

**ASPECTS OF VOLCANISM AND METAMORPHISM OF THE
ONVERWACHT GROUP LAVAS IN THE SOUTH-WESTERN
PORTION OF THE BARBERTON GREENSTONE BELT**

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DECLARATION

I, Marthinus Cloete, declare that this dissertation is my own unaided work, and has not been presented to any other University for the purpose of obtaining a degree.

A handwritten signature in dark ink, appearing to read 'M. Cloete', is written over a horizontal line.

M. CLOETE

ABSTRACT

A volcanological study of the Onverwacht Group in the southwestern part of the Archaean (~3.5 – 3.2 Ga) Barberton greenstone belt (BGB), South Africa, shows that volcanic extrusion rates of the Komati and Hooggenoeg Formations must have been high to have maintained the degree of submarine sheet flooding that is evident. It is concluded that the volcanic attributes of the Komati and Hooggenoeg Formations are not typical of MOR crust, as has been claimed, but rather closely resemble those of modern oceanic plateaus. The shear-zone-bound basal contact of the Komati Formation suggests that the top of the ancient oceanic plateau was allochthonously emplaced and delaminated from its basal (intrusive) part.

The alteration petrography and alteration geochemistry of the statically metamorphosed pillows of the Komati and Hooggenoeg Formations show that they have undergone high- and low-temperature seafloor metasomatism, respectively. Element fluxes between rock and sea water are found to be similar to high-temperature (350 – 425°C) natural and experimental sea-water-basalt interaction, with only minor differences apparent. However, style and temporal development of alteration sequences are different to that of MORB, suggesting that convective flow was probably regulated differently. It appears that subseafloor convection was regulated mechanically by the emplacement of lava sheet flows that inadvertently sealed the recharge and discharge zones of the convective system, only to be re-opened by subsequent thermal tension cracking.

A detailed evaluation of the regional static and dynamic styles of metamorphism was made to determine their respective origins in the light of the tectonic evolution of the BGB. The metamorphic assemblages and the mineral chemistry of the statically metamorphosed rocks suggest that the seafloor metamorphic imprint is cryptically overprinted by prograde burial/load metamorphism. Medium pressures and greenschist-amphibolite transition-zone grades ($P \sim 4$ kb and $T \sim 520$ °C) were reached in a hydrostatic stress field. Along strike, however, the rocks become dynamically overprinted. The conditions of dynamic metamorphism are slightly lower ($P \sim 2.5$ kb and $T \sim 350$ °C), with mineral chemistries supporting distinct retrograde tendencies.

PT vectors (paths) have been established by linking the PT conditions of seafloor metamorphism, the PT(max) of static metamorphism and the retrograde conditions of dynamic metamorphism. The PTt path is established which has two segments: an early prograde anti-clockwise part and a latter retrograde clockwise part for the statically and dynamically metamorphosed rocks, respectively. Overall the path has a sharp inflection which coincides with the main D2 event of crustal shortening and rotation to vertical of the BGB lithologies. The PTt position of the inflection corresponds to the sedimentation of the uppermost sediments of the BGB which have a distinct orogenic character.

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Fig. 4.39

Schematic diagram illustrating rock displacement by means of rotation associated with isoclinal folding. (a) Rocks undergoing burial metamorphism prior to isoclinal folding and crustal shortening. Isotherms (T_1 , T_2) lie parallel to the stratigraphy and increase in value with increasing stratigraphic depth. The relative axis of rotation of the fold limb is situated close to sample b. Sample d is at the same stratigraphic elevation as a and therefore, of the same grade of burial metamorphism. Due to d's position farther away from the axis of rotation greater displacement and hence change in P and T may be expected. Samples MC210, MC60 and MC150 are simulated by sample positions a, b and c. (b) During rotation of the strata to vertical the isotherms are also transposed. The arc of rotation (stippled curves) shows the magnitude of rock displacement, i.e. structural burial relative to the pressure datum. (c) Subsequent to rotation the isotherms re-equilibrate. Sample b maintains P-T conditions similar to pre-rotational times, and a and c undergo only minor changes in PT. Sample d, however, experiences a major increase in pressure due to structural burial and a sharp rise in T due to (upward) thermal relaxation of the isotherms. It should be emphasised that simple isoclinal folding is used here merely to illustrate rock displacement. Metamorphism resulting from rock displacement can be produced by both simple or complex structural processes.

Fig. 4.40

PT diagrams intended to demonstrate and emphasize the differences in the general magnitude of the tectonic forces involved between (Phanerozoic) regional metamorphism and (Archaean) greenstone belt metamorphism. The schematic PT diagrams show selected PT-paths (a, c) produced by a variety of continental collisional settings which result in typical regional metamorphic assemblages and fabrics, as discussed by Thompson and

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1. INTRODUCTION

1.1 Geological Setting and Significance of the Barberton Greenstone Belt

The Barberton Mountain Land (BML), an area of 120 x 50 km, is situated in the eastern Transvaal Lowveld of South Africa (Fig. 1.1) and consists of mid-Archaean (3500 - 3200 Ma) granitoid-greenstones. It is a strongly folded ENE-trending volcano-sedimentary remnant, entirely surrounded by a variety of intrusive, structurally emplaced and basement granitoids of the Kaapvaal Craton (Anhaeusser *et al.*, 1969; Kröner *et al.*, 1991). The volcano-sedimentary portion of the Barberton greenstone belt (BGB) has been named the Swaziland Supergroup (Anhaeusser, 1975), the Barberton Sequence (SACS, 1980) and the Jamestown Ophiolite Complex, i.e. JOC (de Wit *et al.*, 1987a).

Geologically, the BGB is noteworthy because: (1) it is one of three mid-Archaean greenstone belts known (the others being the Warrawoona Group of the Pilbara Block, W. Australia and the less well-preserved Sebakwean greenstone belt of Zimbabwe, e.g. Windley, 1984), (2) it comprises an unique sequence of some of the best preserved, first-formed lithologies on the planet (e.g. Anhaeusser *et al.*, 1969; Viljoen and Viljoen, 1969a; Saager and Köppel, 1976; Williams and Furnell, 1979; de Wit *et al.*, 1987a, 1992; Byerly and Palmer, 1991), (3) apart from having served Archaean greenstone geology in southern Africa, Canada, Australia and elsewhere as a general working model, it has also, directly and indirectly, assisted world-wide advances in the understanding of primeval life forms and the crustal and geochemical development of the early Earth (e.g. Hart, 1970; Glückson, 1971; White *et al.*, 1971; Green, 1972; Anhaeusser, 1973; Schopf, 1974; Nesbitt *et al.*, 1979; Weaver and Tarney, 1979; Groves and Batt, 1984; Windley, 1976, 1984; Condie, 1980; de Wit, 1987a, 1987b, 1992; Taira *et al.*, 1992).

1.2 Stratigraphy

The BGB is traditionally seen to consist of three major lithostratigraphic units (Fig. 1.2). In ascending order, these are: (1) the Onverwacht Group, including largely ultramafic, mafic and basic volcanic rocks with minor units of felsic volcanic, volcanoclastic and silicified sedimentary units, (2) the Fig Tree Group, comprising a complicated succession of greywacke, shale, chert, dacitic flows and fragmental volcanic rocks, and (3) the Moodies Group, consisting largely of feldspathic to quartzitic sandstone, chert-clast

conglomerate, lesser amounts of siltstone and shale (Lowe, 1991).

Recent structural field studies indicate that the BGB stratigraphy probably consists of several tectonic units and that it should be appraised as a litho-tectonic complex (Williams and Furnell, 1979; de Wit, 1982; Lamb, 1984; Lowe *et al.*, 1985; de Wit *et al.*, 1987a; de Ronde, 1991). The BGB can be further subdivided into a Northern and Southern terrane separated by the regional late-stage, post Moodies, Inyoka Fault (e.g. de Wit *et al.*, 1992; *cf* Lowe, 1991). Kröner *et al.* (1991), however, show that the stratigraphic elements of the BGB essentially young upwards and that although the structures are complex, it basically represents an overall intact sequence.

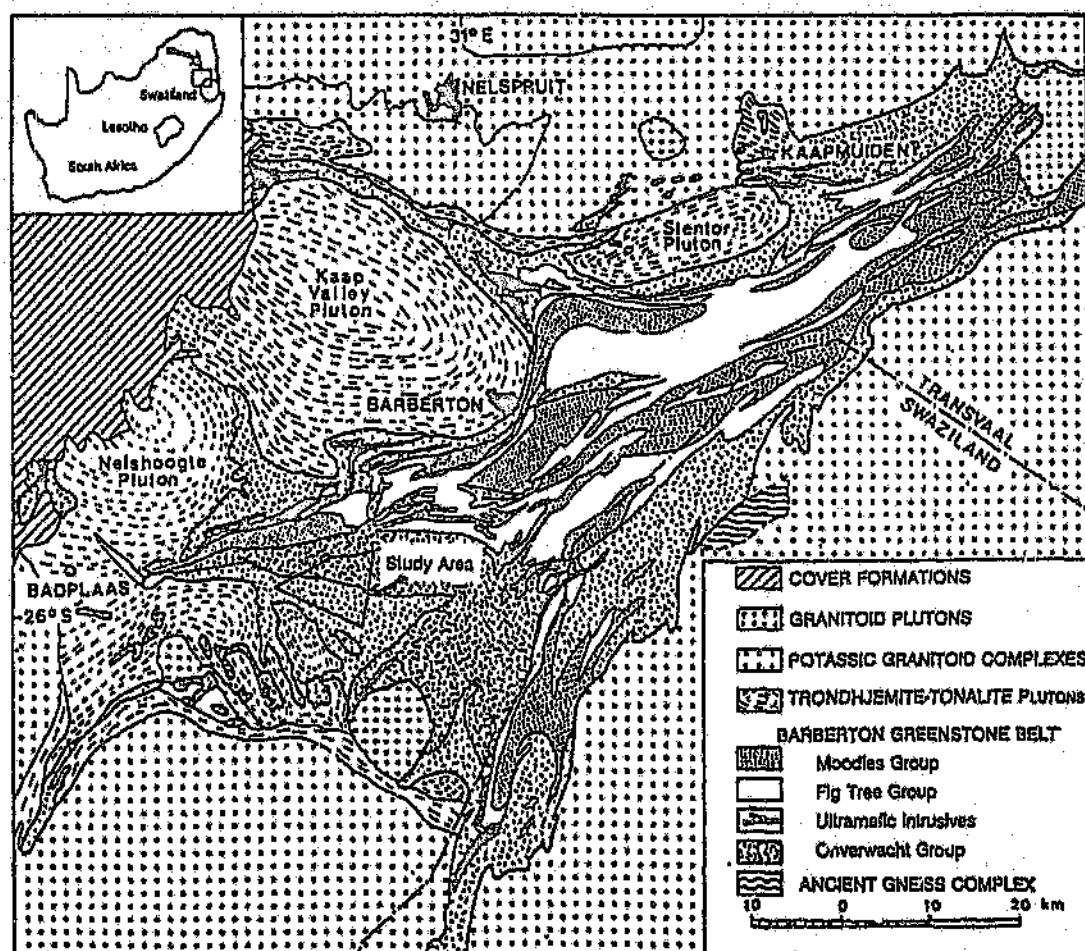


Fig. 1.1 Geological sketch map of the Barberton granite-greenstone terrain (modified after Kröner and Todt, 1988). The arrow in the inset points to the locality of the BGB and that of the main map to the setting of the study area.

1.2.1 The Onverwacht Group

The Onverwacht Group in the southern part of the BGB is revealed in a sequence of large, tight isoclinal, upright, northeast-trending, steeply plunging folds, consisting of the Onverwacht Anticline, Kromberg Syncline, and Steynsdorp Anticline. Here the Onverwacht Group has been divided into six primary formations (Viljoen and Viljoen, 1969a, 1969b; and revised by Lowe, 1991): the Sandspruit, Theespruit, Komati, Hooggengoeg, Kromberg and Mendon Formations (Fig. 1.2). However, due to structural complexity, only the top four formations have in recent times been argued to represent an intact sequence, up to 10 km thick (e.g. Lowe *et al.*, 1985; Lowe, 1991). Armstrong *et al.* (1990) also show by means of age dating that the Komati Formation structurally overlies a younger Theespruit Formation, implying the presence of a major structural discontinuity along the Komati Fault (cf Williams and Furnell, 1979; de Wit *et al.*, 1983), a ~25 km long strike fault folded by the Onverwacht Anticline, the significance of which had been overlooked by earlier workers. Armstrong *et al.* (1990) and Kamo and Davis (1994) confirm the presence of ~3538 Ma tectonic wedges of gneissic tonalite along thrust faults of the Theespruit Formation. Kröner (1993) shows the oldest units of the BGB to be situated in the relatively less well studied Steynsdorp Anticline: the felsic units just north of the core pluton (of the same name) yielding an age of 3547 ± 2 Ma (Table 1.1)

1.2.1.1 *The Sandspruit Formation* is detached from the overlaying Theespruit Formation, and occurs as ultramafic to mafic amphibolitic xenoliths and stringers enveloped by the intrusive granitoids of the Theespruit and Stolzberg Plutons.

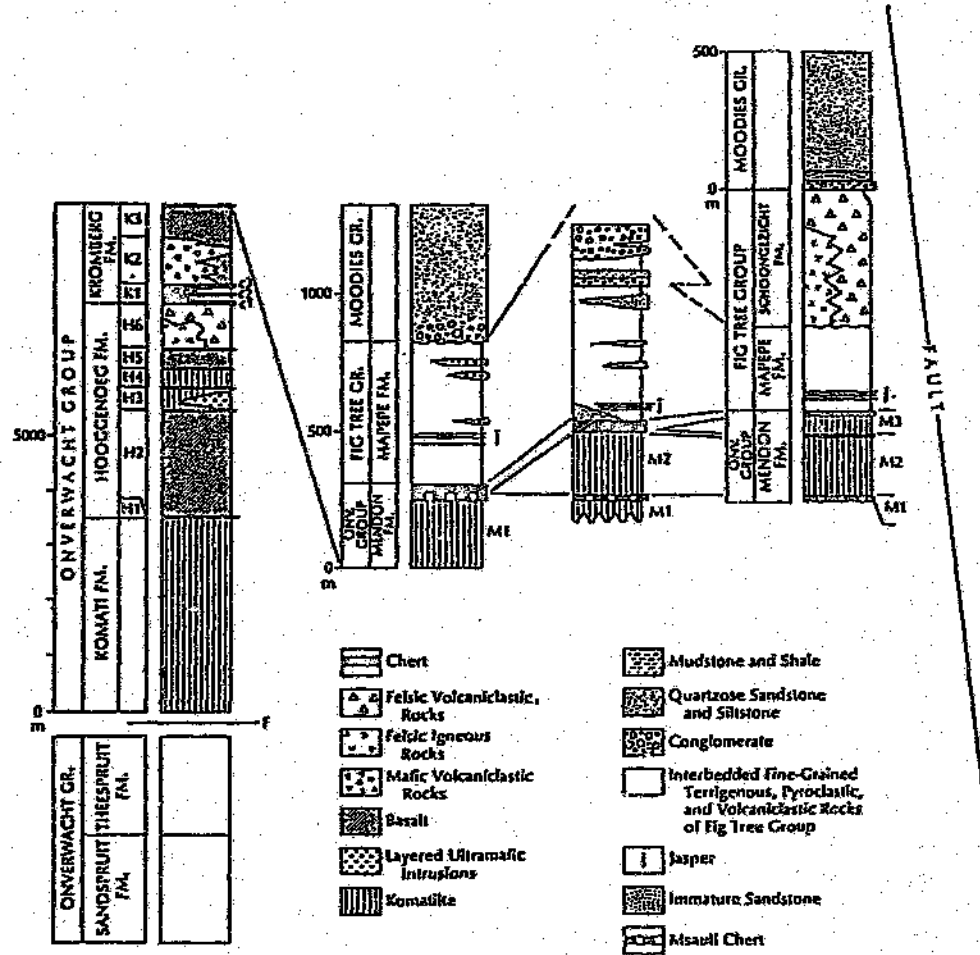
1.2.1.2 *The Theespruit Formation* consists of a varied assemblage of metatholeiitic lavas, tuffs and more seldom of pillows and ocelli-rich rocks. Rare metamafites also occur as serpentinites in the form of conformable and persistent bands, but also as pods and lenses. A diagnostic feature of the formation is the presence of felsic pyroclastic, volcanoclastic and autoclastic units (Lowe, 1982) forming prominent and persistent siliceous horizons. The Theespruit Formation has a lit-par-lit basal contact with the Theespruit Pluton (Viljoen and Viljoen, 1969f).

The Theespruit Formation is variably deformed and consists of a number of litho-

**SOUTHERN
ONVERWACHT ANTICLINE**

FACIES

CENTRAL PARTS OF BELT



N. FACIES

**STOLZBURG
SYNCLINE**

**MOODIES
HILLS**

**EUREKA
SYNCLINE**

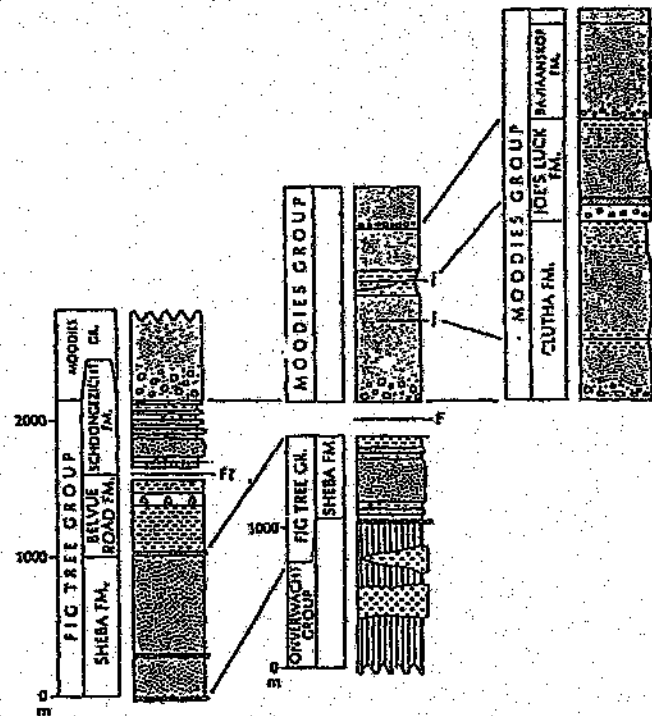


Fig. 1.2 Stratigraphic columns for the western part of the Barberton Greenstone Belt (after Lowe, 1991).

tectonic packages. The northern part is intensely deformed and the mafic and felsic lithologies in places adopt mylonitic fabrics. This zone, 600-800 m wide, is referred to as the Komati Schist Zone (KSZ) and has the Komati and Theespruit Faults bounding it to the north and south respectively (de Wit *et al.*, 1983). Horizontal shortening of the Theespruit Formation (D2 of de Wit *et al.*, 1983) is linked to the emplacement of the trondhjemites (to the south) which have oblate to planar fabrics, similar in geometry and orientation to those of the adjacent greenstones (Viljoen and Viljoen, 1969b; Anhaeusser and Robb, 1980, 1981; de Wit *et al.*, 1983).

1.2.1.3 *The Komati Formation* in the well preserved type-locality area on the western limb of the Onverwacht Anticline attains a maximum thickness of ~3.5 km between its faulted base and the conformably overlying Middle Marker, (Fig. 1.3). Towards the east and the fold hinge of the Onverwacht Anticline thickening occurs due internal disruption and further south along the eastern limb the formation is faulted out altogether. West of the type-locality a ~80% reduction in the thickness of the Komati Formation occurs, which was initially viewed to be stratigraphic and thus an original extrusive feature (Viljoen and Viljoen, 1969). However, due to an abundance of flattened pillows and ocelli in the area (north of the Theespruit Pluton) de Wit *et al.* (1983) ascribe it rather to vertical extension related to emplacement of the nearby granitoids.

The Komati Formation, initially defined and described by Viljoen and Viljoen (1969a,c), consists predominantly of metamorphosed komatiite, komatiitic basalt with minor meta-tholeiite and small intrusive bodies of soda-rich porphyry. The virtual absence of pyroclastic, chemical or detrital sediments interlayered with the lava flows is noteworthy. The komatiites and komatiitic basalts form a steeply dipping, NNE-younging sequence of flows with an intricate compositional development. The lower half of the formation comprises komatiite (MgO = ≥ 24 wt%, olivine-dominated, massive and spinifex-textured) interlayered with low-MgO komatiitic basalt (MgO = 8-16 wt%, pyroxene-dominated, massive, pillowed, and seldom ocelli-bearing) whereas the upper half essentially contains only high-MgO komatiitic basalt (MgO = 18-24 wt%, pyroxene-dominated, massive and spinifex-textured with occasional small irregular-shaped pillows). For a more complete description of the geochemistry of the lavas the interested reader is referred to Smith (1980), Smith and Erlank (1982) and Viljoen *et al.* (1983).

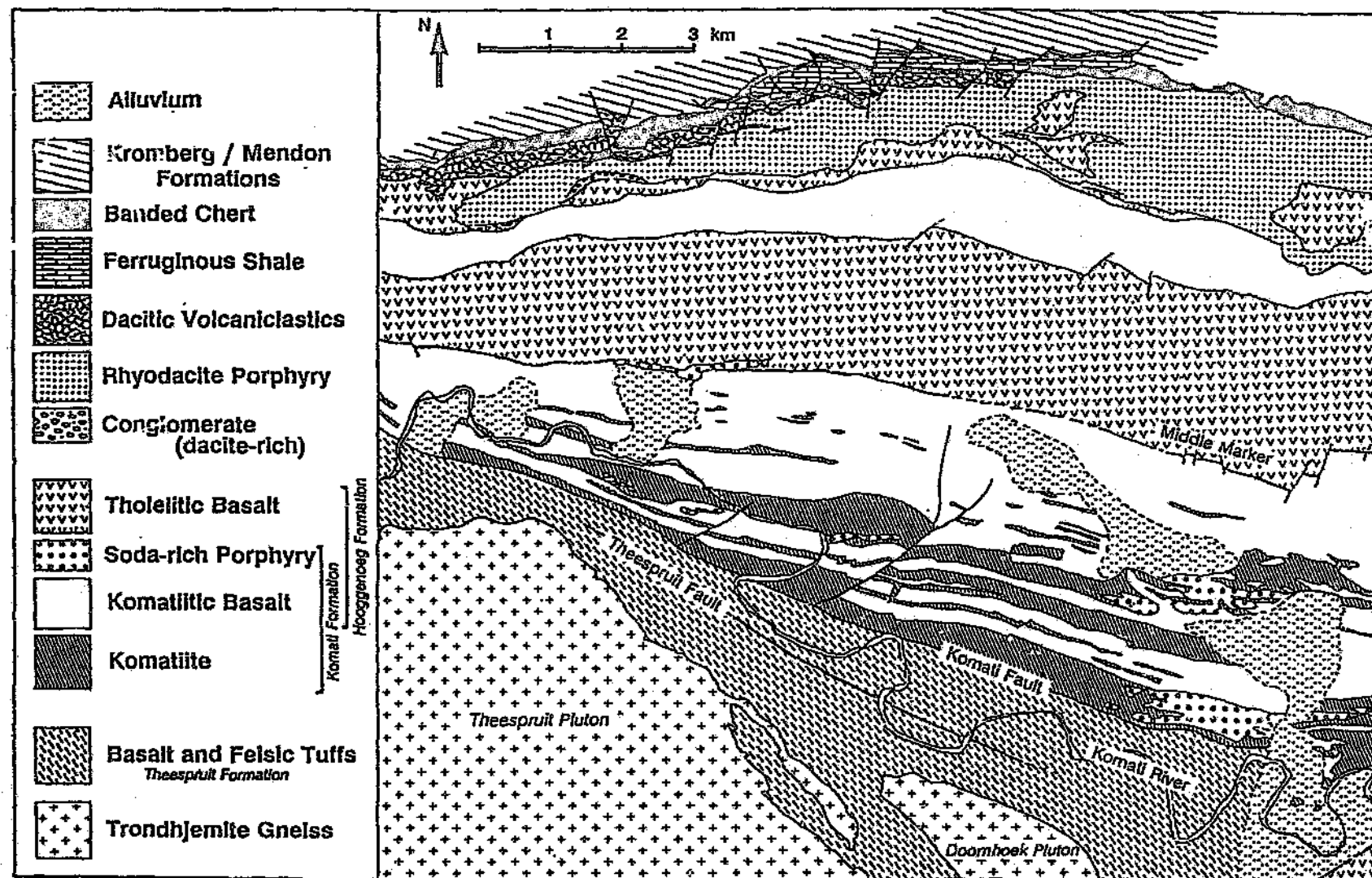


Fig. 1.3 Generalized geology of the Komati and Hooggenoeg Formations on the western limb of the Onverwacht Anticline showing the distribution of the komatiitic and tholeiitic volcanics (after Viljoen et al., 1963, Williams and Furnell, 1970 and de Wit et al., 1983).

Remapping of the area by Williams and Furnell (1979) shows the interlayering of komatiite to be less persistent within the komatiitic basalts than initially showed by Viljoen and Viljoen (1969b).

1.2.1.4 *The Hooggenoeg Formation* has been described as a volcanic sequence having a distinct cyclic nature with five or more cycles developed in the type area (Viljoen and Viljoen 1969d). Each cycle is initially formed by a large build up of tholeiitic basalt which progresses upwards into narrower zones of dacitic to rhyodacitic lava capped by chert. Subsequent workers (Duchac and Hanor, 1987; Ranson and Byerly, 1983), however, indicate that for various localities in the BGB the tops of mafic flow sequences are often pervasively silicified, explaining the observed 'felsic transition' from basalt to chert and rendering the cyclic volcanicity of Viljoen and Viljoen (1969d) unnecessary. Williams and Furnell (1979) also recognised a 1 km thick komatiitic basalt horizon in the centre of the formation extending the entire length of the western limb of the Onverwacht anticline. These observations drastically changed the nature of the Hooggenoeg Formation from being 'cyclic and basic to felsic' to a bimodal mafic extrusive sequence like most other komatiite bearing sequences (Williams and Furnell, 1979). Lowe (1991) recognises the Hooggenoeg Formation as consisting of the Middle Marker, a 2 - 5 m thick sedimentary unit of silicified ash, volcanoclastic sediments and black chert (H1) (Lanier and Lowe, 1982), about 1300 m of massive and pillowed tholeiitic basalt topped by a thin unit of grey chert (H2), two 350 m thick units of komatiitic and minor tholeiitic basalt capped by thin horizons of silicified volcanoclastic debris (H3 and H4), basalt and variolitic basalt overlain by a thin unit of grey chert and silicified volcanoclastic sediment (H5), and up to 750 m of dacitic volcanic and volcanoclastic rocks (H6) (Fig. 1.2).

Unit H6, selected to mark the top of the Hooggenoeg Formation (Lowe *et al.*, 1985), is about 20 km long and has an intrusive base and an erosional top suggesting that its extrusive centre extended above sea-level and was partially eroded during volcanism (Lowe, 1980). However, due to its intrusive base and other features (to be discussed further on), it is suggested that H6 cannot be considered part of the Hooggenoeg Formation. For the ease of reference this important unit will be referred to as the Buck Reef Porphyry. The Buck Reef Porphyry and its associated volcanics and volcanoclastics

were subsequently capped, at the onset of the Kromberg time, by the Buck Reef Chert consisting of shallow-marine non-volcanogenic sediment (Lowe, 1991). From the Middle Marker to the base of the Buck Reef Porphyry on the western limb of Onverwacht Anticline, the Hooggenoeg Formation attains thicknesses of between 3 and 4 km. The Buck Reef Porphyry has also been shown to be coeval and cogenetic with the deep-seated (3445 Ma) trondhjemites that intrude the base of the Onverwacht Group (de Wit *et al.*, 1987b; Armstrong *et al.*, 1991).

Inferences of possible structural thickening of the Hooggenoeg Formation were initially indicated by Williams and Furnell (1979) due to presence of fold-like structures in the upper half of the western part of the Onverwacht Anticline. De Wit (1982) and de Wit *et al.* (1983) remapped large parts of the Onverwacht Group and resolved the upper part of the Group to be structurally thickened by large-scale isoclinal folding. In contrast to the upper Onverwacht (Hooggenoeg and Kromberg Formations) forming primarily a continuous, northerly-younging, almost vertical to slightly overturned stratigraphic succession in excess of 8 km thick (Viljoen and Viljoen, 1969a, b), de Wit (1982) argued that it represents a much thinner section, structurally duplicated by large-scale isoclinal folding. Lowe *et al.* (1985) have critically remapped the same area and in a substantial account conclude that despite the intrusion of felsic magma (the Buck Reef Porphyry) the presence of unfaulted sections to the east of the area indicate that the gross stratigraphic order of units within the group is not significantly changed. They conclude: "Structures previously mapped as major fold hinges within the upper part of the Onverwacht Group can be shown to have formed by local deformation and reflect neither large-scale folding nor structural repetition of the rocks". Subsequently de Wit *et al.* (1992) contend thickening of the Onverwacht by means of repetitive thrust stacking.

1.2.1.5 The Kromberg Formation, on the western limb of the Onverwacht Anticline, consists of the very prominent Buck Reef Chert overlying the Buck Reef Porphyry and its associated volcanic and volcanoclastic rocks. Along the type section the Kromberg Formation includes about 1700 m of volcanic and sedimentary rocks of three main lithofacies: (1) massive and pillowed basalt, (2) mafic lapilli tuff and lapillistone, and (3) black and banded chert (Lowe, 1991; Byerly and Lowe, 1991). The top of the Kromberg Formation is placed at the top of the regionally traceable Footbridge Chert.

1.2.1.6 *The Mendon Formation* comprises komatiitic basalt and massive komatiite overlying the Footbridge Chert (Lowe, 1991). Initially these komatiitic rocks were included as part of the Kromberg Formation (Viljoen and Viljoen, 1969b) but regional mapping shows them to be part of cyclic sequences of interbedded komatiitic lavas and cherts that overlie the mainly basaltic Kromberg Formation.

Lowe (1991) and Lowe and Byerly (1991) report that the Mendon Formation volcanics crop out throughout the central portions of the BGB. The central parts of the belt north of the western limb of the Onverwacht Anticline are subdivided into a series of narrow structural blocks by a series of regional, bedding-parallel faults (*cf* also de Ronde, 1991). The fault-bound packages usually contain upper Mendon Formation and lower Fig Tree Group sediments. In places, along the southern parts of the Onverwacht Anticline, the Mendon Formation comprises a single volcanic unit capped by the 20 - 35 m thick Msauli chert and overlying successions of black, banded and ferruginous cherts. The chert is overlain conformably by the Fig Tree sediments (Lowe, 1991).

The deformation of the fault-bound packages (D_1) precedes deposition of the overlying Moodies Group and the deformational event (D_2) responsible for major horizontal shortening and vertical rotation of the entire greenstone belt (de Wit, 1982; de Wit *et al.*, 1983; Lowe *et al.*, 1985; Lamb and Paris, 1988).

1.2.2 The Fig Tree Group

The Fig Tree Group consists of a southern shallow-water facies encompassing thick banded iron formations overlain by coarser clastics and a northern deep-water facies dominated by graywackes and shales and lesser amounts of chert and banded iron formation (Eriksson, 1980; de Wit, 1982; Lamb and Paris, 1988). Lowe (1991) has divided the strata south of the Inyoka Fault (i.e. the southern facies) into the Schoongezicht and Mapepe Formations. The lower Mapepe Formation sediments are considered to consist mainly of terrigenous shale, chert-grit sandstone and chert-clast conglomerate interstratified with thick units of fine-grained dacitic, pyroclastic and volcanoclastic sediments. Chert, jasper and barite constitute minor parts of most sections (*cf* also de Ronde, 1991). Kröner *et al.* (1991) show that the dacitic tuffs in the Mapepe Formation have ages ranging from ~ 3260 to 3225 ± 4 Ma. The Schoongezicht

Formation consists of 1500 to 2000 m of dacitic tuff, coarse volcanoclastic sandstone, conglomerate and breccia (Lowe, 1991).

The northern facies of the Fig Tree Group occurs to the north of the Inyoka fault and is well out of the influence sphere of the study area and is therefore not discussed further. The stratigraphic column of the northern facies, however, is presented for completeness in Fig.1.2.

1.2.3 The Moodies Group

The Moodies Group constitutes the youngest sedimentary rocks of the Barberton Greenstone Belt and consists of felspathic and quartzose sandstone, chert-clast conglomerate and siltstone. Chert-clast conglomerates at the bottom of the Moodies have a wide-spread occurrence in the southern areas and rest with an angular unconformity on the Onverwacht Group (Jackson *et al.*, 1987; Lowe, 1991). Jackson *et al.*, 1987) and others have proposed a foreland basin setting for Moodies sedimentation. Environments of sedimentation include braided alluvial plain, deltaic, shallow-water coastal systems and shelf (Eriksson, 1977, 1978).

1.3 Intrusive Rocks

De Wit *et al.* (1987b) identified two types of early plutonic rock associated with higher-level porphyries and volcanic rocks: a trondhjemitic suite, distinguished by $\text{SiO}_2 > 68\%$; and a tonalitic suite with $\text{SiO}_2 < 68\%$. The ~3440 Ma trondhjemitic suite occurs only along the southwest margin of the BGB, appears to intrude only the Onverwacht Group and seems to be co-magmatic with the Buck Reef Porphyry and allied volcanics. The intrusives of this group include: e.g. the 3448 ± 4 Ma Doornhoek Pluton (Kamo and Davis, 1993), the 3443 - 3438 Ma Theespruit Pluton (Kamo and Davis, 1993; Kröner *et al.*, 1991; Armstrong *et al.*, 1990) and the 3460 Ma (Kamo and Davis, 1993) and 3445 Ma (Kröner *et al.*, 1991) phases of the Stolzburg Pluton. In contrast, the tonalitic suite is significantly younger (~200 Ma) and appears to intrude the entire Barberton Sequence and occurs mostly in the central and northern parts of the belt. Related porphyries intrude the Fig Tree (Kröner *et al.*, 1991; de Ronde, 1991; Kamo and Davis, 1993) and Moodies Groups (Heubeck and Kröner, cited in Kröner, 1993) whereas associated volcanoclastic rocks comprise a significant part of the Fig Tree Group (Kröner *et al.*,

1991, see Table 1.1 for the relevant ages). The main plutonic phase of this suite is the 3227 ± 1 Ma Kaap Valley tonalite and includes the southerly situated, small $3216 +2/-1$ Ma Dalmein granodiorite (Kamo *et al.*, 1990). These granitoids, i.e. the tonalites, trondhjemites and granodiorites, are often referred to as the TTG intrusive suite (Kröner and Todt, 1988; Kröner *et al.*, 1991).

Younger potassium-rich plutons also intrude the granite-greenstone terrane at about 3100 Ma but seem not to have affected the metamorphic (assemblages) history of the study area and are therefore not considered further in this project. The interested reader, however, is referred to Kamo and Davis (1991, 1994) for more information.

1.4 Geochronology

The substantial improvement in geochronological techniques in recent years (e.g. Compston *et al.*, 1986; Kober, 1986), has led to a flurry of new dates for the supracrustals and intrusive rocks of the Barberton Greenstone Belt. The new ages also mark the end of an era of geochronology that was generally useful but not accurate enough to resolve detailed aspects concerning the igneous, deformational, metamorphic and stratigraphic development of Archaean greenstones. A comprehensive review of the geochronology prior to the advent of precise zircon age dating will not be considered as it now seems irrelevant and may confuse the uninitiated reader (reviews are available from e.g. Kröner *et al.*, 1991; Kamo and Davis, 1991; 1994). A selection of the most recent and relevant ages of the BGB and surrounds are summarised in Table 1.1 according to the generalised classical stratigraphy. An extensive discussion involving the age data and the significance of the lithologies with regard their temporal and spatial development is given in Chapter 5 of this work.

1.5 Metamorphism and previous work

1.5.1 Barberton Greenstone Belt Metamorphism

An analysis of the metamorphic history of the Barberton greenstone belt based on detailed petrography and mineral geochemistry has not yet been attempted. Data on the metamorphism of the Barberton Greenstone Belt is therefore sparse and rare accounts often only describe in general the metamorphic minerals encountered but with no mention of assemblages or equilibrium state. In the literature the types of metamorphism

Table 1.1 Ages for selected lithologies of the Barberton Greenstone Belt and surrounds.

Unit	Rock type	Age in Ma	Dating method	Interpretation	Ref.	
Barberton Greenstone Belt, metavolcanic rocks and dykes						
Moodies Group	Felsic dykes cutting Moodies	3205 ± 6	Pb-Pb evaporation	Maximum age for Moodies deposition	1	
		3222			2	
Fig Tree Group	Dacitic tuff & agl'm	3225 ± 3 to 3259 ± 5	Pb-Pb evaporation	Duration of f. volc.	3	
	Undeform. porphyry	3227 ± 3	U-Pb conventional	Min. age of thrust	4	
	Deformed porphyry	3229 ± +4/-3	U-Pb conventional	Max. limit of thrust	5	
Onverwacht Group						
Mendon Fm.	Dacitic lava	3298 ± 6	Pb-Pb evaporation	Felsic volcanism	6	
Hooggenoeg Fm.	Dacitic tuff	3416 ± 5 to 3445 ± 3	Pb-Pb evaporation	Duration of f. volc.	3	
	Felsic intrusion	3445 ± 8	U-Pb SHRIMP	Felsic volcanism	7	
	Felsic lava	3438 ± 12 to 3451 ± 15	Pb-Pb evaporation	Duration of f. volc.	8	
	Inflow. sed. M Marker	3472 ± 5	U-Pb SHRIMP	Detrital material from underlying volcs.	7	
Komati Fm.	Metagabbro	3352 ± 6/-5	U-Pb baddeleyite	Mafic magmatism	5	
	Felsic dyke	3467 ± 12/-7	U-Pb conventional	Min age of Komati F.	5	
	Inflow. sed. Komati U.	3472 ± 5	U-Pb SHRIMP	Penecontemporaneous detrital material	7	
	Komatitite	3486 ± 8 Ma	⁴⁰ Ar/ ³⁹ Ar	Age of Metamorphism	9	
Theespruit Fm.	Volcaniclastic sediment	<3453 ± 6	U-Pb SHRIMP	Min age of felsic volc.	7	
Theespruit Fm.?	Steynsdorp Felsic volc.	3547 ± 2	Pb-Pb evaporation	Felsic volcanism	10	
Tonalite Trondhjemite Granitoid (TTG) gneisses and plutons in the Barberton region						
Dalmela Pluton	Granodiorite	3216 +2/-1	U-Pb conventional	Age of emplacement	5	
Knap Valley Pluton	Tonalitic gneiss	3226 ± 14	U-Pb SHRIMP	Age of emplacement	7	
		3229 ± 5	U-Pb conventional		11	
		3227 ± 1	U-Pb conventional		5	
Theespruit Pluton	Trondhjemitic gneiss	3440 ± 5	Pb-Pb evaporation	Age of emplacement	3	
		3443 ± 3	U-Pb conventional		5	
		3437 ± 6	U-Pb SHRIMP		7	
Doornhoek Pluton	Trondhjemitic gneiss	3448 ± 4	U-Pb conventional	Age of emplacement	5	
		-3215	Pb-Pb sphene	metamorphism	12	
		3191 ± 46	Rb-Sr whole-rock		13	
Stolzberg Pluton	Trondhjemitic gneiss	3445 ± 3	Pb-Pb evaporation	Age of emplacement	3	
		Early phase of plut.	3460 ± 5/-4	U-Pb conventional		12
		Late-stage event	3237 - 3201	Pb-Pb zirc. + sph.	Age of metamorphism	12
Steynsdorp Pluton	Gneissic tonalite	3509 ± 8/-7	U-Pb conventional	Age of emplacement	5	
		3510 ± 4	Pb-Pb evaporation		10	
		3531 ± 3	Pb-Pb evaporation	Zircon xenocryst	3	
Gneissic wedges in Theespruit Fm.	Tonalitic gneiss	3538 ± 6	U-Pb SHRIMP	Pre Onverwacht crustal history	7	
		3538 +4/-2	U-Pb conventional		5	
Ancient Gneiss Complex						
Ngwane Gneiss (NG)	Tonalitic to trondhjemitic gneiss	3500 to 3563 ± 3	U-Pb SHRIMP	Age of emplacement	14,15	
		3644 ± 2	U-Pb SHRIMP	Age of emplacement	16	
References: 1. Heubeck & Kröner, unpubl., 2. Layer et al., submitted, 3. Kröner et al. (1991), 4. de Ronde, 1991, 5. Kamo et al. (1990), 6. Byerly & Kröner, unpubl., 7. Armstrong et al. (1990), 8. Lopez-Martinez et al., 1992, 9. Kröner and Todt, 1988, 10. Kröner & Hegner, unpubl., 11. Tegtmeier and Kröner (1987), 12. Kamo and Davis, submitted, 13. Barton et al., 1983, 14. Kröner et al. (1989), 15. Wendt & Kröner, unpubl., 16. Kröner and Tegtmeier, in press, 17. " = cited in Kröner, 1993.						

that have been recognised for the BGB can be grouped into four categories:

- 1) dynamic metamorphism associated with the emplacement of granitoids,
- 2) pervasive regional low-grade metamorphism remote from the granitoids,
- 3) sea-floor metamorphism and
- 4) burial metamorphism (related to this project).

1.5.1.1 Dynamic Contact Metamorphism. Hall (1918) studying the contact relationships between the Larberton greenstones and the surrounding granitoids pointed out that the granites were intrusive into the greenstones and had produced a contact aureole '...some three miles thick in places'. The most common minerals constituting the aureole were documented as: andalusite, ottrelite (a chloritoid variety), biotite, garnet, sillimanite, tourmaline, zoisite, staurolite, cordierite and corundum, in decreasing order of abundance.

For the southern part of the BML Viljoen and Viljoen (1969) showed that closer towards granite contacts the rocks are of upper greenschist and amphibolite facies metamorphism and commonly are dark green to almost black in appearance. Away from the contacts, metamorphic grade diminishes rapidly to greenschist assemblages, manifested generally by lighter coloured rocks. The more aluminous felsic horizons contained, in addition to quartz and sericite, marked amounts of pyrophyllite and occasional andalusite, chloritoid and staurolite. Close to the Theespruit Pluton contact, de Wit *et al.* (1983) observed almandine garnet in meta-mafic and kyanite in 'meta-pelitic' rocks. The thermal development of these assemblages is thought to be related to granite emplacement which was synchronous with the main event of maximum horizontal shortening (D₂), as the geometry and orientation of fabric developed within the granite and adjacent greenstones are similar (de Wit *et al.*, 1983).

Similarly, in the northern parts of the greenstone belt, Anhaeusser *et al.* (1969) and Anhaeusser (1984) have described the occurrence of muscovite, biotite, amphibole, garnet, andalusite, sillimanite and chloritoid (medium-grade amphibolite facies) occurring in the relatively narrow contact aureole of the intrusive granites. The thermal development of these minerals is intimately associated with successive stages of deformation which was followed by extensive post-tectonic recrystallization and, finally,

late-stage retrogression. Approximately 5.5 km away from the granitoid contact, at Sheba Gold Mine, Schouwstra and de Villiers (1988) calculated a pressure of 4.2 kb and temperatures of 300 - 425°C from coexisting sulphides in mineralized shear zones. Tomkinson and Lombard (1990) describe for the New Consort Mine a process of continued (dynamic) contact metamorphism taking place in the presence of high-temperature hydrothermal fluids. This led to the formation of pegmatites (containing e.g. spodumene $\pm$ holmquistite), skarns (comprising diopside-hedenbergite, amphiboles, vesuvianite, garnet, scapolite, bytownite and scheelite) and the amphibolitization of metasediments. Mineralization is associated with the growth of e.g. phlogopite and tourmaline and sodium metasomatism in a range of 520 - 580 °C.

Anhaeusser (1983) has documented meta-pelite and calc-silicate assemblages in the 12 km long and 2.5 km wide Schapenburg greenstone remnant preserved in the early trondhjemitic granitoids of the southern mountain land. The metapelites contain grunerite, almandine, cordierite, andalusite, staurolite, epidote/zoisite as well as quartz and magnetite. The calc-silicates consist of diopside, garnet, microcline, epidote/zoisite and quartz. Interlayered metagreywackes are reported to show signs of incipient anatexis melting. Pressures and temperatures for the area are estimated to be in the order of 4 - 6 kb and 500 - 700°C (T. Walmach, 1990, personal communication).

1.5.1.2 Regional Low Grade Metamorphism. Anhaeusser *et al.* (1969) were the first to note a pervasive type of greenschist facies metamorphism remote from the peripheral contact aureoles. These authors were uncertain as to the cause of this metamorphism and suspected high mantle heat flow through a thin crust. The previously mentioned dynamothermal contact aureoles were ascribed to rapid cooling as the granites moved up through the crust to higher levels.

1.5.1.3 Sea-floor Metamorphism. This style of metamorphism has been inferred and described for numerous localities in the Barberton greenstone belt. Two types have been implied: (1) a low-temperature type and (2) a high-temperature type. The low temperature type is thought to involve mostly shallow marine convection and temperatures between 60° to 150°C (Lowe and Byerly, 1986; de Ronde, 1991) and is characterised by the pervasive silicification of a wide range of sediments (Lowe and

Knauth, 1977; Lanier and Lowe, 1982; de Wit *et al.*, 1982) and volcanic rocks containing igneous features such as spinifex, pillows and ocelli (de Wit *et al.*, 1982). The silicification is found to be generally related to submarine exhalative activity and seawater convection within the oceanic crust to produce silica-enriched fluids which in turn silicified lava and sediments on re-entering the near-surface environment (Stanistreet and Hughes, 1984; Paris *et al.*, 1985). Duchac and Hanor (1987) similarly attribute the metasomatic conversion of spinifex- and cumulate-textured rocks, now consisting primarily of quartz, mica and chlorite, to the dynamic, hydrothermal upwelling of fluids through original fracture porosity in the ultramafic flows. Lowe and Byerly (1986) maintain that the tops of komatiitic lavas underlying chert beds and regionally marked by thick zones of silicification, formed largely by low-temperature, near-surface alteration during intervals of volcanic quiescence. Also associated with this striking style of open-system metamorphism is the formation of ironstone pods, barite deposits, carbonatization, boron mineralisation and K_2O enrichment (de Wit *et al.*, 1982; Lowe and Byerly, 1986; Duchac and Hanor, 1987; de Ronde, 1991; Byerly and Palmer, 1991).

The high-temperature type seafloor metamorphism is inferred mostly from descriptive field geology, the recognition of stratigraphic sequences analogous to those of Phanerozoic ophiolites, petrography and mode of hydration (de Wit and Stern, 1980; de Wit *et al.*, 1987; Cloete, 1986, 1990) and oxygen isotope profiles (Hoffman *et al.*, 1986). Oxygen isotopes suggest that the metavolcanics of the Komati Formation may have been altered and metamorphosed in an oceanic environment but that a deep-seated water source (juvenile water) could also have played a role (Smith *et al.*, 1984).

1.5.1.4 Burial Metamorphism. Despite abundant geochemical- and petrographic evidence in favour of open-system metamorphism, critical evaluation of the metasomatic assemblages in the Komati Formation shows that they are cryptically overprinted by burial or load metamorphism suggesting the gradual onset of closed-system conditions akin to normal isochemical metamorphism. As this aspect of the metamorphic history of the BGB is the subject of the present study it will be elaborated on in the following chapters.

In view of the scarcity of detailed metamorphic studies in the BGB, a brief summary of

the salient metamorphic features of two other major greenstone terranes is warranted for the purpose of comparison and perspective.

1.5.2 Australian Greenstone Belt Metamorphism

Binns *et al.* (1976) give a significant account of Archaean greenstone belt metamorphism in the Yilgarn Block. Two distinct tectono-metamorphic styles are recognized:

- 1) Areas of static metamorphism, metamorphosed to prehnite-pumpellyite, greenschist or lower amphibolite facies with little internal strain. Primary structures and delicate micro-textural aspects are well preserved.
- 2) Areas of dynamic metamorphism, containing greenschist, amphibolite or even lowermost granulite facies, which are highly strained. Penetrative foliations and lineations are pervasively developed.

The good correlation between the Yilgarn Block and the Barberton greenstone belt regarding the above mentioned metamorphic types seems to suggest a commonality of tectono-metamorphic elements. Other metamorphic characteristics of the Yilgarn Block which are generally similar to those of the BGB and warrant brief mention are:

- 1) Gradational relationships between the static- and dynamic-style domains which exist as narrow zones in which the rocks characteristically contain features of both terrains.
- 2) Metamorphic grade variations within the static-style domains that do not correlate well with those of the outcrops surrounding the granitoid intrusions.
- 3) Superimposed contact effects which vary from narrow zones of massive high-grade hornfels around discordant plutons to wider zones of foliated amphibolite and schist encircling forcefully emplaced intrusions to contacts without any significant alteration.
- 4) Metamorphic recrystallization in the static-style and dynamic-style domains occurred more or less at the same time.
- 5) Gravity anomalies indicate that thick sequences of folded supracrustal rock underlie the synformal low-grade static-style areas and that thinner portions of the sequences underlie amphibolite facies areas.
- 6) It is also suggested that only broad relationships between metamorphic grade and stratigraphic level exist as the evidence of pre-recrystallisation deformation would

invalidate such a relationship.

Bickle and Archibald (1984) use meta-pelitic mineral assemblages to constrain a pressure-temperature trajectory for static and dynamic metadomains in the Yilgarn Block. The static-style domain, containing amphibolite facies assemblages, was heated to 530 - 560°C at 4.2 kb. The dynamic-style domain, comprising mid- to upper-amphibolite facies assemblages, was heated isobarically from the previously-mentioned conditions to temperatures of 600 - 650 °C. The higher-grade metamorphic domain is thought to have resulted from heating due to granitic magmatism on crust already thickened and heated to greenschist conditions. The cause, however, of the (ambient) static metamorphism is uncertain and ascribed to crustal thickening either due to overthrusting or magmatic processes or a combination of both.

1.5.3 Canadian Greenstone Belt Metamorphism

The Wawa, Wibigoon, Uchi and Sachigo Subprovinces of the Superior Province, northwestern Ontario, form isoclinally folded, easterly trending, metavolcano-sedimentary sequences, intruded by composite granitoid batholiths (Ayres, 1978). Stratigraphically, these greenstone belts are not complete, as they have eroded tops and unexposed or intruded bases. The greenstone belts, up to 20 km thick, comprise subaqueous basaltic shield volcanoes capped by complex subaerial to subaqueous, andesitic to rhyolitic stratovolcanoes (Ayres, 1978). Two styles of metamorphism are generally recognised: (1) a static-style and (2) a dynamic-style in which respectively, primary textures and structures of the rocks are preserved as opposed to their destruction and replacement by high strain-features (Thurston and Breaks, 1978). The highest metamorphic grades and gradients are achieved within the contact zones of the granitic plutons which are considered to be the main heat source of metamorphism. This metamorphism results in the amphibolitic to hornfelsic aureoles around the granitoids whereas the inner, low grade parts are ascribed to regional burial metamorphism complemented locally by heat from the intrusive granitoids (Ayres, 1978). The relatively low metamorphic grade in the central portions of the greenstone belts are assigned to relatively shallow burial, possibly up to 10 km (Ayres, 1978; Thurston and Breaks, 1978). The geothermal gradient proposed for the Uchi subprovince is inferred to be at ca. 35°C/km (Thurston and Breaks, 1978).

Similarly, the lavas of the Abitibi subprovince are thought to have undergone two types of metamorphism. Soon after extrusion a regional prehnite-pumpellyite grade was imprinted on permeable parts of the lava flows, followed by contact metamorphism accompanying the intrusion of numerous granitic plutons at peak conditions during the Kenoran orogeny (Jolly 1980; Dimroth *et al.* 1983). Beatty and Taylor (1982) infer that the typical style of Abitibi Greenstone Belt metamorphism is one of sub-seafloor burial metamorphism. This type of metamorphism is not related to the intrusion of granitic batholiths and is characterised by good textural preservation (possibly similar to the static-style domains of Binns *et al.* (1976) and Ayres (1978) and the burial metamorphism of Clute (1990). High geothermal gradients are implied by mineral-chemical and phase equilibria data (Jolly, 1980) and a Barrovian-like geotherm (70°C/km) is implied. An oxygen isotope investigation at Pyke Hill (Beatty and Taylor, 1982) indicates that the komatiites reacted with ocean water which penetrated deep into the volcanic sequence as it became heated due to burial. Prehnite-pumpellyite metamorphism of the metavolcanics is considered to be progressive due to burial that occurred up to depths of 12 km (Jolly, 1980). The pervasive- and high-intensity hydrothermal alteration styles of Dimroth *et al.* (1983) appear to be synonymous with sea-floor metamorphism mentioned for the BGB.

Consensus with regard to the origin of the contact aureoles around granitoid-greenstone terrains is unanimous for the Barberton Greenstone Belt (e.g. Hall, 1918; Anhaeusser *et al.*, 1969; Viljoen and Viljoen, 1969), for the Yilgarn Block (Binns *et al.*, 1976; Bickle and Archibald, 1986) and for the Canadian greenstones (Ayres, 1978; Thurston and Breaks, 1978; Dimroth *et al.*, 1983).

The origin of the regional-scale, static style metamorphism of Archaean greenstone belts remote from contact aureoles, however, is less clearly understood. For example, Binns *et al.* (1976) concluded that the lack of direct association between the static- and dynamic- style domains precluded a direct correlation between plutonism and metamorphism, but that relative influences of crustal thickening and localized heat sources deserve further enquiry. Bickle and Archibald (1984) also remain unclear as to its origin and agree that it is associated with crustal thickening, either due to overthrusting or magmatic processes, or both. The Canadian workers recognise a pre-

orogenic (static) style of metamorphism (i.e. the pre-Kenoran type for the Abitibi subprovince) and attribute it, in some cases, to initial seafloor alteration (Beaty and Taylor, 1982; Dimroth *et al.*, 1983) and/or associated very low grade burial metamorphism (Thurston and Breaks, 1978; Jolly, 1980; Dimroth *et al.*, 1983). The upper limit of burial metamorphism is unanimously placed at the top of the prehnite-pumpellyite facies which usually occurs very remote from the thermal and dynamic influence sphere of the granites. The metamorphic aureoles produced around the granitoids, on the other hand, are relatively narrow. The metamorphism of the voluminous supracrustals between the contact aureoles and the very low-grade prehnite-pumpellyite facies rocks remain unexplained.

1.6 Aims and Motivation

The aim of the project is to determine the principle controls on the metamorphic development of the Komati and Hooggenoeg Formations. The main influences are thought to be: (1) the tectonic environment of volcanism, (2) the conditions of seafloor metamorphism, (3) the conditions of burial metamorphism and (4) the conditions of dynamic metamorphism.

Attempts at resolving these problems using modern volcanological and metamorphic concepts has not been carried out for the BGB where it is thought that such an approach should result in new perspectives and possibly a better understanding of its tectonic evolution.

Investigations were primarily concentrated on the Komati and Hooggenoeg Formation type-localities on the western limb of the Onverwacht Anticline due to the excellent state of preservation. Samples revealing dynamic metamorphism were collected along strike to the west of the Komati Formation type locality because of good field control for the onset of dynamic metamorphism. The most dynamically metamorphosed rock was taken on the eastern limb of the Onverwacht Anticline as a control sample.

Sampling was strongly biased to komatiitic basalt pillow lavas for the following reasons:

- 1) To standardise on rock type and morphology. Massive lavas, for example, can be confused with shallow sills which could lead to erroneous conclusions with regard

to timing of emplacement and mode of water/rock interaction.

- 2) To enable comparison between a wide range of water/rock interaction sites (e.g. pillow cores, contraction crack selvages, pillow rims and interpillow cavities). Massive lavas offer a restricted range in water/rock interaction sites whereas the original nature and composition of tuffs tend to be problematical.
- 3) To have a wide as possible range of metamorphic assemblages. Early seafloor metasomatism of pillow lavas has resulted in rock with more varied bulk composition yielding a larger range of (diagnostic) mineral assemblages than those of the massive flows.
- 4) To have greater control on the degree of rock deformation. Compared with massive lava, pillows generally have an array of structural and textural features which allow for easy assessment of both rock preservation and deformation.

