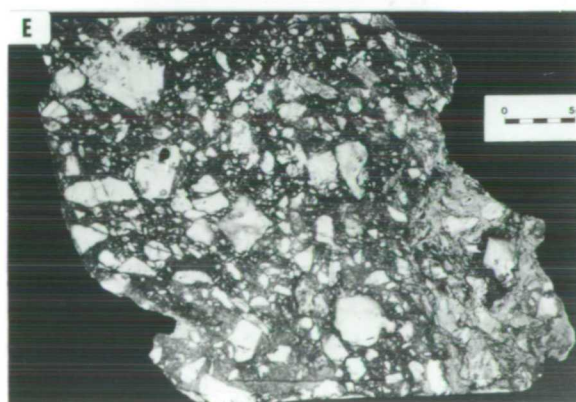
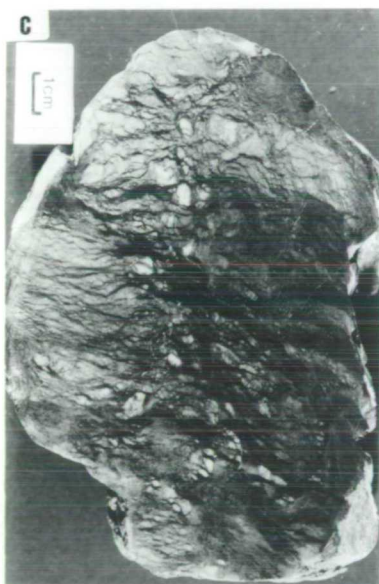
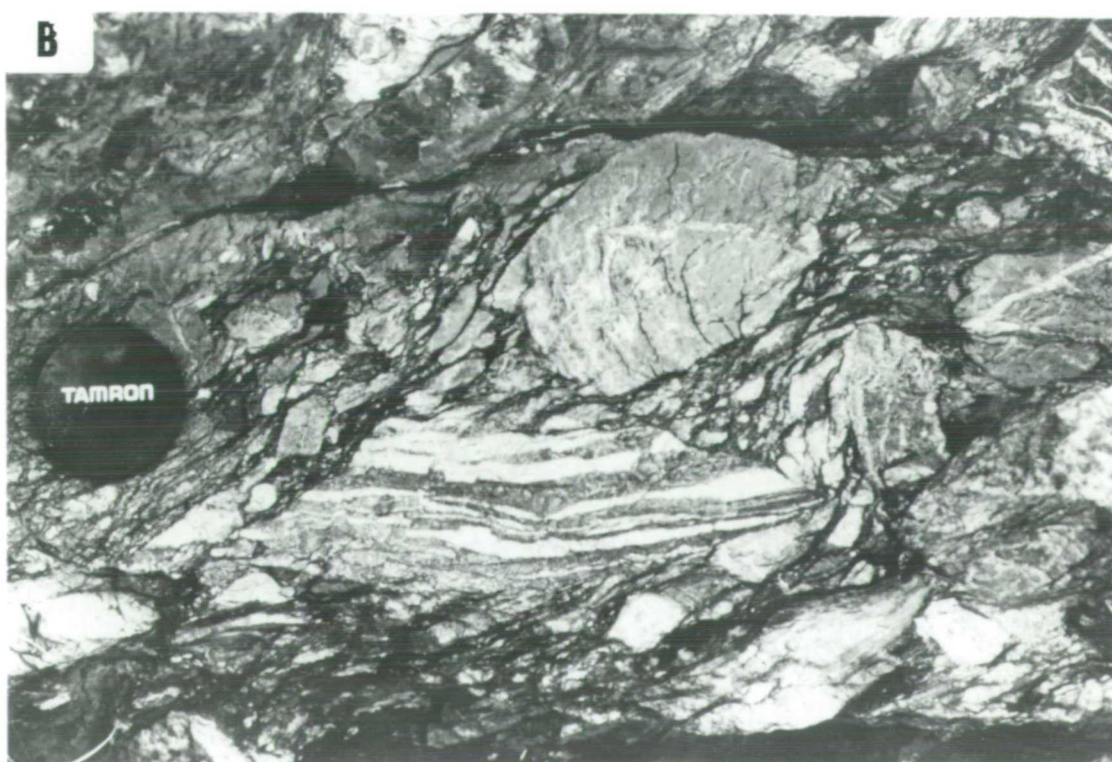


Plate 4.12 Other textures associated with syn-sedimentary contacts.

- 4.12a. Strong pressure solution cleavage (and bedding) preferentially picked out by subsequent tourmaline rich fluids. Non-tourmaline rich pressure solution cleavages are also common. GR X +2 659 000, Y -34 160.
- 4.12b. Large clast of laminated green and white chert. Such chert-type was not observed as a tectonic contact in the SDSZ but was observed in the Kuschke Shear Zone and is described in chapter 8. GR X +2 659 920, Y -34 090.



CHAPTER 5

THE SNYMANSDRIFT SHEAR ZONE – MICROSTRUCTURAL STUDIES

The rocks have "...undergone a more or less intense change in mineralogical composition brought about not merely by the chemical activity of circulating waters but also by pressure due to orogenic forces." Gotz (1887).

5.1 INTRODUCTION

This chapter is concerned with the microscale characteristics of the SDSZ produced during late D2 deformation and in particular the operative deformation mechanisms and environmental conditions that existed during that time. Although fault rocks generally record only the last increments of their deformation history, areas of lower strain within shear zones often preserve textural evidence that can constrain the progressive development of the rock sequences observed (Obée 1985). Such zones do exist in the SDSZ.

The majority of the samples described in this chapter come from the Zandrivier (ZR) prospect located in the central, most highly deformed and metamorphosed, part of the SDSZ (Fig. 4.1). AAC drilled six diamond drill holes on the ZR prospect during this study, one of which, DDH ZR 1 (Fig. 4.1), was drilled right through the the SDSZ and intersected relatively undeformed greenstones of the underlying unit of simatic rocks. This hole has been studied in some detail because it penetrates deep into the SDSZ and places vertical constraints on the evolution of the zone. Fig. 5.1 is a simplified log of the petrology of DDH ZR 1. It shows that a considerable weathering profile extends down to 80-85m in the hole and hence the study is limited by the lack of fresh material in the upper parts of the SDSZ. The borehole also intersected a zone of mineralisation between 90 and 100m depth. This zone is discussed in chapter 7.

The contacts between the different lithologies in DDH ZR 1 are either gradational, usually over less than 1m, or sharp and marked by quartz veins and/or sulphide-rich graphite 'bars'. The contacts show a good correlation with the major lithological contacts seen at surface which dip (50–60°) sub-parallel to the D2 tectonic fabric (Fig. 5.2). The most notable difference between the surface and borehole lithologies is that the latter show a far higher degree of finite strain and primary textures have largely been obliterated. This suggests that *considerable vertical finite strain gradients exist in the SDSZ*. Thus a conformable sequence cannot be observed in drillcore. Table 5.1 summarizes the main microstructural observations and interpretations which show that although there was an initial, strong lithological control on the fault rocks developed, by a combination of grain size reduction and syn-tectonic metamorphism the textural, chemical and metamorphic differences between precursor lithologies were progressively homogenized during D2. Late syn-D2 chloritoid and clinozoisite porphyroblast growth suggest that mid-greenschist facies temperatures of 350–500°C were attained in the SDSZ. Pressure estimates are poorly constrained but were a *maximum* of 3Kb.

5.2 METABASITE UNITS

Fig. 5.1 shows that there are three units of metabasite rocks in DDH ZR 1. Two of these are strongly foliated and lie within the SDSZ whereas the third and lowermost unit intersected is the least deformed and lies outside the main zone of deformation. This latter unit can be correlated with undeformed simatic rocks seen at surface (Fig. 5.2) and in thin section original simatic rock textures are observed (Table 5.1).

Original textures are also found in the other two metabasite units but only in locally preserved areas of relatively low strain. Textural and mineralogical sequences are observed which indicate the simatic rocks were transformed into chlorite-quartz +/- ankerite phyllonites by a combination of cataclasis, diffusive mass transfer and syntectonic metamorphism (Table 5.1 and plate 5.1).

This style of deformation and metamorphism is markedly different to that observed in the undeformed simatic rocks and supports the field and petrological evidence that the metamorphism responsible for the actinolite-albite mineral assemblages *predates* D2 and occurred early in the history of the greenstone belt (the M1 event – chapter 2 5 4 and chapter 3)

5.3 AUGEN C-TYPE MYLONITE-PHYLLONITE

Lying between the two lowermost phyllonite units in DDH ZR 1 is a 60m wide, strongly differentiated tectonite lithology comprised of highly strained domains of iron-rich chlorite which surround lower strained augen (or microlithons) of siderite and quartz (Fig. 5.1). The size and shape of the augen is heterogeneous but many of them have a lozenge or rectangular boudin-like shape, occasionally with thin laminations of siderite +/- magnetite in them (plate 5.2a). The base of the lithology is marked by a 2m wide band of laminated oxide-carbonate facies BIF showing pinch and swell and boudinage structures along some of its layers. These structures and the position of the unit in the borehole (Fig. 5.2) indicate that the lithology can be correlated with the tectono-sedimentary melange seen at surface (chapter 4.4) where the augen represent highly strained clasts originally tectonically derived from a BIF unit. Textural and mineralogical descriptions are given in Table 5.1 and it is evident the clasts have been strongly overprinted by ductile deformation mechanisms, predominantly diffusive mass transfer of quartz and crystal plastic deformation of siderite. There are also many thick layers of chlorite, especially towards the top of the unit, which are phyllonitic. Although these can be correlated with the chloritic mudstones seen at surface (Fig. 5.2) it is not evident in drillcore what the protolith of the phyllonites was. Both plastic and brittle processes played an important role in the development of this tectonite and it is best termed a c-type mylonite-phyllonite (Appendix 3). The elongate tails and general 'smearing out' of the augen were probably caused by pressure solution processes, aided by the initial fine grain size of the chert and influx of fluids along grain boundaries of the quartz to produce the oriented chlorite laths. Some grain size reduction of the quartz probably accompanied the pressure solution.

5.4 DEFORMATION OF SEDIMENTS AND BRECCIA ZONES

A quartz, sericite +/- chlorite bearing unit lies between 92 and 98m in DDH ZR 1 (Fig. 5.1). The lithology, which is also found in DDH ZR 2 and 3 (Fig. 4.1) contains a significant percentage of quartz together with zones of deformed breccio-conglomerate that exclusively contain chert clasts. These zones and the lithology's position in the borehole suggest it can be correlated with the breccia zones (syn-sedimentary faults -chapter 4.8.3) and associated sediments seen at surface (Fig. 5.2).

Like the augen c-type mylonite, in the drillcore the breccias and associated sediments have been strongly overprinted by ductile deformation. On the ZR prospect the syn-sedimentary breccia described at locality R (Fig. 4.1 - chapter 4.8.3,) is found in a low strain lens within the central SDSZ. Along strike in both directions, over a distance of 10 to

50 metres the breccias become c-type mylonites which are similar to those observed in the core. Samples collected over this transition and from the boreholes have been studied to characterise the microstructural evolution of these zones and the surrounding sediments and this is summarized in Table 5.1.

The progressive deformation of mudstone units has not been studied microstructurally. Field observations indicate they become very foliated chloritic phyllonites with no primary textural preservation.

In general, the sediments show a slightly different structural evolution to the melange zone and simatic rocks although the final tectonites developed are texturally similar. The original cataclastic flow textures associated with the soft sediment deformation are overprinted firstly by more localised cataclasis and pressure solution, then by more distributive ductile deformation to produce a quartz-sericite +/- chlorite phyllonite. It is inferred that *diffusive mass transfer* and *DAGBS* were the main operative deformation mechanisms in the lithology, with minor crystal plastic deformation of coarser ribbons, or lenses of quartz, some of which represent deformed quartz veins.

5.5 FOLIATION DEVELOPMENT

The microstructural observations in the augen c-type mylonite and sediments reveal a consistent pattern of foliation development. Original quartz and chert clasts underwent grain size reduction predominantly by cataclasis and pressure solution processes, accompanied or aided by the growth of phyllosilicates in recrystallised or broken zones. The phyllosilicates are found as small (20–50 microns long) discrete laths oriented along grain boundaries of the quartz in the augen mylonite and chert clasts (step 1 and 2, Fig. 5.3) and sub-parallel to the direction of clast elongation. As these ‘mantles’ are traced into the higher strain phyllosilicate domains the ratio of phyllosilicate to quartz increases, with the laths becoming larger, more ubiquitous, and commonly merging. The quartz grains become relatively more elongate (step 3 and 4 Fig. 5.3, plate 5.3d). Eventually the coalescence of phyllosilicate domains isolates clusters or single grains of elongate quartz grains. Within the main phyllosilicate domains there is virtually no quartz (step 5) and they sharply truncate quartz grains lying immediately outside (plate 5.3d and e). Subsequent microscale brittle deformation, kinking and crenulation of the phyllosilicate domains has also occurred. The spatial distribution of the e textures depends on degree and nature of shear band development (plate 5.3). A similar textural sequence can be observed in the *siderite* although the inferred grain size reduction does not seem to have been accompanied by the growth of grain boundary chlorites. However both coarse and fine grained *siderite* are truncated by

the larger chlorite domains. The processes described lead to the development of phyllonitic layers wide, and it can be seen that as deformation progressed, the augen c-type mylonite and sediments as a whole became increasingly phyllonitic. The deformation of simatic rocks also produces phyllonites by a similar process except a chemical replacement of amphibole by chlorite occurred (chapter 5.2).

In common with other descriptions of foliation development in strongly differentiated phyllosilicate-quartz tectonites (White and Knipe 1978, Knipe and Wintsch 1985, Bell *et al.* 1986) the observed textural sequence is interpreted as one in which quartz and carbonate diffused away from and/or along the high strain chlorite domains which developed and grew as deformation progressed. Shearing of the quartz clasts was apparently accompanied by fluid influx into the new finer grained areas along grain boundaries to form oriented chlorite laths. Such laths, as they grow and coalesce, may aid a change in deformation mechanism to DAGBS in the chlorite rich domains (White and White 1981). Therefore the difference between the augen and chlorite domains can be interpreted in terms of temporally and spatially different strain rates with the chlorite domains being strain softened with respect to the microlithons owing to a change in deformation mechanism. Geometric softening due to the preferred alignment of the chlorite grains may also have been important (White *et al.* 1986).

A corollary of these textural interpretations is that the fabric seems to propagate in three dimensions from a central point and thus its development may be analogous to the propagation of an elliptical shear crack (Rutter and Brodie 1985). It is the apparent growth from a series of nucleation points, interaction between and common coalescence of the fabric domains that is responsible for the textural patterns seen in the mylonite. Such microcrack propagation along grain boundaries (and also in the feldspars in the metabasites) is generally non-catastrophic. Stress corrosion due to fluids at the crack tip may therefore have been an important deformation mechanism throughout the deformation history (Atkinson 1982, White and White 1983).

5.6 PHYLLONITES

Both chlorite-quartz+/-sericite and sericite-quartz+/-chlorite phyllonites are found in DDH ZR1 (Fig. 5.1). Their microstructural characteristics are summarized in Table 5.1.

Although quartz-sericite schists are commonly interpreted as deformed sediments (eg Winkler 1979), there are some samples in DDH ZR 1 that suggest that some of these phyllonites may represent deformed and metasomatically altered simatic rocks. Plate 5.7c

shows the contact between a deformed simatic rock and a quartz-sericite-chlorite phyllonite. The simatic rock contains textures identical to those seen in other deformed metabasite units (chapter 5.2, Table 5.1). The contact between the two lithologies is gradational over one to three millimetres and represents a high strain boundary, characterised by the replacement of the albite-chlorite mineral assemblage with a quartz-sericite-chlorite assemblage (plate 5.7d).

Qualitative SEM work reveals that towards the boundary, the albite crystals become chemically altered to quartz around its edges. Consequently the new matrix grains, chemo-tectonically derived from the feldspar, are quartz. It is not clear if the grain size reduction was by brittle or plastic deformation processes, but the chemical change does not appear to be related to fractures and the fact that some of the new grains are optically strain-free suggests some dynamic recrystallisation occurred.

It seems therefore that the quartz-sericite-chlorite phyllonite is a deformed simatic rock rather than a deformed sediment. A possible explanation for these features is that a pore fluid phase of differing composition to the metabasite may have flowed through the unit causing the albite to break down. Hence, at the boundary, the feldspars come into contact with this different fluid and are replaced by quartz, which can deform more readily. This is thought to be a further example of reaction enhanced ductility (White and Knipe 1978) and also indicates that the phyllonites are extensively strain softened.

One final complexity of the chemical reactions in the SDSZ is shown in plate 5.8f. Here a strongly crenulated, chloritoid-bearing, quartz-sericite phyllonite (which probably represents a deformed sediment as it locally contains remnant quartz clasts) has been completely overprinted and obliterated by a fine grained variably oriented mass of chlorite. The documentation of deformed simatic rocks becoming sericite-rich phyllonites, and deformed sediments becoming chlorite rich phyllonites, indicate considerable chemical exchange has occurred during the metamorphism recorded in the SDSZ which was probably caused by syn-tectonic fluid migration through the zone.

The presence of abundant oriented phyllosilicates and the lack of evidence for grain size reduction in relatively high and low strain domains suggests DAGBS was the major deformation mechanism in the phyllonites (Obee 1985, White *et al.* 1986, Brodie and Rutter 1987). The widespread development of phyllosilicates in the SDSZ phyllonites indicates the fault rocks may also be extensively strain softened due to geometric or fabric softening.

5.7 DISCUSSION

The microstructural and field observations outlined in this and the previous chapter indicate that although there was a wide range in precursor lithologies, their textural and chemical differences were considerably reduced by dynamothermal metamorphism involving an intragranular fluid phase. Both BIF and detrital sediments as well as simatic lithologies and various cherts were converted to a series of hydrated, locally carbonated, extensively strain softened phyllonitic fault rocks. This progressive fault rock formation history is summarized in Fig. 5.4 and corroborates the ideas of Gotz (1887). Rast (1958) also documented the progressive textural and chemical homogenization of different protoliths in tectonic slide zones from the Scottish Dalradian schists. Beach (1982) and McCaig (1984) (amongst others) have also suggested that fluid/rock interaction in hydrating shear zones acts to reduce chemical differences between lithologies.

At the microscale in the SDSZ all deformation mechanisms appear to have been operative, often cyclically, and a complex interplay must have existed between them. However; it is emphasized that intracrystalline plasticity *has played a subordinate role in the progressive deformation history*. Brittle and diffusive mass transfer processes, often cyclically operating (in the form of DAGBS), were the dominant deformation mechanisms. Strong initial lithological controls on fault rock development effectively caused intracrystalline plasticity (especially of quartz) to be 'bypassed' as a major deformation mechanism in the SDSZ. Many of the clasts in the sedimentary units and melange were originally fine grained chert in which there was apparently little need to further reduce the grain size. The coarser grained quartz clasts initially had their grain size reduced by pressure solution processes; feldspar and amphibole crystals were susceptible to microfracturing and alteration which produced fine grained and phyllosilicate-rich, strain softened mineral assemblages. Therefore, although the SDSZ is macroscopically 'brittle-ductile' (descriptively defined) it probably formed in the brittle-semi-brittle field (mechanistically defined). The rheology was not primarily controlled/dominated by the brittle-plastic transition in quartzo-feldspathic lithologies. This must be borne in mind when comparing the SDSZ with published models of shear zone development (eg Sibson 1983) and is considered to have important implications for models of mineralized 'brittle-ductile' shear zones and these are discussed in chapter 8.

5.8 PROGRESSIVE D2 OR A D3 EVENT IN THE SDSZ

The presence of crenulations and transposition fabrics in the phyllonites and other lithologies described above, raises the following question; are these microstructural features

a product of progressive D2 deformation or of a separate, later, 'D3' event?

In thin section it is not possible to distinguish between these two postulates. However in surface exposures within the SDSZ there is no evidence for a widespread D3 event such as reorientation of lineations, folding of foliations etc., unless it was strictly coaxial with respect to D2. Only two isolated occurrences of well developed crenulations of the D2 fabric were observed on surface (localities I and K, Fig. 4.1). Therefore, progressive D2 deformation is the favoured cause of the fabric development history within the SDSZ. Hence a time framework for fabric development *within* the D2 event can be formulated as follows;

D2₁: cleavage development in sediments (involving local transposition of bedding (chapter 4.5);

D2₂: crenulation cleavage and phyllonite development;

D2₃: transposition of D2₂;

D2₄?: minor crenulation of D2₃.

This 'fabric history' (summarized in Fig. 5.5) indicates that in the centre of the SDSZ there is a considerable deformation overprint on many early D2 structures. The spatial distribution of the different stages indicates considerable vertical and lateral strain gradients existed in the SDSZ. Early stages are preserved at the edges of the zone and in areas of low strain within it; late stages are found either in localized high strain domains or distributed throughout the lithologies. The spatial distribution is important to consider when investigating timing of metamorphic porphyroblast growth.

5.9 PORPHYROBLAST DEVELOPMENT

5.9.1 Introduction

Both at surface and in drillcore the lithologies present contain variably developed chloritoid and clinozoisite porphyroblasts. Many textural criteria have been proposed for the elucidation of the timing relationships of porphyroblast growth with respect to foliation development in tectono-metamorphic rocks and it is evident that extreme care must be taken in their interpretation (Zwart 1960, Spry 1969, Ferguson and Harte 1975, Bell *et al.* 1986, Prior 1987). Most of the criteria proposed so far are based on the geometric relationship

between inclusion trails (S_i) preserved in the porphyroblasts and the prevalent external matrix foliation pattern (S_e). From these relationships the timing of porphyroblast growth has generally been deduced as pre-, syn- or post kinematic, with pre- to syn-kinematic blasts inferred to have rotated after or during their growth respectively in simple shear deformation (Zwart 1960, Rast 1963, Rosenfeld 1970, Schoeneveld 1978).

In contrast, Ramsey (1962) showed that flattening, or irrotational homogeneous strain, of a schistosity around a *static*, non-compressible, growing crystal could lead to inclusion patterns similar to those envisaged by a simple shear-rotating porphyroblast mechanism. In a series of papers, Bell and his co-workers (Bell 1981, Bell and Rubenach 1983, Bell 1985, Bell *et al.* 1986) have proposed that virtually all porphyroblasts are syn-tectonic and do not rotate during or after their growth (providing they do not internally deform). Bell proposes that porphyroblast growth is strongly linked to crenulation cleavage development and partitioning of deformation into components and domains of pure and simple shear. The porphyroblasts are inferred to grow in the domains of pure shear (a la Ramsey 1962). These models infer that it is the matrix which rotates and not the porphyroblast. De Wit (1976) also proposed a genetic link between crenulation development and porphyroblast growth and concluded that most porphyroblasts in regionally metamorphosed rocks were syn-tectonic. Both de Wit (1976) and Bell and co-workers (*op. cit.*) also stated that previous conflicting interpretations of porphyroblast/matrix relationships arose from an *incomplete understanding of the progressive fabric development in the matrix*. A critique of the timing of porphyroblast growth is not given here and to avoid getting embroiled in long standing arguments, only simple observations on porphyroblast growth in the SDSZ have been made in order to determine their timing with respect to deformation.

5.9.2 Textural observations of porphyroblasts

Two different porphyroblast types have been identified in the SDSZ; clinozoisite and chloritoid.

1. Clinozoisite: The clinozoisites, which were only found in drillcore, are present in the chlorite-quartz phyllonite above the augen mylonite (Fig. 5.1). The phyllonite is at stage D2₂ of fabric development (Fig. 5.5) and contains poorly developed crenulations. The clinozoisites do not contain any microscopically visible inclusion trails and it is difficult to interpret their timing with respect to surrounding structures.

The clinozoisite crystals are generally sub-hedral to euhedral and are found both in chlorite rich domains, where they appear to overprint the fabric, and in quartz-carbonate rich microlithons. They are also associated with unusual orbicular structures with the

clinozoisite radially disposed around them (plate 5.8a). These features indicate the blasts do not predate D2. However, a significant number of the crystals in the microlithons are sharply truncated against the phyllosilicate domains (plate 5.8b) and others have fibrous overgrowths of quartz and chlorite in pressure shadow areas. This is interpreted to indicate dissolution of the porphyroblasts has occurred and hence suggests there was at least localised post-porphyroblast D2 deformation. There is little evidence for rotation of the blasts and in fact it is unlikely that the orbicular structures could have rotated at all without breaking up. It is concluded that porphyroblast growth was late syn-D2₂, and that there was little or no subsequent rotation.

2. Chloritoid: Three different morphologies of chloritoid porphyroblast have been identified;

Type 1a - individual laths found in metasediments that contain original soft-sedimentary structures;

Type 1b - individual laths found in quartz-sericite phyllonites in drillcore;

Type 2 - equidimensional rosettes of polysynthetically twinned crystals;

Type 3 - elongate rosettes, or 'bow-ties', associated with tourmaline.

Types 2 and 3 were only seen at surface and come from the ottrelite schist unit of Hall (1909) and Willemse (1938).

Inclusion trails in the chloritoids are variably developed; it is possible to make the following observations;

Type 1a chloritoids are euhedral (up to 0.5 x 0.25 mm in size), apparently randomly oriented and are not associated with crenulations. No inclusion trails were observed. The blasts overprint soft sediment deformation and stratal disruption structures (plate 4.7c).

Type 2 chloritoids occur in two different areas. In the low strain lens at ZR (locality R, Fig. 4.1) they occur as rosettes (stellate growth) up to 1.5mm in diameter which overprint the D2₁ cleavage (de Wit 1985 - Fig. 21). Away from this lens, the chloritoids exhibit a close spatial relationship with crenulation development in the fabric of their host lithologies, being found only in the quartz-dominated microlithons between the crenulation cleavage (plate 5.8c). In the samples studied the fabric in the host lithology is at stage D2₂ (Fig. 5.5) and, where visible, S_i in the rosettes closely mimics S_e in the microlithons but does not overgrow the crenulation cleavage itself. S_i is also fairly constant between rosettes *on the scale of the thin section* (plate 5.8c) regardless of the orientation or size of rosette. This suggests there has been little or no rotation of the rosettes (Bell 1985). Timing of rosette growth with respect to crenulation development is difficult to establish. However, the S_i at

the edges of the chloritoids is commonly curved and possible truncation of chloritoids against the crenulations was observed locally. These features suggest porphyroblast growth was closely associated and possibly synchronous with crenulation development (Zwart 1960, de Wit 1976, Bell *et al.* 1986).

Type 1b chloritoids were found in the strongly differentiated quartz-sericite phyllonites of stage D23 (Fig. 5.5). They also show a strong association with microlithons (plate 5.8c). In these samples, S_i in most porphyroblasts does not bear any resemblance to S_e in the surrounding microlithons. It can be concluded there has been differential rotation of the matrix and porphyroblast and therefore that the chloritoids pre-date D2₃. This interpretation is supported by the truncation, suggestive of post porphyroblast dissolution, of some of the chloritoids against the sericite domains.

The bow ties (up to 2.5mm by 75 microns in size) that constitute type 3 chloritoids, occur just below the low strain lens at ZR (locality L, Fig. 4.1). The rosettes make up nearly 75% of the rock and lie in a tourmaline-chlorite-sericite matrix (plate 5.8e). The matrix carries an interpreted D2₂ fabric (Fig. 5.5), with mainly wavy, open folded crenulations and occasional sharper defined crenulations. Again the rosettes are associated with the lower strain microlithons and, in some cases, S_e passes into S_i without any significant deflection. The tourmaline found in this lithology appears to be a post-chloritoid hydrothermal mineral which has mimicked a pre-existing fabric. In matrix-rich areas, original phyllosilicates are seen to be overprinted by the tourmaline which, although generally following the fabric, is locally misoriented from it by up to 20°. Furthermore, apparent crenulations of tourmaline show no evidence of deformation such as fracturing or straining. The size of the rosettes, their close packing and the similarity of S_i to S_e in the microlithons make it unlikely that the chloritoids have rotated after their formation.

3. Discussion of chloritoid-matrix relationships:

The relationships described for the chloritoid porphyroblasts suggest they i) overprint both soft-sedimentary structures and D2 cleavage in the sediments, ii) are also closely related to crenulation development in more deformed samples and iii) pre-date the development of a strongly differentiated crenulation cleavage in the most deformed samples. This series of relationships could be interpreted two ways (and expressed schematically in Fig. 5.5); either that;

A) a complex history of repeated chloritoid growth at successive stages of fabric development or;

B) a simpler, one stage, syn-D2₂, chloritoid growth which statically overprinted areas of low strain (D2₁) in certain parts of the shear zone but formed synchronously with

crenulation development elsewhere. Subsequent localised D2₃ deformation gave rise to highly discordant S₁/S₀ relationships.

The latter model is more consistent with the observations and is the preferred interpretation.

Although this interpretation is independent of matrix vs. porphyroblast rotation arguments, it is entirely dependent on the interpretation of progressive D2 fabric development. Samples taken in isolation from the area without recognizing such a history could conceivably be interpreted as a complex poly-metamorphic history involving several stages of chloritoid porphyroblast growth. The proposed syn-D2 origin for chloritoid growth differs markedly from the interpretation of de Wit (1985) who believed the SDSZ had been affected by three periods of metamorphism, designated M2, M3 and M4. The M2 event was syn-tectonic with respect to D2 deformation and resulted in the 'dynamic' development of aligned phyllosilicates which defined the D2 fabric. M3 and M4 were believed to be periods of 'static' recrystallization or 'thermal peaks' under greenschist facies metamorphism which resulted in the development of margarite (a brittle mica) and chloritoid porphyroblasts respectively. No margarite was observed in thin section in this study. The interpretations differ because the samples studied by de Wit come specifically from the low strain lens at ZR where the chloritoids do statically overgrow both D2₁ cleavage and margarite porphyroblasts (Figs. 9 and 21 - de Wit 1985). These relationships lead him to conclude there were post D2, M3 and M4 events. The progressive D2 fabric development recorded in this study was not observed by de Wit (1985). Thus there is no evidence for 'M3' and 'M4' being discrete separate events that post-date D2.

Although margarite was not observed in thin section in this study, in the field it is only found within several metres of a quartz-tourmaline-gold vein in the low strain lens. It is possible the margarite is related to the mineralization and does not reflect a separate metamorphic event. It is evident however that more detailed studies of the timing of porphyroblast growth with respect to deformation (and mineralization) in the SDSZ are urgently required.

5.9.3 Summary and estimation of P-T conditions during D2

Porphyroblast growth in the SDSZ was essentially a syn-D2, and probably a syn-D2₂, phenomenon that was strongly controlled by lithology with clinozoisites forming in chlorite rich phyllonites and chloritoid in sericite rich phyllonites and deformed sediments. These mineral assemblages indicate mid greenschist facies, peak metamorphic conditions were attained during D2. The assemblages constrain the temperature of formation as being between 350 and 500°C, but do not tightly constrain the pressure conditions. Winkler 1979,

Phillips 1987a). The maximum pressure attained during D2 was probably no more than 3–4kb (Turner 1968, Winkler 1979). It is emphasized that both clinozoisite and chloritoid can form at much lower pressures and therefore the actual depth of formation depends on the prevalent geothermal gradient. Given the gradation from tectonic to sedimentary processes observed at surface, the pressure of 3–4kb is considered to be too great. 1–2kb is probably a more realistic estimate. This implies considerable vertical geothermal gradients existed in the SDSZ, perhaps driven by regional magmatic activity. These possibilities are considered further in chapter 8. The high vertical strain gradients in the SDSZ suggests higher than 'normal' thermal gradients could have existed during D2.

The sedimentary units in the SDSZ underwent prograde metamorphism during D2, whereas the pre-existing simatic rocks underwent retrograde metamorphism from upper greenschist/lower amphibolite facies (chapter 3). The originally different metamorphic grades of the precursor lithologies effectively 'equilibrated' at the ambient conditions of dynamothermal metamorphism in the shear zone. In a similar fashion, Jamieson (1986) has suggested that the supposed 'inverted metamorphic gradients' commonly seen immediately below obducted ophiolites are the result of the fortuitous preservation of residual metamorphic assemblages from different crustal levels within epidote-amphibolite facies shear zones.

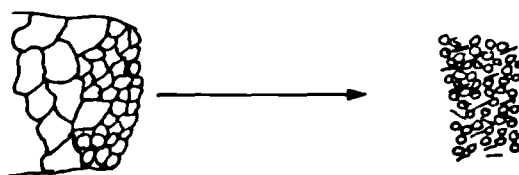
5.10 SUMMARY

The field and microstructural features documented in this study have revealed a progressive sequence of fault rock development in the SDSZ which can be adequately described using Obee's (1985) classification. Although there was a very strong initial lithological control on the fault rock produced, the textural, metamorphic and chemical differences between precursor lithologies were evidently progressively homogenised during continued deformation. The homogenisation was carried out by a combination of grain size reduction and metamorphic reaction, involving the cyclical operation of (minor) crystal plasticity, cataclasis and especially fluid/rock interaction involving an intragranular fluid phase. Original sedimentary and igneous rocks underwent prograde and retrograde metamorphism respectively and were converted to a series of greenschist facies (350–500°C, <3kb) chlorite-quartz and sericite-quartz phyllonites. DAGBS was of prime importance in this deformation sequence. The phyllonites represent extensively hydrated, locally carbonated and strain softened fault rocks in which foliation development was a complex process and involved several intermediate syn-D2 stages.

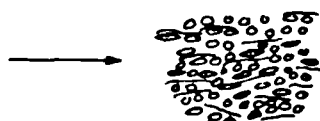
These observations have an important implication for field mapping and

metamorphic studies in general. Unless characteristic *textures* of precursor lithologies can be identified in tectono-metamorphic rocks, either in areas of low strain or by tracing them into deformed zones, then because of possible major chemical reactions involving a through-flowing fluid phase, it could be erroneous to assign all quartz-sericite schists as deformed sediments and chlorite-quartz schists as deformed basic rocks. This is especially true of mineralized shear zones where the attendant hydrothermal fluid flow has the potential to completely obliterate earlier chemical features (eg Chown *et al.* 1984).

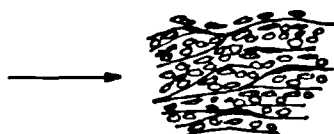
1. DAGBS aided by fine grain size, grain boundary permeability and development of oriented chlorite laths. Annealing forms "pseudo-core-and-mantle" structure.



2. Growth and coalescence of chlorite laths. Micro-cracking important. Quartz begins to develop preferred shape orientation



3. Chlorite domains begin to merge and isolate quartz microlithons



4. Continued chlorite growth leaves isolated elongate quartz grains which are gradually removed by dissolution



These processes overlap temporally. The spatial distribution varies depending on the degree of shear band development.

Fig. 5.3 Simplified process of fabric development

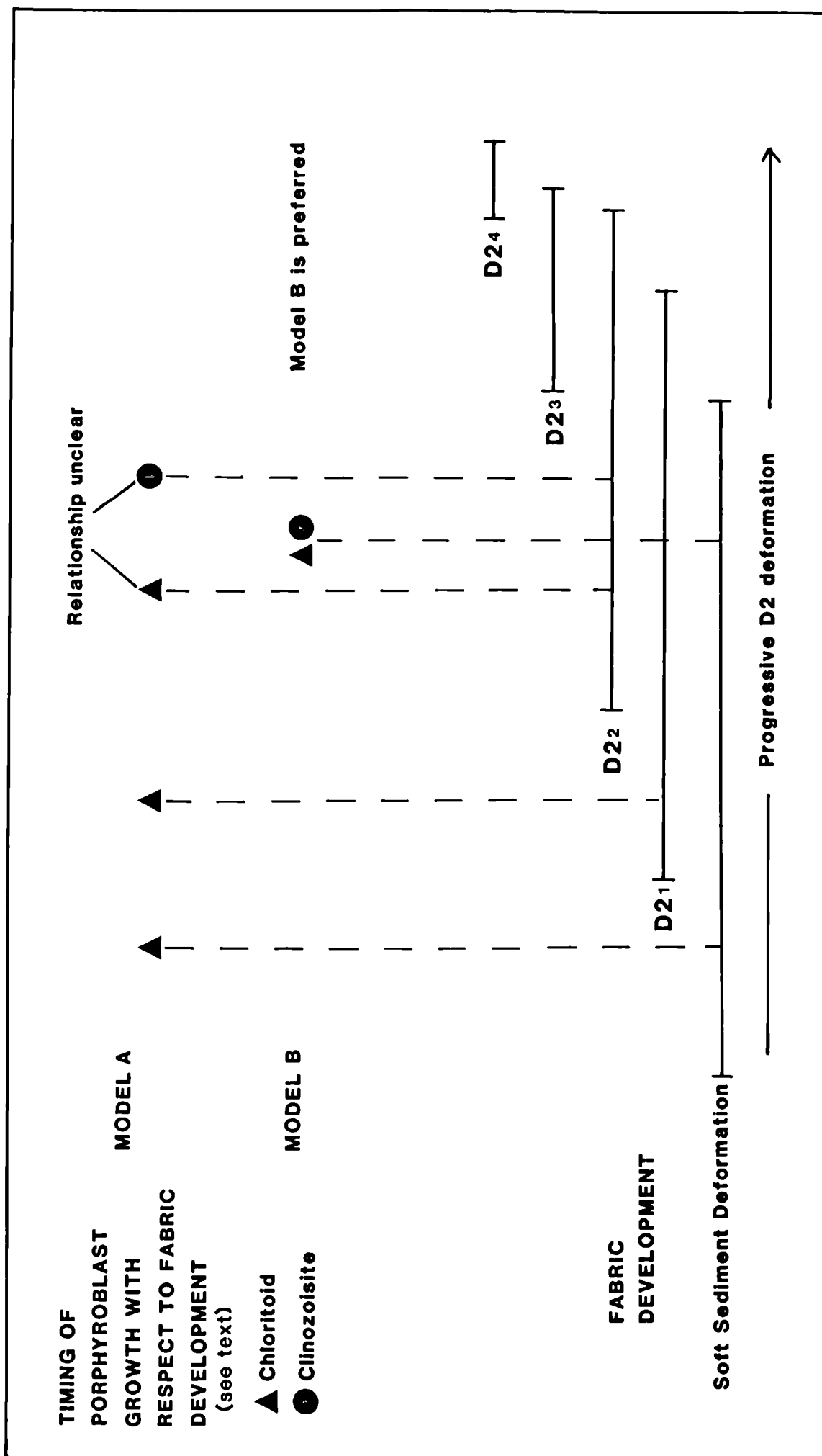


Fig. 5.5 Relative timing of D2 fabric development and porphyroblast growth

TABLE 5.1 TEXTURAL OBSERVATIONS AND INTERPRETATIONS OF MICROSTRUCTURES IN DDH ZR1

Lithology	Location	Strain	Mineralogy and Texture	Deformation Mechanisms	Interpretations
Metabasites	Unit 1 - lowermost unit outside S05Z. Unit 2 - within S05Z. (Figs 6.1, 6.2)		Ab-act-subophitic texture, fibrous act overgrowths identical to textures in metagabbros (Chap 3.2).		Remnant igneous texture.
		Med	Ab - extensive trans- and intra-granular micro cracking accompanied by development of oriented chlorite in pressure shadows and between boudins (Plate 6.1c and d). Some alteration to chl-qtz. Act - strained, fractured, occasionally folded (Plate 6.1b). Altered to chlorite, concomitant reorientation into shear bands (Plate 6.1a). Chl - weak to strong foliation anastomoses around microlithons of ab-act.		Reaction enhanced ductility (White and Knipe 1978) by syntectonic replacement of act by chl. Inferred fluid penetration, aided by cataclasis causes hydration and retrogression of ab-act to chl-qtzank.
		High	Simatic texture progressively destroyed to produce chl-qtz-absank and then chl-qtzank phyllonites.		
Augen c-type mylonite phyllonite	DDH ZR1 - depth	Med	Augen - well developed elongate qtz tails (Plate 6.2b), often reoriented in shear band geometries. Clast centres are annealed (100 micron) but surrounded by a mantle and tail of finer grained qtz (20-50µm) with small chl laths along grain boundaries (Plate 6.2c). These laths become larger and merge with main chl domains (Plate 6.2c and d). Some of the larger clasts are cut by similar zones of fine grained qtz, often exploiting pre-existing laminations (Plate 5.2d).	Diffuse mass transfer. Static recovery.	Pseudo-core and mantle textures. BIF clasts initially had a fine grain size similar to that in the 'mantles' (Chap 2.4). The triple point junctions are annealing or static recrystallisation features. Grain boundary inclusions tend to inhibit static growth (Hobbs et al. 1976) and therefore annealing only affected clast centres. The role of intracrystalline plasticity is subordinate.
			Sid - unit is extensively carbonated, found preferentially in microlithons. Core and mantle structures common (Plate 6.2e). Basal BIF layer is pervasively carbonated due to extensive veining and passive infiltration - often associated with growth of iron-rich chlorites (Plate 6.2f). Chl - defines main foliation, heterogeneous spectrum of textures exists between foliation and clasts (Plate 6.3). Shear band development common, no difference in grain size between shear bands and primary foliation.	Crystal plastic dynamic recrystallisation, metasomatism. Microcracking, diffusive mass transfer. DAGBS.	Siderite probably introduced syntectonically and remobilised by plastic (and probably) brittle processes.
		High	Progressive shearing produces strongly differentiated qtz-chl and chl-qtz tectonites. Both primary and shear band foliations are often arranged en echelon. Lateral terminations of foliation domains also 'step down' en echelon with lozenge shaped steps marked by small chl laths and elongate qtz grains all oriented parallel to main fabric or by euhedral siderite (Plate 6.3f).		Dilatational jogs; seen on all scales in fault zones, e.g. Sibson 1986, McCaig 1987. Indicate syn-tectonic remobilisation of qtz-chl-sid via an intragranular fluid phase.
Sandstones	Surface.	Low	Angular to subangular qtz grains, indolose extinction, but little or no preferred orientation. Extensive trans- and intragranular microcracks define a rough anastomosing domainal fracture cleavage marked by ser-sph.	Cataclasis, Recovery.	
Breccia Zones	Surface.	Low	Plates 5.10a and 6.4a. Chert clasts - moderate to strong preferred geometric orientation associated with strong domainal, anastomosing orientation associated with strong domainal, anastomosing cleavage seams of ser-chl-gph. Strongest development of seams at clast impingement points. Larger clasts (up to 50cm) have microcracks.	Cataclasis, rigid body rotation.	Unoriented clasts become oriented during fabric development.

			Fine grained chert in clasts has diffuse and gradational contacts with matrix over 100-500 microns (Plate 6.4b). No difference in grain size between chert and matrix, phyllosilicates penetrate along grain boundaries. Quartz clasts - within microlithons, little or no preferred orientation but strong undulose extinction and microcracking. Within cleavage seams clasts are cusped elongate. Seams are crenulated with wavelength controlled by qtz clast length and distribution (Plate 6.4c). Local stylolites along crenulate hinges.	DAGBS	Clasts being reworked into matrix without concomitant change in grain size suggests fine grained nature of chert controlled deformation mechanism. Crenulations localised on outcrop scale caused by a late flattening and propagation of instabilities set up around qtz clasts. Fluid flow through the unit has occurred.
		Med	Plate 6.5. Deformation becomes increasingly distributive: increasing degree of preferred orientation of qtz clasts; development of differentiated qtz-serschl tectonic fabric. Quartz grains develop oriented ser-chl-qtz overgrowths in pressure shadows (Plate 6.5a). Some subgrain development and dynamic recrystallisation, often localised at fibre-clast contact. Direct recrystallisation into matrix with micas along grain boundaries in 'mantles' of recrystallised quartz. Grain boundary micas coalesce to form phyllonitic bands. Large chert/qtz clasts often still preserved (Plate 6.5b and c). Possible late annealing of clast centres.	Minor crystal plasticity. Intense pressure solution and diffusive mass transfer processes.	Intense flattening.
		High	Foliation defined by serschl domains which anastomose around and isolate ellipsoidal microlithons of finer grained qtz and occasional remnant clasts. Further deformation leads to development of ser-qtz phyllonite.	Diffusive mass transfer.	
			In DDH ZR3 a 3m wide 'qtz eye' ser schist contains 3mm² eyes of granoblastic aggregates of annealed quartz (Plate 6.5d).		Annealed remnant clasts.
Chl-qtz-ank phyllonites	Fig. 6.1	High	Fine grained folded masses of oriented chl laths (<50 microns wide by 50-500 microns long) define fabric domain up to 3-4cm wide. Domains vary from rough and wiggly to anastomosing and locally continuous, and surround microlithons of fine grained qtz with minor chl and locally ank (Plate 6.6a). Domain-microlithon boundaries are identical to Fig. 6.3. Some microlithons represent deformed and partly transposed qtz-carb-py veins (Plate 6.6d). Shear band development common, often with concomitant metasomatic replacement of chl by ser (Plate 6.6b). Late ser euhedra. Anr - locally up to 10% fine grained <50 microns, difficult to optically observe deformation textures. Coarser grains are little deformed (Plate 6.6c).	DAGBS Dynamic recrystallisation of qtz+carb.	Geometric softening (White and White 1983). Infiltrative metasomatism (Beach and Fyfe 1972) from syn to post tectonic fluids. Demonstrate focused fluid flow on microscale within shear zones, e.g. McCaig 1987. Syn to post tectonic influx of carbonate fluids.
Ser-qtz-chl-ank phyllonite	Central part of DDH ZR1. Fig. 6.1.	High	Moderately differentiated tectonite, local pervasive carbonate (same textures as) chl-qtz-ank phyllonites. Well developed shear bands, often transposing original fabric (Plate 6.7a). No difference in grain size between 'old' and 'new' fabrics. Transposition fabric also has small crenulations. Crenulation cleavage observed in one section (Plate 6.7b) related to high angle shears cutting main fabric. Fibrous ser overgrowths in pressure shadows on large porphyroclasts. Late unoriented ser euhedra.	DAGBS Incongruent pressure solution.	K/Na overprint.

Plate 5.1 Deformation of simatic rocks.

- 5.1a** Fibrous amphibole (long pale coloured fibres - arrowed) being retrogressed to chlorite and reoriented in a shear band geometry. Matrix of fine grained quartz-carbonate and fractured albite. Sample DDH ZR1 310.04. Field of View (FoV) 0.9mm, XPL.
- 5.1b** Folded, fractured and strained amphibole crystal lying in fine grained matrix of foliated chlorite axial planar to fold. Hinge of fold marked by sphene. DDH ZR3 60.63, FoV 2mm, XPL.
- 5.1c** Intensely fractured albite lying in foliated chlorite matrix. Oriented chlorite fibres between boudins are also present. DDH ZR1 191.11, FoV 6mm, PPL.
- 5.1d** SEM micrograph showing fracturing of feldspar and also shear band geometry to the fabric defined by chlorite. Oriented chlorite fibres clearly visible in pressure shadows and between boudins. Euhedral ankerite is also present. Some replacement of albite by chlorite (controlled by fractures) has occurred (arrowed). DDH ZR1 304.59, FoV 200 microns.

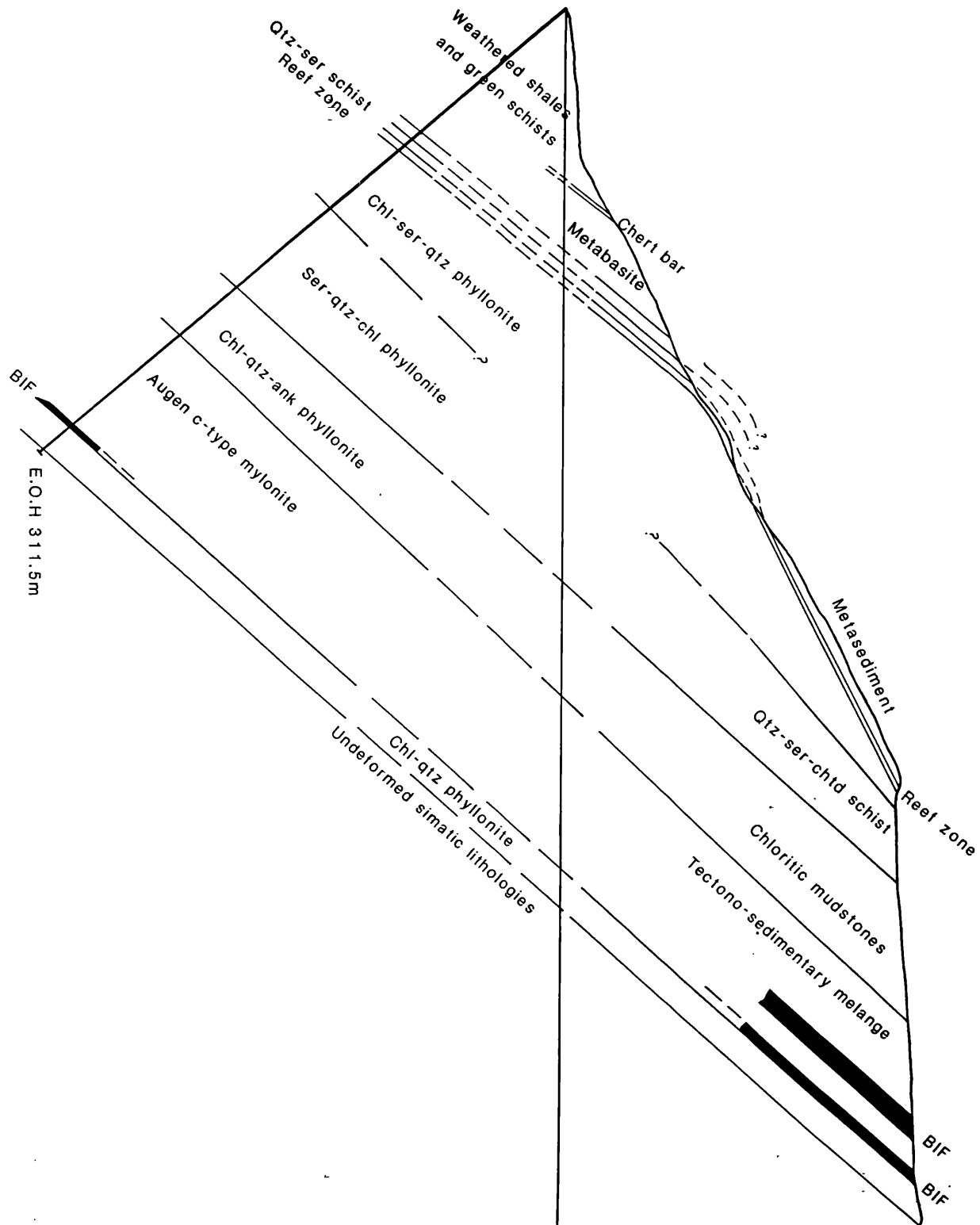


Fig. 5.2 Cross-section along line of DDH ZR 1 looking SE

Scale 1:2000

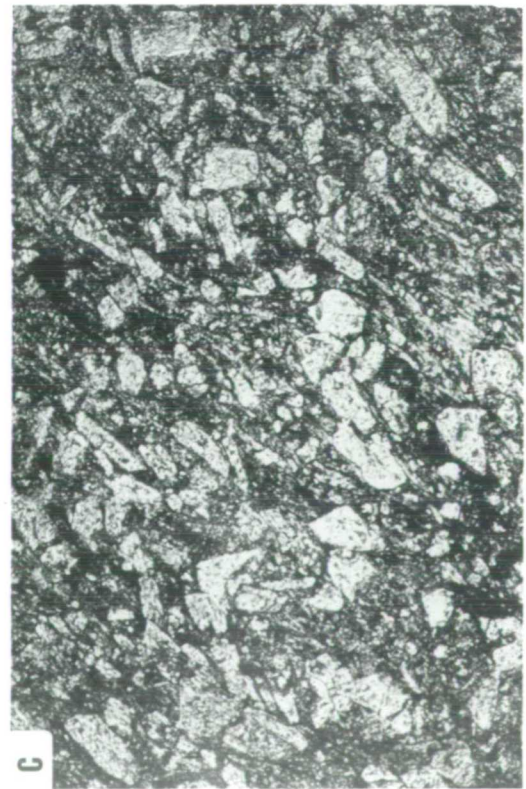
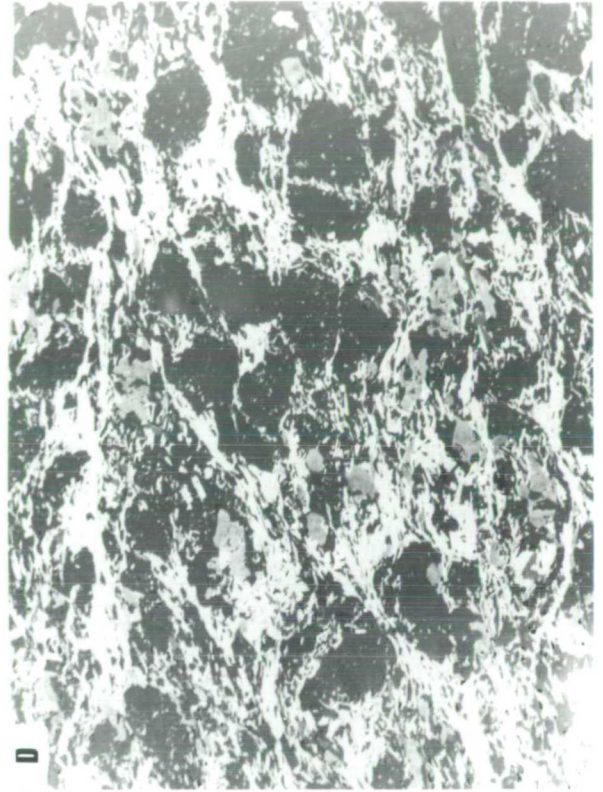


Plate 5.2 Microstructures in augen c-type mylonite.

- 5.2a.** Hand specimen showing ductile deformation of clasts in augen mylonite. Note higher degree of plastic deformation compared to surface exposures (cf. Plate 4.1 and 4.2). DDH ZR1 286.30.
- 5.2b.** Remnant chert clast with a fine grained recrystallised tail. Coarse grained centre of clast contains siderite and quartz and has probably annealed. Chlorite defines the dark foliation planes. DDH ZR1 238.52, FoV 2mm, XPL.
- 5.2c.** Close up of recrystallised tail in 5.2b showing numerous fine grained oriented chlorite laths lying along fine grained, elongate quartz grains. Coarse euhedral siderite is visible. Note how the chlorite grains merge with the main foliation domain. DDH ZR1 238.52, FoV 1.5mm, PPL.
- 5.2d.** Remnant plastic shear zone delineated by fine grained quartz (with small oriented chlorite grains along their boundaries) that aided grain size reduction of a large chert clast. DDH ZR1 238.68, FoV 2.5mm, XPL.
- 5.2e.** Coarse grained, twinned siderite mantled by finer grained recrystallised or possibly cataclastically derived siderite. DDH ZR1 238.68, FoV 2.5mm, XPL.
- 5.2f.** Cross-cutting siderite vein in remnant BIF block at base of augen mylonite. Iron rich chlorite porphyroblasts are also present. DDH ZR1 290.84, FoV 3mm, XPL.

Operative deformation mechanism

Protolithology:		BIF	Cherts	Sediments	Simatic rocks	
Main control on fault rock development	Development stages					
	Pressure sensitive	1.	Cataclasis and fluidization Carbonate influx	Cataclasis and ductile flow Soft sediment deformation Particulate flow and grain breakage	Microcracking of feldspar Retrogression of amphibole to chlorite	
		2.	Cataclastic flow causing rigid body rotation (C-type mylonite)	cyclical devt ↑ ↓	Diffusive mass transport	Reorientation of chlorite, continued microcracking and change of feldspar
		3.	Pressure solution, increasing dynamic recrystn (augen mylonite)			Silicification and/or recrystn Increasing dynamic recrystn
	Temp., grain size, rate sensitive	4.	Continued growth of phyllosilicates along grain boundaries, aided by crack propagation, and in larger domains leads to increasing dominance of diffusion (and minor plastic) accommodated grain boundary sliding			
5.		Chlorite - quartz phyllonites	Continued syn-tectonic fluid/rock interaction and metamorphic reaction		Sericite - quartz phyllonites	

Fig. 5.4 Progressive development of phyllonites in the SDSZ

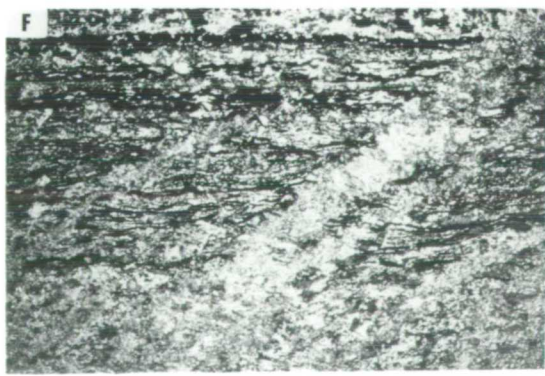
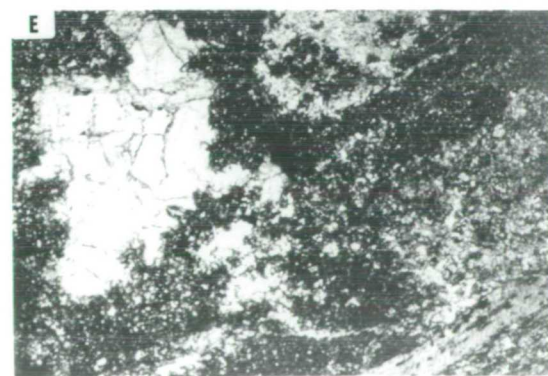
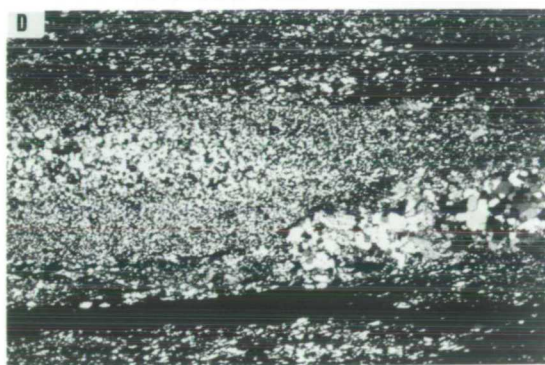
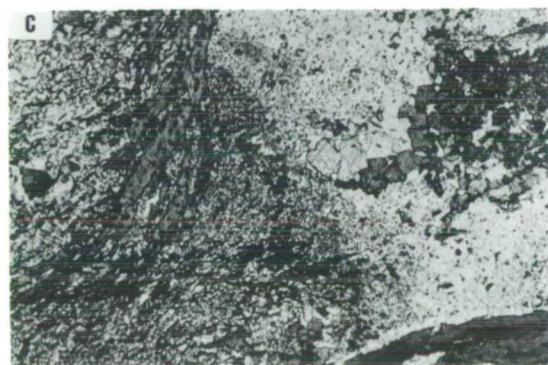
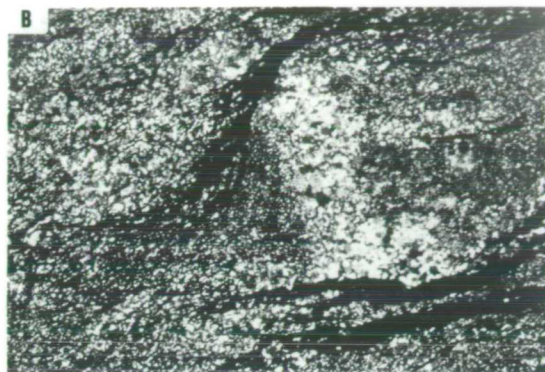


Plate 5.3 Textural variation and foliation development in augen mylonite.

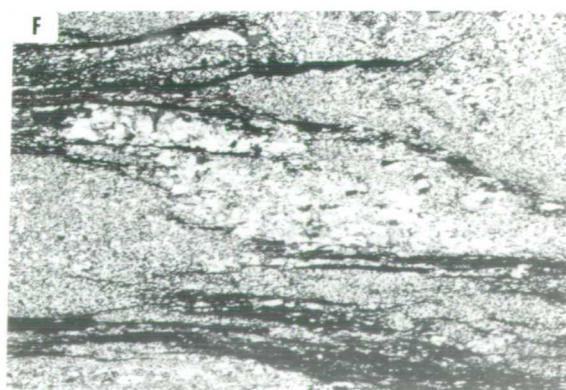
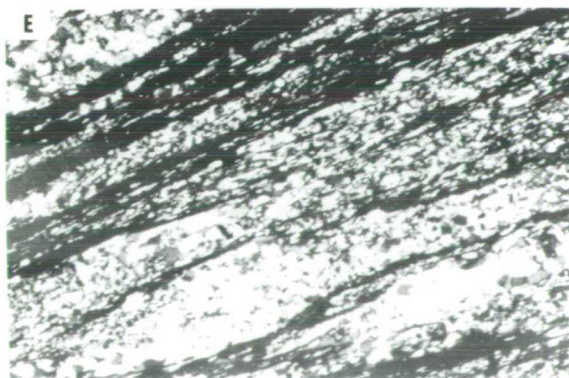
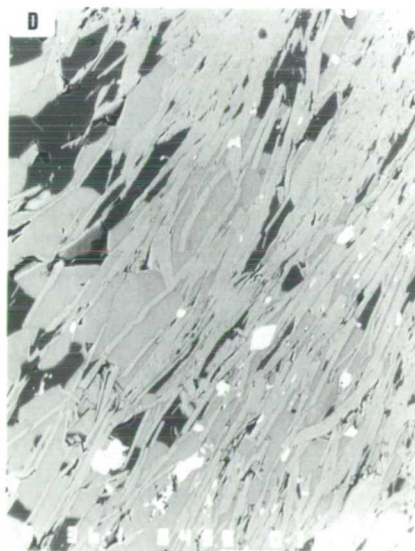
5.3a, b and c. Three examples showing textural variation in augen mylonite. Note how the foliation development varies depending on the clast distribution and amount of grain size reduction. Shear bands are locally better developed than the main foliation. 5.1b and c show a variation from shorter, rough cleavage domains to longer anastomosing domains. Note how the cleavage domains terminate indicating they may represent elliptical planes in three dimensions. Light phase – quartz, grey phase – predominantly carbonate, dark phase – chlorite.

- a. DDH ZR1 238.52, FoV 2cm, PPL.
- b. DDH ZR1 231.68, FoV 2cm, XPL.
- c. DDH ZR1 239.68, FoV 2cm, PPL.

5.3d. SEM micrograph of gradational contact between quartz (dark grey) and carbonate (light grey) rich microlithon and chlorite foliation domain. Note the relative increase in aspect ratio and gradual disappearance of quartz and carbonate probably due to dissolution into an intragranular fluid phase. DDH ZR1 239.68, scale bar 100 microns.

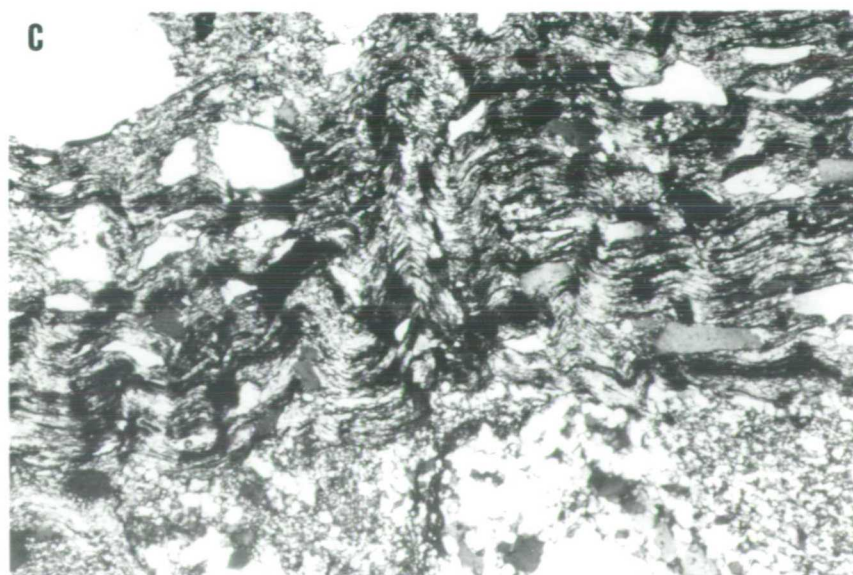
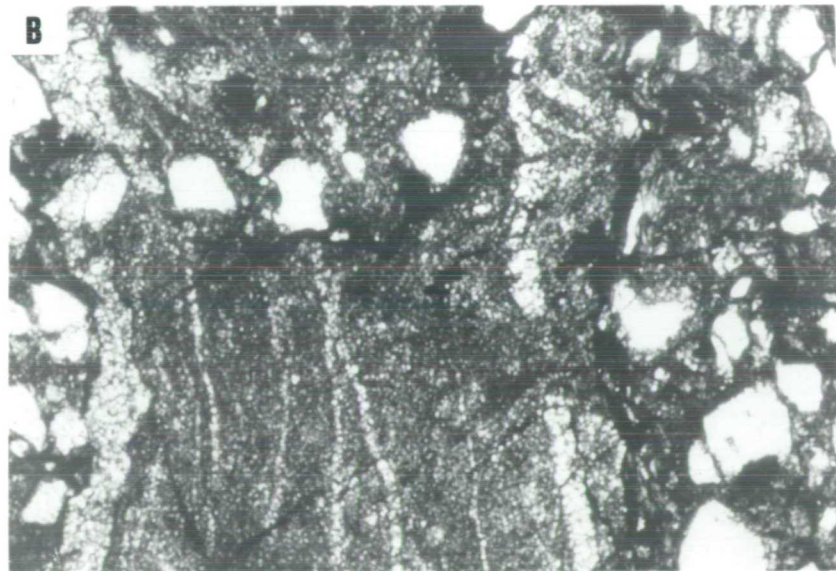
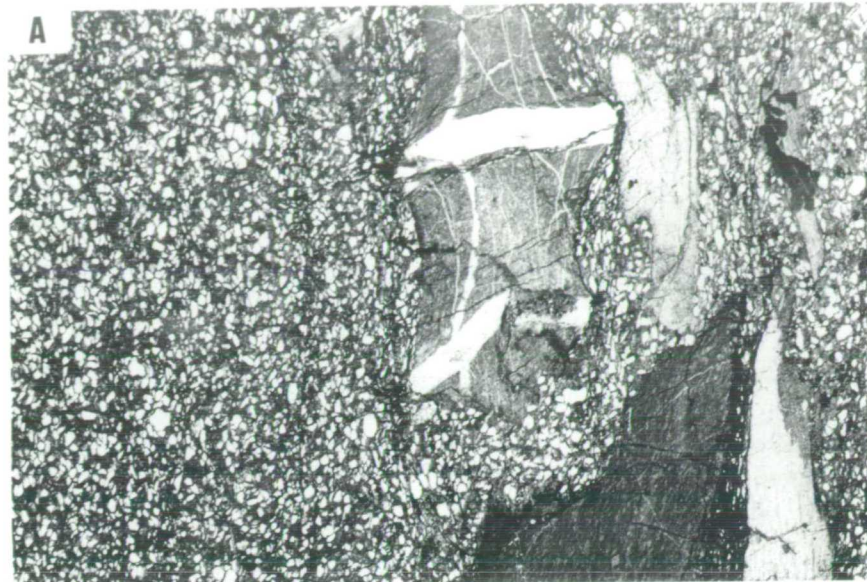
5.3e. View of chlorite domains and microlithons illustrating general sequence of fabric development. Note the relative elongation of some of the quartz and carbonate grains as the chlorite domains are approached and also truncation of coarser, more equidimensional grains against the domains. DDH ZR1 239.68, FoV 2mm, XPL.

5.3f. ‘Dilational jog’ of siderite and quartz marking the termination of two separate chlorite domains (marked). DDH ZR1 239.68, FoV 1mm, XPL.



**Plate 5.4 Microstructural evolution of sediments and breccia zones
(i).**

- 5.4a. Large oriented chert clasts lying in a coarse grained sandstone matrix showing well developed domainal pressure solution seams. Note intense solution seams between large clasts and also the resistant quartz vein within one chert clast. Sample SD 55, from locality R, Fig. 4.1, FoV 1.5cm, PPL.**
- 5.4b. View of end of chert clast showing similarity in grain size between clast and matrix. Oriented phyllosilicates pass from the matrix up to 500 microns into the chert. The fine grained chert is being tectonically reworked into the matrix, probably by DAGBS. SD 56, FoV 2mm, PPL.**
- 5.4c. View of pressure solution seam showing high angle crenulations and elongate nature of quartz clasts with cusped boundaries. Note how the wavelength of the crenulations is strongly controlled by the clast size and the more resistant quartz vein. Narrower pressure solution seams are found along most of the crenulation hinges. Sample SD 56, FoV 2.5mm, PPL.**



**Plate 5.5 Microstructural evolution of sediments and breccia zones
(ii).**

- 5.5a. Sericite domains and oriented fibrous sericite overgrowths on remnant quartz clasts in deformed sediment. DDH ZR1 84.88, FoV 1.5mm, XPL.**
- 5.5b. Hand specimen showing extreme flattening of clasts as a differentiated quartz-sericite domainal fabric develops. Sample SD 59, locality R, Fig. 4.1.**
- 5.5c. Thin section showing reworking of remnant clast into the matrix. Note the elongate quartz grains within sericite domains indicative of intense dissolution and also fine grained quartz rich microlithons that represent very deformed chert clasts. Sample SD 59, FoV 9mm, PPL.**
- 5.5d. Granoblastic quartz-sericite schist. Note the euhedral unstrained quartz grains and triple point junctions suggestive of annealing. The matrix is well foliated evidence and is at a higher degree of finite strain than other samples on this plate. DDH ZR3 83.21, FoV 8mm, XPL.**

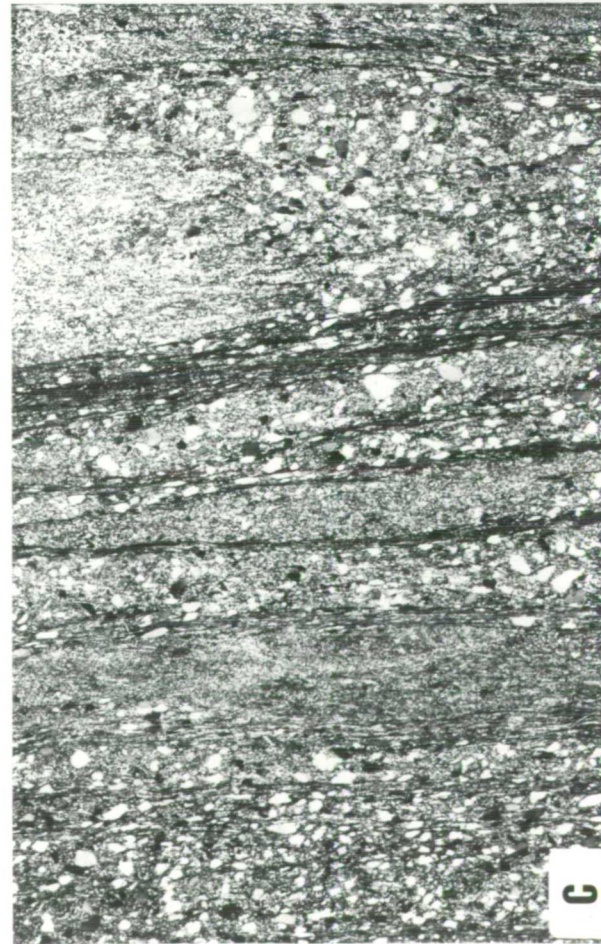
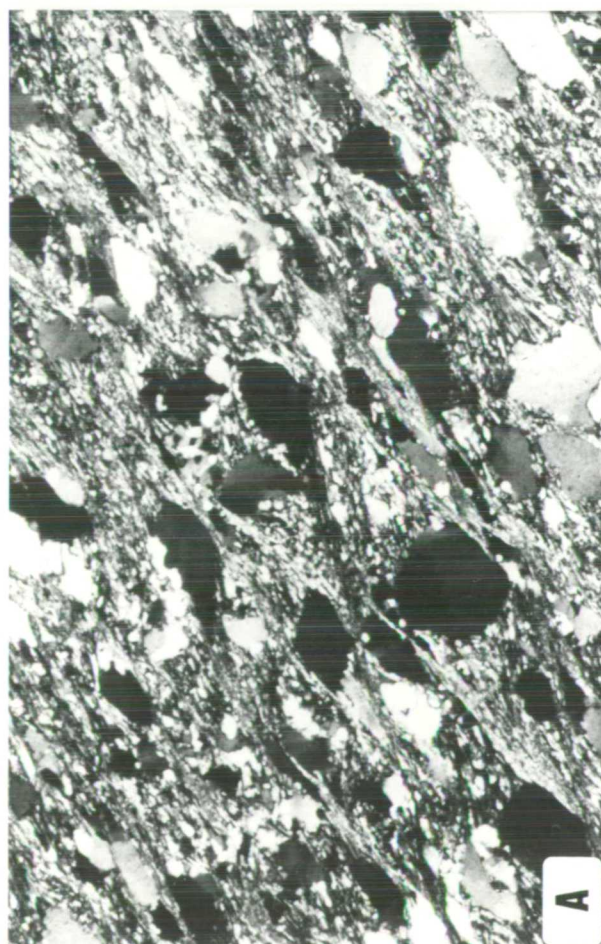
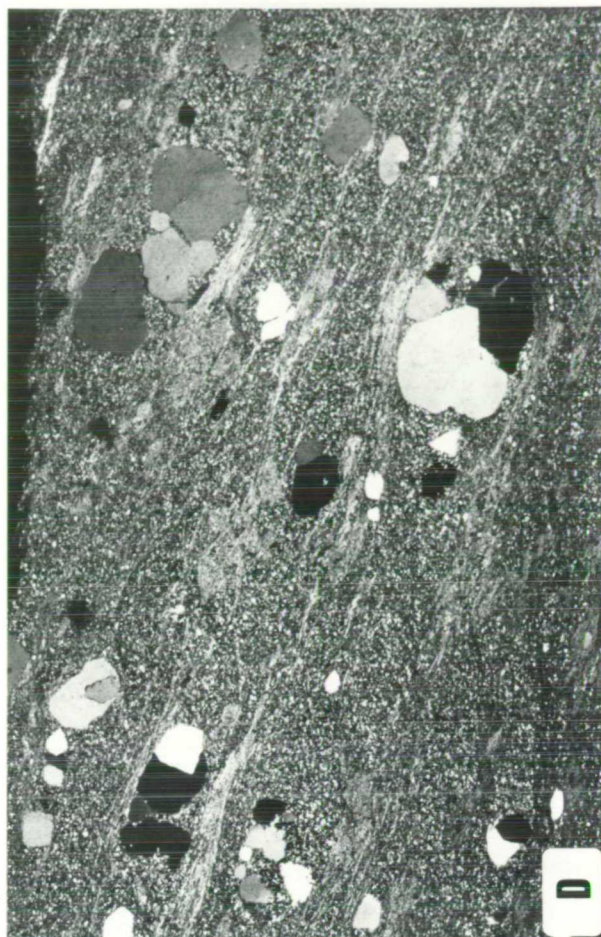
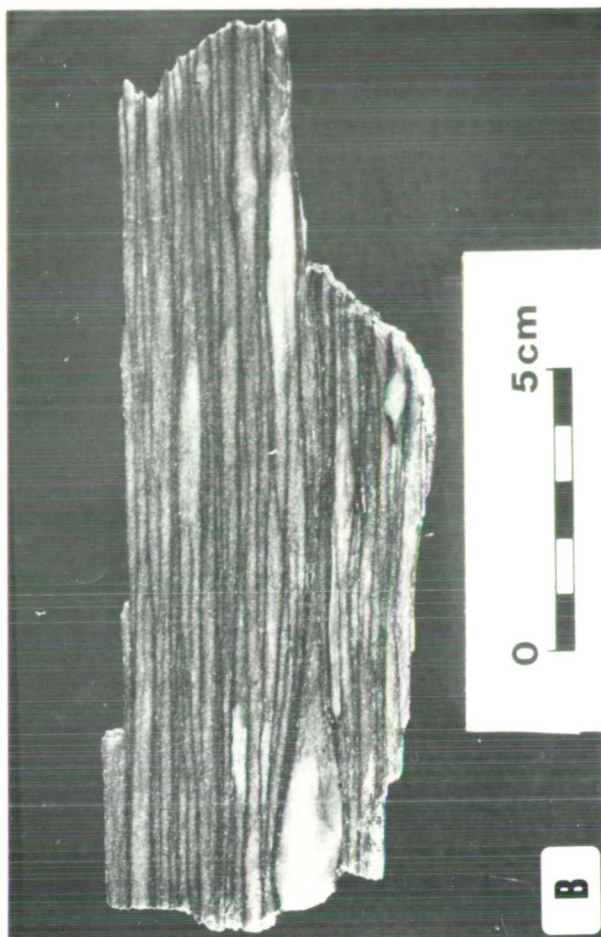


Plate 5.6 Textures in chlorite-quartz-ankerite phyllonites.

- 5.6a. Anastomosing chlorite-rich foliation domains (light grey) that surround and isolate fine grained quartz (and in this case fairly chlorite rich) dominated microlithons (darker grey). DDH ZR1 202.52, FoV 10mm, XPL.**
- 5.6b. Chlorite phyllonite cut by later D2 sericite micas in shear bands. Some replacement of chlorite along the original chlorite layers is also seen. DDH ZR1 210.1 FoV 10mm, XPL.**
- 5.6c. Ankerite-bearing phyllonite with a rough, slightly anastomosing foliation. The carbonate is relatively undeformed. DDH ZR1 231.08, FoV 6mm, XPL.**
- 5.6d. Quartz-carbonate vein that cross-cuts the D2 fabric but is folded axial planar to, and transposed by the same fabric. The vein is therefore syn-tectonic. Grain size reduction of the vein by crystal plastic processes has been important. DDH ZR1 184.00, FoV 6mm, XPL.**

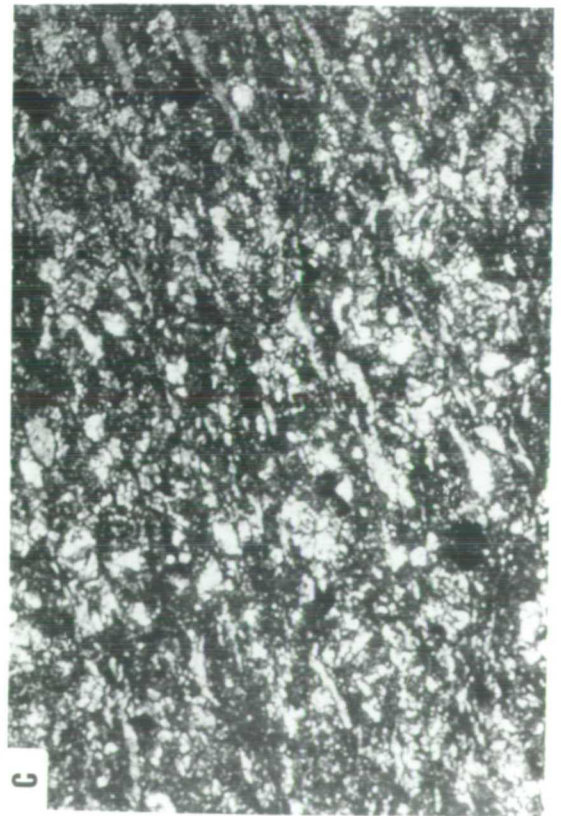
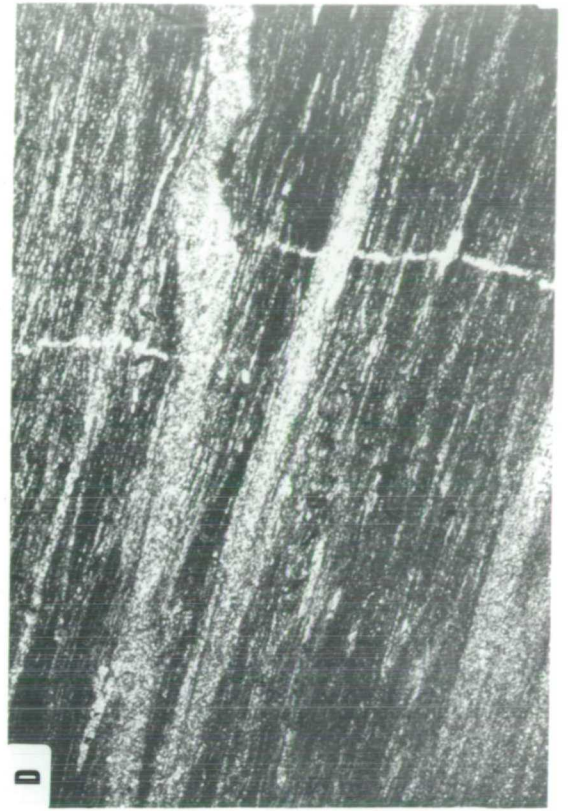


Plate 5.7 Textures in sericite-quartz +/- chlorite phyllonites.

- 5.7a.** Overview of transposition of early D2 quartz-sericite layering by an increasingly differentiated, late D2 quartz-sericite fabric. DDH ZR1 84.88, FoV 1.5cm, XPL.
- 5.7b.** Development of strongly crenulated D2 fabric by narrow through going sericite shear zones. Such shears are extremely localised in core from the ZR prospect. It is not clear if they represent late D2 structures or a separate 'D3' event. DDH ZR2 85.00, FoV 10mm, XPL.
- 5.7c.** Overview of very sharp contact between deformed simatic rock with fractured feldspars in an oriented chlorite matrix, and a higher strain sericite-quartz-chlorite phyllonite. DDH ZR1 191.11, FoV 1.5cm, PPL.
- 5.7d.** Close up of contact in 5.7c showing break down of feldspar to quartz which undergoes dynamic recrystallisation (arrowed). Such textures suggest the sericite-rich phyllonite does not represent a deformed sediment. DDH ZR1 191.11, FoV 4mm, PPL.

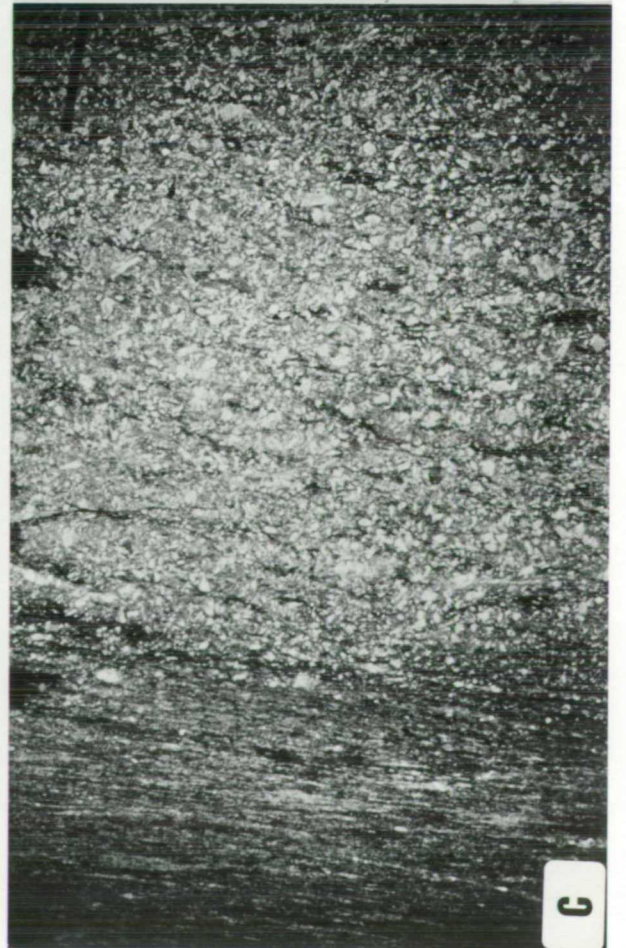
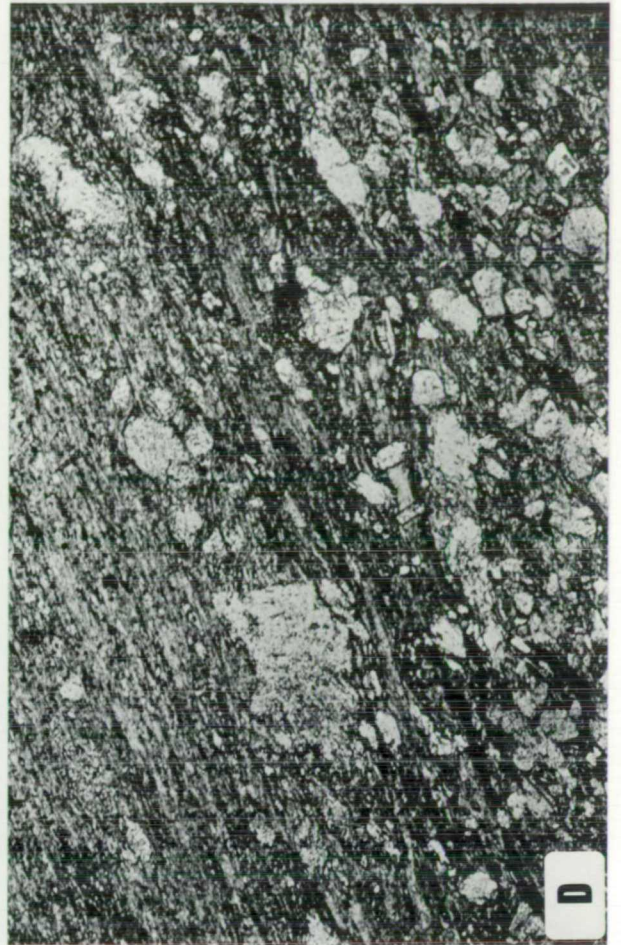
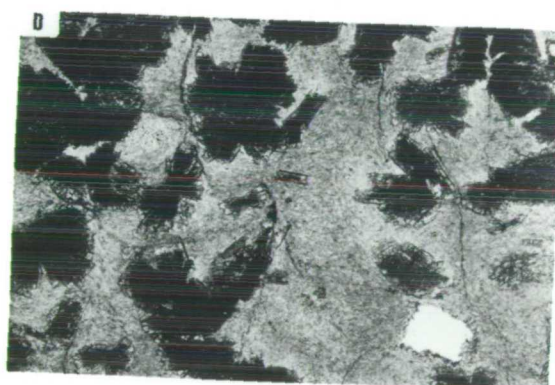
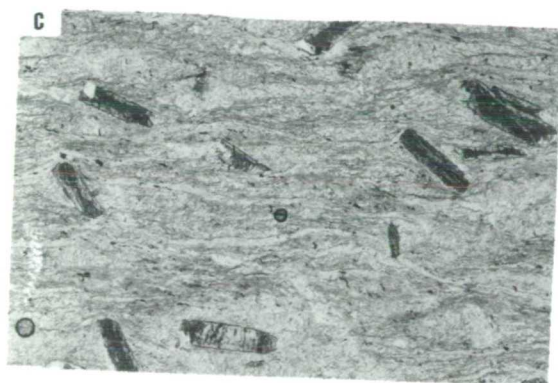
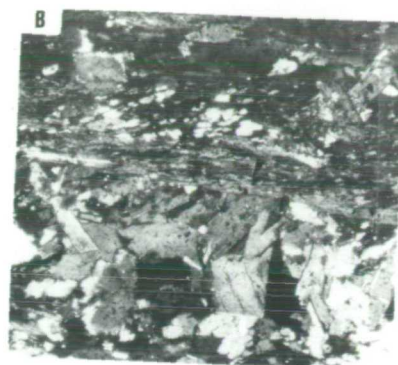
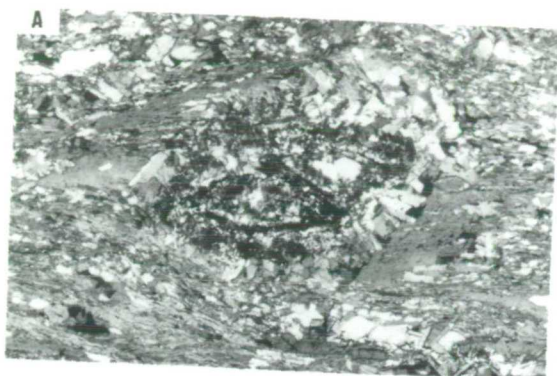


Plate 5.8 Porphyroblasts in the SDSZ.