1.7 Objectives

The primary objectives of the study are:

- 1) To verify previous mapping of the Komati and Hooggenoeg Formation type-areas and describe the volcanology with the aim of reconstructing the emplacement environment.
- 2) To investigate the petrography and geochemistry of the various pillow alteration zones induced by seawater interaction and to gain an understanding of the processes in terms of modern seafloor-metamorphic concepts.
- 3) To compare the volcanological and hydrothermal metamorphic environments of the Onverwacht rocks to those that have been described from ophiolites and mid-ocean ridges.
- 4) To investigate the geochemistry of mineral assemblages to assess metamorphic zoning in the study area, the attainment of chemical equilibrium of assemblages, the prograde and retrograde metamorphic tendencies of minerals, and to obtain geothermometer temperatures.
- 5) To investigate by means of microthermometry the temporal and spatial development of fluid compositions, their relation to the various metamorphic events and to constrain prevailing pressures during metamorphism.
- 6) To construct a PTt-path for the Komati and Hooggenoeg Formation samples by linking the PT conditions of emplacement, seafloor metamorphism, burial

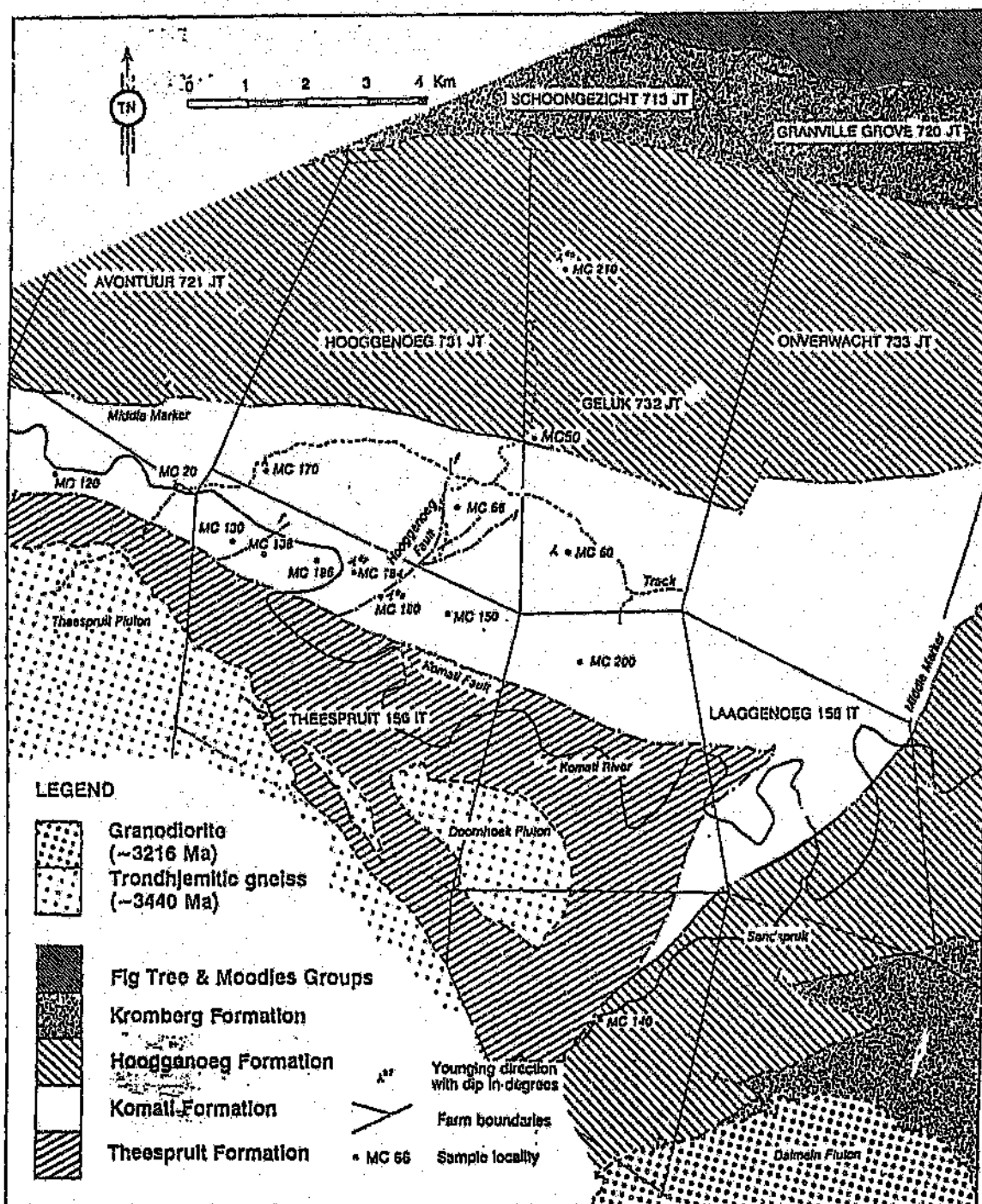


Fig. 1.4 Geological sketch map of the Onverwacht Anticline showing sample localities within the Komati and Hooggenoeg Formations.

metamorphism and by comparing them to the trajectory of those having undergone dynamic metamorphism.

- 7) To construct a geodynamic model for the development of the BGB incorporating the volcanological and metamorphic constraints with the results of other studies.

1.8 Sample preparation and analytical techniques

1.8.1 Sampling

Whole pillows were sampled by means of a portable rock-drill with jack hammer attachment and separating wedges. Sampled pillows weighed between 20 - 50 kg and were sectioned with a dimension-stone diamond saw into 8 to 10 cm thick slabs. This allowed for easy handling, facilitated observation of internal pillow structures, alteration zones and textural domains, and aided petrographic and geochemical sampling by means of diamond sawing. Pillow sample localities are shown in Fig. 1.4.

1.8.2 Petrographic and mineralogical

Petrographic work was mostly performed on thin sections enabling both transmitted and reflected light microscopy. Thin sections were studied with a Leitz Orthoplan petrographic microscope. Mineralogical studies were aided by standard XRD analysis and use of an JEOL JSM-35-Scanning Electron Microscope (SEM) with an EDAX energy dispersive detection system. Mineral compositions were quantitatively determined by means of a JEOL SUPERPROBE 733 with five wavelength-spectrometers. Internationally recognised mineral standards, described by Jarosewich *et al.* (1980) and Jarosewich and Macintire (1983), were used for calibration of the microprobe. The mineral standards and their identification numbers (in parentheses) are: Kakanui hornblende (USNM 143965); Kakanui anorthoclase (USNM 133868); Kakanui pyrope (USNM 133868); Natural Bridge diopside (USNM 117733); Lake Country plagioclase (USNM 115900); Teibaghi Mine chromite (USNM 117075); microcline (USNM 143966); Boyd rhodonite; Boyd rutile (synthetic); siderite (USNM R 2460); calcite (USNM 136321); dolomite (USNM 10057). Corrections were made by ZAF and instrument conditions were: kV = 15; beam current = 20 nA; peak and background counting = 10 sec. each; spot size = $3 - 4 \times 10^{-6}$ mm.

1.8.3 Modal analyses

Modal analyses were performed to establish sulphide distribution within the various

structural and textural domains of the pillows, e.g. amygdales, pillow rims, contraction crack selvages, etc. A method was devised in which the modal percentages of the domains were determined by means of normal point counting but at the same time measuring, with a graduated eyepiece, the surface area of sulphide in each of the domains. The data enables a proportional (percentage) expression of the total sulphide of a polished section belonging to the respective domains. To quantify the proportional values, the sulphur content of the corresponding thin section cut-offs were determined analytically. The cut-off wafers represent the ± 1 mm thick volume of rock adjacent to the surface of the corresponding polished thin section. The assumption was made that the analysed wafers have similar S-concentrations to those of the matching polished thin sections. As this is not always true, due to the inhomogeneous distribution affects of sulphide, the above-mentioned quantification of S for the pillow domains is not fully quantitative. The modal analyses were carried-out on a 'Swift Automatic Point Counter'.

1.8.4 Microthermometry

Microthermometric measurements of fluid inclusions were made using a Chalmers stage as described by Poty *et al.* (1976) and Hollister *et al.* (1981) mounted on a Leitz Orthoplan microscope. The accuracy of measurements was determined to be better than $\pm 0.3^\circ\text{C}$ for temperatures below 100°C , and $\pm 2^\circ\text{C}$ for temperatures up to 400°C . A stabilised power source ($\pm 1\%$ variation) was used and the instrument was calibrated as prescribed in the user's manual using the supplied pure chemicals, and a natural quartz sample containing fluid inclusions with pure H_2O and CO_2 for triple point determinations of the latter at -56.6°C . Distilled water put between coverslip fragments was routinely used for zero point determinations. Additional gas chromatography standards from Carlo Erba, Italy, were also used and showed good agreement with the existing calibration. The gas chromatography standards were: n-tridecane (-5.5°C); carbon tetrachloride (-22.8°C); o-xylene (-25.1°C); m-xylene (-47.86°C); chloroform (-63.5°C); and toluene (-95°C). Minor modifications were made to the microthermometry stage that improved repeatability and ease of operation. They are: (1) insulation of the stage with polystyrene during freezing runs, (2) insulation of the stage with an aluminum-foil shield during heating runs, and (3) the introduction of exhaust nitrogen into the sample chamber to prevent condensation and to increase visibility.

Microthermometric analyses of fluid inclusions were conducted on doubly polished

wafers about 100 μ m thick. All heating and freezing measurements were repeated and accepted only if reproduction within the above mentioned limits were obtained. Freezing measurements were completed prior to the heating measurements. Fluid inclusion maps were used to relocate inclusions so that both heating and freezing data for individual inclusions could be obtained.

1.8.5 Crushing

Rock samples were crushed in a jaw-crusher pre-contaminated with komatiitic basalt. The resulting ≤ 6 mm in diameter chips were then pulverised to a fineness of $\leq 75 \mu$ m with a six-pot swing mill, using carbon-steel vessels. Thereafter the samples were split with a rotary splitter into ± 500 g fractions and contained in plastic bottles.

1.8.6 Major and trace element analyses

Major and trace element analyses were done on fusion discs and powder pellets, respectively. All major elements, with the exception of FeO and the volatiles, were determined by X-ray fluorescence spectroscopy (XRF). The lithium tetraborate fusion technique of Norrish and Hutton (1969) was used to prepare the glass discs. The discs were analysed on a Rigaku Denki sequential X-ray spectrometer which was calibrated using international rock standards. Na and K were analysed along with the trace elements on powder pellets prepared according to the technique of Baird (1961). Trace elements were analysed on a Philips PW1400 sequential spectrometer. The volatile elements, $-H_2O$, $+H_2O$, CO_2 , and S were analysed with Leco-spectrometers utilising infra-red spectroscopy. Ferrous-oxide was determined by a wet chemical method modified after Bennett and Reed (1971). Sample dissolution was done in a microwave oven.

1.8.7 Stable isotope analyses

Stable isotope analyses of 16 sulphide separates were performed on a commercial basis at Kreuger Enterprises Inc., Geochron Laboratories Division, 4 Blackstone Street, Cambridge Massachusetts, USA. The standard employed for these analyses was the Cañon Diablo troilite with $^{34}S/^{32}S = 0.0450045$. The analyses are reported in ‰ and the precision is considered to be ± 0.1 ‰.

2. VOLCANOLOGY OF THE KOMATI AND HOOGENOEG FORMATIONS

2.1 Introduction

The rocks of the Komati and Hooggenoeg Formations have been mapped and described by previous workers (e.g. Viljoen and Viljoen, 1969c,d; Williams and Furnell, 1979; Smith *et al.*, 1980; Viljoen *et al.*, 1983) who placed much emphasis on proof for the extrusive origin of the komatiites as well as their petrogenesis. Very little mention was made, however, of the volcanology and environment of emplacement. De Wit and Stern (1980) and De Wit *et al.* (1987a) subsequently suggested that the pillow lavas of the Komati Formation are either 'intrusive pilloids' or 'remnant screens' resting upon a sheeted-dyke complex. Since then geological models have followed claiming the Onverwacht rocks to have formed at an ancient MOR-spreading centre (de Wit *et al.*, 1987a,b, 1992; de Wit and Tredoux, 1988).

The aim of this chapter is to give an account of the volcanology of the Komati and Hooggenoeg Formations with special emphasis on the pillow lavas and their relationship with the massive units that host them. This is done with the view of maximizing the constraints on the environment of emplacement and to shed light on the above mentioned controversy. Included is a chemical classification, an evaluation of the structural organisation of the pillows relative to their massive counterparts, and descriptions of internal pillow structures, textures and relict igneous mineralogy. The data enable an assessment of the ocean-emplacement depth, lava extrusion rates, palaeoslopes and help to determine if the pillows are in fact intrusive or part of remnant screens.

It will be shown later that the volcanics of the Komati and Hooggenoeg Formations do not resemble typical MOR-type crust as claimed by the above mentioned workers but that they are akin to lavas in the upper parts of oceanic plateaus of modern oceans. Tectonic and stratigraphic implications, however, are discussed in the final chapter subsequent to the presentation of more data.

2.2 Nomenclature

Based on field observations of Viljoen and Viljoen (1969a) and Viljoen *et al.* (1983) it is now generally accepted that the Komati Formation consists of two closely associated major rock types, i.e. komatiite and komatiitic basalt. These terms replaced peridotitic and basaltic komatiite respectively which fell into disfavour due to proposals made at the Penrose Conference and also by Arndt and Nisbet (1982).

In order to define komatiitic basalt, clarification of the term "komatiite" is necessary. Komatiite is defined by two sets of criteria: (1) a volcanic origin (usually deduced from textural and field observations) and (2) ultramafic compositions ($>18\%$ MgO on an anhydrous basis), making it simply an ultramafic lava. Komatiitic basalts are somewhat more difficult to define (Arndt and Nisbet, 1982) as they are essentially magnesian basalts, and only because of their spatial relationship to komatiites, are classified as komatiitic basalts. Together with this special association the above-mentioned authors propose that the definition also contains petrographic and chemical characteristics. The petrographic features include spinifex textures: where olivine type spinifex occurs in lavas with MgO between 18 and ca. 12%, and the pyroxene type occurs in lavas with MgO between 9 and 18%. Basalts with MgO $< 10\%$ have few diagnostic features. Chemically, they have high Mg, Ni, Cr; low alkalis, TiO_2 , Fe/Mg ratios; high SiO_2 at given MgO; and $\text{CaO}/\text{Al}_2\text{O}_3$ ratios $> 0.8 - 1.0$. All these features are characteristic but not necessarily diagnostic.

Selected pillow lavas of the study area have been chemically analysed (Table 3.1) and the relevant data are plotted on a Jensen Diagram (Fig. 2.1) together with published analyses of Onverwacht Group volcanics (Viljoen *et al.* 1982, Fig. 4.6). From the diagram it is evident that the analyses of the rocks selected for this study fall within the field of komatiitic basalt as proposed by Viljoen *et al.*, (1982). It is also evident that the pillows used in the present study are compositionally close to tholeiite. The plot indicates that three pillows selected for this study actually fall within the tholeiite field as defined by the Jensen plot (Fig. 2.1). To obtain a better distinction between komatiitic basalt and tholeiites Viljoen *et al.* (1982) proposed a small boundary shift, as indicated by the arrow in Fig. 2.1. From a genetic point

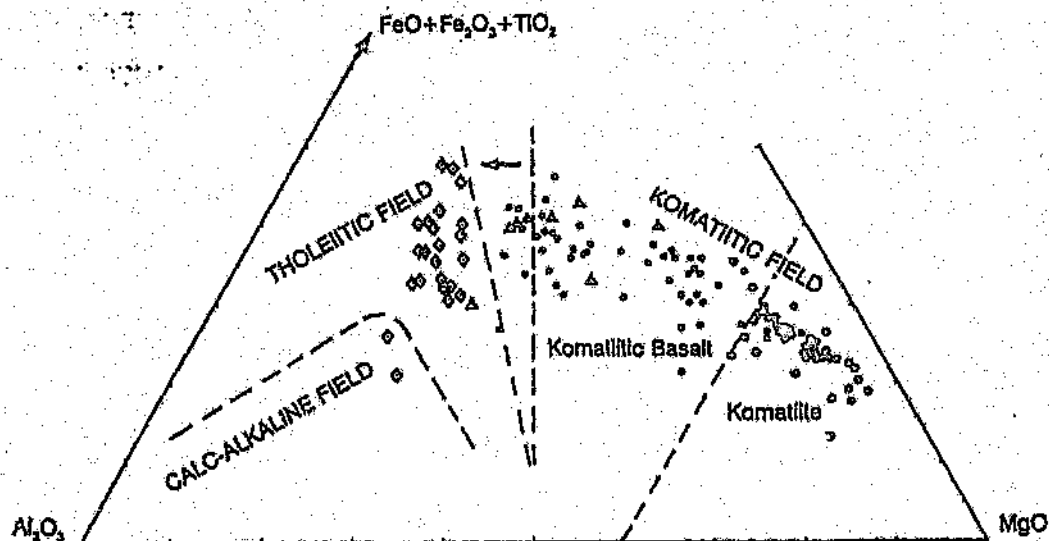


Fig. 2.1 A Jensen plot of the pillow lavas sampled in this study and related volcanics of the Onverwacht Group (after Viljoen *et al.* 1982) with subdivisions of the calc-alkaline, tholeiitic and komatiitic suites (stippled lines). Large dots = komatiite; small dots = komatiitic basalt; triangles = pillows selected for this study; dotted diamonds = tholeiites; arrow indicates shift in the komatiite / tholeiite suite boundary, proposed by Viljoen *et al.* (1982).

of view, the komatiitic and tholeiitic basalts are thought to be derived from the same source material from which the associated komatiites were derived (Smith 1980). The petrogenetic relationships between komatiitic and tholeiitic basalt have been discussed in detail by this author and these characteristics are also compared with similar rocks from other major greenstone belts. It should be stressed, however, that two geochemically distinct suites of komatiite were identified by Smith (1980) who referred to them as Group I and Group II komatiites implying that two different mantle sources were required for the melts. The significance of this observation will be discussed at the end of this chapter.

2.3 Flow Rock Morphology.

Fieldwork for this study has found the general mapping and recording of pillow outcrops by Viljoen and Viljoen (1969c, Fig. 2) to be accurate. This assessment is made for the well-

exposed parts of the Komati Formation which are located on the farms Hooggenoeg and Theespruit (north of the Komati River). Volcanologically significant for this area is the predominance of massive flows over pillowed lava, with an outcrop area ratio of ~97:3 and the virtual absence of bedded terrigenous and volcanoclastic sediments with the exception of a 20 cm thick clastic horizon within the lower half of the formation. This horizon has been referred to by Armstrong *et al.* (1990) as the Komati Unit and may or may not be the same horizon referred to by Viljoen *et al.* (1983) as an 'agglomeratic bed'.

Excluding komatiite flows with MgO $\geq$ 24 wt%, which are not known to form pillows, the remaining komatiitic basalts have a massive to pillow flow ratio of 95:5, emphasizing the tendency of lava to be massive rather than pillowed. Table 2.1 shows the relative area percent distribution of the Komati Formation rock types. Notably absent are any significant volumes of interlayered volcanoclastic sediment, pillow breccias, pyroclastics and tuffs.

Table 2.1: Relative proportions of rock types of the Komati Formation on the farms Hooggenoeg and Theespruit, calculated from the map of Viljoen and Viljoen (1969c, Fig. 2).

Rock type	Area%
Komatiite (massive, MgO $\geq$ 24%)	32.5
Komatiite (poikilitic dykes) *	0.3
Mixed zone (massive)	16.1
Mixed zone (pillowed)	0.2
Komatiitic basalt (massive)	47.7
Komatiitic basalt (pillowed)	2.4
Quartz-feldspar porphyry	0.8
Sediments (hyalotuffs, pillow breccias)	~0

(* includes the dyke subsequently mapped by Williams and Funnell, 1979)

The proportion of pillowed to massive flows for the Hooggenoeg Formation on the western limb of the Onverwacht anticline has been assessed from work of Viljoen and Viljoen, 1969d; Williams and Furnell, 1979; de Wit *et al.*, 1983 and several field traverses. Massive flows greatly exceed those that are pillowed and estimates indicate that $\leq 15\%$ of the area may be underlain by pillows. Williams and Furnell (1979) show that komatiitic basalts occurring in the central part are intimately interleaved with tholeiitic basalts and comprise $\leq 30\%$ of the Hooggenoeg Formation. This value excludes the ≤ 2 km thick Buck Ridge rhyodacitic sill traditionally included as part of the Hooggenoeg Formation.

2.3.1 Massive Flows

Composite massive komatiitic basalt flows containing scattered pillow occurrences attain thicknesses of up to 200 m (Viljoen *et al.*, 1983), whereas the maximum thickness for individual flows is much less, probably not more than 10 m (Smith 1980). These flows are mostly not exposed and soil covered and thickness estimations, therefore, probably represent minimum thicknesses. Thinner flows may be identified by chilled (fine grained) basal margins and incipiently jointed flow tops, marked by fills of hydrothermal quartz or albite, whereas the thicker flows have less discernable contacts due to shed chill effects. Lower contacts may form either against other massive flows, as relatively flat features, or as protruding tongues onto irregular lower pillowed surfaces. In the Komati Formation, no columnar jointing in komatiitic basalt was observed whereas in the Hooggenoeg Formation two horizontally rotated sets (Fig. 2.2A), 50 m upstream of sample MC210 were encountered. The basal portion of the Hooggenoeg Formation is dominated by very thick massive flows and very few pillows. The tops of the massive flows are occasionally buff-coloured, silicified and capped by grey chert. These cappings impart a cyclicity to the Hooggenoeg Formation with cycle thicknesses decreasing upwards (cf Williams and Furnell, 1979). The silicified tholeiites below the chert cappings have lead earlier workers to believe that they are intermediate and felsic volcanics, the tops of the mafic to felsic cycles (Ranson and Byerly, 1983). There is general consensus now, however, that they represent the products of hydrothermal alteration (e.g. Lowe and Byerly, 1986; Duchac and Hannor, 1987).

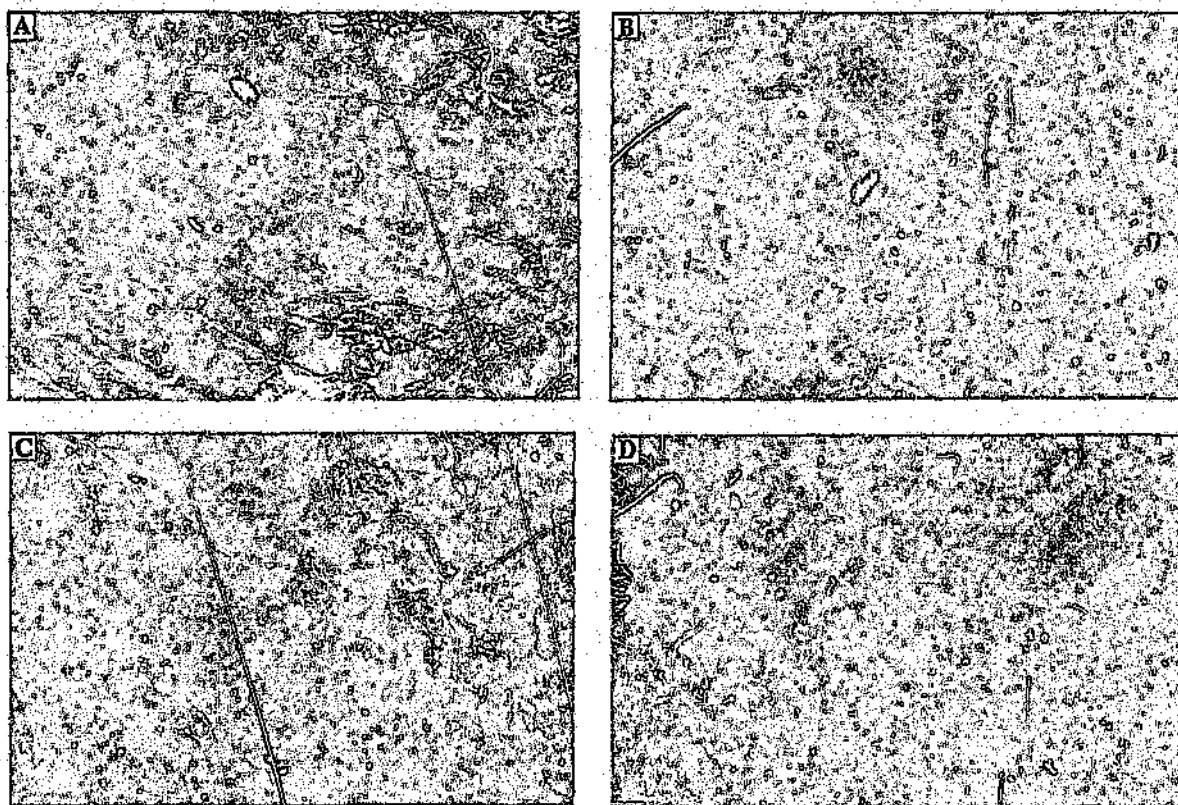


FIGURE 2.2

- A** Horizontally-rotated columnar joints in massive komatiitic basalt of the Hooggenoeg Formation. Columns are ca. 3 meters in length (Locality: $\pm$ 50m north of MC210. Scale: hammer handle = 0.53 m).
- B** Vertically dipping, light-coloured (ocelli-rich) material forming 'inch-scale' rhythmic layering in massive komatiitic basalt. Flame structure suggests younging towards the north. Locality: Spinifex Stream below MC180. Note the superposed crosscutting, subvertical contraction-cracks. Scale: Lens cap diameter = 6 cm.
- C** Komatiitic basalt pillows with 'open stacking'. The interpillow areas are variably filled with metahyaloclastite (dark-coloured material) causing some pillows to be partially suspended (above lens cap). Locality: Stream section to the east of MC170. Scale: Lens cap diameter = 6 cm.
- D** Komatiitic basalt displaying incomplete- or open pillow forms with re-entrant rims (arrows) adjacent to larger pillow with thin pillow crust. Locality: outcrop at MC180. Lens cap diameter = 6 cm.

In hand specimens, the vesicles of massive flows of the Komati Formation are usually small and difficult to see. They are often observed at basal flow contacts (of komatiitic basalts) and become more obvious in thin section. Tony's flow, for example (Smith 1980, Fig. 3.4), contains amygdules 0.1 to 2 mm in diameter which are filled with quartz. In the Hooggenoeg Formation vesicles are generally more abundant and larger and pipe vesicles up to several centimetres long, at basal contacts, are not rare.

Ocelli of massive flows consisting of lighter coloured felsic material 0.5 to 10 cm in diameter often occur close to where pillow lavas are present but they are less common in the absence of pillows (noticed by Viljoen and Viljoen 1969c). Ferguson and Currie (1972), who have studied the petrochemistry and melting phenomena of the ocelli, suggest they owe their origin to an immiscibility loop between the liquidus and solidus allowing for lava emplacement as single liquid. The observed tendency of ocelli to be associated with massive-pillow flow transitions may suggest that cooling of the lavas and decreased flow rates have progressed to optimum conditions for immiscibility to take place.

Occasionally ocelli attain sizes of up to 15 cm as is evidenced in Spinifex Stream below MC180 (Fig. 2.3, and Cloete, 1991). Flow banding in the massive lavas is generally rare but at this locality spectacular bands of lighter-coloured feldspar-rich material occur parallel to bedding, which in places tend to curve out of the flow plane into NE direction (i.e. subperpendicular to bedding). Coincident with curving is the break-up of the bands which result in the formation of large globules and plumes of ocelli, reminiscent of large air bubbles rising in water. De Wit *et al.* (1987, Fig. 14) have interpreted these features as the injection of evolved magma along vertical channels into a pre-existing magma pocket. These intrusive structures are inferred to indicate younging, which is thought to be the same as the present right-way up.

It appears rather that the lighter coloured bands represent flow structures related to liquid immiscibility (*cf* Ferguson and Currie, 1972). The change in flow direction from parallel to near perpendicular to bedding is thought to be due to the buoyant rise of the less dense flow

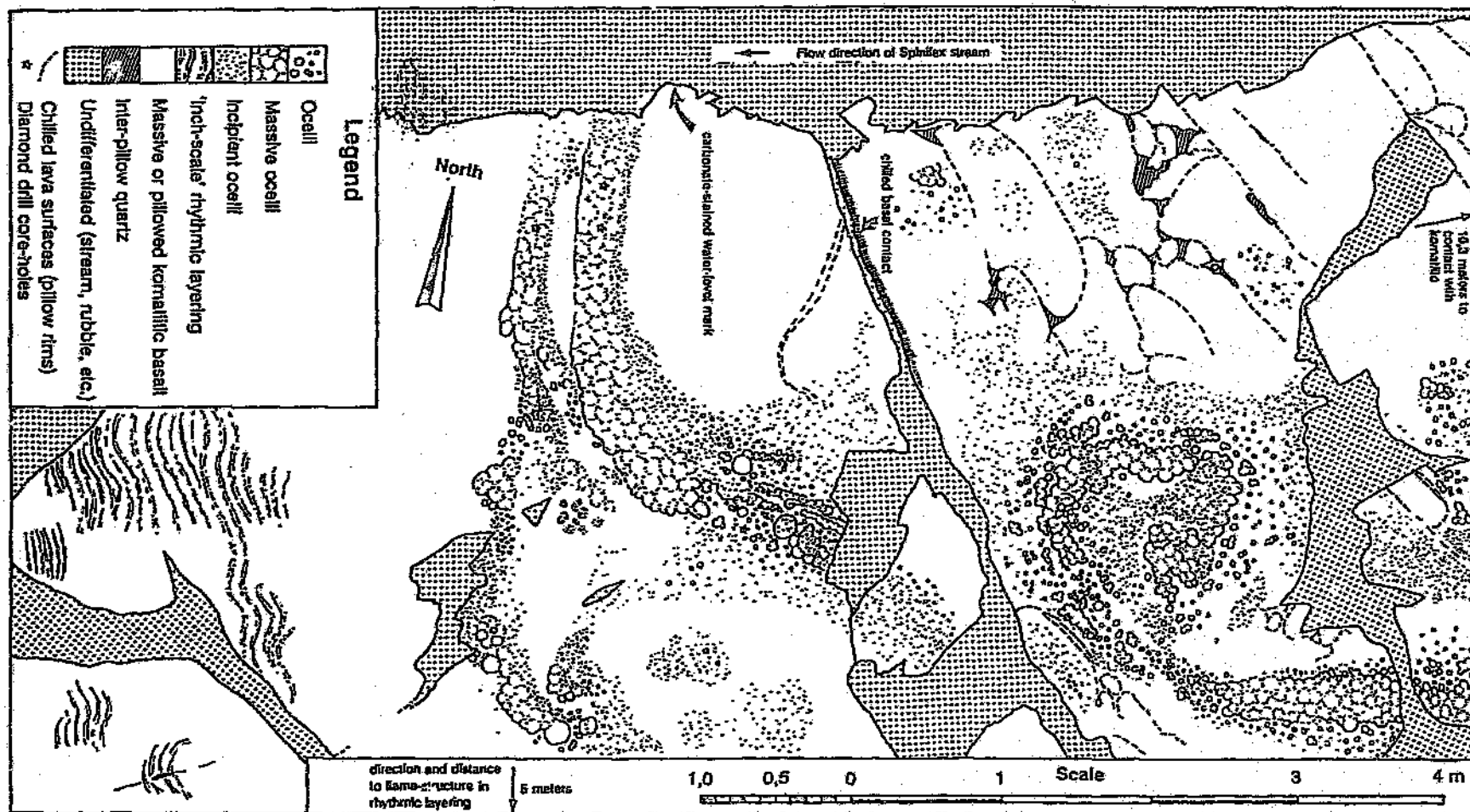


Fig. 23 Detailed sketch map of massive- and pillowed komatiitic basalt containing sinusoidal rhythmic layering, two 'ocelli flow-bands with plumes' and an 'ocelli-spiral'. (The outcrop was mapped at a scale of 1:25. The map plane parallels the surface slope, i.e. $\sim 30^\circ$). Locality: Spinifex Stream below MC180.

bands in a stagnant flow caused by ponding. The break-up of the flow bands and the formation of discrete globules at their upwardly curved terminations is attributed to sustained buoyancy prior to solidification.

At the same outcrop approximately 5 m south of the ocelli plumes, some conspicuous fine inch-scale, rhythmic layering can be found forming metre scale sinusoidal waves. At the centre of the outcrop a sharp inflection in the rhythmic layering is noticed which may be interpreted as an igneous flame structure (Fig. 2.2B). The younging indicated by the structure is to the north. The sinusoidal shapes of the rhythmic layers may be the result of flow 'buckling' as the flow entered the lava pond or due to differential buoyancy and/or convection. De Wit *et al.* (1987) contend that they represent inch-scale, rhythmic layers akin to those of the Skaergaard Intrusion in which the banding is parallel to the floor (or roof) of the intrusion. As the banding is vertically dipping at Spinifex Stream it is difficult to see how the bands should support younging in the present right-way up (as the roof or floor components of dykes are negligible).

Also at this locality (Fig. 2.3), in the next flow up, are some incipiently formed pillows with flat bases and rounded tops overlying a chilled basal contact. Here the pillow-massive flow relationships are characterised by: (1) poorly developed chilled margins; (2) incomplete (open) pillow shapes; and (3) consistent younging directions. The incomplete pillow forms and associated thin selvages reflect relatively rapid flow action which was too large for all the lava to pass through discrete tubes but also too slow to sustain a simple massive flow. The flow rates, however, were high enough to allow only limited chilling of the incipiently formed pillow surfaces. Such relationships (Dimroth *et al.*, 1978) between the pillows and massive flows serve to demonstrate that they are: (1) volcanogenic and not tectonic, and (2) that the surrounding massive units cannot be considered to be dykes or shallow sills but only flows.

Just above the same basal contact a conspicuous ocelli-spiral occurs (Fig. 2.3). From the map it is noticeable that the ocelli-spiral is restricted by the basal contact and pillows on

either side as well as from above, suggesting its presence within a throttle zone or lava channel. The spiral may therefore represent an immiscible flow-sheet caught up in a vortex motion on passing through the flow restriction.

2.3.2 Feeder Dykes

Only two relatively small cross-cutting dykes have been identified for the Komati Formation (Viljoen and Viljoen, 1969d; Williams and Furnell, 1979) which can be described as possible feeder dykes for some of the lava flows. Their limited vertical extent (≤ 200 m) suggests that they were of minor importance considering the large volumes of volcanic rock that make up the Komati and Hooggenoeg Formations. Primary fissures seem not to crop out on the western limb of the Onverwacht Anticline. This may suggest that if large fissures existed they were spaced at relatively large intervals (relative to the size of the study area) or they may be orientated subparallel to the present erosion surface, thus escaping detection.

2.3.3 Pillowed Flows

Pillow outcrops of the Komati Formation consist of stream sections, sporadic outcrops in the veld and rugged showings on the steep-sided hill slopes producing some good individual displays but of limited vertical and lateral extent. The stream platforms are usually two dimensional and not large whereas on the sides of slopes they can be three dimensional. They always occur as clusters rarely larger than a few tens of meters in any dimension. Transitions between massive and pillowed flows are seldom exposed. This is because massive komatiitic basalt is usually soil covered, in contrast to pillows that generally crop out ruggedly.

Similarly, pillows in the Hooggenoeg Formation occur as isolated clusters, often with incomplete selvages marking the transitions to the massive flows. Pillow clusters tend to increase in size towards the top of the formation - at the bottom rarely exceeding a few tens of metres in any dimension whereas at the top several tens of metres in any dimension being more common. Pillows always grade upwards into massive flows and not the other way round.

Pillows vary greatly in size (0.01 - 2 m) and have irregular to rounded shapes. According to the varieties of pillow lava in Table 2.2 (modified after Dimroth *et al.* 1978, Table 1), the types that are frequently observed in the Komati Formation are "normal-size pillows of equal size" (Type A*) and "normal-size pillows with few megapillows" (Type B). "Megapillows with some normal pillows" (Type C) and "megapillows without quenched crust" (Type D) have not been found. A number of large pillows without quenched crust do occur but are

Table 2.2: Varieties of pillow lava for the study area with pillow dimensions modified from Dimroth *et al.* (1978, Table 1).

Type	Description
Normal-size pillows of equal size = Type A (*)	Bun and mattress shaped pillows, 30-100 cm long, are moulded upon each other. Very small pillows (<15 cm) are present, particularly at triple junctions of normal size pillows.
Normal-size pillows and few megapillows = Type B (*)	Few very large pillows (>150 cm) are set in the normal-size pillows, particularly in the flow. Megapillows have a large irregular outline, are surrounded by many incompletely formed pillows and contain many re-entrants of pillow lava.
Megapillows and some normal-size pillows = Type C (#)	Pillows several metres long are mattress-shaped in cross section. A few normal-sized pillows are present, particularly at triple junctions between megapillows.
Megapillows without quenched crust = Type D (#)	Pillows several meters long are mattress-shaped in cross section. They lack a quenched glass crust and their presence is indicated by concentric cooling cracks.
Welded megapillows = Type E (*)	The flow appears to be massive. Traces of concentric cooling cracks here and there betray the presence of pillows.

* = Pillow types observed in both the Komati and Hoeggenoe Formations.

= Pillows observed only in the Hoeggenoe Formation (their absence in the Komati Formation may be due to the poor exposure that generally occurs at the transitions from massive to pillowed flows).

grouped with the Type B category. The chances, however, of Type C and D pillows existing is a distinct possibility, seen in the light that they are intermediate to the two end-member varieties, i.e. Type A and Type E (welded megapillows), both of which occur. Their non-observance could be related to the generally poor exposure of massive- and pillowed flow contacts. These relationships for the tholeiites of the Hooggenoeg Formation, however, are readily observed (e.g. in the stream section just below MC210).

Pillow morphology. Pillow shapes in the Komati Formation seem to be related to pillow size. Small pillows, ≤ 10 cm in diameter are generally sub-circular and pillows 10 to 30 cm long are ellipsoidal and generally have top-bottom symmetry. Larger pillows tend to progress from bun- to mattress shapes and appear draped over underlying irregularities. These features are similar to those noted by Dimroth *et al.* (1978).

At Type A pillow outcrops (localities MC120, MC170 and MC184), the chilled crusts of pillows are well developed and are commonly closed, forming discrete sacks that are moulded more or less tight against each other. These pillows usually have triple junctions that are filled with dark green to dark grey metahyaloclastite or coarsely crystalline milky quartz (if not occupied by very small pillows). Some of the larger pillows may contain thinner selvages and appear more massive than their smaller counterparts. Their stacking is relatively tight and occasionally partially collapsed pillows are seen (river section at MC170). Totally collapsed pillows were not observed. The upper surfaces of these pillows are more rounded than the lower surfaces and downward pointing protrusions are not uncommon. Openly-stacked pillows on the other hand seem to have less downward-facing protrusions, more symmetrical shapes and more metahyaloclastite in the interstices. These pillows probably form a result of a slower lava supply rate, which allows for more cooling to occur before the individual pillows are separated from the flow front and stacked. The degree of open-stacking is limited to 'partial pillow-supported stacking' and never entirely matrix (hyaloclastite or sediment) supported (Fig. 2.2C).

At Type B pillow outcrops (ie. MC184 and MC180), pillows can often be seen to be

incomplete or 'open' and lined by relatively thin selvages and occasionally the chilled crusts are absent (Fig. 2.3). Re-entrant pillow crusts, open pillows (Fig. 2.2D) and ghost selvages within pillows (Fig. 2.4A) are fairly common and imply lava extrusion through pillow selvages to form the adjacent pillows, a process known as budding (Wells *et al.* 1979). The pillows are more frequently bun- than mattress-shaped but are nevertheless tightly moulded against each other. Pillows that commonly contain multiple drainage cavities in the upper halves usually have thin or no chilled selvages at all, i.e. only finely crystallised rims (Fig. 2.4B). These pillows also have fewer or less accentuated cooling cracks than the category described above. The pillows appear to be interconnected and networks of tubes may be expected (hill slope at MC180).

Massive flows with the outlines of welded pillows are rare. Only to the west of locality MC184 were the possible outlines of welded pillows with weakly developed contraction joints observed. The scarcity of these occurrences may be due to the tendency of massive komatiitic basalt to be soil covered.

2.3.4 Pillow Breccia

According to Dimroth *et al.* (1978), pillow breccias consist of complete pillows and/or pillow fragments set in a matrix of hyaloclastite. Three varieties of pillow breccia are known and are distinguished by the petrography of the hyaloclastite. They are: (1) isolated-pillow breccia, (2) broken-pillow breccia, and (3) pillow-fragment breccia. For the Komati Formation this entire class of extrusive rock was not found. Transition of normal-sized pillows into pillow breccias, i.e. occurrences of individual pillows totally suspended in metahyaloclastite was not observed. Only partially suspended pillows of localised extent occur here and there (MC170), not warranting status of pillow breccia and can be referred to as 'locally open-stacked pillows'.

2.3.5 Hyalotuffs

Stratified tuffs consisting of hyaloclastic shards and pillow fragments directly related to flow rocks are known as hyalotuffs (Dimroth *et al.* 1978). They frequently have graded bedding.

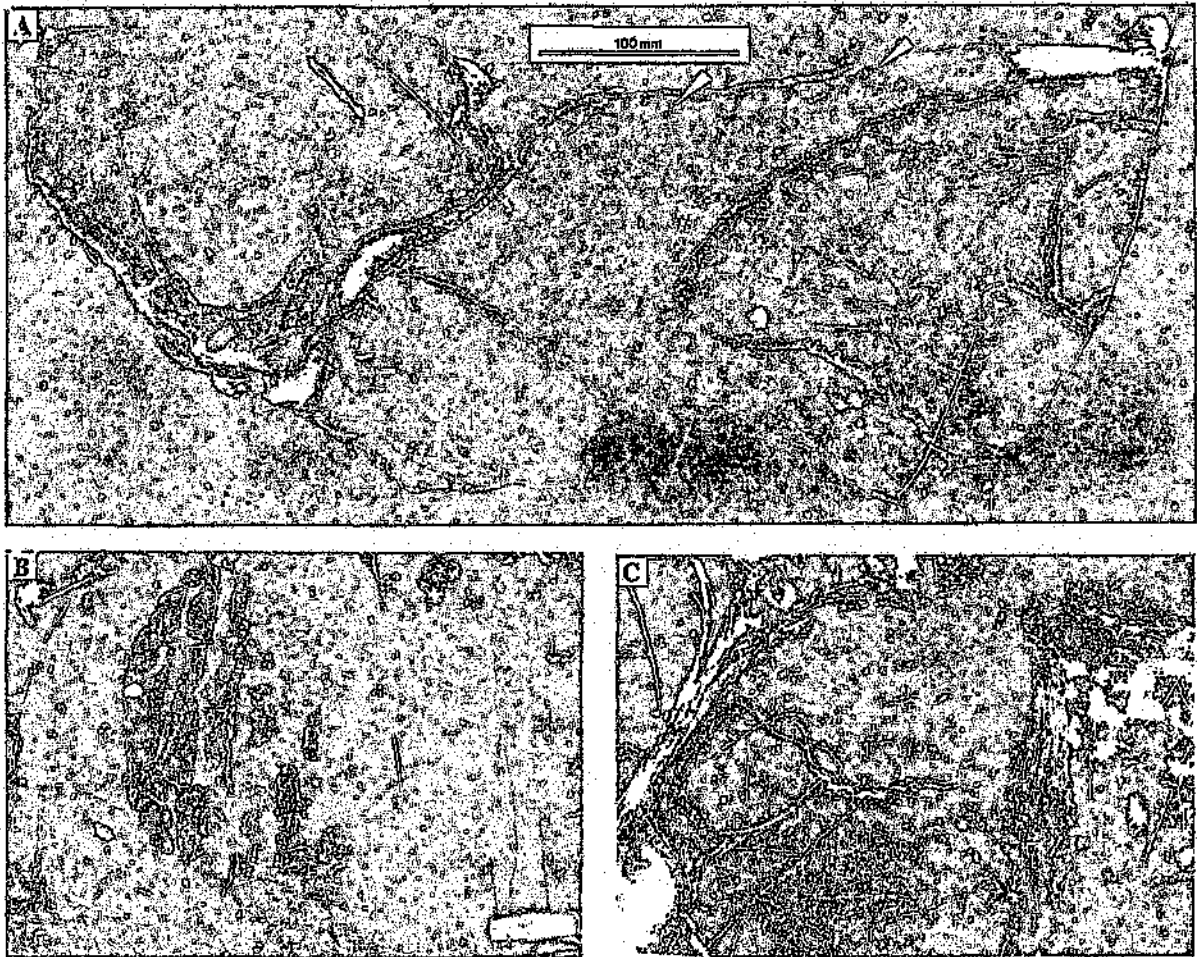


FIGURE 2.4

- A A sectioned slab of komatiitic basalt showing a ghost selvage high-lighting a pillow within a pillow (right) and concentric contraction joints isolating an ocelli-rich inner part of the pillow on the left. The arrow tips, in the upper part of the pillow, point to some rare large amygdales trapped just below the light-coloured (albitised) pillow rim. Locality: MC170.
- B Vertically rotated pillow of komatiitic basalt with multiple drain-away cavities (the pillows' lower protuberance is just left of the middle of the hammer handle). Note the absence of a chilled selvage, especially evident left of the hammer head. The rim has instead a fine crystalline texture. Note the smooth upper-surfaces of the septa, on the side of the coin, and the pendants on their undersides. A dribble mound is visible on the centremost septum on the same height as the coin. Locality: 100m west of MC184. Scale: Length of handle = 53 cm.
- C A sectioned slab of pillowed komatiitic basalt showing within-pillow flow bands both parallel and perpendicular to the rim. The pillow has a well developed selvage which is entirely silicified. The interpillow cavity is predominantly filled with quartz and lesser amounts of albite, clinozoisite and sulphides. The white marks in the core of the pillow represent either microphenocrysts or amygdales. Locality MC180. Scale: diameter of pillow = 16 cm.

Pyroclastic rocks of this nature have not been found or previously reported for the Komati Formation. In the Hooggenoeg Formation few such horizons exist and are thin (≤ 5 m) and usually silicified. The Middle Marker at the base of the formation may be such an example, but the air-fall component of the horizon and possibly its original felsic content (Lanier and Lowe, 1982) disqualify it from having a true hyalotuff affiliation. A poorly stratified horizon about 10 m thick which has not been silicified occurs approximately 200 m below MC210, just below a huge outcrop of buff-coloured silicified pillows.

2.4 Internal Pillow Structures

2.4.1 Drain-away Cavities

Quartz filled drain-away cavities usually occur in larger pillows (≥ 80 cm) and are often present stacked one above the other. They have previously been described by Viljoen and Viljoen (1969c) as 'horizontal gas cavities' and by Dimroth *et al.* (1978) as large 'tubular gas chambers' contained in the upper halves of pillows. This interpretation does not satisfactorily explain the small scale structures that are associated with them nor their flat horizontal shapes and their stacked mode of occurrence.

In the study area the multiple, large quartz-filled cavities contained in the upper halves of individual pillows are thought to represent successive decreasing levels of lava stand, suggesting repeated stages of limited draining (Wells *et al.* 1979). No hollow or hollow-layered pillows similar to those described by Wells *et al.* (1979) were observed in the study area, implying that complete draining of pillows did not occur. When several sets of 'stacked' cavities are present in an outcrop, the stacks tend to have roughly the same number of repetitions in pillows of roughly the same 'stratigraphic height' (at outcrop scale). This suggests that the successive lowering of lava level may have been hydrostatically controlled by an interconnecting plumbing system, suggesting in turn that many of the pillows are lava tubes. The individual cavities may therefore be related to instances of pillow budding at the flow front causing a drop in the lava stand of the interconnected lava tubes. Good examples are exposed at MC180 and at MC184 and moderately exposed tubular pillows with drainage cavities are present to the NE of locality MC130.

The poorly-developed chilled selvages or the lack of them in pillows containing drainage cavities (Fig. 2.4B) is ascribed to slower cooling rates related to the heat supply of sustained lava flow through the tubes. The drainaway cavities of the study area, however, seem to be more continuous and have greater horizontal dimensions (≤ 80 cm) than those illustrated for some Canadian Archaean basalts (Wells *et al.* 1979). In other instances the septa of the drainage cavities appear disrupted whilst the outer pillow shells remain intact, suggesting some form of pulsing of either gas or lava (Fig. 2.4B). The upper surfaces of the septa are always flat and smooth whereas their lower surfaces are irregular with stalactitic pendants (Wells *et al.* (1979). The pendants may be associated with regular-shaped dribble-mounds on the smooth surfaces below them (Fig. 2.4B, for the original examples see Wells *et al.*, 1979). The pendants and associated dribble-mounds may grow towards one another and join-up to form pillars between the septa reducing the original lateral dimensions of the cavities.

Flow banding within pillows is not common. Within pillow MC180 thin flow bands consist of alternating plagioclase and amphibole rich zones which could be the result of flow-deformed ocelli or intersertal glass. They occur both parallel and perpendicular to the chilled margins (Fig. 2.4C). Flow bands truncated by pillow rims are inferred to indicate points of initial pillow growth or sites at which pillows were nipped-off from the main feeder tubes.

On a microscopic scale flow textures are seen in the form of stringers of radiating acicular amphibole set in a fine-grained groundmass. Coarser-grained amphibole pseudomorphs after acicular pyroxene occasionally retain preferred orientations or flow fabrics in the form of fold patterns within undeformed pillows.

Radial and concentric cooling cracks are a common feature of the studied pillows and are usually filled with white hydrothermal material representing mixtures of quartz and feldspar $\pm$ epidote and $\pm$ sulphides. Radial and concentric joints are co-dominant and seem to have been mutually responsible for stress relief (Fig. 2.4A). Alteration of pillow interiors along the cooling cracks (to be discussed in Chapter 3) is visible to the naked eye as either light

TABLE 2.3 Modal analyses of the relative abundances of the various pillow and textural domains and modal sulphide within these domains. (See legend at the bottom of the table, i.e. the last page, for explanation of abbreviations, symbols and formula for calculation of S (ppm) in domain.

Sample No.	Total S (ppm)	Analysis	PTC	RC	VFCC	AHYG	RTM	RH	HFIPA	Counts
MC150B	251	area % of domain	80	15	1	4				834
		modal sulphide (area %)	26	4	11	60				1115
		ppm S in domain	81	59	3012	3417				
MC150C	98	area % of domain	72	21	5	3				864
		modal sulphide (area %)	38	3	4	55				401
		ppm S in domain	52	14	86	1996				
MC150D	125	area % of domain	47	40	11	2				743
		modal sulphide (area %)	25	11	8	57				363
		ppm S in domain	66	33	94	3082				
MC150E	181	area % of domain	70	20	7	3				953
		modal sulphide (area %)	26	4	6	65				829
		ppm S in domain	67	33	141	3783				
MC150G	110	area % of domain	47	41	10	2				772
		modal sulphide (area %)	37	5	7	51				612
		ppm S in domain	87	14	78	2530				
MC150L	132	area % of domain	10	15	2	1	37	15	20	435
		modal sulphide (area %)	0	0	0	1	1	23	75	485
		ppm S in domain	3	0	17	88	4	200	497	
MC150M	261	area % of domain	6	19	2	1	49	8	16	451
		modal sulphide (area %)	0		6	0	44	10	39	405
		ppm S in domain	0		674	0	239	351	633	
MC150P	154	area % of domain	23	15	1	1	16	9	35	379
		modal sulphide (area %)	3	0	0	0	1	2	93	540
		ppm S in domain	17	2	28	56	12	40	413	
MC15JR	216	area % of domain	2	17	1	1	58	21	0	481
		modal sulphide (area %)	0	0	2	9	9	80	0	887
		ppm S in domain	0	3	366	1748	33	838	0	
MC150R	5036	area % of domain						100		
		modal sulphide (area %)						100		
		ppm S in domain						5036		
AVE MC150		area % of domain	39.5	22.4	4.4	2.0	17.7	5.8	7.8	
		modal sulphide (%)	17.1	2.9	5.0	33.0	6.2	12.9	23.0	
		ppm S in domain	63	19	199	2439	87	3446	485	
MC180A	95	area % of domain	48	46	3	3				500
		modal sulphide (area %)	31	15	2	53				1040
		ppm S in domain	61	31	64	1511				
MC180B	195	area % of domain	69	24	4	2				490
		modal sulphide (area %)	47	9	7	38				434
		ppm S in domain	132	71	319	3196				
MC180U	70	area % of domain	69	27	2	2				573
		modal sulphide (area %)	22	5	3	69				31
		ppm S in domain	23	14	118	2195				
MC180R	79	area % of domain	56	41	1	2				447
		modal sulphide (area %)	35	7	0	57				336
		ppm S in domain	50	14	0	2159				
MC180S	81	area % of domain	61	36	1	2				576
		modal sulphide (area %)	41	10	0	49				75
		ppm S in domain	54	23	0	2089				

Sample No.	Total S (ppm)	Analysis	PTC	RC	VFCC	ANYG	RIM	RH	HFIPA	Counts
MC180T	50	area % of domain modal sulphide (area %) ppm S in domain	58 95 81	33 2 3	7 2 16	1 1 36				564 166
MC180C	429	area % of domain modal sulphide (area %) ppm S in domain	11 0 4	23 0 2	7 2 137	2 0 24	52 86 714	3 4 546	4 8 906	332 8758
MC180H	558	area % of domain modal sulphide (area %) ppm S in domain	13 1 60	15 0 11	9 5 321	2 1 453	54 82 847	0 0 418	8 10 727	492 3748
MC180K	102	area % of domain modal sulphide (area %) ppm S in domain	2 1 24	8 1 6	10 13 104	0 22 11424	3 49 205	23 11 49	33 3 9	420 201
MC180G(*)	527	area % of domain modal sulphide (area %) ppm S in domain		15 6 207	10 16 1336		28 57 1085	5 1 77	43 11 139	437 2773
AVE MC180		area % of domain modal sulphide (%) ppm S in domain	38.7 27.3 65	26.9 5.5 20	5.3 8.0 434	1.7 29.0 1689	15.7 27.3 743	3.1 1.6 103	8.7 3.2 178	
MC60A	172	area % of domain modal sulphide (area %) ppm S in domain	23 40 299	71 36 87	2 4 326	4 20 860				476 374
MC60C	259	area % of domain modal sulphide (area %) ppm S in domain	34 18 137	57 19 86	4 30 1856	5 33 1754				497 607
MC60D	378	area % of domain modal sulphide (area %) ppm S in domain	17 7 155	65 16 93	1 7 3355	6 6 353	7 61 3218		4 3 291	471 473
MC60E	241	area % of domain modal sulphide (area %) ppm S in domain	19 2 28	57 5 21	1 0 161	3 1 98	14 45 788		8 47 1504	480 697
MC60F	429	area % of domain modal sulphide (area %) ppm S in domain	6 1 71	9 1 48	1 0 92	1 2 1216	34 93 1159		49 4 33	550 1901
MC60G(*)	2465	area % of domain modal sulphide (area %) ppm S in domain	14 1 176	39 2 128		3 1 758	14 41 7324	30 55 4533		
AVE MC60		area % of domain modal sulphide (%) ppm S in domain	18.8 18.3 158	49.8 6.3 78	1.5 6.9 1295	3.5 10.3 822	11.5 40.0 2542	1.0 9.1 4533	10.0 8.9 232	
MC120A	147	area % of domain modal sulphide (area %) ppm S in domain		95 98 152	1 0 0	5 2 74				436 844
MC120B	97	area % of domain modal sulphide (area %) ppm S in domain	1 0 0	93 93 99	4 7 190	2 0 13				428 614
MC120C	209	area % of domain modal sulphide (area %) ppm S in domain	1 0 0	86 10 23	0 0 0	4 2 82	4 62 3171	1 23 7942	4 4 180	491 3731
MC120D	186	area % of domain modal sulphide (area %) ppm S in domain		71 10 25	1 0 13	5 64 2187	6 18 552	0 0 0	16 9 103	483 14183

Sample No.	Total S (ppm)	Analysis	PTC	RC	VFCC	AMYG	RIM	RH	HFIPA	Counts
MC120F	431	area % of domain modal sulphide (area %) ppm S in domain		44 1 7		1 0 72	17 84 2086	3 2 244	35 13 164	431 9451
MC120H	358	area % of domain modal sulphide (area %) ppm S in domain					48 648	11 319	41 25	391 3 5019
AVE MC120		area % of domain modal sulphide (area %) ppm S in domain	0.3 0.0 0	64.8 35.1 69	0.9 1.1 128	2.9 11.3 684	12.6 42.0 1099	2.5 5.7 810	15.9 4.7 94	
MC128A	104	area % of domain modal sulphide (area %) ppm S in domain		97 23 24		3 77 3092				500 803
MC128B(*)	1713	area % of domain modal sulphide (area %) ppm S in domain		95 94 1709	2 0 101	4 6 2546				463 7710
MC123F	178	area % of domain modal sulphide (area %) ppm S in domain	11 5 71	44 11 46	1 0 41	3 65 3744	40 19 82			451 914
MC128E(*)	4215	area % of domain modal sulphide (area %) ppm S in domain	9 1 527	37 10 1167		2 2 4004	52 87 7028			409 11866
MC128C(*)	2193	area % of domain modal sulphide (area %) ppm S in domain	9 7 1730	86 76 1952	1 12 29971	5 4 2144				430 4543
AVE MC128		area % of domain modal sulphide (area %) ppm S in domain	5.8 2.5 465	71.8 43.0 979	0.8 2.5 6022	3.2 30.9 3106	18.4 21.1 1422	0.0 0.0 0	0.0 0.0 0	
MC20E	148	area % of domain modal sulphide (area %) ppm S in domain		86 99 171	14 1 6					
MC20F	202	area % of domain modal sulphide (area %) ppm S in domain		67 22 67	7 3 0		16 60 776	10 15 286		486 344
MC20H	146	area % of domain modal sulphide (area %) ppm S in domain		71 35 71			17 51 448	13 14 165		486 431
MC20B	381	area % of domain modal sulphide (area %) ppm S in domain		22 21 374			14 47 1321		65 32 187	444 349
AVE MC20		area % of domain modal sulphide (area %) ppm S in domain		61.4 44.4 130	5.3 1.0 4	0.0 0.0 0	11.4 39.5 819	5.7 7.2 217	16.2 8.0 187	

Total S (ppm) = S content of wafer sampled from the opposite surface to which the polished thin section was made (assuming that both contain comparable amounts); PTC = primary textured core; RC = recrystallized core; VFCC = in-filled cooling cracks; AMYG = amygdules; RIM = recrystallized pillow rims; RH = recrystallized hyaloclastite; HFIPA = hydrothermally filled interpillow areas; Counts = number of units counted; S (ppm) in domain = Total S in slide x [modal S in domain (area %) / area% of domain in slide]; (*) = sample is thought to be enriched or mineralized from an external source. NB - modal analyses have been rounded off to the unit although S calculations are based on original values which included the first decimal.

lime-green (silicification), whitish (albitic) and dark green (amphibolitic) selvages to the pillow wall-rock (Fig. 2.4A and 2.6D). Shallow concentric cooling cracks occasionally cause segments of pillow rim to 'exfoliate', exposing outer core to the interpillow cavities.