- 5.8a.** Dark grey euhedral laths of clinozoisite radially disposed around an orbicular structure with a centre of quartz and carbonate. It is unlikely these blasts have rotated. DDH ZR1 239.68, FoV 4mm, XPL.
- 5.8b.** Truncation of clinozoisites and quartz against a chlorite domain indicating there was some post-porphyroblast but syn-D2 flattening. DDH ZR1 239.68, FoV 1.5mm, XPL.
- 5.8c.** Euhedral laths (type 1b) of chloritoid in strongly differentiated and crenulated quartz-sericite-phyllosilicate. Note that S_i bears little or no relationship to S_e and is interpreted as indicating that chloritoid growth was syn-D2 but predating this stage of D2 fabric development. DDH ZR1 161.11, FoV 6mm, PPL.
- 5.8d.** Type 2 chloritoids showing stellate growth and forming equidimensional rosettes. Note the strong association of their growth in microlithons between crenulation cleavage and the sub-parallelism of S_i and S_e in the microlithons. Sample SD 65, 75m north of locality R, Fig. 4.1, FoV 6mm, PPL.
- 5.8e.** Closely packed bow-ties of type 3 chloritoid lying in matrix of crenulated tourmaline and phyllosilicates. Note how the crenulations lie outside the bow-ties and the continuity of S_i and S_e and the parallelism of S_i between bow ties. Sample SD 64, 20m north of locality R, Fig. 4.1, FoV 4mm, PPL.
- 5.8f.** Contact between quartz-sericite-chloritoid phyllosilicate (right) and chloritic phyllosilicate (left). Note how the chlorite is blanketing out the original textures leaving only remnants of chloritoid (arrowed). DDH ZR2 112.54, FoV 15mm, PPL.



CHAPTER 6

STRUCTURAL GEOLOGY OF THE SNYMANSDRIFT SHEAR ZONE

6.1 STRUCTURAL SUB-AREAS WITHIN THE SDSZ

6.1.1 Overview

In order to analyse the structural data collected, the SDSZ has been divided into five structural sub-zones (numbered 1 to 5 on Fig. 4.1) whose major features are described and interpreted in Table 6.1. The sub-areas were chosen on the basis of similar structural and metamorphic style, qualitatively defined in the field. The degree of finite strain and metamorphic overprint is lowest in sub-area 1 but gets progressively greater in sub-areas 5, 2, 3 and 4 (in that order). Fold and lineation terminology is after Fleuty (1964).

Lithological contacts within the SDSZ are variably strained, essentially planar to curvilinear features which can be traced over most of the exposed strike length. Within the sequence, other narrow high strain zones have been identified, usually marked by chert bars or deformed chert breccias (chapter 4.8, Fig. 4.1).

The data in Table 6.1 can be summarized as follows;

- 1) All structures seen in the zone are interpreted as having formed during progressive D2 deformation with early structures being progressively overprinted by fabric development.
- 2) Displacement along minor high strain zones and the SDSZ as a whole, decreases to the west.
- 3) Early syn-sedimentary faults (defined by chert breccias) are locally folded and generally strongly overprinted by fabric development.
- 4) Folds are axial planar to the D2 fabric except in sub-area 2 where the fabric may transect the fold slightly. This is further evidence that the majority of fabric overprint occurred in the later stages of D2.
- 5) Deformation proceeded during compression and minor high strain zones are likely to be thrusts (see below).
- 6) The SDSZ is at its thickest in sub-areas 3 and 4 giving it an overall pod-like or lenticular morphology. This is the result of a central sequence of folded and faulted mudstones

(and minor sandstones and conglomerates) and tectonic slices of simatic rocks (Figs. 1.3, 4.1) introduced during early D2 faulting concentrated between decollement horizons immediately above the melange and immediately below lithofacies A of the Uitkyk Formation.

6.1.2 Evidence for imbricate thrusting within the SDSZ

Determination of the sense of shear across the SDSZ and narrower high strain zones within it is difficult but several lines of evidence suggest that the deformation was essentially thrust-related.

1) At Zandrivier excellent examples of reverse drag into a bedding parallel fault are common.

2) Most folds but especially the upright, sub-horizontal anticline/syncline pair on Mt. Mare has a reverse (NW directed) sense of vergence. Asymmetry of other folds also generally gives an NW vergence.

3) There has been structural thickening (i.e. repetition) within the mudstone unit due to the introduction of slices of older simatic rock. The number of these slices is not known (chapter 4.2 and 4.6) but at least two major units can be inferred from borehole data and surface observations.

4) The development of high strain zones sub-parallel to bedding and slides associated with folding is concurrent with compressive deformation. The faults are thus more likely to be thrusts than normal faults. The possibility that some of the faults may represent inverted normal faults can not be ruled out at this stage.

The development of second order high strain zones and folds within the SDSZ is essentially confined to the lithological package which lies between the melange and lithofacies A of the Uitkyk Formation. Within the package the high strain zones often 'meet' at branch points to form an anastomosing linked fault system. At the edges of the SDSZ both to the west and the east, these faults die out, either linking with sedimentary contacts or dying out completely in slides associated with syn-sedimentary folds (Table 6 1).

Therefore, the high strain zones and intervening lithologies are believed to form a flattened (ie. deformed), syn sedimentary duplex in which imbrication of thrust slices of mudstone/sandstone and simatic rock has occurred. The duplex is imperfect, or 'leaky' (Hyett 1985) because it dies out laterally into sediments without linkage of all the associated faults. Imbrication was occurring only within the chloritic mudstone-sandstones and was

probably related to a lower decollement (sole) at the melange/chloritic mudstone-sandstone unit and an upper decollement (roof) at the mudstone-sandstone/Lithofacies A contact (Table 6.1). The exact timing sequence of individual fault slice movement within the duplex is not resolvable but is assumed to be a foreland-propagating, northwest verging sequence. Later D2, but probably still syn-sedimentary, faults deformed this upper decollement. They were rooted in the developing duplex and cut up-section into the overlying sandstones. This litho-tectonic package was continually steepened, flattened and metamorphosed as deformation progressed. Thus, (at least) the upper parts of the SDSZ represent a deformed and imbricated piggy-back sedimentary basin carried above a propagating fault.

Two outstanding problems need to be resolved; what is the sense of shear in the tectono-sedimentary melange and how does it relate to the development of the overlying duplex?

6.2 CLAST DEFORMATION AND ORIENTATION IN THE TECTONO-SEDIMENTARY MELANGE

6.2.1 Introduction

From field observations (chapter 4.4) it was inferred that the c-type mylonites found towards the base of the melange had been derived from the brecciation of large BIF blocks during progressive D2 deformation. A corollary of this inference was that the clasts in the cataclasites, which initially had variable orientations, must have undergone rigid body rotation to become well oriented in the c-type mylonites. The rotations would have been caused by torque applied to the clasts due to ductile flow in the matrix.

Most clasts exhibit significant degrees of internal deformation of which two distinct styles have been observed: *in situ* cataclasis; and subordinate plastic deformation. The latter has lead to the development of small (1 to 10mm long) asymmetric 'tails' of fine grained quartz flanking some of the clasts. Such features suggest the clasts may have behaved in an analagous fashion to porphyroclasts in mylonites (eg Passchier and Simpson 1986) and thus their potential as kinematic indicators was investigated. The following observations were recorded: clast size, shape, orientation and style of deformation.

6.2.2 Clast deformation

Two end-member styles of brittle deformation of the clasts are seen in the melange

(Fig. 6.3). First, planar to curvilinear spaced fractures which cause the clasts to break up into smaller but still quite euhedral blocks and secondly, more distributive disaggregation forming large numbers of variably shaped fragments (Fig. 6.3a,b,c,d). Both tensile and shear fractures are present but there is no consistent asymmetry to their development in any plane of the strain ellipsoid. Indeed, many of the fractures proved ambiguous (Fig. 6.3d) as to their origin and it was concluded that fracturing of fragments is a very unreliable kinematic indicator in the melange.

Many of the more elongate clasts have asymmetric tails of fine grained quartz attached to them (Plate 4.2a and b). Thin section work shows these tails consist of fine grained quartz with oriented chlorite laths along their grain boundaries. Although there is some difference in grain size between the centres of the chert clasts and these tails, there is no preferred optical orientation to the quartz. In many places the tails have apparently developed at the expense of sharper edges on the clasts (Fig. 6.3e). The tails are also preferentially developed at clast impingement points. Overall it appears the tails are pressure solution features and not the result of crystal plastic recrystallisation. De Wit (1985) also assigned the tail formation to a solution-reprecipitation mechanism.

Although asymmetric pressure solution tails (and shadows) have been quoted as kinematic indicators (Simpson and Schmid 1983), the simple use of the sense of asymmetry to determine the sense of shear has been shown to be unreliable (Takagi and Ito 1988). Opposing tail asymmetries are very common in the Snymansdrift melange and this variability is believed to be directly related to the orientation of the clasts with respect to the tectonic framework.

6.2.3 Clast orientation

The following empirical observations can be made concerning clast orientation in the melange and exemplified in plates 4.1 and 4.2. For descriptive purposes the longest, intermediate and shortest geometric axes of the clasts are designated a, b and c respectively (Fig. 6.4a).

In the c type mylonites a well developed lineation is present in the foliation plane, plunging down dip. If it is assumed the foliation plane represents the XY plane of the finite strain ellipsoid then the lineation defines the principal stretching direction (X) (Ramsey and Huber 1983). Judging from the average clast size variation in each of the three planes of the ellipsoid, the lineation is defined by the preferred orientation of the a axes of the clasts which have come to be aligned sub-parallel to one another in the XY plane. The b axes also seem to preferentially lie in this plane.

In the XZ plane of the strain ellipsoid, although evidence for preferred orientation is seen, there are significant divergences from uniform behaviour for all clasts. The XZ plane generally contains the a and c axes of most clasts. The most elongate clasts were found to lie with their long axes sub-parallel to, or at a slight angle to, the foliation. Of those clasts with axes oriented at a slight angle to the foliation, the majority of them dip in the same direction as the foliation (ie SE) but at a slightly steeper angle (Fig. 6.3f). Less elongate clasts show much more variation in dip. Their long axes may dip at steeper angles, but still in the same direction as the foliation, or they may dip in the opposite direction to the foliation (ie NW). Pressure solution tails are developed on many of these clasts. However, it is *the orientation of the clast which apparently controls the asymmetry of the tails*.

On clasts dipping in the same direction as the foliation the tails exhibit a dextral 'stair-stepping' (Passchier and Simpson 1986) asymmetry, whereas on clasts dipping in the opposite direction to the foliation exhibit a sinistral stair-stepping asymmetry (Fig. 6.3f and g). In the YZ plane of the strain ellipsoid, which generally contains the b and c axes of clasts (plate 4.2), similar orientation patterns are present except that there are a larger proportion of clasts showing opposing dips and tail asymmetries.

The following conclusions are drawn from observations on clast deformation and orientation in the melange;

1. The geometry of the clast appears to have a very strong control on its rotation through the matrix. The preferred orientation of elongate clasts and the marked lack of such clasts with their long axes oriented at a high angle to the foliation plane, suggests these clasts have rotated to the preferred orientation position and remained there as stable objects. More equidimensional clasts show a lesser degree of preferred orientation, their long axes showing a more marked variation in orientation with respect to the foliation.

2. Both cataclastic and diffusive mass transfer processes have resulted in many of the clasts changing shape during rotation. Cataclasis generally produced new clasts that were more equidimensional than their parent clast, whereas the development of pressure solution tails resulted in clasts becoming more elongate during deformation.

3. Both fracturing of clasts and pressure solution tail asymmetry are very unreliable kinematic indicators because of ambiguity involved in interpreting their origin. Specifically, the opposing dips and associated stair-stepping tail asymmetries could be explained by one of two models; the clasts were rotating in the same direction but were 'captured' at different stages of rotation, or the clasts were rotating in opposite directions because of a component of flattening across the foliation.

4. From 3, it is intuitively obvious that if there was a component of pure shear across the melange during deformation it could have a strong influence on the direction of *rotation* of the clasts possibly causing some clasts to rotate in an opposite sense to the overall sense of shear. This implies that the orientation of the clast with respect to the walls of the deforming zone is a further strong control on the direction of rotation. It is not possible from simple observations alone to distinguish between rotations induced by a component of pure shear across the zone or those induced by flattening by the incremental strain axes (Passchier 1988).

Similar conclusions were reached independently on a theoretical basis by Lamb (1987b) and Passchier (1987) who have studied rigid body rotation in deforming masses in some detail. Passchier (1987) has outlined a method to determine the ratio of pure to simple shear in mylonite zones based on such rotations. This analysis has been extended to the melange.

6.2.4 The Kinematic Vorticity Number (W_k) and rigid body rotation

Although natural shear zones are often assumed to have formed under conditions of simple shear alone (non-spinning, non-coaxial rotational flow), in fact, only shear zones with perfectly undeformed wall rocks and absence of volume change can satisfy this assumption. Natural shear zones formed by simple shear alone may in fact be quite rare (Ramsey and Huber 1987, Passchier 1988). The effects on fabric development of simultaneous pure shear leading to flattening or extension across a shear zone have been considered in some detail in the past (Ramsey 1963, Ghosh and Ramberg, 1976) but only recently has quantitative measurement been attempted.

Means *et al.* (1980) have defined (after Truesdell 1954), a kinematic vorticity number as a measure of the degree of non-coaxiality at an instant in the deformation history. They define the term as W_k . W_k is a function of the vorticity of the flow and the instantaneous strain rates along the incremental stretching axes. It is a useful measure of non-coaxiality where the fabric is said to be "steady-state". Means *et al.* (1980) show that for isochoric plane strain, when $W_k = 0$, deformation is one of pure shear, and when $W_k = 1$, simple shear. Values in between 0 and 1 are commonly assumed in structural geology. Values of $W_k > 1$ are possible and correspond to the "pulsating strain histories" discussed by Ramberg (1955). Thus, if W_k can be deduced from geological observations in shear zones then a measure of the 'ratio' of pure shear to simple shear that was operative at the moment deformation ceased in the zone, can be determined. Rigid body rotation, such as inferred in the Snyman'sdrift melange provides one method of estimating W_k .

Passchier (1987) and Lamb (1987) have investigated rigid body rotations in flowing masses on a mathematical basis and shown that it involves complex tensor operations. The mathematics of the system can be simplified if the following assumptions are made; first, that the flow during deformation is non-spinning and isochoric plane strain, and secondly, the rigid objects are isolated, axially symmetric ellipsoids which do not change shape during deformation. If these conditions are met it can be shown that the instantaneous rate of rotation (R) of an ellipsoid in a deforming zone is a function only of

- a) its aspect ratio (k),
- b) the orientation of the long axis of the ellipsoid with respect to the boundaries of the deforming zone (ϕ), and
- c) the kinematic vorticity number (W_k) (Fig. 6.4a).

Fig. 6.4b shows graphs of R against ϕ for different values of k (0 (a material line), 0.25, 0.5, 0.75, 1 (a disc)) and W_k taken from Lamb (1987b). Lamb's analysis was initially developed from the consideration of crustal rotations for palaeomagnetic studies and use a variable Θ instead of W_k . The analysis can be extended to the shear zone problems discussed above; and Lamb's graphical representation is more vivid with respect to expressing the complexities of rigid body rotation than the stereographic representations employed by Passchier (1987). Θ , which can be considered as an angular measure of vorticity, is the angle between the relative velocity vector and the margin of the deforming zone and is thus 90° for pure shear and 0° for pure shear. Negative Θ implies dextral shear and compression. Fig. 6.4b expresses virtually all aspects of rigid body rotation which can be simplified as follows;

1) in true simple shear ($\Theta = 0^\circ$, $W_k = 1$) all objects regardless of ϕ or k , will permanently rotate during deformation and they will rotate with the same sense of rotation as overall sense of shear of the deformation zone. Disks will rotate at a *constant* rate but all other objects with $k < 1$ will rotate at varying rates depending on their orientation (ϕ). Objects will rotate fastest when their long axes are oriented at a high angle to the walls of the deforming zone and slowest when the same axes lie parallel to the walls.

2) In true pure shear, circular disks will not rotate and are thus said to be *stable*. Other objects with $k < 1$ will also be stable if their long axis is either exactly parallel to, or exactly normal to, the deformation zone boundaries. Objects with $k < 1$ at other orientations will rotate (at varying rates) towards these positions of stability but their sense of rotation depends dramatically on ϕ . Both sinistral and dextral rotations would be expected with an initial, randomly oriented population of objects.

3) At intermediate values of Θ (or W_k) then, it is clear that depending on ϕ and k a certain number of objects will rotate with the opposite sense of rotation to the overall sense of shear of the deforming zone. This illustrates the danger of using rigid bodies as sense of

shear indicators. Exactly which clasts will rotate with opposite vorticity is strongly controlled by their aspect ratio, k . Fig. 6.4b shows for a given Θ , objects with k greater than a certain critical value (called k_{crit}) will rotate (at varying rates) with the same sense of vorticity as the surrounding matrix. Objects with $k < k_{crit}$ would theoretically change their rotation sense during deformation. However, in order to change their sense, the rate of rotation must decrease to zero and at which point no torque is applied to the body; it is thus stable.

There are in fact two orientations of such stationary states for each object but the orientation furthest from $\phi = 0$ is generally a quasi-stable state because any small deviation from this position induces a rapid rotation away from that position (Lamb 1987b). The orientation closest to $\phi = 0$ is a true stable state. There is a tendency for objects to rotate into this position and then remain stationary. At this position the long axis of the clast makes an angle to the foliation that is of the same sign as the sense of vorticity. Therefore, in general for a given value of Θ (or W_k), there is one orientation of an ellipsoid with $k < k_{crit}$ at which it is stable and does not undergo further rotation. Thus if k_{crit} can be defined for a population of clasts with variable aspect ratios in a deforming zone it should be possible to determine Θ or W_k .

These theoretical considerations show that the rotation of rigid bodies in a flowing mass is surprisingly complex, a feature not considered in some articles advocating their use as kinematic indicators (eg Simpson and Schmid 1983, White *et al.* 1986).

6.2.5 Methods for determining W_k or Θ .

Lamb (1987b) mathematically defined the orientation of stationary states and plotted them against Θ for different values of k . This graph is shown in Fig. 6.5a. Because Lamb did not extend this method to measuring vorticity, I have replotted Fig. 6.5a as ϕ against k for different values of Θ . The resultant graph (Fig. 6.5b) lends itself directly to field measurements of clast populations. If it is assumed that the finite strain was sufficient to have caused clasts to rotate into stable positions, then for a given outcrop, k and ϕ can be measured and plotted and so information concerning vorticity deduced. The graph is similar to one determined independently by Passchier (1987) which plots η (equivalent to ϕ) against B (an aspect ratio factor) for variable W_k (Fig. 6.5c). In both these graphs, for a given value of W_k or Θ , all objects to one side of the vorticity curves will be permanently rotating and will reach stable positions on the other side.

Theoretically, the stable clasts should actually plot along the relevant vorticity curve. It is thus necessary to determine which clasts are permanently rotating and which had

reached stable positions. Such an analysis was first carried out by Passchier (1987) on mylonites where the shape of recrystallised tails around porphyroclasts can be used to determine whether they were stable (δ type tails) or rotating (δ type tails; Passchier and Simpson 1986). As noted above, in the Snymansdrift melange such tails are poorly developed or absent. However, it is evident from Passchier's (1987) analysis that the vorticity number is defined by a 'cut off point' to one side of which clasts have variable senses of tail asymmetry but have only one sense to the other side. This latter sense also defines the overall sense of shear for the deforming zone.

6.2.6 Application to the Snymansdrift melange

In attempts to apply the theory to rocks, the following six requirements should at least be met (Passchier 1987, Lamb 1987b):

- 1) The fabric and general setting of the samples should indicate that deformation was reasonably homogeneous on the scale of the sample and was non-spinning.
- 2) Grain size in the matrix should be significantly smaller than the size of the objects in order to make reasonable the assumption of homogeneous flow.
- 3) High finite strains are required to rotate sufficient objects to stable positions.
- 4) Object shape should be regular and approach orthorhombic shape symmetry (or approximate an ellipsoid).
- 5) A sample should contain a large number of spatially well dispersed objects (to avoid problems of clast interference during rotation) that have variable aspect ratios.
- 6) Objects should not change shape significantly during the deformation.

The combination of these six factors underscore the fact that the practical application of the method in general is fraught with difficulties and limits the number of outcrops in the melange at which it could be applied. Both Passchier (pers. comm. 1986) and Lamb (pers. comm. 1987) believe their analyses can be applied to the melange if they are used with caution. In the melange only three localities were found to meet the six criteria to a satisfactory degree. Two of these localities exposed the XZ plane, and one the YZ plane, of the inferred strain ellipsoid. The outcrops are marked as localities W1, W2 and W3 respectively on Fig. 4.1. The main factors limiting the selection of outcrops were;

a) in the lower parts of the melange clast content was very high and clast impingement/interference is common. Brittle deformation of clasts is also ubiquitous and therefore the clasts had changed shape.

b) in the upper parts of the melange there are insufficient numbers of large clasts.

At each locality values of k and θ were measured from the clasts, assuming that the foliation plane was sub-parallel to the overall boundaries of the shear zone. The readings were plotted directly onto the both the Passchier and revised Lamb graphs. The results are shown in Fig. 6.6.

6.2.7 Discussion of results

Distinct patterns are present in Fig. 6.6 which can be interpreted in terms of the rigid body rotation model. The clast distributions do not define specific W_k or Θ curves but a definite asymmetric clustering of data is apparent. Thus, despite the practical difficulties in applying the method, it appears a reasonable interpretation of the results of the vorticity analysis can be attempted. In both the XZ planes (Fig. 6.6a to d), inferred cut-off lines have been drawn, to one side of which the clasts show differing signs of long axis orientation, whereas to the other side they show a similar sign of orientation. These lines separate respectively clasts thought to be permanently rotating from those that have reached stable positions. In both cases the sign of similar orientation is positive and corroborates the empirical observation that the longer clasts tend to dip at a steeper angle than the tectonic fabric in the XZ plane. Indeed, the long clasts seem very 'sensitive' to W_k . The vorticity analysis therefore gives an *extensional* sense of shear at these two localities.

The analysis suggests W_k at localities W1 and W2 was in the range 0.7-0.9 with inferred cut-off points of $W_k = 0.76$ and 0.8 respectively (Fig. 6.6, Passchier plots). This corresponds to a range in relative vorticity vector Θ of between 30° and 45° and cut-offs of 40° and 34° respectively on the modified Lamb plots. These results suggest that at these localities deformation, at least during the final stages of flow, was not one of simple shear alone but involved a significant component of pure shear across the tectonic fabric. In the YZ plane (locality W3, Fig. 6.6e and f) no asymmetric pattern of long axis orientation is seen suggesting there is no component of simple shear along the fabric seen in this plane. This supports the interpretation that the down-dip lineation defined by the clasts in the melange indicates there was no simultaneous transcurrent component to movement in the SDSZ.

Given the heterogeneous nature of clast distribution in the melange it would be dangerous to extrapolate these results to the entire unit. However the predominance of clasts parallel to the fabric in the central parts of the melange suggests the extensional

sense of shear may be a consistent feature and that there has been a component of pure shear across the entire zone. This is consistent with the observed irregular chocolate tablet boudinage of chert layers at the edges of BIF blocks (chapter 5.3.3). For isochoric behaviour in the shear zone this type of strain path would produce an oblate strain ellipsoid. De Wit's (1985) analysis of strained ocelli from a deformed lens of simatic rocks within the SDSZ (chapter 5.6) indicated the finite strain ellipsoid has principal strain ratios of 5.8:3.6:1. De Wit (*op. cit.*) did not state Flinn k values for the ellipsoid but it is evident from his data the ellipsoid lies in the field of apparent flattening on a Flinn plot (Ramsey and Huber 1983). The limited ductility contrast between ocelli and host basalt and lack of microstructural evidence for significant volume changes in the ocelli during deformation suggest the strain path was one of true flattening.

6.2.8 Summary: implications for sense of shear criteria

Although clast rotation, deformation and associated tail asymmetry patterns are being used as sense of shear criteria with increasing frequency in the literature, observations in the melange and theoretical studies by Lamb (1987b) and Passchier (1987) have shown that clast rotations are (surprisingly) complex. Such complexities may not be appreciated by many geologists and it may lead to wrong interpretations concerning sense of shear. Three of the most important characteristics of clasts in a deforming zone that may cause the most alarm are;

- 1) the potential for clasts to rotate in the opposite sense to the overall vorticity of the deforming zone wherever a significant component of pure shear is present.

- 2) the effects of changing shape of clast during deformation as was seen in the melange. Such changes can lead to very complex rotations (Lamb 1987b).

- 3) Furthermore, it appears from experimental work (Passchier and Simpson 1986) and field and microstructural observations (Takagi and Ito 1988 and this study) that the development of tail asymmetry may at least initially be deformation mechanism dependent; it is thus necessary to determine if the tails were formed by dynamic recrystallisation or pressure solution as similar-looking tails in hand specimen may have had very different progressive rotational deformation histories. Misidentification of the type of tail may therefore lead to erroneous interpretation of the sense of shear. Passchier and Simpson (1986) showed experimentally that the interplay between rate of rotation, rate of recrystallisation and shear strain rate was a further control on tail development.

Porphyroclasts, or any pre-deformational rigid objects, can therefore only be used as kinematic indicators when subjected to a rigorous microstructural, geometrical and statistical analysis. In outcrop, a simple visual determination of tail asymmetry may often

lead to the wrong interpretation of sense of shear and if they are to be used at all as kinematic indicators, only the sense of stair-stepping of well developed tails of dynamically recrystallised material around circular clasts should be used. This is because the parent clasts are the only ones likely to have permanently rotated in the same direction as the overall vorticity during deformation. If a more detailed analysis is carried out however, the potential exists to obtain a significant amount of further information concerning the strain path in a deforming zone.

If W_k is found to be <1 , there are wider implications for other sense of shear criteria such as shear bands and the orientation of secondary fractures and veins such as Reidel shears and *en echelon* tension gashes. The initial orientation of many of these features is assumed to be controlled by the incremental stretching axes and they commonly initiate at approximately 45° to the boundaries of simple shear zones. Any pure shear across these zones (or +ve or -ve volumetric dilatation - Ramsay and Huber 1983) will cause the orientation of the incremental stretching axes to change. Thus the orientation of these secondary features will also change and they will not correspond to 'traditional' depictions of such features. These structures may then be open to misinterpretation. Such a problem will probably be very prevalent in all shear zones, but especially transpressional or transtensional environments.

The last and probably the most important factor to consider in the application of Passchier's (1987) and Lamb's (1987b) methods is one which warns against the indiscriminate use of such a vorticity analysis. Their models consider only the behaviour of a single area within a deforming zone and not how such areas affect or interact with adjacent areas in that zone. Sanderson (1982), Sanderson and Marchini (1984) and Ramsey and Huber (1987 p. 613) have shown that a combination of simple shear and pure shear as described in the models of rigid body rotation are not strain compatible features unless there is a further accompanying strain feature. Four such features which could be common in real shear zones are;

- 1) deformation of the surrounding wall rocks (Passchier 1986),
- 2) the development of a discrete discontinuity within the shear zone or along the shear zone boundary (Sanderson 1982);
- 3) material may be lost from or extruded out of the zone - the 'cream cake' strain effect (Ramsey and Huber 1987 p. 613) or fluid expulsion,
- 4) the shear zone may have non-parallel or diverging walls e.g. wedge shaped shear zones (Ramsey and Huber 1987 p. 620).

Factor 3 above is probably the most likely feature to have affected the SDSZ and requires some discussion.

6.2.9 Volume changes: an important proviso

The main limitation of this study on the SDSZ arises from the lack of data concerning volume changes within it. Thus, although it has been concluded the strain path was probably one of true flattening other paths are theoretically possible depending on whether or not there was positive or negative dilation in the zone.

From microstructural observations (chapter 6) it is clear gross chemical changes have occurred within the zone. In order to quantify volume changes it would be necessary to obtain undeformed samples of all the fault rock protoliths, their intermediate deformed equivalents and the final fault rocks (McCaig 1984). Given the scale of the shear zone, the very large degree of chemical heterogeneity within individual protoliths and the large number of samples that would clearly be involved, this major geochemical project has not been undertaken here. Unless the quantitative data is available it is difficult to determine which elements have been added or removed from the system. The following qualitative observations suggest that the volume changes were probably very complex.

1. The upper sedimentary units containing soft-sedimentary deformation structures underwent prograde metamorphism. This undoubtedly resulted in the expulsion of pore fluids and considerable volume loss.

2. The simatic units in the immediate footwall to the shear zone underwent retrograde metamorphism from amphibole-feldspar rich assemblages to chlorite-quartz-carbonate assemblages. This generally involves the addition of volatiles although it is not possible to determine if volume loss or increase occurred.

3. The cataclasis of the BIF horizon involves an initial volume increase and influx of carbonate-bearing fluids which converted magnetite to siderite (chapter 5.3). These features probably involved an overall volume increase in the melange. Subsequent deformation had an unknown effect on volume changes but syn-tectonic veining attests to continued fluid flow. A single analysis of the augen mylonite (chapter 6.3) shows there is 15% volatiles in the lithology which was derived from BIF with only 1 to 2%. Fluid influx may therefore have caused at least local volume increases.

The large proportion of sedimentary material within the SDSZ suggests that overall there was probably volume loss during deformation. If this is true, several other strain paths for the zone are possible which would result in $W_k < 1$. The two most likely are a) deformation with volume loss in plane strain simple shear or b) deformation with volume loss in flattening strain. These three alternative strain histories are shown in Fig. 6.7. Other possibilities are discussed by Ramsey and Huber (1983 p.179). Deformation with volume loss in true flattening strain is considered to be the most likely overall strain history of the

SDSZ.

6.3 A MODEL FOR THE EVOLUTION OF THE SDSZ

6.3.1 Introduction

In this chapter and the previous two, a large amount of observational information on geological relationships within the SDSZ has been presented. In this section much of this material is synthesized into a model for its evolution. The zone is clearly complex and given the large data base any model would have to be very tightly constrained. Rather than present a very detailed history, the approach adopted here is to first summarize the features considered most important which need to be explained and then secondly present a simple framework into which these features can be fitted. The summary is given in Table 6.2.

There is an apparent 'contradiction' in the senses of shear within the SDSZ. The melange zone is characterized by an extensional sense of shear but is overlain by a contractional duplex. Furthermore, there is an apparent change in movement sense along strike from the centre of the melange (where the vorticity analysis was carried out) to the east, where the base of the zone overlies younger Uitkyk Formation sediments and is therefore probably a thrust (sub-area 5, Fig. 4.1 - Table 6.1).

This apparent paradox can be explained in one of two ways.

1) The melange, where it is present in the centre of the SDSZ, could have been preserved as a large low strain lens within the lithological sequence. If so, then the melange would represent an early D2 normal fault which was subsequently inverted along much of its length, especially towards the east.

2) The different senses of shear formed synchronously during complex, three dimensional block rotations. Two key points to consider are that;

a) the footwall unit of simatic rocks and BIF is an allochthonous, fault and unconformity bounded block' (Point 1, Table 6.2), and

b) the lateral variation in strain within the SDSZ is inferred to represent variations in displacement (Point 2, Table 6.2).

Therefore, the potential does exist for very complex relative block rotations, including clockwise, have an integral part of D2 deformation.

The field relationships strongly suggest all fault movement was essentially synchronous and that model 2 is the more likely explanation. The apparent paradox of changes in sense of shear can be resolved simply in a model of fault block rotation.

6.3.2 A model for the evolution of the SDSZ

As a conceptual starting point, the development of the zone is considered as a sequence of ductile sediments and associated tectonites deformed between two more rigid blocks or sheets of simatic rocks, both being simultaneously transported north-westwards. The hanging wall sheet is not visible in the field but is inferred to be a thin simatic unit, probably bounded by a lateral equivalent of the Hollandsdrift Shear Zone (Fig. 4.1). The model is shown in Fig. 6.8.

Consider an initial, previously deformed terrain of simatic rocks in which BIF units were already dipping SE, and in which three incipient faults develop. Initiation of D2, NW directed compressional deformation, localized along BIF/simatic rock contacts, results in the uplift of both the footwall and hanging wall blocks along faults, so producing a fault bounded basin. If displacement on the lowermost fault in the sequence increases eastwards from a lateral tip point, then the footwall block will undergo anticlockwise rotation during uplift. *A direct geometrical consequence of this rotation is the development of a normal fault on the upper surface of the footwall block and near surface deformation of the BIF unit* (Fig. 6.8a). This type of fault development is analagous to the formation of lag slides (Hutton 1980).

Uplift of both hanging wall and footwall will cause rapid shedding of detritus into the developing sedimentary trough being carried piggy back on the footwall thrust sheet. Uplift and normal faulting of the BIF on the upper surface of the footwall block, combined with this detrital input would result in an observed gradation from tectonic to sedimentary processes (Point 3, Table 6.2). The tectono-sedimentary melange may therefore have its origin as a deformed disconformity.

The footwall block however, is of limited lateral extent to the east and is in effect a 'paleodome'. Obviously, a lag slide cannot form where the footwall block is not present. If a simple scissor component of movement occurs along the strike of the melange to the east, displacement can be transfered into a thrust motion (Fig. 6.8b). The same fault can then override a further sedimentary trough to the northeast. It is geometrically necessary for the pivot point to occur at locality X (Fig. 4.1), the point where footwall block, underlying sedimentary trough and melange meet.

Whilst scissor faulting dominates at the lowermost boundary of the main sedimentary trough, continued northwest directed compression involving components of pure and simple shear ($W_k < 1$), will cause contractional deformation of the sediments themselves. Early syn-sedimentary imbrication and stacking of thrust slices of mudstone/sandstone and simatic rocks produced a duplex between the melange and an overriding unit of lithofacies A of the Uitkyk Formation. A consequence of this duplexing is that cyclical deformation of early formed sedimentary contacts occurs (Fig. 6.8c). Thus a gradation in lithological contacts would be observed.

The combined effects of stick-slip behaviour along the roof thrust (mudstone/sandstone-Lithofacies A), $W_k < 1$, and rheological contrasts between the mudstones and Lithofacies A, would result in high angle out-of-sequence faults cutting through the roof to form syn-sedimentary folds in the overlying sandstones (Fig. 6.8c - Ramsey 1967). These structures were continually flattened and prograde metamorphosed as deformation proceeded to produce the deformed litho-tectonic sequence that constitutes the SDSZ.

6.4 SUMMARY

The combined field and microstructural relationships within the SDSZ have revealed a complex history of progressive D2 deformation and metamorphism. The zone represents a deformed syn-sedimentary duplex that was carried piggy-back above an allochthonous thrust sheet of simatic rocks. Complex fault block rotations, controlled by varying displacement along the bounding faults to the basin has resulted in scissor rotations along the basal fault contact. This has produced apparent reversals of sense of shear both along and across strike. The deformation history was one of volume loss in true flattening, characterized (at least locally) by W_k values of less than 1.

There are many corollaries to this model, not least of which are the implications for fault movement elsewhere in the Mt. Mare area. These are examined in chapter 7. It is now pertinent to investigate the timing of gold mineralization with respect to the proposed deformation history for the SDSZ.

Fig. 6.3 Examples of clast deformation in the melange

Drawings are approximately actual size and are taken from various samples. In all cases the fabric runs across the page and dips into the page.

- 6.3a). Planar to curvilinear fractures in the YZ plane of the strain ellipsoid.
- b). Planar to curvilinear fractures in the XY plane of the strain ellipsoid.
- c). An example of more distributive cataclasis - YZ plane.
- d). An example of the possible ambiguity of fracture interpretation in the melange. These fractures could be either tensile fractures subsequently rotated and giving a dextral sense of shear or shear fractures giving a sinistral sense of shear.
- e). Evidence for pressure solution in the melange. Clast with rounded edges and tail development. The clast was probably initially lozenge shaped with sharp edges which have been smoothed off by diffusive mass transfer.
- f and g). Ambiguity in pressure solution tail asymmetry - note similar foliation patterns around clasts, both drawings taken in the XZ plane looking southwest. The majority of the clasts seem to dip in the same direction as the foliation but at a steeper angle (f). Others dip in the opposite direction to the fabric (g). Sense of shear cannot be deduced from these structures.

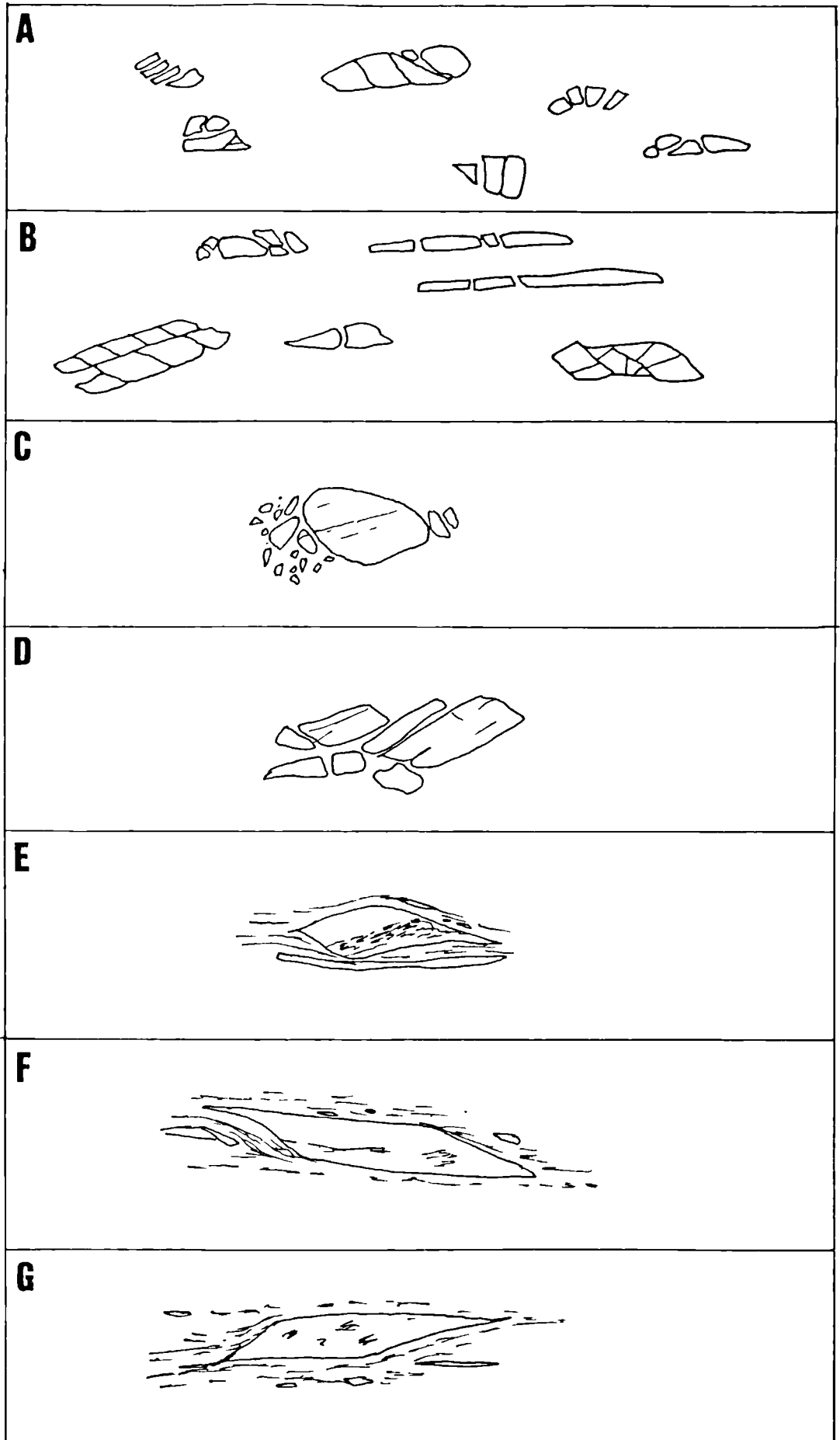


Fig. 6.4 Geometric description of clast orientation and principles of clast rotation in pure and simple shear.

6.4a). Geometric description of clast orientation (from Lamb 1987). The axes x, y are fixed and the velocity field is defined relative to them such that the margins of the deforming zone are parallel to the x axis ie a non-spinning deformation. The z axis is out of the page. The three axes x, y and z coincide with the a, c and b axes of the ellipsoidally shaped clast respectively. The orientation of the ellipse is defined by ϕ , the angle between the long axis of the clast and the x axis. Θ is the angle of the relative velocity vector between the two margins with the x axis. Negative Θ (as shown) implies dextral shear and compression).

6.4b). Plots of normalised rotation rate (R) against ϕ for different values of Θ and various values of k (from Lamb 1987). Note that for any particular value of Θ , the plots share the same values of ϕ for which R is a maximum or minimum and similarly for $R = -1$.

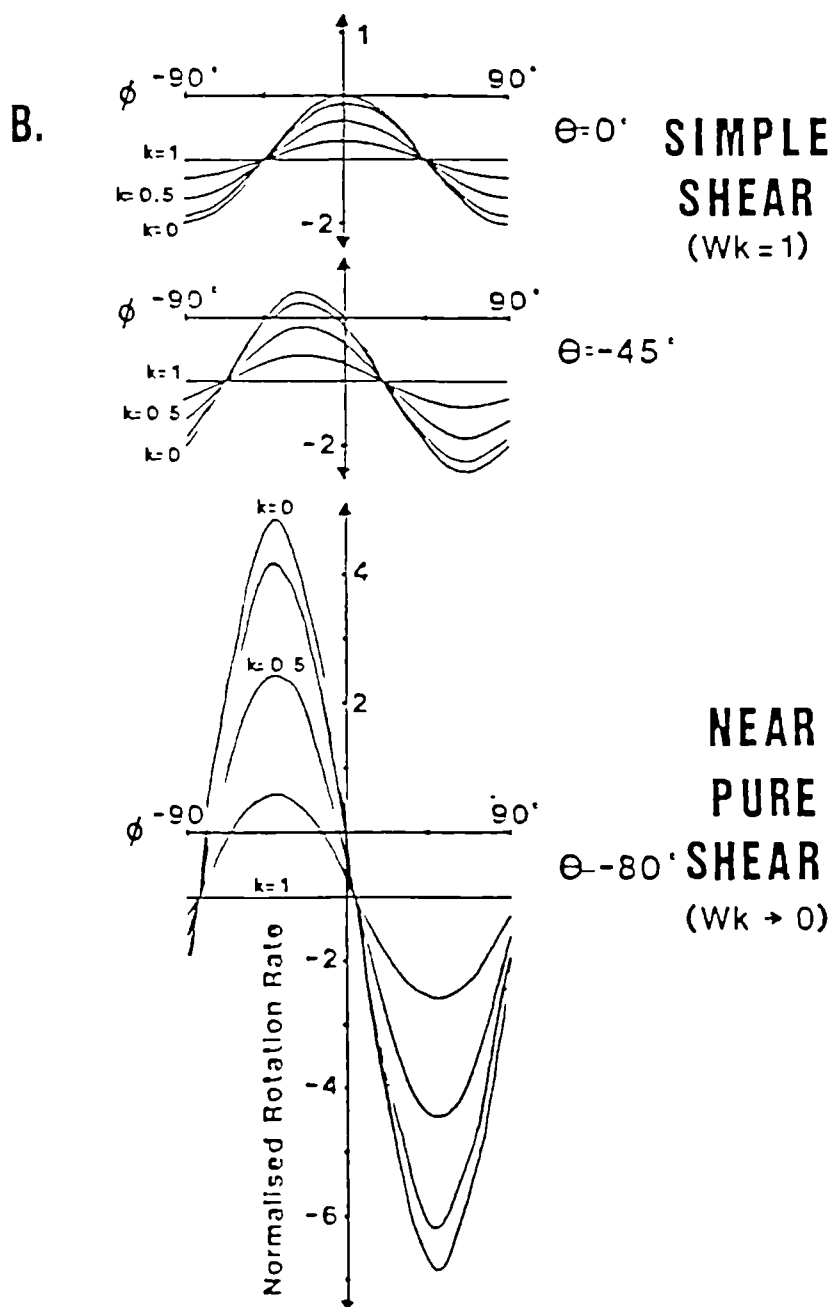
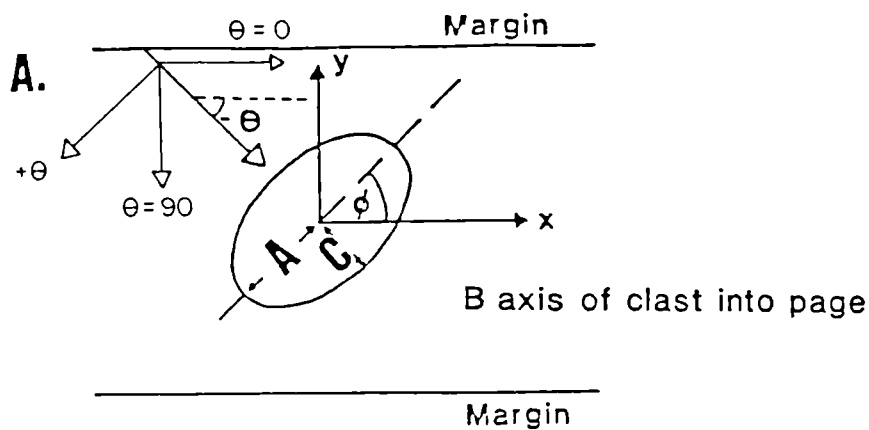


Fig. 6.5 Orientation of Stable States

6.5a). Plot defining orientation of stable states for which such states exist (from Lamb 1987). In general there are two states; the one with the value of ϕ furthest from zero is unstable as there is a tendency for the inclusion to rotate away from it; the other (ϕ closest to zero) is stable as the inclusion rotates towards it. The orientation midway between these two states is that for which R is a maximum or minimum.

6.5b). Plot of k against ϕ for different values of Θ . A simple graphical replot of Fig. 6.5a was used to construct this diagram. For clarity only three curves in the unstable field have been plotted. Clasts with k and ϕ values below the straight line and to the left of the Θ curves should reach stable positions and lie along a specific Θ curve.

6.5c). Plot of B , a shape factor, against n (orientation) from Passchier (1987). Note the similarity in final graph shape to Fig. 6.5b. Passchier's graph was derived from tensor calculations. Note the kinematic framework for this graph shows sinistral shear. Clasts with B and n values to the left of the W_k curves permanently rotate whereas those to the right reach stable positions and theoretically, for a given W_k deformation, should plot along the specific W_k curve.

N.B. All graphs show orientation as a positive angle. It can of course be negative and thus the graphs are symmetrical about their abscissae.

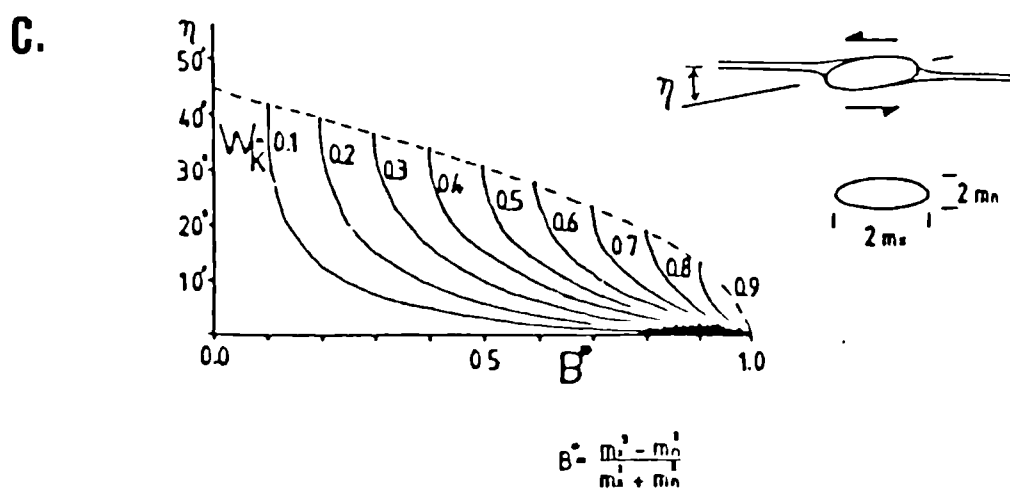
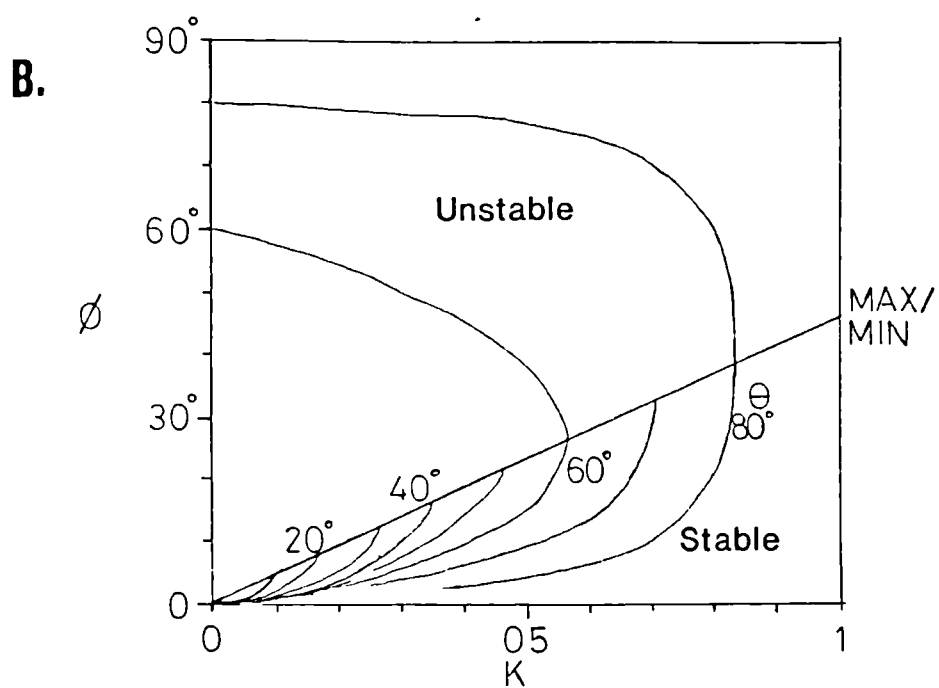
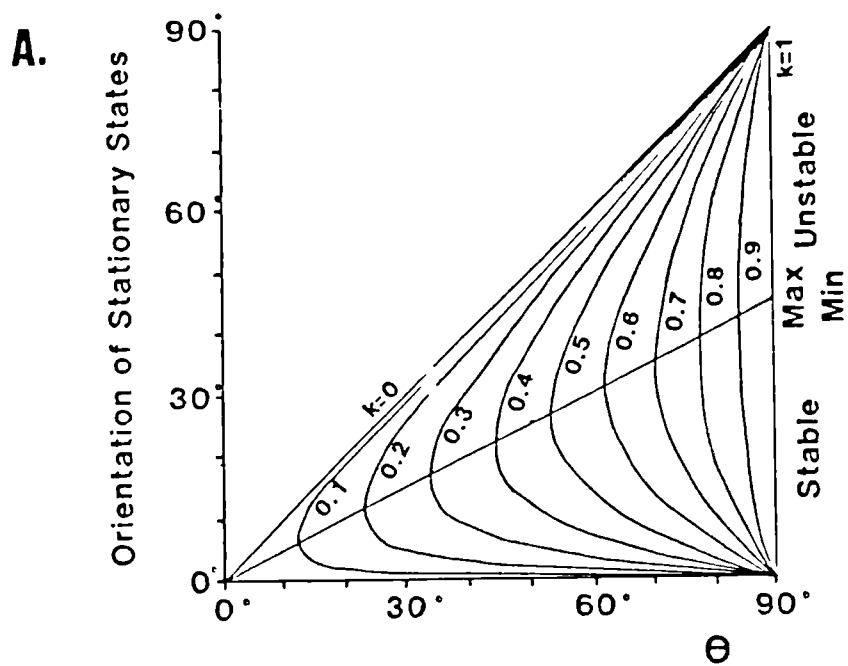


Fig. 6.6 Results of Vorticity Analysis

Graphical representation of clast aspect ratios (B and k) against orientation (α and ϕ) with respect to deformation zone boundaries on the Passchier and Lamb graphs respectively. Analysis was carried out at three localities; W1 and W2 on sections parallel to the XZ plane of the strain ellipsoid, and W3 parallel to the YZ plane (Fig. 4.1). For discussion of results see text.

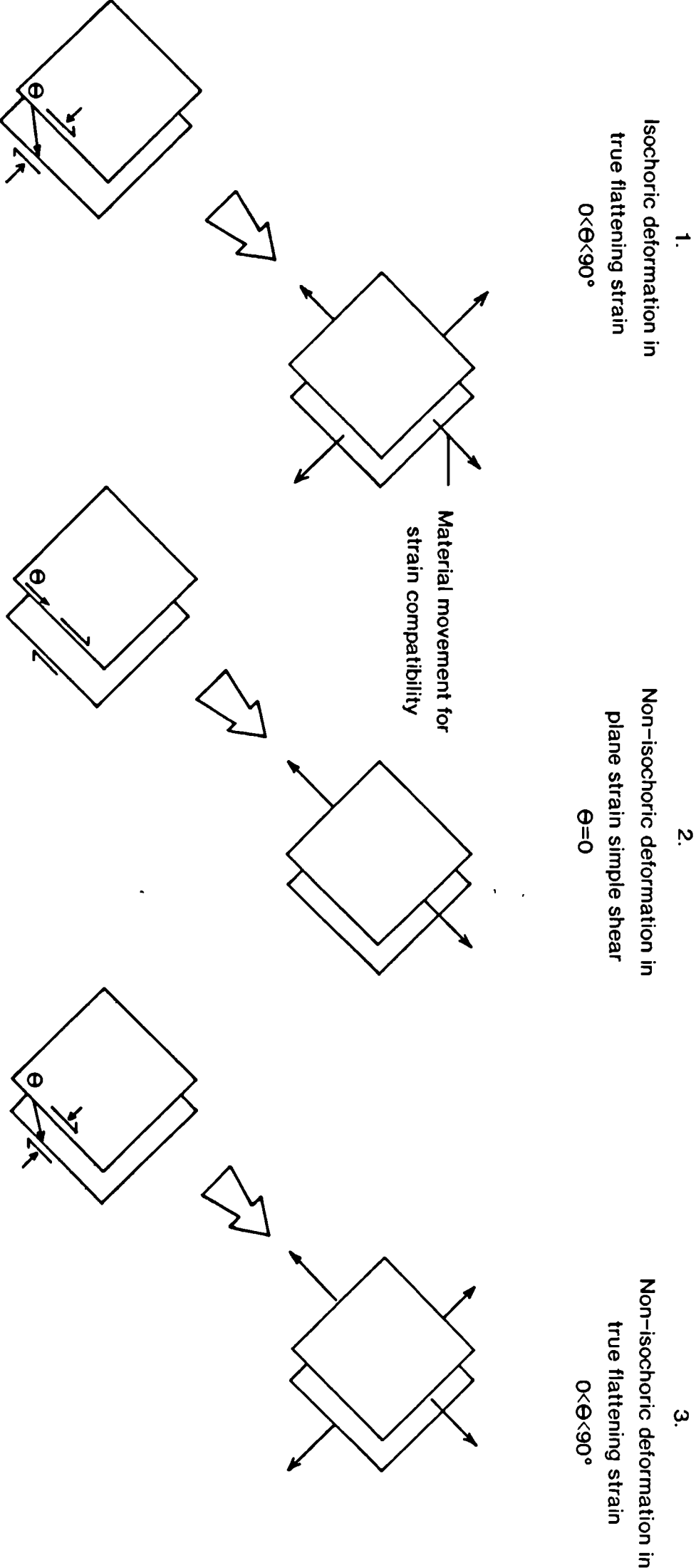


Fig. 6.7 Some possible strain histories for the SDSZ

TABLE 6.1 STRUCTURAL SUMMARY OF SUB-AREAS WITHIN THE SDSZ
(see Fig. 4.1 for location)

SUB-AREA 1 Snymansdrift No. 2 and 3 Hills (Fig. 1.3, 4.1)

Mapping Scale 1:1000 (Fig. 6.1)

State of strain Lowest in the SDSZ. Sedimentary structures and contacts well preserved; only overprinted by moderate cleavage development.

Features of note Change in regional strike of bedding from NE (to the east) to NW, subparallel to unconformity at locality A (Fig. 2.2). Area covers western termination of SDSZ. Melange thins down to a 1-2m wide mudstone layer (containing soft-sedimentary structures described in chapter 4.5) between two packages of sandstone (Fig. 4.1, 6.1). Breccias show little evidence of preferred orientation or plastic overprint.

Main structures Upright (80° SW), E-W trending, open to close anticline, plunging steeply (70°) to the SW. Several associated slides (Hutton 1981) effectively slice out some of the fold limbs. Bedding planes near slides are marked by qtz-tourmaline fibrous shear veins

Conclusions Fold does not coincide with regional change of strike, nor do bedding\cleavage relationships change around the bend. Therefore the change in strike predates the fold and reflects original bedding orientation. Veins and slides indicate flexural slip under high pore fluid pressure may have been an important deformation mechanism (Fyfe *et al.* 1978).

SUB-AREA 2 Snymansdrift No. 4 and 5 Hills (Fig. 1.3, 4.1).

Mapping Scale 1:1000 (Fig. 6.2). Mapping concentrated on well exposed upper parts of melange and overlying sediments

State of strain Medium. Original structures locally well preserved but towards eastern end of sub-area fabric development increases. Strong fabric development also localised along lithological contacts.

Features of note Sedimentary contact between mudstones and Lithofacies A in sub-area 1 is now marked by a chert breccia horizon which clearly truncates bedding in Lithofacies A. This contact is now a fault and supports the conclusions of chapter 4.8.3. The breccias are thought to represent syn-sedimentary faults which propagated along prominent lithological contacts. However, this breccia zone is itself folded around the two main folds and is thus a folded fault. Contacts with lenses of simatic rock are also often marked by such breccias but are often strongly overprinted by fabric development. Hinges of folds have been infiltrated by fluids leading to quartz-tourmaline veining, tourmalinised cross-bedded sandstones and gossans (partly after carbonate).

Main structures ENE-WSW trending, asymmetric, steeply inclined (80°SE) but gently plunging (15-35°SW) synclines. The intervening anticlines are either poorly developed or removed by steeply SE dipping slides. Note the folds do not deform lithological contacts beneath the mudstone layers and that the mudstones have flowed disharmonically into the fold hinges. The folds dies out laterally to the west and also upwards along their axial traces.

Conclusions Features imply the sandstone-mudstone contact was a significant decollement horizon within the tectono-sedimentary package and that two discrete phases of deformation occurred; an earlier syn-sedimentary phase of faulting that propagated along the pre-existing lithological contact and a second phase of faulting also possibly syn-sedimentary that was rooted in the mudstones but which cut up through the decollement horizon deforming the earlier faults. Therefore the folds and faults seen in this sub-area are interpreted as the result of progressive D2 syn-sedimentary deformation but these structures were then progressively flattened and overprinted by fabric development. This may have lead to a slight transection of the folds by the cleavage. Displacement along minor faults and folds also decreases westward. Flexural flow was an important fold mechanism aiding thinning of bedding along breccia zones.

SUB-AREA 3 Eastern Zandrivier (Fig. 1.3, 4.1).

Mapping scale 1:5000 (Fig. 4.1)

State of strain Medium to high - concentrated along lithological contacts but a more pervasive fabric overprint is present as a whole.

Features of note Stereonet data shows two domains of markedly different bedding orientations: Domain 1 - beds dipping steeply SE (sub-parallel to cleavage) come predominantly from BIF blocks in the melange. Clast lineations, minor fold hinges and extension vein fibres all plunge sub-parallel down dip (50-70°SE) of the fabric: Domain 2 - shallow dipping beds in sandstone and conglomerate beds associated with folds and slides in a central area above the melange. Slides are often marked by unusual quartz-tourmaline breccia zones and chert breccias (chapter 4.8.3).

Main structures Two major folds in sandstone-conglomerates - gently to open, steeply inclined SE and gently plunging (value).

Conclusions Folds, faults (and lenses of simatic rock) are confined between melange-mudstone contact and uppermost mudstone-lithofacies A contact. This suggests both these contacts may be decollement horizons.

SUB-AREA 4 Zandrivier to Mt. Mare

Mapping scale 1:5000

State of strain High - strong fabric overprint throughout sub-area but especially at Zandrivier

Features of note A significant lateral gradient from east to west in the degree of finite strain and metamorphism is present. Zandrivier has the greatest overprint of plastic deformation in the SDSZ. Deformation is penetrative throughout the lithologies. However towards Mt Mare deformation again becomes localised (as slides) along lithological contacts. Cleavage and bedding are sub-parallel throughout most of the sub-area (Fig. 4.1). The range in cleavage values is primarily due to the anastomosing fabric (chapter 4.5). Lithological units are found as discrete tabular units but are penetratively deformed in the east

Main structures At the top of Mt. Mare the melange grades up into a lens of sandstone and conglomerate. The upper contact of the lens is highly deformed by an upright, tight anticline-syncline pair. The folds show a slight E-W variation in plunge from 55° to sub-horizontal respectively. An associated slide is marked by a narrow (50cm wide) quartz-tourmaline breccia zone.

SUB-AREA 5 Mt. Mare to Greenwalds. Area covers eastern end of SDSZ plus an underlying unit of Uitkyk Formation sediments, in turn underlain by intercalated BIF horizons and greenstones. The eastern end of the Kuschke shear zone has also been mapped.

Mapping scale 1:5000

State of strain Moderate

Features of note The ENE striking BIF horizons to the north are truncated by Uitkyk Formation sediments along an east dipping unconformity (locality U, Fig. 4.1). This is opposite to that seen at the western end of the BIF horizons (locality A, Fig. 2.1). The SDSZ now overlies a structurally upward facing unit of Uitkyk Formation sediments which also dip and young to the SW. Although slightly foliated, there is no clear structural break between the sediments and underlying BIF horizons. At locality V (Fig. 4.1) well developed breccio-conglomerates contain BIF and chert clasts apparently derived *in situ* from the underlying BIF. This indicates the sediments are also *in situ* and that the contact is a deformed disconformity.

Main structures Cleavage and bedding are predominantly sub-parallel (75-90°), lineations plunge down dip of the fabric. Two major folds are present; an upright, steeply plunging syncline that has deformed the main unconformity and an upright, moderately plunging anticline in the uppermost unit of sandstones.

Conclusions The opposite dips of unconformities at either end of the Kuschke BIF suggest the central allochthonous unit of simatic rocks is an isolated uplifted block or dome surrounded by faults and unconformities. Paleotopography may therefore have been a significant control on deformation. The base of the SDSZ is marked by deformed simatic rocks and BIF which overlie younger Uitkyk Formation sediments. This suggests the SDSZ is a thrust related feature at its eastern end.

CHAPTER 7

MINERALIZATION IN THE SNYMANSDRIFT SHEAR ZONE

"During the final manifestations of igneous activity associated with the intrusion of the...granite, fissures formed in the Swaziland System and in the granite. These were filled predominantly with barren vein quartz, but occasionally they were reopened and sulphides, gold and other minerals were deposited in them." Willemse (1938).

7.1 INTRODUCTION

The SDSZ is host to several ex-gold workings which are currently being explored by AAC to evaluate their sub-surface potential. Although each of the deposits has its own unique geological characteristics they have two common denominators; they are structurally controlled and they are associated with abundant tourmalinisation. Four of the deposits, Zandrivier, Mount Mare Valley (MMV), Snymansdrift # 3 Hill, and Greenwalds (Figs. 1.3, 4.1) have been studied in order to highlight the main similarities and differences that exist between them. The information from each of the deposits can be integrated in a model for mineralization in the SDSZ as a whole.

7.2 ZANDRIVIER-MMV PROSPECT

The two deposits at Zandrivier (ZR) and MMV, both lie along the same narrow (2-5m wide) high strain zone, within the central portion of the SDSZ (Fig. 4.1). The zone is marked by the variably deformed chert bars (chapters 4.8 and 5.4) and two differing styles of mineralization are associated with it. The first is seen at surface and is best developed around the ZR area itself, and the second style is the sub-surface equivalent, and has been intersected in drill core from the prospects.

7.2.1 Surface mineralization at Zandrivier

At surface, the host lithologies to the ZR deposit are sediments, at a notably

lower visible state of strain than the surrounding rocks and sedimentary structures, such as bedding, and evidence for early soft-sediment deformation of the chert bars are well preserved (chapter 4.8.3, plate 4.10). The brecciated chert bar represents an early D2, bedding parallel fault at the contact between hanging wall sandstones and footwall chloritic mudstones. It is this horizon that carries gold-bearing quartz veins. The host sediments around the 'reef' are grey in colour due to the presence of tourmaline and some minor graphitic. Both along and across strike, the sedimentary structures and breccias become strongly overprinted by intense ductile (predominantly diffusive mass transfer) processes (Table 5.1). Surface and adit mapping indicates the entire ZR deposit occupies a low strain lens of sandstone within the central SDSZ that has escaped subsequent deformation and is approx. 300m long, 20-50m wide and at extends for at least 70m down dip (Fig. 4.1). It may represent a fault bound sliver (or horse) of sandstone from the early D2 syn-sedimentary duplex.

In and around the original bedding-parallel fault zone, the breccia clasts show a strong preferred orientation (Fig. 7.1b.). There is a clear reorientation (and associated elongation of the clasts due to pressure solution deformation) into the main fault zone. This indicates there is later D2 localized strain along the earlier syn-sedimentary fault. It is this zone that hosts the quartz vein. The lineation defined by the orientation and elongation of the chert clasts plunges SE in the same direction as the lineations elsewhere in the SDSZ (Fig. 4.1), but at a lower dip (30-40°). The same lineation direction is recorded in fibrous quartz and quartz-tourmaline veins found on exposed bedding planes. These are also interpreted as D2 fabrics and suggest that overall, the low strain lens may have a shear band-like geometry. A possible schematic representation of the geometric orientation of the lens based on detailed mapping of the prospect is shown in Fig. 7.1a.

In detail, the quartz vein anastomoses in three dimensions and in general *cuts* the fabric in the fault zone. It also tends to 'pinch and swell' both along strike, and up and down dip, giving the veins as a whole a podiform appearance with the pods being a maximum of 1m thick. The veins are not deformed by the fault, rather they appear related to movement along the plane which anastomoses in three dimensions (as illustrated in Fig. 7.1b). Therefore, introduction of the quartz vein was related to a brittle event that occurred *late in the history of the low strain lens*.

Detailed sampling of the adits at ZR by the writer and AAC has shown that all the gold found to date was located in the quartz vein. However, in hand specimen, the

quartz veins are intensely fractured and it is these late stage fractures that contain the gold (plate 7.1a). Therefore, the quartz predates introduction of gold. In thin section the quartz shows well developed transgranular and intragranular cracks. However, the majority of the cataclasis seen is concentrated around grain boundaries and grain size reduction is very evident. Zones of more intense cataclasis are directly associated with the influx of tourmaline stringers and gold (plate 7.1a) and there is probably a genetic link between them. There is an extensive arsenic anomaly in the adits (AAC unpubl. data) and some very weathered, red, footwall mudstones suggesting that there may have been some arsenopyrite associated with the mineralization although no original sulphides were observed.

Overall, it would appear that the gold and tourmaline were introduced late in the history of the low strain lens and were associated with minor brittle fracturing of an earlier quartz vein. The tourmaline also passively infiltrated the surrounding sediments to give them their distinctive grey colour. Therefore, both fluid flow events were associated with *repeated* brittle faulting along a ductilely reactivated bedding parallel fault. This is consistent with the stick-slip model of fault propagation proposed for these horizons in chapter 4.8.4 and corroborates the ideas of Willemse (1937 - chapter 1.4). The brittle deformation is clearly restricted to the low strain lens and may have been caused by the build up of stresses around the edges of the low strain lens if it acted as a rigid obstacle whilst ductile deformation continued around it (Fig. 7.1a). Hence at the prospect scale, there was late D2 localised brittle failure in the lens and more distributive ductile deformation around it.

7.2.2 Borehole mineralization

In borehole intersections the gold bearing horizons exhibit varying degrees of post-mineralization deformation. The drillcore from directly below the ZR prospect showed the lowest strain and that at MMV the maximum. However, as noted in chapter 5, the fault rocks that host the borehole mineralization in the drill core are at a higher degree of finite strain than the surface. This is consistent with the dying out of the low strain lens at depth (Fig. 7.1). Features associated with the ZR borehole mineralization are described first. The following mineral assemblage is associated with the borehole mineralization:

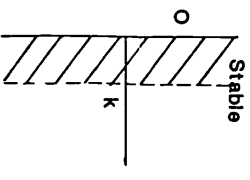
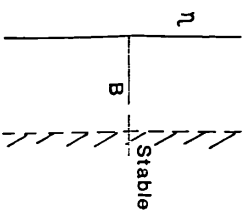
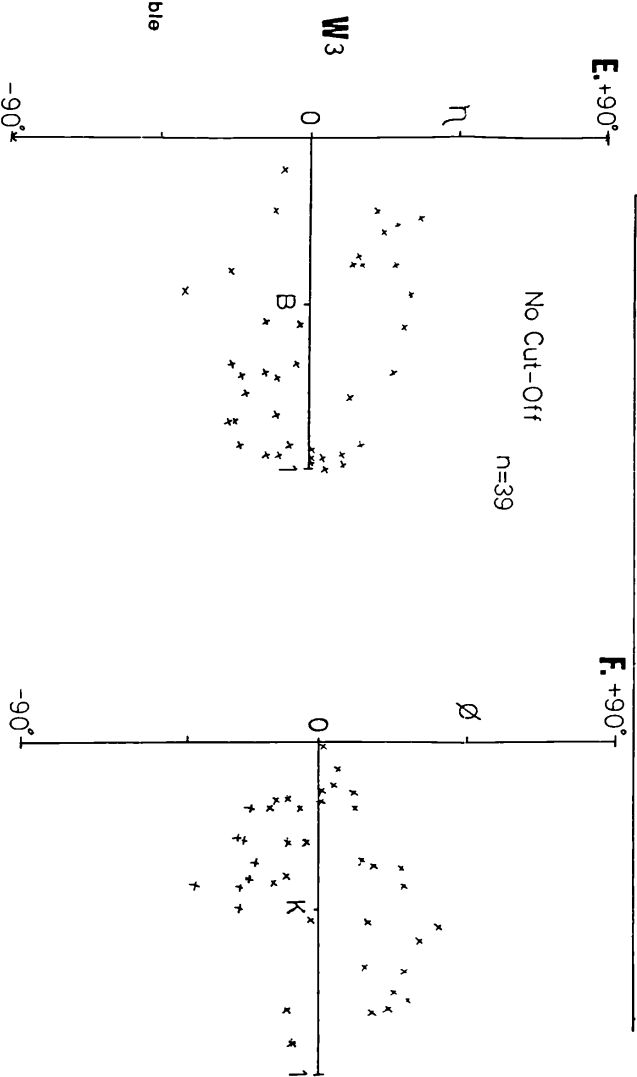
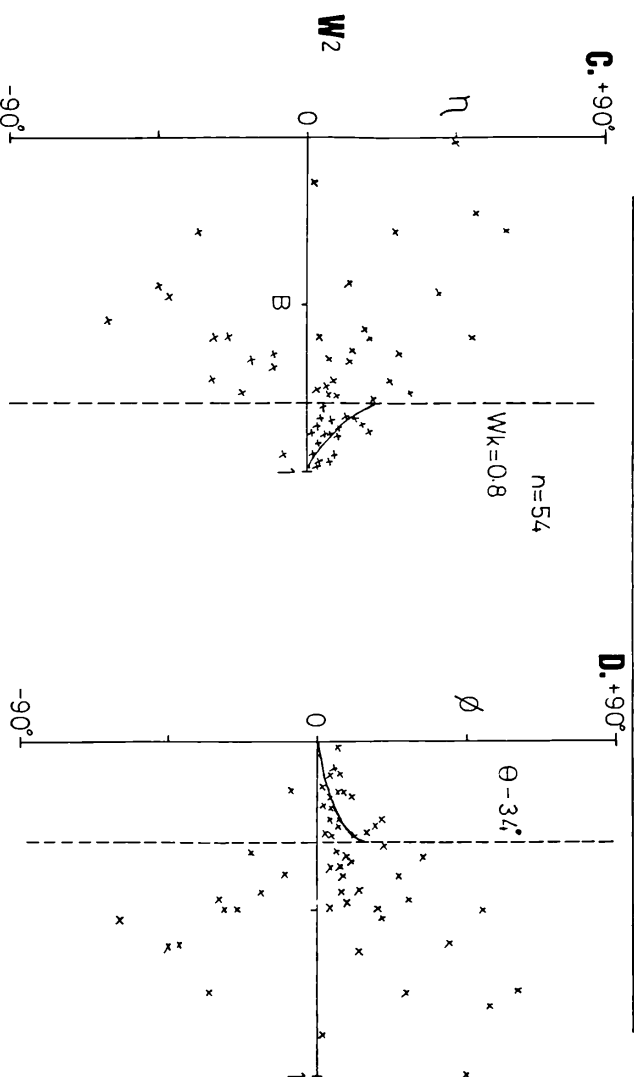
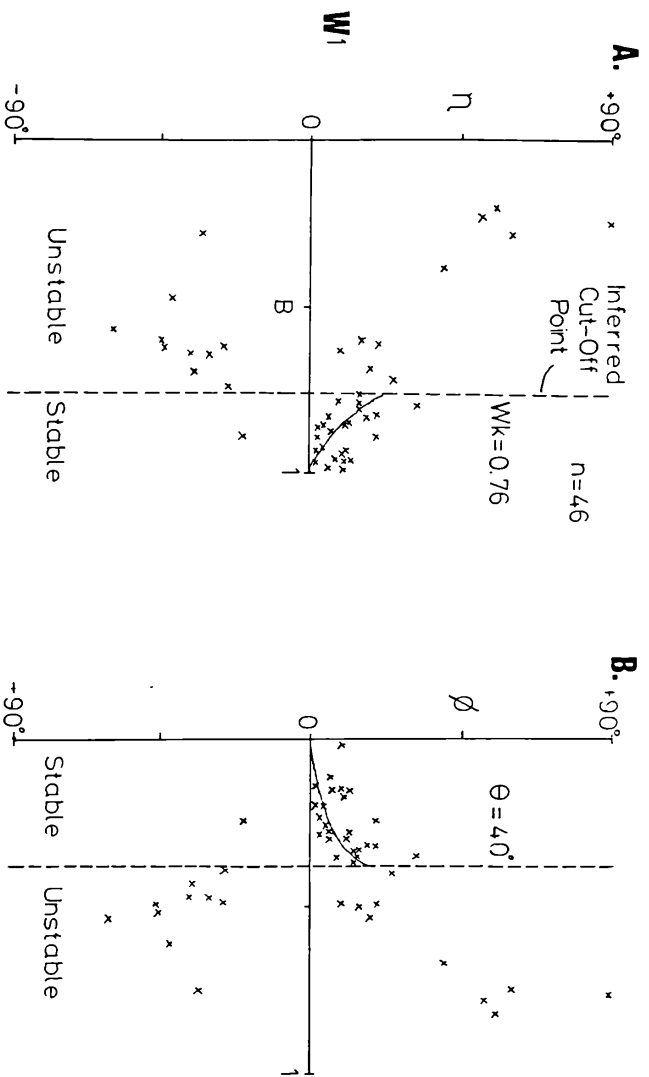
quartz-tourmaline-arsenopyrite-pyrite-carbonate-chlorite and gold.

Quartz and ankerite: The ore zones are marked by extensive quartz +/- ankerite veining which seems to have exploited the contact between deformed sediments (quartz-sericite phyllonites - chapter 5.5) in the hanging wall and a chloritic phyllonite in the footwall (Fig. 5.1). The deformed sediments contain variably strained chert clast breccias, often overprinted by pyrite and tourmaline (plate 7.1b). Although no original bedding is visible, the unit and the mineralized contact are directly correlatable with surface mineralization (Fig. 5.2). The chloritic phyllonite contains extensive, deformed syn-tectonic quartz-carbonate-pyrite veins and these represent a fluid flow event that predates the mineralized veins. This could correlate with the main quartz veining event seen at surface in the low strain lens.

The quartz-ankerite veins disrupt the D2 tectonic fabric and also contain fragments of the wall rock lithologies suggesting that hydraulic fracturing, sub-parallel to the tectonic fabric, was responsible for fluid emplacement (plate 7.1d). In thin section the quartz can be seen to be very deformed and small fractures and optical strain features are well developed. No evidence for dynamic recrystallisation was seen and no gold was observed in the fractures. The deformation produced an abundance of very fine grained secondary fluid inclusions and although several fluid inclusion wafers were prepared they contained hundreds of inclusions per square centimetre and it was thought that their study was not justifiable (plate 7.1c). Only a small percentage of ankerite (<2%) is present in the ZR core but it makes up to 50% of the host veins in MMV core.

Arsenopyrite and pyrite: The main sulphide mineral in the ore zones is euhedral arsenopyrite. It is found overprinting the wall rock or in stringers and fragments of host rock in the veins (plate 7.1d). Euhedral arsenopyrite also occurs entirely within the veins and intergrown with the quartz, as well as cutting across vein-wall rock boundaries (plate 7.1d). These features and the porphyroblastic nature of the arsenopyrite are interpreted as indicating that it was introduced at the same time as the quartz. The fluid penetrated the rocks by hydraulic fracturing but this was accompanied by simultaneous passive infiltrative metasomatism of the surrounding wall rocks and fragments in the veins.

The arsenopyrite crystals have locally undergone intense cataclasis, usually related to late brittle faults. All the pyrite observed in the mineralised zones in this study occurs as infillings along the associated fractures (plate 7.2a). Chlorite is also found in the fractures as well as locally forming rims around the arsenopyrite (plate



PASSCHIER PLOT

MODIFIED LAMB PLOT

7.2b). The chlorite also locally replaces ankerite. In other samples, the arsenopyrite is being replaced by quartz (plate 7.2c) as well as fractured.

Gold: Only a few grains of gold were observed in thin section. They all occur as occluded grains within undeformed arsenopyrite. They are interpreted as having been introduced with the quartz and arsenopyrite (plate 7.2d). All these three minerals predate the introduction of pyrite.

Tourmaline: The wall rock immediately surrounding the mineralised horizons in the ZR core is commonly a distinctive dark brown colour in hand specimen. Thin section work indicates that these rocks are tourmalinites and are made up of 90-100% schorl-related tourmaline. A gradation is seen between these tourmalinites and the original wall rock phyllonites suggesting the tourmaline is a metasomatic replacement of those original rocks. The majority of the tourmaline occurs as fine grained turbid masses that have replaced the original minerals (quartz-chlorite-sericite) and obliterated the tectonic fabric. In other samples the tourmaline is aligned and boudinaged sub-parallel to the fabric, with chlorite fibres between the boudins, indicating there has been some post-mineralization deformation and fluid flow. Thus all the ore-related minerals show evidence of microcracking and infiltration of chlorite-pyrite bearing fluids. Tourmaline is found in the wall rocks, cutting the vein-wall rock boundary and within the quartz veins intergrown with the quartz and arsenopyrite (plate 7.3a). This suggests that the tourmaline was introduced along with those minerals during hydraulic fracturing and concomitant passive infiltrative metasomatism of the wall rocks.

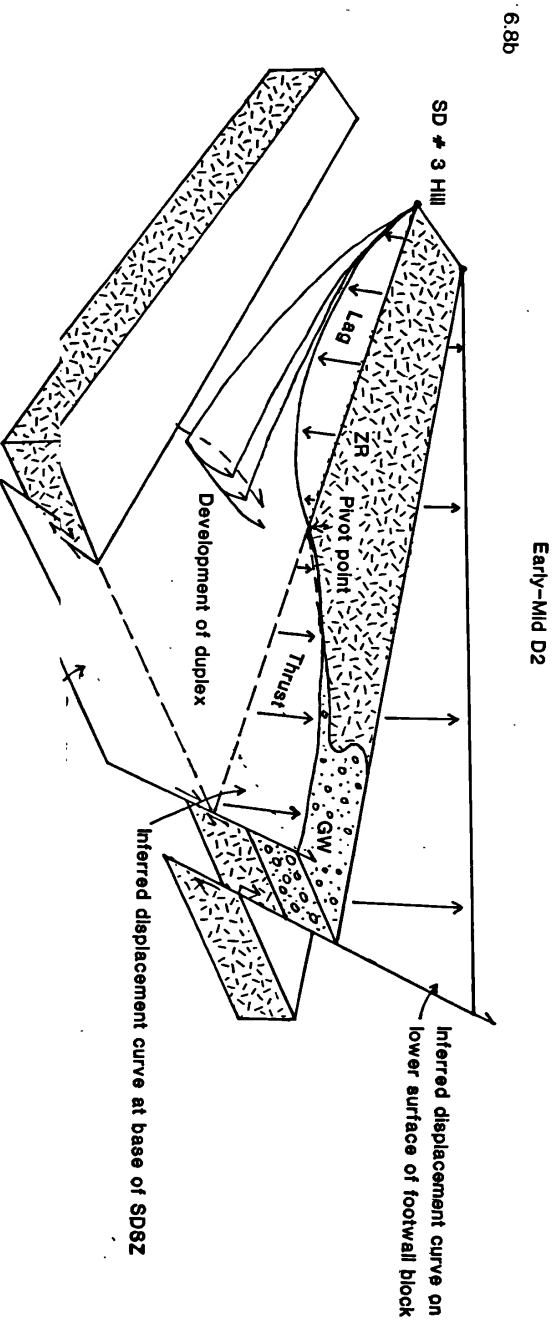
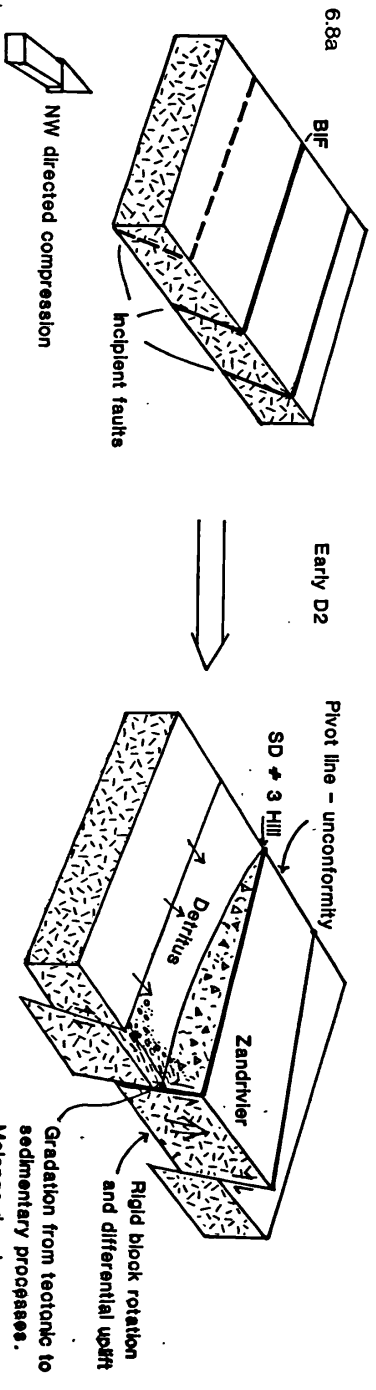
A sequence of progressive deformation of the mineralization has been recognised in the drillcore, with samples from the MMV core generally being more deformed than those at ZR (and described above). The initial microcracking of the ore minerals becomes more prevalent and is succeeded by an increasing degree of plastic deformation of the quartz, ankerite and sulphide minerals. The mineralization is being *progressively reworked* into the host chloritic phyllonites during post-mineralization deformation (plate 7.3b). In hand specimen, the most deformed veins seen in this study are fine grained, cherty-looking and consist of bands of quartz and ankerite containing elongate, smeared out blebs of arsenopyrite and pyrite. These textures are cut by several later phases of quartz-carbonate veining. In thin section both the quartz and ankerite have preserved textural evidence for dynamic recrystallisation and tectonic reworking, although annealing has also occurred (plate 7.3c). The sulphide minerals have also begun to deform in a ductile manner, to become smeared out parallel to the

fabric. Much less tourmaline is associated with the MMV core than ZR and it is not clear what effects deformation may have had on it. The most interesting feature about the deformed mineralization is that *as the degree of strain increases, there is a corresponding decrease in the gold grades* (AAC unpubl. information). Hence the tectonic overprinting apparently has an effect of downgrading the deposit. Thus as a whole, mineralization in the ZR-MMV prospects is preserved in areas of lower strain.

In summary, the field and textural relationships described above indicate that there was only one influx of gold-bearing fluids into this part of the SDSZ. It was a syntectonic, transient fluid flow/brittle failure phenomenon, preceded and succeeded by D2 plastic deformation. The fluid responsible appears to have been under high pressure and propagated by hydraulic fracture (from an unknown source at depth) through the shear zone sub-parallel to the tectonic fabric and into higher crustal levels. There, it exploited a pre-existing plane of weakness within, and passively infiltrated, the low strain lens. The fluid precipitated gold-arsenopyrite-tourmaline-quartz and ankerite in both environments (Fig. 7.1c). The variable quantities of ore-related minerals seen both along strike and down dip, suggest that the variables controlling mineral deposition from the hydrothermal fluid such as pressure, temperature, fluid composition and host rock were complex. The details remain to be investigated. Post-mineralization plastic deformation has tectonically reworked and downgraded the majority of the deposit but has preserved a low strain, higher grade "ore shoot" below the ZR prospect that plunges SE, to an unknown depth, parallel to the main D2 lineation direction.

7.3 SNYMANSDRIFT # 3 HILL

The deposit at SD # 3 Hill is located at the western end of the SDSZ and the structure of this area has been described in Table 6.1. Detailed mapping has shown that the deposit here comprises a series of four, slightly overlapping, slightly *en echelon* quartz veins that lie (similar to the ZR deposit) sub parallel to, and at the contact between, the chloritic mudstones and overlying sand tones and conglomerates (Fig 6 1). The mudstone horizon contains evidence for the soft-sediment disaggregation of chert clasts (chapter 4 6). A strong fibre lineation on the top surface of some of the veins suggests they plunge parallel to the main fold structure in the area (Fig. 6.1). The depth extent of the SD # 3 Hill deposit is not known.



6.8c Development of syn-sedimentary duplex above a lag scissor fault in combined pure and simple shear

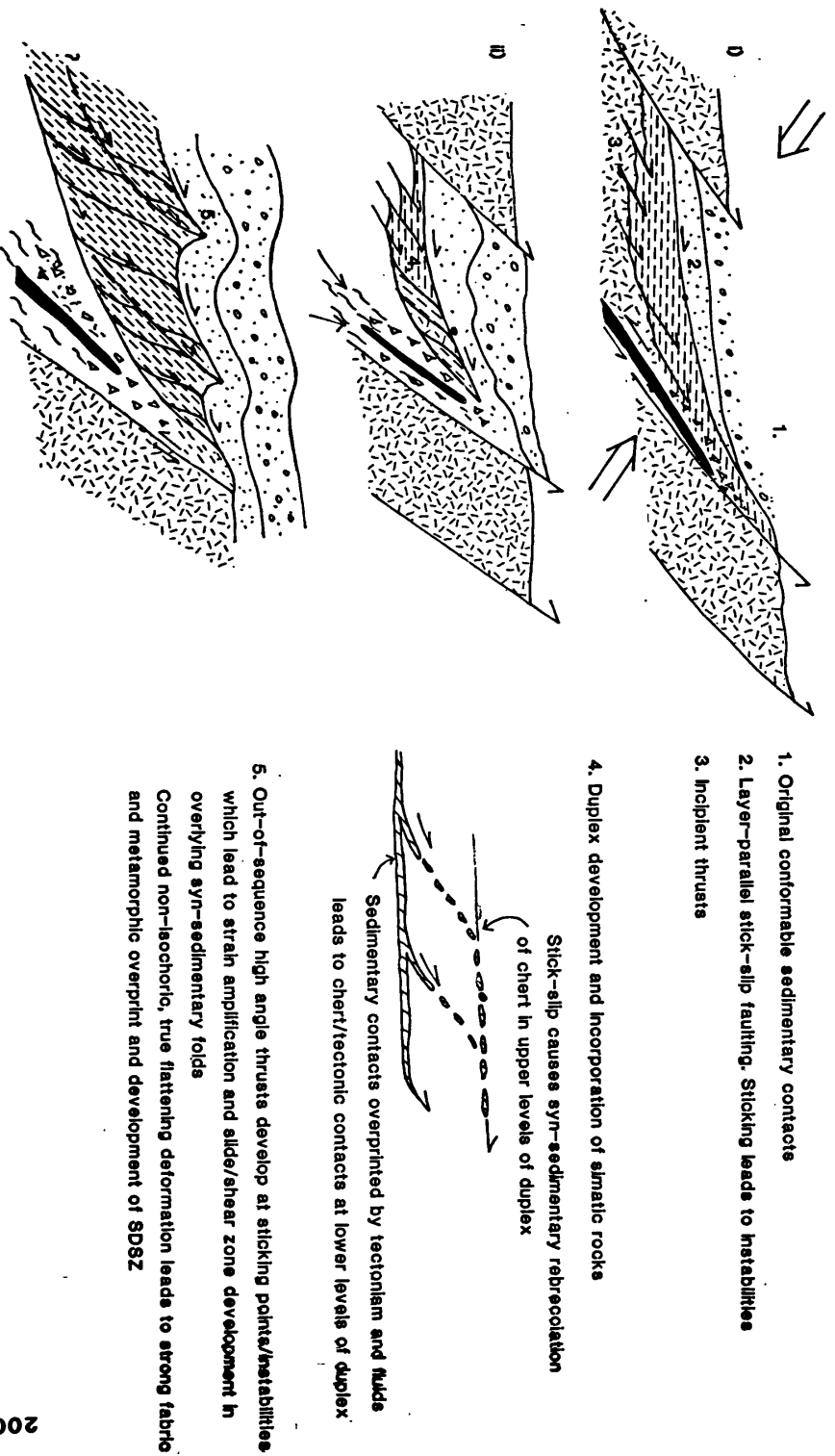


Fig. 6.8 Model for the evolution of the SDSZ

The veins themselves contain fragments of wall rock sediments, some of which are still attached to the wall rock (plate 7.4a), and suggest that hydraulic fracturing parallel to bedding was responsible for fluid flow along the horizon. The fragments are commonly tourmalinised and there are alteration halos of tourmaline that extend for up to 10cm around the veins. Minor sericite is sometimes seen at the immediate vein/wall margins. Occasionally, where the veins narrow towards their extremities, the tourmaline has penetrated the hanging wall sandstones for up to 30cm along cleavage planes (plate 7.4b). These textures suggest that the majority of the tourmaline was introduced along with the quartz during hydraulic fracturing and although most of it precipitated in the vein, it also passively infiltrated the surrounding wall rock. Similar hydraulic fracturing textures at the same mudstone/sandstone interface have been found in the hinges of D2 folds at SD # 4 and 5 Hills (plate 7.4c). The quartz veins are very white and contain secondary fractures filled with tourmaline. Although no gold grains were observed in this study, Kenyon (1973) reports the presence of gold in the fractured quartz vein, and sample grades increase dramatically (unpubl. data - AAC). It is likely therefore, that similar to ZR, the gold (and further) tourmaline were introduced subsequent to the quartz veining event.