2.4.2 Pillow Textures

The relative proportions of the structural and textural features of the pillows have been determined by modal analyses (Table 2.3). Many of the early features related to the emplacement of the pillows are revealed, for example, the degree of vesiculation, the extent to which cooling-crack formation developed and also the sulphur distribution between the various structural units

Pillow rims have well-preserved early textures. The earliest are displays of convex and concave thermal tension cracks that have resulted from strain during quenching of pillow margins (Fig. 2.5A). Pseudomorphs of phenocrystic clinopyroxene, olivine and vesicles are not uncommonly present. Original grains of euhedral and skeletal chromite occur scattered throughout the pillow rims (Fig. 2.7C). The pillow rims have been subjected to distinct styles of alteration, the petrography and geochemistry of which is discussed in Chapter 3.

The outer core is characterised by the first appearance (from rim to core) of isolated microlites which rapidly pass into densely spaced microlites. The microlites consist of fibrous needles of pyroxene (now pseudomorphed by amphibole) and have variable shapes, commonly being spherulitic (radiating from a central point) or S-shaped, classifying them as axiolites (Fig. 2.5B). The individual fibres of the microlites appear to be accentuated by finely disseminated titanite which may have formed during initial hydration. The zones of densely packed microlite, juxtaposed with the pillow rims, may attain thicknesses of up to 2 cm and are usually recognised under the microscope by their collective granular habit exhibiting 'propeller-like' extinction crosses (under low magnification). Broadly similar spherulitic textures adjacent to metabasalt pillow rims, have been reported by Gélinais and Brooks (1972, Plate IVA). When situated in ocelli the fibres characteristically have a skeletal outline set in a devitrified plagioclasic mesostasis. This is not visible with a normal

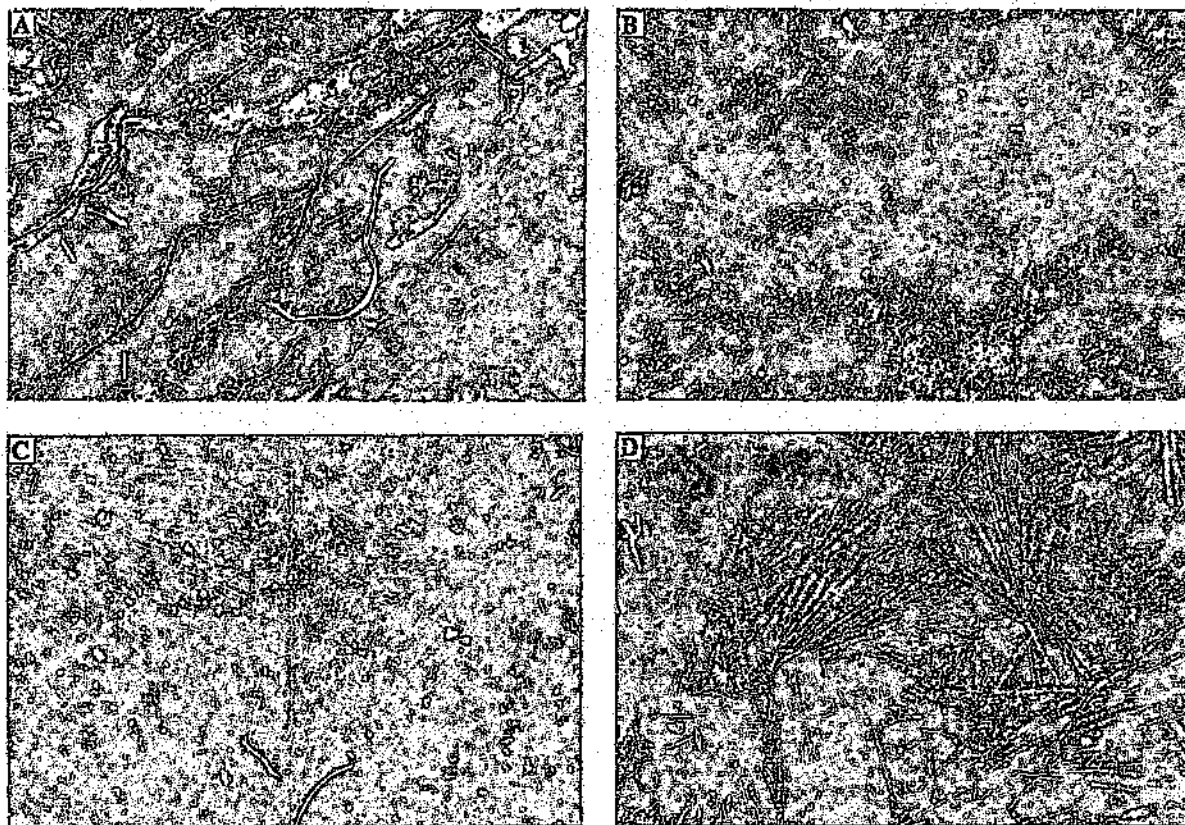


FIGURE 2.5

- A** Photomicrograph of an albitised pillow rim showing thermal tension cracks. The pillow margin surface is indicated by the dark bands (lower right) whereas the core is toward the upper left. The thin dark axial lines of both the cloudy and vermiciform domains consist of stringers of very fine grained titanite which may have been precipitated along early hairline tension fractures. Sample: MC60m. Light: Plane polarized. Length of field of view = 6 mm.
- B** Photomicrograph of spherulitic and axiolitic microlite pseudomorphed by amphibole after augite in the outer-core of pillowed komatiitic basalt. Sample: MC180m. Light: Plane polarized. Length of field of view = 300 micron.
- C** A SEM photomicrograph of skeletal amphibole (light) after pyroxene set in a plagioclasic mesostasis (dark). Sample: MC150. Length of field of view = 200 micron.
- D** Photomicrograph illustrating randomly-orientated and fan-shaped arrays of acicular amphibole after augite. The texture is referred to as 'microspinel' in the text. The darker geometric patches are ocelli and consist of acicular amphibole and 'inclusion-rich plagioclasic mesostasis'. The ocelli can also be described as intersertal pools of devitrified plagioclasic glass which partially include the randomly-orientated and fan-shaped arrays of acicular amphibole. Sample: MC150c. Light: Plane polarized. Length of field of view = 0.7 mm.

petrographic microscope and is resolved only with the aid of an SEM (Fig. 2.5C)

The pillow core/outer core boundary is a gradual transition from fibrous microlitic aggregates to interlocking, randomly orientated and fan-shaped arrays of amphibole after acicular augite. This texture will be referred to as microspinfex (Fig. 2.5D). The individual needles, which range up to 2 mm in length and 0.05 mm in width, are skeletal in nature and tend to have hollow, straw-shaped habits. These features are especially evident with SEM-imagery and in reflected light illumination utilizing polished surfaces. In the ocelli-rich phases the skeletal character is more accentuated (Fig. 2.6A).

In pillow MC60 a microspinfex-textured outer core grades into a micro-orthocumultic central core still consisting of augite suggesting that it was not chilled to the same degree as the cores of other pillows. The textures suggest slower cooling rates or higher heat flow. This may be explained by the central portion of the pillow serving as a lava conduit, resulting in an micro-orthocumulate rather than a skeletal-texture dominated core. The micro-orthocumulate textures in ocelli- and non-ocelli-bearing domains differ by containing more or less devitrified intersertal/interstitial glass. The ocelli and non-ocelli patches tend to have more subhedral and anhedral augite respectively, the latter case tends to give the rock a micro-mesocumulate appearance (Figs. 2.6B and 2.6C).

Ocelli in pillowed komatiitic basalt are light coloured, felsic looking, sub-spherical to angular bodies, 0.1 - 4 cm in diameter, occur throughout pillows but tend to be concentrated in the cores. They have previously been described as variolites or spherulites with distinct zonal arrangements within pillows and have the ability to merge (Viljoen and Viljoen 1969c, Plate IVa + IVb).

Due to the absence of a radiating aspect to these orbicular patches, the non-genetic descriptive term ocelli has been adopted, as proposed by Ferguson and Currie (1972). The ocelli seem to consist of globules of intersertal devitrified glass ingressed by randomly-orientated, acicular, or fan-shaped arrays of interlocked amphibole pseudomorphs after

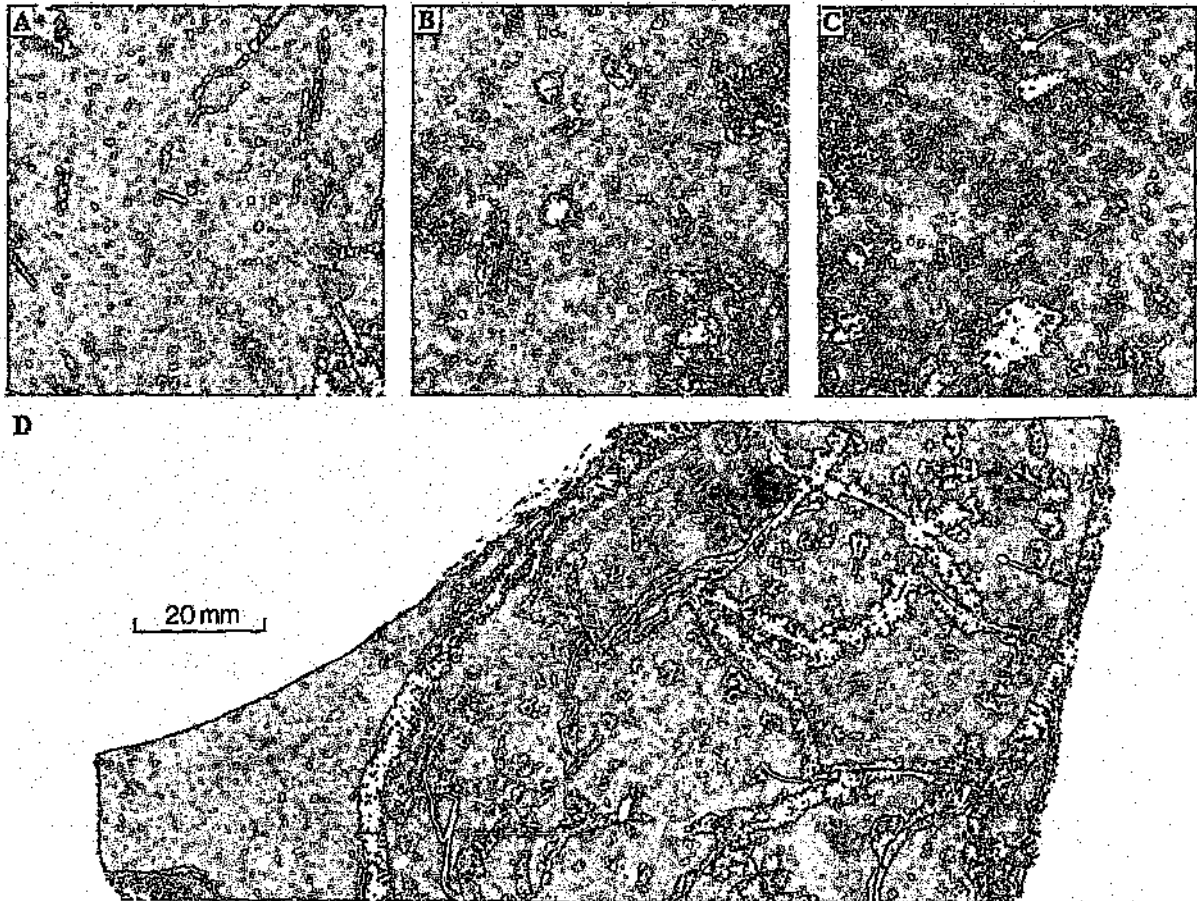


FIGURE 2.6

- A A photomicrograph showing the skeletal nature of augite (now amphibole) and olivine (pseudomorphed by quartz) in an ocelli (dark background). Olivine has a 'disc-shape' (diamond with two tails) whereas some of the amphibole needles have hollow habits as shown by their tendency to be split down their centres (eg. the needle pointing to the centre of the olivine pseudomorph. Sample: MC150c. Light: Plane polarized. Length of field of view = 2 mm.
- B Micro-orthocumultic, open-spaced subhedral to euhedral augite with intersertal, turbid (black), devitrified glass. The light-coloured patches probably represent subhedral olivine microphenocrysts replaced by assemblages of quartz $\pm$ plagioclase $\pm$ chlorite $\pm$ epidote. Note that a few grains of the close-spaced equivalent of this texture, shown in the next figure, occur in the bottom right corner. Sample: MC60a. Light: Plane polarized. Length of field of view = 1.2 mm.
- C Micro-orthocumultic, close-spaced anhedral to subhedral augite with interstitial, black, devitrified glass. The augite grains have thin coronas of actinolite. Sample: MC60a. Light: Plane polarized. Length of field of view = 1.2 mm.
- D A polished slab of pillowed komatiitic basalt showing an albitised pillow margin set against a sulphide-bearing interpillow hyaloclastite (left). The core of the pillow has a 'myrmekite-like' arrangement of ocelli set in a dark matrix upon which a network of contraction cracks with doubly zoned alteration halo's are superposed. Sample: MC60

pyroxene (Fig. 2.5D). The ocelli are different from the look-a-like textures of the metatholeiites of the overlying Hooggenoeg Formation which comprise needles radiating from a central point, warranting the descriptions variolitic or spherulitic as described by Gélinais *et al.* (1976). As mentioned previously, Ferguson and Currie (1972) suggest that the ocelli owe their origin to an inter-liquidus/solidus immiscibility loop based on melt experiments in which two distinct glasses were observed to form.

In pillow MC60 ocelli appear to have coalesced and have a dark vermiform matrix (Fig. 2.6D). The group of coalesced ocelli is central to the flanking pillow rims. Some of the ocelli have stretched shapes paralleling the flanking margins which suggest flow deformation and, therefore, within-pillow flow. Such a result confirms an earlier conclusion of within-pillow flow for the centrally situated micro-orthocumulate-textured augite. Petrographically, the ocelli are cloudy patches whereas the non-ocelli parts are translucent and light-green in colour, with little contrast between individual actinolitic grains. Precursor features, such as microspinfex, are contrasted more vividly by the turbid ocelli. Under high magnification the needle-like grains accentuate the light brown, cloudy, angular devitrified zones, best described as intersertal. The devitrified glass has plagioclastic compositions (determined by SEM) and contains numerous submicroscopic inclusions of titanite, amphibole and occasionally sulphide. This is in accord with Donaldson's (1982) observation that komatiitic lavas seldom crystallise plagioclase despite the common occurrence of plagioclase normative glass. In well preserved sections mesostases are characterised by thin, impurity-free zones next to the acicular grains suggesting that they represent nutrient-depletion zones in the original liquid prior to chilling. Closer towards the pillow rims the glass, which has a plagioclastic composition, adopts an interskeletal distribution accentuating the feathery crystal shapes of pseudomorphed augite (Fig. 2.5C). Around the micro-orthocumulate augite grains, described earlier, the plagioclastic mesostases appear as a chilled 'intercumulus liquid' phase.

Pseudomorphic replacements of euhedral olivine and clinopyroxene microphenocrysts by quartz and feldspar are frequently observed (less commonly by amphibole, chlorite and epidote). Silicification of olivines has also been documented by Gélinais and Brooks (1972)

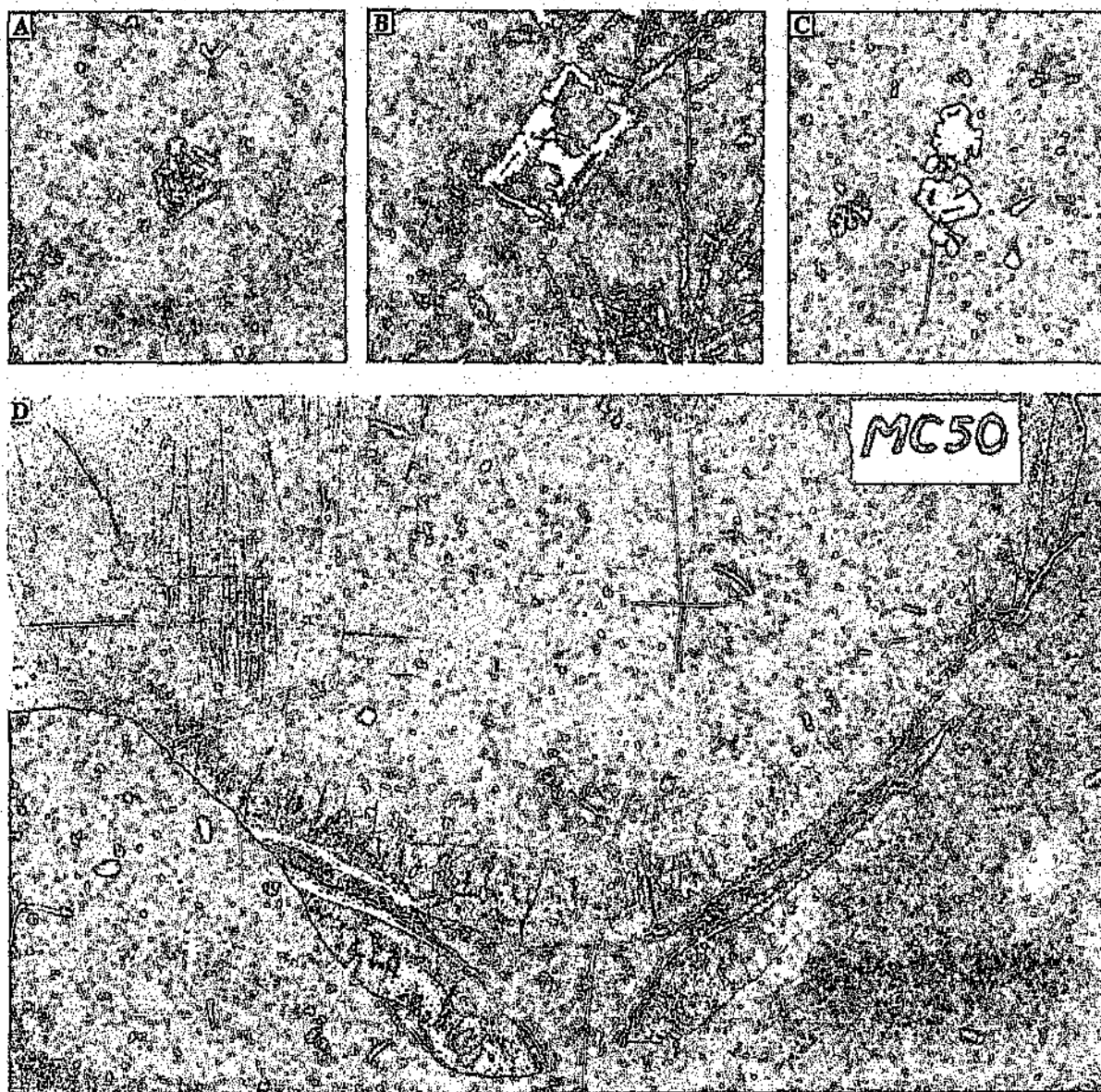


FIGURE 2.7

- A + B** Photomicrographs of lantern-shaped skeletal olivine pseudomorphed by quartz and feldspar. Sample: MC150c. Plane polarized light. Length of field of view = 250 micron each.
- C** Photomicrograph of pristine hopper chromite. Sample: MC180c. Reflected plane polarized light. Length of field of view = 250 micron.
- D** Partially silicified tholeiitic pillow basalt showing radially orientated elongate amygdales 5 - 20 mm long and 1 - 3 mm wide. Sample situated at the base of Hooggenoeg Formation. Note epidotised pillow rim. Sample MC50. Length of field of view = 35 cm.

in high-magnesian basalts and is attributed to isochemical replacement, on a volume for volume basis, involving the devitrification of a Si-enriched glass caused by the rapid growth of olivine. The pseudomorphs range from relatively large euhedral (1 mm) to small skeletal (≤ 0.3 mm) 'lantern' and disc shaped forms (Figs. 2.6A; 2.7A,B), the latter mainly after olivine (Donaldson 1982; Gélinais and Brooks 1972). The euhedral pseudomorphs suggest an early history of relatively slow cooling, probably early on in the melts (flow-) history, whereas the latter skeletal shapes imply relatively high degrees of supercooling towards the end.

Amygdales are generally small, spherical and rarely visible in outcrop specimens. On cut specimen surfaces amygdales range from 0.5 - 5 mm in size, with diameters larger than 2 mm being quite rare (Fig. 2.4A). The larger amygdales appear to be concentrated at the tops of pillows beneath the upper chilled margins, whereas the smaller ones are characteristically more evenly spread throughout the entire pillow. In general, the Komati Formation pillows appear relatively un-vesiculated and features such as ripe amygdales or amygdales with zonal, radial or vertical orientations were not observed. The low degrees of vesiculation of the komatiitic basalt pillows are consistent with the descriptions for the massive flows as furnished by Smith (1980) and Viljoen and Viljoen (1969a). This, however, is in contrast with the well vesiculated tholeiitic pillows of the overlying Hooggenoeg Formation. A detailed account of the degree of vesiculation and its significance in terms of environment of emplacement will be discussed below.

The amygdales are usually similar in size and composition to the pseudomorphed microphenocrysts except that they tend to be less regular in shape and rounder. Small euhedral chromite grains occasionally cluster around the vesicles (see also Fleet and MacRae 1975, Fig. 5). Interestingly, modal analyses indicate that up to 33% of sulphides contained in the pillows occur in the amygdales which constitute only between 1 - 3 volume percent of the pillows (Table 2.3).

2.4.3 Relict Igneous Minerals

Chromite. Grains of chromite occur in all pillows, unless highly carbonated (ca. > 10% CO₂). It forms mainly single euhedral (octahedral) grains, up to 150µm in diameter, that are microphenocrystic in contrast to others that are glomerocrystic and have small euhedral grains between 10-50µm in size. Optically chromite appears to be unzoned and skeletal chromite is occasionally observed closer toward the pillow rims (Fig. 2.7C).

Table 2.4 Relict igneous mineral compositions of pillowed komatiitic basalt

SAMPLE	MC150C	MC150C	MC150C	MC150C	MC150C	MC150C	MC200A	MC200A	MC170A	MC170A	MC170A	MC60C
ANALYSIS	34c01	34c01	34c01	34c01	32c02	32c02	03.e2	03.e5	04.a1	03.a2	02.a1	05a01
MINERAL	CHR	CHR	CHR	CHR	CHR	CHR	CPX	CPX	CPX	CPX	CPX	CPX
COMMENT	rim1	intrm.2	intrm.3	core4	intrm.1	core2	intrm.	intrm.	core	rim	core	rim
SiO ₂	0.09	0.11	0.10	0.11	0.10	0.10	51.69	52.19	51.76	54.14	52.10	53.30
TiO ₂	1.97	0.31	0.34	0.45	0.38	0.42	0.34	0.34	0.50	0.05	0.27	0.09
Al ₂ O ₃	12.04	6.94	6.97	7.19	7.78	7.73	2.70	2.53	3.34	0.69	1.27	0.96
Cr ₂ O ₃	39.06	54.14	55.04	55.26	52.99	53.29	0.52	0.90	0.25	0.70	0.18	0.11
FeO*	41.40	35.63	34.65	32.89	35.53	34.39	8.50	7.05	7.46	5.70	8.51	9.67
MnO	1.19	0.69	0.67	0.49	0.70	0.80	0.17	0.19	0.24	0.14	0.31	0.31
MgO	0.47	0.81	1.82	2.48	1.05	1.19	17.95	17.73	16.85	19.63	14.81	12.97
CaO	0.06	0.00	0.61	0.00	0.02	0.01	17.33	18.90	19.40	18.58	21.45	22.10
Na ₂ O							0.19	0.21	0.17	0.22	0.29	0.39
TOTAL	96.28	98.63	99.40	98.87	98.55	97.93	99.39	100.05	99.97	99.80	99.19	100.00
Number of ions on the basis of 32(O)						Number of ions on the basis of 8(O)						
Si	0.026	0.032	0.029	0.032	0.029	0.029	1.914	1.917	1.905	1.972	1.956	1.994
Ti	0.489	0.077	0.083	0.110	0.094	0.104	0.009	0.009	0.014	0.001	0.008	0.003
Al	4.149	2.396	2.367	2.4.7	2.673	2.662	0.118	0.110	0.145	0.030	0.056	0.042
Cr	9.026	12.532	12.532	12.569	12.207	12.308	0.015	0.026	0.007	0.020	0.005	0.003
Fe ³⁺												
Fe ²⁺	10.131	8.734	8.354	7.922	8.667	8.411	0.263	0.217	0.230	0.174	0.267	0.303
Mn	0.241	0.140	0.134	0.098	0.141	0.162	0.005	0.006	0.007	0.004	0.010	0.010
Mg	0.205	0.354	0.782	1.064	0.456	0.519	0.991	0.970	0.924	1.066	0.829	0.723
Ca	0.019	0.000	0.003	0.000	0.006	0.003	0.688	0.744	0.765	0.725	0.863	0.886
Na							0.014	0.015	0.012	0.016	0.021	0.028
CHR = chromite												
CPX = augite												
FeO* = All iron as FeO												

Two chromite grains have been analysed by electron microprobe (Table 2.4). The analyses represent traverses from rim to core. Reasonable analyses of the cryptically zoned rims were difficult to obtain due to their thinness. The larger octahedra are characterised by gradual

compositional changes across their inner portions and by sharp changes right at their outer edges. Cr, Mg, Al and Ti decrease gently in concentration towards the zoned edges with a concomitant increase of Fe and Mn. Marked reversals for the decreasing trends of Ti and Al occur at grain rims, whereas Cr and Mg continue to decrease but sharply. The gradual increases of Fe and Mn toward the grain margins become steep at the fringes.

The chromites are exceptionally enriched in Cr compared to chromites from stratiform intrusions, alpine complexes, spinel peridotite xenoliths and kimberlite and even for spinifex komatiite (Fleet and MacRae 1975). They are very low in Mg, Al, and high in Fe making them the purest 'ferrochromites' (FeCr_2O_4) observed in the literature (Deer *et al.* 1962; Fleet and MacRae 1975; Evans and Wright 1972; Donaldson 1982).

The distinct compositional changes at grain margins are cryptic as they are not visible in reflected light. The gradual compositional changes for the inner parts are probably related to igneous fractionation whereas the sharp compositional changes at the edges, may be related to the superimposed effects of metamorphism. Observations of similar Fe-rich rimmed chromites have been reported for the spinifex rocks of Munro Township (Fleet and MacRae 1975) and quenched ocean floor basalts (Bryan 1972). Both workers, however, note thin selvages of magnetite around the chromites which was not seen in this study. Metamorphism is thought to have influenced chemical zoning of chromite in alpine-type intrusives in India (Mitra, 1972).

Augite. Relicts of augite are often preserved in pillows that do not exhibit microspinifex textures (Table 2.4). The mineral occurs as equant, euhedral to anhedral grains between 100-200 μm in diameter and is rimmed by thin coronas of actinolite. It is common, however, for augitic pyroxenes to be preserved in massive flows of Archaean and Phanerozoic greenstones but the mineral is rarely found in pillows.

The pillow augite compositions are compared with those published for komatiite flow-rocks of the Komati Formation, Munro Township and Belingwe (Fig. 2.8). Generally, there seem

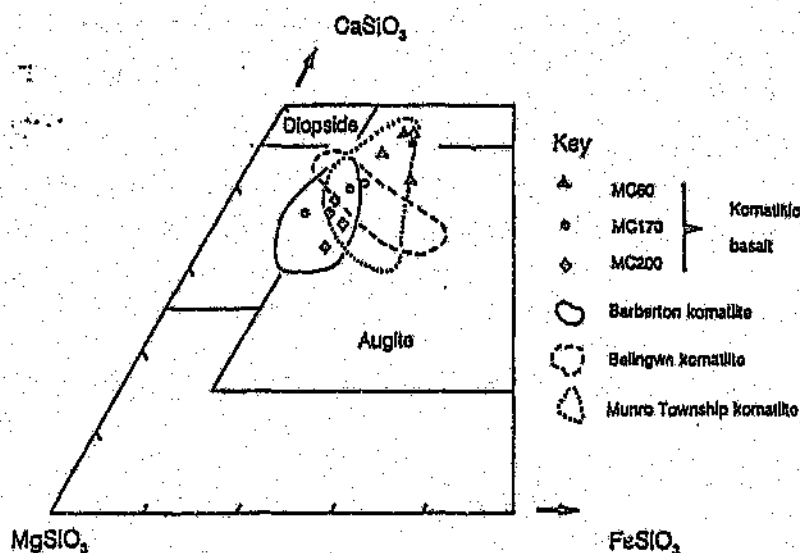


Fig. 2.8

Augite compositions of pillowed komatiitic basalt of the study area compared with augite from massive komatiite from Belingwe, Munro Township and the Komati Formation.

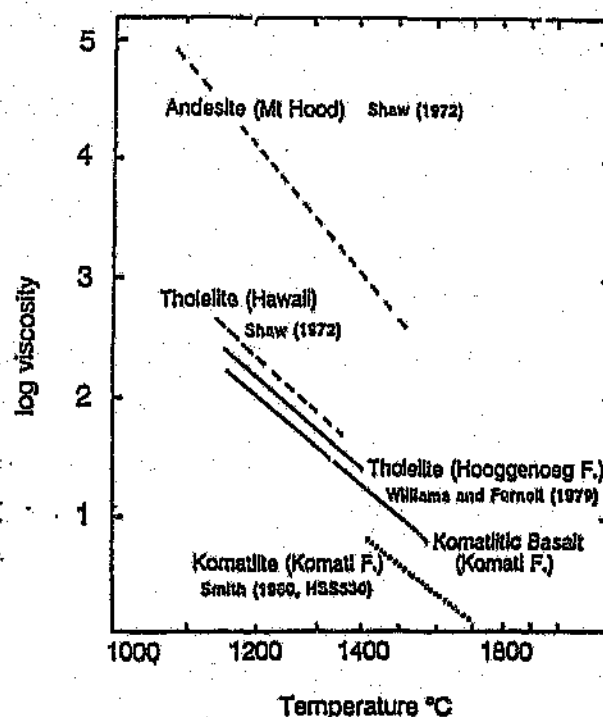


Fig. 2.9

Comparison of viscosities calculated from chemical analyses of komatiitic and tholeiitic basalt and komatiite of the Onverwacht Group. Sources of chemical analyses are indicated next to curves and in Table 2.5. Also included for comparison are the viscosity curves for tholeiite from Hawaii and andesite from Mt Hood. For these rock types the curves have been adopted from the cited author.

to be only minor differences in terms of MgO, FeO and CaO between the various groups and interestingly their ultramafic counterparts from the Komati Formation have only slightly higher MgO and lower CaO and FeO. They are close to the Munro Township komatiite augites but distinctly more CaO-rich than those from Belingwe. Compared with augites from normal basalts in terms of Ti and Al, they plot on the low Ti side of the basalt field but on the komatiite trend as demonstrated by Fleet and MacRae (1972, Fig. 10). On the komatiite trend, however, the pillow augites have half as much Al as the acicular pyroxenes from Munro Township.

It appears that skeletal augites in komatiites, which grow rapidly, are characterised by low Si and high Al contents. Fast growth induces incorporation of abundant Al before diffusion away from the crystals occurs, and the inability of plagioclase to nucleate implies that growing augites do not have to compete for Al, as the Al content of the melt remains high (Donaldson, 1982). On the contrary, the preserved pillow augites of the study area, which have euhedral to anhedral crystal shapes, have higher Si and lower Al than those of the spinifex (skeletal) textured komatiite which confirms a slower growth rate as implied earlier.

2.5 Lava Rheology

Relatively little volcanological data exist for komatiitic basalts. If komatiitic basalts can be shown to have similar rheological properties to tholeiitic basalt, analogies can then be readily made to the wealth of data that exist for this class of rock. Thus, to estimate the environment of emplacement of the komatiitic basalts an understanding of its rheology is necessary, i.e. viscosity, volatility and vesiculation. Viscosity, for example, determines discharge rates, flow thicknesses, flow structure and the ease with which volatiles are lost.

2.5.1 Viscosity

Factors governing magma viscosity are temperature and composition (Bottinga and Weil 1972; Shaw 1972; Urbain *et al.* 1982). The control that composition has on the viscosity of a silicate melt is due to the creation of molecular chains by means of covalent bonding

between silicon and oxygen atoms. Viscosity rises as the degree of polymerization of the melt increases. Alkalies and alkaline-earth elements exist as cations within silicate melts and serve to break up the silicate chains, forming weaker ionic bonds and thus lessening the viscosity of the melt. Generally, low silica melts have low viscosity (Basaltic Volcanism Study Project 1981). Various empirical methods for the prediction of magmatic viscosities exist in the literature (Bottinga and Weil 1972; Shaw 1972; Urbain *et al.* 1982).

Table 2.5: Analyses used to calculate viscosities for Onverwacht Group tholeiitic and komatiitic basalt and komatiite.

Weight % oxide (Analyses normalised to 100%)	Tholeiitic basalt (Anal. no. 11) (Williams and Furnell, 1979)	Komatiitic basalt (Average of MC60, MC120, MC150 and MC180)	Komatiite (HSS530) (Smith, 1980)
SiO ₂	50.10	51.86	47.35
Al ₂ O ₃	13.91	9.79	3.99
Fe ₂ O ₃	2.78	3.07	2.00
FeO	12.09	9.81	9.99
MgO	6.90	10.52	28.89
CaO	9.67	11.97	7.05
TiO ₂	0.86	0.69	0.36
Na ₂ O	2.50	1.76	0.12
K ₂ O	0.27	0.14	0.02
H ₂ O	0.60	0.04	0.00
MnO	0.24	0.18	0.19
P ₂ O ₅	0.07	0.16	0.02
Total	99.99	99.99	100.00
Mean slope: $s =$ (Shaw, 1972)	2.17	2.09	1.88

The method of Shaw (1972) is used to calculate the viscosities of the three major lava types of the Onverwacht Group and are compared with a Hawaiian basalt and an andesite from Mount Hood (Fig. 2.9). Despite some compositional differences between the oxide constituents, on a one-for-one basis (Table 2.5), the calculated viscosities for the two basalt types are relatively close (Fig. 2.9). Values for basalts in general vary between a 100 and

1000 poises at $\sim 1200^{\circ}\text{C}$ (Basaltic Volcanism Study Project 1981). These values bracket the calculated values of Onverwacht komatiitic and tholeiitic basalt, i.e. ca. 120 and 200 poises respectively. Komatiite on the other hand has a very low viscosity of ca. 10 - 15 poises at an emplacement temperature of ca. 1550°C . These values are in agreement with those calculated by Smith (1980) who suggests that komatiite melt viscosity is similar to that of warm glycerol.

It is evident from these calculations that the small differences in viscosity between komatiitic and tholeiitic basalt allow direct comparison of flow structures. The small differences suggest that equally small deviations regarding size and shape of flow forms should result with all else being equal. Komatiite on the other hand would be expected to produce very flat or sheetlike flow-rock forms.

2.5.2 Vesiculation

Vesiculation is controlled by the exsolution of gas from a lava melt. The volume of vesicles that form is a function of temperature, pressure, solubilities of the various gaseous species in the melt and the mass fraction of each gaseous component originally in solution (Macpherson, 1984).

Early assumptions (e.g. McBirney, 1963) that confining pressures brought about by ocean depth regulate to a large extent the degree of vesiculation in submarine lavas has been substantiated by the findings of Moore (1965, 1970) and Moore and Schilling (1973).

Moore and Schilling (1973) show that a definite relationship exists between vesiculation and sampling depth (depth of emplacement) of mid-ocean ridge basalt (MORB) pillows. They record vesiculation and gas phase separation at dredge depths up to 5000 m, but they note that significant migration of volatiles out of pillows occur only above 200 m depth. Their data also indicate that samples formed at a depth of 1000 m contain 5% vesicles; at 700 m 10% and 500 m 16%. Pillow samples collected at depths of 700 m and less, usually contain radially elongate (pipe) vesicles in their outer layers, 1-3 mm in diameter and 5-20 mm long.

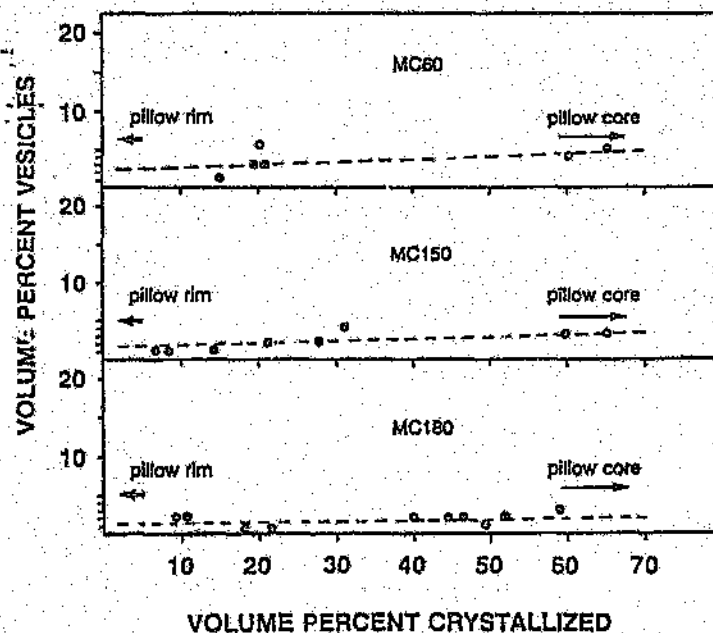


Fig. 2.10 Measured variation in volume percent of vesicles in the pillows MC60, MC150 and MC180 versus degree of crystallisation (= 0% at pillow rim and ~70% in pillow core - see text for explanation).

Macpherson (1984) shows that the volume percentage of vesicles in submarine pillow basalt can be modelled by the exsolution of H_2O and CO_2 vapour using the available solubility determinations for these components in basalt. In agreement with studies on natural samples, CO_2 is calculated to be the major vapour component in vesicles at eruption depths of $> 1-2$ km, confirming the findings of Moore (1970). Small initial contents of CO_2 (0.1 - 0.2 wt%) are compatible with the vesicle volumes observed in most deep-sea pillow lavas. The demonstration by Moore (1970) that H_2O becomes a significant vesicle-forming component at relatively shallow extrusion depths (< 2 km) is also confirmed by the model predictions of Macpherson (1984). Furthermore, it is reasoned that gaseous sulphur is only an important vesicle-forming component at depths of less than 200 m ($P \leq 20$ bar) and that its contribution at greater depths is insignificant. Consequently it seems that vesicles in submarine lavas are due mostly to CO_2 at ocean depths of > 2 km, $CO_2 + H_2O$ at depths between 2 km and 200 m and $CO_2 + H_2O + SO_2$ within 200 m of sea level (Macpherson, 1984).

The degree of vesiculation of the pillows in the study area has been determined by means of modal analyses (Table 2.3). For all the pillows analysed, including two from the slightly deformed area north of the Theespruit Pluton, an average of 2.6 ± 1 area % vesiculation (from rim to core) is indicated. These analyses were performed on 37 polished thin sections with an average count of 518 points per section and between 6 and 10 sections per pillow, representing traverse from core to rim. In an attempt to constrain the depth of emplacement, only the undeformed pillows MC150 and MC180, which occur on the same stratigraphic level, and MC60 occurring 1.6 km higher up, are discussed.

Figure 2.10 illustrates the distribution of vesicles from pillow rim to core and is expressed as volume percent vesicles against volume percent of pillow crystallized. The volume percentage of crystals is assumed to range from zero at the chilled margin and up to $>70\%$ in the pillow core. At 70% crystals (crystallisation) a lava is thought to be effectively rigid and that no further increase in vesicle volumes occur (Wright and Okamura, 1977). For the samples studied here the degree of crystallization has been calculated by equating the dimensions of the coarsest textures in the central core to 70% (a point at which pillow rigidity would inhibit grain-length increases) and expressing the intermediate grain dimensions between rim and core as a percentage of this. None of the studied pillows have holocrystalline cores, an aspect favouring this approach.

Data presented in Fig. 2.10 for pillows of the same stratigraphic elevation show similar degrees of vesiculation in their chilled margins i.e. $1.5 \pm 0.2\%$. A slight increase in the degree of vesiculation is observed with increasing crystallisation i.e. from ca. 1.5 to 3% and 1.5 to 2% for pillows MC150 and MC180 respectively. In essence the data curves are very flat and have no inflections. The stratigraphically higher pillow MC60 has slightly higher value of 2.2 and 5% for the pillow margin and core respectively.

The slight increases in vesicle volumes towards pillow cores appear to be the result of cooling and increased crystallization on the fractionation of volatiles within the individual pillow melts. Macpherson (1984) found that inversely proportional degrees of chilling and

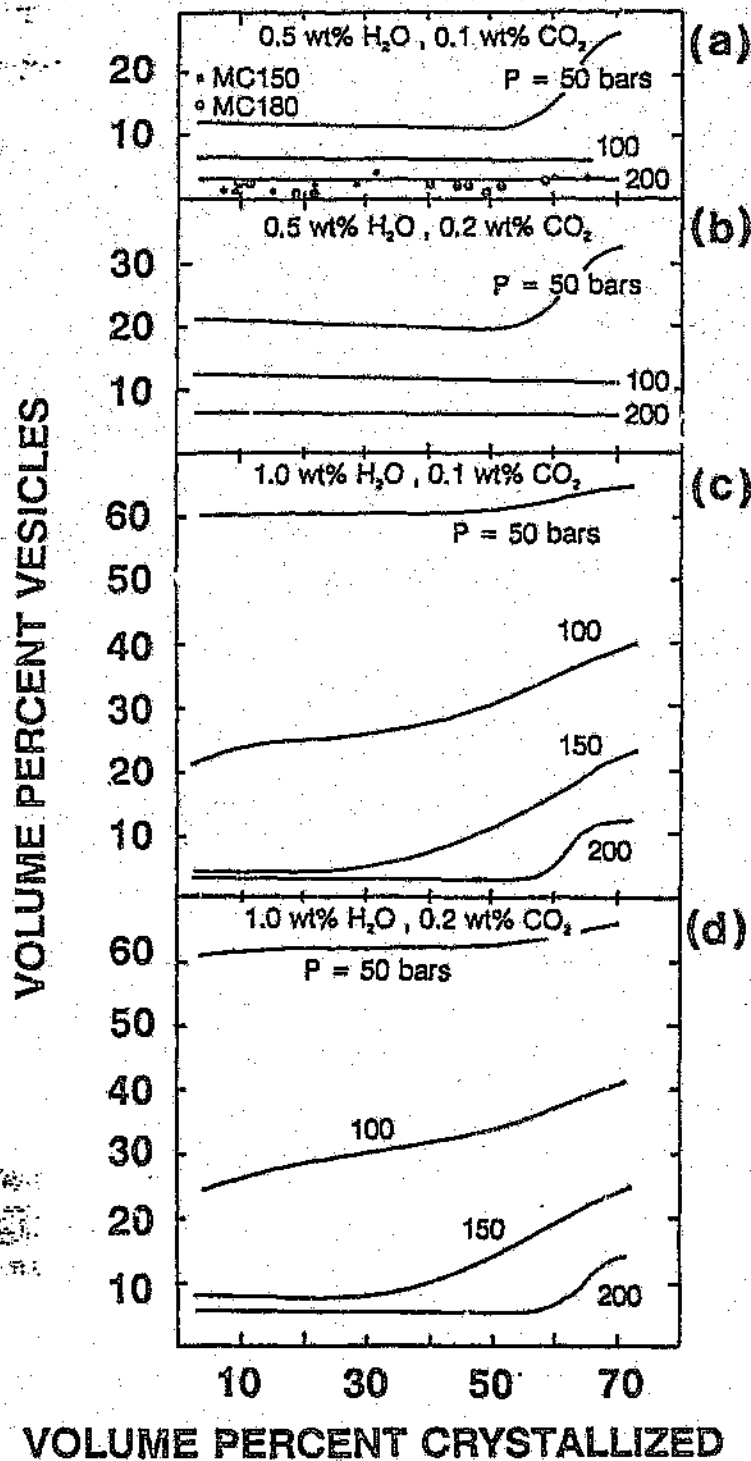


Fig. 2.11

Calculated variation in volume percent of vesicles for individual pillows versus crystallinity (= 0% at pillow rim and ≈ 70% in pillow core), for a range of eruption depths. Assumed initial volatile contents are as shown. The measured values for MC150 and MC180 are plotted in (a).

crystallisation occur at rims and cores respectively, so that volatiles become relatively concentrated in the residual melts, and hence larger vesicle volumes may be found in the cores of pillows.

In contrast to the flat trajectories of the study area pillows, cases arise where decreasing pressure (ocean depth) and/or crystallisation causes water saturation to occur within the residual melt, resulting in sharp increases in vesiculation between rim and core. Such instances have been modelled by Macpherson (1984, Fig. 5-7) showing the variation of vesiculation in individual pillows as a function of crystallinity for both a range of eruption depths and initial volatilities in Fig. 2.11.

2.6 Estimation of Emplacement Depths

2.6.1 Ocean Depth during Komati Formation Time

To estimate the ocean depth of extrusion of the komatiitic basalts it is necessary to know the degree of initial vesiculation and the volatility that caused it. The first requirement of initial vesiculation has been obtained by means of modal analysis (Fig. 2.10) of which the data for pillows MC150 and MC180 have been transferred to Fig. 2.11(a). The figure shows the nature of vesiculation as a function of crystallisation and initial volatility where graphs (a) to (d) respectively represent melts with increased volatility. (From the graph headings, which indicate initial volatility, it is evident that for a fixed degree of vesiculation greater depths are implied by increasing initial volatility).

The second requirement of initial volatile content, however, cannot be determined directly as pristine unaltered rock or chilled melt no longer exists. This can be assessed through indirect means:

- (1) From the flat profiles of the samples studied here it is evident that H_2O saturation/exolution did not occur and that CO_2 was the only gaseous species controlling vesiculation.
- (2) The degree of light rare earth element (LREE) enrichment of basalts, as characterised by the chondrite-normalized ratio La/Sm_n , is thought to give a general

indication of initial volatility (Macpherson 1984). The komatiitic basalts of the study area have a La/Sm_n ratio of 0.7-1.5 (see samples 5080, 5084 of Jahan *et al.* 1982). Such a value is slightly higher than that for average MORB, implying a volatility slightly greater than 0.5% H_2O and 0.1% CO_2 (Macpherson 1984).

- (3) Due to the somewhat more primitive composition of komatiitic basalt compared to MORB and lower alkali-content, it may be expected, on the other hand, that their melts may be less evolved with respect to volatile content. If the lower limit of 'dry' basalt is adopted as the lower limit of the komatiitic basalt it would place them alongside the driest MORB melts which have an initial volatility of 0.4% H_2O and 0.05% CO_2 (Macpherson 1984).

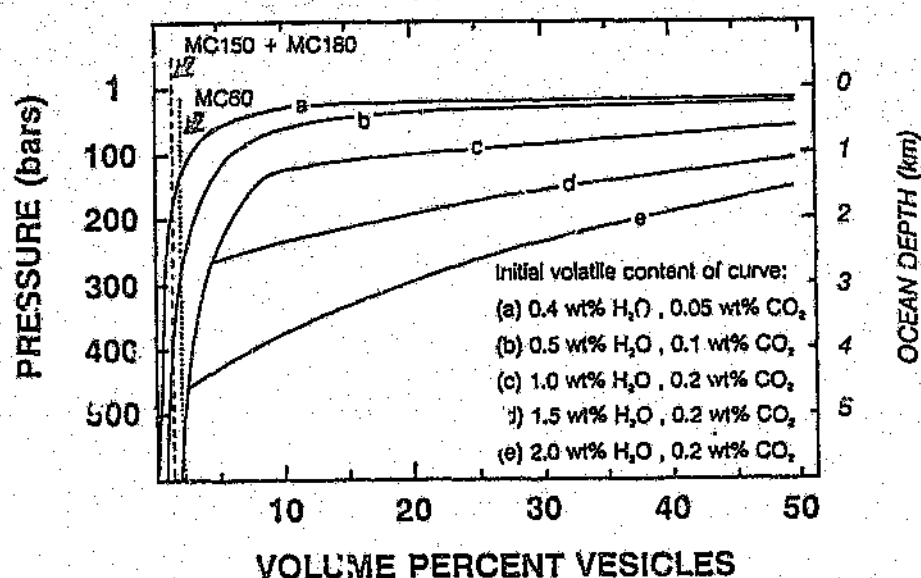


Fig. 2.12 Calculated volume percentage of vesicles in glassy pillow basalt rinds as a function of eruption depth for a range of initial dissolved H_2O and CO_2 contents as indicated, after Macpherson (1984, fig. 4). The measured volume percent vesicles for the studied pillows are indicated as vertical dashed and dotted lines.

Therefore, the lower and upper limits of initial volatility for the komatiitic basalts are selected to be 0.4% H_2O and 0.05% CO_2 and 0.5% H_2O and 0.1% CO_2 . These values are in accordance with La/Sm_n ratio and primitive composition of the komatiitic basalts and comparable with the range found in MORB.

To estimate the depths of emplacement of the Komati Formation pillows the relevant data of Macpherson (1984, Fig. 4) is summarised in a plot of volume percent vesicles in pillow margins vs depth of eruption for a range of initial volatilities (Fig. 2.12). If the measured vesiculation of the komatiitic basalts is plotted, extrusion depths of 4 and 3 km are derived respectively for pillows MC150 and MC180, and MC60 utilizing the higher volatility of 0.5% H₂O and 0.1% CO₂ (curve b) whereas depths of 1800 m and 1300 m are derived using the lower volatility of 0.4% H₂O and 0.05% CO₂ (curve a).

The validity of the selected upper- and lower volatilities can be evaluated independently by assessing the degree of isostatic adjustment implied by the two extremes. Control on the degree of isostatic compensation is given by knowing the difference in stratigraphic height between the lower (MC150 and MC180) and upper (MC60) pillows (Fig. 1.4, ie. = 1.6 km). The degree of isostatic settling that occurred between emplacement of the lower and upper pillows is given by the relationship:

$$I = \left[1 - \left(\frac{e_1 - e_2}{S} \right) \right] \times 100$$

where I = the degree (%) of isostatic adjustment after emplacement of the upper pillows, S = difference in stratigraphic elevation of the lower and upper pillows and e₁ and e₂ are the estimated ocean depths respectively for the first and second emplacements. Applying the depths derived from the above mentioned higher initial volatility (i.e. 4 and 3 km), isostatic compensation amounting to 37.5% is indicated or alternately the ocean floor shoaled by 1000 m relative to the 1600 m of effusive rocks that were laid down. Application of the lower-volatility derived depths (i.e. 1.8 and 1.3 km) shows that isostatic compensation amounted to 60.5% or that the ocean floor shoaled only by 500 m in response to the emplacement of 1600 m of lava.

Isostatic compensation for the Komati Formation could not have been too rapid or complete (i.e. 100% or stayed at a constant depth) because shallow-water sediments occur at its stratigraphic top, the Middle Marker (Lanier and Lowe, 1982). With the higher-volatility,

implying an initial emplacement depth of 4 km and ~40% isostatic compensation, shallow water conditions would not have been attained, given the 3.5 km thickness of the Komati Formation, suggesting this option to be unrealistic. The known rapidity with which isostatic adjustments occur, for example the 2.6 mm/y for the loading of oceanic crust by Hawaiian extrusives (Ludwig *et al.*, 1991), tends to support the lower-volatility derived depth as a better approximation of the two.

2.6.2 Ocean Depth during Hooggenoeg Formation Time

Shallow depositional environments have been predicted for the sedimentary units of the Hooggenoeg and overlying formations. These estimates are based on sedimentary structures such as thin evaporite beds (Lowe and Byerly, 1986) and stromatolites (Byerly and Palmer, 1991) which were formed during periods of volcanic quiescence. Estimates based on the stromatolites and mudpool structures, are for sequences overlying the Hooggenoeg Formation. The presence of tourmaline-bearing sandstone in the Middle Marker and evaporitic cherty sediment slightly higher up in the sequence are interpreted to be indicative of shallow marine conditions for the Hooggenoeg Formation (Byerly and Palmer, 1991 Fig. 3a,b). Boron isotopes of the tourmaline indicate a major evaporitic component, thought to be consistent with a shallow environment. Sedimentary structures of the Middle Marker are locally current/wave influenced with no diagnostic evidence for tidal activity suggesting relatively shallow conditions (Lanier and Lowe, 1982).

It is notable, however, that pillow vesiculation at the base of the Hooggenoeg Formation is not scorlaceous or indicative of having degassed. Degassing of tholeiitic basalt typically occurs at depths shallower than 200 m (Moore and Schilling, 1973) and may therefore be expected if the above-mentioned shallow-water environments prevailed. Figure 2.7D shows the degree of vesiculation of a tholeiitic pillow basalt (MC50) situated just above the Middle Marker. Along the inner margin of the pillow selvage, radially elongate amygdalae up to 3 mm in width and between 5 and 20 mm long occur. Such vesiculation, however, is typical of emplacement depths of about 700 m (Moore and Schilling, 1973). The apparent variance of the data may suggest that ocean depth was not constant and that it may have fluctuated

between sedimentation and subsequent volcanicity or that the initial volatility of the lavas were markedly lower than average. The possibility that the lavas degassed subaerially prior to emplacement into the shallow sea is discounted on the grounds that no evidence for this could be found, e.g. spongy pahoehoe (Walker, 1989) and lava overlying foreset-bedded pillow breccia (Moore *et al.*, 1972).