Spatially associated with the SD # 3 deposit is a significant percentage of pyrite. At surface it is found pseudomorphed by boxworked goethite up to 50cm away from the veins and can locally make up to 50% of the rock. In the drillcore it is found in the sandstones immediately above the mudstone/sandstone contact (plate 7.4d). These are the 'placer' deposits of de Wit (1985) and Franey (1987). In thin section, although both the quartz and pyrite show evidence of fracturing, it is evident that the pyrite is generally very euhedral and also cross-cuts many quartz clast/matrix boundaries (plate 7.4e) and the D2₁ cleavage. Hence the pyrite is interpreted as a post depositional hydrothermal overprint. It is associated with tourmaline and patchy ankerite and sericite. Although no gold was found in the pyrite in this study, Kenyon (1983) did report the presence of gold flakes in the boxworked goethite. Muff and Saager (1979) also reported gold inclusions in fresh pyrite from Uitkyk sediments from the Mt. Robert deposit on the farm Amatava (Fig. 1.2). Hence at least some of the gold was probably introduced into the sediments with pyrite as well as with the tourmaline. Reconnaissance work by the writer at the Mt. Robert deposit indicates it lies on a fault plane that cuts the sediments and that all the pyrite is spatially associated with it. Although clear detrital pyrites do exist in the Uitkyk sediments it is likely that much of the pyrite at Mt. Robert is also hydrothermal in origin.

Overall, the SD # 3 deposit shows similar characteristics to the ZR deposit. It is found at the lithological contact between sandstones and mudstones which show evidence for *early* soft sediment deformation. The gold probably occurs, along with tourmaline, in late stage fractures that cut a pre-existing quartz (and tourmaline) vein and which was introduced during very late D2 brittle deformation along pre-existing planes of weakness. The deposit also plunges parallel to the main local fold axis. It differs from the ZR deposit in that the primary sulphide mineral associated with it is pyrite (which probably also contains gold inclusions), rather than arsenopyrite which is absent. The sub-surface extent and geometry of mineralization at depth is not known.

7.4 GREENWALDS (GW)

The deposit at GW, the third reef zone studied, lies at the eastern end of the SDSZ (Fig. 4.1, Table 6.1) and is also a fractured, quartz vein-hosted gold deposit. It differs from the other two reefs in that it lies entirely within sandstones and conglomerates and lies in the immediate footwall of an unusual bedding-parallel, tourmaline-bearing cataclasite. Similar cataclasites have been found at several localities in the SDSZ and they usually lie in high strain zones, commonly cutting and anastomosing through the early chert-breccia horizons. They form discrete, planar to thin tabular zones (3–15cm wide) that can be traced up to 75m along strike and have often been the focus of early prospecting in the area. The best 3D exposure is found at GW where a 12m x 8m plane has been exposed by the mining.

The cataclasites are characterised by very smooth, black planar bases with more irregular contacts with the hanging wall sediments, and they superficially resemble some reported pseudotachylytes (plate 7.5a and b, Sibson 1975, Maddocks 1987). Commonly, a discrete, well defined fragmental rock layer lies at the base of the zone and constitutes the main fault plane (plate 7.5c and d). The clasts in the breccia are predominantly quartz with subordinate, extremely small (<100micron) mica flakes that lie in a matrix of tourmaline. The bases are cut by conjugate sets of extension veins whose fibres plunge moderately to steeply SE, parallel to the regional lineation.

The cataclasites record a complex history of fluid flow that involved at least three distinct phases of metasomatism. The sandstones around the cataclasites are

silicified (plate 7.5b) and the related quartz is highly strained, fractured and cut by foliated sericite. The quartz, followed by the mica, both predate later tourmaline-filled fractures which are connected to the main fault zone (plate 7.5d). Some passive replacement of the mica by tourmaline is also seen (plate 7.6a). The anastomosing nature of the tourmaline-bearing fractures causes a spalling, or plucking, of the host rock, predominantly from the hanging wall, into the main fault where the clasts undergo intense grain size reduction (plate 7.5d and 7.6b). Commonly the clasts show quite good degrees of rounding and sphericity and resemble quasi-conglomerates seen in pseudotachylytes (Sibson 1975). In detail, the quartz fragments have ill-defined, almost cusped boundaries and the pieces do not fit back together in a neat fashion. SEM investigation reveals that these features are caused by tourmaline cross-cutting and replacing the quartz, mainly at clast margins (plate 7.6c and d). The majority of the matrix tourmaline is very fine grained and only locally are individual crystals seen. They do not show any evidence of deformation but occasionally show a slight alignment between clasts.

The textures seen are probably best explained by the concomitant cataclasis, movement along the fault and chemical interaction of the quartz-mica host rock with boron rich fluids. Hydraulic fracturing probably aided the cataclasis which was followed by rapid precipitation of tourmaline and sealing of the fault plane. Although no visible gold was observed in this study at GW, it is likely that the gold was introduced into the earlier quartz vein at this stage, as tourmaline stringers are also present in the fractured quartz vein (similar to ZR and possibly SD # 3 Hill. The tourmaline related deformation event post-dated at least two phases of fluid flow along the same horizon (quartz and mica) and at least one period of deformation which resulted in the development of the foliation seen in the sericite.

7.5 DISCUSSION

Although the different gold deposits found in the SDSZ exhibit their own unique set of geological characteristics, there are a number of important similarities /that allows their formation to be integrated into the evolution of the SDSZ as a whole. The most important similarities are; the D2 structural control, their association with zones showing evidence for repeated fluid flow and/or brittle deformation, and the gold-tourmaline association.

TABLE 6.2

SUMMARY OF FIRST ORDER OBSERVATIONS WITHIN THE SNYMANSDRIFT SHEAR ZONE

OVERALL SETTINGS OF SHEAR ZONE	
1. Footwall unit of sinistral rocks and BIF is allochthonous and bounded by faults or shear zones, unconformities and one disconformity.	
2. Upper and lower boundaries constrained and discrete, but lateral terminations are gradual and not discrete.	
DEFORMATION CHARACTERISTICS	
3. Deformation created a basal tectono-sedimentary melange showing a well defined gradation from tectonic processes to sedimentary processes both across and along strike. The sediments form part of the Ulitkyk Formation.	Unit may have behaved as an uplifted rigid block carried above a thrust.
4. Ulitkyk Formation sediments in the melange can be traced into an unconformity to the west.	Interpretation
5. Ulitkyk Formation in upper parts of zone and its edges shows evidence of 'soft sediment' deformation including syn-sedimentary faults and stratal disruption.	Deformation may be explained if shear zone represents a sequence of sediments deformed between two more rigid blocks. Lateral tips of zone are gradual, suggesting deformation and displacement are also decreasing laterally.
6. Variation observed in lithological contacts from sedimentary, to syn-sedimentary faults (including brecciation of pre-existing tectonites) to discrete tectonic breaks.	Sedimentation was occurring above a shallow dipping syn-sedimentary fault plane. The resulting package was subsequently rotated to a steeper angle.
7. At surface early syn-sedimentary features are ubiquitously overprinted by brittle faults and cleavage development at shear zone edges, and by brittle faults, intense fabric development and ductile deformation in the central most highly deformed parts of the zone. The sediments undergo prograde metamorphism to mid-greenschist facies. A progressive fault rock development sequence of cataclastic - c-type mylonite and possible mylonite is seen.	As a whole, sedimentation was semi allochthonous.
8. The sinistral lithologies in the footwall block undergo retrograde metamorphism to mid-greenschist facies in the shear zone.	Sedimentation and deformation were contemporaneous. The synsedimentary faults form part of an imbricate fan.
9. Fault bounded blocks or lenses of sinistral rock are found within the tectono-sedimentary package, interleaved with packages of mudstone/sandstone.	Variances explained in terms of stick-slip model of fault propagation and decollement along pre existing lithological contacts.
10. Microstructural studies of garnet from the zone revealed no gross overprinting of deformation at depth than seen at surface.	Sediments are undergoing burial due to increasing sedimentation and an overlying thrust sheet. A progressive outward migration of deformation mechanism "fronts" from the central parts of the shear zone towards its lateral tips.
11. Progressive homogenisation of different fault rock protoliths produces a sequence of greenschist facies phylionites.	There was a pre D2 metamorphic event. The retrograde nature of the metamorphism can not be used to infer structural throw.
12. Extensive quartz, quartz carbonate and quartz-tourmaline is found throughout the zone and they contain textures indicative of hydraulic fracturing. Extensive carbonate metasomatism accompanied development of melange. "Pseudo conglomerates" found in cherty tectonites. Syn sedimentary faults and soft sedimentary structures.	Sequence constitutes a deformed synsedimentary contractional duplex formed between melange and lithofacies A of the Ulitkyk Formation
13. Syntectonic metamorphism produced near equilibrium assemblages of sericite quartz-chloritoid +/- chlorite and chlorite quartz-chloritoid in the phylionites	Significant fluid flow and elemental transfer between lithologies and possibly out of the system must have occurred to result in chemical homogenisation.
14. Clast orientation analysis in the melange suggested flow had a local Wk values of 0.76	Zone was one of significant fluid flow throughout its history. Fluid pressures are inferred to have been close to lithostatic most of this time.
15. Reverse sense of shear obtained from deflected cleavage in minor faults within sediments (deflection caused by flexural slip). Extensional sense of shear obtained from clast orientation analysis.	Metamorphism in the shear zone reached mid greenschist facies, corresponding to temperatures between 350 and 450°C and less well constrained pressures of 2 to 3 kbar.

The strain path was probably one of deformation with volume loss in true flattening and involved components of pure and simple shear. Steepening and flattening of the zone may be caused by back-rotation and/or ramping imbrication by underlying faults.

Model invoked - SDSZ can be explained as a package of sediment deformed between two rigid blocks both being translated northwards. The melange can then be explained as a lag slide (Hutton 1980) formed on the upper surface of the lowest block. The melange may have its origin as a deformed disconformity.

The following observations can be made concerning the timing relationship of gold introduction.

1) In the centre of the SDSZ, gold introduction was synchronous with late D2 ductile deformation, but passively infiltrated areas of lower strain. Mineralization was tectonically reworked and downgraded.

2) At the edges of the shear zone, where penetrative, late D2 ductile deformation is absent and mineralization was associated with the late stages repeated brittle faulting.

These features could be explained as follows;

1) there were either two (or more) discrete phases of gold introduction into different parts of the shear zone, or,

2) there was one phase of gold introduction but the style of associated deposit and relationship to surrounding deformation was strongly controlled by the spatial variation in D2 strain *prior* to the introduction of mineralization. Thus if mineralization was introduced late in D2 it would have been synchronous with D2₃ deformation in the central and deeper parts of the shear zone, but synchronous with D2₁ deformation at the edges. This is the preferred interpretation and is the most consistent with the combined structural and metamorphic observations.

The tectonic reworking of the mineralization at ZR-MMV replaces the associated assemblages with ones identical to those seen in the shear zone away from gold-rich areas (chapter 6). From this it is concluded that the mineralization event was simply a transient fluid flow and associated alteration phenomenon that was superimposed on the ambient dynamothermal metamorphism prevalent in the SDSZ at the time. Hence the gold was introduced during the peak metamorphic conditions recorded in the shear zone (300–400°C and <3 kb chapter 6), during late D2 deformation.

Many different sources for Archaean gold bearing fluids have been proposed in the past (see MacDonald 1986 for review). Although detailed isotopic, fluid inclusion and mineral chemistry studies are needed before the source, or the nature of the mineralising fluid and possible factors causing mineral deposition in the SDSZ can be characterised, some pertinent observations can be made. Combined with the field

and petrographic data, a simplified paragenetic diagram of vein related minerals in the SDSZ (Fig. 7.2) shows that the mineralization was preceded and succeeded by at least several generations of syntectonic quartz +/- carbonate +/- pyrite veins. This indicates that there were in fact many episodes, or pulses, of fluid flow through the zone during D2 and that this assemblage may represent the ambient syntectonic pore fluids. Therefore, the most unusual feature about the mineralization event is simply that the fluid concerned contained different minerals. This difference is probably best explained by one of two models; either there was a change in the processes (e.g. mineral reactions) at the source providing the fluids, or, there was an input of fluids from a different source. At present, the second model is favoured. The ambient pore fluids probably represent some form of 'metamorphic' fluid whereas the gold-tourmaline-arsenopyrite-ankerite association (coupled with the presence of tourmaline in local granites - Hall 1908) suggests the related fluids may be derived from a granitic source. Hence the shear zone must have 'tapped', at least once, a magmatic reservoir.

Because hydraulic fracturing seems to have been important during the emplacement of the fluid, it is inferred that the speed of fluid flow and the (associated metasomatic reactions) seen were probably geologically very rapid processes and may have occurred in a matter of seconds or minutes (Gold and Soter 1975). The virtually instantaneous drop in fluid pressure that would accompany hydraulic fracturing, would probably cause phase separation (or 'boiling') in the fluid as well as precipitation of quartz. This is likely to have been just as an important control on precipitation of the gold as decreases in temperature (associated with the ascendancy of the fluids into the upper crust) or changes in fluid chemistry caused by host rock interaction or even mixing with fluids from different sources. Phase separation has been reported in several Archaean gold deposits from veins that show textures interpreted to be caused by hydraulic fracturing (Wood *et al.* 1986, Groves *et al.* 1984). Drummond and Ohmoto (1985) and Spooner *et al.* (1987) have discussed in detail some of the important effects boiling and hydraulic fracturing may have on ore deposition.

In summary, the geological relationships within there was only one phase of gold introduction into the SDSZ, and it occurred late in D2. The mineralizing fluid was overpressured and travelled by hydraulic fracturing from deeper, perhaps a magmatic source, to higher crustal levels but was only one of several transient fluid flow events.

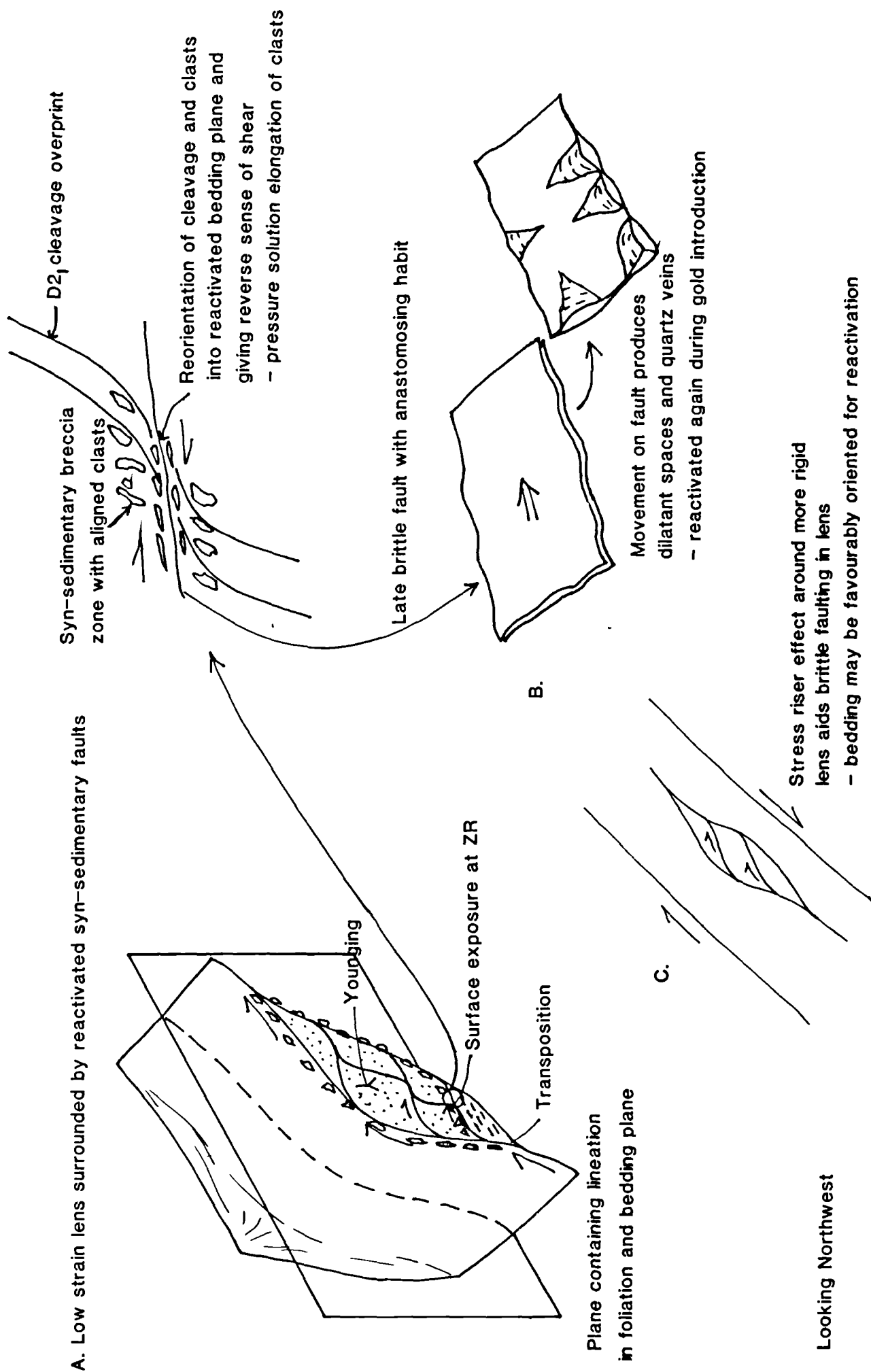


Fig. 7.1 Schematic interpretation of the geology and mineralisation at Zandrivier

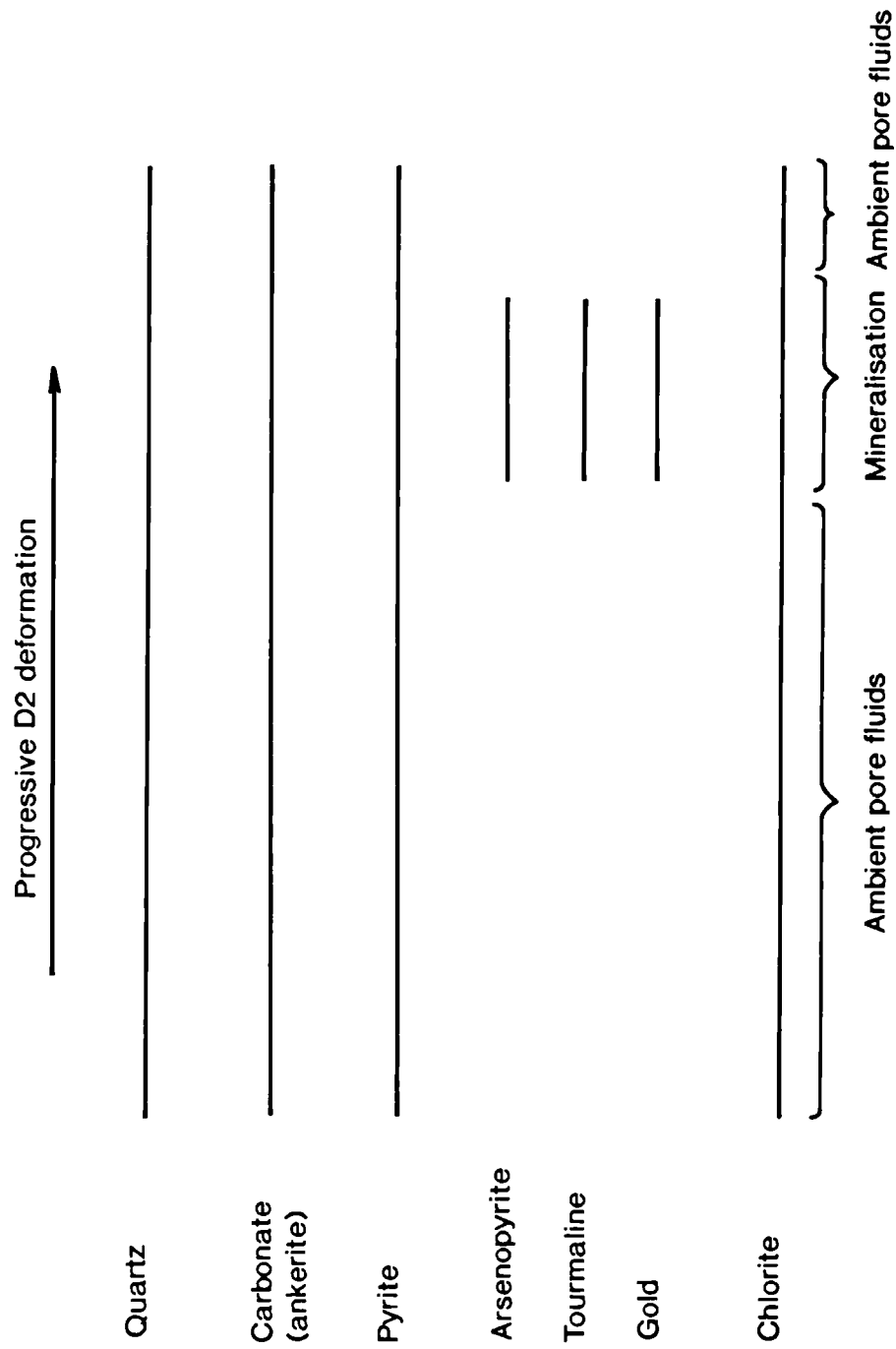


Fig. 7.2 Simplified paragenetic sequence of vein minerals in the SDSZ

Plate 7.1 Borehole mineralisation - reef textures.

- 7.1a. Gold and tourmaline stringers in intensely fractured quartz vein from surface reef zone at ZR. Both the gold and tourmaline, which appear to have been co-precipitated, post date the quartz veining. FoV , Reflected Light (RL).**
- 7.1b. Euhedral pyrite and minor ilmenite overprinting a deformed breccio-conglomerate. DDH ZR2 80.95, FoV , RL.**
- 7.1c. Reef zone from DDH ZR1 showing euhedral arsenopyrite overprinting horses of wall rock and entirely within the quartz vein.**
- 7.1d. Fluid inclusion wafer showing abundant fine grained predominantly secondary inclusions. FoV 4mm .**

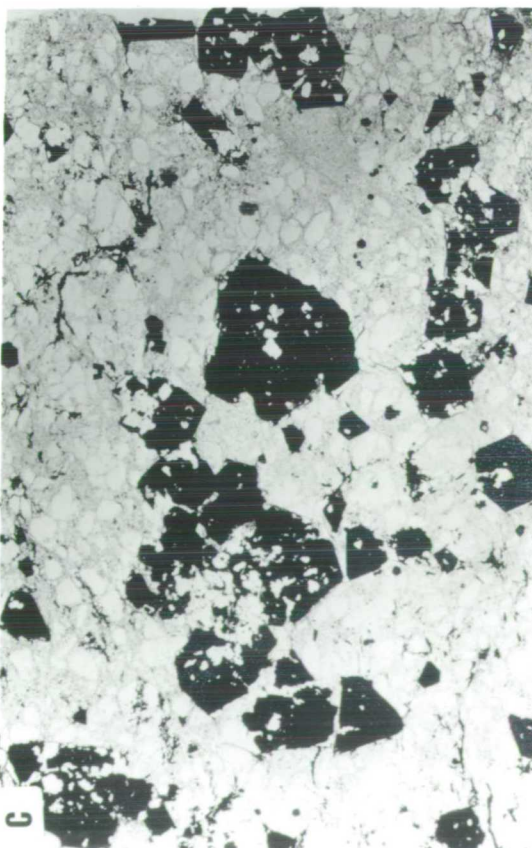


Plate 7.2 Borehole mineralisation - ore minerals and paragenesis.

- 7.2a. Intense cataclasis of arsenopyrite with later pyrite infilling the fractures. Such cataclasis marks the first stages of deformation of the reef zones. DDH ZR1 97.76, FoV 9mm, RL.**
- 7.2b. Further example of fracturing of pyrite showing chlorite (dark grey) infilling fractures. DDH ZR1 97.76, FoV 7mm , RL.**
- 7.2c. Evidence for fracturing of arsenopyrite accompanied by its replacement by quartz. Euhedral arsenopyrite outlines are still visible but are locally mimicked by quartz. DDH ZR1 96.42, FoV 4cm, PPL.**
- 7.2d. Gold grain occluded in arsenopyrite grain suggesting the two minerals were co-precipitated. DDH ZR1 97.76, FoV 4cm, RL.**

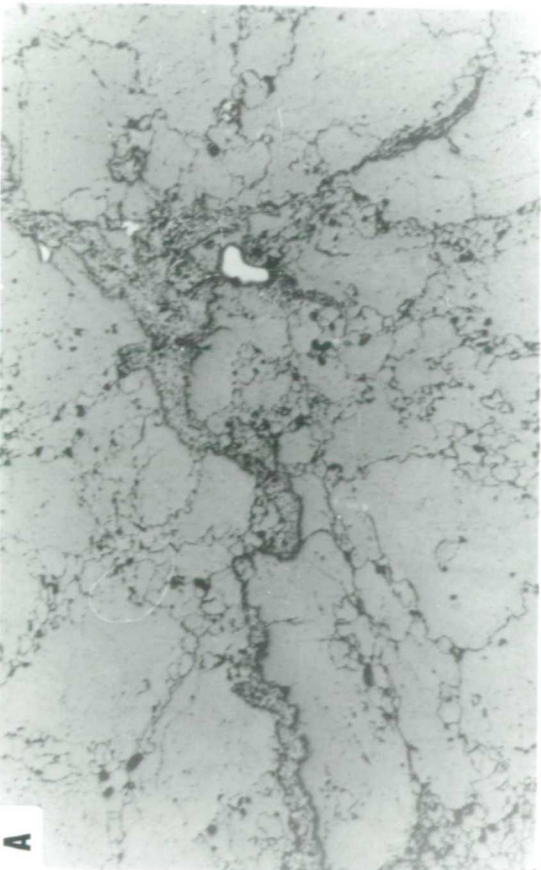
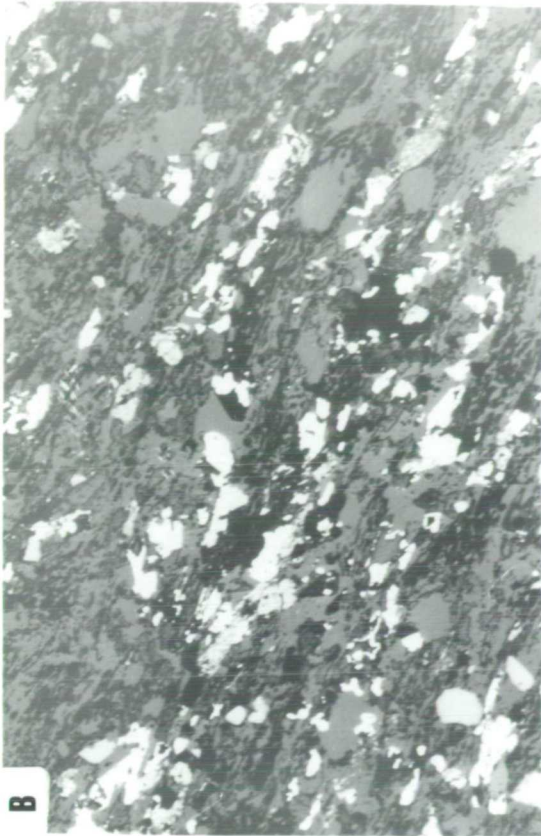


Plate 7.3 Textures in gangue minerals.

- 7.3a.** Thin section showing tourmalinised wall rock (left), quartz vein filled with euhedral tourmaline laths and euhedral arsenopyrite (right). Such a texture indicates all three minerals were co-precipitated during hydraulic fracturing. DDH ZR1 96.72, FoV 1.5cm, PPL.
- 7.3b.** Edge of reef vein in DDH ZR3 showing transposition by D2 fabric. The vein is being tectonically reworked by the same fabric it cross-cuts and is a syntectonic fluid flow phenomenon. DDH ZR1 98.58.
- 7.3c.** More advanced deformation of reef veins in DDH MMV 5 showing transposition. The quartz and carbonate show evidence for grain size reduction, partly obliterated by annealing in their coarser grained centres. FoV 2cm, XP.

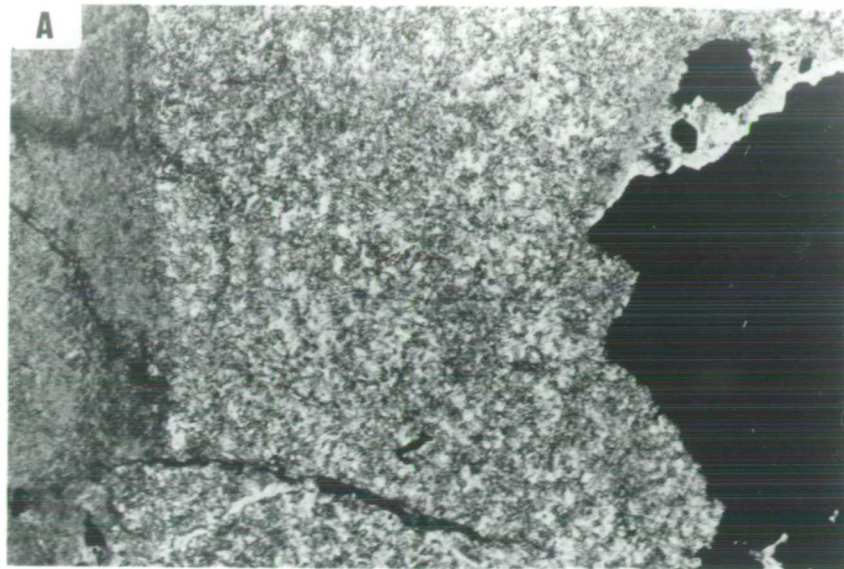


Plate 7.4 Reef zone textures from SD # 3 Hill.

- 7.4a. Surface exposure of reef zone showing a 75cm wide quartz vein at contact between hanging wall sandstone and footwall mudstone. Note the large black, tourmalinised horse of sandstone that is still attached to the hanging wall (arrowed) suggestive of bedding-parallel hydraulic fracturing.**
- 7.4b. Tourmaline infiltrating sandstone along cleavage planes and related to a quartz vein containing tourmalinised horses.**
- 7.4c. Hinge of main D2 fold on SD # 4 Hill showing bedding-parallel hydraulic fracturing and tourmalinisation of horses of wall rock.**
- 7.4d. Euhedral pyrite in coarse grained sandstone from DDH SD1 (location on Fig. 6.2). The sulphide distribution is not related to sedimentary structures and in general appears to overprint the sediment.**
- 7.4e. Thin section showing euhedral pyrite overprinting the sediment matrix and cross cutting quartz grains. Such textures are ubiquitous and are unlikely to have survived sedimentary transport. DDH SD1 67.36, FoV 9mm, PPL.**

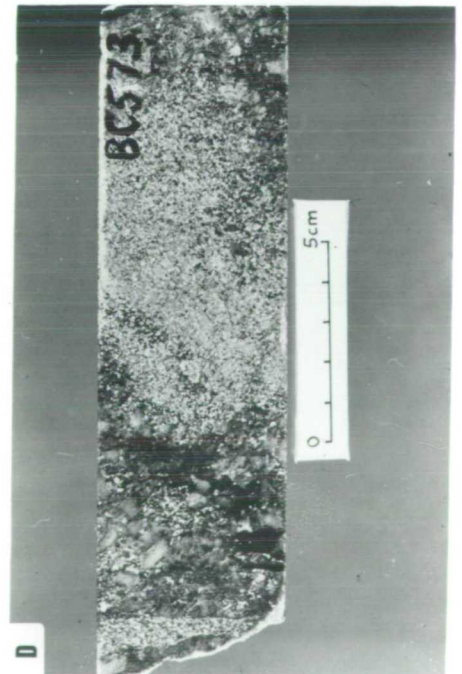
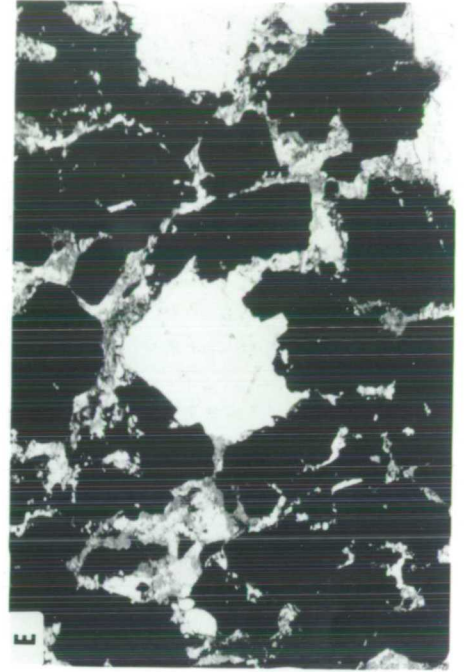


Plate 7.5 Tourmaline-bearing cataclasites (i) field observations.

- 7.5a.** Small remnant sub-vertical exposure of tourmaline-bearing cataclasite showing the smooth, black, planar base. The base is cut by at least two generations of fibrous extension veins whose fibres plunge parallel to the regional D2 lineation. GR X 2 657 150, Y -36 750.
- 7.5b.** Cross-section of cataclasite showing fragments of quartz spalling-off the hosty rock into the black, tourmaline-rich matrix. Relatively undeformed sandstones are seen at the base of the photograph and can be traced into the quartz rich areas, suggestive of silicification, which predated the tourmaline and brecciation. Note the sub-parallel tourmaline-bearing fractures in the wall rock. Sample GW 1. GR X 2 657 100, Y -37 700.
- 7.5c.** Hand specimen of cataclasite from Greenwalds showing quartz (silicified sandstone) cut by mica and tourmaline bearing fractures. The main zone of cataclasis, at the base, is now only 1cm wide and is quite planar. Sample GW 1.
- 7.5d.** Whole thin section showing fragmental layer in detail. Note in upper part of photograph how quartz grains are spalling off into a grey matrix which is sericite. The quartz was therefore post-dated by sericite influx. Note the rounded nature of many clasts in the fragmental layer and the planar nature of the boundaries. Sample GW 1, FoV 2.5cm, PPL.

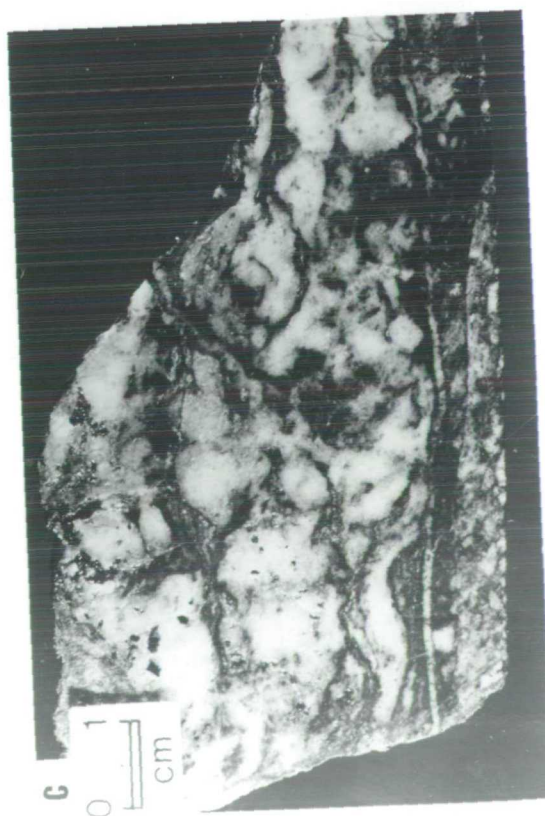
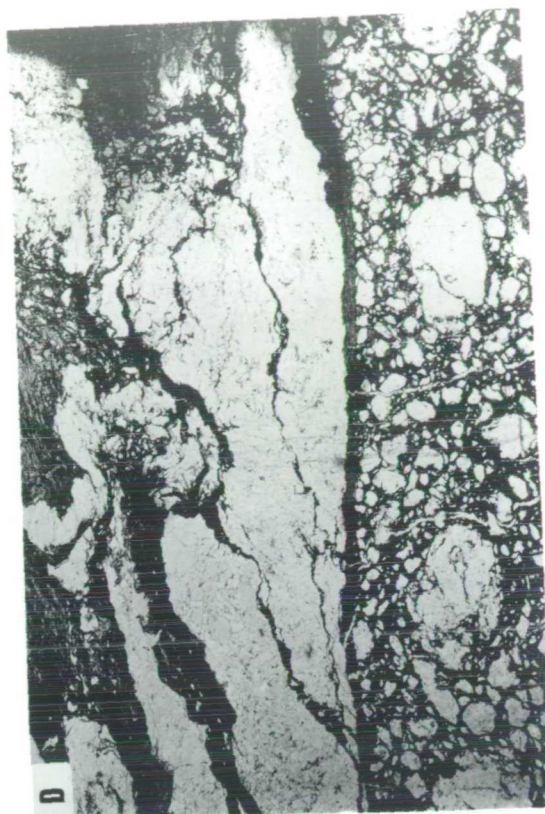
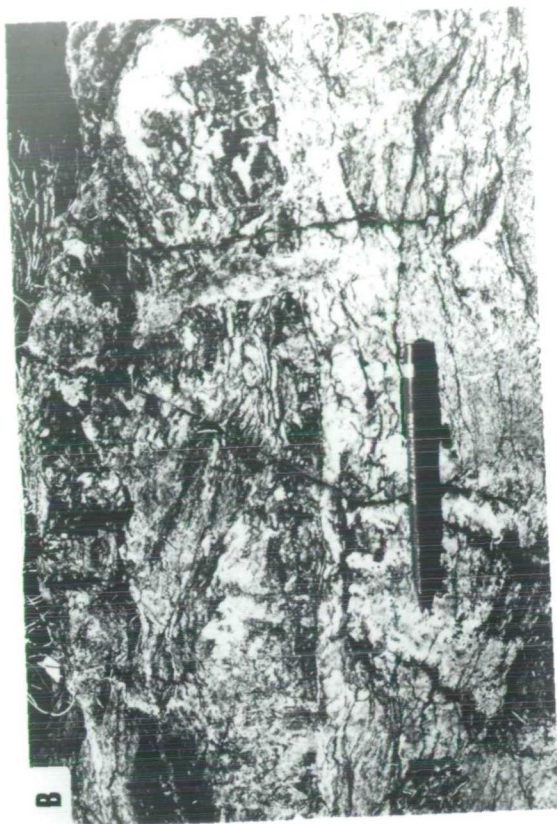
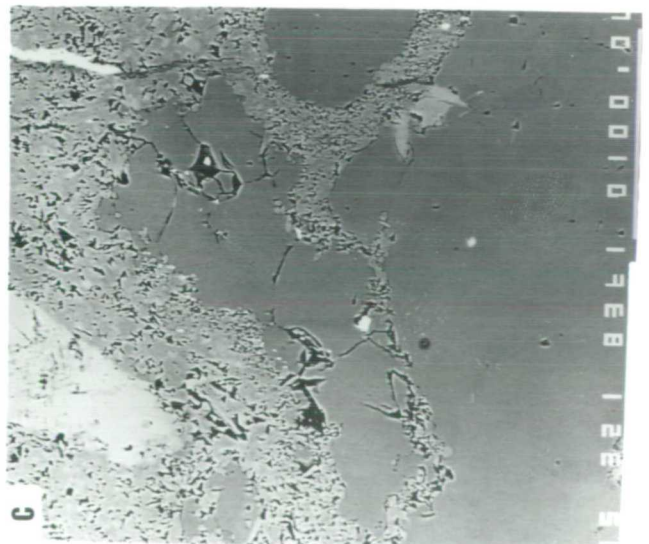


Plate 7.6 Tourmaline-bearing cataclasites (b) microstructures.

- 7.6a. SEM micrograph of mica associated with wall rocks of cataclasite showing evidence for pseudomorphing and replacement of sericite (light grey) by later tourmaline (dark grey). The mica is fractured locally. Sample GW 1, scale bar 10 microns.
- 7.6b. View of quartz clast in main cataclasite layer showing intense breakage and spalling-off of smaller fragments. A later stage quartz vein is also present. Sample GW 1, FoV 1cm, PPL.
- 7.6c SEM micrograph showing cataclasis of quartz grains. Note how the smaller fragments do not fit back together in a regular fashion. A lighter grey fragment of sericite is also present. Sample GW 1, scale bar 100 microns.
- 7.6d. SEM micrograph of small quartz fragment being replaced by euhedral tourmaline suggesting that cataclasis and fluid-rock interaction were simultaneous. Sample GW 1, scale bar 10 microns.



CHAPTER 8

OTHER D2 SHEAR ZONES IN THE MT MARE AREA AND A DYNAMIC CONCEPTUAL MODEL OF A MINERALIZED FAULT ZONE

8.1 INTRODUCTION

This chapter is primarily concerned with field observations from other D2 shear zones in the Mt. Mare area and also discusses their relative timing relationships. The Kuschke (KSZ) and Hollandsdrift (HDSZ) Shear Zones contain units of an unusual schistose-gneissose fuchsite carbonate-quartz lithology apparently formed by tectono-metasomatic processes. The shear zones are also characterised by a combined mechanical/chemical breakdown of phyllonitic tectonites and, similar to the SDSZ, an early D2 gradation from tectonic to sedimentary processes is observed in both zones. Combined, the field observations documented in this study indicate the Mt. Mare area is a foreland propagating, emergent imbricate fan thrust complex which developed during progressive D2, northwest verging thrusting. The thrusting resulted in stratigraphic repetition and the successive stacking or accretion of allochthonous sheets of simatic (and related) rocks and small molasse-filled piggy-back basins. The deformation style may be analagous to 'foreland basin' environments found in Phanerozoic collisional terrains.

A dynamic conceptual model of a D2 mineralized shear zone in the Mt. Mare area has been constructed by interpreting two of the D2 shear zones as different spatial levels of one fault zone. The implications of several features of the model are discussed, some of which suggest the depth of formation of some Archaean gold-quartz vein systems may have been overestimated in the past;

8.2 THE KUSCHKE SHEAR ZONE

8.2.1 Introduction

Within the allochthonous simatic rocks above the Kraal Thrust (chapter 2.3) is a NE-SW striking SE steeply dipping high strain zone called the Kuschke Shear Zone (KSZ - de Wit 1985), which separates a lower unit of predominantly volcanic rocks from an overlying unit of intrusives and associated xenoliths of BIF (Fig. 1.3 - chapter 2.4). The zone is marked by a variety of fault rocks including various greenschist facies chlorite-quartz-ankerite and talc chlorite-carbonate phyllonites derived from the simatic rocks, variously coloured 'cherts' and an unusual fuchsite carbonate-quartz lithology interpreted as a tectonite. The geology of the western end of the KSZ is shown in Fig 2.2.

The upper boundary of the KSZ is generally well defined and lies immediately beneath the train of BIF xenoliths. The lower boundary is less well defined and is marked by a gradual increase in strain gradient over 20 to 50m, passing northwards from the extrusive lithologies. The gradient is most noticeable in the gradual flattening and elongation of pillow lavas and associated ocelli (plate 8.1a), to produce a foliation and lineation. Strain analysis on the ocelli by de Wit (1985) suggested they had been extended by a minimum 180% parallel to the lineation they define in the XY plane of the strain ellipsoid, extended by over 25% in the YZ plane and flattened by over 70% in the XZ plane. These strains, which correspond to a strain ellipsoid ratio of 10.1:4.6:1, represent minimum values only because the original textures are commonly completely transposed and the lithologies are converted to a sequence of chlorite-quartz + - ankerite phyllonites within the central and most highly deformed parts of the shear zone.