The degree of vesiculation of lavas higher up in the Hooggenoeg Formation appears to increase, suggesting upward shoaling. This data is consistent with the sedimentological and evaporite data suggesting that shallow conditions existed. The pillows, nonetheless, appear not to be degassed. According to the sedimentological and vesiculation data for the upper part of the Hooggenoeg Formation, it is concluded that emplacement depths also fluctuated between periods of sedimentation associated with volcanic quiescence (i.e. quiescence sedimentation) and volcanism. At the base of the Hooggenoeg oscillations may have ranged between 0 - 700 m (0 - 70 bar) with the range of depth decreasing and the period of quiescence increasing with time, as is evidenced by (1) the upward increase in the frequency of chert horizons (e.g. Williams and Furnell, 1979; Lowe, 1991), and (2) the vesicle-based observations for upward shoaling.

Temperature conditions during quiescence-sedimentation and seafloor metasomatism (chertification) have been determined by other workers. Seawater and hydrothermal conditions are generally thought to have ranged between 60° - 150°C (Lowe and Byerly, 1986; de Ronde, 1991). Knauth and Lowe (1978) have demonstrated by means of oxygen isotopes that cherts with fine-grained quartz after gypsum have formed through exchange with seawater at about 70°C.

2.7 Age of Volcanism

Various attempts have been made at dating the Komati Formation and are summarised in Table 1.1. Lopez-Martinez *et al.* (1992) reported a 3486 ± 8 Ma ^{40}Ar - ^{39}Ar age for a komatiite sample which is thought to represent the age of metamorphism (imprinted shortly after volcanic emplacement). The reported age is difficult to evaluate because the sample

yielded several ages ranging from 3486 ± 8 Ma to 3417 ± 36 Ma. Furthermore, only one of several komatiite samples produced this age. Camo and Davis (1990, 1994) have dated a soda-porphyry at $3467 \pm 12/-6$ Ma and 3458 ± 2 Ma respectively which represent minimum ages for the Komati Formation. Armstrong *et al.* (1990) report a 3472 ± 5 Ma zircon age from an interflow sediment more or less at the same stratigraphic level as the mentioned soda porphyry. This interflow sediment is known as the Komati Unit and is thought to represent a penecontemporaneous volcanoclastic sediment, therefore, defining the maximum age of the Komati Formation. It is thus apparent that the emplacement span of the Komati Formation may be as short as 5 Ma and as long as 16 Ma depending on which of the above mentioned minimum ages are selected.

The age of the Hooggenoeg Formation is less tightly constrained. The case of the Hooggenoeg Formation, containing the Middle Marker as described above, constrains the maximum age of the formation at 3472 ± 5 Ma (Armstrong *et al.*, 1990). The minimum age of the Hooggenoeg Formation is constrained by the maximum age of the overlying Buck Ridge volcanoclastics dated at 3445 ± 3 Ma (Göner *et al.*, 1991) (Table 1.1). The duration of Hooggenoeg volcanism is thus constrained to ~25 Ma. The maximum ages of the Komati and Hooggenoeg Formations, however, are identical (at 3472 ± 5 Ma) as defined by the Komati Unit and the Middle Marker (Table 1.1). The significance of this concurrence has received little attention in the literature and is discussed below.

Preliminary palaeomagnetic work on the Komati and Hooggenoeg Formations show that their magnetic poles coincide (Hale and Dunlop, 1984; Layer, 1991). Coincident magnetic poles imply that the two formations were temporally and spatially close at the time of palaeomagnetic imprinting. Imprinting is thought to have resulted during volcanism and/or seafloor metamorphism which are assumed to be coeval. Coincident poles for the two formations suggest that they may have been part of the same volcanological process and that emplacement was spatially focused and temporally rapid, negating the effects of plate motion. Rapid emplacement favours the (above-mentioned) 5 Ma emplacement span of the Komati Formation and may suggest that Hooggenoeg Formations also has a short

emplacement duration. A short volcanic emplacement span of 15 Ma has been demonstrated for komatiite volcanism in the Norseman-Wiluna greenstone belt, Western Australia (Olson *et al.*, 1988).

2.7.1 Age of Tectonic Emplacement

An allochthonous origin has been postulated for the Komati Formation on the basis of field evidence (Williams and Furnell, 1979; de Wit *et al.*, 1983) and geochronology (Armstrong *et al.*, 1990). As the Komati and Hooggenoeg Formations represent an intact volcanic sequence it is concluded that at least both formations constitute the allochthon. The top of the allochthon, however, is marked by the contact between the mafic and basic Hooggenoeg Formation and the felsic Buck Ridge porphyry and associated volcanics (Fig. 1.3). The felsic and mafic volcanics have traditionally been grouped together as the Hooggenoeg Formation but is impracticable on the grounds that, (1) it would seem that there is overwhelming evidence in favour of an allochthonous origin for the Komati and Hooggenoeg Formations as suggested above, and (2) that the Buck Ridge porphyry/volcanics are essentially autochthonous as is evidenced by their cogenetic relationship with the deep-seated trondhjemites that intrude the Theespruit Formation (de Wit *et al.*, 1987b). The Buck Ridge porphyry/volcanics can therefore not also be part of the allochthon.

The maximum and minimum ages of tectonic emplacement of the Komati/Hooggenoeg allochthon are constrained by the minimum 3453 ± 6 Ma age of the Theespruit Formation and the maximum age of the Buck Ridge porphyry which is 3445 ± 8 Ma. This results in a well constrained ~8 Ma time-span for the tectonic emplacement of the allochthon.

2.8 Discussion

The early hypothesis of the propagation of pillows by bulbous budding (Lewis, 1914), which forms the basis of most current thinking on the formation of pillow lavas, has been confirmed by Moore *et al.* (1972), Moore (1975) and Ballard and Moore (1977). They observed and documented for the first time the process of subaqueous pillow formation. Moore (1975) shows that actively growing pillow lavas consist of elongate, interconnected

flow tubes that are elliptical or circular in section. The bulbous flows are fed from upstream by a larger network of lava tubes that maintain the lava pressure within the growing lobes. Discrete pillow sacks are rare but occasionally form when a flow tube is nipped off on a steep flow front and plunges down to its base. Moore (1975) also observed that pillow buds do not grow by the expansion of an outer plastic selvage. They were seen to expand at incandescent cracks within the pillow crust where new crust was continuously produced. They branch by the opening up of cracks on the sides or tops of tubes through which new lobes emerge, expanding the new incandescent zone. Depending on the speed of expansion specific structures are produced in the glassy rim known as corrugations and fault slivers for fast and slow dilating pillows, respectively.

Ballard *et al.* (1979) observed after eruptions on Hawaii and the Galapagos Rift that sheet-flows represent early, brief but voluminous eruptions. They are followed by slower but more sustained eruptive phases that produce pillow basalts. They noted that the voluminous early flows spread as open sheet flows that tend to pond in low lying areas. Extended fissure vents were observed to be left open when eruptions ended and that vent edifices were small or lacking. The latter eruptions were observed to produce broad fields of tube-fed pillows that tend to cover the earlier sheet-flows. Interleaved with minor sheet-flows these areas grow into small shields and, with the continued extrusion of many flows, the formation of channelised flows emerge.

Comparison of Archaean with modern subaqueous lava flows in recent years has shown that they have similar morphologies and that their genesis may be similar to their modern sea-floor counterparts (Dimroth *et al.* 1978; Wells *et al.* 1979; Cosineau and Dimroth 1982). These authors indicate that the ancient lavas are also associated with an abundance of massive lava that grades laterally into pillowed and adjoining volcanoclastic units. Moore *et al.* (1973) show that the difference between subaerial (littoral) and subaqueous flows in a tidal zone, is that subaerial massive lava (pohoe) overlies pillow breccia with large-scale delta foreset bedding. Dimroth *et al.* (1978) and Cosineau and Dimroth (1982) contend that ideal subaqueous flows grade from massive lava bases upwards into pillowed lava,

brecciated pillow lava and stratified hyalotuff (with one or several of the subunits being missing in less ideal flows but the vertical progression is always the same). Of the various permutations, the most common vertical transition is massive lava - pillow breccia $\pm$ hyalotuff and pillow lava - pillow breccia $\pm$ hyalotuff. The transition from massive to overlying pillowed lava in the same flow is considered to be rare.

In contrast to this, Baragar (1984) shows that when both massive and pillowed facies are present in the same flow, massive lava always overlies a pillowed base. This is demonstrated for the Archean komatiitic and basaltic tholeiites of eastern Hudson Bay where new insight was gained into the mechanism of pillow development. Baragar (1984) demonstrates that most flows with clearly recognisable boundaries are pillowed at their bases and have massive upper parts. The pillows are envisaged to form at the leading edge of a massive flow where the frontal convex-shaped crust ruptures and lava is fed into a number of self sealing tubes while the remainder of the frontal surface stabilizes. At this stage the flow passes from open channel to restricted tube conditions and the rate of flow is decreased. As a result the lava level rises behind the stabilized flow front and lava overflows the impediment just to repeat the process farther ahead. Reoccurrences of tube formation at the flow fronts and successive overriding and interference of the approaching lava results in a entangled carpet of lava tubes over which the massive phase passes. The ratio of pillowed to massive lava within flows partly depends on the flow velocity. A swiftly oncoming lava front would be inclined to overtake the propagating tubes forming at its base and pillows would thus be few or absent. On the other hand, a near-stationary flow front could build a mound of pillows as the lava slowly rises and spills repeatedly over the growing barrier.

2.8.1 Quenched Magma

Microscopic investigation of the komatiitic metabasalt pillow lavas of the Komati and Hooggenoeg Formations reveal that they are quenched magmas. The pillows generally exhibit both external and internal textural features indicative of magma quenching. These are (1) relicts of thermally-strained glassy selvages as revealed by curved hairline fractures; (2) pillow rims passing into microlitic border zones; (3) skeletal olivine, chromite and

pyroxene, similar to those reported for flow-rocks of the Canadian greenstones and modern ocean-floor; (4) the common presence of plagioclasic mesostasis in the cores of pillows; and (5) the occurrence in most pillows of microspinel, a texture comprising axiolitic- to acicular skeletal augitic-pyroxene pseudomorphed by amphibole. Only occasionally do pillow cores comprise micro-orthocumultic augite, textures indicative of slower cooling but which are always enclosed in a mesostasic intercumulus, suggesting cooling rates still high enough for quenching. The slower cooling of such cores is attributed to a higher heat-supply linked to within pillow flow and/or formation of crystals in a slow cooling massive flow which is ultimately drained into pillow tubes.

The possibility that the Komati Formation pillows could be intrusive, as suggested by de Wit and Stern (1979), is doubtful on the basis of the commonly observed evidence of lava quenching. Moore (1975) shows that due to the rapid growth of crystals in mafic melts, basalt must cool from liquidus to solidus temperature in about 3 seconds for glass to occur (without crystals). On the other hand for any glass to be preserved, cooling from liquidus to solidus must be less than 13.5 seconds. The occurrence of glassy pillow surfaces, by analogy, would hardly be possible in a high heat-flow sheeted dike environment. The question of the pillows being so-called pilloids, a term used by de Wit *et al.* (1987) to describe hydrothermally altered, polygonally jointed komatiite, mistakenly seen by Viljoen and Viljoen (1989) to represent pillowed komatiite, has been disputed by Cloete (1991b).

The small difference in viscosity between the low-MgO komatiitic basalts and the tholeiites of the study area have not resulted in any significant difference in pillow shapes, structures and other flow forms. A statistical analysis, however, may reveal that the tholeiites have, on average, marginally larger dimensions than their komatiitic counterparts.

2.8.2 Subaqueous Emplacement of Flows

The absence of massive lavas overlying foreset-bedded pillow breccia or pillow lava in the study area suggests the absence of littorally emplaced lava for the Komati and Hooggenoeg Formations. The absence of spongy pahoehoe and scoriaceous material and the presence

of pillow lavas eliminates the possibility that the Komati and Hooggenoeg Formations formed subaerially. On these grounds it is inferred that emplacement of the Komati and Hooggenoeg Formations was entirely subaqueous and that very shallow conditions (≤ 200 m), at which lava degassing would have occurred, did not arise during active volcanism. This contrasts with evidence for shallow water sedimentation and evaporites which seem to have formed during periods of volcanic quiescence of the Hooggenoeg Formation (e.g. Lanier and Lowe, 1982; Byerly and Palmer, 1992). Ocean depth is thus inferred to have oscillated with deeper water conditions coinciding with periods of volcanism.

2.8.3 Sheet-Flood Flows vs Channelised Flows

Cousineau and Dimroth (1982) envisage sheet-flood flows to propagate radially away from overflowing lava fissures and all parts of the flow front being progressive and dynamic. The sheet-flood mechanism operates as long as flow rates remain high enough to counteract the effects of differential cooling. Such differences are thought to be initiated and enhanced by topographical features which collectively effect viscosities and in turn cause flow speed differences, ultimately allowing hotter parts of flows to propagate more efficiently than cooler parts. Ultimately, such flow perturbations are thought to grow by positive feed-back of which the final product is one in which lava flow is confined to channels. Dimroth *et al.* (1978) also associate the emergence of channelised flows with the increased gradients of shield (volcano) settings.

The 3.5 km thick Komati Formation is characterized by an extraordinarily high proportion of massive to pillowed flows. A large proportion of these ($\sim 35\%$), however, are komatiites with MgO contents greater than 24 wt%, which are known to form spinifex-textured flows but not pillows. The other lavas in the area do form pillows (Viljoen and Viljoen, 1969c). Of these the Geluk-type komatiites and other high-MgO komatiitic basalts probably form pillows less readily than their more viscous low-MgO counterparts. In the study area $\leq 5\%$ of all komatiitic basalts or alternatively $\leq 10\%$ of all the low-MgO komatiitic flows are pillowed, implying very high massive- to pillow flow ratio's which, in turn, imply high extrusion values.

The komatiites that are interleaved with the low-MgO komatiitic basalts seem to mimic this high extrusion rate attribute. Spinifex-textured flows are thought to represent the thin-flow facies of large lava flow-systems (Barnes *et al.*, 1988). The spinifex-textured flows occur peripheral to the thick massive flows, implying that they probably lose heat more rapidly and as a result have lower flow-rates due to increased melt viscosity. The loss of heat leads also to olivine supersaturation and the formation of spectacular spinifex textures (Donaldson, 1983; Barnes *et al.*, 1988). High and low abundances of spinifex-textured flows, therefore, indicate low and high extrusion/flow rates. The Komati Formation is well known by local geologists for the sparse occurrence of spinifex-textured komatiite (Anhaeusser, 1991). The high proportion of thick, orthocumulate-textured vs spinifex-textured komatiite thus corroborate the high flow rates of the komatiitic basalts.

The dominance of massive-, over pillowed and spinifex textured flows in the study area undoubtedly points to lava sheet-flooding as the predominant mechanism of lava emplacement, to the exclusion of channelised flow or tube fed emplacements. The low- to very low viscosities of the melts suggest that topographical depressions would have been filled early-on and that paleoslopes would have become very flat. This is supported by the observed thicknesses of the sheet-flood flows which would have been thinner in case of steeper paleoslopes. It is evident, therefore, that the lavas were extruded predominantly as sheet flows on a relatively flat submarine plain or plateau involving large volumes of lava as indicated by the thicknesses of the flows (*cf* maps of Viljoen and Viljoen 1969). Dimroth *et al.* (1975) have interpreted paleoslopes of $\leq 1^\circ$ for basalt plains in Canada. In contrast, the volcanological features common to shield type settings where slopes may attain maximums of 10° , are absent in the study area, e.g. channelised flows flanked by pillows, brecciated pillows and hyalotuffs (Dimroth *et al.* 1978). Drainaway cavities occurring in the upper parts of pillows also attest to emplacement into flat surroundings as is consistently the case for the study area. At mid-ocean ridges where considerable slopes are present pillows are commonly completely drained (*cf* Wells *et al.* 1979) and massive flows show collapse pits and pillar structures (Ballard *et al.*, 1979).

A special type of channelised flow i.e. sinuous rilles (Wilson *et al.*, 1987) caused by large scale outpourings of hot lava which thermally erodes the surfaces they flow on and ultimately lead to the formation of huge lava rivers (Huppert and Sparks, 1985; Barnes *et al.* 1988), appears not to have occurred in the Komati Formation. Although similar looking bodies also described as layered ultramafic intrusive complexes (Anhaeusser, 1983) occur elsewhere in the upper Onverwacht and especially on the northern flank of the BML, they have clearly not formed in the Komati Formation (*cf* maps of Viljoen and Viljoen, 1969; Anhaeusser, 1983). The absence of lava rilles or rivers seems to contradict the inferred high extrusion/flow rates that are indicated for the Komati Formation. Their absence however, may be ascribed to the thermal 'inertness' of the komatiitic lithologies which seems to have withstood thermal erosion unlike the sediments and tholeiites underlying the komatiitic flows of the Yilgarn Block volcanics (Barnes *et al.*, 1988).

The lateral extent of the flows in the Komati Formation are indeterminate due to the Hooggenoeg Fault situated in the centre of the study area (Fig. 1.4). tectonic thinning toward the west (de Wit *et al.* 1983), discontinuous outcrop related to soda-porphyry intrusions toward the east (Fig. 1.3) as well as poor outcrop conditions in the hinge zone of the Onverwacht Anticline. Furthermore, due to the faulted base of the Komati Formation, it is evident that only a limited (upper) section (15 x 3.5 km) of a possibly much larger submarine lava plain is preserved. The continuance, however, of the Hooggenoeg Formation flows across the entire length of the western limb of the Onverwacht anticline of ≥ 15 km, and its ~ 3.5 km thickness (e.g. Viljoen and Viljoen, 1969; Williams and Furnell, 1979), attest to a relatively large platform of lava. The platform architecture corroborates the absence of steep paleoslopes and by implication predicts high extrusion rates which is consistent with the observed dominance of sheet-flooding ; w formation.

2.8.4 Emplacement Mechanisms

The predominance of sheet flows and the virtual absence of channelised flows and the evidence for pillows at the basal contacts of a massive flows (Fig. 2.3) as opposed to massive flows translating upwards into pillows, supports the emplacement model of Baragar (1984)

rather than that of Dimroth *et al.* (1978).

The pillows are thus envisaged to have formed at the leading edge of massive flows where the convex-shaped crust ruptures and lava is fed into several self sealing conduits while the flow front stabilizes. Lava flows from open channel to restricted tube conditions resulting in an overall decrease in flow rate. As a result, the lava level behind the stabilized flow front rises and lava spills over the barrier just to repeat the process farther ahead. Succeeding reoccurrences of tube formation at the flow fronts and consecutive overriding and interference on the advancing lava results in an entangled carpet of lava tubes and the build-up of a lava barrier. The ratio of pillowed to massive lava within flows is thus determined by flow velocity and extrusion rate. A rapidly approaching lava front would be inclined to overtake the propagating tubes forming at its base and pillows would be scant or absent. On the other hand, a near-stationary flow front could build a mound of pillows as the lava slowly rises and spills repeatedly over the growing barrier.

The study area pillows occur as isolated clusters (commonly 2 - 10 m high and 3 - 40 m in lateral extent) at the base of massive flows (Fig. 2.3). It is likely, therefore, that the isolated occurrences of pillows in massive units represent stabilized flow fronts which developed from entangled tube build-up during reductions or brief hiatuses in extrusion and/or flow rate. The ponding of lava by such barriers is probably what gave rise to the massive units containing the ocelli plumes and spirals, as evidenced in Spinifex Stream (Fig. 2.3).

Previous workers have considered the possibility that some of the massive units actually represent shallow intrusive sills (Viljoen and Viljoen, 1969c,d; Viljoen *et al.*, 1983) or dykes (de Wit *et al.*, 1987). However, the random occurrence of pillow clusters within the massive units, also seen as stabilised flow fronts, defines the massive units as extrusive. The consistent facing-directions of pillows and the open-pillow transitions between pillows and massive flows attest to intimate volcanic relationships. De Wit *et al.* (1987a) suggest that the vertically orientated units of the Komati Formation represent a sheeted dyke complex, presently in the right way-up position. According to de Wit *et al.* (1987a) the Komati

Formation forms the type area of the sheeted dyke component of the Jamestown Ophiolite Complex (JOC) pseudostratigraphy. Pillows that occur in the Komati Formation are thought to represent "pillow screens" and others "intrusive pilloids". The pillow screen and intrusive pilloid origin for the pillows conveniently seem to invalidate pillow facing which in turn allows for the interpretation of the rotated flows as sheeted dykes. However, a sheeted dyke complex model for the Komati Formation has to resort to intricate explanations which are not scientifically obvious, and raises a number of questions: (1) why have the pillow lavas consistent facing directions; (2) why are drainage cavities consistently situated on the northern side of pillows; (3) why are flow-top breccias found in a hypabyssal environment; and (4) why should such rocks show extrusive flow structures, typical of modern volcanic environments.

The apparent lack of interflow sediments from the Komati Formation may initially have assisted with the formulation of the sheeted dyke model. Recently however, a 20 cm thick interflow sediment in the lower half of the Komati Formation has been dated by Armstrong *et al.* (1990, Fig. 1). This clastic rock, referred to as the Komati Unit, is the only known sediment to occur within the formation. The presence of the clastic unit within the so-called sheeted dyke complex not only questions the feasibility of such an interpretation but must also cast doubt on the whole JOC pseudostratigraphy concept (de Wit *et al.*, 1992) as will be shown in Chapter 5.

2.8.5 Depth of Emplacement

The degree and style of vesiculation of the pillows from the lower half of the Komati Formation have enabled assessment of the minimum and maximum ocean depths of emplacement. Based on composition, upper and lower initial volatilities for komatiitic basalt were selected using the intermediate and lower volatility values of modern MORB. The emplacement depths indicated for the pillows using the upper volatility values yielded depths of 4 and 3 km for the stratigraphic base and middle portion of the Komati Formation. The lower volatility values give depths of 1.8 and 1.3 km for the base and middle part of the Komati Formation. By employing rates of isostatic adjustment the feasibility of the estimates

can be tested. The higher and lower initial volatilities indicate ~40% and ~70% isostatic compensation. A maximum rate of 100%, for example, implies constant ocean depth as the rate of sagging (isostatic compensation) keeps pace with the rate of volcanic loading. The indicated values, however, suggest that shoaling occurred as is evidenced by the marginal increase in vesiculation of the upper pillows. Upward shoaling is also supported by the presence of shallow-water sediments at the stratigraphic top of the Komati Formation (Lanier and Lowe, 1982) and the well vesiculated (but not scoriaceous) pillows that overlie them (Fig. 2.7D). Such shallow water conditions, however, cannot be achieved when modelling the high initial volatility values. This option implies an emplacement depth of 4 km, deposition of the 3.5 km thick Komati Formation of which ~40% of the thickness is isostatically compensated and therefore, a final ocean depth of 1.9 km. The low volatility value option implies an initial emplacement depth of 1.8 km, deposition of the 3.5 km thick Komati Formation of which ~70% of the thickness is isostatically compensated and, therefore, a final ocean depth of 0.75 km. An ocean depth of 150 m, however, is in good agreement with the previously mentioned 700 m depth indicated for the pillows overlying the Middle Marker (Fig. 2.7D), implying the low volatility value option as the most viable choice.

2.8.6 Emplacement Environment

The emplacement environment of the Komati and Hooggenoeg Formations is constrained by various volcanic and petrographic data which have a definite bearing on the tectonic setting of the volcanism. They are:

- 1) Volcanism appears to have commenced in a ≤ 2 km deep marine environment for the present faulted base of the Komati Formation and gradually shoaled upwards as indicated by the vesiculation of the stratigraphically intermediate pillows and shallow water sediments of the Middle Marker. The somewhat deeper water conditions (≤ 700 m) indicated by the pillows that overlie the Middle Marker seem to suggest that ocean depth may have fluctuated between periods of (volcanic) quiescence sedimentation and active volcanism. Episodic drowning of the lava platform during Hooggenoeg Formation time is implied.

- 2) The volcanics of the Komati and Hooggenoeg Formation appear to have been extruded onto a relatively flat lava plain with slopes $\leq 1^\circ$. This is evidenced by the absence of paleoslope indicators such as completely drained pillows and massive flows, channelised lava flows flanked by pillow breccias and rift-related scarp breccias. In contrast, the study area lavas have relatively consistent thicknesses resulting in volcanic layering extending throughout the entire length of the western limb of the Onverwacht Anticline (~15 km).
- 3) The lavas appear to have been buried relatively rapidly due to near-complete isostatic adjustment. This was probably around 70% during Komati Formation times and around 100% during Hooggenoeg times as seen by the more or less perpetuated shallow water conditions. (The estimates do not include the Buck Ridge volcanics at the top of the Hooggenoeg Formation for reasons explained earlier).
- 4) Compared to mid-ocean ridge settings the structural make-up of the study area volcanics are distinctly different. The lava plain setting of the study area is inconsistent with the ridge topographies of both fast- and slow-spreading centres. The moderate to steep slopes and relatively low extrusion rates found at mid-ocean ridge spreading centres (Ballard *et al.*, 1979; Van Andel 1983) collectively give rise to a characteristic suite of flow structures which have not been identified in the study area (i.e. the above mentioned collapse pits and pillar structures, the extensive development of channelised flows, hollow pillows, pillow and scarp breccias and high pillowed-to-massive flow ratio's). MOR-settings are typified by extrusive sequences up to 500 m thick whereas those of the Komati and Hooggenoeg Formations are each 3.5 Km thick. The Komati and Hooggenoeg Formations have shoaled with continued volcanism, probably as a result of the isostatic compensation of the thickened oceanic crust which may have included a significant hypabyssal component which seems to have been faulted out. In contrast, MOR crust usually drowns subsequent to formation due to the thermal cooling of the underlying lithosphere.
- 5) The high ratio of massive- to pillowed flows suggest mostly sheet-flood emplacement of lavas and to maintain the observed flow thicknesses, high flow rates are suggested. Such high-flow rates imply that volcanism occurred in a high heat-flow setting.

Compared with MORs, which are characterised by intermittent and unfocused volcanism both spatially and temporally (Ballard *et al.*, 1979), it is concluded that heatflow in the study area must have been much greater.

2.8.7 Komatiite Geochemistry

Smith (1980) identified two suites of komatiite, namely Group I and Group II which are geochemically distinct and originate from melts of different mantle sources. Recent geochemical studies have shown that Barberton-type komatiite melt (i.e. low Al_2O_3 and depleted HREE) formed by moderate degrees of mantle melting at depths greater than 15 Gpa, i.e. below 400 km (Arndt and Lesher, 1992). The komatiites are thought to have risen to the surface by means of mantle plume action (Jarvis and Campbell, 1983; Campbell *et al.*, 1989). Campbell *et al.* (1989) postulate that the geochemically distinct basalts and komatiites are produced by thermal plumes in the mantle. Fluid dynamics show that plumes embody a hot axial stream capped by a mushrooming head into which cooler mantle is entrained. The komatiites are thought to form in the high-temperature axial jet of the plume and the basalts by melting in the cooler heads. The sudden onset and short-lived duration of volcanism in the Western Australian greenstone belts is explained by this model (Campbell *et al.*, 1989). Standard mid-ocean ridge crust, on the other hand, has been modeled in terms of decompressional melting of mantle with potential temperatures of 1280 °C and 1350 °C (McKenzie and Bickle, 1988; Chen, 1990), implying depths of 40 and 80 km respectively. Only unimodal basic melts (MORB) are produced.

2.8.8 The Onverwacht Oceanic Plateau

The lava plain aspect of the Komati and Hooggenoeg Formations (each comprising effusive thicknesses of ca. 3.5 km), the high indicated flow rates, the gradual upward shoaling of the sequences and the quenched characteristics of the effusives corroborate the volcanic make-up of oceanic plateau rather than standard MOR-crust. An oceanic plateau origin for the studied rocks realistically explains many of the volcanological characteristics of the Komati and Hooggenoeg Formations as outlined above. The possibility that the Onverwacht rocks in question represent an allochthonous relict of an ancient oceanic plateau is supported on

a much broader geological base than studied here. Before such comparisons can be made, a discussion on the nature and origin of these relatively poorly understood volcanic phenomena appears appropriate.

Oceanic plateaus are anomalously thick volcanic piles typically elevated to between 2 - 3 km above the ocean floor and consist of oceanic crust ranging in thickness from 20 - 40 km, which is two to five times thicker than standard oceanic crust (Hussong *et al.*, 1979; Klaver, 1987 cited in Storey *et al.*, 1991). Due to their large crustal thickness oceanic plateaus are relatively buoyant and are inclined to resist block subduction, as is thought to be the case

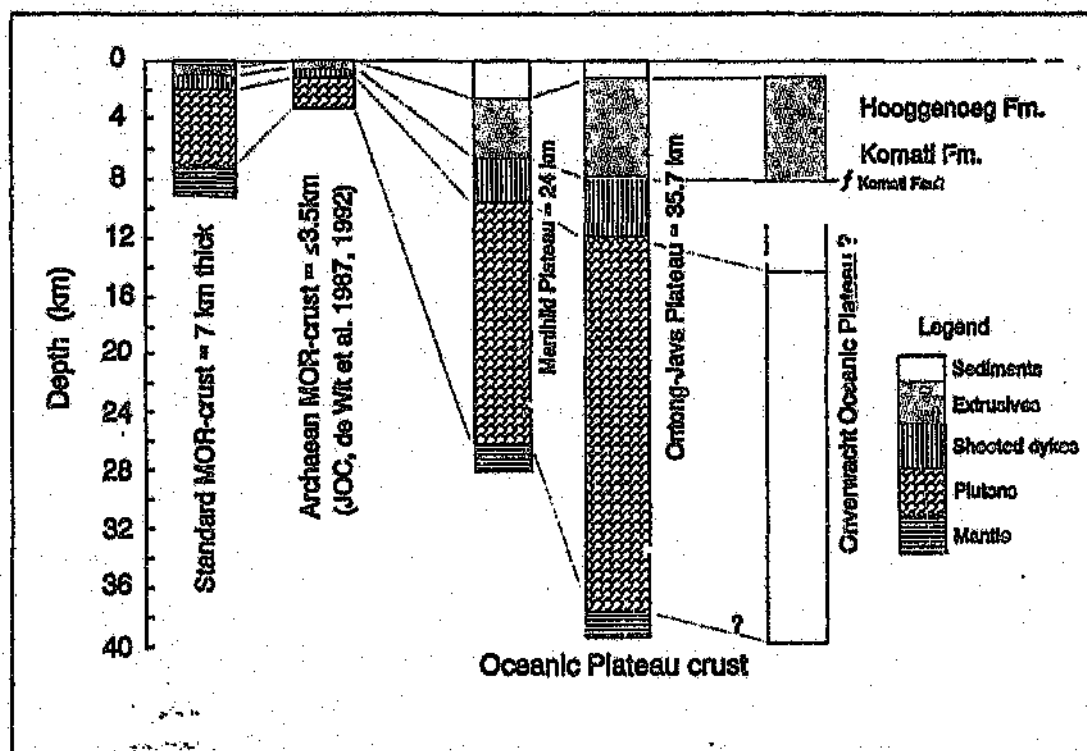


Fig. 2.13 Schematic representation of various oceanic crustal profiles versus those proposed for the Onverwacht Group or parts thereof. (See text for discussion)

for the Ontong-Java and Caribbean Plateaus (e.g. Storey *et al.*, 1991 and references therein). They are often covered by relatively shallow water and seldom rise to above sea level (e.g. Kerguelen island) and, therefore, are difficult to sample. Oceanic plateaus comprise an

estimated 3% of the Earth's ocean floor (Sandwell and MacKenzie, 1989) of which the largest, the Ontong Java Plateau, spans 1,500,000 km² (nearly the total land surface of South Africa). They tend to have weak or absent linear magnetic anomalies (Mahoney and Spencer, 1991). It is thought that they form cataclysmically when the bloated heads of emerging mantle plumes invade the base of the lithosphere, whereas the sometimes associated seamount chains or extended aseismic ridges reflect the much narrower but long-lived plume tails that follow (Richards *et al.*, 1989).

Recently, predictions concerning rapid emplacement seem to have been confirmed as it has been shown that most of the Ontong Java Plateau formed, within a time span of 3 million years (Tarduno *et al.*, 1991). The massive outpouring of lava is inferred to be several times larger than that of the Deccan Traps of India. The magma emplacement rates required to produce the Ontong Java Plateau in 3 Ma (conservatively estimated) range from 8 - 22 km³/year, and are compatible with an origin requiring the incursion at the base of the oceanic lithosphere of the head of a large mantle plume. This volcanic event is considered to be the largest of the past 200 Ma.

Storey *et al.* (1991) have also considered oceanic plateaus to be sites of komatiite formation. They suggested that the middle Cretaceous komatiites of Gorgona Island, which have MgO contents up to 24%, are part of an obducted fragment of the Caribbean oceanic plateau. Plate reconstructions show that the Caribbean Plateau is related to the Galapagos hotspot and that the different isotopic compositions of the Gorgona komatiites and tholeiites require the simultaneous tapping of two different mantle sources. The coeval tapping of two mantle sources is effectively explained by the mantle plume hypothesis (e.g. Campbell *et al.*, 1989; Griffiths and Campbell, 1990). These workers also contend that some greenstone komatiite-tholeiite sequences are remnants of hotspot-related Archaean oceanic plateaus of which the upper portions were delaminated and obducted over continental crust at ancient convergent margins.

2.9 Summary

The Komati and Hooggenoeg Formations show good correspondence with the known characteristics of oceanic plateau, namely volcanic structure, lithology, sedimentology, geochemistry and rapid emplacement history, and it is proposed that they collectively

represent the upper portion of an ancient oceanic plateau, hereinafter referred to as the Onverwacht Oceanic Plateau (OOP). Evidence corroborative of an OOP is as follows (summarised in Table 2.6):

- 1) Comparison of the internal structure of normal oceanic crust and oceanic plateau shows that the extruded component (layers 2a + 2b) of normal crust is distinctly less than 1 km (e.g. Alt *et al.*, 1986) whereas the equivalent portion of an oceanic plateau is considerably thicker, namely ≤ 6.5 km (Hussong *et al.*, 1979). The thickness of the corresponding segment of the OOP crust, consisting of the Komati and Hooggenoeg Formations (6.5 - 7.0 km) is in good agreement with those of the Ontong-Java and Manihiki Plateaus (Fig. 2.13, Table 2.6). The Middle Marker separates the formations and has traditionally been seen to represent an important break during Onverwacht volcanism (Viljoen and Viljoen, 1969d). The stratigraphic significance of the Middle Marker has been questioned subsequent to the discovery by Williams and Furnell (1979) of a 1 km thick komatiitic basalt horizon within the Hooggenoeg Formation. These authors maintain that the broad scale alternation of komatiitic and tholeiitic rock made it similar to many other komatiitic greenstone sequences elsewhere and that the Middle Marker therefore does not indicate a major change in the origin and nature of melts. Subsequent to the discovery of the komatiitic horizon the effects of hydrothermal silicification were also recognised resulting in the negation of the diagnostic "basic to felsic volcanic cycles" of the Hooggenoeg Formation. Although the Hooggenoeg Formation volcanics are collectively less mafic than those of the Komati Formation, distinctions between the two on a qualitative basis are small. Similarities include their komatiitic affinity, bimodal maficity, volcanic architecture, high inferred extrusion rates, lack of associated felsic volcanics and sediments, a common palaeomagnetic pole and short-lived emplacement. The features are reconcilable with a plume origin. The fact that they are separated by the Middle Marker chert horizon cannot exclude the possibility that both formations are related to the same mantle-plume melting event. This would imply that the Middle Marker does not mark a major change in mantle melting as suggested by Williams and Furnell (1979). The Middle Marker could therefore be seen as having formed during a short-lived hiatus in a broader period of plume related volcanism. The periods of volcanic quiescence, which progressively increase in frequency upwards, may also be seen to mark the waning of OOP volcanism (i.e. H1, H2, H3, H4 and

Table 2.6 A summary of characteristic volcanological features of the OOP, modern MOR-crust and oceanic plateau.

Volcanological feature	Mid-ocean ridge	Oceanic Plateau	Onverwacht Oceanic Plateau
Crustal structure (seismic stratigraphy) Layer: 1 - sediments 2a + 2b - pillowed and massive flow units 2c - sheeted dyke complex 3 - gabbroic magma chambers Total crustal thickness	Modern MORB ≤ 0.3 km ⁽¹⁾ ± 0.6 km ⁽²⁾ ± 1.0 km ⁽²⁾ ≤ 5.0 km ⁽²⁾ ~ 6.9 km Archaeon MOR-crust (JOC ⁽²⁰⁾) 0.3 km (see text for explanation) 0.5 km 2.53 km <3.5 km	OJP ⁽⁴⁾ 1.2 km 6.5 km 5.0 km 23.0 km 35.7 km Manihiki Plateau ⁽⁴⁾ 2.7 km 4.0 km 3.1 km 14.3 km 24.1 km	(not preserved) ~6.5 km Hooggenoeg + Komati Fm. ⁽⁵⁾ faulted out of section ⁽⁵⁾
Extrusive rock type(s)	Unimodal effusive suite: MORB	Bimodal effusive suite: OJP ⁽⁴⁾ OJB + MORB Gorgona ⁽⁷⁾ KOMT + THOL	Bimodal effusive suite: Hooggenoeg Fm. ⁽⁵⁾ THOL + KBAS Komati Fm. KBAS + KOMT
Extrusion rates and style	Relatively low; outflows are sporadic, short-lived and widely separated in space and time ⁽⁶⁾ . Extrusion rates vary from slow to fast spreading centres but individual flow units are restricted to the axial rift valley and range up to tens of km in length ⁽⁹⁾ .	Very high (cataclysmic); very extensive eruptions which are highly focused in space and time (~3 Ma for OJP). Flows cover large surfaces: i.e. OJ, Manihiki, and Kerguelen plateaus exceed several hundreds of km in both X and Y dimension ^(6,10) .	High flow rates are indicated by; (1) single, ~200 m thick ⁽⁵⁾ low viscosity komatiite flows, (2) the dominance of sheet-flooding over pillowed flows (of all pillowable lava $\leq 5\%$ is pillowed), and (3) a lack of interflow sedimentation. Period of Komati Fm. volcanicity < 10 Ma ⁽¹⁸⁾ .
Extrusive flow structures	Pillow lavas dominate at both slow and fast spreading centres, although at the latter massive flows become an important component ⁽⁶⁾ . Drainage cavities commonly occur in both flow forms: pillows are commonly hollow or collapsed ⁽¹²⁾ and massive flows comprise collapse pits + pillars ⁽⁸⁾ . Pillows overlie massive flows in individual cycles.	Massive flows dominate pillowed flows. Relatively little is known about the volcanology of oceanic plateaus due to an access problem. A few drill holes have penetrated the tops of some plateaus and confirm the mentioned relationship ⁽¹³⁾ .	Massive flows dominate pillowed units. Pillow clusters occur as isolated patches within the massive flows of which the contacts are volcanological, i.e. typified by open pillows with re-entrant crusts. Pillows underlie massive flows. Drainage cavities are restricted to the tops of pillows suggesting low paleoslopes.
Origin of melt	Decompressional melting of upwelling mantle with potential temperature of 1280 °C ⁽¹⁴⁾ or 1350 °C ⁽¹⁵⁾ ; corresponding to ~depths of 40 and 80 km resp. Normal Archaeon mantle is considered to have been only marginally (50-100°C) hotter than at present ⁽¹⁷⁾ .	Geochemical studies indicate hot-spot and deep-seated mantle plume origins ^(6,7) .	Geochemical studies suggest a deep, hot mantle plume melt origin initiated at a depth ≥ 400 km ⁽¹⁶⁾ .
References and Abbreviations	(1) Anderson et al., 1982; (2) Alt et al., 1986; (3) Brown et al., 1991; (4) Hussong et al., 1979; (5) Williams and Furnell, 1979; (6) Mahoney and Spencer, 1991; (7) Storey et al., 1991; (8) Ballard et al., 1979; (9) Ballard and Francheteau, 1983; (10) Tarduno et al., 1991; (11) Bradley and Frey, 1988; (12) Wells et al., 1979; (13) Kroenke et al., 1990; (14) McKenzie and Bickle, 1988; (15) Chen, 1990; (16) Arndt and Lesher, 1992; (17) Campbell and Griffiths, 1992; (18) Armstrong et al., 1990 - both the Komati (clastic) unit within the lower half of the Komati formation + Middle Marker = 3472 ± 5 Ma. THOL = tholeiite; OJB = ocean island basalt; (19,20) de Wit et al., 1987a, 1992. MORB = mid-ocean ridge basalt; KOMT = komatiite; KBAS = komatiitic basalt; OJP = Ontong Java Plateau; JOC = Jamestown Ophiolite Complex.		

H5 of Lowe, 1991, also represented in Fig. 1.2). The somewhat reduced maficity of the Hooggenoeg Formation relative to the Komati Formation may be attributed to the waning of the mantle plume due to continued entrainment of cooler mantle which explains the larger proportions of tholeiite vs komatiite in the upper part of the OOP.

- 2) Extrusion rates at MOR centres are irreconcilable with those envisaged for the OOP. Ballard *et al.* (1979) have documented lava eruptions at ridges to be sporadic, short-lived and widely separated in space and time (Table 2.6). In contrast, volcanological data indicate that the Komati and Hooggenoeg formation effusive rates had to be high to maintain the degree of submarine sheet-flooding that is evident. The relatively short time-span of Komati and Hooggenoeg Formation emplacement suggests that volcanism may have been near-continuous and highly focused in space and time. Such high extrusion rates are also implied for Phanerozoic oceanic plateaus where cataclysmic eruptions have been predicted for the OJP which was emplaced within a span of 3 Ma (Tarduno, *et al.*, 1991). Such flow rates do not seem inconsistent with those conceived for the formation of the OOP. The 3 Ma emplacement span determined for the Ontong Java Plateau is comparable to the less well constrained 5 - 25 Ma period predicted for the Komati and Hooggenoeg Formations. In the light of the relatively short time spans necessary for plume related volcanism to be completed (i.e. 3 Ma and 15 Ma for the OJP and Kambalda volcanics respectively, e.g. Tarduno *et al.*, 1991; Campbell *et al.*, 1989) it would not be surprising if the Hooggenoeg Formation emplacement span is reduced by future precise dating to one similar to the ~5 Ma period of the Komati Formation.
- 3) Comparison of the effusive flow structures of normal oceanic crust with that of the OOP shows a very low degree of correspondence (Table 2.6). Relatively little is known about the volcanology of oceanic plateau due to a general access problem. A few drill holes, however, have sampled the uppermost flows of some plateau (Kroenke *et al.*, 1990). The results show that massive flows significantly dominate pillowed flows, as is the case for the OOP. The relatively simple volcanological make-up of the OOP seems to be the direct result of high discharge rates, which has led to the formation of a flow, lava plain-like environment in which komatiitic and tholeiitic basalt sheet-flooding precluded extensive pillow formation. The preponderance of thick, orthocumulitic komatiite over spinifex-textured komatiite is

consistent with the supposed high discharge rates.

- 4) Comparison of melt origin (i.e. depth) between MOR crust and that of the OOP also shows low degrees of correspondence (Table 2.6). The origins of the unimodal (tholeiitic) melts that comprise MOR crust are considered to be generated at shallow depths, namely between 40 and 80 km (McKenzie and Bickle, 1988; Chen, 1990). In contrast, geochemical studies show that Barberton-type komatiites are most likely to have formed due to hot mantle plume melting at depths in excess of 400 Km (Arndt and Leshar, 1992). The envisaged plume mechanism readily explains the bimodal mafic volcanicity of the OOP (e.g. Campbell *et al.*, 1989; Campbell and Griffiths, 1992) a feature which is not accomplished by MOR-related decompressional melting.

2.10 Conclusions

The important conclusions made in this chapter concerning the volcanology of the Komati and Hooggenoeg Formations can be summarised as follows:

- 1) The general morphology, structure and volcanological characteristics of the studied lava flows are not akin to typical oceanic crust generated at a MOR-systems as suggested by previous workers, e.g. de Wit *et al.* (1987a, 1992).
- 2) The Komati and Hooggenoeg Formations comprise a coherent volcanological entity which was most probably the result of an ancient mantle plume impacting on the ocean lithosphere at about 3472 ± 5 Ma. This lead to the formation of the Onverwacht Oceanic Plateau probably within a period of ≤ 15 Ma.
- 3) The faulted basal contact of the OOP, a 800 m wide shear zone, suggests that only the upper part of the OOP was allochthonously emplaced and delaminated from its basal part. The large felsic, subvolcanic sill at the top of the Hooggenoeg Formation is seen to have been emplaced shortly after the allochthonous emplacement of the OOP at 3445 ± 8 Ma.

A more holistic and detailed account of the geological relationships of the OOP to the under- and overlying sequences of the Barberton greenstone belt is given in Chapter 5, subsequent to the presentation of more data.

3. SEAFLOOR METAMORPHISM

3.1 Introduction

The petrography and geochemistry of the metavolcanics of the Komati Formation have been described by several workers (e.g. Viljoen and Viljoen, 1969 b, c, d, e; Williams and Furnell, 1979; Smith *et al.*, 1980; Viljoen *et al.*, 1983; de Wit *et al.*, 1987). These accounts focus mainly on the igneous petrogenesis of the komatiites and to a lesser degree on the effects of alteration to assess the usefulness of certain elements for classification and modelling (Viljoen and Viljoen, 1969b; Smith, 1980; Condie *et al.*, 1977). Apart from their igneous characteristics, other features such as preservation of texture, mode of hydration and Mg-uptake (Cloete, 1986; de Wit *et al.*, 1987) and oxygen isotope compositions (Smith *et al.*, 1984; Hoffman *et al.*, 1986) have lead to suggestions that the rocks of the study area have interacted with Archaean seawater. The petrography and geochemistry of the Komati Formation metabasites have not yet been evaluated in terms of modern seafloor metamorphic concepts, e.g. alteration petrography, water/rock ratio's, elemental fluxes and sulphur isotope evolution.

Seafloor metamorphism is best preserved in the statically metamorphosed parts of the study area, referred to as Zone A and B Chapter 4 (Fig. 4.1). All seafloor-metamorphosed rocks have also undergone burial metamorphism and those to the west (Zone D, Fig. 4.1), additional dynamic metamorphism. Seafloor metamorphism is well preserved in the statically metamorphosed area due to the isochemical and hydrostatic stress environment of burial metamorphism. Seafloor metamorphism is thus cryptically overprinted by burial metamorphism, only affecting mineral but not whole rock compositions. This allows seafloor metamorphism to be studied by means of chemical and petrographic techniques. Evidence for burial metamorphism, based on mineral chemistry and metamorphic assemblages, will be discussed later in Chapter 4.

The aim of this chapter is to document the alteration petrography, alteration geochemistry and sulphur isotope evolution of the seafloor metamorphosed komatiitic pillow basalts. For the evaluation and interpretation of the data much reliance is placed on comparison with modern seawater-basalt reaction studies. These studies, however, have focused mainly on simulating mid-ocean ridge (MOR) sub-seafloor conditions,

whereas in the previous chapter it was concluded that the Komati and Hooggenoeg Formations appear not to have formed at such a tectonic setting. The basic processes involved in hot-seawater-basalt interaction are considered universal and not restricted to MORs only and the differences in tectonic setting may be reflected in the style and development of mineral parageneses. It is also the aim of this chapter to reconstruct the development of convective flow in the Komati and Hooggenoeg Formations.

3.1.1 Review of some concepts of seafloor metamorphism

Ocean-floor metamorphism has, until quite recently, been poorly understood. This is evident from the debate surrounding the so-called 'spilite problem' (Vallance, 1960, 1965; Cann, 1969; Floyd, 1976). The main arguments were, whether spilites (ie. seafloor metamorphosed basalts) were derived from special spilitic magmas, autometamorphism or reaction of cooling basalt with seawater. Miyashiro *et al.*, (1971) noticed some distinct aspects of mid-ocean ridge metamorphism and called for the recognition of a new type of metamorphism. As the metamorphism at mid-ocean ridges occurred on a grand scale it could not be called regional metamorphism as it was not associated with regional schistosity (as is distinctly the association with orogenic belts). This type of regional metamorphism was later to become known as seafloor metamorphism.

The importance of hydrothermal fluid circulation in the ocean crust was recognised by the occurrence of manganese- and iron-oxide nodules on the seabed (e.g. Rona, 1973; Scott *et al.*, 1974; Thompson *et al.*, 1975) and the first candid observations of active hydrothermal systems were made on the Galapagos Ridge in 1977 (Corliss *et al.*, 1979). The first geochemical constraints on high-temperature seawater-basalt interaction were established by means of elemental flux determinations shortly thereafter (Humphries and Thompson, 1978).

Two broad types of seafloor metamorphism are recognised: ie. low temperature and high temperature seafloor alteration (Scott and Hajash, 1976; Thompson, 1983). Low-temperature alteration includes ocean-floor weathering at bottom temperatures of 0° - 3°C, through diagenesis and sub-seafloor interactions up to 100°C. The high temperature type clearly manifests itself only above 150°C. Low-temperature alteration is referred to also as seafloor weathering, seafloor oxidation and halmyrolysis, which implies

palagonitisation and smectite formation (Staudigel *et al.*, 1981), and all of which lead to the formation of brownstones. High temperature alteration, on the other hand, is synonymously used with seafloor metamorphism, ocean-floor metamorphism, hydrothermal metamorphism, seafloor alteration, and seafloor metasomatism. The alteration of the Komati Formation metabasites essentially fall within the high-temperature type.

Sub-seafloor metamorphism however, is inaccessible to direct observation and is studied indirectly by three methods, (a) direct sampling of hydrothermal vent fluids along ocean-ridge systems, (b) experimental simulation of the sub-seafloor hydrothermal systems in the laboratory and (c) sampling of oceanic crust by either dredging fault scarps or drilling.

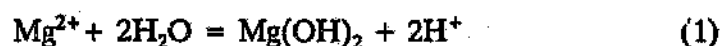
3.1.1.1 *Sampling of hydrothermal vents* along ocean ridge systems undoubtedly confirms the far reaching effects of heat- and mass-exchange mechanisms related to high-heatflow convective circulation (Von Damm *et al.*, 1985; Von Damm and Bischoff, 1988; Edmond *et al.*, 1986). In contrast to fresh seawater, the chemistry of venting hydrothermal fluids is characteristically depleted in MgO and SO_4^{2-} , enriched in H^+ , SiO_2 , H_2S and contains a wide diversity of transition metals especially Fe, Mn, Zn, Cu (Von Damm *et al.*, 1985) and dissolved gasses (Craig, 1980). These fluids vent at up to 350° and 400°C (Von Damm *et al.*, 1985; Delaney *et al.*, 1984), are reducing and are thought to be buffered by pyrite-magnetite $\pm$ pyrrhotite equilibria at subsurface conditions (Janecky and Seyfried, 1984; Janecky *et al.*, 1986). They are clearly much evolved with regard to the chemical, redox and physical state of the precursor fluid.

3.1.1.2 *Experiments simulating seafloor metamorphism* have successfully reproduced the compositions of submarine hot-spring fluids (Seyfried *et al.*, 1988, 1991). These studies, which involve the reaction of powdered tholeiitic rock with heated seawater, take into consideration five primary variables, i.e. temperature, pressure, crystallinity of the starting material, seawater/rock ratio and reaction time-span (Mottl 1983; Seyfried *et al.*, 1988).

An important method used to study the reaction path of the solution vs. that of the rock is the seawater/rock ratio (Seyfried and Bischoff, 1977; Seyfried *et al.*, 1978). In the

experiments this is defined as the starting mass of seawater present in the reaction bomb divided by the mass of fresh rock. For natural hydrothermal systems the water rock/ratio is equated with the mass of seawater which has progressed through the system, integrated through time, divided by the total mass of altered rock within the system (Mottl, 1983). The secondary mineral assemblages of the laboratory experiments were generally found to resemble those found in nature (Mottl, 1983; Alt *et al.*, 1986; Seyfried and Janecky, 1985; Seyfried *et al.*, 1988). Some discrepancies exist, however, which are attributed to kinetics, but the experiments succeed in closely duplicating the solution chemistry of natural systems (Mottl, 1983; Rosenbauer and Bischoff, 1983; Seyfried and Janecky, 1985; Seyfried *et al.*, 1988).

3.1.1.3 *The behaviour of heated seawater.* Striking changes in the composition of seawater occur upon heating. This includes the precipitation of anhydrite at temperatures greater than 150°C and 500 bars, and at a pH buffered by acetic acid (Bischoff and Seyfried, 1978). Bischoff and Dickson (1979) show that at the withdrawal of Mg^{2+} from seawater takes place in the form of Mg^{2+} as a component which is assimilated by the interlayers of secondary silicates (e.g. smectite) leaving H^+ behind in solution which is subsequently consumed by silicate-hydrolysis reactions.



The removal of Mg^{2+} from solution is swift and so is the associated formation of H^+ . At temperatures $\geq 150^\circ C$ and water/rock ratios of ≤ 50 the removal of Mg^{2+} from seawater is complete whereas at higher ratios removal is less complete (Mottl, 1983). Mottl and Seyfried (1980) refer to these as "rock-dominated" and "seawater-dominated" systems respectively.

In rock-dominated instances, where seawater loses all its Mg, a complex assemblage of secondary silicates forms including zeolites, albite, tremolite-actinolite, titanite and epidote (Mottl, 1983). In comparison, seawater dominated conditions are characterised by very high acidity at which not many secondary minerals can exist and accordingly only smectite, chlorite, quartz, hematite and anhydrite are stable. It is also shown that in these systems adequate amounts of dissolved Mg remain to convert the basalt entirely to a Mg-

rich assemblage of secondary phases (probably similar to the hyalospilites of Cann, 1969).

Mottl (1983) demonstrates that the removal of Mg^{2+} from seawater is compensated with regard to the electrical charge in solution, by the leaching of mainly Ca^{2+} from the rock. At 150°C roughly half of the leached Ca^{2+} reacts to form anhydrite whereas the remainder stays in solution.

3.1.1.4 Constraints from seafloor metamorphosed rocks. The mineralogy and geochemistry of the earlier-mentioned 'spilite problem' has been satisfactorily resolved by seawater/basalt experimentation (Seyfried *et al.*, 1978) and serves to illustrate the application of these findings (Mottl, 1983). It is suggested that the inter-pillow cavities represent the hydrothermal channelways for the flow of heated seawater. The juxtaposed pillow rims and interpillow hyaloclastite/breccia would react with larger volumes of heated seawater than would the relatively more restricted pillow interiors. Interaction with MgO-rich heated seawater in a recharge zone, for example, will lead to the chloritisation of pillow rims, a common feature of seafloor metamorphosed pillows (i.e. hyalospilites). Smaller volumes of heated seawater react with the pillow interiors via the thermal contraction cracks and give rise to Na-enrichment and the formation of so-called orthospilites. The development of mineral assemblages as basalt reacts with increasing amounts of heated seawater is schematically shown in Fig. 3.1.

Recent basalt/seawater studies using evolved seawater show that Mg-, Ca-, and Na-fixation reactions result from different physicochemical conditions which are probably linked to specific regions within the submarine geothermal system. Seyfried *et al.* (1988) show that:

- 1) Mg-addition to oceanic basalt is entirely prograde and is presumably restricted to downwelling portions of sub-seafloor convection cells, generating chlorite-quartz-rich alteration assemblages.
- 2) Ca-addition is a prograde reaction which occurs in response to the continued influx of (Mg-depleted) evolved fluids to the hottest parts of the hydrothermal system, increasing their temperature but sustaining moderate pressures, i.e. not more than 450 bar (enhanced by shallow-level magma chambers at spreading centres). In such environments Ca is removed from solution, accompanied by H^+ -

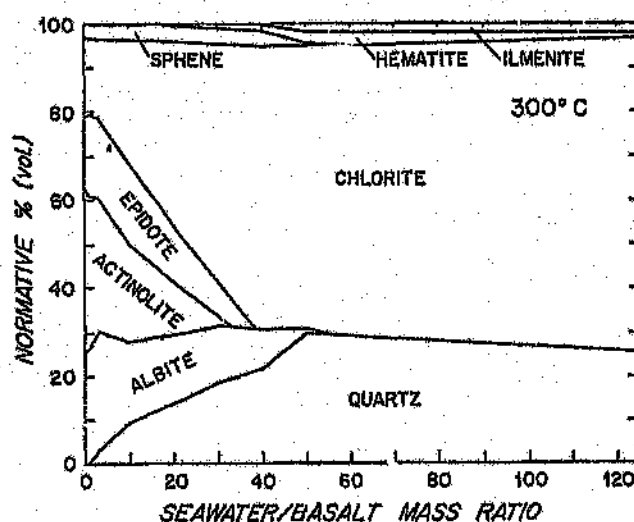


Fig. 3.1.

A model estimating mineral assemblages and proportions produced when basalt reacts with seawater in different water/rock ratios within the greenschist facies (from Mottl 1983). The model is based on experimental data but is close to actual observed assemblages in metabasalts retrieved from the ocean-floor. See Mottl (1983) for details.



Fig. 3.2

A sectioned slab showing partially-flattened pillows from the west part of the study area. Flattening is especially evident from the deformation of the rims of the two smaller pillows. An incipient fabric is developed in the upper part of the larger pillow due to realignment and deformation of ocelli. Vertical and horizontal contraction cracks are shortened and stretched, respectively. Embayments into the pillow rim by the interpillow filling (above the vertical cooling crack of the lower pillow) suggests replacement of the rim. Note the lack of metahyaloclastite. Dark material in the hydrothermal quartz and albite is mostly carbonate (right hand edge of photograph). Locality MC120.

production, high heavy-metal solubility, the formation of clinozoisite—epidote and calcic plagioclase solid-solutions.

- 3) The addition of Na, on the other hand, shows retrograde affinities and is thought to dominate reaction zones affected by upwelling hydrothermal fluids which may cool adiabatically and by conduction.

3.2 Petrography of Pillow Alteration Zones

The effects of seafloor metamorphism can easily be seen in the field on river-polished exposures but at other outcrops these features are obscured by lichens and oxidation. For ease of sample description and detailed geochemical and petrographic sampling, large (15 - 50 kg) pillow samples were collected and sectioned into 8 to 10 cm thick slabs. These surfaces reveal the pillow structure and clearly show the effects of hydrothermal metamorphism in the interpillow areas, adjoining rims and cores. Included in this section are descriptions of pillows that subsequently underwent dynamic metamorphism which were collected west of the statically metamorphosed area (Zone D, Fig. 4.1).

In the following descriptions mineral identifications are based on electron microprobe analyses. (Aspects concerning mineral nomenclature and chemistry, however, will be dealt with in Chapter 4).

3.2.1 Interpillow Cavities

The pillow cavities are always filled with either a darkish, amphibole-rich material which is assumed to be metahyaloclastite (Fig. 2.2C, 2.6D), or white to smoky coloured coarse-grained hydrothermal infilling (Fig. 2.4C). The dark amphibole-rich hyaloclastite zones are usually fine-grained, they have no primary volcanic or shard-like textures and often contain specs of sulphide. The light-coloured hydrothermal material consist predominantly of medium- to coarse-grained milky quartz, white inter-granular albite and light-green epidote. Irregularly-shaped fragments of mafic material occasionally "float" in these zones and appear to be exfoliated and brecciated portions of pillow rim. In one case such fragments were pervasively silicified (MC180). Sulphides are rare. Due to the complex nature and wide range of textures and phase proportions that occur for the interpillow metahyaloclastite and hydrothermal fillings, the petrography has been summarised in Table 3.1.

Table 3.1. Minerals, textures and abundances found in the interpillow domains of pillows of the Komati Formation.

Interpillow infilling	Mineral	Texture	Grain size	Abundance
Matahyaloclastite	Amphibole (actinolite, magnesio-hornblende, pargasite, edenite and cummingtonite).	Equigranular to acicular aggregates.	Fine-grained.	2 - 50%
	Chlorite with dark interference colours.	Xenoblastic and intergranular, dusted with titanite and amphibole micro-inclusions.	Fine-grained.	15 - 40%
	Epidote-clinzoisite.	Idio- to hypidioblastic and occasionally as replacement relicts.	Fine- to medium-grained.	5 - 20%
	Albitic plagioclase.	Clots of untwinned xenoblastic microcrystalline (sugar-textured) aggregates or interstitial hypidioblastic grains.	Fine- to medium-grained.	0 - 5%
	Titanite	Cryptocrystalline exsolutions (dustings) of other minerals and occasional clusters around new growths of xenomorphic limonite.	Cryptocrystalline to fine-grained.	0.1 - 1%
	Sulphides (pyrite, pyrrhotite, chalcopyrite, $\pm$ pentlandite, $\pm$ cobaltite, $\pm$ bornite).	Disseminated and grains rarely occurring in contact with each other.	Fine-grained.	0.1 - 1%
	Minor quartz, microcline, and calcite.	Interspersed and granular.	Fine-grained.	0 - 2%
Hydrothermal material	Quartz	Occurs as microcrystalline patches or encrustations on pillow rim or hyaloclastite surfaces. Spherulitic-textured quartz develops further away from the nucleation surfaces. Large columnar crystals nucleate onto fine-grained substrates and grow into adjacent open spaces with impinging textures (Fig. 3.3).	microcrystalline and spherulitic quartz = fine-grained. Columnar quartz = medium- to coarse-grained (≤ 12 mm).	30 - 80 %
	Albite	Occurs in sugar-textured form encrusting pillow and hyaloclastite surfaces. Also as untwinned xenoblastic grains forming rosettes (Fig. 3.4) and larger hypidioblastic laths occasionally impinging on columnar quartz (Fig. 3.5). Albite commonly shows checker-board twinning (Fig. 3.6).	Granular and rosette albite = fine-grained (≤ 1 mm), laths ≤ 5 mm.	10 - 40%
	Epidote-clinzoisite	Epidote develops as hypidioblastic to ididioblastic porphyroblasts with Fe-poor rims and is commonly clinzoisite mantled. Clinzoisite occurs occasionally as cryptocrystalline fringes to alteration fronts.	Fine- to coarse-grained (0.5 - 8 mm).	1 - 10%
	Amphibole	Acicular to fascicular clots ≤ 10 mm in size representing either altered hyaloclastite relicts or hydrothermal amphibole. No primary features are preserved. The presence or absence of titanite may be the only discriminating factor between a hyaloclastite or hydrothermal origin as TiO_2 is usually considered to be immobile.	Fine-grained.	0 - 5%

The mineralogy and petrography of the dynamically overprinted interpillow fillings is essentially similar to that described above but for the presence of incipient cataclastic textures and associated intergranular recrystallisation of quartz. Hosted by clear albite and quartz are a few large porphyroblasts of late deformation-twinning calcite containing acicular chlorite and chromite, probably representing inert relicts of earlier metahyaloclastite.

3.2.2 Pillow Rims

Pillow rims differ markedly in appearance, both texturally and mineralogically from the cores they envelop. Microscopically, many of the primary textural differences between the once holohyaline rims and hyalocrystalline cores can still be recognised but the once corresponding compositions are now conspicuously different. Three main types of alteration are recognised, namely amphibolitisation/chloritisation, albitisation and silicification (Fig. 2.4C, 2.6D, 3.4A). The pillow rims may show signs of composite alteration consisting of simple permutations of the above mentioned types. Alteration at the rims is usually confined to the once glassy precursors and only rarely is the outer core also involved. The thickness of the alteration zones are thus variable and in most instances determined by the original thickness of glassy selvages which may range between 0 to 3 cm.

3.2.2.1 Chloritised/amphibolitised pillow margins (Example: Pillow MC150, Fig. 3.4A). Mineralogically, the rims consist of xenoblastic chlorite with anomalous black or blue interference colours. The chlorite contains numerous inclusions of micro-crystalline titanite which occur as small glomerocystic patches and stringers which may be branched. Also embedded in the chlorite (matrix) are small (≤ 1 mm) porphyroblasts of randomly orientated fascicular to idioblastic pargasite (Mg and Al-rich amphibole) which is often zoned, sometimes curved and comprise up to 40% of the outer rim. The pargasite cores are dusted with titanite inclusions whereas the outer zones are clear with a distinct blue and green pleochroism. Very fine-grained (0.04 mm) parallel growths of acicular magnesiohornblende occur as vermiform stringers sub-perpendicular to the pillow surfaces (Fig. 3.4B). These vermiform features can be traced inward towards the inner rim where they progressively comprise less amphibole and more plagioclase, until clear plagioclase 'veinlets' with distinct central traces emerge (Fig. 3.4C). Anomalous

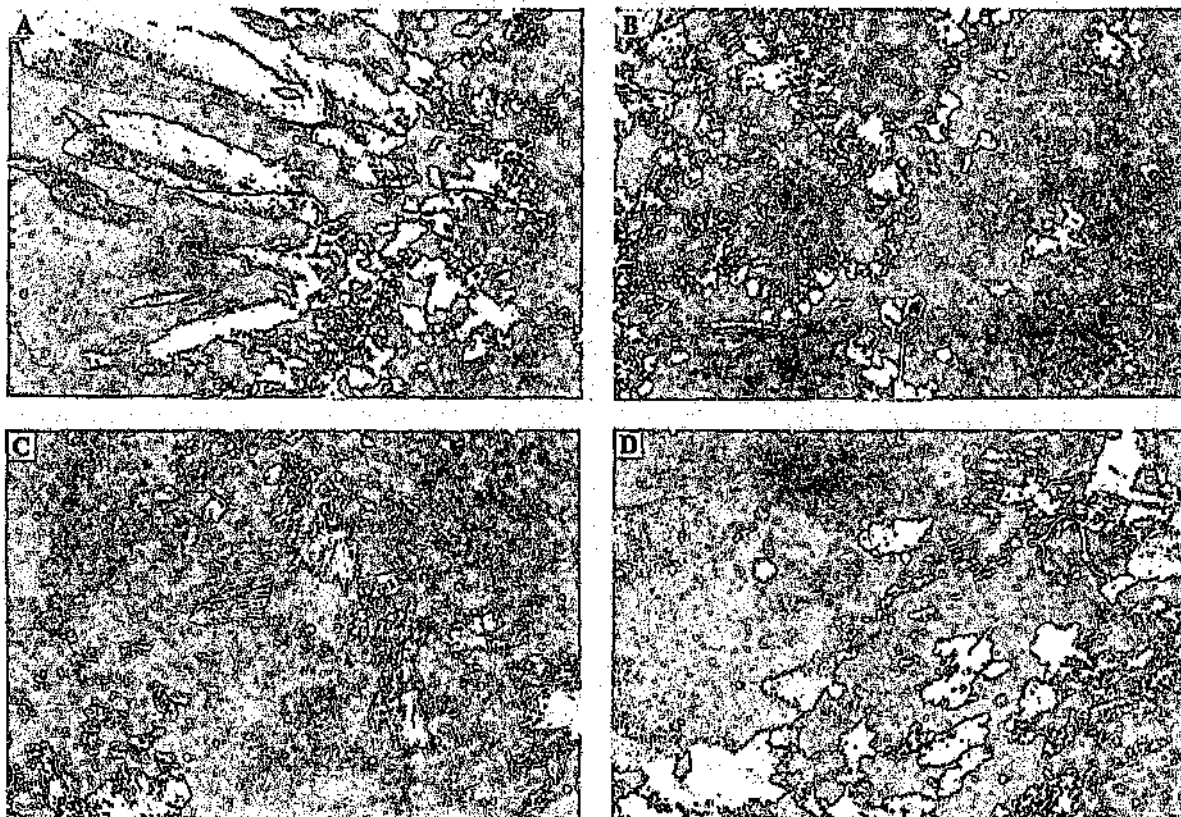


FIGURE 3.3

- A A photomicrograph of typical coarse columnar quartz nucleated onto a finer-grained substrate at the one end and extending into tri-cusate open-space at the other. The fine-grained material encrusts a pillow rim surface. Sample MC160d, Crossed nicols, Field of view = 12mm.
- B A photomicrograph of rosettes of untwinned albite developed around microcrystalline centres of albite. The dark material is anomalously-black chlorite which appears to be replaced by the albite rosettes. Sample MC60f, Crossed nicols, Field of view = 4 mm.
- C A photomicrograph of relatively coarse-grained randomly orientated to radiating hypidioblastic laths of albite impinging upon quartz (dark phases at lower centre). The albite laths are nucleated onto two converging pillow-rim surfaces which meet just above the upper edge of the photograph. The albite grains show checker-board twinning and quartz is idioblastic. Sample: MC180s. Crossed polars. Length of field of view = 21 mm.
- D A photomicrograph of checkerboard albite crosscutting and replacing columnar quartz. The grains have short alternating Albite-Law twins which do not traverse the whole width of grains which results in a checkered appearance. Such albite is thought to be representative of albitisation (Spry 1969). Sample: MC180f, Light; crossed nicols. Scale: length of field of view = 4 mm.

blue clinozoisite may also be observed sporadically as relicts of a former assemblage associated with Na-plagioclase. Small amounts of pyrite and chalcopyrite are scattered throughout the rims and usually occur in association with clinozoisite.

Juxtaposed with the veinlets (at the inner-rim) are dark angular and/or embayed isotropic relicts (Fig. 3.4C). Under high magnification the less isotropic examples show greater degrees of recrystallisation and comprise a fibrous texture. Semiquantitative SEM analyses show that the most isotropic relicts are non-crystalline except for very small ($\leq 0.5 \mu\text{m}$) dispersed grains of titanite. Although these relicts resemble glass, the analyses show them to have undergone a definite change in composition and they are now depleted in mafic constituents. They are probably best described as chemically-modified, isotropic, cryptocrystalline relicts after komatiitic glass.

The above mentioned "micro-veinlets" (Fig. 3.4C) consist of a dark central band made up of microcrystalline titanite and is straddled on either side by selvages of clear albite and/or oligoclase. The traces appear to represent hairline tension cracks of the original komatiitic glass which were infilled with titanite shortly after crack development. During this time the adjacent glass appears to have reacted to form the albitic selvages. The micro-veinlets appear to be preserved due to the stability of the replacing minerals, namely titanite and albitic plagioclase, at greenschist conditions, whereas other glass was devitrified and/or altered at subsequent conditions leading to the formation of chlorite and amphibole. In hand-specimen the albite relicts show-up as a fine white speckled fringe marking the inner part of the chloritised pillow rim (Fig. 3.4A).

3.2.2.2 Albitised pillow rims (Example: Pillow MC60, Fig. 2.6D). Mineralogically the rims consist predominantly ($\leq 85\%$) of un- to poorly twinned, turbid xenoblastic albite. The twins display a checker-board texture. The turbidness is caused by titanite dust which may also form small glomerocrysts and stringers not unlike those described above (Fig. 3.4D). The rims also contain ($\leq 15\%$) small porphyroblasts ($\leq 1 \text{ mm}$) of hypidiodioblastic epidote which are often zoned with iron-poor rims. Small xenoblasts of chlorite ($\leq 5\%$) occur scattered through the rim, as do minor amounts ($\leq 5\%$) of xenoblastic quartz, acicular amphibole and sulphide. Turbid albite is often transformed to clear albite adjacent to tension cracks suggesting that titanite is hydrothermally

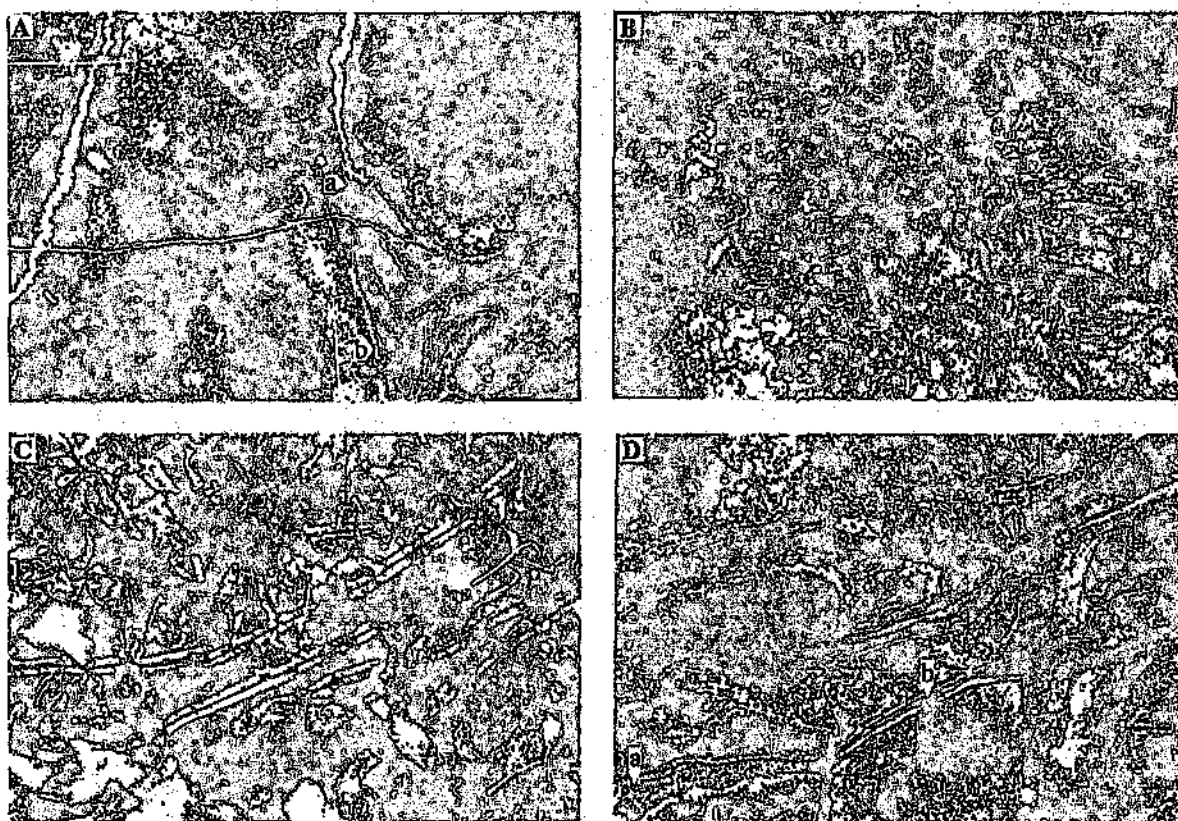


FIGURE 3.4

- A A slab of pillowed komatiitic metabasalt with a chloritised pillow rim, amphibolised cooling-crack selvages juxtaposed to an veinlet filled with albite. Note the termination of the albite veinlet at the chloritised rim (a) and also the whitish dusting at the inner rim (b). A late thin $\pm$ horizontal veinlet of microcline crosscuts all the alteration-related features. Sample: MC150. Field of view = 17 cm.
- B A photomicrograph of vermiform stringers subperpendicular to the pillow surface (top right) consisting of fine sub-parallel growths of magnesiohornblende set in a more transparent matrix of chlorite. Toward the left of centre is a complex arrangement of thin plagioclase veinlets and black, glassy looking relicts of which more detail is presented in the next figure. Sample: MC150r. Plane polarised light. Field of view = 11 mm.
- C A photomicrograph of an inner pillow rim displaying vermiform plagioclase 'veinlets' with distinct central traces consisting of microcrystalline titanite. These traces are thought to represent tensional hairline fractures along which seawater entered and albitised the adjoining glass. Adjacent to some of the albitised selvages dark material is still visible reminiscent of opaque glass which now is cryptocrystalline and has a felsic composition. Sample: MC150r. Plane polarised light. Field of view = 1.7 mm.
- D A photomicrograph of a pervasively albitised pillow rim. Most of the turbidness is caused by titanite dust which also forms microcrystalline glomerocrysts (a) and stringers (b) similar to the hairline tension fractures shown above. The lower hairline fracture is partially replaced by chlorite (bottom left) in which the small sphene glomerocrysts (a) occur as relicts of the central trace (b). Sample: MC60m. Plane polarised light. Field of view = 1.7 mm.

remobilised. Mottled albite (Fig. 3.5A) suggests intermediate stages of titanite removal. The removal of titanite may imply that titanium is mobile during albitisation.

3.2.2.3 *Silicified pillow rims* (Example: Pillow MC180, Fig. 2.4C). Microscopically, silicified rims are compositely altered and consist of a pervasively silicified outer skin up to 10 mm thick and a dark turbid microcrystalline inner-rim of which both inner and outer portions may be replaced by albite and minor clinozoisite (Fig. 3.5B).

The silicified outer-portions comprise up to 98% xenoblastic to spherulitic quartz in which minor amounts of microcrystalline titanite occur as dark traces, marking the original shard- and amygdale outlines. Toward the inner-rim inclusions become more numerous, which impart a turbid appearance to the host (Fig. 3.5B). The murky inner-rim displays in plane polarized light fine zonal segregations of leucoxene reminiscent of Liesegang rings. The individual rings may also be traced into the clear, quartz-rich, outer parts suggesting that the zoning was there prior to silicification (Fig. 3.5C). Botryoidal Liesegang-like zonations have been observed in palagonite of 'freshly' altered MORB and appear to have a complicated chemistry (Scott and Hajash, 1976). Coliform and concentric banding resulting from the hydrolytic devitrification of glass is reported from the Turner-Albright ophiolite (Zierenberg *et al.*, 1988). The zonations may also have originated from the initial devitrification or replacement of the glass via a chemical mechanism similar to that proposed for Liesegang banding by McBirney and Noyes (1967).

Albitic overprinting of the silicified- and murky rim portions are especially noticeable at the tips of exfoliated fragments and shards. Here checker-board albite alteration fronts occur within the original fragment boundaries which are delineated by the seemingly inert titanite encrustations (Fig. 3.5B). Most of the open spaces formed by the tension fractures of the rim are also filled by xenoblastic albite which may have checkers or not. Small cross-cutting veinlets of anomalous blue clinozoisite represent late hydrothermal infilling. The dynamically deformed pillows MC120 and MC20 are only moderately deformed, a feature accentuated by the shape of the rims (Fig. 3.2). These appear to have accommodated most of the strain as they are clearly thinned parallel to the foliation and thickened perpendicular to it, i.e. in the pressure shadows. The mineralogy

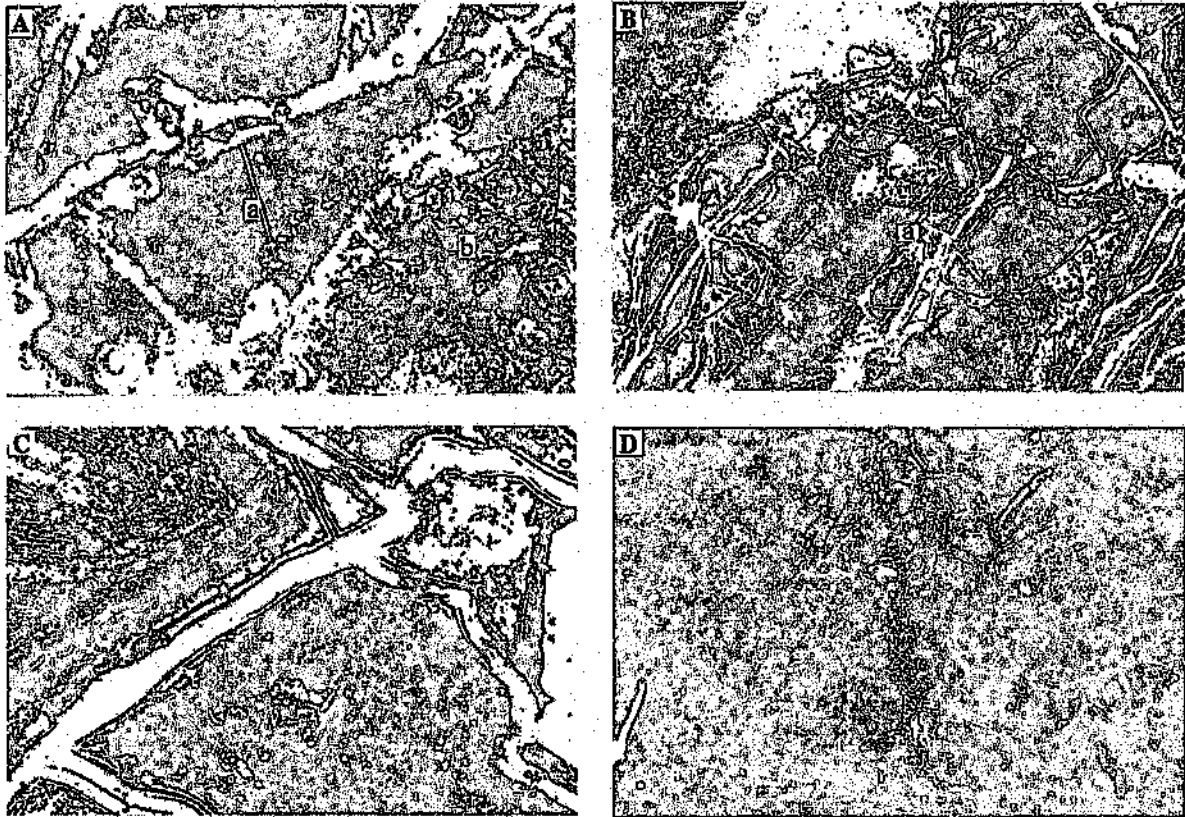


FIGURE 3.5

- A** A photomicrograph of an albitised pillow rim showing very murky albite (dark triangular shard = a), mottled albite (b) and clear albite (c). The turbidness of the albite is caused by microcrystalline titanite. The titanite-free selvages (c) of the hairline fractures suggest that TiO_2 has been mobile. Sample: MC60e. Plane polarized light. Field of view = 3 mm.
- B** A photomicrograph of a pillow rim consisting of a pervasively silicified outer portion and a dark (turbid) microcrystalline inner-part of which both portions are partially replaced, especially at (discoloured = a) fragment tips by albite and lesser clinozoisite. Sample: MC180n. Plane polarized light. Field of view = 16 mm.
- C** A photomicrograph of fine zonal segregations of leucoxene reminiscent of Liesegang zoning, in a silicified pillow rim (top right of Fig. 3.5B). Botryoidal Liesegang-like zonations have been observed in altered MORB (Scott and Hajash 1976). Sample: MC180n. Plane polarized light. Field of view = 4 mm.
- D** The core of a komatiitic pillow basalt showing dark (green) selvages adjacent to an albite filled tension fracture (top right of lens cap). The dark (amphibolitised) selvages overprint incipiently formed ocelli which range from massive to disseminated. Locality: close to the Doornhoek-Laaggenoeg farm boundary on the northern bank of the Komati River (courtesy: Vic King). Lens cap = 6cm.

of the rims of uncarbonated pillows is generally similar to that of the statically metamorphosed ones, but they are distinctly different with regard to texture. Pillow MC120 for example, has chloritised rims similar to the type-area equivalent MC150, but has no primary textures preserved such as the vermiform micro-veinlets, titanite stringers or fine-grained xenoblastic sulphide-associated clinozoisite. There is rather a prevalence of large (1 mm) titanite glomerocrysts, large (2 mm) pyrite idiomorphs and a tendency for chlorite to sustain fabric. The microlitic border zone or outer-core appears as a thin compact granular zone which may be in sheared contact with the rim or interpillow cavity material. In places chlorite of the rim is replaced by albite-rich rim-assemblages akin to the juxtaposed interpillow cavity filling. The granular nature of the albite and quartz-rich rim-assemblages are post seafloor in nature as they appear recrystallised and do not contain the intricate textures described earlier (Fig. 3.5B). Calcite is late as it overprints the imposed tectonic fabric.

The carbonated pillows (MC20) have tectonically flattened shapes. The rims of these pillows are entirely overprinted by calcite, titanite, chlorite, quartz and ilmenite. The new rim assemblage is post-deformational as is evidenced by their fabric-free, felt-textured minerals. Very fine grains of xenoblastic chalcopyrite occur scattered throughout the rim.

3.2.3 Pillow Cores

Pillow interiors are less intensely altered than the enclosing rims. The most conspicuous alteration, however, is confined to the selvages of the radial and concentric cooling cracks which invade pillow cores (Fig. 2.6D, 3.5D, 3.6A, 3.6B). For the purpose of this discussion, pillow-core alteration is subdivided into three parts, namely the cooling-crack veinlets, cooling-crack selvages and pillow core. Distinction between vein infilling and selvage alteration is made as it appears that the alteration processes are not always related. This perception is based on the observation that virtually all cooling-crack veinlets are white in colour and consist mostly of albite whereas three distinct types of alteration selvage are developed. Only half of the observed cases of selvage alteration seems mineralogically related to veinlet infilling, suggesting that selvage alteration is pre- to syn-infilling.

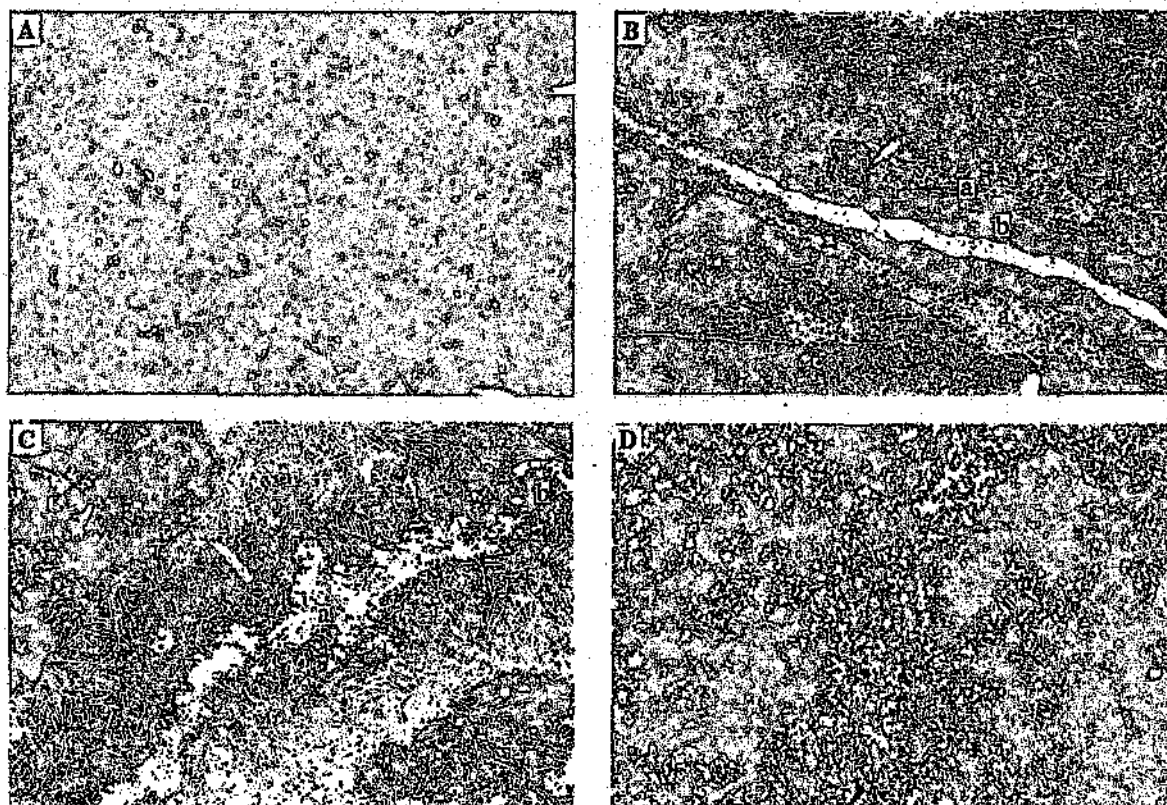


FIGURE 3.6

- A The core of a komatiitic pillow basalt showing light (green) selvages adjacent to albite-filled tension fractures. The light-coloured (albitised) selvages overprint ocelli. Note the veinlet closest to the hammer tip displacing an ocelli (lower down) suggesting that they formed before the start of tension fracturing. Locality: close to the Doornhoek-Laaggenoe farm boundary on the northern bank of the Komati River (courtesy: Vic King). Lens cap = 6 cm.
- B A photomicrograph of a cooling-crack veinlet exhibiting composite alteration. An outer (transparent) amphibolitised zone (a) is replaced by an inner (turbid) albitised zone (b). Note the gradual lateral shift in alteration (amphibolitised wallrock in contact with the veinlet on the left makes way to the right for the albitised wallrock. Sample: MC18^c 20c. Plane polarised light. Field of view = 21 mm.
- C A photomicrograph of a cooling-crack with amphibolitised selvages showing incipient degradation of the microspinel-textured part adjacent to the albite-infilled tension fracture. The light-grey granular phase in the veinlet is epidote (a) and the dark grey murky patches (v) consist of very fine-grained clinozoisite (upper right). The dark grey zones (upper left and lower right) with microspinel show the typical appearance of ocelli. Sample: MC150e. Plane polarised light. Field of view = 12 mm.
- D A photomicrograph of a disrupted cooling-crack veinlet in the core of a dynamically influenced (flattened) pillow. The veinlet is only preserved in the more competent ocelli-phases which also rarely preserve geometrical features reminiscent of microspinel (lower right). No sign of any alteration associated with the veinlet is observed. Sample: MC120c. Plane polarised. Field of view = 3 mm.

3.2.3.1 Cooling crack veinlets. These features of the statically metamorphosed pillows comprise hydrothermally filled tension fractures in the cores of pillows and vary from ~0.05 - 5 mm in thickness. Mineralogically, they consist of (in descending order of abundance) albitic plagioclase, epidote, quartz, $\pm$ sulphide, $\pm$ chlorite, $\pm$ sericite and $\pm$ carbonate. Volumetrically, fine-grained xenoblastic checker-board albite comprises the most abundant phase (70 - 95%) of the veinlets. Both subidioblastic epidote and clinozoisite occur either as individual grains or single zoned crystals, usually with the Fe-poorer clinozoisite at the rims. In places, fascicular arrays of elongate epidote-clinozoisite crystals poikiloblastically enclose fine-grained (≤ 0.3 mm) granular quartz (MC150b). In places fine-grained clinozoisite, forming cloudy turbid patches with a more or less coherent extinction, replaces sections of the albitic veinlets (Fig. 3.6C). Clinozoisite also tends to be more often associated with sulphides than with any other phase of the veinlet assemblage. In some pillows close to the Hooggenoeg Fault and those towards the west of the study area sericite is a common late-stage alteration product. However, in most of the statically metamorphosed area sericite is rarely present and its initial involvement with the filling process is doubted. A few calcite and chlorite-rich veinlets are formed in MC60 which have crosscutting relationships relative to the cooling-crack veinlets suggesting a late origin.

3.2.3.2 Cooling-crack selvages. These are the alteration haloes straddling the cooling-crack veinlets described above, approximately 1-7 mm thick and represent altered pillow core. Megascopically, three distinct types are observed; normal dark-green amphibolitisation, light-green albitisation and light-lime-green silicification (Fig. 2.6D, 3.5D, 3.6A). The silicification alteration type is usually faint and occurs in combination with the light-green albitic type. Simple combinations forming composite dark- and light-green selvages, are also observed (Fig. 3.6B) representing finite overprinting amphibolitised veinlet wallrock.

The cooling crack alteration margins reveal microscopically that alteration has imparted only subtle textural and modal changes to the adjacent pillow core. The dark-green selvages, consisting mostly of amphibole, are similar to the unaffected host except that the igneous texture (usually microspinel) is partly or wholly overprinted by a fine-grained xenoblastic granular habit (Fig. 3.6C). When ocelli patches are overprinted by

this type of alteration the most noticeable changes are that the turbid plagioclase-rich mesostases become progressively amphibolitised and translucent (Fig. 3.6C).

When the light-green albitic selvages overprint non-ocelli-bearing core they become cloudy and murky (at a glance similar to ocelli) (Fig. 3.6B). The igneous features of these zones also become texturally degraded and amphibole is dusted with micro-inclusion replacements of albite. When albitic alteration overprints ocelli-bearing core the mesostases of affected ocelli progressively lose their murky brown appearance and are replaced by clear albitic plagioclase which retains the precursors' geometric habits. This renders the turbid ocelli less cloudy. Amygdales and phenocrysts that fall within the altered selva also become albitised.

The light-lime-green alteration type representing silicification is not often found and seems not to affect the original petrography except that the amygdales and phenocrysts within these selvages are replaced by quartz, suggesting incipient silicification (Fig. 2.6D).

The corresponding structural and textural features of the pillows in the dynamically influenced area are texturally significantly degraded. Deformation of the pillow cores at MC120 is accentuated by the way the radial and concentric cooling-cracks have been shortened and stretched depending on their initial orientation (Fig. 3.2). The associated alteration selvages are no longer visible. In the amphibole-rich portions of the cores the veinlets are in places very thin or discontinuous and are more readily preserved in the competent ocelli-rich phases (Fig. 3.6D). In the carbonated pillows both the cooling-crack veinlets and selvages are totally erased, as a result of both earlier deformation and CO₂-metasomatism.

3.2.3.3 Pillow core. The pillow core comprises the remainder of the pillow, i.e. those portions excluding the compositionally transformed rims, cooling-crack selvages and the microlitic outer-core as defined in Chapter 2. Petrographically, the pillow cores in comparison to the other alteration units are (1) texturally the most faithfully preserved parts of pillows, (2) they show the least amount of replacements, and (3) they have undergone the least degrees of hydration (i.e. between *ca.* 1.5 - 3.0 wt% H₂O). In some instances kernels of original aegirite are preserved, suggesting only incipient alteration and

incomplete hydration. The nature of chemical modification of pillow cores is, therefore, assumed to be isochemical except for the mentioned hydration. This however, is not strictly true as most olivine and augite phenocrysts are pseudomorphed by mineral phases that collectively do not approximate the precursor compositions suggesting some local redistribution.

The cores of the dynamically influenced pillows to the west of the static-style area are in all petrographical aspects more disturbed and altered than their counterparts described above. Their cores however, are less altered than their respective exterior portions. Pillow MC120 still has noticeable ocelli which are distinctly flattened and, on a microscopic scale, partially homogenized with the matrix. Pseudomorphs of primary phenocrysts are often obliterated or reshaped. Some amygdales appear to have grown into large ellipsoidal quartz 'blasts' (4 - 8 mm) through metamorphic accretion.

For the carbonated pillows the internal pillow structure, mineralogy and texture, is entirely and pervasively replaced by a homogenous felty textured rock consisting of carbonate, chlorite, stilpnomelane, ilmenite and $\pm$ talc.

3.3 Sulphide Petrography

The sulphur content of the varied portions of the pillows ranges between 30 to 400 ppm for the pillow-core samples and between 100 to 5000 ppm for the pillow-exterior samples (i.e. rims and hyaloclastite). The sulphides which are essentially ubiquitous but sparsely distributed have been grouped into two categories, namely authigenic and allochemical, based on their association with the host silicates. The authigenic sulphides, which occur only in the pillow cores, have a more restricted mineralogical variation compared with the allochemical sulphides which dominate the pillow-exterior parts and some mineralized parts of the cores. The authigenic sulphides are considered to be secondary as no (primary) pyrrhotite globules or pyrite/magnetite atolls akin to those described for fresh submarine basalt (Gitlin, 1985) were observed.

The authigenic group comprises (in decreasing order of abundance) chalcopyrite, pyrite and pyrrhotite. The sulphides are located within the pillow cores, usually occur where the host silicate textures are well preserved and grain sizes are generally $\leq 40 \mu\text{m}$. A large

part of these sulphides is located within the amygdales which are hosted by texturally well preserved microspinel. The remainder of the authigenic sulphides is contained within mesostasis and amphibole grains of the microspinel-textured parts. The grains are seldom in contact with each other and are usually irregular in form whereas some of the larger grains within the amygdales have their shapes determined by idioblastic clinzoisite and, to a lesser extent, amphibole laths. Texturally, pyrite has two generations, which occur as solid subidioblastic kernels and a younger, porous to lamellar pyrite, similar in appearance to the spongy pyrite described by Gitlin (1985). The younger type is termed spongy for convenience and is occasionally seen to surround the solid type.

The allochemical sulphides are, (in decreasing order of abundance) pyrite, pyrrhotite, chalcopyrite, pentlandite, cobaltite and rare spinel. These sulphides are mostly situated in the pillow-exterior parts, i.e. pillow rims, hyaloclastite and hydrothermally infilled interpillow cavities. Occasionally isolated sulphide enrichments (mineralization) within the pillow cores, contraction-crack selvages and veins suggest a reintroduction of sulphide and they are also classified allochemical. They are usually hosted by entirely recrystallised clinzoisite-epidote, amphibole, $\pm$ chlorite, $\pm$ feldspar, of which the original primary textures are degraded. Sulphides occasionally replace silicates. The pillow margin sulphides are contained largely in metasomatic silicates (described earlier) whereas those in the interpillow areas occur in hydrothermally infilled silicates. The allochemical sulphides are generally between one and two orders of magnitude larger than their authigenic counterparts and tend to be more idioblastic. Two generations of pyrite, similar to those of the pillow cores, also occur within metahyaloclastite. Relatively compact pyrite kernels with spongy fringes do occur as well as patches of spongy pyrite with solid idiomorphic rims.

In both the authigenic and allochemical sulphide groups the association between (Fe-poor) clinzoisite and sulphide is quite apparent. There is a marked preference for sulphide to occur with clinzoisite rather than with (Fe-rich) epidote. This suggests that sulphides and silicates competed for Fe, implying a syngenetic origin.

The sulphides of the dynamically metamorphosed pillows are also iron-rich and contain

at the rims relatively large (≤ 2 mm) idiomorphic pyrite porphyroblasts. In contrast to this, the carbonated pillow MC20 is Fe-sulphide depleted relative to all other pillows and only appears to contain finely disseminated xenoblastic chalcopyrite. Unlike in the other pillows, here the sulphide is only marginally concentrated in the rim.

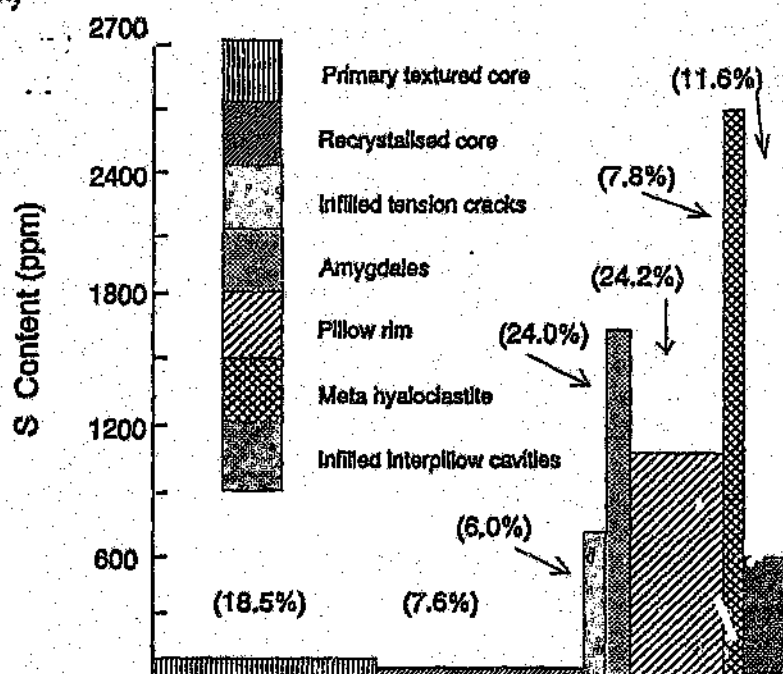
Occasionally both pillow cores and interpillow areas contain limonite and bornite either intermixed or as disjunct zones around chalcopyrite. Due to the limited and patchy distribution close to weathered surfaces these minerals are thought to represent late to recent oxidation and/or weathering, not warranting further consideration.

3.3.1 Modal Distribution of Sulphides

The distribution of sulphides within the various structural and textural domains of the pillows has been determined by modal analysis of which the results are summarized in Table 2.3 (see analytical techniques and short explanation at the bottom of the table for method of analyses). Polished thin sections (traverses from the core to the pillow rim and into the interpillow area) were selected for each pillow and are listed in this sequence.

The data in Table 2.3 are summarised in Fig. 3.7. For the statically- and dynamically metamorphosed pillows the data have been grouped into the Figs. 3.7a and 3.7b respectively. The main features of the data are: (1) the concentration of sulphide within the pillow exteriors relative to the cores; and (2) the concentration of sulphide in the amygdales in the pillow cores. The amygdales of the statically metamorphosed pillows are conspicuously enriched with sulphides relative to the rest of the pillow core. Although they comprise a very small proportion of the pillows, i.e. ≤ 3 volume%, they contain on average *ca.* 24% of all analysed sulphide. This may be related to the phenomenon that sulphur initially tends to gather as tiny spherules on the walls of vesicles of newly-formed ocean-ridge pillows. Gaseous sulphur is thought to exsolve from the melt (Moore and Schilling, 1973; MacRae *et al.*, 1979) and is then transformed into Fe-sulphides on reaction with the glassy vesicle surfaces (Moore and Calk, 1971). It is probable, therefore, that the biased sulphur distribution of the komatiitic pillow basalts may be related to similar early exsolved sulphur which during later hydration of the cores and infilling of the vesicles, attracted as nuclei, some of the locally redistributed sulphur and transition metal cations. The preservation of this sharply biased sulphide distribution is

a) Average for statically metamorphosed pillows



b) Average for dynamically metamorphosed pillows

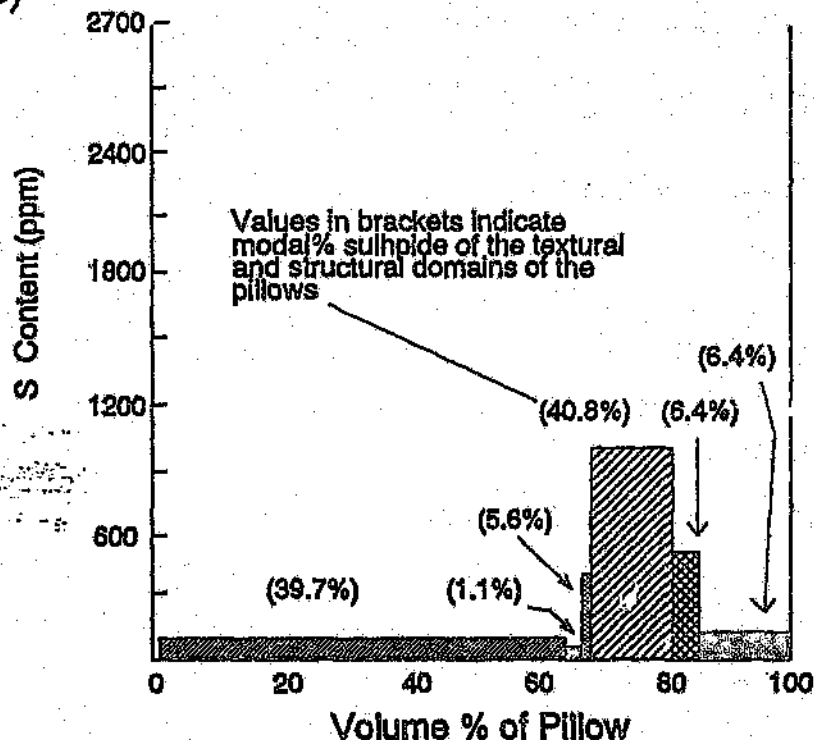


Fig. 3.7

A schematic representation of sulphur distribution between the various structural and textural domains of the statically (a) and dynamically (b) metamorphosed pillows (based on modal analysis averages of pillows: Table 2.3).

thus thought to be a primary feature and would also imply that the sulphides are secondary but authigenic, i.e. they were only locally redistributed (*cf* Gitlin, 1985). Comparison of the amygdale contents before and after dynamic metamorphism reveals a substantial decrease in S in terms of average concentration (from 1650 to 342 ppm) and a decrease in average modal content (from 24% to 5.6%, Fig. 3.7).

The metahyaloclastites and pillow rims of the statically metamorphosed pillows are the most enriched in sulphur of all the pillow domains. This is apparent if it is considered that initially rims and hyaloclastite must have contained similar amounts of S as the associated cores but are now enriched up to five and ten times these levels (Fig. 3.7). The enrichments seem to be concentrated by hydrothermal activity as the sulphides occur now in what used to be the fluid channelways. Comparison of rim and metahyaloclastite sulphur contents before and after dynamic metamorphism, reveals a drastic decrease for metahyaloclastite (i.e. from ca. 2500 ppm S down to 500 ppm S) but not for pillow rims.

3.3.2 Sulphide Chemistry

3.3.2.1 Chalcopyrite (CuFeS_2). Electron microprobe analyses of both the authigenic and allochemical chalcopyrites show that they approach the ideal chalcopyrite formula (*cf* number of atoms on basis of 2S, Table 3.2). Despite this the allochemical samples, which do not contain (visible) inclusions, seem to accommodate more readily limited substitution of Cu and Fe by Ni, Co, Zn and Sb than the authigenic ones (Fig. 3.8a). This imparts a tendency for the authigenic chalcopyrite to plot closest to the ideal position ($\text{Cu} + \text{Fe} = 2 \text{ basis } 2\text{S}$) whereas the allochemical ones plot on a tight substitution trend further away with a overlap zone in between.

Craig and Scott (1982) indicate that chalcopyrite deviates very little from its ideal formula. Although a wide range of trace elements have previously been found in this mineral their presence is generally thought to result from an admixture of impurities (Deer *et al.*, 1962e). The tendency of chalcopyrite to stay close to the ideal composition was demonstrated by Gavlin and Gabrielson (1947) who showed that the uptake of Co and Ni into chalcopyrite to be the least when compared to associated pyrite, pyrrhotite and arsenopyrite (reported in Deer *et al.*, 1962). The study area chalcopyrites however, show limited substitution close to the ideal formula and confirm the tendency to contain

lower trace abundances than associated pyrite and pyrrhotite (Tables 3.3, 3.4).

3.3.2.2 Pyrrhotite ($Fe_{1-x}S$). Due to scarcity and small grain size no XRD analysis of pyrrhotite was attempted to unequivocally determine the hexagonal or monoclinic crystal structure of the mineral. Several microprobe analyses were performed and show a narrow compositional range, i.e. 46.87 - 47.56 atomic% Fe (Table 3.3). Only one pyrite-pyrrhotite pair with a mutual boundary was encountered (cf analysis S150k2) and yielded pyrrhotite with 47.07 atomic% Fe + (Co + Ni). According to the T-X diagram of the pyrite + hexagonal pyrrhotite solvus in equilibrium with vapour at ca. 4 kb (Arnold, 1962; Touret and Barton, 1964) the sulphides appear to have equilibrated at 430°C which is well below metamorphic peak conditions of the host rocks. As the host silicates are not retrogressed, and are known to have formed at $520 \pm 25^\circ\text{C}$ (Fig. 4.37), it is evident that the sulphide blocking temperature is ca. 100°C below that of the silicates. This implies that pyrrhotite originally formed as the hexagonal variety, since monoclinic pyrrhotites are only stable up to 254°C, (Craig and Scott, 1982). The sulphur fugacity of the interpillow areas can also be constrained to between -7 and -6 log f_{S_2} (Craig and Scott, 1982) at the time of sulphide blocking, i.e. the temperature below which sulphide re-equilibration ceased.

Substitution of Ni and Co for Fe in pyrrhotite is demonstrated in Fig. 3.8b. The trace element content of authigenic pyrrhotite is generally less than those rated as allochemical suggesting restricted substitution for the former type, similar to that of chalcopyrite.

3.3.2.3 Pyrite (FeS_2). Several pyrite analyses are reported in Table 3.4. Fig. 3.8c demonstrates the range of Ni and Co substitution for Fe in the analyzed pyrite. Discrimination between the authigenic and allochemical types is not as clear as it is for pyrrhotite and chalcopyrite. Here it is evident that substitution of Ni and Co in allochemical pyrite is twice that of the authigenic ones and that the authigenic range is entirely included within the allochemical range. Also listed, apart from the authigenic and allochemical classification, are the textural labels idioblastic or spongy (described earlier). Idioblastic- vs spongy pyrite trace element plots (not shown) show no specific relationships or differences.

From the total amount of substitution, i.e. up to 4 wt%, it is obvious that pyrite contains

TABLE 3.2 MICROPROBE ANALYSES OF CHALCOPYRITE FROM CORES AND INTERPILLOW AREAS.

Sample	S6001	S6002	S60C1	S60C3	S60C2a	S60C2b	S60H1	S60H2	S15CA1	S150A5a	S150A5b	S150R3a	S150R4a	S150R4b
Mineral	Chalpy.	Chalpy.	Chalpy.	Chalpy.	Chalpy.	Chalpy.	Chalpy.	Chalpy.	Chalpy.	Chalpy.	Chalpy.	Chalpy.	Chalpy.	Chalpy.
Origin	AUTH	AUTH	AUTH	AUTH	ALLOCH	ALLOCH	ALLOCH	ALLOCH	AUTH	AUTH	AUTH	ALLOCH	ALLOCH	ALLOCH
Texture	XENOB	XENOB	XENOB	XENOB	XENOB	XENOB	XENOB	XENOB	XENOB	XENOB	XENOB	XENOB	XENOB	XENOB
Text. unit	CORE	CORE	CORE	CORE	VEIN	VEIN	HYALC.	HYALC.	CORE	CORE	CORE	HYALC	HYALC	HYALC
Cu wt%	34.10	34.16	34.36	33.80	34.18	34.28	34.39	34.20	34.04	33.99	34.25	33.89	33.44	33.50
Fe wt%	30.25	30.52	29.94	29.96	30.35	30.20	30.30	30.07	30.37	30.63	30.62	30.02	30.46	30.11
As wt%	0.00	0.00	0.02	0.04	0.00	0.00	0.06	0.10	0.00	0.00	0.00	0.00	0.00	0.02
Co wt%	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.05	0.00	0.00	0.00	0.00	0.00	0.00
Zn wt%	0.01	0.00	0.01	0.00	0.20	0.03	0.02	0.00	0.00	0.00	0.00	0.02	0.01	0.00
Ni wt%	0.02	0.00	0.00	0.02	0.00	0.01	0.01	0.08	0.01	0.00	0.00	0.04	0.02	0.02
Sb wt%	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.02	0.00
S wt%	34.66	34.63	35.52	35.13	35.24	35.12	35.14	35.42	35.04	35.30	35.03	35.32	35.33	35.14
Total	99.04	99.31	99.85	98.95	99.97	99.64	99.94	99.92	99.46	99.92	99.90	99.29	99.28	98.79
Atomic %														
Cu	24.84	24.84	24.74	24.57	24.63	24.79	24.81	24.63	24.66	24.49	24.72	24.54	24.20	24.37
Fe	25.08	25.25	24.53	24.78	24.89	24.85	24.87	24.64	25.03	25.11	25.15	24.73	25.08	24.93
As	0.00	0.00	0.01	0.02	0.00	0.00	0.04	0.06	0.00	0.00	0.00	0.00	0.00	0.01
Co	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.04	0.00	0.00	0.00	0.00	0.00	0.00
Zn	0.01	0.00	0.01	0.00	0.14	0.02	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.00
Ni	0.02	0.00	0.00	0.02	0.00	0.01	0.01	0.06	0.01	0.00	0.00	0.03	0.02	0.02
Sb	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00
S	50.05	49.91	50.70	50.61	50.34	50.34	50.25	50.56	50.31	50.41	50.12	50.69	50.68	50.67
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
(S+As):(Fe+Ni+Co+Sb)	1.00	1.00	1.03	1.03	1.02	1.01	1.01	1.03	1.01	1.02	1.00	1.03	1.03	1.03
Number of atoms on the basis of 2S														
Cu	0.99	1.00	0.98	0.97	0.98	0.98	0.99	0.97	0.98	0.97	0.99	0.97	0.96	0.96
Fe	1.00	1.01	0.97	0.98	0.99	0.99	0.99	0.97	1.00	1.00	1.00	0.98	0.99	0.98
Co + Zn + Ni + Sb	0.001	0.000	0.000	0.001	0.006	0.001	0.001	0.004	0.000	0.000	0.000	0.002	0.001	0.001

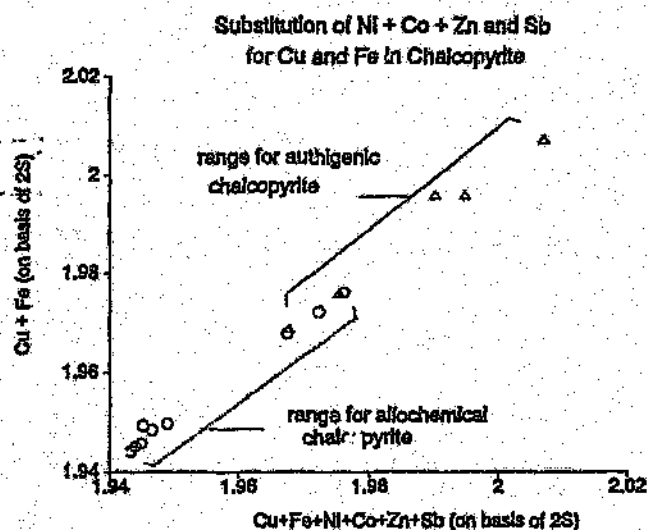
TABLE 3.3 MICROPROBE ANALYSES OF PYRRHOTITE AND PENTLANDITE FROM CORES AND INTERPILLOW AREAS.