The strained ocelli define a lineation which, to the east of locality D (Fig. 2.2 - and along the entire exposed length of the KSZ to the NE) plunges down dip of the tectonic fabric and parallel to fibres in small extension veins. To the west of locality D there is a marked change in strike of the KSZ and along this section there is evidence for localized crenulation of the D2 fabric and development of a horizontal lineation eg. locality E (Fig. 2.2). However, this section of the KSZ does not cut the Kraal Thrust but appears to meet it at a branch point (Fig. 2.2). The parallelism of the KSZ to the SDSZ (and Kraal Thrust) and their similar lineation orientations indicate the KSZ is also a predominantly dip slip D2 shear zone. The section west of locality D probably represents a (late?) D2 lateral ramp structure.

Although no reliable sense of shear criteria were observed, three lines of evidence suggest the KSZ is a major structural break and is probably a thrust; first, geochemical evidence (chapter 3.5) suggested that the extrusives were derived by fractional crystallisation and/or partial melting from the intrusives and thus were originally above the intrusives in the postulated early simatic complex. Secondly, the extrusives are at a lower metamorphic grade (low to mid greenschist facies) than many of the overlying intrusives (mid greenschist to lower amphibolite, Table 3.2), both units interpreted to have been metamorphosed during the 'M1' metamorphic event (chapter 3.2). Thirdly, at locality B (Fig. 2.2) a small low strain lens has been identified lying within the KSZ which contains a fuchsite-carbonate-quartz tectonite that carries a shallow dipping (25 to 40° SW) foliation. This foliation (possibly a 'D1' fabric - chapter 2.4) becomes transposed by the steeply dipping D2 foliation around the edges of the lens. Similar tectonites have been reported by de Wit (1982, 1986) in the Barberton Mountain Land where they were assigned to early regional extensional tectonism. Although the original sense of shear along this early tectonite is not known, its presence suggests the D2 KSZ may have exploited an earlier structural horizon and is thus a reactivated fault.

8.2.2 Fuchsite +\ - quartz +\ - carbonate (FQC) tectono-metasomatic fault rock

An unusual fuchsite +\ - quartz +\ - carbonate (siderite and\ or ferroan dolomite) lithology (FQC) occurs as an integral part of the KSZ fault rocks. It consists of schistose to gneissose (up to 2cm wide), anastomosing, alternating layers of foliated phyllonitic fuchsite, variably deformed quartz and carbonate, plus minor chlorite. A gradation exists in the predominant minerals present from fuchsite and quartz layers only, through to fuchsite and carbonate layers only. The lithology varies in overall width from 1.2m up to 30m and occurs as layers and lenses of varying length (Fig. 2.2).

Plate 8.1b is an example of the alternating layers of foliated fuchsite and quartz. In detail, (plate 8.1c) it appears many of the quartz layers are quartz veins which, although trending sub-parallel to the foliation in the fuchsite layers, have also locally cross-cut that foliation. Indeed several generations of veining can be seen as evidenced by the cross cutting relationships of separate quartz veins. However, some of the veins have also been ptymatically folded axial planar to the fabric in the fuchsite and, in the field, a gradation is seen from such textures to areas in which the original quartz veins have been significantly deformed and possibly recrystallized to produce a very parallel banded gneissose phyllonite-mylonite (plate 8.2d).

A second example of the relationships seen in this fault rock is shown in plate 8.2, a fuchsite-carbonate end member. Multiple generations of carbonate (with only minor quartz) veins (up to 70cm long in outcrop) can be seen to lie sub-parallel to the tectonic foliation, defined by green fuchsitic layers. A minimum of 6 distinct and sequentially cross-cutting veining events are present in this specimen. However, earlier formed veins eg vein A (plate 8.2), show *progressively more signs of deformation than later formed veins* eg vein B. These later veins show evidence of fibrous morphologies suggestive of hydraulic fracturing and extension normal to the tectonic fabric (Ramsey and Huber 1983). Inclusion trails of wall rock material are not common in these veins. Vein C, one of the least deformed veins, has a texture suggestive of 'open space filling' in its centre.

Deformation of earlier formed veins has apparently proceeded by boudinage (ie brittle microfracturing) and/or pinch-and swell (eg vein D), followed by continued ductile flow and 'smearing out' of the veins parallel to the D2 fabric until they become virtually unrecognisable having been almost completely transposed eg vein E. In many cases the transposition is incomplete and earlier veins are only preserved as low strain lenses in the tectonite (eg vein F). Occasional shear folding of the veins axial planar to the D2 fabric (vein G) is also seen.

Thin section work shows the majority of the carbonate in deformed veins is very fine grained (< 20-30 microns) and shows no evidence of preferred orientation. Coarser grained, fibrous carbonate in less deformed veins occurs as cores surrounded by mantles of the fine grained carbonate. Grain size reduction by crystal plasticity is inferred to be the dominant deformation mechanism in the recrystallization of the carbonate although the role of initial microcracking is difficult to judge (cf. Hadizadeh and Rutter 1985, Simpson 1987). The lack of preferred orientation and fine grain size of carbonate in the highly strained veins suggests much of the recrystallized carbonate may have behaved in a superplastic fashion and that DAGBS may have been operative (Behrmann 1980). This is supported by experimental deformation of carbonate (calcite) at mid greenschist facies conditions, similar to those inferred for the KSZ (Schmid 1983)

The flow of large quantities of hydrothermal fluid through the KSZ implied by the large number of veining events present has the potential to cause major geochemical changes on the surrounding fault rocks. In similar lithologies in the Barberton greenstone belt, Duce and Hanor (1987) demonstrated the fuchsite was probably derived from the intense hydrothermal alteration of ultramafic lithologies. It is likely the fuchsite in the KSZ has a similar origin, representing a highly deformed and altered ultramafic precursor

lithology.

At locality E (Fig. 2.2), an old gold working is present in the KSZ. The reef that was mined is a large quartz vein which itself forms an integral part of the FQC tectonite. It lies sub-parallel to the tectonic fabric but has also been deformed to form pinch and swell structures. Thus, in a similar fashion to the gold-bearing veins in the SDSZ (chapter 7), the gold-bearing fluid appears to represent only one of many pulses of fluid that passed through the shear zone during D2 tectonism.

8.2.3 Faulting mechanism

The textures described above suggest that *cyclical* hydrofracture dilatancy, metasomatism and layer parallel ductile shear deformation were responsible for the development of the FQC tectonite. The extreme length of many of the veins within the lithology and the sub-parallelism of later, less deformed veins to the D2 fabric, suggests the veins did not form at a high angle to the foliation (in classic *en echelon* tension gash fashion) and were subsequently rotated into parallelism with the fabric. Rather, it appears the majority of the veins initially formed sub-parallel to the fabric due to multiple pulses of fluid which periodically migrated through the shear zone causing hydraulic fracturing normal to the tectonic fabric. Each fluid flow event was apparently succeeded by layer parallel shear and progressive deformation of the resultant (and pre-existing) vein or veins *before* the next pulse of fluid passed through the zone. This implies some form of 'stress cycling' of σ_1 and σ_3 occurred during tectonite development and that fault movement was episodic.

Sibson (1981a and b) has investigated the conditions under which such cyclic hydrofracture dilatancy could occur adjacent to an intermittently slipping fault. He suggested that it was a low differential stress phenomenon and for it to occur at anything but the shallowest levels in the crust, the pore fluid pressure had to be greater than hydrostatic in and around normal and strike slip faults, and greater than lithostatic in and around thrust faults. In 1959, in what is regarded as a classic paper, Hubbert and Rubey first stressed that thrusting mechanisms require that fluid pressures are close to or exceed lithostatic pressure in and around the fault. They were intrigued by the apparent mechanical paradox that the size of the large thrust nappes involved in mountain building processes, were so great that the force required to overcome friction at their bases in order that they could have moved along the observed low angle fault planes would have been

greater than the limited strength of the block itself. Hence the block would break up internally by cataclasis rather than slide. They proposed an elegant hypothesis that high fluid pressures in and around the thrust plane would reduce the frictional constraints (in accordance with the law of effective stress, Fig. 8.1) so the thrust block could effectively 'glide' on the fault plane.

Although modifications of the initial hypothesis have been made, and many other thrusting mechanisms proposed (see review by De Bremaeker 1987), high fluid pressures have been reported from thrust faults in various oilfields (Hubbert and Rubey 1959) and active accretionary prisms, including the direct intersection of an overpressured thrust fault by the DSDP (Anderson 1981, Davis *et al.* 1983, ODP Drilling Group 1987). Therefore it would seem that high fluid pressures aiding thrusting is an attractive hypothesis and one for which field evidence should exist. The FQC lithology is interpreted as direct evidence for high fluid pressures beneath the allochthonous thrust sheet of simatic rocks as predicted by the Hubbert and Rubey (1959) hypothesis. Similar zones of fibrous extension veins have been reported from below thrust sheets in the Spanish Hercynides (Price 1977) and the Briançonnais region of the European Alps where they can be traced over distances up to a hundred kilometres along strike (Cunningham 1988).

Such a complex mixture of brittle failure under inferred conditions of high fluid pressure, and high strain ductile flow, is very suggestive of stick-slip behaviour during formation of the FQC. Furthermore, the repeated fluid flow events are also suggestive of seismic pumping (Sibson *et al.* 1975) and a combination of this hypothesis and the Hubbert and Rubey (1959) hypothesis is used here to propose a stick-slip, earthquake-related faulting mechanism for the development of the observed textures in the FQC.

During pre-failure stress rises that occur as precursors to earthquake activity, fluids can be 'sucked' into resultant dilatant cracks in and around the fault zone (Fig. 8.2a). When failure occurs along the fault, the dilatant zones collapse due to the sudden release of stress and can thus pump fluid away from the epicentral area, usually towards the surface. This fluid will be under high fluid pressure and is most likely to migrate along the fault plane itself where it can aid slip in accordance with the law of effective stress (Fig. 8.2b) (Sibson *et al.* 1975, Sibson 1981, 1986, White and Baxter 1987). The speed of fluid flow would probably be in the order of metres per second, possibly on a par with the rate of seismic wave propagation (Gold and Soter 1985, De Bremaeker 1987).

have a podiform morphology with a leading and trailing edge. Such 'pods' have been called fluid domains by Gold and Soter (1985). As this domain travels under high fluid pressure along the zone, dilatancy collapse at the rear of the domain would cause hydraulic fracturing at the leading edge thus creating an increase in permeability and allowing the fluid to continue its passage (Fig. 8.2c). During the fluid flow, two possible scenarios can be envisaged; the dilatancy collapse at the trailing edge may be complete thus leaving little or no trace of its passage (Etheridge *et al.* 1984) or the fluid may encounter the correct P-T-x conditions to precipitate minerals. Such vein formation would cause self-sealing of this hydrothermal system. Spooner *et al.* (1987) have suggested H₂O-CO₂ phase separation during fluid pressure fluctuations associated with hydraulic fracturing can self-sustain such fluid flow and cause mineral precipitation over considerable vertical distances.

Regardless of which of these two possibilities occur, it is evident that at the instant of hydraulic fracturing, in accordance with the law of effective stress, the overlying thrust sheet above the domain will be almost floating on a cushion of fluid and will be free to move. This is in accordance with the seismic valve model of fault movement (Sibson 1981) and the fluid diffusion-earthquake hypothesis of Nur (1972) and Scholz *et al.* (1973).

A corollary of the fluid domain concept is that it is obvious that the whole of the thrust sheet is not being uplifted simultaneously; only small segments at a time are moving. An elegant analogy for this type of movement was proposed by De Bremacker (1987) who studied seismic wave propagation along thrust planes. He noted that in the seismogenic regime, thrust sheets do not move as single entities because it takes a finite length of time for the initial earthquake movement to be transferred along the length of the fault zone. As the seismic wave travels along the zone De Bremacker (1987) proposed it uplifted the overlying sheet in an analogous way to shaking a rug from one end thus causing a rippling effect. This type of movement would be a direct consequence of the passage of a fluid domain through a fault zone as proposed here for the KSZ. The movement of the fluid would itself be analogous to squeezing an orange pip between ones fingers. Thus it is seen that layer normal hydraulic fracturing is very quickly followed by layer parallel shear, the main feature observed in the FQC tectonite. Post- vein formation ductile shear may have actually been *promoted* by the introduction and precipitation of the carbonate into the KSZ as this mineral was inferred to have been extremely ductile at the prevalent temperatures in the KSZ (see above). The carbonate introduction could therefore be considered as an example of strain softening induced by hydrothermal mineral precipitation.

8.2.4 Progressive D2 deformation of the FQC tectonite

Once formed, the FQC has locally undergone two different styles of subsequent, but still D2, deformation. As shown in Fig. 2.2, the FQC occurs as discrete layers or lenses within the shear zone passing along and across strike into chlorite-quartz phyllonites. Often at the edges of these lenses, or occasionally entire lenses eg locality F (Fig. 2.2), the FQC can be traced directly into a sequence of laminated green and white cherts (plate 8.3a), and grey, white and yellow-brown cherts which are cut by stylolites that trend sub parallel to the D2 fabric elsewhere in the KSZ. The 'chertification' process is therefore probably the result of a syn-D2, more passive fluid flow and silicification event (and possible recrystallisation). The orientation of the stylolites suggest D2 compression of the cherts continued after their formation. Importantly, all the textures seen in the brecciated chert horizons along syn-sedimentary faults in the SDSZ (chapter 4.8) are seen *in situ* in these cherts in the KSZ. This is consistent with the interpretation of the breccias as fault rocks that had undergone a long and complicated faulting history (chapter 4.8.4).

The second style of later deformation and alteration to have affected the FQC can be seen on a traverse along strike to the west of locality G (Fig. 2.2). The FQC (and also the surrounding phyllonites) can be traced into very banded chlorite-siderite-quartz rich phyllonitic lithologies which become very difficult to distinguish from sedimentary material (siltstones and shales). At locality H (Fig. 2.2) large blocks of BIF from the hanging wall have been caught up in this material and undergone brittle-ductile deformation, resulting in the spalling apart of the blocks to produce breccias (plate 8.3b). Plate 8.3c shows one such block disintegrating *in situ* to produce a ductile, almost slurry-like breccia protruberance into the surrounding matrix. Such a feature is suggestive of fluidization (Reynolds 1954, Murphy 1985). Although not well exposed, it appears in the field that the fault rocks in the KSZ grade directly along strike to the west into sediments of the Uitkyk Formation (Fig. 2.2). It is suggested such features represent an example of an early to mid D2 gradation from purely tectonic to purely sedimentary processes on the scale of the shear zone but has also been overprinted by D2 cleavage development, similar to that observed in the SDSZ.

It appears that continued, but more passive infiltrative fluid migration through the FQC and surrounding phyllonites after their formation, resulted in the extreme hydration and mechanical and chemical breakdown of the phyllonitic material to form silt and clay grade material. Such processes could have caused significant strain softening within the KSZ and allowed it to widen. Such widening would lead to the incorporation of blocks of

BIF from the hanging wall into the shear zone where they could undergo brittle-ductile deformation and fluidization to form the observed breccias. Subsequent deformation has locally lead to the orientation of the chert clasts and formation of c-type mylonites. It is important to emphasize that initially the processes of phyllonite breakdown and brecciation occurred entirely within the KSZ, towards its lateral tip.

8.3 OTHER SHEAR ZONES

8.3.1 The Kraal Thrust (KT)

As noted in chapter 2.3 and shown in Fig. 1.3 and 2.2, the Kraal Thrust, which also strikes NE-SW and dips moderately SE, separates Uitkyk Formation sediments to the north from overlying, older simatic rocks to the south. Only a few lineations were observed associated with the KT and dip slip movement only is inferred.

Although marked as a discrete line on the geological maps, the Kraal Thrust is difficult to define accurately in the field because its boundaries are quite diffuse over a 30 to 50m wide zone. Passing downwards, the overlying volcanic rocks become increasingly deformed towards the thrust zone and the original pillow lavas and tuffs quickly become highly strained and are converted to chlorite-quartz +\ - carbonate phyllonites. The lowermost boundary of the KT is the most difficult to identify because the fault rocks again begin to look very similar to mudstones. Indeed, it is clear the thrust has been passed only when large lenses of coarse, cobble to boulder grade detrital material are encountered eg locality J (Fig. 2.2). The detritus, which occurs as a channel-like feature in mudstones (which dip under the Kraal Thrust), is predominantly clast supported, poorly sorted and shows an extremely heterogeneous clast size variation (plate 8.4a) upto 5m³ in size. The majority of the clasts are sub-angular to very angular and many of them show evidence of *in situ* brecciation (plate 8.4b and c).

Although not discussed in chapter 2, these coarse sediments and the surrounding mudstones grade directly into Lithofacies D of the Uitkyk Formation (horizontally stratified conglomerates). The coarse detritus, itself derived from as greenstone provenance (plate 8.4a, Table 2.3) is probably best explained as pre-cursory olistostrome material (Elter and Trevisan 1973) derived, syn-D2, from the allochthonous simatic rocks during tectonism,

laid down in a small foreland depository and subsequently overridden by the thrust sheet.

Fig. 1.3 shows that this area of sediments is underlain by a further unit of simatic rocks. The contact between the two is not exposed but must be either an unconformity or fault contact.

8.3.2 The Hollandsdrift Shear Zone (HDSZ) and the origin of siderite nodules

The HDSZ is a moderate to high strain zone which sharply truncates the top of the Uitkyk Formation in the central west of the study area (Figs. 1.3 and 2.2) and trends approximately WNW-ESE and dips moderately to steeply SSW. Lineations, defined locally by deformed ocelli and the preferred geometric orientation of clasts, are moderately to steeply plunging SE or S. Strain analysis on the ocelli by de Wit (1985) indicated the HDSZ had been affected by minimum strains of 85% extension, 9% extension and 50% shortening in the X:Y:Z directions respectively of the strain ellipsoid and this equated to an ellipsoid ratio of 3.7:2.2:1. The Uitkyk Formation sediments which underlie the HDSZ are structurally upward facing and the right way up but young towards and dip under the HDSZ. In turn, the HDSZ, the upper boundary of which is not exposed, dips under an overlying package of older, metamorphosed simatic rocks and BIF. Combined, these features indicate the HDSZ is probably a D2, slightly dextral oblique thrust fault. It may be a lateral ramp structure to the more regional trending NE-SW trending fault zones already documented (SDSZ, KSZ, KT). The overlying unit of simatic rocks and BIF is therefore allochthonous and is the highest exposed structural lithotectonic package in the Mt. Mare area. The BIF is isoclinally folded and forms a south-west steeply plunging, upright D2 antiform (de Wit 1985, Fig. 1.3).

A variety of lithologies are found within the HDSZ including fault rocks, relatively undeformed lenses of simatic rock and fine to coarse grained Uitkyk Formation sediments (Fig. 2.2). De Wit (1985) proposed that the lenses of sediment had been structurally incorporated into a larger schist zone and were preserved as low strain lenses. However, there are a number of field relationships which suggest this may not be the case in many instances.

The basal lithology within the shear zone varies from a chlorite-quartz phyllonite in the south-east to a gneiss with fuchsite quartz-carbonate tectonite in the north west identical to the FQC described from the KSZ (chapter 8.1.2). The FQC in the HDSZ is not different from the FQC in the KSZ, as with other sericite zones in the Mt. Mare area,

the HDSZ contains evidence suggesting there is a combined mechanical chemical breakdown of the fault rocks within it.

At locality K (Fig. 2.2), the FQC grades upwards over 3 to 4m from a predominantly fuchsite-quartz rich 'end-member' to a fuchsite-siderite-chlorite 'end member' (plate 8.5a). Towards the top of this unit the siderite layers become noticeably deformed, predominantly by pinch-and-swell development, within the (inferred) slightly more ductile phyllonitic matrix (plate 8.5b). Progressive breakdown of the matrix and continued pinch-and-swell development produces less foliated mud grade material containing isolated, elongate lenses and 'pods' of remnant siderite 'swells' (plate 8.5b).

A further example of this breakdown is shown in plate 8.5c. Here, the D2 tectonic fabric is being *truncated* at a low to moderate angle by a poorly sorted, angular, breccio-conglomerate lithology which also carries the D2 fabric as a spaced cleavage. The clasts consist predominantly of banded chert and BIF, and along strike range in size up to large blocks 10m long by 1m wide. These blocks show evidence for *in situ* brecciation and appear to be the parent material for much of the clastic material. The overall appearance of the breccio-conglomerates is again very suggestive of fluidization (plate 8.5c). There are also a significant number of siderite nodules in the breccio-conglomerate and examples are seen of pinch-and-swell structures breaking down directly to form them (plate 8.5c). These nodules are identical to those seen in Uitiky Formation sediments outside the HDSZ (Table 2.3, plate 2.5) and it is proposed they are remnant 'swells' formed by progressive breakdown of D2 tectonites. Similar truncation relationships were observed along the contacts of most of the coarse clastic layers within the lowermost parts of the HDSZ (Fig. 2.2) and they are thus not tectonically-bound lenses of sediment (de Wit 1985) but are probably breccio-conglomerate tectonite horizons. At locality L (Fig. 2.2) *in situ* brittle break up of fuchsite layers is seen at the base of a sandstone layer and this produces cobble grade, angular fuchsite clasts again identical to clasts in debris flows of the Uitiky Formation (Table 2.3). Murphy (1985) has also recorded simultaneous fluidization phenomena and foliation development in sediments from the Irish Caledonides.

Passing 50 to 150m upwards through the HDSZ, similar zones of breakdown of the phyllonitic material are common and these are interlayered with lenses of less deformed simatic rock. Overall the distribution of lithologies and style of deformation is very heterogeneous and it can be termed a melange zone (Fig. 2.2). However the number of recognisable phyllonites and simatic rocks also decreases upwards and the zone *grades* into a series of slaty to phyllitic mudstones which are interlayered with and completely surround

lenses and layers of conglomerate and sandstone of the Uitkyk Formation. This gradation is considered to represent a further example of tectonic processes grading into sedimentary processes on the scale of the shear zone. The sediments, which are similar to lithofacies B and A respectively (Table 2.2) contain cross-beds indicating a south-west younging direction (Fig. 2.2). The mudstones are commonly isoclinally folded axial planar to the D2 fabric (plate 8.5d) and have clearly been considerably deformed. There is locally intense syn- to late D2 chloritoid development in the lower 100 to 200m of the zone.

The HDSZ is best explained as a zone of deformed Uitkyk Formation sediments which represents a 'piggy back basin' that initially formed and rode above a high level, possibly emergent, thrust. This thrust itself directly overrode similar aged sediments (Fig. 2.2). As D2 progressed the basin was progressively deformed, steepened and tightened as it was overridden by a package of syntectonic rocks and BIF. The sedimentary conglomerate layers within the mudstones probably represent debris flows and channel-fill deposits that cut across a developing alluvial plain.

The siderite nodules in Uitkyk Formation debris flows *outside* the HDSZ were apparently formed from the breakdown of D2 tectonites. This strongly suggests material (in an inferred fluidised state) may have been periodically *expelled* from the HDSZ and that the same fault overrode that detritus during progressive D2 deformation. Cyclical stick-slip fault activity and periodic pulses of fluid flow as inferred from the FQC tectonite, would have considerably aided such forceful expulsion. Such expulsion would also explain two features of the sedimentology of the Uitkyk Formation; first, the rounded and 'resedimented' nature of some of the coarse detrital sediments (Table 2.2) as they may have been initially worked close to or within an emergent fault zone. Fluidization has been shown both experimentally (McCallum 1985) and from field observations (Shelnutt and Noble 1985, Delaney *et al.* 1987) to be an efficient 'rounding agent' of clastic material. It negates the need for the clasts in the Uitkyk Formation to have travelled considerable distances to become rounded. Secondly, it is also consistent with the observed coarsening upward sedimentary cycles and inferred allogenic influence on sedimentation (Fig. 2.1). Syn-D2 carbonate fluid release from the fault zones into a near surface environment would also explain the cross-cutting carbonate alteration visible in the Uitkyk Formation (plate 2.5).

8.4 GRANITE-GREENSTONE RELATIONSHIPS

The Pietersburg greenstone belt marks a part of the linear contact between two different granitoid terranes. Although chemical data is sparse, it appears tonalite-trondjemite lithologies dominate to the north of the belt and more potassic granites to the south (SACS 1980, Franey 1987). The linear feature is a major crustal structure in the northern Kaapvaal craton and can be traced for 250km north-east of the study area to the Sutherland-Giyani greenstone belt (Figs. 1.1, 1.2). In the south-west of the Pietersburg belt the structure is marked by a major sinistral strike to oblique reverse slip fault called the Willemse Shear Zone (WSZ - de Wit 1985) that has offset the Mt Mare and Rietvley areas by approximately 10km horizontally and by 1 to 4km vertically (de Wit 1987). This shear zone is not well exposed but truncates the D2 thrusts in the Mt. Mare area and is therefore interpreted as a 'D3' fault.

Granite-greenstone contacts are generally poorly exposed in the south-west Pietersburg belt but large scale cross-cutting relationships noted by previous workers (Hall 1908, Willemse 1938, SACS 1980, de Wit 1987, 1988) both north and south of the WSZ, suggest in general the granites are intrusive into the belt. SACS (1980) report the age of some of these granites as being between 2.3 and 2.6 Ga, and is the only published time constraint on belt evolution. However, despite these cross-cutting relationships there is evidence to suggest at least some D2 deformation also post-dated granite intrusion perhaps indicating on a gross scale that intrusion was in part a syn- to late D2 event.

In the northern most part of the Mt. Mare area good exposures of tonalite (-trondjemite?) are found and they have a variable character ranging from pegmatites to aplites. Complex cross-cutting relationships exist between these different types suggesting the magmatic processes were polyphase and thus may have operated over an extended period of time. The granites contain large xenoliths of simatic rock up to 500m long and 100m wide but both are locally cross-cut by a NE-SW trending foliation. Narrow, 3m wide quartz-sericite schist-gneiss zones cut the granite. These were first observed by Hall (1908) and are interpreted here as plastic shear zones. One such zone marks the contact between the granite and an overlying unit of simatic rocks (Fig. 1.3). It is clearly a significant tectonic break and I propose to call it the Witkop Fault. Although no lineations were observed, the fabric orientation in the granite is sub-parallel to D2 elsewhere in the Mt. Mare area and is inferred to be of the same age. Therefore some D2 deformation apparently post-dated granite intrusion. Similar conclusions were reached by de Wit (1988) in the Landsberghoek area (Fig. 1.2). If granite intrusion was syn- to late D2 then D2 is

also constrained to have occurred between 2.3 and 2.6Ga.

As noted in chapter 5, the southern, uppermost boundary of the SDSZ is not exposed but rock chips from government water boreholes showed granitic rocks to be present within 500m south of the last surface outcrop of the SDSZ. The nature of this contact is not known but geophysical data indicates it may also be a fault contact. De Beer and Stettler (1986) investigated the deep structure of the Pietersburg belt by gravity measurements some 20km to the northeast of the Mt. Mare area. The location of the survey, their results and interpretations are shown in Fig. 8.3.

The results indicated to de Beer and Stettler (1986) that the Pietersburg belt is only a shallow crustal feature which, despite an across strike thickness of 7km at surface is less than 3km thick (Fig. 8.3). They interpreted the the basal low angle discontinuity between the simatic and granitoid rocks as a shear zone thus implying this part of the belt as a whole is allochthonous. The southern, uppermost granite-greenstone contact is also a low-angle (dipping 20° SE) planar contact for much of its length but has a marked change in dip to 60° SE close to the surface. It is proposed here this contact is the along strike extension of the upper boundary of the SDSZ and is probably a major D2 thrust. It is implicit in de Beer and Stettlers work that many of the steeply dipping faults seen at surface must become shallower at depth. If this is the case then the basal granite-greenstone contact may be the decollement layer for the D2 faults.

8.5 RELATIVE TIMING OF D2 FAULTS

A total of 4 observed and 2 inferred D2 faults/shear zones have been documented in this study (Table 8.1). Of these, at least 5 are thought to be thrusts or thrust related features formed during approximately NW-SE shortening across the Mt. Mare area. The relative timing between these faults must now be established and this, combined with other geological information documented in this study, may aid interpretations of the overall geologic environment during D2

The following observations can be made concerning relative timing of faults,

The HDSZ and its overlying allochthonous unit of simatic rocks and BIF truncates and terminates the a pa ka f Uitkyk Formation sediment which themselves are

in part conformable with the tectono-sedimentary melange within the SDSZ (Figs. 1.3, 2.2, 4.1). The same sediments also unconformably overlie a second allochthonous unit of simatic lithologies and BIF. The Kraal Thrust separates this second unit from underlying Uitkyk Formation sediments and both this fault and the Kuschke Shear Zone die out along strike into the same sediments.

These features and also the intimate relationship between tectonism and sedimentation documented in this thesis suggests that at least early in D2 the faults did not move sequentially in a strict foreland or hinterland propagating sequence. Rather it appears all the faults must have been operative more or less simultaneously. Indeed, if stick slip fault behaviour was common as proposed here for the SDSZ, KSZ and HDSZ, it is likely that 'out of sequence' faulting would have occurred. However, the final lithological distributions and related gradations from tectonic to sedimentary processes do imply the HDSZ has moved the furthest of the exposed faults and the KT the least. Thus, during D2 the Mt. Mare area is considered to represent a supracrustal, foreland propagating, or trailing, emergent imbricate fan thrust complex.

The proposed thrusting has resulted in the stacking, or accretion, of successive and alternating packages of simatic (and related) lithologies and detrital sedimentary rocks which represent fault- and unconformity-bounded early D2 piggy-back basins carried above the emergent thrusts. The alluvial debris flows and associated coarse sediments within these basins are manifestations of contemporaneous deposition ahead of the migrating thrust sheets which in turn subsequently overrode those same sediments later in D2. Such shortening across the area has resulted in the migration towards each other of two granitoid terrains in which some magmatic activity overlapped with D2. This geologic history is schematically outlined in Fig. 8.4.

There is insufficient evidence in the Mt. Mare area to infer an overall tectonic setting for D2. The style of thrusting, inferred alluvial nature of sediments and relationship between faulting and sedimentation do show strong similarities to foreland basin environments such as those in the European Appennines (Ricci Lucci 1985) and Pyrennes (Valdez 1985 - see also Allen and Blackwood 1985). It is proposed here the Mt. Mare area did form in such an environment but it is as yet unclear how this relates to possible terrane collision, subduction zones etc. Until the effects and geometry of D2 deformation are better understood on a regional scale, it will be difficult to propose specific environments of formation for the simatic and related rocks themselves.

8.6 A DYNAMIC CONCEPTUAL MODEL OF A MINERALIZED D2 FAULT ZONE

8.6.1 Introduction

The field data compiled in this study from the separate faults and shear zones within the Mt. Mare area can be used to construct a simple dynamic conceptual model of a mineralized shear zone within the Mt. Mare area. The model is shown in Fig. 8.5.

The key to establishing fault zone models is to determine the changes in macroscopic faulting style and associated deformation mechanisms that occur at different crustal depths for the appropriate host rocks (Sibson 1983, Maddocks 1987). It is therefore necessary to establish whether or not for a given set of fault rocks if their *current* spatial distribution reflects a 'snapshot' of a developing shear zone in which all the fault rocks were active simultaneously, or; if the distribution reflects a tectonic telescoping or juxtaposition of fault rocks from different crustal levels and which were not active simultaneously (eg Sibson *et al.* 1980, White and White 1983).

The model presented here is based primarily on observations from the Snymansdrift Shear Zone and the Kuschke Shear Zone and hinges mainly on two previous interpretations; first, the gradation from tectonic processes to sedimentary processes and associated mechanical/chemical breakdown of tectonites to form fluidised clastic material suggests the current spatial distribution of fault rocks, *on the scale of the Mt. Mare area*, may reflect close to an original spatial distribution in mid to late D2. Clearly within some of the zones tectonic telescoping has occurred eg the low strain lens centred on locality K, SDSZ (Fig. 4.1). Secondly, the Kuschke Shear Zone could possibly represent a deeper level expression of the SDSZ. The rationale for this interpretation comes from the fact that the FQC tectonite is converted, syn-D2, into a series of variously coloured and laminated cherts (chapter 8.1). Similar cherts are found as breccias in the syn-sedimentary fault breccias found within the SDSZ. Compare plates 4.12b and 8.3a. Therefore at depth within the SDSZ these chert breccias may be part of an FQC tectonite and associated phyllonite zone not unlike the KSZ.

8.6.2 Description of model

The model represents a mid- to late D2 fault zone which is characterized by cyclical slip on veins at all levels and is assumed to develop at the base of the greenstone belt.

Although drawn as a narrowing downward zone, it is likely the lower levels forms part of a broader schist zone. This movement is intimately associated with episodic fluid flow and has resulted in the development of a complex hydrothermal plumbing system and release of fluids from depth into a near surface environment. This 'linked' tectono-metasomatic system was driven by earthquake-related fault movement at depth, probably related to a complex interplay of magmatic, metamorphic as well as tectonic processes.

Movement in the lower levels of the zone is characterized by cycles of hydraulic fracturing and fluid flow and high strain ductile flow and this is transferred upwards by a 'seismic valve' mechanism into brittle and syn-sedimentary deformation close to the palaeosurface. As the fluid travels across and interacts with the tectono-sedimentary interface, fluidization, tectonite break down and melange development occur. Fluid release into the overlying sediments results in significant carbonation and silicification.

Creep between discrete brittle faulting events was primarily controlled by ductile flow of phyllonites at depth and incompletely lithified sediments near surface. In the middle and lower levels of the zone such flow is accompanied by infiltrative metasomatism involving an intragranular fluid phase. This aids the progressive textural, chemical and metamorphic homogenization of different precursor lithologies to produce a sequence of hydrated, extensively strain softened phyllonites. A complex juxtaposition of lithologies develops at these levels primarily related to deformed sediments being 'dragged down' into the zone and deformed simatic rocks being uplifted. Much of this movement occurs along discrete faults, many of which are localized along pre-existing sedimentary contacts. Subsequent silicification of these contacts and also of deeper level tectonites produces cherts which themselves can be subsequently uplifted to near surface and brecciated along syn-sedimentary faults.

Of the many pulses of fluid flow only one was carrying gold, together with arsenopyrite and tourmaline perhaps reflecting a magmatic fluid input into an otherwise metamorphic fluid regime. The release of fluids into the upper crust has the potential to cause significant fluid mixing. Subsequent tectonism of the resultant orebody has preserved mineralization in lenses or pods of low strain.

8.7 DISCUSSION

Several features of the fault zone model, which are based on observations made throughout this study, have implications for studies of mineralized Archaean shear fault zone hosted gold quartz vein deposits and perhaps also the deformation of greenstone belts in general.

1) It is clear that many such studies have been greatly influenced by models of fault zones in *continental* crust whose rheological behaviour is believed to be predominantly controlled by the mechanistically defined brittle-plastic transition in quartzo-feldspathic rocks (eg Sibson 1983, 1986). The combined field and microstructural observations from the Mt. Mare area has shown that crystal plasticity of quartz has played a minor role in the deformation history and therefore the model presented here can not be compared directly to pre-existing models. The main operative deformation mechanisms throughout D2 were brittle and diffusive mass transfer (fluid flow) processes which affected very variable mineral assemblages of quartz, amphibole, feldspar and phyllosilicates. Only crystal plasticity of carbonate was important in the FQC. Therefore, although the zone is, on a descriptive basis, macroscopically brittle-ductile it can not be equated to the mechanistically defined brittle-plastic transition. It is more likely to have formed in the 'brittle-semi-brittle' field (White and White 1983, Scholz 1988). Many studies of Archaean shear hosted gold deposits have not recorded in sufficient detail the deformation mechanisms operative in the shear zone itself but have implicitly equated hydraulic fracturing (as recorded in veins) and ductile (but non-crystal plastic) flow in the surrounding rocks with the brittle-plastic transition. Hydraulic fracturing, although more common in the upper crust, is essentially depth independent as it relies only on the 'ratio' of fluid pressure to lithostatic pressure (Sibson 1981, Phillips 1973). Therefore the depth of orebody and fault rock formation may have been overestimated on this basis.

Problems with depth estimates may be further compounded in many areas by the possible misinterpretation of the typical greenschist amphibolite facies metamorphic mineral assemblages in undeformed simatic rocks in greenstone belts as regional burial metamorphic assemblages. Many of the assemblages are not pressure diagnostic and could be the result of early 'ocean floor' metamorphism' (chapter 3.2).

2) It is the vertical depth over which the gradation from tectonic processes at depth to sedimentary processes at surface occurs which is in fact one of the more poorly constrained parameters of the model presented here. Metamorphic petrology (chapter 5)

suggested that pressure estimates were poorly constrained but that a *maximum* of 3kb was attained in the SDSZ. This corresponds to a depth of between approximately 9–12km and, for the proposed temperature range of 350–500°C, a vertical geothermal gradient of between approximately 25–55°C. This implies a considerable amount of overburden has been removed. However, this is not consistent with distance over which the gradation from tectonic to sedimentary processes and breakdown of tectonites occurs in the SDSZ, HDSZ and KSZ. Furthermore, both chloritoid and clinozoisite can form at much shallower levels (Deer *et al.* 1979) than 3Kb. Consider therefore the effects of changes in inferred geothermal gradients on the depth of fault rock formation.

Published fault zone models generally assume normal geothermal gradients of 25–30°C existed in the crust (Davis 1983, Sibson 1977, 1986). For the SDSZ this implies a depth of burial of between 11 and 20km. This is considered unreasonable because the Uitkyk Formation sediments outside the SDSZ were only at upper zeolite–lower greenschist facies during M2 (150–250°C – chapter 2.5). This implies a lateral geothermal gradient to the centre of the SDSZ of between 50° and 100°C. A similar vertical gradient may not be unreasonable if the effects of the following two factors are considered; first, the documented fluid flow has the potential to disturb ‘normal’ geothermal gradients and cause them to be enhanced in areas of upwelling fluid. Similarly if widespread magmatic activity involving high level granitoid plutons was a syn-D2 phenomenon, geothermal gradients may again have been abnormally high on a regional scale. If such a gradient existed in the SDSZ this would imply a depth of between 3.5 and 10km, or 1–3kb maximum. Such a high temperature gradient would also explain the observed high vertical strain gradient (chapter 5.1), and therefore the postulated 3–7Km depth for the model is not unreasonable.

The effects of high geothermal gradients have generally not been considered but clearly have the potential to displace upwards or cause ‘telescoping’ of the brittle-plastic transition or seismogenic–aseismogenic boundary to higher crustal levels than is inferred in existing fault zone models. It is vital that further work in the Mt. Mare area (and other greenstone belts) attempt to quantify vertical temperature gradients that existed during D2 and more tightly constrain the pressure and depth conditions of fault rock formation. Recent deeper drilling within the SDSZ below DDH ZR 1 may prove invaluable in testing the hypotheses presented here.

3) The ore-bearing fluid which precipitated gold was essentially only one of many syn-tectonic fluid flow events within the shear zone, implying either a change in source or change in source process (chapter 7). The orebodies (and other earlier formed veins) have

been reworked at their edges and downgraded by subsequent tectonism. The orebodies have therefore been *preserved as low strain lenses* in the shear zone. It is the writers personal experience and is implicit in much of the literature (although generally not specifically stated), that such a feature is common to many shear-zone hosted gold deposits. Orebodies die out because they have been tectonised around their edges. This would explain the commonly observed sub-parallelism of orebody shoots to the main lineation direction in the surrounding fault rocks eg Zandrivier deposit; Morning Star Mine, Western Australia; Dalny Mine, Zimbabwe; Kolar Schist Belt, India (Viswanath *et al.* 1960), Red Lake greenstone belt, Canada (Andrews *et al.* 1988). It is therefore essential in exploration and mining to understand the size, geometry and controls on formation and subsequent deformation of these lenses. One problem in determining such features arises in phyllosilicate-rich shear zones where diffusion accommodated grain boundary sliding may have been an important deformation mechanism. There is unlikely to be significant textural differences, such as grain size variation, in relatively low and high strained areas. Furthermore, the size of a lens could conceivably be greater than the mine scale and therefore may not be recognized. Thus it is necessary to obtain information from the surrounding fault rocks away from the mine and also from the edges of orebodies. Although mineralization may appear late in the history of the low strain lens, it could have been postdated by significant but unrecognised deformation on a larger scale.

In summary, the depth of formation of Archaean mineralized shear zones and also depth of burial of some greenstone belts is considered to be poorly constrained and may have been consistently overestimated in the past. Further work should therefore be concentrated on;

- 1) Discrimination of possible ocean floor vs regional metamorphic vs dynamothermal metamorphic mineral assemblages
- 2) Identification of the operative deformation mechanisms in macroscopically brittle-ductile or ductile styles of deformation.
- 3) The effects of non-quartzo-feldspathic controlled rheology and abnormally high geothermal gradients on fault zone models
- 4) The controls on syntectonic fluid flow and orebody formation, the role of hydraulic fracturing (and associated H₂O-CO₂ phase separation).

Fig. 8.1 The mechanical role of fluids and their relationship to faulting

The principal mechanical effect of fluids in the earth's crust is embodied in the simplified law of effective stress (Fyfe *et al.* 1978). It is held that if a given volume of rock is subjected to an external stress σ_n , then for a given pore fluid pressure, p , the effective stress, σ_n^i acting on the rock is given by;

$$\sigma_n^i = \sigma_n - p$$

Thus the pore fluid pressure reduces the normal stresses acting on a rock volume on an arbitrary plane. However it has no direct effect on the shear stress but will modify the commonly used Navier-Coulomb criteria of brittle failure in rocks as follows;

$$T = C_o + \mu_i(\sigma_n - p)$$

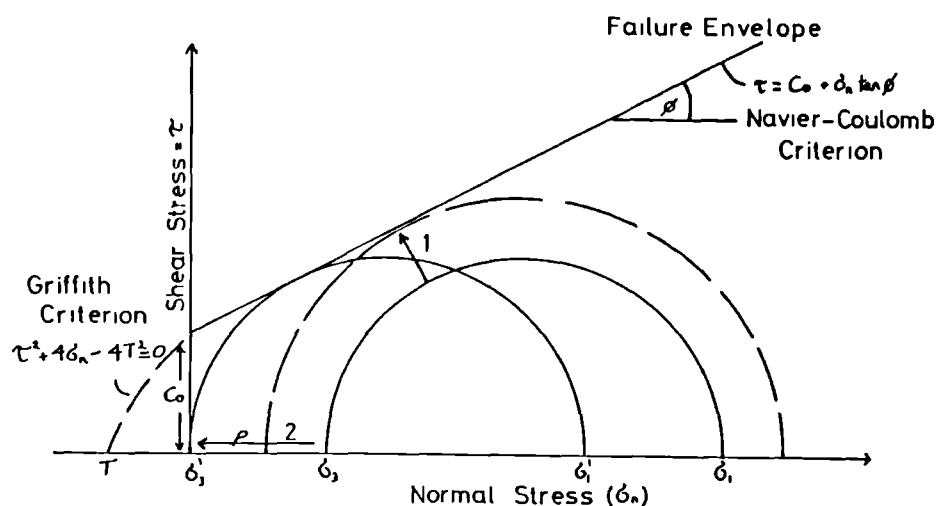
where T is the shear stress acting on a potential failure plane, and σ_n is the normal stress acting on the same plane, C_o is the cohesive strength of the material and μ_i , the coefficient of friction.

This relationship can be expressed on the Mohr Circle diagram as shown in Fig. 8.1a. The effect of a relative increase in the pore fluid pressure in a rock is to move the Mohr Circle towards the brittle failure envelope and render the rock more susceptible to cataclasis. If the pore fluid pressure increases sufficiently then hydraulic fracturing will occur. This condition is satisfied when;

$$\sigma_3 - p = t$$

where σ_3 is the least principal stress and t is the tensile strength of the rock (a negative quantity). One of the implications of this criterion is that sufficiently high fluid pressures can induce brittle failure at any level in the earth's crust, although it is probably most common in the elasto-frictional regime.