Sample	S60H2a	S60Hb	S60H3a	S60H3b	S150A1a	S150A6a	S150A6a	S150K2	S150R4	Sample	S150K2
Mineral	Pyrrht.	Pyrrht.	Pyrrht.	Pyrrht.	Pyrrht.	Pyrrht.	Pyrrht.	Pyrrht.	Pyrrht.	Mineral	Pentl.
Origin	ALLOCH	ALLOCH	ALLOCH	ALLOCH	AUTH	AUTH	AUTH	ALLOCH	ALLOCH	Origin	ALLOCH
Text. unit	HYALC	HYALC	HYALC	HYALC	CORE	CORE	CORE	HYALC	HYALC	Text. unit	HYALC
Fe wt%	60.49	60.38	60.47	60.65	59.83	59.84	59.85	60.05	59.73	Fe wt%	29.35
As wt%	0.05	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Co wt%	2.32
Co wt%	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	Ni wt%	34.34
Zn wt%	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01	S wt%	33.83
Ni wt%	0.35	0.50	0.38	0.46	0.13	0.06	0.18	0.40	0.65	Total	99.84
Sb wt%	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	Atomic %	
S wt%	39.00	39.18	38.50	38.73	39.01	39.10	39.04	39.01	39.30	Fe	23.83
Total	99.89	100.07	99.35	99.84	98.98	99.01	99.08	99.46	99.69	Co	1.79
Atomic %										Ni	26.53
Fe	46.97	46.76	47.28	47.18	46.77	46.74	46.74	46.77	46.37	S	47.85
As	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Total	100.00
Co	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00		
Zn	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01		
Ni	0.26	0.37	0.28	0.34	0.10	0.04	0.13	0.30	0.48		
Sb	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00		
S	52.75	52.86	52.44	52.48	53.13	53.20	53.11	52.93	53.14		
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00		
Fe+(Ni+Co):S+(As)	0.89	0.89	0.91	0.91	0.88	0.88	0.88	0.89	0.88		
Fe+(Ni+Co)	47.22	47.14	47.56	47.52	46.87	46.80	46.89	47.07	46.86		

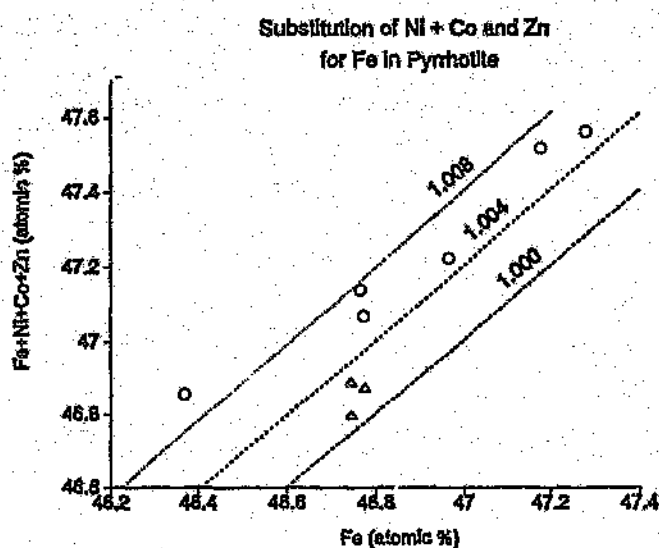
TABLE 3.4 MICROPROBE ANALYSES OF PYRITE FROM CORES AND PILLOW RIMS

Sample	S60H4a	S60H4c	S60H4d	S150A1a	S150Ab	S150Ac	S150A2a	S150K1a	S150K1b	S150K1bs	S150K1bs	S150R2
Mineral	Pyrite	Pyrite	Pyrite	Pyrite	Pyrite	Pyrite	Pyrite	Pyrite	Pyrite	Pyrite	Pyrite	Pyrite
Origin	ALLOCH	ALLOCH	ALLOCH	ALLOCH	AUTH	AUTH	AUTH	ALLOCH	ALLOCH	ALLOCH	ALLOCH	ALLOCH
Texture	IDIOM	SPONGY	SPONGY	IDIOM	SPONGY	SPONGY	IDIOM	IDIOM	IDIOM	SPONGY	SPONGY	IDIOM
Text. unit	HYALC	HYALC	HYALC	CORE	CORE	CORE	CORE	HYALC	HYALC	HYALC	HYALC	HYALC
Fe wt%	46.27	44.03	46.24	46.11	43.21	42.99	46.11	42.53	42.42	40.55	40.69	45.50
As wt%	0.00	0.33	0.00	0.03	0.02	0.00	0.03	0.00	0.00	0.01	0.04	0.02
Co wt%	0.00	2.12	0.00	0.00	0.68	1.13	0.00	0.53	0.00	1.43	0.00	0.00
Zn wt%	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.01	0.00
Ni wt%	0.03	0.01	0.01	0.00	2.32	2.01	0.00	3.64	4.33	4.43	5.59	0.82
Sb wt%	0.00	0.00	0.00	0.01	0.02	0.04	0.01	0.00	0.02	0.01	0.00	0.00
S wt%	53.80	52.82	53.62	53.04	52.69	52.70	53.04	53.28	53.14	52.96	52.73	53.53
Total	100.10	99.31	99.88	99.19	98.94	98.87	99.19	99.98	99.91	99.41	99.06	99.87
Atomic %												
Fe	33.05	31.84	33.11	33.29	31.34	31.20	33.29	30.53	30.49	29.30	29.51	32.61
As	0.00	0.18	0.00	0.02	0.01	0.00	0.02	0.00	0.00	0.01	0.02	0.01
Co	0.00	1.45	0.00	0.00	0.47	0.78	0.00	0.36	0.00	0.98	0.00	0.00
Zn	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00
Ni	0.02	0.01	0.01	0.00	1.60	1.39	0.00	2.49	2.96	2.04	3.66	0.56
Sb	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.01	0.00	0.00	0.00
S	66.93	66.53	66.88	66.70	66.34	66.62	66.70	66.62	66.54	66.66	66.61	66.82
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
S:Fe	2.02	2.00	2.02	2.00	1.99	2.00	2.00	2.00	1.99	2.00	2.00	2.02

a)



b)



c)

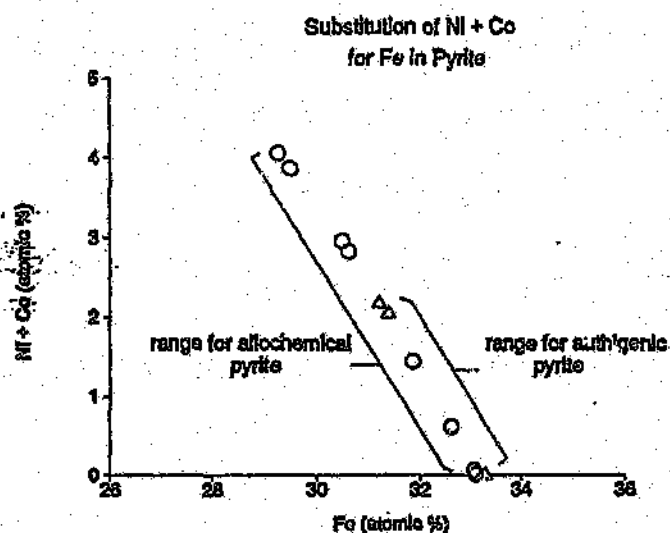


Fig. 3.8

Plots to show trace element substitution of major components in chalcopyrite (a), pyrrhotite (b), and pyrite (c). Triangles (Δ) = authigenic sulphides; circles ($\circ$) = allochemical sulphides. Data from Table 3.2, 3.3 and 3.4.

by far the largest amount of non-essential metallogenic elements when compared to chalcopyrite and pyrrhotite. It is doubtful, however, that these trace contents are due to inadvertently analysed inclusions beneath the grain surfaces as the trend in Fig. 3.8c forms a coherent substitution trend. On the other hand, data scatter would support the presence of inclusions. The capacity of pyrite to take up Ni and Co appear to make it a metallogenic-element sink for these elements, especially when other sulphides such as pentlandite and/or cobaltite are not in the vicinity to use these as essential components. Pyrite compositions are, furthermore, quite heterogeneous within individual grains.

Deer *et al.*, (1962) report that Fe in the pyrite structure may easily be replaced by Ni or Co and that Co preferentially concentrates in pyrite and Ni in pyrrhotite when these sulphides are associated. On the other hand, Ni and Co are more readily concentrated in pyrite than in chalcopyrite when these are associated. The authors also report a pyrite analyses containing 13.9 wt% Co, suggesting that the degree of substitution observed here is not large. Optically anisotropic pyrite was also found to occur (S130A1b, composition not shown), containing *ca.* 0.95 wt% Co and 2.5 wt% Ni. Pyrite, which is usually isotropic, is also reported to occasionally show anisotropic characteristics for no clear reasons (Ramdohr, 1969).

3.4 Geochemistry of Pillow Alteration

Geochemical sample selection strongly reflects pillow structures which have controlled early fluid access. Element (re)distributions of the pillows that have resulted from fluid interaction are shown in three ways:

- 1) Overall distributions of selected elements between pillow interiors and -exteriors from both the statically and dynamically metamorphosed areas.
- 2) Detailed distributions of elements between pillow cores and veinlet-margins.
- 3) Detailed elemental flux determinations for comparison with those established for seawater-basalt interaction.

A sample set from each pillow comprises; pillow core; 2 - 3 cooling-crack alteration selvages; outer-core; pillow rim and metahyaloclastite (in the text, the latter two sample types are also referred to as pillow exterior samples).

3.4.1 Pillow Core vs Pillow Exterior Element Plots

The bulk chemical compositions of the pillows are reported in Table 3.5 and some of the data is shown in Fig. 3.9. These plots are intended to depict the general trends and differences between pillow interiors and exteriors and those changes that progressively occur between the statically and dynamically metamorphosed pillows (including the carbonated ones).

Al_2O_3 , TiO_2 and Cr_2O_3 , generally considered to be immobile (eg. Humphries and Thompson, 1978a,b; Morrison, 1978; Smith, 1980; Thompson, 1983; Mottl, 1983), clearly show redistribution within pillows. Al_2O_3 , Cr_2O_3 and TiO_2 are consistently enriched at the rims of pillows relative to the cores. Hyaloclastite, however, is inconsistently enriched or depleted suggesting variable mobilization. Variations in Al_2O_3 and TiO_2 content are sympathetic with the petrographically observed distribution of albitic plagioclase and titanite whereas Cr_2O_3 enrichments are not related to increases in the presence of chromite. The chromium anomalies for the pillow exteriors seem to suggest that it may be contained within silicate phases.

The H_2O -plot (Fig. 3.9) shows the overall levels of hydration of the pillows and a crude tendency for the pillow exteriors to be more hydrated than the corresponding cores. The pillow cores have ca. 2 wt% H_2O which essentially reflects the crystalline water content of amphibole which predominate in them. The slight increases that occur toward the pillow exteriors is attributed to the additional presence of chlorite. By contrast, mid-ocean ridge basalts (MORBs) contain up to 7 wt% H_2O (Humphries and Thompson, 1978a) which is due to a much higher content of chlorite with ca. 12 wt% H_2O (Deer *et al.*, 1974). The carbonated pillows have a reversed hydration pattern as a result of carbonate extensively replacing hydrous minerals at the rims and to a lesser degree in the cores, implying metasomatism via the interpillow channelways.

CO_2 is very low for the (statically-) seafloor metamorphosed pillows and variable for the dynamically metamorphosed pillows (CO_2 -levels of MC60 is elevated due to a late event). The virtual absence of CO_2 in the statically metamorphosed pillows strongly supports a high-temperature seafloor alteration origin as CO_2 is usually added in low temperature off-axis environments (Staudigel *et al.*, 1981; Mottl and Holland, 1978;

TABLE 3.5 MAJOR AND TRACE ELEMENT ANALYSES OF KOMATIITIC PILLOW BASALTS.

SAMPLE NUMBER COMMENT	HC020 01 CORE	HC020 03 OC	HC020 04 RIH	HC020 05 HYDF	HC020 06 HYALC	HC060 01 CORE	HC060 2A1 CCWR	HC060 2A2 CCV	HC060 2B1 CCWR	HC060 2B2 CCV	HC060 2C1 CCWR	HC060 2C2 CCV	HC060 03 OC	HC060 04 RIH	HC060 05 HYDF	HC060 06 HYALC	HC120 01 CORE	HC120 03 OC	HC120 04 RIH	HC120 05 HYDF	HC150 01 CORE	HC150 2A1 CCWR
S102	51.27	52.50	46.98	91.11	56.29	50.46	57.57	59.14	56.07	55.87	60.11	55.67	52.64	57.61	94.76	51.03	51.82	53.57	55.28	89.73	53.30	51.42
T102	0.56	0.56	0.61	0.06	0.30	0.76	0.78	0.62	0.79	0.72	0.68	0.77	0.81	0.89	0.07	0.61	0.58	0.54	0.73	0.07	0.71	0.73
AL2O3	6.12	6.26	6.72	0.41	4.30	9.21	7.79	7.99	8.48	10.47	9.24	8.81	9.41	10.11	1.68	8.62	8.09	7.69	9.43	3.47	10.79	6.60
FE2O3	3.80	3.14	4.19	0.04	1.14	2.18	1.17	0.97	1.34	1.57	0.49	3.88	1.36	1.09	0.01	2.20	3.26	2.56	3.70	0.01	1.20	1.94
FeO	10.90	11.94	9.31	0.68	6.55	10.29	9.90	6.88	10.23	8.27	9.23	7.55	10.58	9.94	0.66	13.53	10.43	10.57	8.18	0.66	10.23	12.15
MNO	0.170	0.180	0.010	0.010	0.020	0.100	0.177	0.156	0.185	0.155	0.175	0.150	0.010	0.010	0.010	0.230	0.010	0.010	0.010	0.010	0.191	0.267
MgO	16.88	17.14	13.04	0.51	6.22	10.78	9.54	7.01	9.56	7.95	7.82	9.61	10.28	8.81	0.34	11.95	11.60	11.71	9.20	0.40	9.15	11.02
CaO	8.17	6.14	16.91	6.16	23.21	14.22	10.15	14.25	9.88	10.48	9.19	9.46	11.86	9.27	0.19	10.89	11.96	11.78	11.01	3.62	10.43	14.03
Na2O	0.88	0.80	0.85	0.61	0.92	1.17	2.07	2.88	2.55	4.21	2.76	3.73	1.23	2.04	1.45	0.40	1.12	1.03	1.44	1.50	2.82	1.41
K2O	0.52	0.62	0.64	0.01	0.42	0.02	0.01	0.04	0.06	0.09	0.20	0.06	0.13	0.13	0.01	0.01	0.01	0.04	0.01	0.01	0.16	0.07
P2O5	0.14	0.14	0.07	0.09	0.01	0.12	0.01	0.03	0.01	0.01	0.01	0.01	0.15	0.12	0.05	0.13	0.12	0.07	0.13	0.11	0.18	0.01
CR2O3	0.20	0.18	0.32	0.01	0.08	0.05	0.13	0.01	0.10	0.18	0.06	0.15	0.04	0.11	0.01	0.31	0.09	0.10	0.15	0.01	0.01	0.10
TOTAL	99.61	99.70	99.85	99.70	99.46	99.36	99.30	99.98	99.25	99.98	99.97	99.85	98.50	100.13	99.24	99.69	99.31	99.67	99.27	99.60	99.17	99.75
S(ppm)	53	15	117	7	103	40	94	218	157	264	130	206	32	789	1	4350	104	550	2740	25	208	121
CO2	5.930	4.380	11.300	4.860	15.200	1.140	0.540	1.270	1.150	1.720	0.930	1.592	1.610	1.950	0.160	3.490	2.250	1.933	3.380	1.940	0.100	0.420
H2O	5.62	5.67	4.53	0.01	2.42	3.10	2.38	1.70	2.50	1.77	2.53	2.07	2.44	3.40	0.03	4.52	3.53	3.14	3.44	0.07	1.76	1.67

Trace Elements (ppm)

CO	75	66	72	0	42	55	68	37	68	51	45	48	53	58	5	103	62	61	65	3	58	76
NI	384	410	363	0	195	159	178	117	176	132	179	149	189	239	6	721	174	404	430	8	152	143
ZN	99	100	85	22	51	92	99	64	99	80	92	80	105	93	15	128	93	96	88	9	87	116
CU	61	43	112	48	118	76	81	112	120	82	114	184	16	436	4	321	70	84	451	3	147	72
RB	13	16	14	5	12	4	5	5	5	8	4	6	5	5	6	3	4	3	3	3	6	5
SR	38	33	75	14	93	108	57	75	52	90	55	77	75	64	23	29	58	41	56	17	149	88
Y	13	12	16	8	13	19	16	13	17	14	16	13	21	18	8	14	16	14	16	8	19	18
ZR	34	35	37	11	20	62	69	52	67	60	64	61	63	73	9	39	47	40	53	9	61	63
HB	0	0	0	6	0	1	2	2	3	4	2	3	0	1	2	0	1	0	1	1	3	3
HG	0.093	0.108	0.031	0.000	0.030	0.142	0.220	0.024	0.135	0.083	0.340	0.255	0.040	0.108			0.054	0.061	0.007	0.000	0.097	0.205
AU	0.018	0.021	0.008	0.002	0.011	0.009							0.007	0.016	0.009	0.010	0.006	0.011	0.027	0.003	0.017	
SB	0.134	0.767	0.117	0.087	0.236	0.101							0.074	0.150	0.071	0.029	0.050	0.280	0.743	0.130	0.109	
BA	63.00	73.00	121.00	12.06	75.32	39.80	52.00	82.00	68.00	68.00	60.00	88.00	109.00	54.00	16.84	84.00	44.30	47.00	43.00	51.00	56.00	
AS	2.82	72.97	1.39	0.90	9.93	0.35							1.76	111.00	0.43	66.00	0.35	1.06	20.70	0.47	0.78	
V	0.00	1.38	0.91	0.69	4.56	0.00															9.92	

Trace Elements of sulphide extracts (ppm)

AU	1.930		71.780		0.033								0.035		0.021	0.958	0.311	0.516		1.345	
SB	65.020		126		0.405								0.400		0.310	59.560	7.150	6.143		151.93	
AS	1259.5		14771		25.2								1.2		1.6	2993.3	735.3	1588.0		210.44	

TABLE 3.5 (CONTINUED)

SAMPLE NUMBER COMMENT	HC150 2A2 CC+HR	HC156 2B1 CC+HR	HC150 2B2 CC+HR	HC150 03 OC	HC150 04 RIM	HC150 06 HYALC	HC180 01 CORE	HC180 2A1 CC+HR	HC180 2A2 CC+HR	HC180 2B1 CC+HR	HC180 2B2 CC+HR	HC180 2C1 CC+HR	HC180 2C2 CC+HR	HC180 03 OC	HC180 04 RIM	HC180 05 HYDF	HC180 06 HYALC
S102	53.65	51.79	53.93	53.62	48.79	48.31	51.63	48.52	50.08	49.91	54.86	50.51	51.65	51.40	53.43	89.78	53.98
TiO ₂	0.46	0.72	0.40	0.70	0.78	0.54	0.71	0.74	0.70	0.46	0.32	0.73	0.67	0.73	0.82	0.02	1.03
Al ₂ O ₃	10.27	7.49	10.91	9.38	12.50	10.54	10.91	8.56	9.72	8.64	10.37	10.50	9.99	9.13	14.00	5.34	8.41
Fe ₂ O ₃	1.51	1.07	0.46	1.49	2.54	3.11	1.83	3.48	1.04	2.69	0.58	1.10	0.23	3.05	1.07	0.01	1.10
FeO	8.90	12.86	9.67	10.51	11.76	12.44	11.07	11.28	12.02	11.39	9.11	12.99	11.34	10.98	9.67	0.70	13.18
MnO	0.237	0.244	0.231	0.160	0.250	0.170	0.213	0.237	0.220	0.236	0.221	0.207	0.186	0.290	0.160	0.040	0.160
MgO	7.12	11.02	6.71	9.46	10.71	11.36	9.58	11.62	9.91	11.47	7.41	10.60	8.54	10.00	7.13	0.05	9.39
CaO	13.33	13.17	12.93	11.42	9.09	10.11	11.39	12.64	13.08	12.27	13.66	10.56	12.64	12.03	9.21	0.51	10.28
Na ₂ O	3.46	1.53	3.59	2.11	1.52	1.21	1.72	1.30	1.96	1.41	2.53	2.09	2.82	1.36	3.01	3.36	1.38
K ₂ O	0.87	0.07	0.85	0.10	0.64	0.46	0.37	0.16	0.27	0.52	0.81	0.16	0.59	0.27	0.69	0.06	0.32
P ₂ O ₅	0.01	0.02	0.01	0.16	0.24	0.15	0.23	0.01	0.06	0.01	0.01	0.01	0.01	0.21	0.25	0.11	0.16
CR ₂ O ₃	0.02	0.06	0.04	0.02	0.07	0.13	0.04	0.07	0.06	0.19	0.07	0.06	0.07	0.02	0.05	0.01	0.17
TOTAL	99.84	99.74	98.93	99.13	98.89	99.13	99.69	98.62	99.16	99.20	99.95	99.52	98.74	99.47	99.49	99.99	99.56
S(ppm)	64	101	63	101	120	1160	95	46	50	135	85	54	257	105	646	142	1070
CO ₂	0.120	0.280	0.130	0.180	0.080	0.110	0.310	0.310	0.280	0.400	0.460	0.180	0.023	0.160	0.160	0.290	0.240
H ₂ O	0.78	1.67	0.21	1.67	3.47	3.08	2.02	2.00	1.57	1.75	1.32	1.79	1.50	2.03	2.18	0.02	2.44

Trace Elements (ppm)

CO	31	64	33	43	66	53	71	80	66	67	54	75	65	74	60	8	56
HI	101	160	104	136	278	566	154	160	146	170	139	166	145	150	198	7	369
ZN	80	118	90	97	116	122	92	121	105	72	89	105	99	105	89	12	125
CU	21	74	21	81	177	328	84	23	17	127	92	37	259	62	316	40	772
RB	17	6	18	4	17	14	11	6	7	13	18	7	12	9	19	7	13
SR	182	102	185	132	103	66	171	118	188	173	175	150	221	127	229	45	108
Y	13	18	14	19	19	16	19	19	18	19	17	20	18	19	21	7	21
ZR	46	63	46	62	67	40	63	63	63	65	61	65	61	60	74	10	74
NB	1	2	2	0	2	0	3	2	2	2	2	3	3	2	4	1	1
KG	0.131	0.185	0.308	0.196	0.000	0.000	0.071	0.150	0.145	0.150	0.095	0.195	0.185	0.138	0.003	0.000	0.002
AI				0.007	0.007	0.006								0.006	0.004	0.004	0.014
SB				0.009	0.100	0.091	0.060							0.320	0.070	0.066	0.066
BA	209.00	74.00	193.00	71.00	87.00	92.30	105.00	25.00	209.00	107.00		45.00	88.00	76.00	145.00	32.00	44.00
AS				6.140	0.283	0.420	0.190							2.320	2.040	0.470	2.071
W				3.39	5.73	4.87	6.70							6.20	5.29	2.97	13.40

Trace Elements of sulphide extracts (ppm)

AU				7.430	0.511	25.900								0.251	12.885	1.090
SB				1095.6	24.580	29.990								17.550	2414	4.240
AS				425.5	104.5	155.4								161.5	20380	108.2

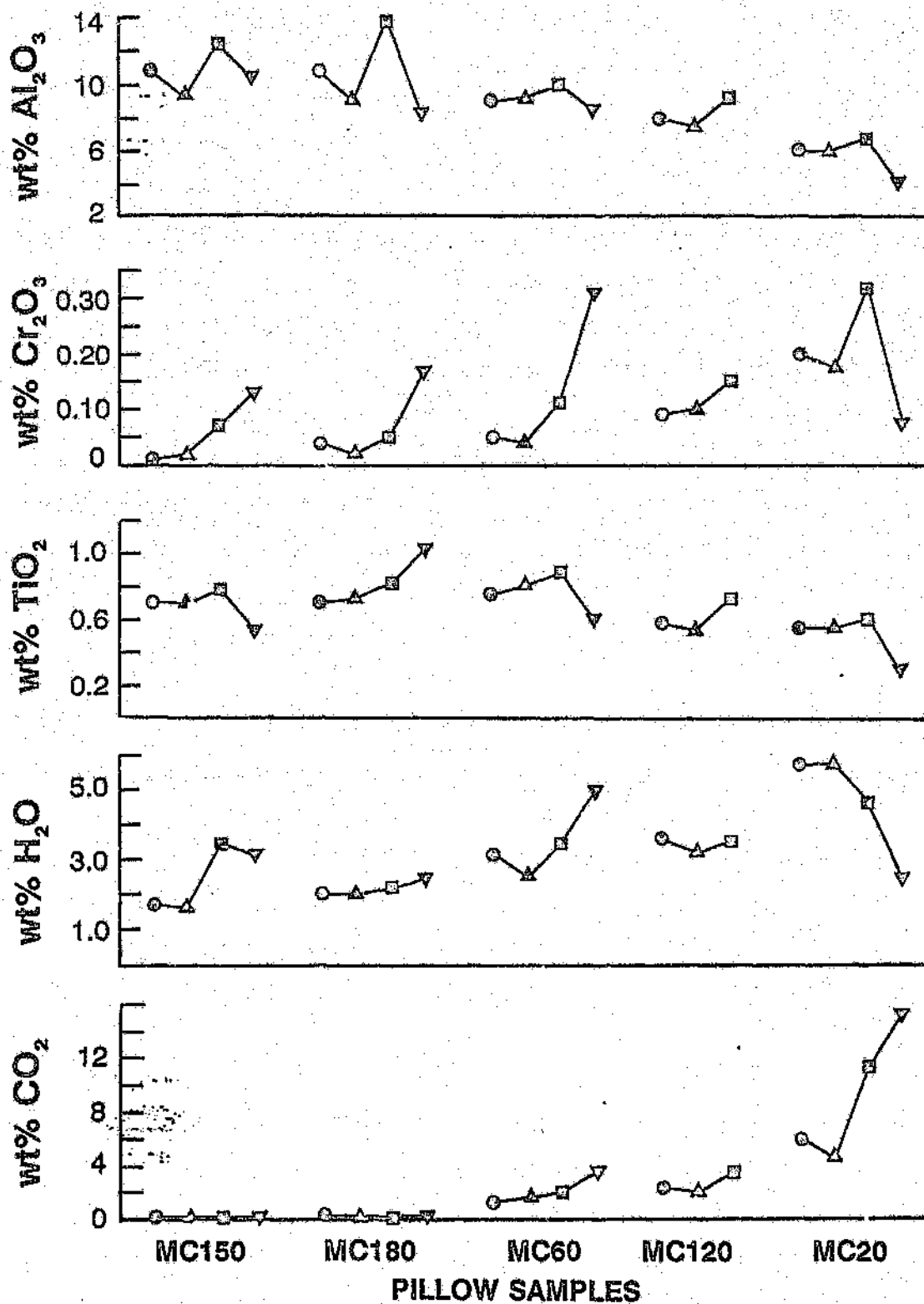


Fig. 3.9

Element abundances vs textural domains of pillows, indicated by sample number and symbol. The plot shows the distribution of some key elements within the pillows and differences between the statically- and dynamically metamorphosed pillows. Pillow core = $\bigcirc$; outer core = $\triangle$; pillow rim = $\square$; metahyaloclastite = ∇ .

Seyfried and Bischoff, 1981).

3.4.2 Pillow Core - Cooling-Crack Selvage Abundance Plots

Detailed element distributions between pillow cores and veinlet-margins are shown in Fig. 3.10 and the data are presented in Table 3.6. The samples were selected from three representative pillows from the statically metamorphosed area. (Corresponding element distributions for the dynamically influenced pillows were not obtained due to the general tectonic-related degradation of veinlets). Selection of the three pillows is based on the range of alteration displayed by the cooling-cracks: ie. MC150 = amphibolitised selvages (A-selvages), MC180 = albitisation overprinting amphibolitised selvages (AA-selvages) and MC60 = albitised and silicified selvages (AS-selvages). The two end-members of the range are represented by the amphibolitised- and albitised + silicified selvages respectively whereas the amphibolitised ones with the albitic overprint are inferred to represent a composite intermediate member.

From each pillow two to three cooling-crack selvage- and cooling-crack veinlet pairs were sampled by means of sawing out the veins with their attendant alteration selvages. The samples were then lightly crushed and sieved to retain as much sample as possible with a grain-size between *ca.* 1 - 2 mm. The grains were immersed in water and hand picked under a binocular microscope and sorted into "selvage only" and "veinlet $\pm$ selvage" for chemical analyses. The analytical error is given at a standard deviation of 2σ for each of the elements as indicated by vertical bars in the plots.

SiO_2 is lost from the A-selvages and gained by both the AS- and AA-types (Fig. 3.10). Although the plots show both losses and gains for the veinlet selvages the variations are notably systematic for the individual pillows suggesting that distribution of the component may have been mineralogically controlled. Most workers show that SiO_2 is usually lost to seawater during basalt alteration (e.g. Hart, 1973; Scott and Hajash, 1976; Keays and Scott, 1976; Thompson, 1983) and that in some instances both losses and gains occur (eg: Cann, 1969; Humphries and Thompson, 1978a).

Al_2O_3 is depleted in all veinlet margins relative to the pillow core by up to 4 wt% except for one sample which is close to parity (within analytical range). The A-selvages are most

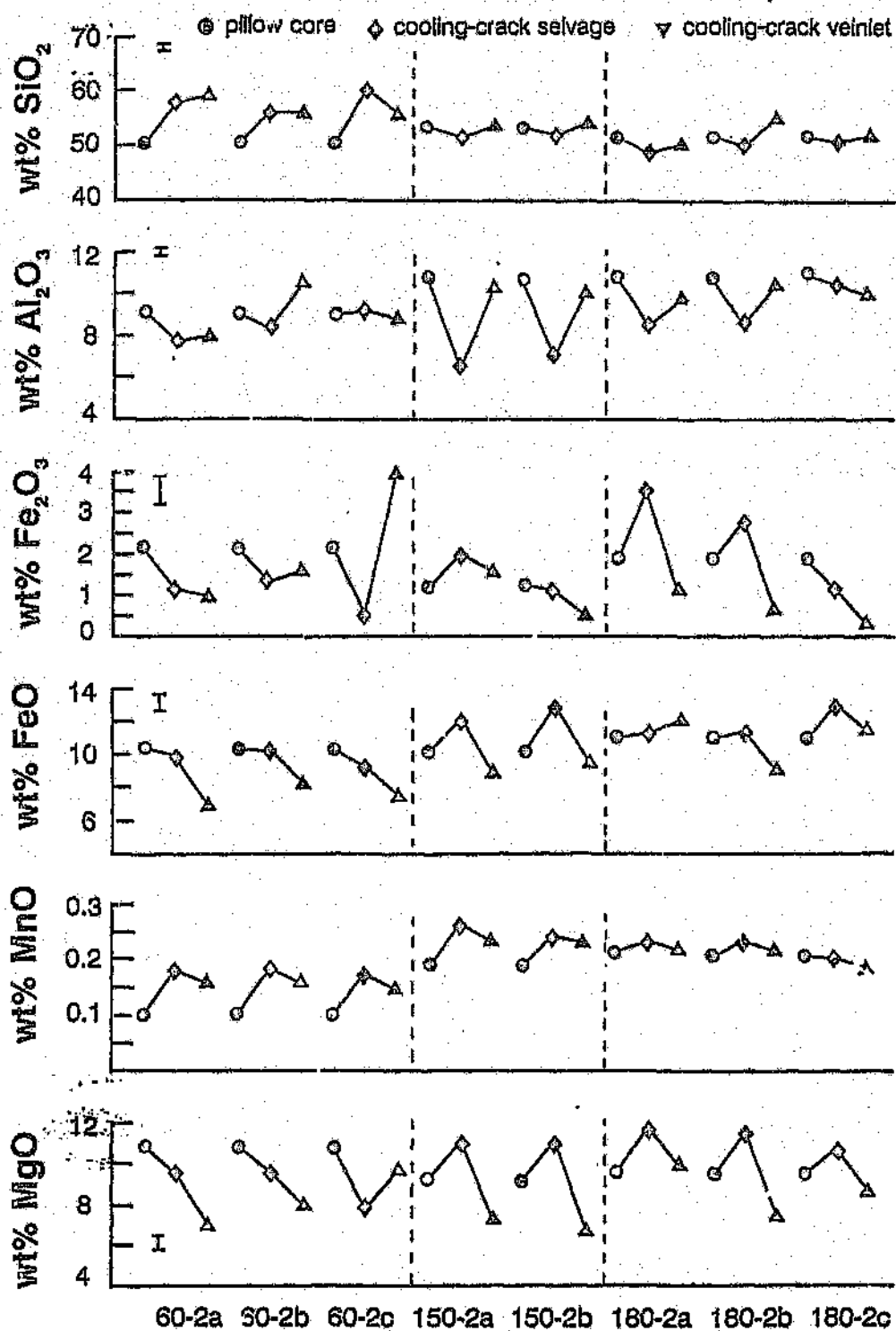


Fig. 3.10

Elemental abundances vs incipiently altered pillow structures: $\odot$ = pillow core; $\diamond$ = cooling-crack selvage; ∇ = cooling-crack veinlets indicated by sample number and/or symbols for comparisons between the seafloor metamorphosed pillows only. (Analytical standard deviation = 2σ indicated by vertical bar).

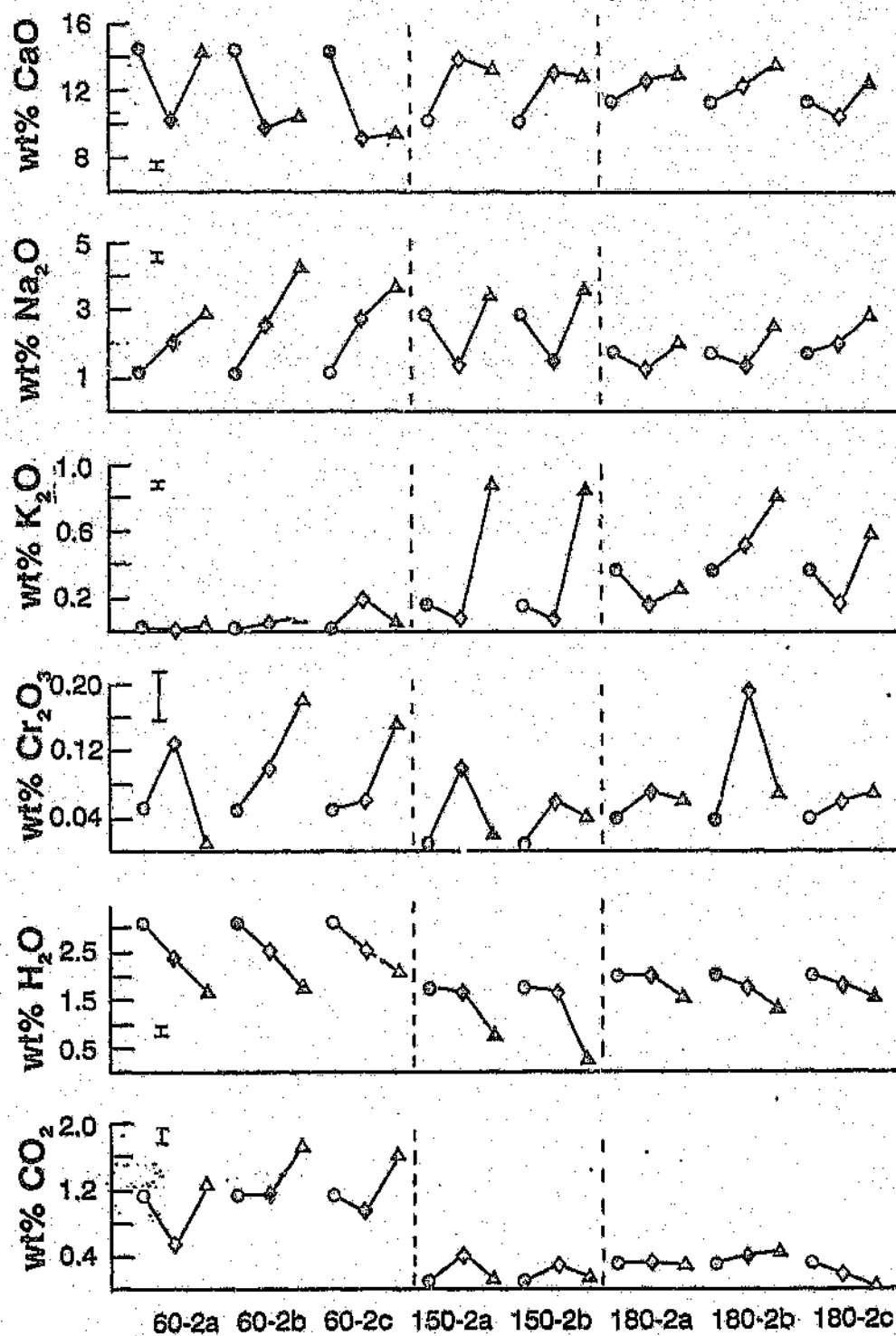


Fig. 3.10

(Continued...)

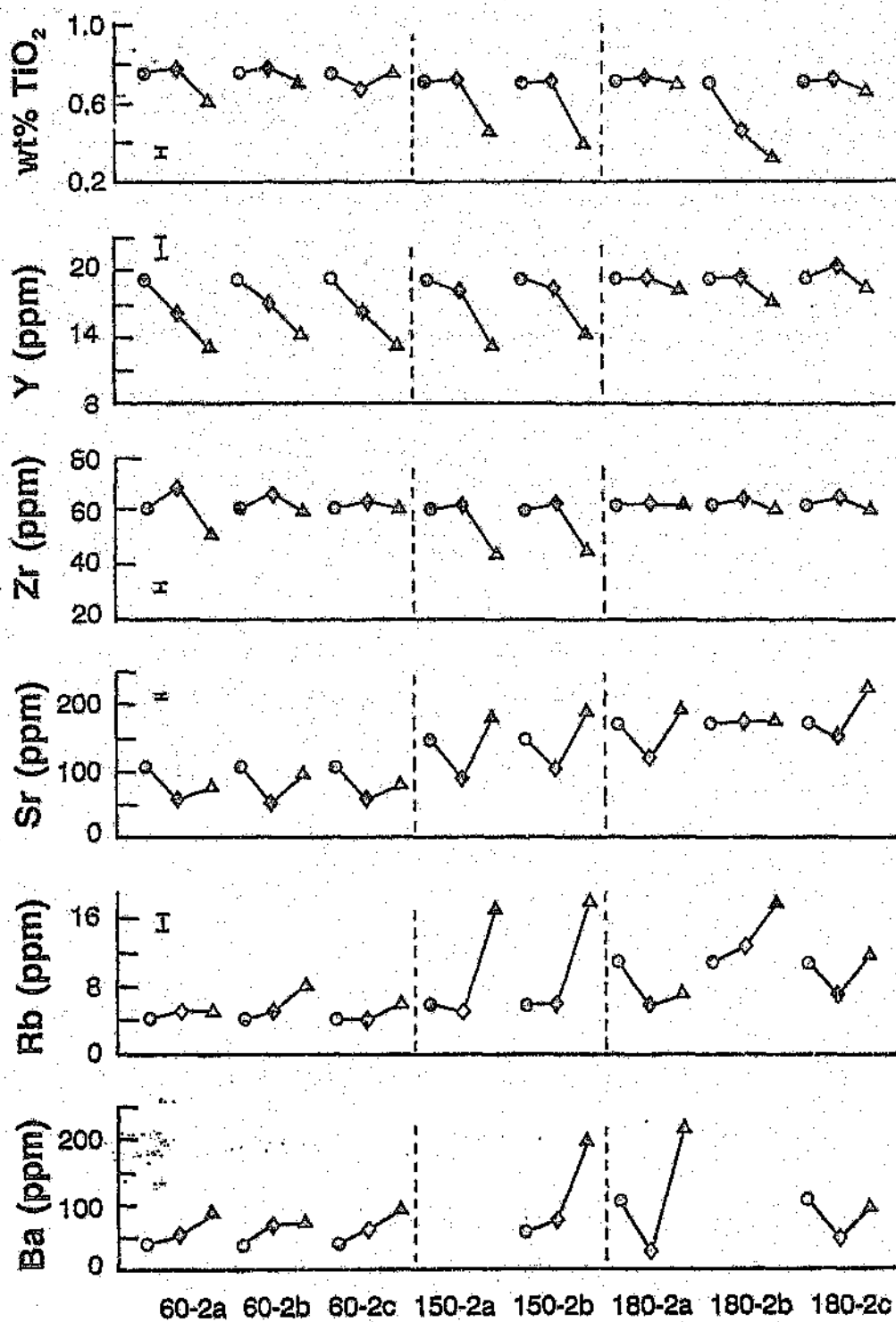


Fig. 3.10

(Continued...)

Table 3.6 A summary of element distribution between pillow cores and veinlet walls as shown in Figure 3.10.

Category	Element	Element distribution	Distribution control
Group A	SiO ₂ , FeO, MgO, CaO, Na ₂ O.	Distribution patterns are consistent within pillows but vary between pillows. Distribution varies in sympathy with alteration type.	Mineralogically controlled, due to the formation of amphibole at the cost of albitic plagioclase or vice versa.
Group B	Al ₂ O ₃ , MnO, Cr ₂ O ₃ , TiO ₂ , H ₂ O, Zr, Sr, Y, Ba	Distribution patterns are consistent within and between pillows regardless of alteration type.	Other than mineralogical (e.g. physicochemical = T, P, pH, aH, bulk rock composition, environment, etc.). Variable but minor mineralogical control evident, e.g. Al ₂ O ₃ , TiO ₂ , Cr ₂ O ₃ .
Group C	Fe ₂ O ₃ , K ₂ O, CO ₂ , Rb	Distribution within and between pillows is irregular.	Control unclear. Younger geological processes expected.

depleted, the AA-selvages intermediately and the AS-selvages least. The levels of depletion in the individual pillows, however, are more or less similar and coherent, which suggests that there may have been a degree of mineralogical control on the distribution of alumina.

During seafloor metamorphism of MORB, Al₂O₃ is considered by most workers to be immobile (eg: Humphries and Thompson, 1978a; Motl, 1983; Thompson, 1983). Many of these workers, however, document veins and amygdales filled with chlorite which contains appreciable amounts of Al₂O₃, implying that it is mobile under certain conditions. This observation is supported by few recordings confirming its mobility (Cann, 1969; Evarts and Schiffman, 1983; Alt *et al.*, 1986).

Fe₂O₃ is depleted in five of the eight veinlet selvages relative to pillow core. There is a rough tendency for Fe₂O₃ to be depleted in the albitised + silicified selvages and enhanced in the amphibolitised zones suggesting a mineralogical control. The irregular content of ferric iron may further be enhanced by the patchy distribution of epidote on the one hand, and by the common association of sulphide with clinozoisite on the other.

For hydrothermally altered MORB it is found that both (pyrite-poor) chlorite- and

epidote-rich rocks lose iron and that the ferric/ferrous ratio's either remain constant or increase for the former and latter respectively. In pyrite-rich, chloritic rocks Fe^{2+} is gained which lowers the $\text{Fe}^{+3}/\text{Fe}^{+2}$ ratio (Humphries and Thompson, 1978a).

FeO is slightly depleted in the AS-selvages and marginally and moderately enriched in the A- and AA-selvages, respectively. These systematic distributions for the various selvage types within each of the pillows suggest that FeO has a mineralogical control, probably corresponding with the degree of amphibolitisation.

Hyaloclastites are considerably enriched with FeO relative to pillow cores suggesting that Fe^{3+} it is mobilized to the pillow exteriors (i.e. rims and hyaloclastites). This is in agreement with most high temperature submarine alteration studies (Scott and Hajash, 1976) whereas only a few show variable gains and losses (Cann, 1969; Humphries and Thompson, 1978a). The latter authors ascribe gains to the bonding of ferrous iron with S to form sulphide. The study area hyaloclastites are also distinctly sulphide-enriched (Fig. 2.6D; Fig. 3.7a).

MnO enrichment of the cooling-crack selvages relative to cores is consistent, both within and between pillows (except for 180-2c which is marginally depleted). The consistency of MnO enrichment in the selvages despite them being either albite or amphibole dominated shows that the enrichment is not mineralogically controlled. Two of the three pillow rims and all three hyaloclastites are MnO depleted relative to pillow core suggesting an overall scenario of MnO mobilisation from the pillows (Table 3.5). The enrichment of the veinlet selvages relative to pillow cores and pillow exteriors pose somewhat of a problem suggesting, at a glance, that the cores were expelling MnO faster than it could be removed by the veinlet fluids, whereas in the pillow exteriors it was effectively leached.

Petrographic evidence suggests that alteration of the pillow exteriors outlasted that of the pillow cores due to clogging of the cooling-crack veinlets and that temperatures seem to rise toward the waning stages of alteration (to be demonstrated below). On the other hand, Seyfried and Janecky (1985) demonstrate that Mn shows increasing solubility with increasing acidity, which in their experiments is induced at high temperatures (425°C).

It could, therefore, be possible that the pillow exteriors were more effectively leached than the cooling-crack selvages due to a thermally progressive fluid environment.

Humphries and Thompson (1978a) report that MnO of oceanic basalt has a wide range of abundances and, despite the spread, show a tendency for rims to be marginally enriched relative to cores. They also point out that inconsistent trends have been reported more frequently than consistent ones, concluding that MnO is mobilized during hydrothermal alteration and then redistributed within the altered basalts.

MgO is gained by both the A- and AA-selvages but lost by the AS-selvages. MgO is clearly gained by the veinlet walls which have undergone amphibolitisation and lost by those that are silicified and albitised, exhibiting a mineralogical control for its distribution.

Although the addition of seawater-MgO to basalt is probably the best-known attribute of basalt-seawater interaction (Mottl, 1983) it does not happen to all altered basalt. Seyfried *et al.*, (1988) suggest that circulating fluids may become MgO-depleted due to continued consumption and that fluids thereafter could become progressively sodium- and/or calcium enriched. Exposure to such evolved fluids could easily check or suppress MgO-addition to basalt.

CaO in the AS-selvages is depleted by up to 4 wt%, the A-selvages are moderately enriched and the AA-selvages are either marginally enriched or -impoverished relative to the pillow core. The analyses of the latter selvages seem to reflect the result of the composite alteration (Fig. 3.6B), i.e. somewhere between the two end-member types which are represented by the AS- and A-types. Enrichments and depletions of CaO, however, seem linked to amphibolitisation and albitisation respectively, suggesting a mineralogical control.

During the initial alteration of oceanic basalt CaO is exchanged for seawater MgO (Hart, 1973; Bischoff and Dickson, 1975; Humphries and Thompson, 1978a; and Mottl, 1983). During deeper circulation Mg^{2+} becomes depleted and fluid temperatures rise, after which Ca^{2+} addition to basalt may take over (Seyfried *et al.*, 1988). CaO-depletions or -

additions to basalt thus occur depending on the evolved state of the fluids and the fluid temperature.

Na₂O is systematically enriched relative to pillow core at the AS-veinlet selvages and moderately depleted at the A-selvages. The AA-veinlet wallrocks are either marginally enriched or depleted which essentially reflects the result of the previous two alteration types (Fig. 3.6B) implying that sodium distribution of the selvages is mineralogically controlled.

Humphries and Thompson (1978) show that pillow cores and rims of altered oceanic basalt have inconsistent sodium trends. Basalt-seawater experimentation show that albite is fixed at low- and removed at moderate water/rock ratios (*ca.* 10, Mottl, 1983). Seyfried *et al.* (1988) emphasize that localized effects inducing low Mg/Na fluid ratio's bring about albitisation, rather than mere water/rock ratios.

The observations here not only confirm the variable tendencies observed for natural and experimentally altered basalt but show that a progression from Na-depletion to Na-enrichment can occur within a single pillow (MC180, Fig. 3.6B). This reflects an evolutionary trend for hydrothermal fluid development within the study area consistent with the findings of both Mottl (1983) and Seyfried *et al.* (1988).

K₂O is depleted in five of the eight cooling-crack selvages suggesting an overall but inconsistent loss from the pillows. For MORB Humphries and Thompson (1978a) find no consistent trends for the distribution of potassium between the cores and rims of their samples. Experimental studies (e.g. Mottl, 1983; Rosenbauer and Bischoff, 1983) and other reports of naturally altered basalt (e.g. Melson and Van Andel, 1966) show potassium to be leached from the rock. Depleted levels are thought to be maintained for the duration of high-temperature alteration, with possible addition later during low-temperature alteration (Scott and Hajash, 1976).

Cr₂O₃ data have a large spread which is probably related to an inhomogeneous distribution of chromite microphenocrysts (*cf* Smith, 1980). Despite this, the veinlet selvages show consistent enrichments relative to pillow cores (Fig. 3.10). Pillow rims and

most hyaloclastite are also enriched relative to the veinlet selvages which suggests either mobilisation of chromium from the core toward the exteriors (discussed earlier, Fig. 3.9) or enrichment from the exterior inward. The first alternative would mean that the pillow cores discharged chromium to the cooling crack selvages at a rate greater than it could be removed by the veinlet fluids and the veinlets would in turn be introducing Cr_2O_3 into the interpillow areas at a rate greater than it could be removed by the fluids (otherwise the progressive outward increase in concentrations could not be accounted for assuming that the core is being leached). The same distribution could be modelled by invoking " Cr_2O_3 mineralisation" of the pillow from an external source. Such a model could not be supported by the leaching of chromium from other komatiitic basalt pillows as they consistently show the same enrichment patterns (Fig. 3.9). There is, however, a large volume of ultramafic rocks, containing in excess of 2200 ppm Cr (Smith, 1980, Table 4.2), which are interleaved with the pillows in question, which contain ≤ 700 ppm Cr (Smith, 1980, Table 2.2). It is thus conceivable that the circulating fluids could have redistributed the element within the volcanic pile enriching rocks with lower Cr values. Smith (1980) also reported clinopyroxene containing between 0.55 - 0.80 wt% Cr_2O_3 , and actinolite containing as little as 0.01 wt% Cr_2O_3 suggesting that considerable amounts of the element were expelled during hydration of the primary minerals (cf also Table 2.4 and Appendix: AMPHIBOLE). Smith (1980) also considered chromium to be a relatively mobile trace element.

Humphries and Thompson (1978b) show that due to the wide range in the chromium content of fresh MORB it is difficult to observe subtle changes which may be due to hydrothermal alteration. The authors show that there are no consistent trends within altered oceanic basalts except for a few instances where the rims of pillows are exceptionally enriched relative to their cores.

The consistent enrichments of the pillow selvages observed here seem to favour a redistribution of chromium within the volcanic pile, and the source is considered to be the closely associated ultramafic lavas. This view is supported by (1) the absence of similar chromium enrichments for MORB where ultramafic volcanics are not developed; and (2) evidence that igneous phases such as pyroxene lose considerable amounts of Cr upon hydration to amphibole. The lack of correspondence between chromium

enrichment and selvage alteration supports the absence of a mineralogical control and favours rather an external and/or physicochemical control for the distribution of the element.

H₂O. The cores of the pillows are more hydrated than their associated veinlet wallrocks, irrespective of alteration type. The AS-type selvages show the lowest degree of hydration which may be linked to the presence of anhydrous minerals (albite and quartz) that characterise the alteration. The A- and AA-type selvages consistently contain slightly less crystalline water than the cores. The fact that the pillow exteriors are generally more hydrated than the cores imparts a typical seafloor trend (Humphries and Thompson (1978a), which may be attributable to the added presence of chlorite which is more hydrous than amphibole. A mineralogical control is thus evident for the distribution of water.

CO₂ contents of pillows MC150 and MC180 are very low, ie. ≤ 0.4 wt% for all textural domains, whereas MC60 has as much as 1.2 wt%. For the latter sample there is petrographical evidence supporting a late introduction of CO₂. All the veinlet margins, on the other hand, have no consistent enrichment or depletion trends relative to their cores. Of significance here is not so much the inconsistent trends but the very low overall abundances measured for MC150 and MC180 and the observed late paragenetic relationship of carbonate veining in MC60.

The CO₂-content of altered basalt dredged from oceanic ridges (0 - 3 Ma) is not normally reported in the literature. Several workers show that the small amount of primary (igneous) CO₂ the rocks contain is leached during high-temperature basalt-seawater interaction (Mottl and Holland, 1978; Seyfried and Bischoff, 1981). The seemingly insignificant role CO₂ has in MOR-vent fluids is further demonstrated by its exclusion from being listed as an analysed species (Von Damm *et al.*, 1985). Only basalts that are sampled off-axis which are older than ~3 Ma are noticeably enriched in CO₂ suggesting that its addition to basalt occurs relatively late and in a low-temperature environment (Staudigel *et al.*, 1981).

The degree of carbonatisation of the pillows in the statically metamorphosed area is very

low and where it does occur it appears unrelated to the main alteration of the cooling-crack selvages. Carbonate addition to the statically metamorphosed parts of the Komati Formation occurs along the Hooggenoeg fault (Williams and Furnell, 1979) which displays a late and structural association. The absence of CO_2 from seafloor-related alteration in the static-style area suggests, firstly, that the pillows have been exposed to high-temperature seafloor metamorphism and secondly, that they have largely escaped pervasive low-temperature CO_2 -addition. This, however, is a common trend in the pillowed flows of the overlying Hooggenoeg Formation.

TiO_2 distribution between the various pillow cores is remarkably consistent (Fig. 3.10). Relative to the cores the veinlet selvages seem only marginally but consistently enriched with the exception of samples MC602c and MC1802b. Deviation of MC602c from the mentioned trend is small and close to the analytical margin of error but sample MC1802b shows a large titanium depletion. The other elements, except for Cr_2O_3 and to a lesser extent K_2O and Rb, show no corresponding distributions which seem to excluded the possibility of attributing the anomalies to analytical error. The chromium anomaly, 5 times higher than the core, seems to suggest that an above average number of chromite grains may have been present in the sample. Chromite, however, has been described previously as often clinging to amygdalae (ie. to melt vesicles) as large single grains or grain clusters, suggesting also the presence of an amygdale. If true, the infilling of the amygdale with hydrothermal minerals characteristic of the main alteration type would not notably disturb the relative abundances of the major elements except TiO_2 which is relatively immobile. The relative immobility of TiO_2 compared with the other major elements infilling such a vesicle would have the effect of diluting the original TiO_2 content of the sample. The unusual composition of the sample is thus attributed to a local inhomogeneity of the sample, variable mobility of elements and the small mass of the hand-picked samples.