Sibson (1981b) has summarised the evidence and models proposed for fluid flow accompanying seismic faulting. He noted that transitory high fluid pressures were associated with modern earthquakes and suggested that crack-seal and hydrothermal mineral veins associated with ancient faults (showing evidence for incremental mineral deposition) may have been related to seismic activity. Sibson also suggested there may be a cause and effect relationship between fluid flow and faulting that could be explained in terms of two end-member models (Fig. 8.1b). In the first model fluid is 'sucked' into dilatant fractures that grow as elastic strain accumulates, to be rapidly expelled as a pulse every time slip occurs on the main fault and the dilatancy collapses, the so-called seismic pumping mechanism. In the second model, it is the slow build up of fluids that induces fault slip (in accordance with the law of effective stress). Once failure has occurred, the newly created fracture permeability in and around the fault allows the fluid to escape. Progressive deposition of hydrothermal minerals re-seals the fracture system and the whole cycle may be repeated. This 'seismic valve' mechanism requires the presence of some form of fluid reservoir.



For a given state of differential stress ($\sigma_1 - \sigma_3$), to cause brittle failure either:

1. Increase the magnitude of the differential stress ($\sigma_1 - \sigma_3$)
- or 2. Increase the pore fluid pressure (ρ) causing a translation of the Mohr Circle to the left ($\sigma_1' - \sigma_3'$).

When $\sigma_3 - \rho = T$, the criterion for hydraulic fracturing is satisfied.

Fig. 8.1a Mohr Circle construction showing effect of pore fluid pressure on brittle failure using a modified Navier-Coulomb-Griffith Failure Envelope (after Fyfe et al. 1978)

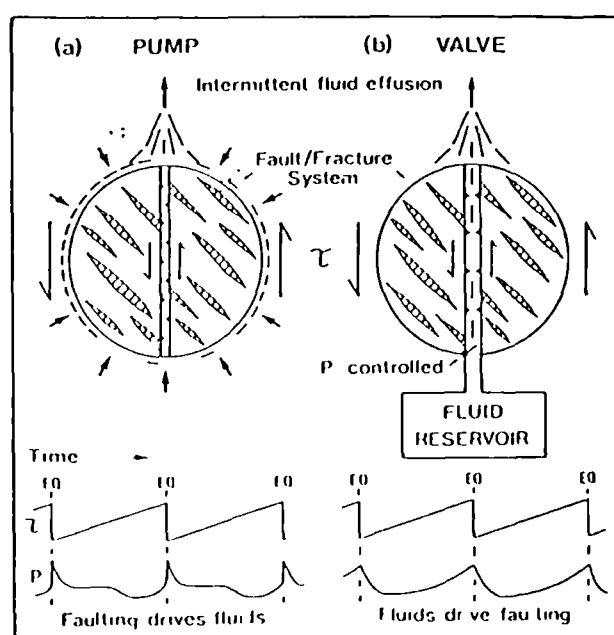
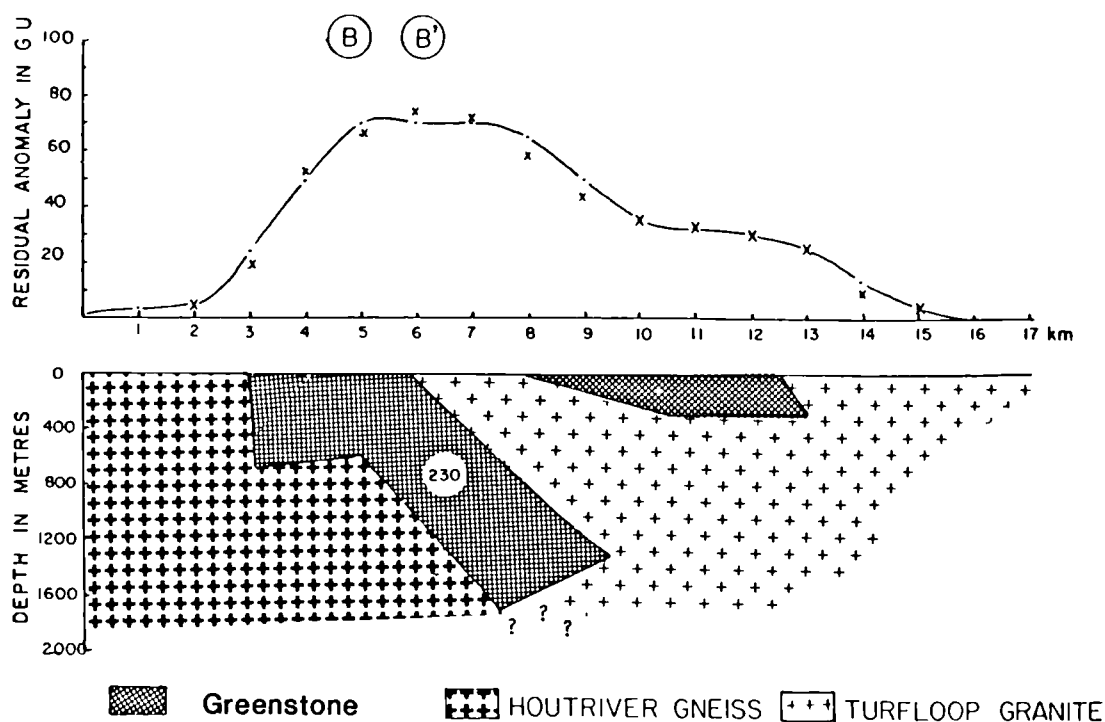


Fig. 8.1b Fluid flow accompanying faulting (from Sibson 1981b).



Note how vertical scale is greatly exaggerated. At depth the belt is actually a thin, shallowly dipping sliver of more dense greenstone within the granite. Such features suggest major shallow dipping structures are a significant part of the deep structure of the Kaapvaal Craton

Fig. 8.3 Deep structure of the Pietersburg Greenstone Belt (after de Beer and Stettler 1986)

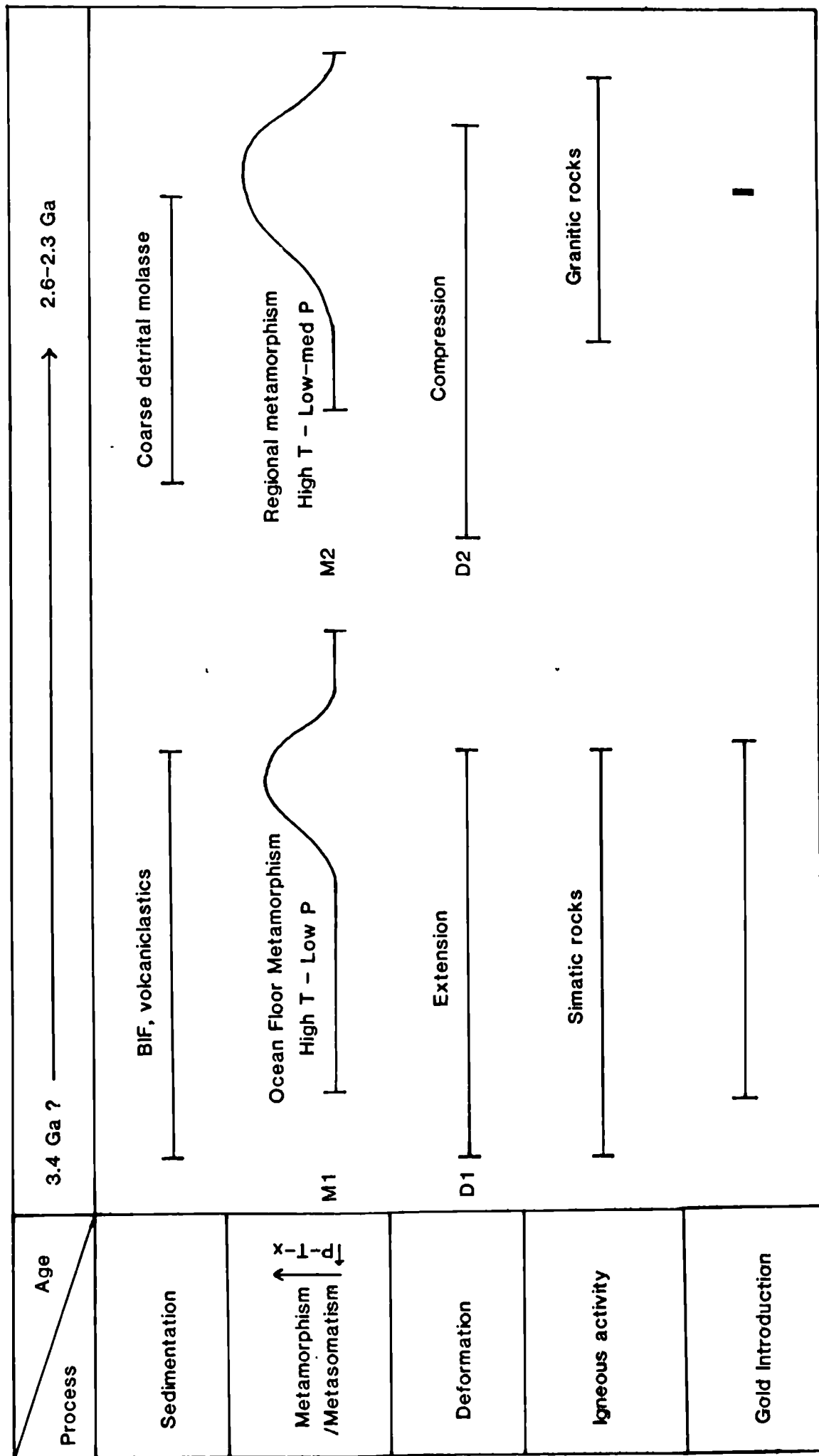


Fig. 8.4 Summary of the geological history of the Mt. Mare area as documented in this study

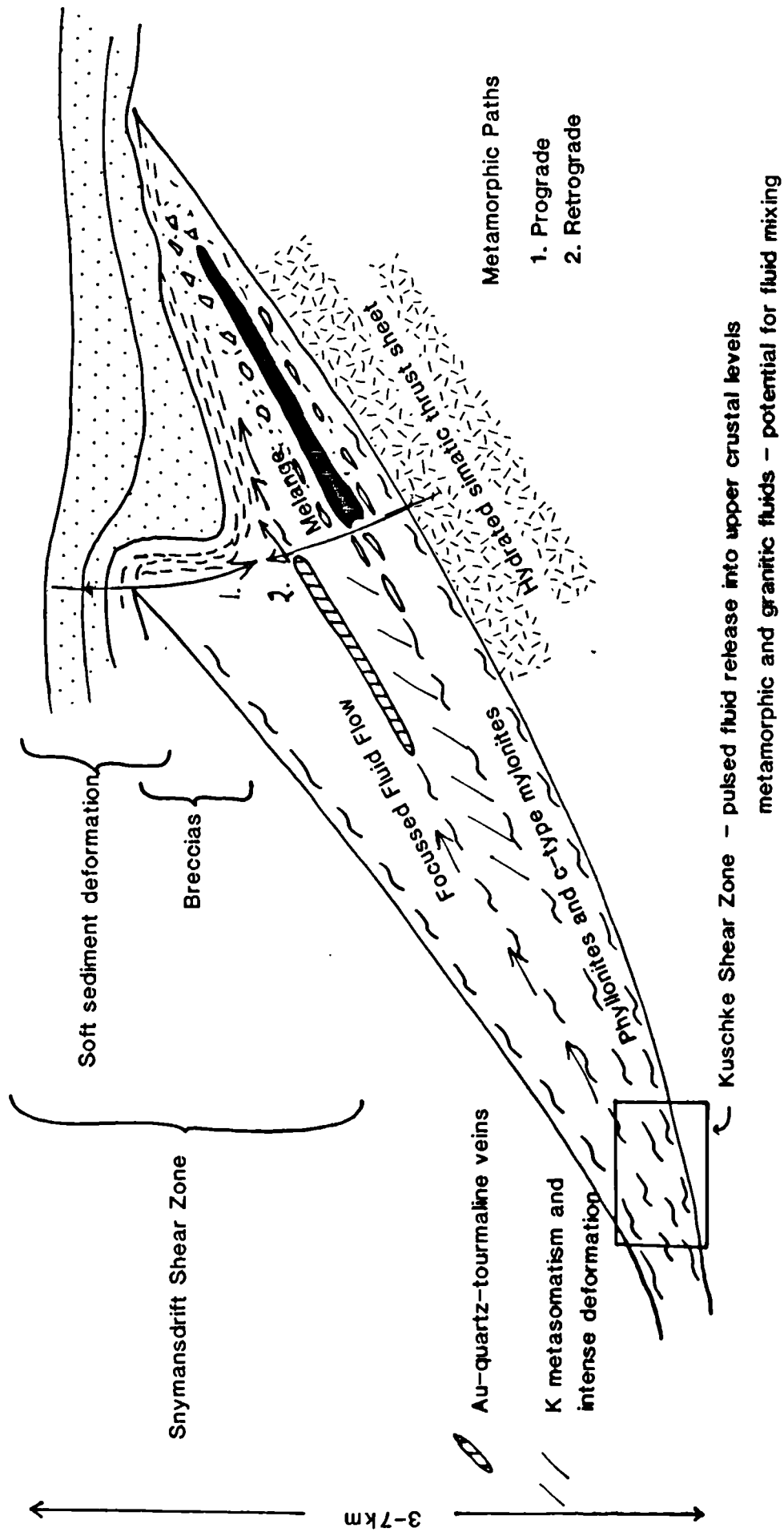


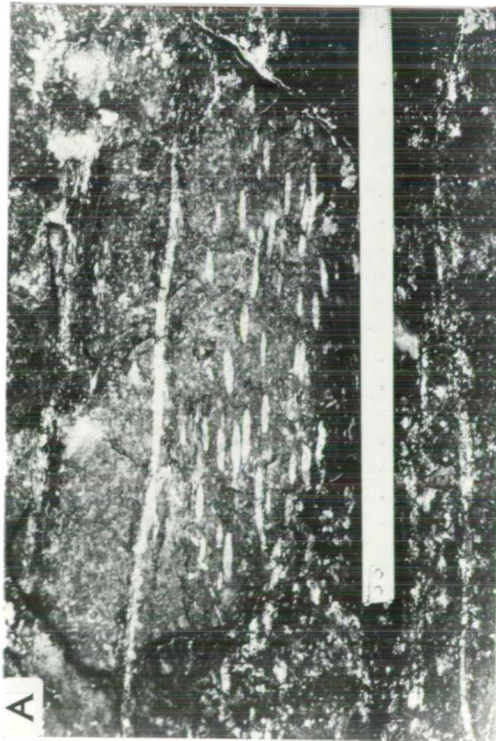
Fig. 8.5 Conceptual dynamic model of a mineralised fault zone

TABLE 8.1 OBSERVED AND INFERRED FAULTS IN THE MT. MARE AREA

	Lower Contact	Upper Contact
Hollandsdrift Shear Zone	Discrete high strain zone.	Inferred fault between Uitkyk sediments and overlying simatic rocks.
Sny mansdrift Shear Zone	Tectono-sedimentary melange.	Inferred fault between sediments and granite - possible lateral extension of HDSZ
Kuschke Shear Zone	Discrete high strain zones - separates simatic intrusives from extrusives.	
Kraal Thrust	Discrete high strain zone - separates Uitkyk sediments from overlying simatic rocks.	
Witkop Fault	Discrete high strain zone - inferred fault separating granite and simatic rocks.	

Plate 8.1 Strain features and fuchsite-quartz tectonite in the KSZ

- 8.1a.** Strained ocelli from southern boundary of the KSZ. Pillow lavas and ocelli can be traced into the shear zone and a gradual straining and 'flattening' of these primary textures is clearly observed. YZ plane of strain ellipsoid.
GR X 2 657 950, Y -33 350, Fig. 2.2
- 8.1b.** Fresh exposure of schistose to gneissose fuchsite-quartz end-member of the FQC tectonite. Note the coarseness of the layers and sub-parallelism of dark phyllonitic fuchsite layers and white quartz layers. Many of the quartz layers may represent veins intruded along the phyllonitic layers and so 'jacking' them apart -during episodes of hydraulic fracturing. GR X 2 657 950, Y -33 000, Fig. 2.2.
- 8.1c.** Large hand specimen showing detail of fuchsite-quartz tectonite. Many of the quartz layers are clearly veins and, furthermore, several cross-cutting generations of veins are present. Note however that earlier formed veins are deformed eg folded more than later veins.
GR X 2 657 950, Y -33 000, Fig. 2.2.
- 8.1d.** Parallel banded gneissose fuchsite-quartz tectonite. Areas containing relatively undeformed quartz veins as exemplified in plates 8.1b and c can be traced along strike where they have visibly undergone greater post-vein strain and possible recrystallization to form this parallel banded, possibly mylonitic tectonite. Locally this unit passes becomes a laminated green and white chert.
GR X 2 657 900, Y -33 600, Fig. 2.2.



**Plate 8.2 Schistose to gneissose fuchsite-carbonate
tectono-metasomatic fault rock.**

An example (true scale) of a fuchsite-carbonate end-member of the FQC tectonite showing the coarse to fine alternating layers of fuchsite and carbonate. The carbonate layers are veins formed by hydraulic fracturing and fluid flow sub-parallel to the foliation. At least six (and probably many more) different sequential cross-cutting veining events can be found in this specimen. Note however, that early formed veins eg vein A, are more deformed than later veins eg B. Many veins, especially younger generations that have not been severely deformed,, show evidence of internal fibrous growth indicative of extension normal to the foliation or even open space filling eg vein C. Deformation of veins apparently proceeds by asymmetric boudinage or pinch-and-swell eg vein D, followed by continued layer parallel ductile flow and 'smearing out' the veins parallel to the D2 fabric eg vein E. Crystal plasticity and superplasticity of carbonate is inferred to have been an important deformation mechanism in the tectonite. In many cases early formed veins are found preserved as pods or lenses that have been preserved in areas of low strain which have escaped deformation eg vein F. Many of them could be remnant 'swells'. On a larger scale the mineralization at ZR is similarly thought to lie in a low strain lens (chapter 7).

Sample from GR X 2 657 950, Y -33 000, Fig. 2.2

Plate 8.3 Progressive D2 deformation in the KSZ

- 8.3a.** Laminated green and white chert formed by the silicification and possible recrystallization of the FQC tectonite. Similar textures are found in the syn-sedimentary chert breccias found in the SDSZ (chapter 4.8.3). All textures observed in the breccia clasts were observed *in situ* in the chert bars in the KSZ.
GR X 2 657 000, Y - 33 500, Fig. 2.2.
- 8.3b.** Large BIF block showing *in situ* brecciation and ductile flow at BIF-phyllosilicate contact.
GR X 2 657 000, Y -32 400, Fig. 2.2.
- 8.3c.** Very heterogeneous breccia in fine grained, unoriented mudstone matrix lying within the KSZ. A breccia-filled, slurry-like protruberance into the underlying phyllosilic-mudstone matrix is arrowed. These breccias can be traced along strike into Utkyk Formation sediments.
GR X 2 657 900, Y -32 400, Fig. 2.2.



Plate 8.4 Coarse Uitkyk Formation detritus from immediately below the Kraal Thrust

- 8.4a.** An example of coarse, poorly sorted, angular to sub-angular detritus found in a channel-like deposit in mudstones that underlie and dip under the Kraal Thrust.
GR X 2 657 650, Y - 33 000, Fig. 2.2.
- 8.4b and c.** Two examples of *in situ* brecciation of large chert clasts in the detritus. These sediments and surrounding mudstones pass along strike into lithofacies D of the Uitkyk Formation.

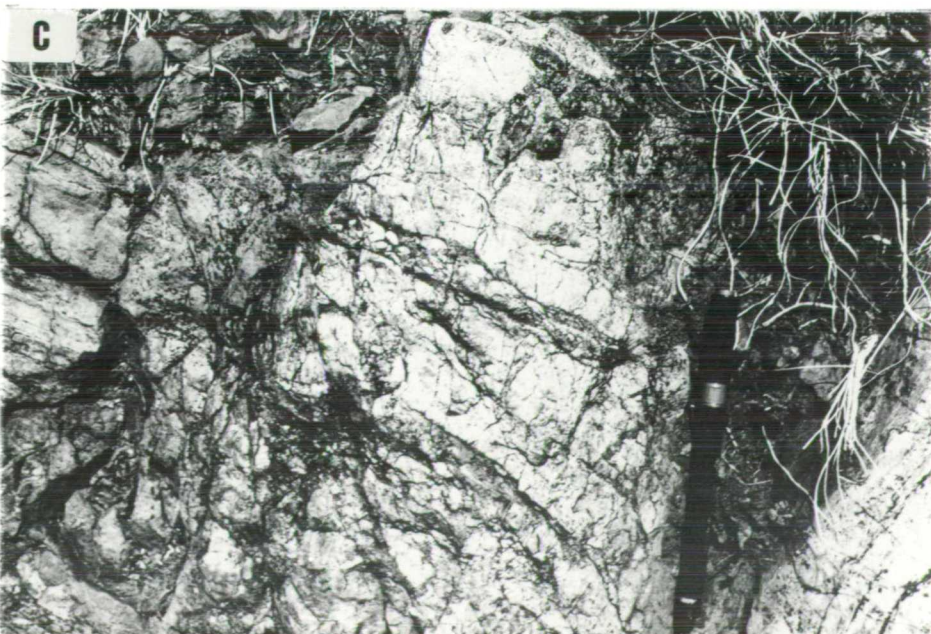
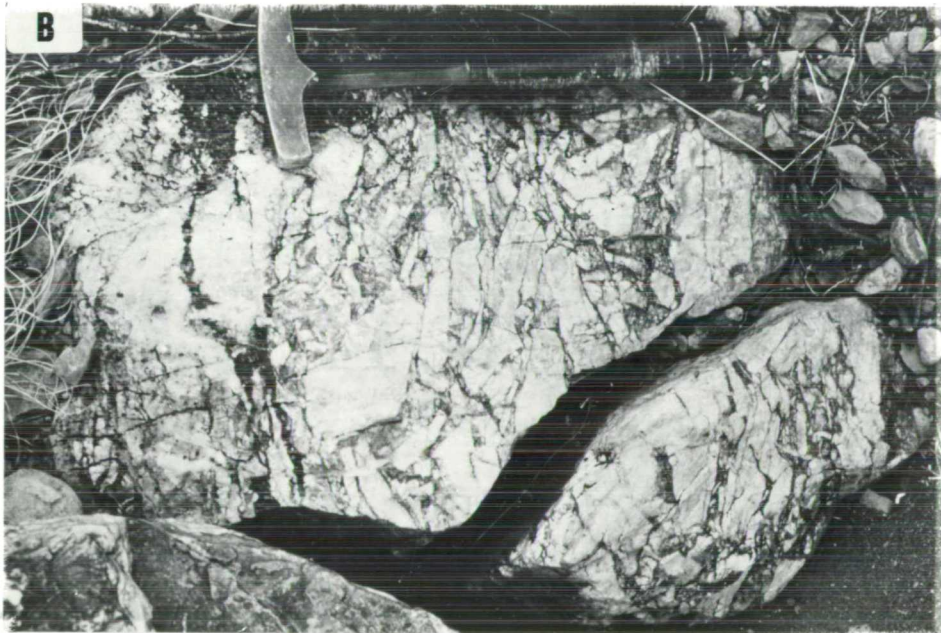
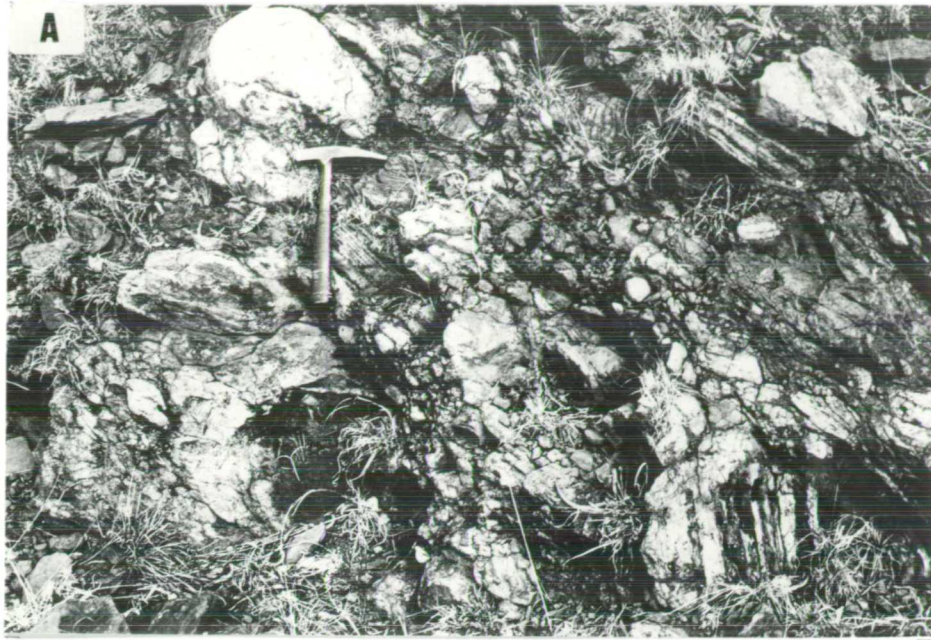


Plate 8.5 Hollandsdrift Shear Zone - field observations

8.5a. Example of the FQC tectonite from the base of the HDSZ. The lithology grades over the outcrop, a distance of 2-3m, from a fuchsite-quartz end-member at its base to a fuchsite-siderite (+\-chlorite) end-member. The unit then grades into a clastic lithology to the left of the photograph. GR X 2 659 100, Y -31 900, Fig. 2.2.

8.5b. A close up of the FQC-clastic lithology contact. Note there is progressive deformation of the siderite layers to form pinch-and-swell structure. The lithology grades into the clastic unit which contains unoriented but occasionally rounded chert clasts lying in a mudstone matrix. It is the breakdown of the phyllonitic layers that forms the mudstone matrix, preserving occasional layers and remnant swells of siderite close to the contact, but within the clastic lithology. GR X 2 659 100, Y -31 900, Fig. 2.2.

8.5c. Contact of phyllonite-siderite tectonite and breccio-conglomerate layer. The breccio-conglomerate truncates at a low to moderate angle the D2 tectonic foliation, but carries itself the D2 fabric. Note to the left of the hammer a single pinch-and-swell layer of siderite is present. Just below the hammer a similar layer has broken down to leave siderite nodules, which are remnant swells, within the breccio-conglomerate. These nodules are identical to those observed in the Uitkyk Formation (chapter 2.4). GR X 2 659 100, Y -31 600, Fig. 2.2

8.5d. Isoclinally folded mudstone layers, axial planar to the D2 fabric in the HDSZ. Such features indicate the sediments have been considerably flattened during D2 deformation. GR X 2 659 075, Y -31 500, Fig. 2.2.



CHAPTER 9

SUMMARY

9.1 THESIS OVERVIEW

The combined field, microstructural and geochemical data presented in this study has shown the Mt. Mare area is a poly-deformed and poly-metamorphosed, supracrustal tectono-stratigraphic complex. It is comprised of distinct lithological units of older simatic rocks and BIF which were deformed and metamorphosed prior to the deposition of younger, detrital sediments. The units are bounded by faults, fault-related melanges and well defined unconformities.

Circa 3.4 Ga simatic rocks and associated BIF horizons are now found as allochthonous thrust sheets which, together with similar lithologies throughout the southwest Pietersburg belt, constitute the tectonically dismembered remnants of a single, 3.5–10 km thick, intrusive-extrusive igneous complex. Parts of this complex constitute a Steinmann Trinity. The original stratigraphic relationships between the different simatic lithologies have been obscured by subsequent deformation but incompatible/immobile element data indicates the extrusive rocks (basalts and pillow lavas) were derived from the intrusive rocks (gabbros and peridotites respectively), by low pressure crystal fractionation and/or partial melting. This suggests the sequence basalt-gabbro-peridotite represents increasing depth in the complex. Gabbroic magma chambers may have played an important role in this igneous evolution.

Formation of the igneous complex overlapped in time with a regional-scale, high temperature-low pressure (upper amphibolite to lower greenschist facies, <3Kb) 'ocean floor' metasomatic event (M1). This resulted in the extensive hydration (and local carbonation) of the majority of the simatic rocks and widespread 'open system' major and minor element mobility. This involving at least the gross addition of magnesium to the simatic pile and loss of silica, iron and gold. The metasomatically induced geochemical changes may have mimicked primary igneous differentiation trends and obscured others. Syndeformational magmato-metasomatic interaction was responsible for the development of spinifex-textured amphiboles in cross-cutting fibrous extension veins or dykes. Little is known about the age or geometry of early deformation of the simatic pile, but evidence exists for (at least) one period of deformation before the deposition of the Uitkyk Formation.

The igneous complex was emplaced as allochthonous units during a period of NW-SE directed horizontal shortening (50-70%) across the area, possibly between 2.6-2.3 Ga. This is the main deformation event D2, seen in the area. It resulted in significant thrust-related stratigraphic repetition by the successive stacking of rigid sheets of simatic rocks (and BIF) and small, syn-D2, molasse filled sedimentary depositories in a foreland basin environment. A sequence of six D2 faults have been identified and, although much of the fault movement was characterised by stick-slip, out-of-sequence movement, they formed part of a northwest verging, foreland propagating, imbricate fan thrust complex. The fan was related to a flat decollement layer at the base of the belt which as a whole is therefore allochthonous. Granitoid intrusion accompanied D2 as two magmatically active terrains were translated towards each other.

Many of the D2 faults were emergent and thus a gradation from tectonic to sedimentary processes is commonly observed. In the SDSZ the gradation is marked by a distinctive tectono-sedimentary melange formed by the progressive brittle-ductile deformation and fluidization of BIF units. The melange probably formed as a lag slide above an underthrust block of simatic rocks carried above the Kuschke Shear Zone. Both the Kuschke and Hollandsdrift Shear Zones are also marked by the fluidization of phyllonites and BIF/chert bars. Such fluidization may allow active expulsion of debris from within or close to the emergent fault tips to be deposited as sedimentary debris flows.

The debris flows form part of the Uitkyk Formation, a sequence of coarse detrital molasse-like sediments which were shed from the emergent simatic nappes during thrusting into small unconformity and fault bounded basins. Sedimentation lead to the development of prograding alluvial deposits which were carried piggy-back above the migrating thrusts. Syn-sedimentary faulting and soft-sediment stratal disruption accompanied this early deformation.

Continued D2 deformation resulted in the piggy back basins becoming steepened and flattened between the more rigid simatic thrust sheets and lead to the development of the Snymansdrift and Hollandsdrift Shear Zones. Both pure and simple shear were important components in the deformation history, with W_k estimated to be in the range 0.76-0.8 in the SDSZ. Although deformation was initially localised along pre-existing lithological contacts, it quickly became more distributive at depth due to high vertical strain and temperature gradients, influenced by syn-D2 convective fluid flow and granitoid activity

Although there was an initial, strong lithological control on fault rock development and deformation style in the area, subsequent syn-D2 dynamothermal metamorphism lead to the progressive homogenization of the textural, chemical and metamorphic differences between precursor lithologies. Original sedimentary and igneous lithologies underwent

hydrated, (locally carbonated) and extensively strain softened, chlorite-quartz and sericite-quartz phyllonites. Syn-D2 chloritoid and clinozoisite porphyroblast growth suggest that greenschist facies metamorphic conditions were attained in the SDSZ and HDSZ. This corresponds to temperatures between 350-500°C and maximum pressures of 3Kb. The pressure estimate is poorly constrained and could conceivably be too great, depending on the geothermal gradient. Crystal plasticity of quartz and feldspar played a minor role in the progressive deformation history and brittle deformation and diffusive mass transfer (and related) processes were predominant. Overall, deformation occurred in the brittle-semi brittle field.

Stick-slip fault movement was associated with episodic hydraulic fracturing and movement of fluids through the shear zones. This tectono-metasomatic activity produced a distinctive gneissose fuchsite-carbonate-quartz tectonite in the HDSZ and KSZ and allowed significant fluid release from deeper crustal levels into a near surface environment. One such fluid flow event was mineralizing and resulted in the precipitation of quartz-tourmaline-arsenopyrite-pyrite-gold veins. Subsequent deformation has preserved much of the mineralization in low strain lenses.

9.2 DISCUSSION

The conclusions of this study allow answers to questions raised at the outset (chapter 1) to be proposed.

The main philosophy of the original direction of the study was to trace the history of gold mineralization through the geological evolution of the belt. This approach proved very useful because it specifically directed thought towards understanding the complexities that arise from the spatial and temporal interaction of different geological processes (chapter 1.3). In the Mt. Mare area two or more of the following processes interacted at some time in its geologic history; deformation, igneous activity, sedimentation, metamorphism, metasomatism and associated gold mobility. Although gold was introduced early in the geologic history in BIF units, subsequent tectonism and metamorphism appear to have downgraded these concentrations. Introduction of 'new' gold later in the deformation history was apparently of greater importance.

The overlap of processes has led to a complex geological evolution in the study area which agrees with some, but differs with other, conclusions of earlier workers. Five of these conclusions, were outlined in chapter 1.5 and can now be compared to the conclusions of this study:



0 5
cm

A

F

D

C

G

B

1) The percentage of extrusive rocks may have been overestimated in the past. As a result the relationships between different simatic lithologies, further complicated by subsequent deformation, are obscure and indicate the simatic stratigraphy of SACS (1980) requires revision.

2) There was a period of high temperature-low pressure metasomatism which only affected the simatic lithologies soon after their formation. It predates deposition of the Uitkyk Formation, and therefore granite intrusion. The M2 event appears to be due regional metamorphism but possibly occurred under high geothermal gradients induced by widespread magmatic activity.

3) Thrust-related stratigraphic repetition has played a major role in the current lithological distribution in the Mt. Mare area.

4) Mineralization may have been related to granitic fluids migrating along structural pathways

5) Although similarities do exist between Pietersburg and Barberton greenstone belts, such comparison has in the past hindered geologic thought in the Pietersburg area. Comparisons are not preferable at this time.

In conclusion, there is as yet little evidence in the south-west Pietersburg greenstone belt to suggest crustal processes were markedly different in the Archaean to those thought to operate today (chapter 1.3). A better framework for further research in the area may be to consider the possible effects differences in the rates of operation of crustal processes, or their distribution in the earth's crust, could have had on the geological record.

9.3 RECOMMENDATIONS FOR FURTHER WORK

Further work in the area should now test the results of this study. There are four main areas that require urgent work;

1) Further petrological work on the mineralization and also the associated dynamothermal metamorphism in the surrounding shear zone rocks are needed to confirm the models presented here. Serious consideration must be given to the potential effects of fluid mixing within the shear zone before isotopic work is carried out.

2) Further mapping both within the Mt. Mare area and the rest of the Pietersburg belt should continue with geomorphology and extent of D2 deformation with the aim of

structurally restoring the belt. Comprehensive metamorphic studies are needed to more tightly constrain the P-T conditions attained during D2 and estimate the vertical geothermal gradient.

3) Structural restoration of D2 will hopefully allow a better understanding of the nature and geometry of D1 and its effect on the simatic igneous complex. More detailed documentation of the relationship between simatic lithologies within D2 thrust sheets and mapping of metasomatic mineral assemblages are needed. Quantitative petrological and geochemical data, including rare earth element and stable isotope analyses, are required to test the proposed igneous differentiation and metasomatic mass transfer models. Combined, such work may allow further comparisons of Archaean simatic igneous processes with those seen in Phanerozoic ophiolites and the modern ocean floor.

4) The absolute ages of lithological units are very poorly constrained and should be more firmly established. This will allow the geologic history to be incorporated into a coherent framework and may aid the estimation of the rates of geological processes in the Archaean.

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APPENDIX 1 : ANALYTICAL TECHNIQUES

1. Whole Rock Analysis by ICP

Major Elements analysed; SiO_2 , Al_2O_3 , TiO_2 , FeO_3 (total iron), MnO , CaO , Na_2O , K_2O , P_2O_5 .

Trace Elements Analysed; Ba, Be, Cr, Cu, La, Ni, Sr, V, Zn, Zr, Co

Samples to be analysed were crushed and then ground to pass 200# in a tungsten carbide Tema mill. The moisture content (H_2O^-) and loss on ignition (LOI) were then determined prior to fusion of the samples. $\%\text{H}_2\text{O}_-$ was determined by heating a measured mass of the powder overnight at 110°C and then measuring the mass loss. LOI was similarly calculated after heating to 950°C and maintaining this temperature overnight before weighing the next day.

Sample solutions for analysis were prepared by carefully weighing $0.2500 \pm 0.0001\text{g}$ of sample into a clean porcelain crucible and then mixing intimately with $0.7500 \pm 0.003\text{g}$ of Johnson Matthey Spectroflux 100B. This was then transferred to a graphite crucible and placed into a pre-heated muffle furnace at 1000°C . The samples were fused for 20 minutes. The molten bead was then transferred into a 400ml plastic beaker containing 25ml of 2N HNO_3 and 150ml of de-ionised water. The shattered bead was dissolved by using a magnetic stirrer at room temperature for 30 minutes. This solution was then transferred to a clean 250ml volumetric flask. The beaker was rinsed thoroughly, and the rinsings added to the flask. This was then diluted to 250ml and filtered through a Whatman 12.5cm No. 41 filter paper into a clean plastic bottle, which was then sealed with a piece of plastic placed under the cap, and stored until the analysis was undertaken.

In every set of 30 samples, at least four in-house reference materials were included, along with one blank solution and at least five duplicate preparations. The samples were analysed in a set order which included drift samples and standard solutions for calibration. Standards used were BR, GA, GH, NIM-N, SY-2, UB-N, NIM-G, DTS-I, AN-G and MICA-MG.

The following running order was adopted.

Blank

Drift

10 Standards

Blank
 Drift
 10 Samples
 Blank
 Drift
 10 Samples
 Blank
 Drift
 5 Standards
 Blank
 Drift
 10 Samples
 Blank
 Drift
 10 Samples
 Blank
 Drift
 10 Samples
 Blank
 Drift
 5 Standards

The data was then reduced by normalisation to a chosen drift value which compensated for long term fluctuations in the machine sensitivity. Calibration lines were set up using the standard solution results, and the element concentrations calculated from this line using the normalised data values.

2. Accuracy and precision

The following duplicates were analysed during the various runs on the ICP in order to test the accuracy and precision of the machine and the reliability of the results presented in table 3.1. Some of the samples used were done in the early stages of the project and were not part of the main greenstone data set presented in Table 3.1.

	GR3		ZR1S5		ASB 2	
	115.02					
SiO ₂	53.81	53.37	55.42	55.89	42.82	42.46
Al ₂ O ₃	13.14	13.00	12.84	12.99	5.47	5.43
Fe ₂ O ₃	15.51	15.44	20.94	21.05	7.97	7.90
MgO	3.61	3.58	3.96	3.97	31.68	31.40
CaO	6.43	6.43	0.00	0.00	3.02	2.96
Na ₂ O	4.43	4.37	0.02	0.02	0.12	0.10
K ₂ O	0.54	0.53	0.06	0.08	0.01	0.00
P ₂ O ₅	0.19	0.19	0.03	0.03	0.02	0.02
TiO ₂	1.30	1.29	0.93	0.94	0.09	0.09
MnO	0.23	0.23	0.09	0.09	0.36	0.36
H ₂ O-	0.27	0.27	0.24	0.24	0.36	0.36
LOI	1.19	1.19	4.54	4.54	9.08	9.08

Total 100.65 99.89 99.07 99.84 100.75 99.91

Ba	186	182	37	37	10	8
Be	1	1	0	0	0	0
Cr	35	38	0.43%	0.42%	0.54%	0.54%
Cu	184	200	16	14	12	7
La	15	18	1	7	9	0
Ni	20	29	574	590	1433	1383
Sr	322	320	7	10	15	14
V	196	202	218	214	99	94
Zn	150	159	219	228	66	56
Zr	142	142	42	33	4	7

	SPIN E		BAS 8		SPIN A	
SiO ₂	52.84	53.05	37.70	38.29	52.06	52.47
Al ₂ O ₃	7.93	7.88	1.92	1.95	7.65	7.70
Fe ₂ O ₃	13.38	13.57	12.41	12.66	13.40	13.40
MgO	10.59	10.69	35.37	36.08	10.52	10.60
CaO	8.93	9.04	0.14	0.18	10.01	10.08
Na ₂ O	2.88	2.81	0.08	0.10	2.46	2.42
K ₂ O	0.17	0.16	0.00	0.00	0.12	0.12
P ₂ O ₅	0.08	0.09	0.04	0.05	0.09	0.08
TiO ₂	0.84	0.85	0.44	0.47	0.86	0.86
MnO	0.19	0.20	0.16	0.17	0.21	0.21
H ₂ O-	0.11	0.11	0.18	0.18	0.17	0.17
LOI	1.57	1.57	10.85	10.85	1.27	1.27

Total 99.51 100.02 99.29 100.99 99.82 99.49

Ba	54	62	47	49	34	33
Be	1	1	0	0	0	0
Cr	681	726	0.69%	0.70%	687	695
Cu	59	52	40	51	133	136
La	6	8	0	4	4	6
Ni	187	210	0.29%	0.40%	217	173
Sr	48	48	10	12	94	95
V	234	243	83	96	225	227
Zn	114	109	80	92	104	95
Zr	43	53	3	12	45	47

These results which were obtained during the three separate analytical runs undertaken in this project indicate both the major and minor elements are within acceptable limits of accuracy. At high levels of MgO the accuracy for that oxide is reduced because the ICP is not the ideal analytical technique for such values. XRF may have produced more accurate results.

The following sample, ZR1 3 0.04 was analysed in each run undertaken and repeated in one of those runs in order to test for the long term precision of the ICP. The results again show they fall within acceptable limits.

SiO ₂	55.31	55.26	54.71	55.38
Al ₂ O ₃	14.54	14.54	14.23	14.43
Fe ₂ O ₃	10.64	10.63	10.50	10.73
MgO	7.46	7.47	7.74	7.43
CaO	3.08	3.06	3.05	3.17
Na ₂ O	4.77	4.75	4.64	4.72
K ₂ O	0.11	0.10	0.12	0.10
P ₂ O ₅	0.05	0.04	0.06	0.05
TiO ₂	0.53	0.53	0.50	0.50
MnO	0.14	0.13	0.14	0.13
H ₂ O-	0.36	0.36	0.36	0.36
LOI	4.27	4.27	4.27	4.27

Total	101.26	101.27	100.32	101.27
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Ba	27	30	21	17
Be	0	0	0	1
Cr	220	224	227	230
Cu	14	12	16	16
La	21	18	18	6
Ni	117	115	104	147
Sr	54	56	58	59
V	218	218	217	236
Zn	54	59	59	46
Zr	20	16	32	21

APPENDIX 2

MICROPROBE ANALYSES OF MINERALS

REPRESENTATIVE MICROPROBE ANALYSES OF FELDSPARS FROM METAGABBROS

	GC	1	GC	2	G8a	1	G8a	2	G8a	3	G3	1	G3	2	G3	3
Na2O	9.033		8.632		10.992		11.320		10.942		11.077		10.919		10.816	
Al2O3	22.426		22.912		19.691		19.524		19.960		19.794		19.700		19.633	
MgO	0.000		0.000		0.223		0.000		0.042		0.183		0.079		0.174	
SiO2	63.970		62.708		68.458		68.330		67.821		68.011		67.438		67.662	
CaO	4.677		5.093		0.054		0.106		0.000		0.000		0.000		0.000	
K2O	0.115		0.114		0.179		0.172		0.157		0.060		0.129		0.577	
TiO2	0.070		0.143		0.086		0.048		0.062		0.070		0.209		0.018	
Cr2O3	0.000		0.116		0.000		0.006		0.017		0.036		0.000		0.000	
MnO	0.036		0.000		0.000		0.151		0.000		0.116		0.017		0.000	
FeO	0.001		0.334		0.070		0.018		0.119		0.000		0.000		0.034	
TOTAL	100.328		100.052		99.753		99.675		99.120		99.347		98.491		98.434	

NS: All elements except Si, Al, Ca and Na were below detection limits.