TiO_2 fluctuations between cores and selvages are small and fall within the analytical error range suggesting that an evaluation of its mobility is not possible (Fig. 3.10). It should be kept in mind, however, that the overall trend for this element is to be mobilized towards the pillow rims as is demonstrated in Fig. 3.9. Most workers, however, regard TiO_2 immobile during hydrothermal alteration of basalt (Mottl, 1983; Thompson,

1983; Humphries and Thompson, 1978a) whereas the latter authors (contrastingly) also note increases in the rims of their pillow samples. It thus appears that TiO_2 mobilisation during incipient alteration, as shown by the pillow core/veinlet selvage distributions, can be expected to be negligible, but that significant alteration may result in the exterior portions of pillows where greater hydrothermal activity occurs.

Y contents of the AS- and A-veinlet selvages are slightly depleted whereas those of the AA-selvages are slightly enriched, relative to pillow core. Y, however, is generally considered to be an immobile element during hydrothermal alteration (Pearce and Cann, 1973; Humphries and Thompson, 1978b; Smith, 1980). The observed Y depletion of the cooling-crack selvages of MC60 may be real and/or the result of dilution which has resulted from silica addition. The small Y enrichments of the compositely altered AA-veinlet margins (Fig. 3.10, MC180) do not exhibit an intermediate result relative to the end-members, as has been found for most other elements, suggesting that factors other than mineralogy have influenced its relocation. This may be linked to the fact that Y is concentrated into residual melts during crystallization and inclined to fractionate into the mesostasis of glassy rocks (Humphries and Thompson, 1978b). Varying degrees of crystallisation versus chilling of pillows would bring about varying amounts of mesostasis which would be exposed to alteration alongside the thermal tension cracks. As noted earlier, it is apparent that the mesostasis phases are preferentially attacked during cooling-crack selvage alteration, implying that more glass-rich pillow cores could release Y more readily than their more holocrystalline counterparts. This might explain the variable rim compositions the pillows have and the somewhat depleted Y-levels of the metahyaloclastites of MC150 and MC60 (Table 3.5).

Zr-content of the pillow cores is remarkably consistent, whereas most of the veinlet-selvages appear marginally enriched relative to them. Notable enrichments occur for 60-2a and 60-2b. These enrichments seem significant as they counteract the dilution effects of silica addition (Fig. 3.10).

Zr is also fractionated into residual liquids during (fractional) crystallization and thus is concentrated into the mesostasis of hyalocrystalline rocks. The appreciable enrichments of Zr in the AS-selvages may suggest that redistribution within the pillow has occurred

due to the vulnerability of mesostasis to alteration. The degree of redistribution of Zr in MORB during ocean-floor metamorphism is thought to be negligible (Humphries and Thompson, 1978b; Mottl, 1983; Thompson, 1983).

Sr is depleted in virtually all the veinlet selvages relative to the cores (Fig. 3.10). The degree of depletion is also larger than should be expected from dilution caused by other element additions. The distribution of Sr between pillow cores and pillow exteriors (Table 3.5) shows that most of the rim and hyaloclastite samples are appreciably depleted. One pillow rim, i.e. MC180, which is extensively silicified and albitised and locally clinozoisitized, shows Sr enrichment. The depleted veinlet selvages and interpillow samples implies that Sr is generally leached from the fluid-exposed pillow parts, except when they contain epidote-clinozoisite.

Earlier workers show that Sr-depletions of seafloor altered basalts are sympathetic to Ca-depletions (Melson *et al.*, 1968; Cann, 1968). Humphries and Thompson (1978b) point out that altered basalt is not always depleted in Sr, especially not when epidotised, suggesting that uptake from circulating fluids can also occur. Reasonably good correlations between Sr and Ca for pillow rims lead the latter workers to conclude that Sr concentration is predominantly controlled by the same chemical processes as Ca.

Rb is inconsistently distributed between pillow cores and the adjacent veinlet walls. The Rb contents of the komatiitic basalts, however, are very low and close to the detection limit of the analytical technique used. The distribution patterns for Rb closely mimic those of K suggesting a similar control for the elements. Rb distributions for hydrothermally altered MORB are not commonly reported in the literature.

Barium (data set incomplete due to insufficient sample) shows enrichments for the AS- and A- selvages but depletion at the AA-veinlet margins. Evaluation of whole rock Ba data (Table 2.1) show that pillows with enriched cooling-crack selvages also have enriched pillow rims and that the pillow with the depleted cooling-crack selvages correspondingly has a depleted rim (ie. MC180). The pillow core compositions, relative to each other, are quite variable. A large spread for Ba in the komatiitic basalt pillows of the study area has previously been demonstrated by Smith (1980).

Humphries and Thompson (1989) show Ba concentrations of fresh ocean basalts to be low, close to analytical limits, and that values of altered basalts fluctuate threefold compared with fresh samples. They concluded that Ba is not greatly affected by hydrothermal alteration although their data suggest that some Ba is mobilized to enrich pillow interiors.

Although incomplete, the data set for Ba shows that its distribution within the individual pillows is systematic. The compositely altered AA-veinlet selvages, however, do not reflect intermediate values between those of the end-member AS- and A-selvages suggesting that the element is not mineralogically controlled. As control on Ba distribution is both uniform within individual pillows but with no apparent mineralogical control it seems that it may be linked to other physicochemical controls.

3.4.3 Element Fluxes

Element fluxes between pillow cores and veinlet selvages have been calculated from the major and trace element data listed in Table 3.5. The element fluxes are listed in Table 3.7. The calculations were made by (1) assuming the least-altered samples to be the pillow cores and (2) assuming constant TiO_2 according to the method of Grant (1986) and an additional procedure involving mineral compositions. Assumptions made and the additional mineral composition method are discussed below. Having assessed these factors, the flux calculations are made by normalizing the compositions of the altered pillow domains to constant TiO_2 of the respective pillow cores, similar to the method used by Humphries and Thompson (1978).

3.4.3.1 Selection of least-altered sample. Pillow cores and outer cores were initially selected as the textural domains representing the least altered material as both showed excellent textural preservation and lowest degrees of hydration (1.67 - 3.10 wt% H_2O). However, selection of the outer core as the least-altered sample resulted in the element fluxes of samples being inconsistent with the petrographically observed degrees of alteration whereas selection of pillow core resulted in element fluxes corroborating the petrographic observations.

3.4.3.2 Selection of immobile elements. Geochemical methods for studying the changes

TABLE 3.7 ELEMENTAL FLUXES: GAINS AND LOSSES FOR HYDROTHERMALLY METAMORPHOSED KOMATIITIC BASALT ASSUMING CONSTANT TiO_2 (expressed in g/100g).

SAMPLE NUMBER	HC060 2A1	HC060 2B1	HC060 2C1	HC060 03	HC060 04	HC060 06	HC150 2A1	HC150 2B1	HC150 03	HC150 04	HC150 06	HC180 2A1	HC180 2B1	HC180 2C1	HC180 03	HC180 04	HC180 06
COMMENT	CCWALLR.	CCWALLR.	CCWALLR.	OUT.CORE RIM		HYALOCCL.	CCWALLR.	CCWALLR.	OUT.CORE RIM		HYALOCCL.	CCWALLR.	CCWALLR.	CCWALLR.	OUT.CORE RIM		HYALOCCL.
SiO ₂	5.63	3.48	16.72	-1.07	-1.26	13.12	-3.29	-2.23	1.09	-8.89	10.22	-5.08	25.40	-2.50	-1.64	-5.37	-14.42
Zr	5.23	2.46	9.53	-2.89	4.50	-13.41	0.27	1.12	1.89	-0.01	-8.41	-2.55	37.33	0.22	-4.64	1.07	-11.99
TiO ₂	0.00	0.00	0.00	0.00	0.05	0.00	0.00	0.00	0.00	-0.00	-0.00	0.00	-0.00	0.00	0.00	0.00	-0.00
Al ₂ O ₃	-1.62	-1.05	1.12	-0.38	0.00	1.53	-4.37	-3.70	-1.28	0.59	3.07	-2.70	2.43	-0.70	-2.03	1.21	-5.11
Fe ₂ O ₃	-1.04	-0.89	-1.74	-0.90	-1.19	0.56	0.69	-0.14	0.31	1.11	2.89	1.51	2.32	-0.76	1.14	-0.90	-1.07
FeO	-0.64	-0.45	0.18	-0.36	-1.23	6.57	1.59	2.45	0.43	0.47	6.13	-0.25	6.51	1.56	-0.39	-2.70	-1.98
MnO	0.07	0.08	0.10	-0.09	-0.09	-0.09	0.07	0.05	-0.03	0.04	0.03	0.01	0.15	-0.01	0.07	-0.07	-0.10
HgO	-1.48	-1.58	-2.04	-1.13	-2.75	4.11	1.57	1.72	0.45	0.60	6.58	1.57	8.12	0.73	0.15	-3.41	-3.11
CaO	-4.33	-4.72	-3.95	-3.09	-5.78	-0.65	3.22	2.56	1.15	-2.16	2.86	0.74	7.55	-1.12	0.31	-3.42	-4.30
Na ₂ O	0.85	1.28	1.91	-0.02	0.69	-0.67	-1.45	-1.31	-0.68	-1.44	-1.23	-0.47	0.46	0.31	-0.40	0.89	-0.77
K ₂ O	-0.01	0.04	0.20	0.10	0.10	-0.01	-0.09	-0.09	-0.06	0.42	0.44	-0.22	0.43	-0.21	-0.11	0.23	-0.15
P ₂ O ₅	-0.11	-0.11	-0.11	0.02	-0.01	0.04	-0.17	-0.16	-0.02	0.04	0.02	-0.22	-0.21	-0.22	-0.03	-0.01	-0.12
Cr ₂ O ₃	0.08	0.05	0.02	-0.01	0.05	0.34	0.09	0.05	0.01	0.05	0.16	0.03	0.25	0.02	-0.02	0.00	0.08
CO ₂	-0.61	-0.03	-0.10	0.57	0.64	3.21	0.31	0.18	0.08	-0.03	0.04	-0.01	0.31	-0.13	-0.15	-0.17	-0.14
H ₂ O	-0.78	-0.69	-0.27	-0.81	-0.00	3.03	-0.14	-0.11	-0.07	1.40	2.29	-0.10	0.68	-0.28	-0.05	-0.13	-0.34
S	51.59	111.04	105.29	-9.98	678.76	5379.67	-90.32	-108.40	-105.56	-98.77	1317.19	-50.86	113.37	-42.48	7.12	464.34	642.57
Cl	11.26	10.42	-4.71	-5.27	-2.16	73.33	15.92	5.11	-14.39	2.08	11.69	5.76	32.41	1.95	0.97	-19.05	-32.40
Br	4.44	0.32	31.06	8.33	48.72	729.30	-12.92	5.78	-14.06	101.05	592.19	-0.49	108.39	7.45	-8.11	17.44	100.36
Zn	4.46	3.24	10.82	6.52	-7.28	67.48	25.82	29.36	11.39	18.59	73.41	24.09	50.00	10.12	10.12	-14.94	-5.83
Cu	2.92	39.44	51.41	-50.99	321.19	323.93	-76.97	-74.03	-64.84	14.12	284.26	-61.93	112.02	-48.01	-23.70	189.61	448.16
Rb	0.87	0.81	0.47	0.69	0.55	-0.26	-1.14	-0.08	-1.94	9.47	12.41	-5.24	9.07	-4.19	-2.25	5.45	-2.04
Sr	-52.46	-57.97	-46.53	-37.63	-49.70	-71.87	-63.41	-48.42	-15.11	-55.24	-62.22	-57.76	96.02	-25.11	-47.48	27.28	-96.55
Y	-3.41	-2.65	-1.12	0.70	-2.60	-1.56	-1.49	-1.25	0.27	-1.71	2.04	-0.77	10.33	0.45	-0.52	-0.82	-4.52
Nb	0.95	1.89	1.24	-1.00	-0.09	-1.00	-0.08	-1.03	-3.00	-1.18	-3.00	-1.08	0.09	-0.08	-1.05	0.46	-2.31
Hg	0.07	-0.01	0.24	-0.10	-0.04	-0.14	0.10	0.09	0.10	-0.10	-0.10	0.07	0.16	0.12	0.06	-0.07	-0.07

in volume and concentration of metasomatically altered rocks have been previously described (e.g. Gresens, 1967; Grant, 1986). The method selected here for estimating the least-mobile elements in the metasomatised veinlet selvages described above is the "isocon" method of Grant (1986). This is done by plotting the concentrations of the altered samples against those of the least-altered samples. The immobile components are thought to generate a straight line through the origin of the graph with a slope reflecting the ratio of equivalent masses before and after alteration (i.e. M^o/M^a of Grant, 1986). This line is called the isocon and is determined by inspection of the graph to find the best fit line through a series of points.

The isocon method is used on two veinlet selvage samples representing the end-member alteration types of the komatiitic basalt pillows to determine the least-mobile elements (Fig. 3.11). From the figure it is evident that TiO_2 and Zr are the only components that generate a straight line through the origin of each graph, thus defining the isocon. The other elements clearly move to opposite sides of the isocon depending on the alteration type (i.e. Na_2O , FeO , SiO_2 , CaO , and MgO). Al_2O_3 , MnO and Y remain on the same side of the line but vary within the field with Y being the least affected. The isocons in Fig. 3.11 have slopes close to unity (or ratio of equivalent masses before and after alteration) suggesting that small mass decreases of between 2 - 3% have occurred. Flux calculations, therefore, can be done on the basis of assuming either constant concentration of TiO_2 or Zr or constant mass (cf Grant, 1986).

Selection of TiO_2 above Zr as the "best" least-mobile element, is based on evaluation of the element flux values in terms of the observed alteration petrography and by comparing element flux ratios with the ratios required for the growth of the alteration products. Mineral compositions resulting from the alteration have been determined by electron microprobe and are listed in the respective mineral appendixes. As Al_2O_3 is most often used in MORB studies as the least-mobile element (e.g. Humphries and Thompson, 1978a; Mottl, 1983; Thompson, 1983), it is also tested below (Fig. 3.12) with TiO_2 and Zr to show its suitability to serve as basis for flux determinations.

Normalization to constant TiO_2 was selected for the following reasons:

- 1) Normalization to constant TiO_2 (Fig. 3.12b) of the pillow core produces the

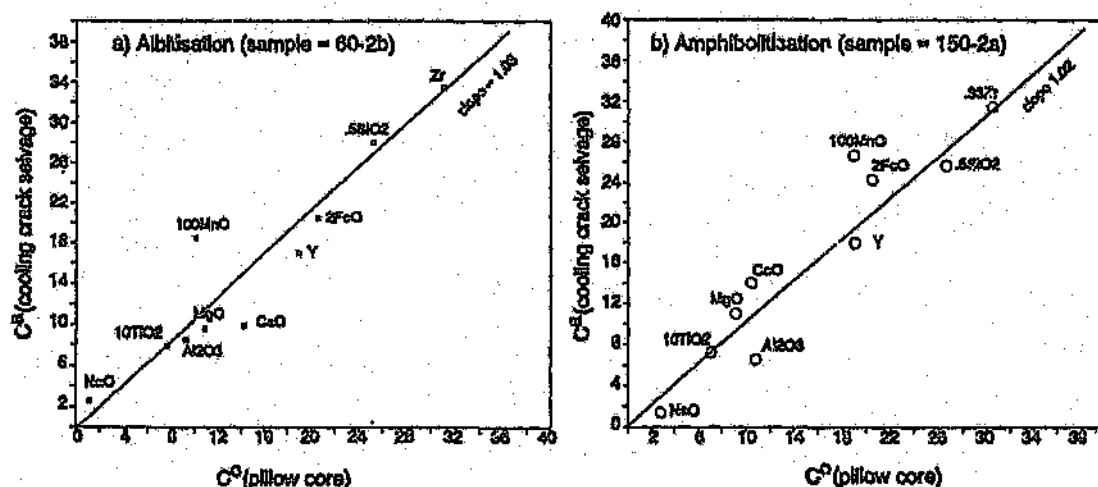


Fig. 3.11 An isocon diagram for albitised (a) and amphibolised (b) cooling-crack selvages of pillow cores MC60 and MC150, respectively. Oxides are plotted in wt% and traces in ppm according to the indicated scaling factors. The isocon is selected to pass through constant TiO_2 and Zr as for each alteration type most of the major elements switch sides. C^A and C^O = element concentrations of altered and least-altered samples, respectively.

correct relative positioning of the samples. For example, the expected least-altered outer core samples plot closest to the origin of the graph, then the veinlet selvages and then the most altered rims and hyaloclastites. Due to the observed petrographic mobility of titanite at albitised pillow rims (Fig. 3.5A) and the analysed migration of TiO_2 towards the pillow rims (Fig. 3.9) it is to be expected that the determined fluxes for these parts is less accurate than those of the cooling crack selvages (where evidence for titanite mobility was not found). For the constant Zr-plot (Fig. 3.12c) a similar correct positioning is indicated, although outer-core compositions are further from the origin when compared to the TiO_2 -normalised plot. Due to the somewhat poorer positioning and Zr being mobilised in some of the selvages as previously mentioned (Fig. 3.10), TiO_2 was selected for normalisation.

- 2) Flux calculations assuming constant Al_2O_3 are unsuitable for the reasons that the location of samples along the regression line of the Al_2O_3 normalised plot (Fig. 3.12a) appear in a nonsystematic sequence in relation to the expected amount of

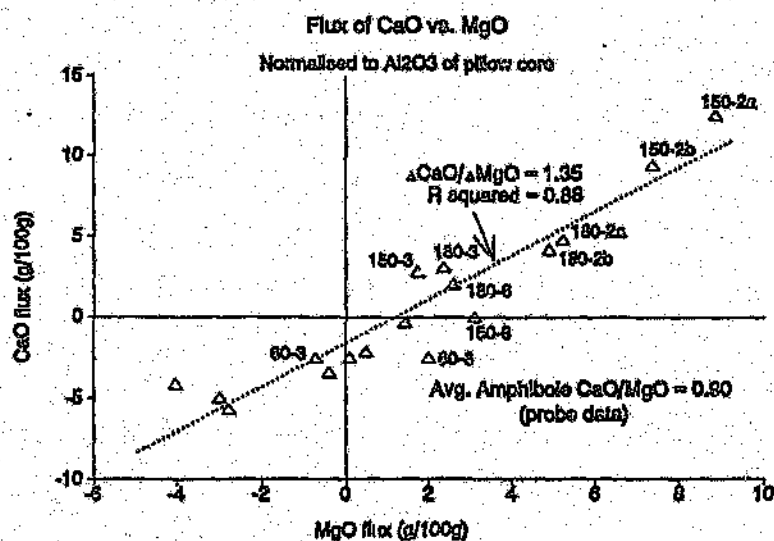
flux for the samples. The outer-core samples, which are petrographically the least altered, plot far removed from the origin of the graph and in fact lie close to the hyaloclastite. Those expected to be second least-altered, the cooling-crack selvages, are shown to be more enriched or altered than their respective rims and hyaloclastites. These inconsistencies are best explained by assuming a loss of Al_2O_3 from the zones of amphibolitisation.

- 3) The positive rate of change for CaO/MgO normalised to Al_2O_3 (Fig. 3.12) has a steeper slope and a better correlation coefficient than normalised to TiO_2 and Zr, respectively. However, mineralogically, all the samples in the positive quadrant, i.e. those that have gained both CaO and MgO, predominantly consist of amphibole. The CaO/MgO ratios of amphibole determined from the electron microprobe data (Appendix: AMPHIBOLE) are about ~ 0.90 which is similar to the slope of the regression curve in the TiO_2 - and Zr-normalized plots but not to that of Al_2O_3 .

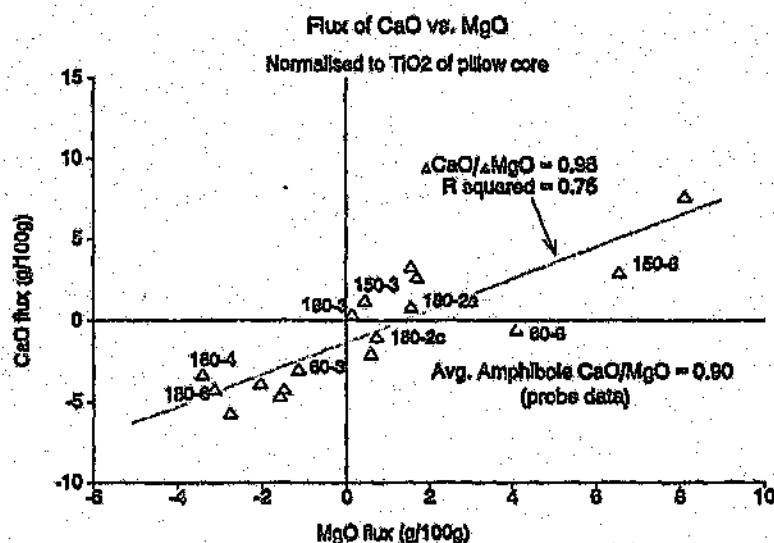
3.4.3.3 Element flux relationships. Detailed element flux relationships for the major elements are shown in Figures 3.12, 3.13, 3.14 and 3.15 and are summarised and compared with those from MORB in Table 3.8. To assist in the identification of sample origin, suffixes to the respective sample numbers denote textural domains, i.e. -2 = cooling crack selvage; -3 = outer core; -4 = pillow rim; -6 hyaloclastite. For evaluation of the flux relationships, sample origin with respect to textural domain is considered important as more significance should be attached to samples with pillow interior origins as opposed to pillow exterior origins. The reason for this is the previously mentioned petrographic and geochemical evidence for the (fractional) mobilisation of titanite in the exterior portions of pillows (Fig. 3.5A; Fig. 3.9; Fig. 3.10).

CaO/MgO. Hydrothermal alteration of the komatiitic metabasalts has brought about a positive rate of uptake of CaO and MgO as indicated above (Fig. 3.12). The dominant alteration products for the cooling-crack selvages are actinolitic amphibole or albitic plagioclase, whereas for pillow rims it is quartz, albite or chlorite, Mg-rich amphiboles (tremolite and cummingtonite) and Ca-amphibole. Veinlet wallrocks showing gains of CaO and MgO have formed amphibole at the expense of plagioclasic mesostasis whereas those that have lost CaO and MgO have increased modal albite at the expense of

a)



b)



c)

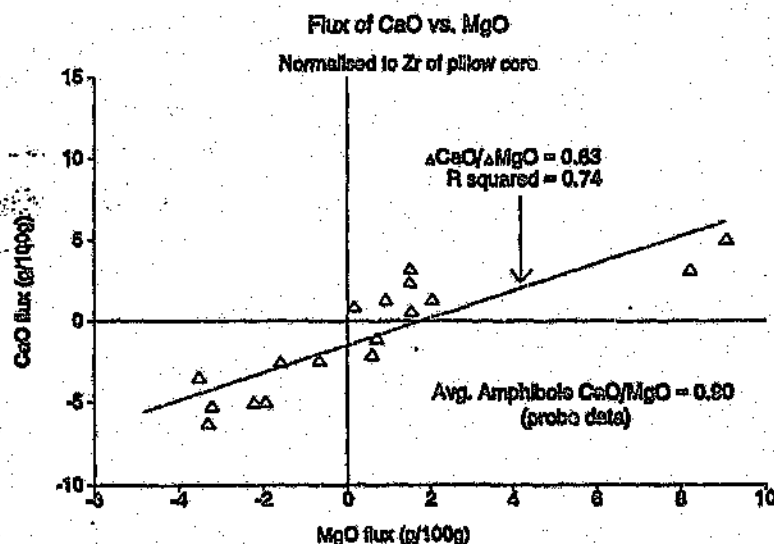


Fig. 3.12

Plots of CaO vs MgO normalised to constant Al₂O₃ (a), constant TiO₂ (b) and constant Zr (c) of pillow cores. See text for discussion.

amphibole. The gains and losses demonstrate a clear mineralogical control and may be represented by reaction (3) to be discussed below.

In contrast, hydrothermally metamorphosed oceanic basalt shows a negative relationship for these elements as seawater-Mg is consumed by basalt at the cost of Ca (on approximately a 1:1 molar basis) and the production of H^+ . This leads to the formation of chlorite and increased acidity of the circulating fluids and has been demonstrated for both natural and experimentally altered basalt to be one of the main characteristics of high-temperature basalt-seawater interaction (Mottl and Holland, 1978; Humphries and Thompson, 1978a; Mottl, 1983; Thompson, 1983). Rosenbauer and Bischoff (1983), for example, demonstrate the chemical and mineralogical changes during Mg-uptake at the expense of Ca and the production of chlorite and H^+ .



For komatiitic basalts, however, Mg-enrichment appears not to be manifested by chlorite as readily as in case of tholeiite. Incipient alteration, as demonstrated by the cooling-crack selvages, starts off in response to Mg-addition with the formation of amphibole and only in more advanced stages, as shown by some pillow rims, does chlorite and or Mg-rich amphibole appear.

H₂O/MgO. The uptake of H₂O vs. MgO during hydrothermal alteration is demonstrated by Fig. 3.13a. The rate of change of the components H₂O and MgO yields two broad positive relationships with one slope of *ca.* 0.4 and a second slope of 0.12. The corresponding average values of the microprobe data (Appendix: AMPHIBOLE and CHLORITE, in which the H₂O content was calculated from the stoichiometry) show amphibole and chlorite to have H₂O/MgO ratios of ~0.12 and 0.64, respectively. The flux ratios of the samples plot between the mineralogical "end-member" values with a strong bias toward the amphibole ratio, indicating the cooling-crack selvages comprise mainly amphibole whereas the externally-situated samples contain additional chlorite. This emphasizes the observation that the MgO-enriched external samples do not exclusively consist of chlorite, as is the case for tholeiite. Those samples that show dehydration and MgO losses represent pillow portions that have become albitised and

silicified.

For hydrothermally altered oceanic basalt a similar positive correlation for H_2O/MgO is documented by Humphries and Thompson (1978). Their average slope is steeper, indicating the uptake of larger quantities of H_2O relative to MgO , which confirms the preponderance of chlorite in their samples.

Na_2O/MgO . The flux of sodium versus magnesium between cores and cooling-crack selvages is negative (Fig. 3.13b). Mg-addition to the veinlet selvages, as indicated above, has resulted in amphibolitisation whereas gains in Na_2O have resulted in albitisation. The outliers of the plot are mostly pillow-rim and hyaloclastite samples suggesting a complex alteration history and/or redistribution of TiO_2 . Note the veinlet selvage sample (MC1802b1) with the anomalous TiO_2 and Cr_2O_3 values (related to the possible presence of an amygdale, as discussed earlier) also plots as an outlier.

The corresponding slope for hydrothermally and experimentally altered MORB is also negative, corroborating the findings shown here. For the oceanic basalts, however, there is an initial stage of concomitant sodium and magnesium addition, i.e. up to water/rock ratios of about 10 (which is equal to additions of $\leq 2g\ MgO/100g\ rock$), after which Na_2O is lost and MgO gained. For the study area such an initial uptake of both Na_2O and MgO is not evident. Alteration may have proceeded beyond such initial stages for the concomitant addition of sodium and magnesium. The range of Na_2O flux for the komatiitic basalt nonetheless, is virtually identical to that of altered MORB, i.e. -2 to +1.5 vs -2.5 to +1.5 (g/100g).

FeO/MgO . The flux of FeO vs MgO for the cooling-crack selvages is positive implying similar addition or depletion of the elements (Fig. 3.13c). Regression of the data gives $R^2 = 0.87$ and a slope of 0.82, which is close to the microprobe data average for amphibole = 0.84 (Appendix: AMPHIBOLE). This suggests an amphibole controlled elemental uptake. Samples showing negative FeO and MgO fluxes suggest a diminution of amphibole due to albitic and silicic alteration.

Hydrothermally metamorphosed oceanic basalts show trends and flux ranges similar to

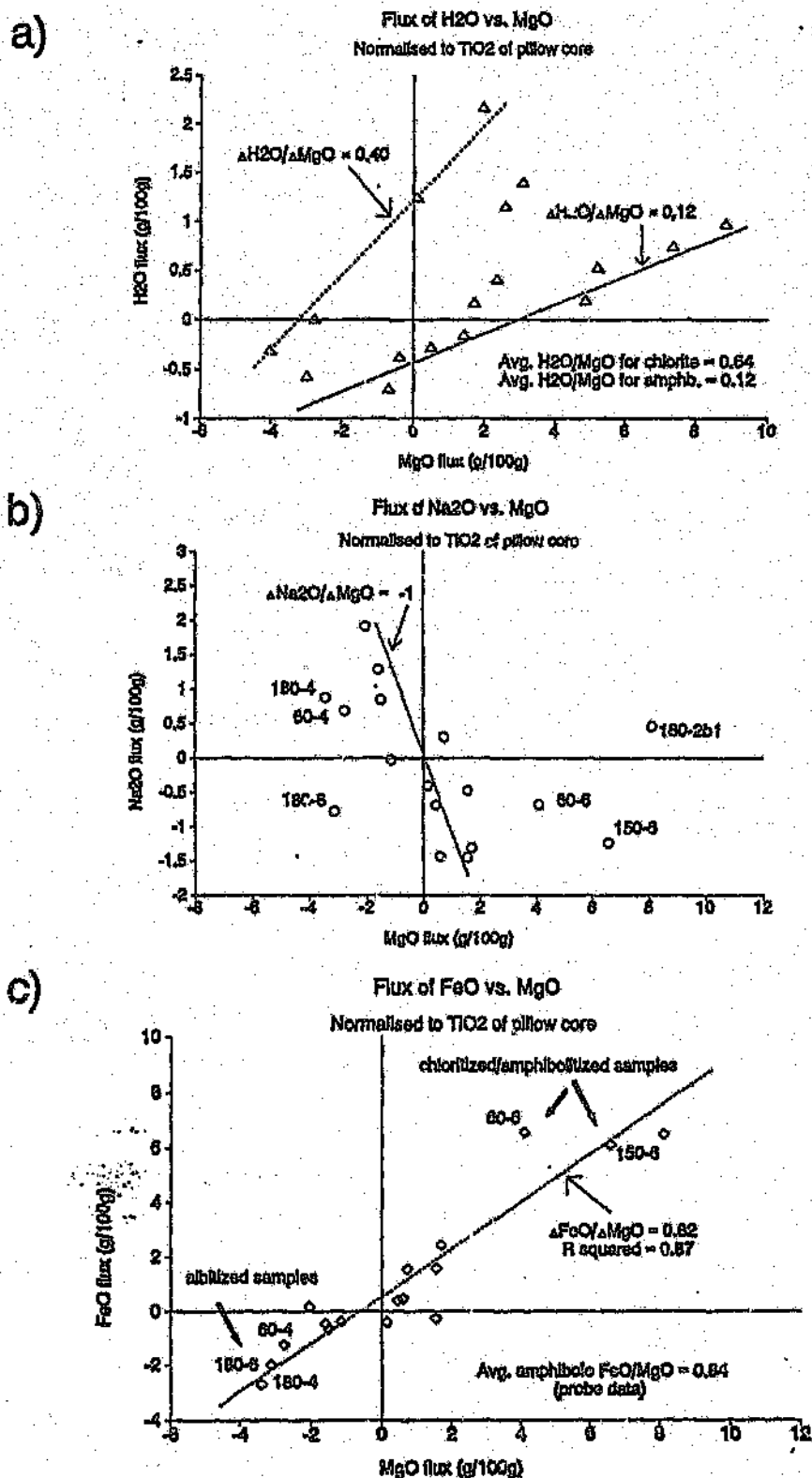


Fig. 3.13

Flux of H₂O vs MgO (a), Na₂O vs MgO (b) and FeO vs MgO (c) all normalised to constant TiO₂ of pillow cores respectively. See text for discussion.

the komatiitic basalt pillows studied here (ie. komatiitic basalt = -3 to +7g FeO/100g vs -4 to +8g MgO/100g and MORB = -4 to +6g FeO/100g vs -2 to +7g MgO/100g in Fig. 3.13c and Mottl (1983), respectively). For MORB-alteration ferrous iron is thought to be supplied by local redistribution within the rock/fluid system whereas Mg is taken up from circulating seawater (Mottl, 1983). In the komatiitic basalts it is anticipated that FeO is supplied by the mesostasis of the pillow core (*cf* reaction 3; Fig. 2.5C and 2.5D).

$\text{Al}_2\text{O}_3/\text{MgO}$. The flux of alumina versus magnesium at the alteration selvages is negative. The insert-diagram (Fig. 3.14a) offers a possible explanation for the observed relationship. On that figure $a' = \pm 0.33$, represents the rate at which the addition or removal of amphibole (average ratio from microprobe data, Appendix: AMPHIBOLE) would increase or decrease the flux of $\text{Al}_2\text{O}_3/\text{MgO}$ of the pillows samples. Likewise, $b' = \pm 0$, represents the rate at which the addition or removal of the average albite of the pillows, would increase or decrease the $\text{Al}_2\text{O}_3/\text{MgO}$ flux of the samples. The insert figure shows that by adding amphibole (+ a') for example, albite has to be removed at a rate of - b' , to satisfy the observed relationship (- c'). Such reactions, however, imply redistribution of alumina on an interpillow scale as some pillows show only albitisation or only amphibolitisation.

Although Cann (1969) and Evarts and Schiffman (1983) report Al_2O_3 removal from basalt, the element is generally considered to be stable during hydrothermal alteration of MORB and is usually utilised as an immobile element in flux determinations (Humphries and Thompson, 1978a; Mottl, 1983; Thompson, 1983).

SiO_2/MgO . At first glance there seems to be no obvious correlation between the flux of silica versus that of magnesium, however, the wide distribution of data suggest that both elements were mobilized, possibly independently of each other (Fig. 3.14b). On closer inspection it would seem that a positive trend exists. The positive trend, however, is largely the result of pillow exterior sample fluxes, which for reasons previously mentioned are not as reliable as those of the pillow interior samples. If the pillow exterior samples (marked by sample numbers) are disregarded, a grouping of data points with a negative trend becomes apparent (marked by the stippled line in Fig. 3.14b).

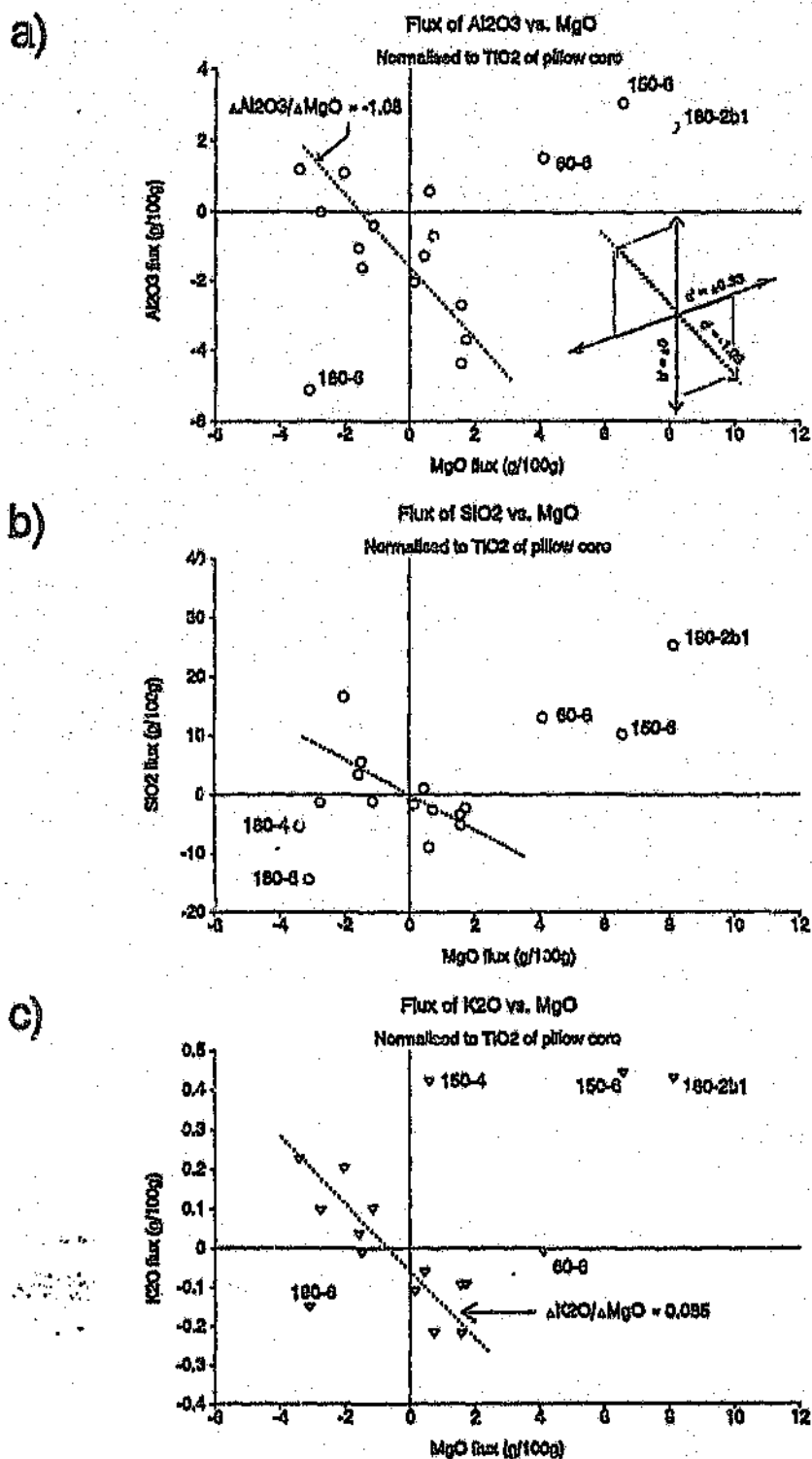


Fig. 3.14

Flux of Al_2O_3 vs MgO (a), SiO_2 vs MgO (b) and K_2O vs MgO (c) all normalised to constant TiO_2 of pillow cores respectively. Note that a positive trending regression line could be modeled through the marked sample points (180-6, 60-6, 150-6, 180-2b1). The marked samples, however, have disturbed TiO_2 values as previously mentioned. See text for discussion.

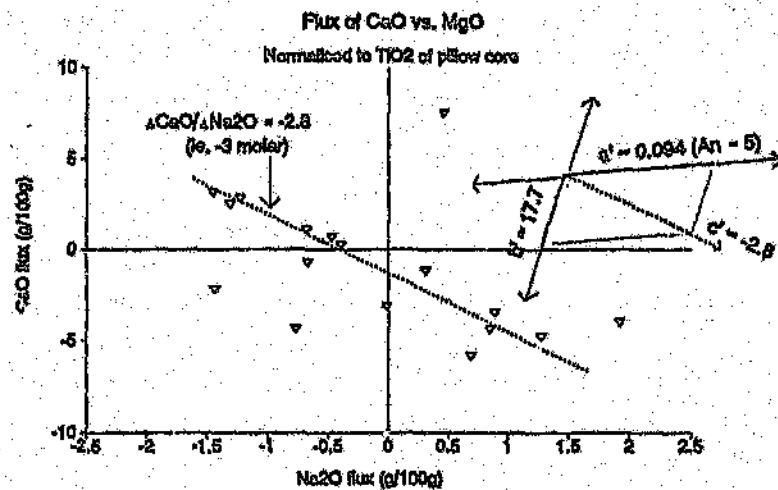
The SiO_2 contents of seafloor altered basalt show that losses from chloritised rims do occur (Humphries and Thompson, 1978a) implying a negative relationship. From experimental studies it is manifest that the degree of silica saturation in solution is usually between high to just above saturation suggesting essentially a continued gradual loss from the rock. Mottl (1983) also records a loss of silica from his experimentally altered samples but states that there is no correlation with Mg-flux.

$\text{K}_2\text{O}/\text{MgO}$. The hydrothermally altered komatiitic basalt veinlet-salvages show a negative relationship for the flux of potassium versus magnesium (Fig. 3.14c). A regression of the data produces a slope of -0.085 suggesting that K_2O is lost when MgO is gained and *vice versa* implying potassium-losses for those samples that have been amphibolitised and gains for those that have undergone albitisation + silicification.

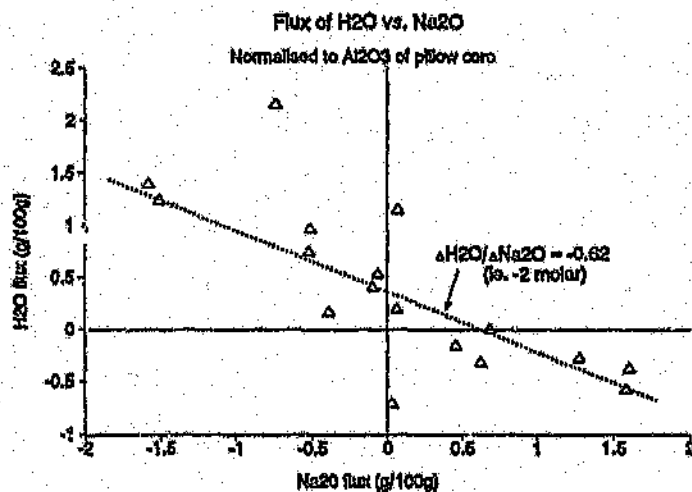
It is well established for MORB alteration, that potassium is leached during high-temperature metamorphism and that its taken up again during low-temperature alteration or weathering of the ocean-floor (Scott and Hajash, 1976; Rosenbauer and Bischoff, 1983; for a summary cf Thompson, 1983). Komatiitic basalt mimics MORB's loss of potassium during initial Mg-addition but appears to reincorporate it again during albitisation which is not the case for tholeiite. K_2O -addition, however, seems unrelated to low-temperature alteration or weathering as low-temperature phase overprints or associated textures are not observed. This is further supported by the lack of CO_2 addition as discussed earlier. However, the distribution of potassium appears to be unsystematic in Fig. 3.10 which suggests that its introduction may be linked to fluids derived from younger felsic intrusives of the area (Viljoen and Viljoen, 1969; de Wit *et al.*, 1986). Such late fluids may have selectively interacted with the more compatible albite and quartz rich assemblages rather than the less compatible Mg- and Ca-rich phases, imparting the impression that it was reincorporated during albitisation. Step-heated komatiite samples have produced $^{40}\text{Ar}/^{39}\text{Ar}$ ages ranging from 3486 ± 6 Ma to 3425 ± 10 Ma which are thought to represent the age of metamorphism of the Komati Formation and subsequent felsic activity (Lopez-Martinez *et al.*, 1992). Such activity supports the presence of K-rich fluids in the Komati Formation.

$\text{CaO}/\text{Na}_2\text{O}$. The flux relationship of CaO versus Na_2O of the veinlet salvages is negative

a)



b)



c)

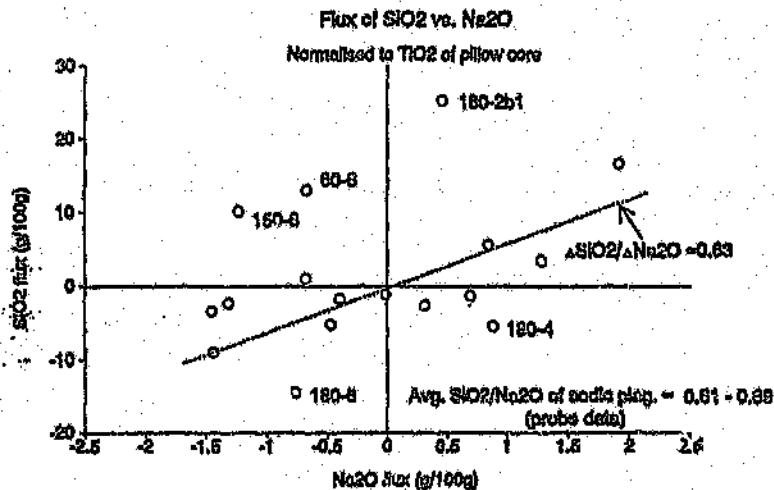


Fig. 3.15

Flux of CaO vs Na₂O (a), H₂O vs Na₂O (b) and SiO₂ vs Na₂O (c) all normalised to constant TiO₂ of pillow cores respectively. See text for discussion.

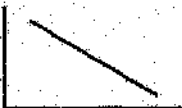
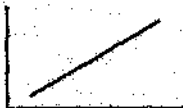
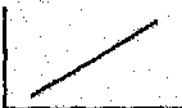
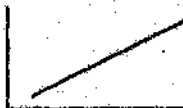
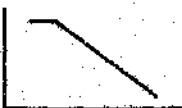
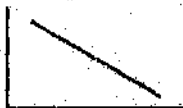
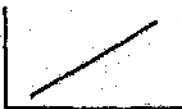
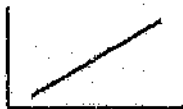
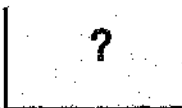
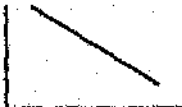
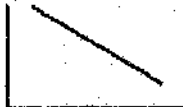
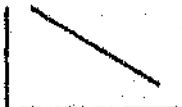
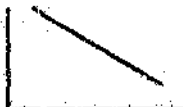
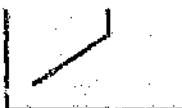
with a ratio of 3:1 (molar basis), suggesting that relatively large quantities of calcium were gained or lost (redistributed) at the expense of small quantities of sodium (a). The insert-diagram offers an explanation for the negative correlation, $a' = \pm 0.094$ represents the rate at which the addition or removal of albite ($An = 5$) would increase or decrease the CaO/Na_2O flux of the samples. Similarly $b' = \pm 17.7$ represents the rate at which the addition or removal of the average amphibole of the pillows would increase or decrease the CaO/Na_2O flux of the samples. The insert shows that by adding albite, for example, amphibole has to be removed (replaced) to satisfy the observed relationship (-/-).

The CaO/Na_2O flux relationships for hydrothermally altered MORB is not normally described in the literature as neither component is considered to be an independent variable (these variables usually being H_2O and MgO , ie. those primarily gained from seawater cf Mottl, 1983). It is well established in the literature, however, that CaO of basalt is exchanged for seawater MgO on a 1:1 molar basis until all the MgO in the fluid is consumed. During this interval, Na_2O is initially taken up, to water/rock ratio's of about 10 and thereafter is lost to the circulating fluids (Mottl, 1983). Such relationships suggest that the flux of CaO/Na_2O for hydrothermally-altered basalt will essentially be sympathetic with a kink in the low flux region parallel to the CaO axis, suggesting that CaO is lost while Na_2O is initially retained whereafter both are depleted (schematically represented in Table 3.8). Such an relationship however, would be to the contrary of what has been established for the study area.

H_2O/Na_2O . The H_2O/Na_2O flux for the studied veinlet selvages of the pillow cores is negative (-0.62). Such a flux ratio implies that 2 moles of H_2O is either taken up or lost at the expense or gain respectively of 1 mole Na_2O (b). This shows that albitisation causes dehydration and that the degree of hydration of these rocks is potentially a misleading indicator of the extent of alteration if one does not know the mineralogy.

SiO_2/Na_2O . The SiO_2/Na_2O flux for the pillow veinlet selvages is positive with a slope of 0.63. This suggests that the flux is controlled by albite formation as microprobe analyses of Na-plagioclase range between 0.61 - 0.69 SiO_2/Na_2O (c). From the graph it is evident that relatively large quantities of silica are added to the rock during albite formation. It is also petrographically evident from the cooling-crack selvages of MC60

Table 3.8 Summary of elemental flux for the study area with corresponding fluxes for naturally and experimentally altered MORB (see text for discussion).

Elemental flux	Seawater-Basalt interaction (from experimental and natural samples)	Hydrothermal alteration of Komatiitic Basalt	Degree of correlation
CaO/MgO			X
H ₂ O/MgO			✓
Na ₂ O/MgO			✓
FeO/MgO			✓
Al ₂ O ₃ /MgO			
SiO ₂ /MgO			✓
K ₂ O/MgO			✓
CaO/Na ₂ O			X

(Fig. 2.6D) and the rim of MC180 (Fig. 2.4C) that preceding fronts of silicification appear to "prepare the ground" for albitic alteration. Seyfried and Janecky (1985) point out that high levels of silica in their experimental solutions stabilised the formation of albite. The high-silica concentrations are thought to result from the high SiO_2 -content of the East Pacific Rise (EPR) basalts. Komatiitic basalts have similar silica contents (~ 1 wt% more than MORB, Table 2.5, Smith, 1980; Donaldson, 1982), suggesting a similar origin for silica in the Archaean fluids. The behaviour of silica and sodium in the komatiitic basalts thus seems to be in good agreement with basalt-seawater experimentation at temperatures between $350^\circ - 425^\circ\text{C}$ and pressures between 375 - 400 bar.

3.5 Sulphur Isotopes

Sulphur isotopes of bulk sulphide separates from pillow cores, cooling-crack veinlets, outer cores, rims and hyaloclastites were analysed, and the results are reported in permil in Table 3.9. The Cañon Diablo troilite (CDT) was used as a standard. The results are plotted in Fig. 3.16 together with other data for the Archaean and those of fresh basalt, present seawater and an ocean-floor hydrothermal-vent fluid. The aim of this section is (1) to investigate the origin of the pillow sulphides; and (2) to evaluate the $\delta^{34}\text{S}$ values of the allochemical interpillow sulphides in terms of a seafloor-metamorphic origin.

The $\delta^{34}\text{S}$ values for the pillow-interior and the pillow-exterior sulphides in the statically metamorphosed area show no significant variation and range between -2.0 to -0.2 (n = 10). The pillow cores of the dynamically metamorphosed pillows also fall within this narrow range but their pillow exteriors are marginally heavier and range between +0.4 to +1.9 permil. The entire $\delta^{34}\text{S}$ range of the pillow sulphides fall well within the confined -3 to +2 permil-range indicative for mantle-derived sulphur (Ohmoto and Lasaga, 1982) thus confirming a mantle origin. The enrichments of allochemical sulphides in the interpillow channelways also bear this imprint, implying that the sulphur isotopes were not fractionated by the fluids that remobilised them. The reasons for the lack of modification and the extent to which the values are compatible with a seafloor metamorphic origin is given below as a part of the interpretation and discussion of results.

Table 3.9 A list of S-isotope values for bulk sulphide extracts of the various textural domains of the Komati Formation pillows.

Sample Number	Sample Origin	Metam. Environment	$\delta^{34}\text{S}\text{‰}$
MC 60-1	pillow core	static (hydrothermal)	-0.2
MC60-2	c. crack veinlet	static (hydrothermal)	-1.5
MC60-4	pillow rim	static (hydrothermal)	-0.9
MC60-6	hyaloclastite	static (hydrothermal)	-1.0
MC150-1	pillow core	static (hydrothermal)	-1.0
MC150-3	outer core	static (hydrothermal)	-2.0
MC150-6	hyaloclastite	static (hydrothermal)	-0.9
MC180-1	pillow core	static (hydrothermal)	-0.7
MC180-4	pillow rim	static (hydrothermal)	-0.8
MC180-6	hyaloclastite	static (hydrothermal)	-0.7
MC120-1	pillow core	dynamic	-1.3
MC120-3	outer core	dynamic	-0.1
MC120-4	pillow rim	dynamic	+0.4
MC20-1	pillow core	dynamic + CO ₂ met.	-0.7
MC20-3	outer core	dynamic + CO ₂ met.	+1.9
MC20-6	pillow rim	dynamic + CO ₂ met.	+1.4

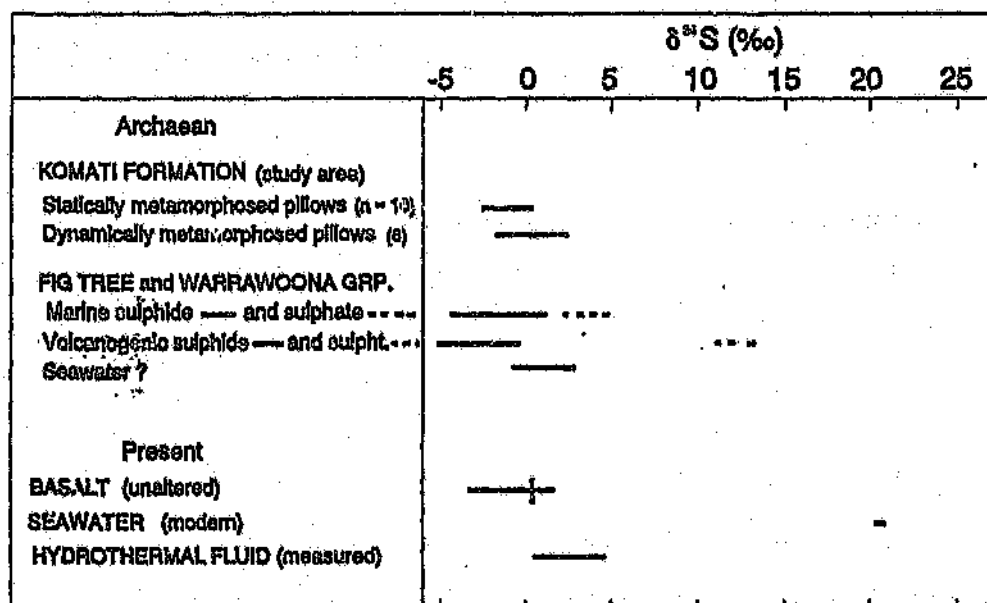


Fig. 3.16 Sulphur isotopes of sulphides from the study area compared with other early Archaean sulphides and sulphates and also present day sulphur values. See text for discussion.

3.6 Interpretation of Results

3.6.1 Alteration petrography

The komatiitic basalts of the study area have conspicuously altered pillow-core tension cracks, pillow rims and metahyaloclastites, suggesting fluid-rock interaction in an open-system. The presence of silicate and sulphide precipitates in the interpillow passageways indicates that extensive solution and mass transfer occurred as a result of fluid-rock interaction. The alteration sequences of individual pillows systematically reflect the changes of the fluid environment that created them. The alteration parageneses thus represent clues to the solution of how the Komati Formation submarine-hydrothermal system developed and operated. A reconstruction of the Archaean hydrothermal environment will be based on the well-established manner in which MOR-fluid compositions evolve and the assemblages they create (Seyfried *et al.*, 1988; Alt *et al.*, 1986).

3.6.1.1 Physical Parameters of Alteration. Reference to the wealth of information accrued from natural and experimental basalt/seawater interaction studies is therefore both desirable and essential. An preliminary evaluation of the physicochemical conditions that prevailed for the Komati Formation will establish the degree to which comparisons can be made with those experiments simulating natural environments. The main physical variables are: (1) pressure, (2) temperature, (3) permeability and (4) water/rock mass ratios.

Pressure. For the study area, pressure estimates for hydrothermal convection can be derived from the previously determined emplacement depths of the Komati Formation (determined in Chapter 2, Fig. 2.12). Pressures of hydrothermal convection can be assumed to be hydrostatic and therefore equal to ocean depth plus the depth of the alteration zone in question below seabed. The minimum pressure, for example, experienced by MC150 and MC180 near the base of the Komati Formation during the emplacement of MC60 equates with the minimum ocean depth of emplacement of MC60 (~1300 m) and the difference in stratigraphic elevation between it and the mentioned pillows at the base of the formation (~1600 m). These depths add up to a total depth of 2900 m which equates with a $P_{H_2O} = 290$ bar. By the same method, maximum pressures experienced by MC150 and MC180 (using the higher initial volatility values),

at the time of emplacement of MC60 equates with a P_{H_2O} of 460 bar. Recent basalt seawater experiments have been using pressures between 300 - 400 bar, simulating conditions at the tops of magma chambers at the East Pacific Rise (Seyfried and Janecky, 1985; Seyfried *et al.*, 1988).

Temperature. The hydrothermal-silicate parageneses of the study area are typical of the greenschist facies and compare favourably with those of MORB (e.g. Cann, 1969; Miyashiro *et al.*, 1971; Humphries and Thompson, 1978a; Alt *et al.*, 1986) and those formed by basalt/seawater experimentation within a temperature range of 300 - 425°C (e.g. Mottl, 1983; Seyfried and Janecky, 1985; Seyfried *et al.*, 1988, 1991).