REPRESENTATIVE MICROPROBE ANALYSES OF SPINIFEX-TEXTURED AMPHIBOLES

	SPIN _C			A	B	C	D	E	F	G	H	I	SPIN _A			A	B	C	D	E	F	G	H	I
Na2O	0.577	0.575	0.862	0.483	0.840	0.526	0.520	0.697	0.600				0.840	0.581	0.684	0.773	0.839	0.728	0.842	0.807	0.565			
Al2O3	1.367	1.601	2.330	1.836	1.945	1.455	1.958	2.042	2.560				1.721	2.390	2.805	2.234	2.034	2.244	2.774	2.949	1.996			
MgO	14.534	14.448	14.423	14.354	14.195	14.139	14.583	14.664	14.315				14.774	15.027	14.682	14.651	15.017	14.532	14.504	14.464	14.529			
SiO2	54.611	54.314	53.5 9	53.507	54.150	54.284	54.537	53.777	53.610				53.597	53.403	53.289	53.656	53.225	53.241	52.819	52.112	53.361			
CaO	12.326	12.421	12.082	12.490	12.117	12.376	12.603	12.427	12.062				12.455	12.616	12.550	12.411	12.457	12.540	12.368	12.641	12.356			
K2O	0.063	0.052	0.032	0.035	0.052	0.042	0.000	0.001	0.021				0.058	0.000	0.100	0.067	0.092	0.000	0.059	0.110	0.078			
TiO2	0.046	0.197	0.091	0.095	0.076	0.151	0.032	0.094	0.188				0.027	0.066	0.038	0.000	0.027	0.147	0.081	0.044	0.118			
Cr2O3	0.000	0.000	0.000	0.080	0.132	0.000	0.093	0.230	0.059				0.290	0.475	0.457	0.327	0.532	0.333	0.350	0.367	0.272			
MnO	0.236	0.300	0.231	0.188	0.218	0.211	0.116	0.299	0.214				13.960	14.008	13.808	14.300	13.348	13.921	14.274	14.384	14.489			
FeO	14.993	14.408	14.393	14.428	15.055	14.533	13.864	14.030	15.099				13.960	14.008	13.808	14.300	13.348	13.921	14.274	14.384	14.489			
Total	98.753	98.316	97.953	97.496	98.780	97.717	98.306	98.261	98.728				97.722	98.566	98.413	98.419	97.571	97.686	98.071	97.878	97.764			
Si	7.850	7.831	7.747	7.789	7.795	7.873	7.825	7.754	7.718				7.777	7.682	7.673	7.735	7.724	7.725	7.653	7.589	7.750			
Al iv	0.150	0.169	0.253	0.211	0.205	0.127	0.175	0.246	0.282				0.223	0.318	0.327	0.265	0.276	0.275	0.347	0.411	0.250			
Al vi	0.082	0.103	0.145	0.104	0.125	0.122	0.156	0.101	0.152				0.071	0.087	0.149	0.115	0.072	0.109	0.127	0.095	0.092			
Na	0.161	0.161	0.242	0.136	0.234	0.148	0.145	0.195	0.167				0.236	0.162	0.191	0.216	0.236	0.205	0.237	0.228	0.159			
Mg	3.114	3.106	3.113	3.115	3.046	3.057	3.119	3.152	3.072				3.195	3.222	3.151	3.149	3.249	3.143	3.133	3.140	3.146			
Ca	1.898	1.919	1.874	1.948	1.869	1.923	1.938	1.920	1.861				1.936	1.944	1.936	1.917	1.937	1.950	1.920	1.973	1.923			
K	0.012	0.009	0.0 6	0.006	0.010	0.008	0.000	0.000	0.004				0.011	0.000	0.018	0.012	0.017	0.000	0.011	0.020	0.014			
Ti	0.0 5	0.021	0.010	0.010	0.008	0.016	0.003	0.100	0.020				0.003	0.007	0.004	0.000	0.003	0.016	0.009	0.005	0.013			
Cr	0.000	0.000	0.000	0.009	0.015	0.000	0.011	0.026	0.007				0.033	0.054	0.052	0.037	0.061	0.038	0.040	0.042	0.031			
Mn	0.029	0.037	0.028	0.023	0.027	0.026	0.014	0.036	0.026				1.694	1.685	1.663	1.724	1.620	1.689	1.730	1.752	1.760			
Fe	1.8 2	1.737	1.743	1.757	1.812	1.763	1.664	1.692	1.818				1.694	1.685	1.663	1.724	1.620	1.689	1.730	1.752	1.760			
Mg/Mg+Fe	0.633	0.641	0.641	0.639	0.627	0.634	0.652	0.651	0.628				0.654	0.657	0.655	0.646	0.667	0.650	0.644	0.642	0.641			

	G3	A	B	C	D	E	F	G	H	I	J	GBA	B	C	D	E	F	G	H	I	J	K	L
Ma20	0.607	0.615	0.857	0.051	1.222	0.842	0.649	0.902	0.912	1.756	1.133	1.250	1.133	1.073	1.037	0.727	3.130	1.109	1.016	1.055	1.145	1.092	1.083
Al2O3	1.816	1.738	1.849	3.818	6.670	3.438	3.065	5.171	3.484	7.184	3.435	3.195	3.435	3.435	1.589	2.962	6.170	3.139	3.322	3.502	3.987	2.716	3.063
MgO	10.004	9.220	9.583	7.995	6.929	8.674	9.866	7.657	6.467	5.803	12.383	11.725	12.165	13.243	11.903	11.903	10.123	12.600	12.158	12.950	13.261	12.769	12.993
SiO2	54.515	56.078	49.559	47.154	50.553	50.772	47.066	50.210	43.640	50.831	50.799	50.152	53.013	51.650	54.021	51.359	51.059	51.559	51.059	50.668	49.606	52.268	51.001
CaO	11.653	11.236	11.265	11.539	11.427	11.987	11.685	11.333	11.333	12.354	12.354	12.354	12.354	12.057	12.527	12.215	10.132	12.232	12.043	11.788	11.593	12.128	11.988
K2O	0.000	0.000	0.008	0.161	0.450	0.215	0.127	0.542	0.175	1.025	0.091	0.157	0.000	0.789	0.009	0.102	0.094	0.058	0.043	0.041	0.069	0.078	0.095
Na2O	0.970	1.221	0.657	0.204	0.137	0.476	0.334	0.266	0.107	1.528	0.104	0.114	0.114	0.012	0.000	0.000	0.004	0.000	0.450	0.300	0.179	0.161	
Cr2O3	0.075	0.132	0.013	0.148	0.161	0.000	0.000	0.000	0.000	0.000	0.104	0.104	0.114	0.012	0.000	0.000	0.004	0.000	0.450	0.300	0.179	0.161	
MnO	0.287	0.286	0.332	0.195	0.172	0.050	0.349	0.229	0.241	0.277	0.314	0.314	0.293	0.335	0.374	0.149	0.230	0.014	0.223	0.188	0.202	0.184	0.243
FeO	19.033	17.698	19.958	23.558	23.967	21.639	21.814	24.172	22.801	26.803	16.688	17.994	17.392	15.766	17.972	13.905	16.954	17.563	16.975	17.836	16.171	17.049	17.049
Total	99.050	98.224	98.686	98.128	98.269	97.874	98.611	97.245	92.730	98.349	97.482	97.986	97.520	97.596	97.876	97.901	97.759	97.427	97.665	98.075	97.854	97.875	97.875
Si	7.759	7.802	7.707	7.189	6.890	7.233	7.334	6.805	7.307	6.554	7.538	7.538	7.461	7.784	7.629	7.767	7.572	7.572	7.517	7.420	7.658	7.658	7.658
Al IV	0.261	0.198	0.293	0.653	1.101	0.580	0.522	0.881	0.598	1.272	0.462	0.462	0.462	0.539	0.216	0.371	0.233	0.426	0.423	0.484	0.560	0.560	0.560
Al VI	0.064	0.057	0.017	0.000	0.049	0.000	0.000	0.000	0.000	0.000	0.096	0.139	0.139	0.063	0.059	0.145	0.813	0.118	0.158	0.128	0.127	0.175	
Mg	0.192	0.166	0.236	0.267	0.347	0.233	0.182	0.253	0.257	0.511	0.359	0.326	0.326	0.310	0.295	0.208	0.873	0.317	0.292	0.304	0.308	0.308	0.308
Mg	2.123	1.612	2.027	1.729	1.511	1.850	2.127	1.650	2.054	1.299	2.738	2.594	2.698	2.899	2.621	2.170	2.769	2.685	2.864	2.864	2.864	2.864	2.864
Ca	1.777	1.675	1.717	1.793	1.791	1.838	1.810	1.743	1.767	1.666	1.963	1.961	1.922	1.970	1.933	1.561	1.932	1.918	1.874	1.874	1.874	1.874	1.874
Ca	0.000	0.000	0.001	0.030	0.084	0.039	0.023	0.100	0.033	0.196	0.017	0.030	0.021	0.007	0.020	0.017	0.011	0.008	0.008	0.006	0.013	0.018	
Ti	0.104	0.128	0.070	0.022	0.015	0.051	0.036	0.029	0.012	0.173	0.030	0.000	0.000	0.088	0.001	0.021	0.001	0.033	0.000	0.000	0.009	0.010	
Cr3+	0.008	0.014	0.001	0.017	0.016	0.000	0.000	0.000	0.000	0.000	0.012	0.013	0.013	0.012	0.046	0.019	0.028	0.002	0.028	0.024	0.020	0.019	
Mn3+	0.035	0.034	0.390	0.024	0.021	0.006	0.043	0.028	0.030	0.035	0.039	0.037	0.042	0.046	0.019	0.028	0.028	0.002	0.028	0.024	0.024	0.023	
Fe2+	2.266	2.059	2.374	2.858	2.932	2.589	2.638	2.923	2.775	3.366	2.070	2.233	2.164	1.936	2.220	1.672	2.090	2.180	2.106	2.183	1.982	2.100	2.100
Mg/Mg+Fe	0.064	0.481	0.377	0.340	0.417	0.446	0.361	0.425	0.278	0.278	0.569	0.537	0.535	0.600	0.541	0.565	0.570	0.552	0.576	0.570	0.588	0.562	0.562
Ma20	0.336	0.511	0.309	0.559	0.349	0.279	0.431	0.470	0.608	1.302	1.588	1.057	1.588	1.411	1.025	1.559	1.643	1.849	1.709	1.358	1.011	1.900	1.900
Al2O3	1.721	1.606	2.277	1.108	1.690	1.480	1.308	1.652	2.435	7.723	8.248	5.413	8.107	6.381	13.178	6.830	6.905	7.351	7.875	7.720	6.043	6.300	6.300
MgO	18.404	18.190	18.067	16.041	18.825	18.861	18.456	18.717	18.707	11.872	12.128	12.891	11.433	13.178	13.155	13.341	13.353	13.353	13.097	11.631	13.601	12.718	12.718
SiO2	55.115	55.157	55.158	55.981	55.833	54.947	56.148	55.394	53.635	46.273	46.824	50.866	46.001	48.857	47.621	47.713	46.537	46.197	46.233	49.837	46.118	46.118	46.118
CaO	12.766	12.808	12.820	12.655	12.655	12.535	13.075	12.450	11.832	11.621	12.025	12.587	12.587	12.222	12.046	11.900	12.325	11.630	12.583	11.614	11.747	11.747	11.747
K2O	0.031	0.042	0.079	0.051	0.000	0.136	0.000	0.031	0.074	0.400	0.246	0.195	0.353	0.396	0.379	0.379	0.463	0.528	0.553	0.530	0.232	0.478	
Na2O	0.951	1.275	0.651	0.205	0.137	0.476	0.334	0.266	0.107	1.528	0.104	0.114	0.114	0.012	0.000	0.000	0.004	0.000	0.450	0.300	0.179	0.161	
Cr2O3	0.076	0.136	0.065	0.148	0.161	0.000	0.000	0.000	0.000	0.000	0.104	0.104	0.114	0.012	0.000	0.000	0.004	0.000	0.450	0.300	0.179	0.161	
MnO	0.287	0.286	0.332	0.195	0.172	0.050	0.349	0.229	0.241	0.277	0.314	0.314	0.293	0.335	0.374	0.149	0.230	0.014	0.223	0.188	0.202	0.184	0.243
FeO	8.603	8.714	7.935	8.638	7.926	8.200	8.095	7.570	8.706	15.987	16.702	15.215	16.586	14.667	15.873	14.350	14.350	14.786	14.552	15.533	14.248	15.081	15.081
Total	97.081	97.808	97.423	98.550	97.873	97.156	98.782	97.163	97.023	97.035	99.397	98.909	97.951	97.834	99.246	97.945	97.867	97.867	97.102	97.309	97.761	98.784	98.784
Si	7.427	7.421	7.352	7.592	7.473	7.311	7.638	7.351	7.184	6.922	6.857	7.375	6.853	7.163	6.853	7.163	7.024	6.867	6.878	6.907	7.276	6.793	6.793
Al IV	0.277	0.275	0.358	0.177	0.252	0.232	0.210	0.258	0.362	1.078	1.143	0.625	1.147	0.837	1.034	0.976	1.103	1.103	1.122	1.093	0.724	1.207	1.207
Al VI	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.285	0.284	0.300	0.276	0.266	0.144	0.144	0.222	0.181	0.260	0.267	0.316	0.250	0.250
Ma	0.088	0.133	0.103	0.147	0.098	0.072	0.114	0.121	0.157	0.436	0.451	0.297	0.408	0.291	0.436	0.469	0.531	0.493	0.493	0.393	0.286	0.543	0.543
Mg	3.657	3.648	3.590	3.849	3.796	3.737	3.742	3.702	3.714	2.650	2.647	2.784	2.535	2.880	2.869	2.928	2.902	2.902	2.635	2.960	2.792	2.792	2.792
Si	1.812	1.812	1.812	1.812	1.812	1.787	1.806	1.827	1.850	1.844	1.887	1.956	2.008	1.920	1.888	1.877	1.957	1.855	2.214	1.817	1.817	1.854	1.854
Ca	0.004	0.011	0.013	0.009	0.000	0.023	0.010	0.005	0.004	0.076	0.066	0.036	0.067	0.074	0.071	0.087	0.100	0.105	0.101	0.043	0.090	0.090	
K	0.004	0.008	0.007	0.003	0.000	0.040	0.110	0.006	0.001	0.144	0.133	0.017	0.140	0.106	0.135	0.115	0.125	0.115	0.115	0.115	0.115	0.115	0.115
Cr3+	0.082	0.062	0.065	0.000	0.060	0.017	0.010	0.013	0.044	0.000	0.000	0.000	0.022	0.000	0.017	0.016	0.004	0.002	0.001	0.016	0.000	0.005	0.005
Mn	0.020	0.011	0.014	0.036	0.011	0.219	0.000	0.033	0.018	0.053	0.050	0.041	0.049	0.000	0.017	0.016	0.049	0.040	0.046	0.035	0.066	0.064	0.064

APPENDIX 3

GLOSSARY OF FAULT ROCK TERMINOLOGY

Modified from Obee (1985)

Fault Breccia:	angular fragments of wall rock set in a finer grained matrix of crushed material. They may be cohesive or incohesive. Cementation of the rock may occur after faulting due to mineral precipitation from ground water or hydrothermal fluids.
Fault Gouge:	very fine grained incohesive crushed rock, often with an abundance of clay minerals. Gouges may be foliated or unfoliated.
Cataclasite:	cohesive fault rock consisting of angular to sub-angular fragments within a finer grained matrix, formed by micro-cracking and frictional sliding of grains. A foliation defined by the preferred alignment of clasts or minerals produced by neomineralisation may be present. Cataclasites may be sub-divided according to the proportion of matrix (Sibson 1977): protocataclasite 10-50%; cataclasite 50-90%; ultracataclasite 90-100%.
Mylonite:	cohesive fault rock in which grain refinement is achieved through both brittle and crystal plastic processes and in which neomineralisation contributes to varying degrees. Mylonites are foliated rocks and can be sub-divided according to the proportion of matrix (Sibson 1977): protomylonite 10-50%; mylonite 50-90%; ultramylonite 90-100%.
C-type mylonite:	applied to those mylonites in which brittle processes predominate over crystal plastic deformation processes and in which neomineralisation contributes significantly. First introduced by White (1982) the term has received little usage in the literature but it is suggested that it is a useful term which covers a large group of fault rocks.
Phyllonite:	cohesive fault rocks containing hydrated, phyllosilicate rich assemblages and in which grain refinement has occurred through crystal plastic and brittle processes. Phyllonites appear similar to phyllites in the field. They are usually accompanied by or are indicative of retrogression, although by definition this is not necessarily the case.
Pseudotachylite:	cohesive fault rock generated by melting of the host rock. It is often observed in veins surrounding angular or rounded fragments of host rock.

ACKNOWLEDGEMENTS

In this 'saga of four continents' there have been many many people who have ultimately and finally helped me complete this thesis. First and foremost I must thank Maarten de Wit and Dennis Buchanan who were instrumental in starting up the project and trying to steer me in the right directions. Maarten's enthusiasm and tremendous insight into many aspects of geology, especially in the field, inspired many of the ideas in this thesis. Dennis, with his pragmatic approach to what is required in a thesis, kept me (just, at times!) on the straight and narrow and more than once absorbed the flak for me. I thank you both.

Secondly I must thank Anglo American Corporation of South Africa for fully sponsoring the Pietersburg project. Their willingness to cooperate in all aspects of the study was greatly appreciated and it was a privilege to work with them. The following were also invaluable;

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I dedicate this thesis to three people; my mother and father for their unstinting support and faith in me and to Lia, without whose constant urging I would probably have never finished.

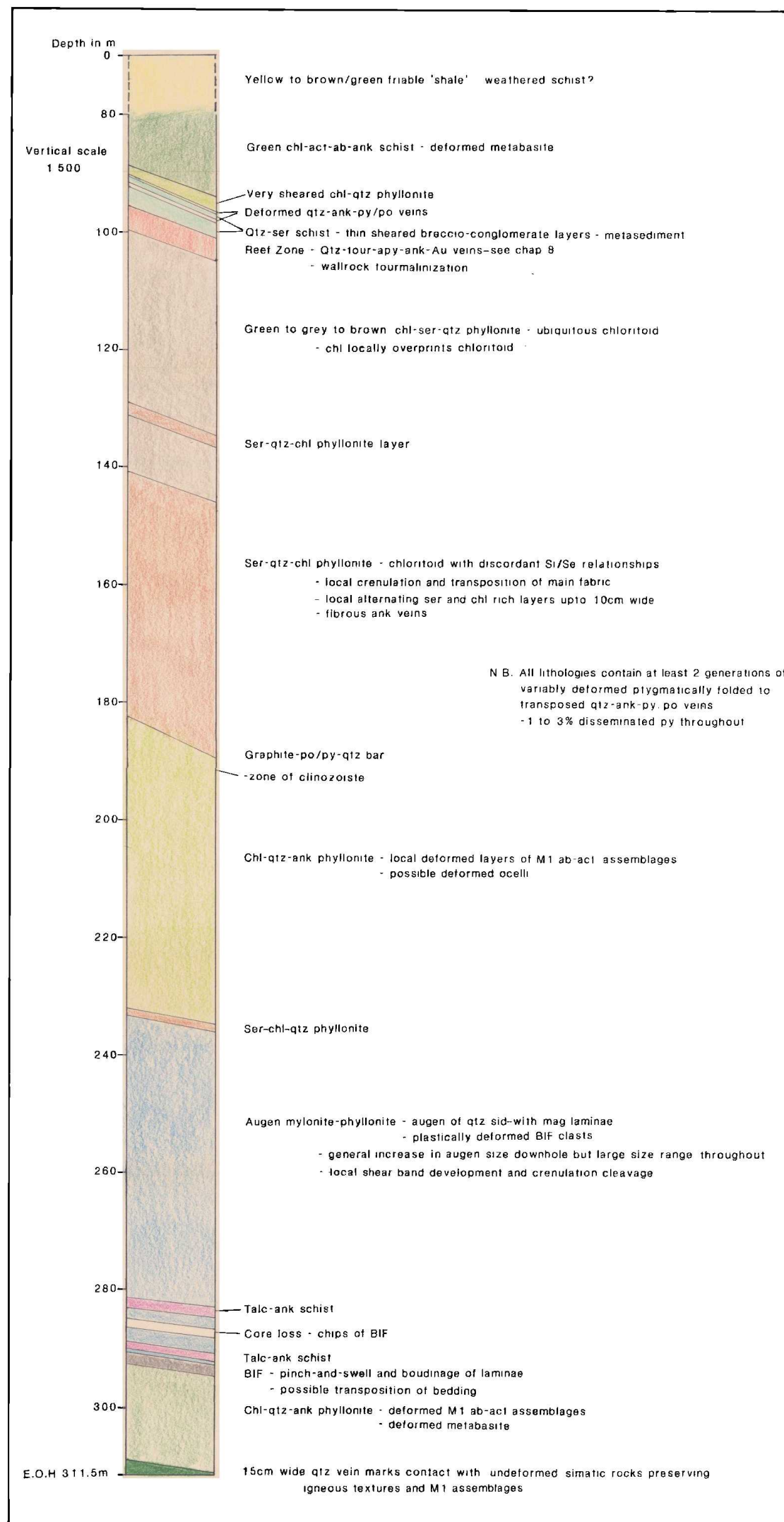


Fig. 5.1 Log of DDH ZR 1

KEY

LITHOLOGY

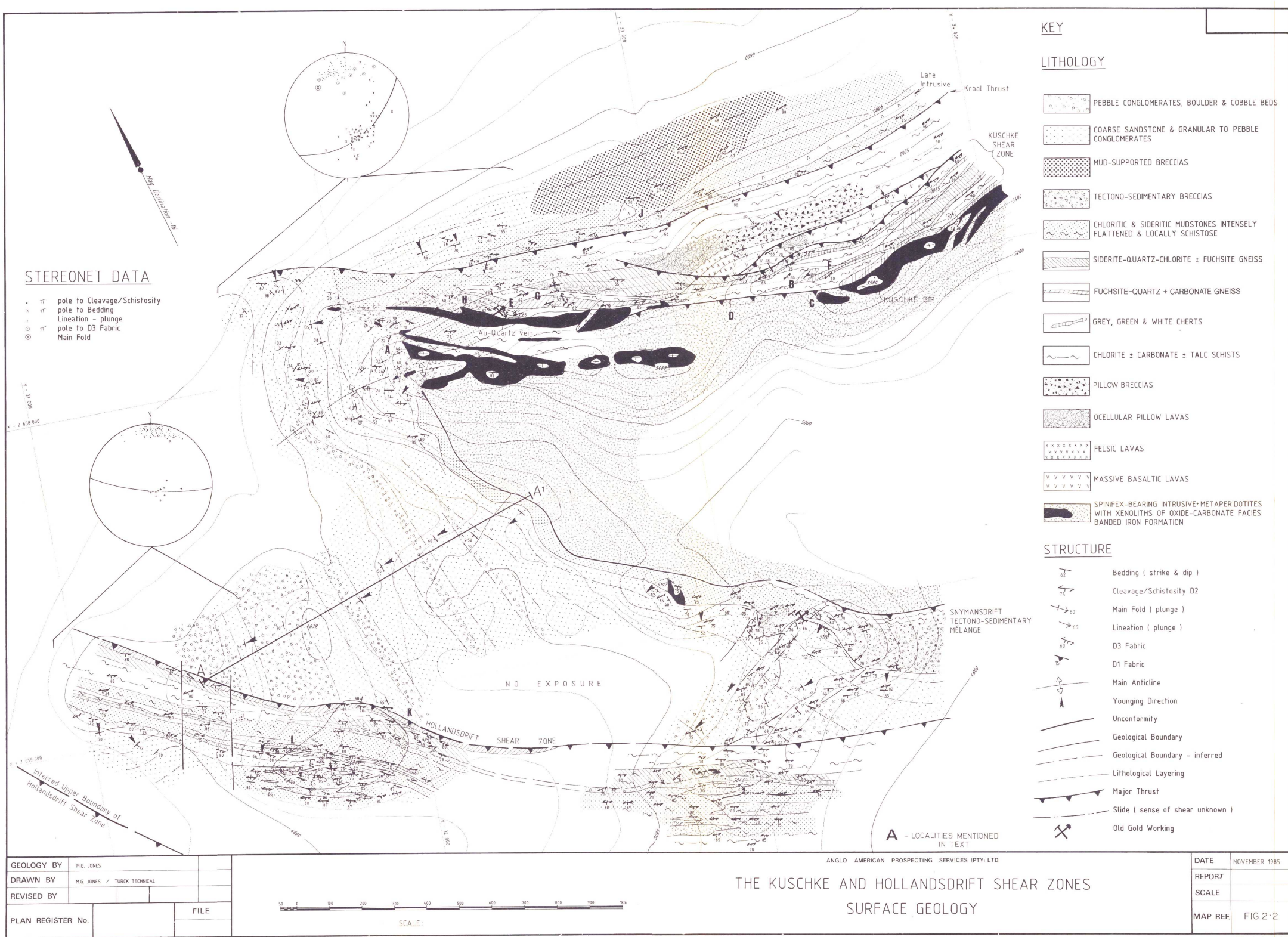
-  PEBBLE CONGLOMERATES, BOULDER & COBBLE BEDS
-  COARSE SANDSTONE & GRANULAR TO PEBBLE CONGLOMERATES
-  MUD-SUPPORTED BRECCIAS
-  TECTONO-SEDIMENTARY BRECCIAS
-  CHLORITIC & SIDERITIC MUDSTONES INTENSELY FLATTENED & LOCALLY SCHISTOSE
-  SIDERITE-QUARTZ-CHLORITE ± FUCHSITE GNEISS
-  FUCHSITE-QUARTZ + CARBONATE GNEISS
-  GREY, GREEN & WHITE CHERTS
-  CHLORITE ± CARBONATE ± TALC SCHISTS
-  PILLOW BRECCIAS
-  OCELLULAR PILLOW LAVAS
-  FELSIC LAVAS
-  MASSIVE BASALTIC LAVAS
-  SPINIFEX-BEARING INTRUSIVE-METAPERIDOTITES WITH XENOLITHS OF OXIDE-CARBONATE FACIES BANDED IRON FORMATION

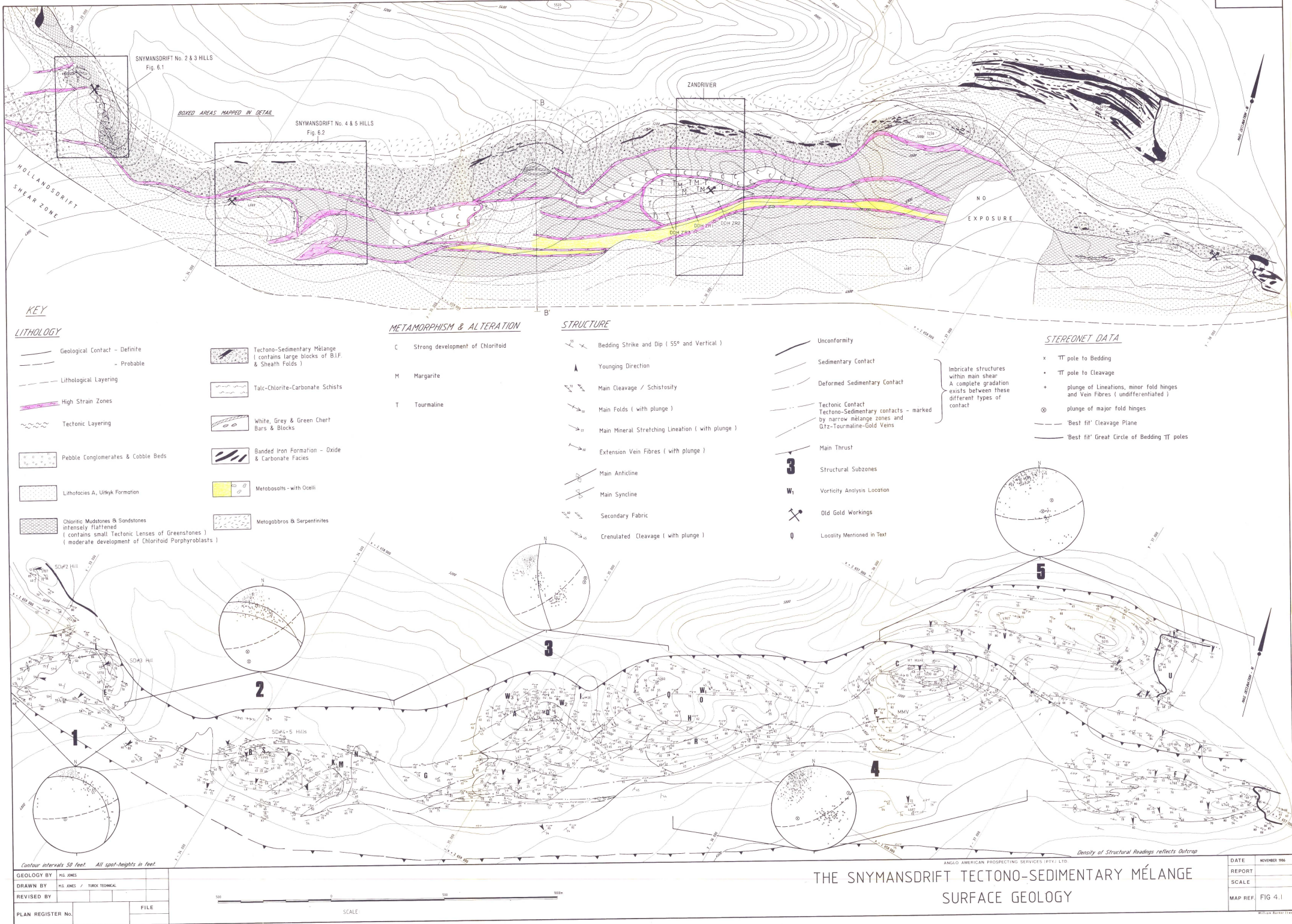
STRUCTURE

-  Bedding (strike & dip)
-  Cleavage/Schistosity D2
-  Main Fold (plunge)
-  Lineation (plunge)
-  D3 Fabric
-  D1 Fabric
-  Main Anticline
-  Younging Direction
-  Unconformity
-  Geological Boundary
-  Geological Boundary - inferred
-  Lithological Layering
-  Major Thrust
-  Slide (sense of shear unknown)
-  Old Gold Working

STEREONET DATA

-  pole to Cleavage/Schistosity
-  pole to Bedding
-  Lineation - plunge
-  pole to D3 Fabric
-  Main Fold





KEY

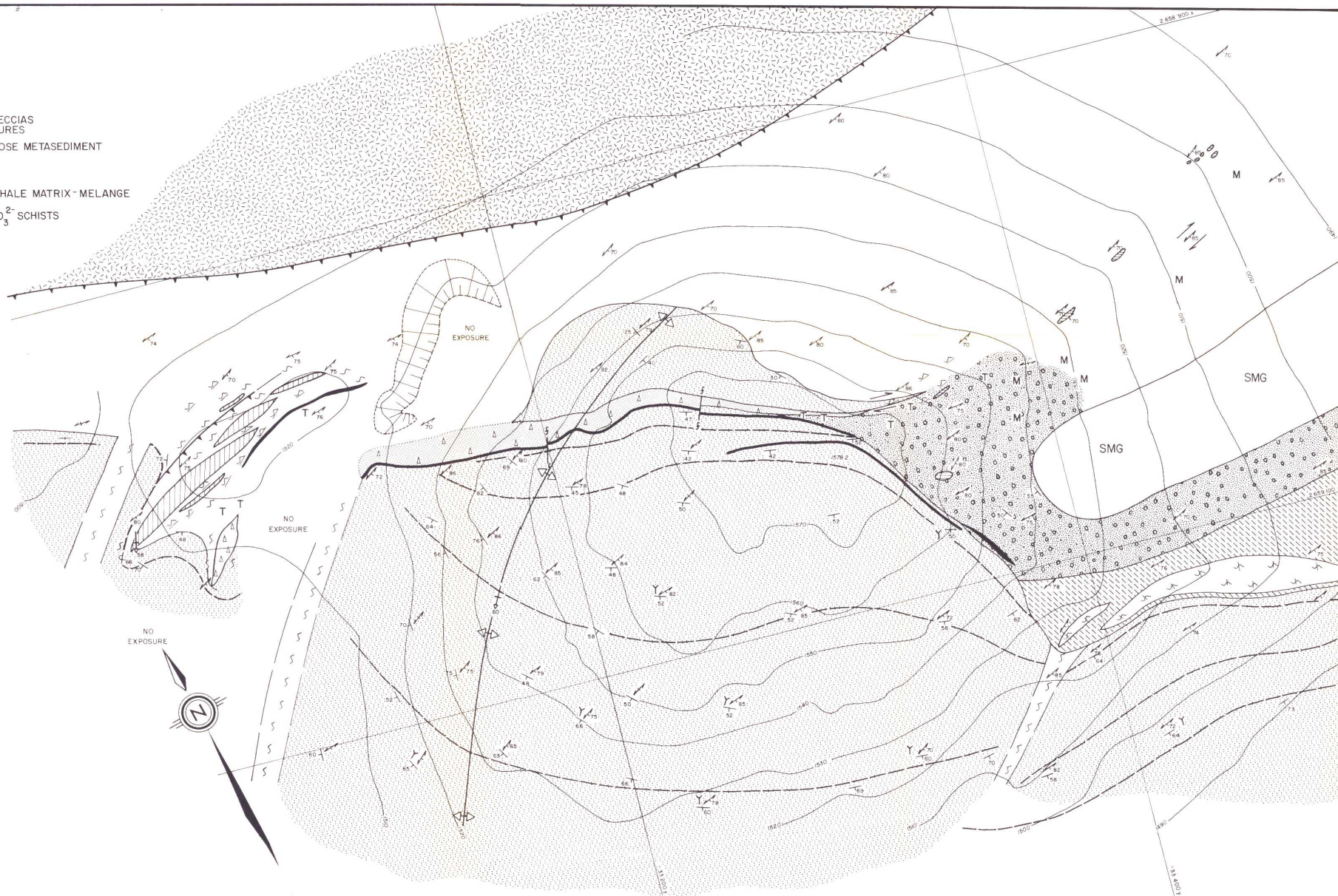
LITHOLOGY

- LITHOFACIES A - UITKYK Fmⁿ
- Chl-Qtz-Lim-Sid-Tour MUDSTONE WITH BRECCIAS AND SOFT SEDIMENT DEFORMATION STRUCTURES
- CHLORITIC CHERT-QUARTZ CLASTIC SCHISTOSE METASEDIMENT
- CHLORITIC FERRUGINOUS MUDSTONES
- BRECCIA - B.I.F. + CHERT CLASTS Chl-Sid-SHALE MATRIX-MELANGE
- CHLORITE, CHLORITE-QUARTZ, Talc-Chl-CO₃ SCHISTS
- UNDEFORMED SERPENTINITE
- WHITE, GREY, GREEN "CHERTS"
- TOURMALINE - CHERT BRECCIA
- SILICEOUS SCHIST
- QUARTZ - TOURMALINE - Au VEINS
- TOURMALINE, MAGNETITE RICH ZONES
- B.I.F. UNITS Sid-mag-qtz
- SMG SHEARED TO UNSHEARED METAGABBRO

STRUCTURE

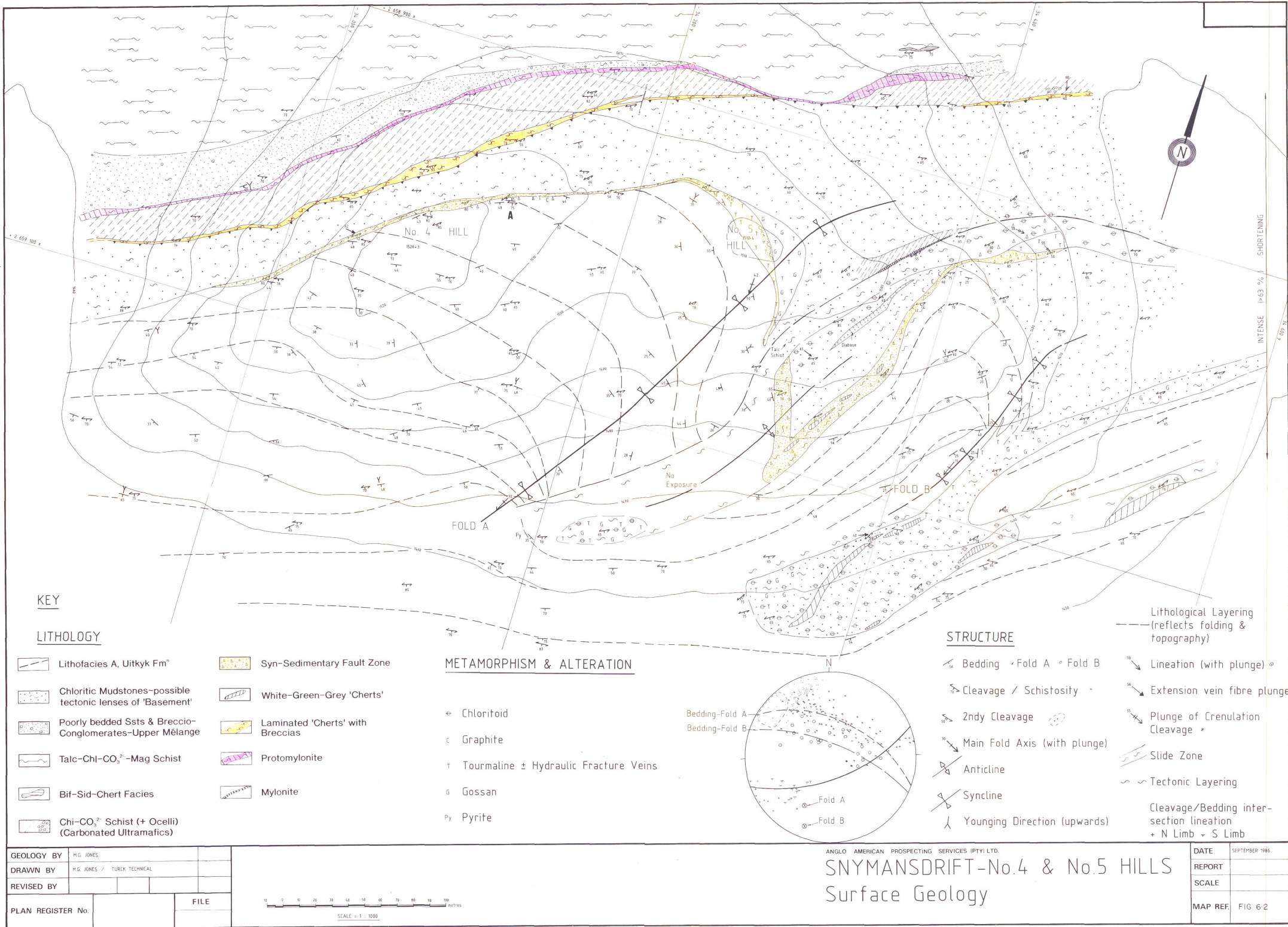
- BEDDING - OVERTURNED
- CLEAVAGE / SCHISTOSITY
- MAIN FOLDS (with plunge)
- ANTICLINE
- MINOR FAULT
- YOUNGING DIRECTION (upwards)
- SENSE OF SHEAR
- LITHOLOGICAL LAYING
- SLIDE ZONE
- GEOLOGICAL BOUNDARY
- THRUST (teeth on upper plate)

CONTOUR INTERVALS = 10m.



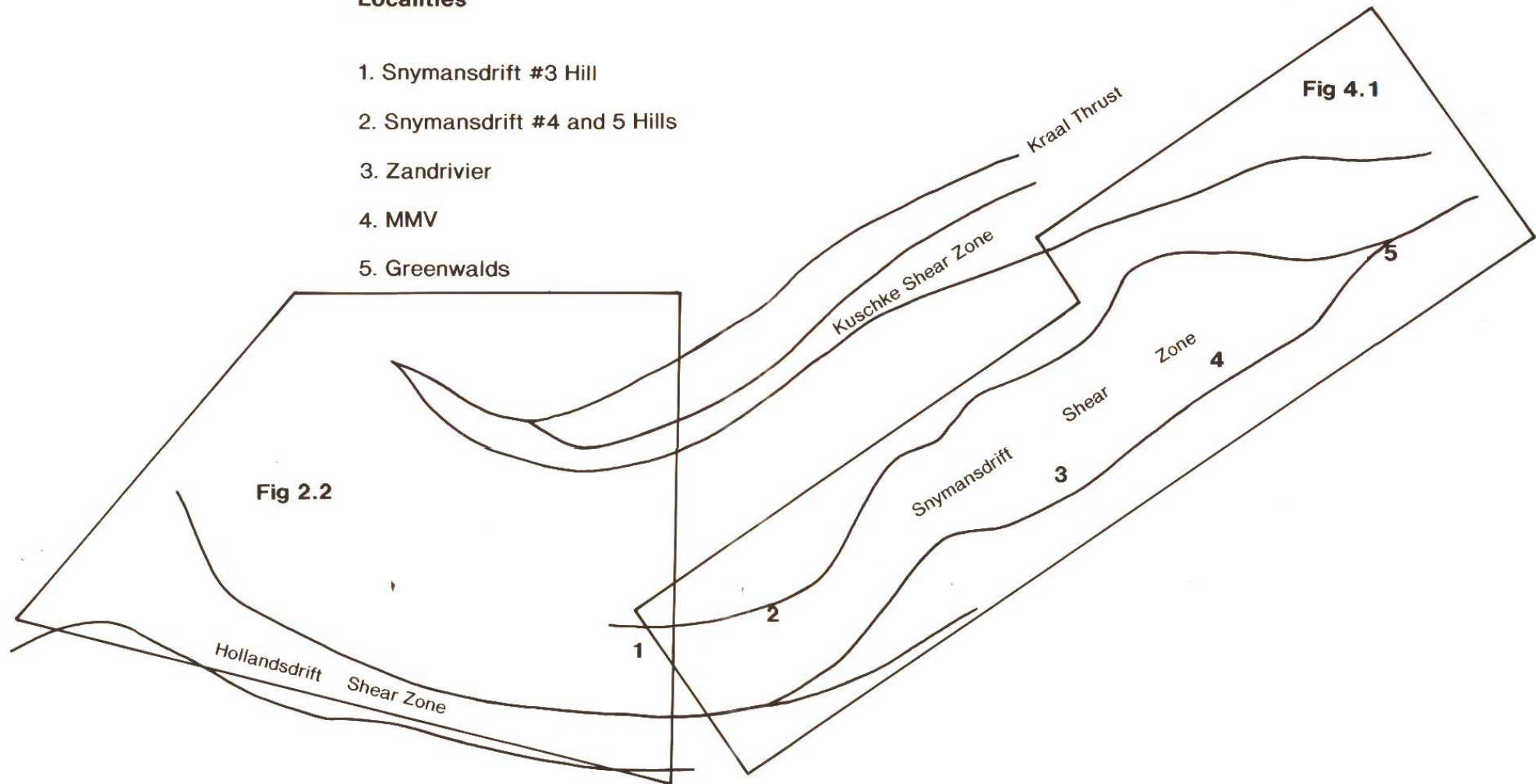
SNYMANSDRIFT No. 2 and 3 HILLS SURFACE GEOLOGY

GEOLOGY BY	M.G. JONES	DATE	SEPTEMBER 1986
DRAWN BY	M.G. JONES / TURCK TECHNICAL	DATE	OCTOBER 1986
REVISED BY	M.G. JONES / C.E. DINES	DATE	JULY 1990
SCALE	1:1000	MAP REF.	FIGURE N° 6-1



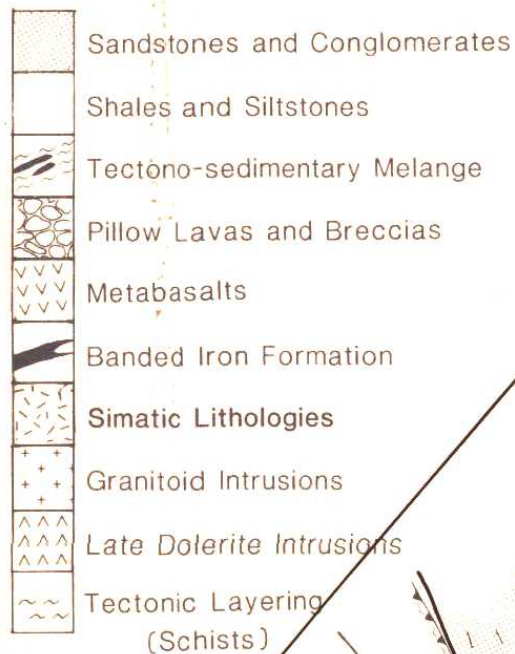
Localities

1. Snymansdrift #3 Hill
2. Snymansdrift #4 and 5 Hills
3. Zandrivier
4. MMV
5. Greenwalds



Overlay for Fig 1.3 showing areas mapped in this study.

Legend



Localities

1. Snymansdrift #3 Hill
2. Snymansdrift #4 and 5 Hills
3. Zandrivier
4. MMV
5. Greenwalds

N

Fig 4.1

Fig 2.2

Structure

- | | |
|--|-------------------------|
| | Younging Direction |
| | Bedding |
| | Cleavage |
| | Lineation |
| | Anticline (with plunge) |
| | Syncline (with plunge) |
| | Main Thrust |
| | Other Faults |
| | Lithological Layering |

Scale
0 0.5 1 1.5 2Km

Figure 1.3 Geology of the Mt. Maré area, Pietersburg Greenstone Belt. Overlay for Fig 1.3 showing areas mapped in this study. Only basic structural elements shown. Area to south of Kuschke Shear Zone mapped by de Wit (1985). Area to north of Hollandsdrift BIF mapped in this study.