Permeability and fluid access. The permeability and the ease at which hydrothermal fluids may have circulated in the Komati and Hooggenoeg Formations is difficult to assess. Models of convective circulation within Archaean ocean crust are often compared to those that occur at fast- rather than slow-spreading centres, an analogy based on the assumption that higher spreading rates existed due to a hotter Archaean mantle (Nisbet and Fowler, 1983; Royden and Parsons, 1982). Relative to slow-spreading ridges, fast-spreading centres have higher massive to pillowed flow ratios and lower degrees of (rift-related) brecciation (e.g. Spooner and Fyfe, 1973; Ballard *et al.*, 1981; Evarts and Schiffman, 1983; Alt *et al.*, 1986) which collectively give rise to shallower and less efficient convective cooling. For the Komati Formation, however, lower permeability but deeper convection is predicted. Lower permeability would have resulted from the higher massive to pillowed flow ratios and the absence of rift-related breccias. Deeper convection is postulated on the grounds that contraction cracking of lava flows occurred at the ocean/seafloor interface and that the cracked lavas were rapidly buried to depths of ≤ 7 km. In contrast, at modern spreading centres, both fast and slow, such depths are equivalent more or less to the base of the oceanic crust where cracking fronts develop very slowly (into magma chambers) due to slow heat dissipation which is initially only conductive (Lister, 1983). Not all plutonic sections are hydrated (e.g. Ito and Anderson, 1983) suggesting for some, that cooling is only conductive as opposed to convective. Hydrothermal circulation within the study area lavas may thus have been different to that of fast-spreading centres due to the burial of a "ready made" fracture network to depths of up to 7 km, resulting in low surficial permeability but increased permeability at depth.

Water/rock ratio. The flux calculations (Table 3.8), especially those of MgO, Na₂O CaO and FeO have virtually identical ranges to those demonstrated by Mottl (1983, Fig. 3), suggesting that element exchange capacity of the rocks during alteration was similar to modern MORB. Using the MgO-flux method of Mottl as an indicator of water/rock ratio, sample MC150 may be used as an example because its alteration is relatively simple. The average MgO gain for the veinlet selvages is 1.6 g/100g, whereas the gain made by the interpillow hyaloclastites is 6.58g/100g (Table 3.7). These values translate into water/rock ratios of *ca.* 7 for the veinlet selvages and 30 for the interpillow channelways (assuming that Archaean seawater also contained 0.22 wt% MgO as the present oceans do, *cf* Holland, 1976). A more integrated view is gained by also considering the bulk value of the pillow as well. The outer-core of MC150 has gained only 0.45g of MgO per 100g of rock suggesting a water/rock ratio of 2. The water/rock ratio of the bulk of the pillow may therefore be assumed to be between 0 - 2. Water/rock ratios of whole integrated systems have been estimated both from oxygen isotope data (Craig *et al.*, 1980) and abundances of soluble elements (Li, B and K) and indicate low values in the range of 0.28 - 3 at 21°N EPR (Von Damm *et al.*, 1985; Spivack and Edmond, 1987). Ito and Clayton (1983) demonstrate from oxygen isotope data that metagabbros from deep-seated levels of the Cayman Trough were altered at similarly low water/rock ratios of 0.2 - 2.0. Seyfried and Janecky (1985) carried out their seawater-basalt experiments at water/rock ratios between 0.5 and 1.0 to simulate conditions in the high temperature portions of the EPR. The range of water/rock ratios determined for MC150, a typical study-area pillow, shows good agreement with values obtained from natural samples.

The main physical conditions of hydrothermal metamorphism to affect the rocks in the study area, as outlined, compare favourably with those of natural samples and those used to simulate the deep-seated parts of convection cells. The broad similarities suggest that the information that exists for basalt/seawater experimentation is both relevant and applicable to the altered portions of the pillows in the Komati Formation.

3.6.1.2 Alteration Sequences. A summary of the alteration petrography of the statically metamorphosed pillows is presented in Table 3.10. The table shows the development of the alteration parageneses for the various pillow domains. Relatively low and high

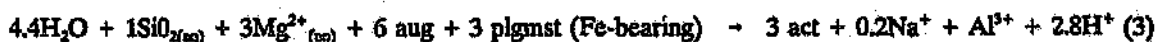
Table 3.10 Mineral parageneses of hydrothermally-metamorphosed pillow domains

Pillow No.	Alteration Stage	Veinlet Wallrock	Vein Infill	Pillow Rim	Interpillow Infill
MC150	1 st (Mg-addition)	Amphibolitization (reaction 2)	(zero → open space)	Alteration record not preserved (albitization of hairline fractures?)	(zero → open space)
	2 nd (Na ± Ca-addition)	unchanged (probably due to clogging)	Alb ± Epd/Clz ± Qtz	Na-plag ± Clz ± Sph (present as mineral relicts)	Alb/Olg ± Epd/Clz ± Act (locally preserved)
	3 rd (Mg-addition)	unchanged	unchanged (probably due to clogging)	Chl + Prg/Hbl + Sph	± Hbl
	4 th (filling of new fracture set)	microcline	—	—	muscovite
MC180	1 st (Mg-addition)	Amphibolitization (reaction 2)	(zero → open space)	Hydrolytic devitrification (Liesegang zoning + albitification?)	(zero → open space)
	2 nd (Si + Na-addition)	Albitization (reaction 3)	(open space)	Silicification	Qtz (columnar grains)
	3 rd (Na ± Ca-addition)	unchanged (probably due to clogging)	Alb ± Epd/Clz ± Qtz	Alb ± Clz	Alb ± Clz
	4 th (filling of new fracture set)	microcline	cericite	—	—
MC80	1 st (Si-addition)	Silicification (filling of amygdalae)	(zero → open space)	Devitrification + Alb + Sph (albitization of hairline fractures)	(zero → open space) (incipient silica precipitation)
	2 nd (Na ± Ca-addition)	Albitization (reaction 4)	Alb + Epd ± Qtz	Alb + Sph ± Epd ± Sph (with continued Albitization Sph is lost)	Qtz + Alb (columnar grains)
	3 rd (filling of new fracture set)	Cc + Chl	Cc + Chl	Cc + Chl	Cc + Chl

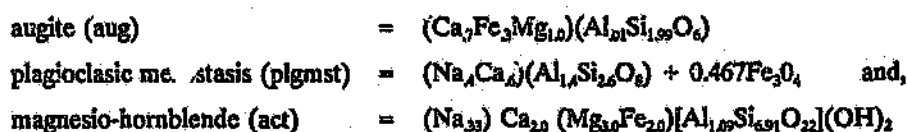
degrees of fluid/rock interaction are thought to have occurred in the cooling crack and interpillow channelway settings, respectively. Wallrock replacement and fluid-passage infill are indicated alongside each other in the table if thought to be coeval whereas the consecutive stages of alteration represent either abrupt or gradual changes in fluid-composition depending on observed parageneses textures. The principle cation species added to the rock by the fluids, inferred from the mineral compositions, are indicated in parentheses in the column "Alteration Stage".

It is petrographically evident that rims of pillows are far more altered than the veinlet margins and that veinlet-infill parageneses are less complex than those of the interpillow areas. These observations relate directly to the intensity and duration of alteration. It is contended here that early alteration records of pillows are preserved more accurately by the cooling-crack selvages and vein infill parageneses. Due to early clogging they were isolated from further hydrothermal activity and overprinting. Early records of the pillow exterior domains, on the other hand, are probably continuously overprinted or replaced allowing only the latest assemblages to be preserved, before the interpillow channelways were sealed. Alteration reactions are described below and have been quantified by reaction formulae which provide an insight into the losses and gains of chemical components and give an indication of how fluid compositions might have evolved. The reactions integrate the petrography with the geochemistry and also simulate the elemental fluxes discussed earlier.

Alteration stage 1, pillow MC150. The cooling-crack selvages of pillow MC150 have been amphibolitised (Table 3.10, Figs. 3.4A, 3.5D). The relatively intact microspirifex habit (Fig. 3.6C) of the amphibolitised selvages suggest that the protolith wallrock was probably pristine at the onset of alteration. The reaction responsible for amphibolitisation may be represented by reaction 3, as follows:



where:



The reaction suggests that amphibolitisation is likely associated with Mg-addition to the rock at the expense of Na^+ , Al^{3+} and H^+ . The losses and gains of elements on either side of the reaction are also confirmed by the geochemical fluxes of the elements in Figs. 3.12b; 3.13a,b,c; 3.14a,b. The concomitant uptake of CaO relative to MgO as is indicated in Fig. 3.12b may be derived from the anorthite component of the plagioclasic mesostasis which is eventually amphibolitised. Mg-addition to the rock and H^+ -production, however, are very typical of alteration in the seawater-dominated recharge zones of modern MOR-hydrothermal systems (e.g. Mottl, 1983; Seyfried *et al.*, 1988).

During Mg-addition to the cooling crack selvages, the cooling cracks remained open. The albite-dominated infill of the cracks is inferred to have followed amphibolitisation because the elemental flux diagrams consistently show that the two types of alteration are mutually exclusive of each other. As Mg-fixation was active at the cooling-crack selvages, similar fluids must have been associated with the pillow rims and those parts (hyaloclastite) lining the hydrothermal vent.

Although the pillow rim (of MC150) is chlorite- and amphibole-rich, it is doubtful on textural evidence (discussed below for Stage 2), whether this assemblage developed during initial amphibolitisation of the cooling-crack selvages. The only early textures preserved are the albitised hairline tension cracks at the inner rim of the pillow (Fig. 3.4C). The albitisation of the hairline cracks cannot be linked to the albitisation of alteration stage 2 of this pillow but can be tenuously linked with the Mg-addition of cooling crack selvages discussed above. They have been previously described as texturally early and from this description it is evident that the fractures are hairline in nature and occur along the inner rim. As shown by basalt-seawater experimentation, it is possible that Mg-rich fluids gained access into the low-permeability fractures and soon lost their magnesium to the glass, probably giving rise to amphibole and/or chlorite and the formation of H^+ . As the fluids penetrated still further into the fractures towards the inner rim, compositions would progressively become rock-dominated and depleted in Mg^{2+} . As a result the Na/Mg ratio would decrease and favour the onset of albitisation (cf Seyfried *et al.*, 1988). At the same time, due to the more dilated (≤ 7 mm) nature of the thermal contraction cracks in the pillow core, fluids are more readily replenished, maintaining the Mg/Na ratios which are favourable for amphibolitisation.

Alteration stage 2, pillow MC150. The termination of Mg-rich fluid circulation and albite infilling appears to have been rapid. This is shown by a lack of albitisation of the veinlet wallrocks. They comprise only actinolitic- and magnesiohornblende and have no albite overprint adjacent to the albite-dominated infill (Fig. 3.6C). This suggests that after the protolith had experienced a sustained period of intermediate-intensity amphibolitisation (microspinel texture only partially degraded), it was interrupted by a short-lived period of albitisation, infilling only the veinlets without affecting the wallrock. Infilling of the veinlets effectively preserved the parageneses from further direct hydrothermal metamorphism associated with the interpillow areas. The onset of albitisation is inferred to mark a dramatic change in fluid composition, brought about by a sharp change in the direction of convective flow in the network of interconnecting channelways and contraction cracks.

Introduction of Na-rich fluids to the pillow exteriors is preserved only as relict patches of sodic-plagioclase and clinozoisite $\pm$ sulphide. This assemblage was subsequently extensively replaced by chlorite and amphibole of Stage 3. Na-metasomatism of Stage 2 may be the reason why the water/rock ratio of ~ 3 (shown by Mg-uptake) for the rim is low compared to a water/rock ratio of ~ 7 for the veinlet selvages and a ratio of ~ 30 for the hyaloclastite (discussed earlier). A value similar to that of the hyaloclastite would normally be expected. The low water/rock ratio of the rim, however, can be explained by Stage 2 Na-metasomatism which removed much of the Mg-added during Stage 1, leaving the rim deficient in Mg but Na-rich. Subsequent Mg-addition of Stage 3 may then have largely replaced the Na-plagioclase and clinozoisite of Stage 2, effectively causing only a small gain in Mg. The veinlet selvages and hyaloclastite seem to have been protected from Stage 2 Na-metasomatism, maintaining earlier (Stage 1) Mg-gains.

Alteration stage 3, pillow MC150. The third alteration stage is characterised by the infiltration of fluids that once more add Mg to the pillow rims and hyaloclastite. The sealed-off tension cracks preserve the pillow interior from interacting with these fluids and they are themselves only partially replaced by chlorite and amphibole where they approach the pillow rims (Fig. 3.4A). Because Mg-addition is characteristic of downwelling hydrothermal flow at MOR-systems, it is inferred to be the case for fluids interacting with MC150.

Alteration stage 4, pillow MC150. This stage of alteration is marked mainly by fracture filling with microcline. The fractures crosscut all the alteration features described above, they are sparsely spaced and usually very thin (≤ 1 mm, Fig. 3.4A). Because microcline is not a typical seafloor-metamorphic mineral, the veins are thus interpreted to be unrelated and of post-seafloor age. The veins may be associated with the intrusion of several small granitoid plugs of the area (Viljoen and Viljoen 1969; Kamo and Davis, 1994).

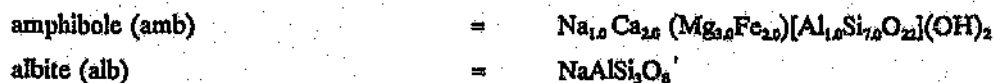
Alteration stage 1, pillow MC180. The cooling-crack selvages of pillow MC180 show composite alteration where albitisation laterally overlaps amphibolitised wallrock (Figs. 2.4C; 3.6B; Table 3.10). Textural relationships suggest, as in the case of MC150, that the protolith wallrock was pristine when it was initially amphibolitised (reaction 3) and then gradually albitised. The fact that both alteration types are laterally exposed to the veinlet with a transgressive relationship suggests that it may be the physical expression of a hydrothermal fluid-composition crossover point.

During initial amphibolitisation prior to albitisation, the tension fractures were obviously still open suggesting that the seawater-dominated fluids must have interacted with the outer portions of the pillow as well. The Liesegang-like zonations that formed at the pillow rim (Fig. 3.5C), are tentatively inferred to be of the same age as the amphibolitised selvages, simply on the basis of their early textural development. Liesegang zonations in devitrified glass have been noted in "freshly" altered palagonitised pillow rims and are described as signs of hydrolytic devitrification (Zierenberg *et al.*, 1988).

Alteration stage 2, pillow MC180. This stage is marked by the progressive change from seawater- to rock-dominated fluids causing the replacement of amphibole by albitic plagioclase along the tension-fractured wallrocks. The relationship is essentially a metasomatic reaction where amphibole rather than the pristine protolith, is albitised and this can be represented by reaction 4.



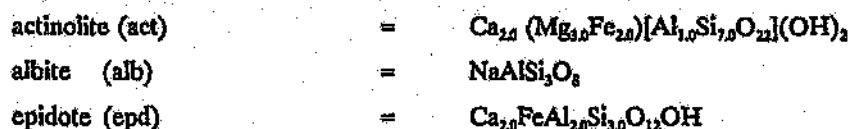
where:



Occasionally small amounts of epidote accompany albite and the reaction may be represented by reaction 5.



where:



The losses and gains of elements on either side of reactions 4 and 5 are supported by the geochemical flux relationships of the corresponding elements in Figs. 3.12b; 3.13b,c; 3.14a,b; 3.15a,b,c. Reactions 4 and 5 require the chemical species released by reaction 3, suggesting that they may operate in tandem as two end-member reactions. Interestingly, reaction 4 consumes H^+ and generates H_2O , supporting the dehydration that is often observed for albitisation (Fig. 3.15b). On the other hand, if only small amounts of additional silica and alumina are available in solution, additional epidote can be formed (reaction 5), a product that is often seen in the veinlet infills (Fig. 3.6C). H^+ is also produced which is essential to perpetuate hydrolysis.

The compositely altered cooling-crack selvages of MC180 (Fig. 3.6B) may, therefore, be described in terms of reaction 3 gradually being replaced by reaction 4 as a result of the of H^+ , Na^+ and Al^{3+} build-up in the fluid. The build-up of these elements suppressed the Mg/Na ratio sufficiently to bring on the albitisation of amphibole.

In contrast to the rapid change of fluids (indicated above) for the veinlets of MC150, the parageneses here suggest a compositional evolution of the circulating fluids which was

sustained sufficiently to albitise the wallrocks before fracture infilling started (Alteration Stage 3).

The pillow rims (Alteration Stage 2, MC180) show pervasive silicification which can be only tenuously linked to albitisation of the veinlet wallrocks. This is because there is no textural-domain link with the veinlet selvage parageneses. Seawater/basalt experiment records, however, show that a genetic link between albitisation and silicification does exist. It is shown that a high α_{SiO_2} is required to stabilise albite (Seyfried and Janecky, 1985). It is also shown that more silica is taken up into the solutions that have reacted with the SiO_2 -rich EPR basalt than any of the other basalts. The solutions also have the lowest pHs and as a result inhibit the precipitation of quartz even though the solutions become supersaturated (Seyfried and Janecky, 1985). These workers also report that at 400°C and 400 bars the high silica levels of the EPR basalt solutions appear to cause a change in the general reaction mechanism reflected by Na- and Ca-fixation. At 425°C and 400 bar they maintain that Na-fixation exceeds Ca-fixation, Na uptake being great and balanced in part by Cl^- uptake (Seyfried and Janecky, 1983; Seyfried *et al.*, 1985). The greatly reduced solubility of Na and Ca-rich alteration phases limits dissolved SiO_2 concentrations to values at or below quartz saturation, even though the fluids of the experiments were extremely acid.

The relatively high SiO_2 -content of the komatiitic basalts may have generated fluids with similarly high silica levels. Such fluids appear to have reacted with pillow MC180, leading to the pervasive silicification of the already partially devitrified rim (Fig. 3.5B). The high fluid acidity may have prevented direct quartz precipitation but may also have induced silicification of the hydrolytically altered glass. The Na-sink that emerges at higher temperature experimentation (i.e. $\geq 400^\circ\text{C}$, reported on by Seyfried and Janecky, 1985) may, in the case of MC180, be represented by the albitisation of the amphibolitised cooling-crack selvages. It is interesting to note that reaction 4, which is thought to represent the parageneses in Fig. 3.6B (with no epidote), consumes H^+ causing solution pH to go up and induce silica precipitation, which may be evidenced by the columnar quartz that has nucleated at pillow surfaces similar to those in Fig. 3.3A. It might be argued, on the other hand, that the quartz could have been precipitated at lower temperatures in a cooling environment, e.g. $<300^\circ\text{C}$ as is often the case (Rimstidt and

Barnes 1980). The formation of lower temperature quartz, however, is not in accordance with subsequent development of clinozoisite which is not known to form at such low temperatures. The formation of clinozoisite in the 350 - 425°C experiments of Seyfried and Janecky (1985) is very significant as epidote minerals generally do not form in experiments conducted at lower temperatures.

Alteration stage 3, pillow MC180. The third alteration stage is inferred to be represented by the infilling of the tension veinlets with sodic plagioclase $\pm$ epidote and the albitisation of the silicified pillow rims and replacement of columnar quartz. This stage of fluid interaction is not marked by a distinct break in composition but is perceived to represent the progressive evolution of the fluids due to continued wallrock interaction and/or increasing temperature. As indicated by Seyfried and Janecky (1985), at temperatures of 425°C sodium fixation exceeds that of calcium. The association in MC180 of small amounts of clinozoisite with albite seems to verify the process and thus the temperature ($\sim$ 425°C) for MC180.

SiO₂-precipitation and -addition to basalt is conspicuous in the upflow zones of active submarine hydrothermal systems (Delaney *et al.*, 1987; Zierenberg *et al.*, 1988), where it is also characteristic for the fluids to be Na-rich, Mg-free and Ca-poor. Such fluids, based on experimentation and sampling of MOR-vent fluids, are thought to form during high-temperature reaction in the deeper parts of convection systems before they rise and are cooled adiabatically and conductively (von Damm *et al.*, 1985). It is also thought that the interaction of such fluids with fresh or previously altered basalt is prone to result in extensive albitisation (Seyfried *et al.*, 1988). Due to the formation of such parageneses in the study area, physical conditions were obviously similar to experimental conditions of a hydrothermal upflow zone, i.e. 350 - 425°C, 350 - 400 bar, and with water/rock mass ratios of around 1.

Alteration stage 4, pillow MC180. The same as Stage 4 of pillow MC 150 (described above).

Alteration stage 1, pillow MC60. The alteration selvages of MC60 are compositely altered showing an outer zone of incipient silicification and a more pronounced inner zone of

albitisation (Fig. 2.6D). Silicification appears thus to have preceded albitisation but the transition between the two stages is not clear. Due to its incipient petrographic nature (discussed earlier) it is presumed that the main stage of alteration and hydration of the core selvages might have occurred only during subsequent albitisation (of Stage 2). (This is partly supported by the fact that the pillow core still has many kernels of augite preserved).

The fluids responsible for incipient silicification of the tension crack selvages appear to have hardly affected the outer parts of the pillow, except possibly for the albitisation of the hairline fractures found at the pillow rim (Fig. 3.4D). Albitisation of these fractures can be ascribed to small volumes of fluid that penetrate the cracks and soon became rock-dominated with resultant $\text{Na}^+_{(\text{aq})}$ -addition to the selvages. It is also evident from Figure 2.5A that the albitised hairline cracks are situated close to the pillow surface suggesting that the fluids may already have been evolved (rock-dominated) prior to interaction with the rock. This contrasts with the Mg-rich fluids of MC 150 which became rock-dominated subsequent to entering the hairline fractures and affected albitisation at the inner rim, i.e. further along the fractures. In a previous discussion it was noted that fluids with a high a_{SiO_2} are usually also Na-rich, Mg-depleted and Ca-poor, and it is suggested that the transitional period between this and the ensuing alteration stages might have been small or even part of the same process.

Alteration stage 2, pillow MC60. The onset of fluid conditions favourable for the albitisation of an ocelli-bearing protolith progressively replace the fluids responsible for incipient silicification mentioned above. The fact that relict kernels of augite are still present in the selvages suggests that the protolith may still have been largely unaltered subsequent to incipient silicification and prior to albitisation which, if correct, may be represented by reaction 6, as follows:



where:

augite (aug)	=	$(\text{Ca}_2\text{Fe}_3\text{Mg}_3)(\text{Si}_2\text{O}_6)$
plagioclasic mesostasis (plgmst)	=	$(\text{Na}_4\text{Ca}_4)(\text{Al}_{1.4}\text{Si}_{2.6}\text{O}_8) + 1.4(\text{FeO})$
actinolitic hornblende (act)	=	$(\text{Na}_4)(\text{Ca}_{2.0})(\text{Mg}_{3.0}\text{Fe}_{2.0})[\text{Al}_4\text{Si}_{7.6}\text{O}_{22}](\text{OH})_2$
albite (alb)	=	$\text{NaAlSi}_3\text{O}_8$

Apart from silica, that may be obtained either from the incipiently altered phases of Stage 1, mesostasis, or from solution, the reaction requires few cations, relative to the amount of water, to alter a considerable amount of komatiitic basalt to amphibole, albite and to add H^+ to the solution. The reaction is, thus very much a hydration reaction due to the large amount of water added to the rock. It is interesting to note that the stage of incipient silicification may have served to stabilise albite formation as suggested by Seyfried and Janecky (1985). It is also notable that both Mg^{2+} and Na^+ are added simultaneously to the sample, contrary to the flux relationships discussed earlier (Fig. 3.13b). It has been established experimentally for MORB, however, that during incipient alteration (at water/rock ratios ≤ 10) that both magnesium and sodium are concomitantly consumed (Mottl, 1983).

The change in fluid composition from Stage 1 (Si-addition) to Stage 2 (Na-addition), is inferred to be gradual rather than abrupt (Table 3.10 and Fig. 2.6D). This is presumed from (1) the sympathetic geochemical association there generally appears to be for silica and sodium fluxes (Fig. 3.15c); (2) the similar relationship observed for alteration of MC180; (3) the close association shown by Si and Na flux by basalt/seawater experimentation (Seyfried and Janecky, 1985), and (4) the close to pillow-surface positioning of the albitised hairline cracks mentioned earlier, suggesting the fluids were already Na-rich early on in the alteration history of the pillows.

Only subsequent to imparting a distinct zone of albitised wallrock do the cooling-crack veinlets clog. During this time the pillow margin is pervasively albitised with a scattering of epidote, resulting in a paragenesis broadly similar to that of MC180. Differences being less clinozoisite and more epidote for MC60, and silicification affecting the veinlet selvages versus pillow rim of MC60 and MC180 respectively (prior to albitisation). Silicification preceding the main stage of albitisation is, nonetheless, quite similar to the paragenesis of MC180, reflecting a coherent compositional trend for fluid development in both instances, namely from SiO_2 -rich to Na_2O -rich. Such fluids are characteristic of the upwelling parts of hydrothermal systems.

Alteration stage 3, pillow MC60. This stage is marked by fracture infilling with chlorite and calcite. The fractures have a crosscutting relationship with the features described

above and it is not possible to indicate whether they are related to seafloor metamorphism or not. They are, however, not widespread or common and may therefore be related to local phenomena such as the intrusion of a Proterozoic dyke (approx. 100m) to the west of the locality.

3.6.1.3 Significance of alteration sequences: fluid circulation. The recognition of alteration parageneses in the study area similar or equivalent to those of naturally and experimentally altered MORB, suggests that broadly similar fluid-types convected in the Komati Formation during seafloor metamorphism. The fluids added Mg, SiO₂ and Na to the rocks. The patterns in which the circulating fluids superpose their alteration onto newly formed crust at MOR-spreading centres has been established by Alt *et al.* (1986). The patterns seem to differ markedly from those established for the study area despite individual alteration processes being similar. A better understanding of the processes that gave rise to the assemblage sequences (or patterns) at spreading axes is germane to understanding hydrothermal circulation of the Komati Formation.

Two types of hydrothermal convection develop in newly formed ocean crust due to different heat flow distributions. Adjacent to the ridge axis one or two convective cells are fixed (into position) by the heat supplied by the intrusion process at the spreading centre while cells further away from the spreading centre migrate laterally with the spreading plate, driven by conductive heat from below (Fehn *et al.*, 1983). As a result, segments of crust formed at the spreading centre will pass through distinct parts of the flanking cells and undergo various alteration processes as they move away from the axis through seafloor spreading. A petrographic study demonstrating the sequential alteration of oceanic crust has been reported from an evaluation of hole 504B of the Deep Sea Drilling Project (DSDP, Alt *et al.*, 1986). It was drilled into 1075 m of hydrothermally altered, 5.9-m.y. old, oceanic crust south of the Costa Rica rift. Rocks at the bottom of the drill-hole, representing the pillow lava/sheeted dyke transition zone, experienced four superimposed stages of hydrothermal alteration and veining related to the movement of crust away from the spreading axis: (a) on-axis prograde alteration in the upwelling zone during initial heating of the system, which generated chlorite, actinolite, albite and pyrite, (b) on-axis (re)fracturing of the dikes with the resultant formation of quartz, epidote, and sulphide representing maximum (hydrothermal) PT-conditions, (3) off-axis formation of

anhydrite in fractures due to deep influx of fresh seawater into the hydrothermal system, and (4) more distant off-axis alteration developing zeolites, calcite and prehnite. At the same time, the upper pillow section is also exposed to different environments ranging from the (sub-seabed) fluid mixing zone where the hot ascending evolved waters intermix with cold seawater and finally right at the top where to ocean-floor weathering of fresh basalt occurs (Alt *et al.*, 1986).

The pillows of the study area have also recorded sequential alteration (summarised in Table 3.10) but in a manner which is distinctly different to that recorded for MORB. The differences are:

- 1) Hydrothermal metamorphism of the Komati Formation is observed only in early volcanic-related and not in tectonic-related fracture porosity. The original volcanic porosity includes thermal tension fracture systems in massive and pillowed flows linked to more porous (but limited in size) interpillow-channelway networks and flow contacts. MORBs are undoubtedly more commonly associated with repeated (multiple) fracturing of the host rock due to rift-related deformation (e.g. Spooner and Fyfe, 1973; Liou and Ernst, 1979; Ballard *et al.*, 1981; Ito and Anderson, 1983; Alt *et al.*, 1988). The difference in fracture porosity between the Komati Formation and MORB are ascribed to different tectonic settings, a feature which has been elaborated on in the previous chapter, i.e. oceanic plateau versus mid-ocean ridge.
- 2) The parageneses of the study area rocks is mainly due to (high-temperature) greenschist-grade hydrothermal metamorphism and has no low-temperature overprinting. This is in contrast to MOR-metamorphism which is characterised by on-axis and off-axis high-temperature and low-temperature alteration respectively, the latter being superimposed onto the former.
- 3) The alteration sequences or patterns established for the study area are irregular (haphazard) compared to the more systematic (predictable) style found in MORB. At a ridge axis newly formed segments of crust may be expected to be emplaced in an upwelling portion of a hydrothermal cell and thus be exposed to compositionally evolved fluids. Due to lateral movement away from the spreading axis the segments will pass through the downwelling part of the cell where heated, unevolved seawater would fix Mg^{2+} and possibly form anhydrite. Thereafter, the

segment will be successively exposed to cooler convective fluid circulation as it moves away along with the cooling oceanic plate where low-temperature assemblages are formed comprising minerals zeolite, calcite and smectite. Such assemblages (or textural evidence for their earlier existence) have not been found in the Komati Formation, corroborating findings that the komatiites were not emplaced in a MOR-type setting.

The distinctly irregular alteration patterns of the study area (relative to MORB), implies that the assemblages have been overprinted by hydrothermal processes that have not evolved as systematically as those of MOR-systems. The observed parageneses suggest: (1) for MC150, Mg-addition followed by Na-addition followed by Mg-addition, which interpreted in terms of MOR-fluid evolution, would imply initial exposure to downwelling fluids, passing abruptly into upwelling-, and back into downwelling fluids again at high temperature; (b) for MC180, the parageneses indicate initial downwelling proceeding gradually into upwelling followed by continued upwelling but possibly slightly more deep-seated (as reflected by late, high-temperature clinozoisite-bearing assemblage); (c) for MC60, the parageneses suggest continuous exposure to upwelling fluids for its entire alteration history, the only shift being toward a deeper-seated position (as is evidenced by the presence of pyrrhotite and pentlandite, to be discussed below). Such an upwelling-dominated assemblage (without low-temperature overprinting), implies that spreading could not have occurred. In light of the hydrothermal convection model proposed by Fehn *et al.*, (1983) it would appear that the location of the hydrothermal cells in the Komati Formation were either not fixed relative to a specific heat source or that hydrothermal circulation was regulated by other means, distinct from those of MOR-systems. Such a flow environment, however, can be modeled in terms of hydrothermal convection within an oceanic plateau where fluid inlet and outlet is continuously disrupted (capped) by the emplacement of successive lava flows. This will be elaborated on below, subsequent to more relevant discussions.

3.6.2 Sulphide Petrography

Unlike the silicate phases of the komatiitic basalts, sulphides reveal very little about their pre- to subsolidus and subsequent alteration history. Sulphide alteration, however, is closely linked to that of the host silicates (*cf* Seyfried and Janecky, 1985; Gitlin, 1985)

which form a framework from which constraints for hydrothermal conditions may be derived. The evolution of fluid composition during basalt alteration is well established by means of experimentation. The downwelling limbs of subseafloor convection cells are characterised by the formation of Mg-rich silicate minerals which simultaneously generate sharp increases in the acidity of the reacting hydrothermal solutions (e.g. Bischoff and Dickson, 1975; Mottl and Holland, 1978; Seyfried and Bischoff, 1979, 1981). It is also evident that if the fluids continue to permeate the ocean crust, aqueous Mg will become exhausted and if temperatures rise to above 350°C and pressures remain sufficiently low (≤ 400 bar), then Ca- and/or Na-fixation dominate the role of H^+ -production. Continued H^+ -production is essential for sulphide and base metal dissolution (Rosenbauer and Bischoff, 1983; Seyfried and Janecky, 1985; Seyfried *et al.*, 1988). These workers also contend that MOR-vent fluid chemistry is best reproduced by basalt-seawater experimentation at PT-conditions between 350° - 425°C and ≤ 400 bars respectively.

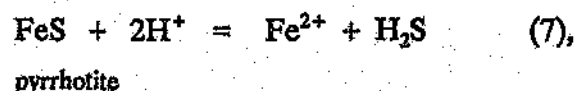
3.6.2.1 *Pillow core sulphides.* Due to the nonexistence of fresh komatiitic basalt, original sulphur content and sulphide species are not known. A general impression of these characteristics, however, may be gained from submarine basalts which have been well studied (e.g. Moore and Fabbi, 1971; Mathez, 1976; Czamanske and Moore, 1977). Czamanske and Moore (1977), for example, show that three sulphide phases coexist in cooling basalt at 610°C, which form initially directly from the melt as immiscible globules. They are monosulphide solid solution (MSS), intermediate solid solution (ISS) and pentlandite. Between 600 and 300°C the intermediate solid solution breaks down to (among others) chalcopyrite and the MSS evolves a second generation of pentlandite (*cf* Craig and Scott, 1982). Pyrrhotite, however, which is a major sulphide constituent of fresh submarine basalt may form from both MSS and ISS in this T range along with other sulphides in globular shapes up to 100 μ m in diameter (Gitlin, 1985). Shanks and Seyfried (1987) suggest pyrrhotite is probably the most reactive phase during basalt/seawater interaction and that the sulphide system is fundamentally related to pyrrhotite-pyrite conversion during hydrothermal alteration.

Modal analyses (Fig. 3.7) clearly show a strongly biased sulphide distribution toward the amygdalae in the pillow cores of the statically metamorphosed pillows. The sulphide enrichment of the amygdalae has been ascribed to early sulphur exsolution from the

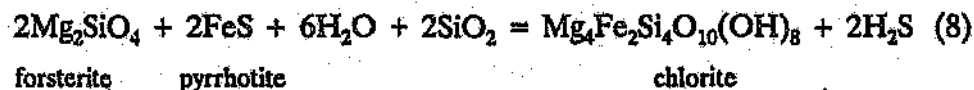
melt/hot rock and later redistribution related to hydration and infilling of vesicles (cf Moore and Schilling, 1973; MacRae *et al.*, 1977).

The scarcity of sulphides in the recrystallised parts of pillow core in the vicinity of the altered cooling-crack wallrocks as opposed to its somewhat more widespread occurrence in the texturally well preserved parts (Fig. 3.7) suggest that sulphide removal was associated with the even $\sim$ seafloor alteration. The uneven distribution of sulphides in the pillow cores as a whole also shows that removal was not pervasive but patchy. The assemblage that has survived partial sulphide removal is chalcopyrite, pyrite and pyrrhotite. Pyrrhotite, however, as shown earlier, is very sensitive to alteration and several mechanisms for its breakdown have been proposed:

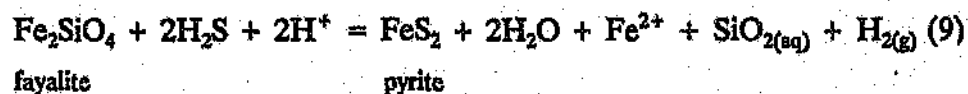
- a) The experimental work of Seyfried and Janecky (1985) indicate that pyrrhotite dissolution may be attributed to either high acidity or to iron silicate or -oxide formation at the cost of iron sulphide according to the following reactions:



or hypothetically, for the silicate relationship:

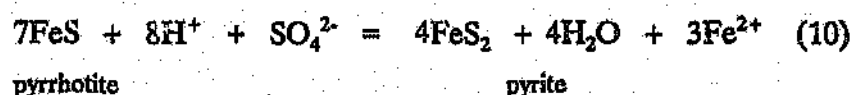


The concentrations of H_2S in solution in the 400 and 425°C experiments (utilising the EPR-samples) dropped rapidly as a result of pyrite precipitation with a concomitant increase in Fe due to probable silicate dissolution according to the following reaction:



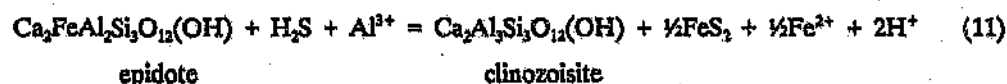
At temperatures below 400°C these reactions did not take place.

- b) The pyrrhotite-pyrite conversion during hydrothermal alteration, described above, is thought to need seawater-derived sulphate¹ as the oxidizing agent. The required acidity to drive the sulphide oxidation-sulphate reduction couple comes from hydrothermal alteration reactions that form sodic plagioclase and epidote (Shanks and Seyfried, 1987). They suggest that the $\delta^{34}\text{S}$ of H_2S of hydrothermal fluids of deep reaction zones appear to satisfy the conversion which may be expressed as follows:



Judging from the alteration parageneses found in the study area and keeping in mind the hypothetical nature of the proposed reactions which have been deduced more from experimental solution compositions than solid reactant compositions, it appears that aspects of both the above-mentioned reactions can be recognised in the komatiitic pillows:

- a) The common occurrence of clinozoisite containing blebs of chalcopyrite and/or pyrite as opposed to epidote without sulphide, may support the first proposal suggesting pyrite formation in response to H_2S reaction with a silicate phase, as follows (similar to reaction 9):



It is evident that the reaction requires aluminium which appears to have been in abundant supply.

- b) The very common occurrence of sodic plagioclase with minor epidote suggest that sufficient acidity was generated to drive the sulphide oxidation-sulphate reduction couple (reaction 10). Sulphur isotope data of the pillow core sulphides do not specifically support or reject such a mechanism of formation (discussed below).

¹ Archaean oceans are thought to have contained sulphate, e.g. Lambert *et al.* (1978).

- c) As no Fe-oxide minerals were found in the Komati Formation seafloor metamorphosed pillows, pyrrhotite decomposition related to oxide formation is not likely to have happened.

The presence of pyrrhotite among the pillow core sulphides demonstrates that the hydrothermal conversion of pyrrhotite to pyrite was incomplete suggesting very low water/rock ratio for the pillow interiors and relatively low f_{O_2} (supported by the widespread absence of magnetite). The suggested placid alteration of the pillow interiors is also consistent with the preservation of the (previously mentioned) amygdale-biased S-distribution which is a subtle feature that could easily have been overprinted had they been subject to more intense alteration. The chemistry of the pillow-core sulphides labelled as "secondary authigenic" comply with the general good preservation of the cores, which is also supported by the textural interpretations of the silicate phases.

3.6.2.2 Interpillow sulphides. The sulphides situated in the pillow exterior samples are always hosted by textural domains and mineral phases suggesting intense alteration and metasomatism. Accordingly, they are marked with the genetic label "allochemical". The modal analyses (Fig. 3.7) clearly show that the metahyaloclastites are the most enriched pillow domains containing on average close to 2500 ppm S. The strong enrichments may suggest that sulphides leached from the pillow cores were precipitated in the interpillow zones (channelways) or that both pillow-core sulphide and seawater sulphide (reduced from sulphate) co-precipitated. The general lack, however, of oxidised minerals such as magnetite or reversely zone epidote in the channelway samples argue against any oxidation (which may have been required to reduce sulphate).

The parageneses of the interpillow zones is pyrite + pyrrhotite + chalcopyrite + pentlandite $\pm$ cobaltite $\pm$ sphalerite (very rare). Deep-seated MOR-stockworks and veins, on the other hand, may comprise pyrite + chalcopyrite + sphalerite $\pm$ galena $\pm$ millerite (Alt *et al.*, 1986; Kawahata and Shikazono, 1988). Implications of the presence of pyrrhotite and pentlandite and the virtual absence of sphalerite in the study area point to an important difference in the evolution of the convective systems of the Komati Formation (oceanic plateau) and MOR-settings respectively.

As pointed out earlier, Shanks and Seyfried (1987) suggested that due to the reactivity of pyrrhotite it would be the first sulphide to disappear during normal MORB alteration. Fluids that retain pyrrhotite in basalt may thus be modeled to be strongly rock-dominated and as a result very reduced. Such fluids for example, are thought to have existed in incipiently-altered MORB in a cooling environment which allowed for pyrrhotite to re-equilibrate at progressively lower temperatures (Gitlin, 1985). The study area interpillow pyrrhotites, however, are allochemical and originate from fluids that have caused severe alteration. This implies that the hydrothermal system which altered the Komati Formation was very rock-dominated and reduced to precipitate these sulphides. It is proposed that the rock-dominated composition of the fluids may have arisen from multiple circulatory passes within individual convection cells, due to clogging for example, so that conditions finally favoured (re)precipitation. Seyfried and Janecky (1985) have established a "maximum solubility band" for base metals (Cu, Zn and Fe) between 350 - 425°C. Precipitation at 425°C occurred despite very high acidity, leading them to conclude that increasing temperature was the dominant control of the upper limit of the solubility band. The high sulphide content of the interpillow areas suggest that the temperature range of the solubility band was exceeded as the lower limit of 350°C is required to achieve base metal dissolution in the first place. It is interesting to note that the pyrrhotite geothermometer (employed in sub-section 3.3.2.2 of this chapter) yielded a temperature of ~430°C which is just above the upper limit of the solubility band. This would seem to support the idea that precipitation was caused by increasing rather than decreasing temperatures.

Pentlandite is commonly found in pristine, unaltered, glass-rich portions of MORB as a primary sulphide (e.g. Czamanske and Moore, 1977). It is not, however, usually reported for basalt that has undergone only slight degrees of alteration (cf Carn, 1969; Humphries and Thompson, 1978b; Gitlin, 1985; Alt *et al.*, 1986). It is also not a common hydrothermal mineral (Ramdohr, 1969). By analogy, if pyrrhotite is sensitive to alteration, then pentlandite is most probably even more sensitive. The presence of pentlandite in the interpillow zones of the study area pillows reflect the degree to which the hydrothermal fluids became rock-dominated, probably also as a result of multiple circulatory passes.

The general scarcity of sphalerite in the interpillow zones may reflect on the temperature of deposition. Sphalerite in MORB is known to form preferentially from cooler fluids than chalcopyrite, both at hydrothermal vent sites (Styrt *et al.*, 1981; Thompson *et al.*, 1988) and deep-seated vein and stockwork systems which result in metal zonations (Kawahata and Shikazono, 1988; Alt *et al.*, 1986; Seyfried and Janecky, 1985). However, in the study area, only very little sphalerite was found supporting the view that a progressive heating sequence was responsible for the precipitation of the sulphides and not a cooling one. This conclusion is also consistent with that of pyrrhotite mentioned above.

The relatively high trace-element content in the allochemical sulphides compared with the authigenic sulphides also support a scenario where the fluids were becoming progressively hotter. Basalt/seawater experimentation shows that these base metals achieve greatest solubility in fluids between 375 - 400°C and that precipitation occurs around 425°C (Rosenbauer and Bischoff 1983; Seyfried and Janecky, 1985). The low trace element content of the authigenic sulphides (in the pillow cores) may thus have resulted from initial fluid-sulphide interaction and remobilisation below maximum solubility (i.e. $\leq 370^\circ\text{C}$), whereas precipitation in the inter pillow channelways occurred once maximum solubility had been achieved ($>400^\circ\text{C}$), resulting in the precipitation of sulphides relatively enriched in trace elements.

3.6.3 Sulphur Isotopes

The results of sulphur isotope analyses (Fig. 3.16) show the $\delta^{34}\text{S}$ -values of all sulphides to be between +1.9 and -2.0 and that the remobilisation of the allochemical sulphides did not significantly affect their $\delta^{34}\text{S}$ -values. Two questions may be asked: 1) what is the significance of such a close to zero value subsequent to fluid transport and b) is such an imprint compatible with high-temperature seafloor metamorphic origin?

To answer these questions an understanding of the effects of experimental high-temperature seawater-basalt interaction is relevant. When seawater enters into the high heatflow zones of newly formed oceanic crust, significant compositional changes take place which affect the nature of sulphides that form from discharging fluids. Shanks *et al.* (1981), who studied seawater sulphate reduction and sulphur isotope fractionation in

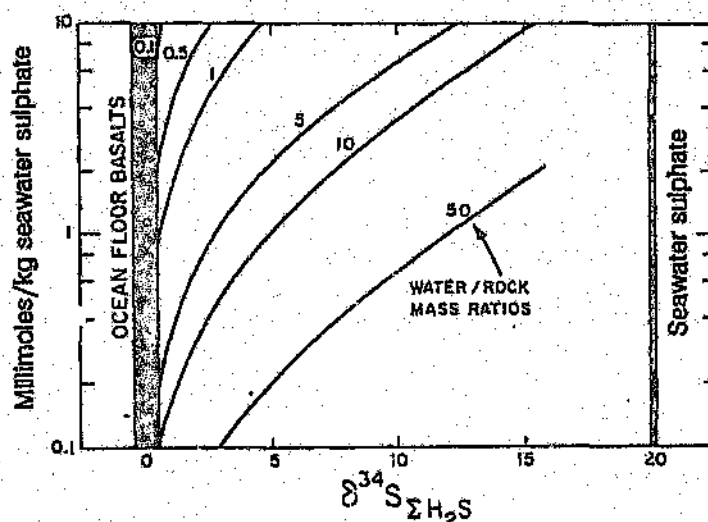


Fig. 3.17 Sulphur isotope values of total dissolved H_2S for various values of water/rock ratio. Present day seawater sulphate which enters the system is quantitatively reduced and homogeneously mixed with basaltic sulphide (after Shanks *et al.*, 1981, Fig. 11).

basaltic systems at temperatures of 200 - 350°C, show that only a small portion of the original sulphate in convecting seawater is transferred into the high temperature part of the hydrothermal system. Sulphate is largely precipitated as anhydrite on the initial heating of seawater to above 15°C. Sulphate at this stage combines with Ca^{2+} initially dissolved in seawater as well as that liberated from basalt during magnesium uptake (Bischoff and Dickson, 1975; Bischoff and Seyfried, 1978; Mottl, 1982). The small portion of sulphate entering the high temperature part of the system is entirely reduced to sulphide which utilizes only a small amount of ferrous iron in the basalt leaving behind essentially unoxidised rocks (Shanks *et al.*, 1981). Shanks *et al.* (1981) maintain that in the deeper parts of a system, direct mobilisation of basaltic sulphide remains the principal mode of interaction at most practical water/rock ratios. Only at $w/r \geq 140$ does seawater sulphate start entering the system, implying that sulphate-sulphide isotopic equilibrium is initially not established.

The sulphur isotope values of hydrothermal sulphides thus represent a mixture of reduced seawater- and basaltic sulphide, ranging from 0 to +20.3 (Claypool *et al.*, 1980; Schneider, 1970; Sakai *et al.*, 1984). The deviation of the isotopic composition of remobilised igneous sulphide, from close to zero, will thus progressively increase with

increasing water/rock ratios as demonstrated in Fig. 3.17. The findings of Shanks *et al.* (1981) corroborate the sulphur isotopic ratios of sulphides from basalts dredged from numerous seamounts and ridges of the ocean-floor (Kawahata and Shikazono, 1988). For example, a fluid-basalt system with a water/rock ratio of 5 would typically sustain a H_2S -bearing solution with a $\delta^{34}\text{S}$ value of +3 permil which in turn would give rise to pyrite with a $\delta^{34}\text{S}$ of +3.6 permil (Kawahata and Shikazono, 1988, Fig. 6; and isotopic fractionation factor of 1.2 between H_2S and pyrite at 300°C - quoted in Ohmoto and Rye, 1979).

To evaluate the extent to which Archaean seawater sulphate may have influenced the sulphides of the komatiitic basalts, it is necessary to consider whether the Archaean oceans had sulphate and what their $\delta^{34}\text{S}$ value may have been. Sulphate appears to have formed in the Archaean oceans as a result of surficial oxidation of reduced sulphur which was introduced into the hydrosphere by exhalative activity (Lambert *et al.*, 1978). Oxidation is thought to have occurred via blue-green algae, sulphur bacteria or with the aid of oxygen produced by photolysis of water vapour. The $\delta^{34}\text{S}$ values of evaporitic marine sulphates are rarely ever 1.5 permil heavier than the brine they derive from (Thode and Monster, 1965). It is thus thought that $\delta^{34}\text{S}$ values of Archaean evaporites would similarly reflect the isotopic composition of oceanic sulphate (Lambert *et al.*, 1978). These authors present data for evaporitic marine gypsum pseudomorphed by barite from Barberton (Fig Tree Group) and North Pole (Warrawoona Group) which are remarkably similar and thought to be isotopically representative of the seawater it originated from (Fig. 3.16). These marine sulphates have been shown to be distinctly different from hydrothermal sulphate of volcanogenic origin (from Big Stubby, hosted by felsic volcanics of the Duffer Formation, Warrawoona Group, Western Australia). The $\delta^{34}\text{S}$ value of Archaean seawater sulphate is thus estimated to have been close to zero (Fig. 3.16), a value slightly lighter than that of the Archaean marine sulphate precipitates (cf Lambert *et al.*, 1978; Schidlowski *et al.*, 1983; Ohmoto, 1986).

The two questions posed at the start of this section can now be answered: (1) The close to zero $\delta^{34}\text{S}$ values of the authigenic (pillow interior) and the allochemical (pillow exterior) sulphides, despite experiencing a wide range of water/rock ratios (e.g. ~ 0.3 -

30 for MC150), suggests that the S-isotope values of the circulating fluids must have been close to zero. This implies that the small amounts of Archaean seawater sulphate that entered the high-temperature system (usually at about 1 mmol/kg; cf Shanks *et al.*, 1981), must also have had a close to zero value. Such a value implies that the Archaean seawater sulphate ratio has been correctly predicted by those workers mentioned above. If the $\delta^{34}\text{S}$ of Archaean seawater sulphate had not been zero, but similar to that of modern seawater (at +20.3), a value of around +10, for example, could have been expected for the hyaloclastite sulphides of MC150 (with a water/rock mass ratio of 30 and assuming 1 mmol/kg of sulphate entered the system) (cf Fig. 3.17). (2) A close to zero $\delta^{34}\text{S}$ value for the pillow sulphides is consistent with a high-temperature seafloor metamorphic origin. If temperatures had not reached 150°C, seawater sulphate would not have been quantitatively extracted from the hydrothermal system, implying that sulphate-sulphide isotopic equilibrium would have become an important fractionation factor. This would have resulted in a grouping of the $\delta^{34}\text{S}$ values of only the allochemical sulphides closer to +5 (similar to that of marine sulphate precipitates, Fig. 3.16).

3.6.4 Alteration Geochemistry

3.6.4.1 Element Distributions. The element-distribution plots (Fig. 3.10) that show the relationships between pillow cores and the veinlet wallrocks can be divided into three groups on the basis of intra- and interpillow trends. They are: (a) Group A elements which have distribution patterns that are consistent within but vary between pillows. Elements that fall within this group are SiO_2 , FeO , MgO , CaO , and Na_2O . These elements seem to represent the main mineral components suggesting that their distributions are largely controlled by the alteration mineralogy, (b) Group B elements which have distribution patterns that are consistent within- and between pillows regardless of alteration type. Elements belonging to this group comprise Al_2O_3 , MnO , Cr_2O_3 , TiO_2 , H_2O , Y, Zr, Sr, and Ba. Possible exceptions here are Al_2O_3 , Y and Ba which seem to have complex distribution controls making them possibly intermediate to Groups A and B. Barium for example, has the distribution of the Group A elements yet appears to have no mineralogical control whereas Y is similar but its distribution seems linked to the vulnerability of mesostasis alteration, (c) Group C elements which have distribution patterns that lack consistency and are therefore classed as irregular. This group includes Fe_2O_3 , K_2O , Rb and CO_2 .

The behaviour of Group A elements is essentially the same compared with those of hydrothermally altered MORB. For MORB these element distributions are often referred to as "variable". In the study area, however, these components vary on an interpillow basis, but show good consistency on an intrapillow scale. Such detailed intrapillow comparison have not been realised for MORB. The consistency of chemical trends within individual pillows suggests that specific reaction zones within the fluid environment exist. Such variations within the fluid environment appear to confirm fluid composition evolution, similar to what is seen during basalt/seawater experimentation. Seyfried *et al.* (1988), for example, suggest that it is likely that the different fixation reactions dominate specific regions within the submarine geothermal system, as Mg-fixation characterises the downwelling limbs of seafloor convection systems, and Ca and Na-fixation those in the hydrothermal upflow zones.

The behaviour of the group B elements can be divided into two subgroups, namely those with similar behaviour to MORB and those that differ from hydrothermally altered MORB. The chemical species that appear to behave similarly are TiO_2 , Zr, Y, Sr, MnO, and $\pm\text{Ba}$ whereas Al_2O_3 and Cr_2O_3 respond differently. The general lack of a predominant mineralogical control on the distribution of these constituents suggest that they are additionally controlled by physicochemical factors such as P, T, pH, Eh and also bulk rock composition. For example, the high chromium values of pillow rims and hyaloclastite in the study area, compared to MORB, is thought to be related to the association of the pillow lavas with high Cr-bearing ultramafic lavas. The mobility of Al_2O_3 , as indicated by the widespread presence of hydrothermal and metasomatic albite, is not clearly understood but is probably the result of a combination of the above mentioned physicochemical factors including a subordinate mineralogical control. Seyfried and Janecky (1985) show that only small differences in basalt composition and crystallinity result in large differences in fluid composition.

The subgroup of elements that show similar behaviour to their counterparts in MORB comprise MnO, Sr and Ba. MnO is initially contained in minerals such as olivine and pyroxene and its mobility during basalt-seawater interaction is effected mainly by fluid acidity and temperature (Seyfried and Janecky, 1985). This supports conclusion drawn earlier that Mn distributions as well as Sr and Ba are controlled by factors other than

alteration mineralogy.

The behaviour of Group C elements (K_2O , Rb, CO_2 and Fe_2O_3 ; Table 3.6), noted for their lack of mineralogical control and inconsistent distribution suggest that their dispersion may be controlled by geological processes that occurred subsequent to seafloor metamorphism. For example, the irregular potassium and rubidium distribution may be related to the sparsely-spaced thin (≤ 0.5 mm) veinlets consisting only of microcline which crosscut the regular features (Fig. 3.4A). As microcline is not a typical seafloor alteration mineral it could have been introduced synchronous with the emplacement of numerous soda-rich granitoid plugs of the area (Fig. 1.3). The carbonate content of the statically metamorphosed rocks are thought to represent typical (background-level) values characteristic of high-temperature seafloor-metamorphosed rocks that have not yet undergone low-temperature CO_2 -addition (Staudigel *et al.*, 1981; Alt *et al.*, 1986). The controls on Fe_2O_3 distribution are not entirely clear but seem to have both local and regional geological controls. The local reduction of ferric iron appears to be marked by the patchy distribution of clinozoisite (reduced epidote), possibly via reaction 11 and regionally due to the onset of burial metamorphism (to be discussed in Chapter 4).

3.6.4.2 Elemental Fluxes. Element fluxes estimated for the alteration of the study area komatiitic basalts are summarised in Table 3.8 and are compared with the fluxes calculated for naturally and experimentally altered MORB (Humphries and Thompson, 1978a; Mottl, 1983). The figure demonstrates that most of the major oxides (ie. MgO , H_2O , Na_2O , FeO , SiO_2) of the two rock types show the same flux relationships with MgO and Na_2O , whereas two (CaO and Al_2O_3) have reversed trends. For the alteration of komatiitic basalt and MORB it is interesting to note that the similar trending oxides show similar amounts of elemental flux (*cf.* and Mottl, 1983; Humphries and Thompson, 1978a). Compared with komatiitic basalt, however, the degrees of MORB hydration are greater due to the ready formation of more hydrous chlorite the more restricted formation of less hydrous amphibole. Dehydration reactions have also occurred for the komatiitic basalts as a result of albitisation and silicification which does not seem so common for MORB.

The important negative-flux relationship between CaO and MgO established for MORB

alteration is positive for komatiitic basalt. The reversed trend of CaO is also confirmed by plotting its flux against that of Na₂O. The antipathetic relationship between MgO and CaO during MORB alteration arises from seawater derived Mg²⁺ being added to the basalt, a loss of Ca²⁺ to seawater and a concomitant increase in fluid acidity. This converts basalt to Mg-rich secondary assemblages such as chlorite. (Bischoff and Dickson, 1975; Seyfried and Bischoff, 1977; Mottl, 1983). For the study area the addition of seawater Mg²⁺ to komatiitic basalt occurred sympathetic with CaO which lead to amphibolitisation and H⁺-production. Chlorite formation is never evident in incipiently altered komatiitic basalt but is found in more extensively altered zones such as at some pillow rims.

The flux of alumina which is assumed to be zero during MORB alteration, has a marked negative flux relationship with MgO during komatiitic basalt alteration (demonstrated by reaction 3, Fig. 3.14a, Table 3.7). Seeing that all the other fluxes of the two rock-types in question correspond relative to that of MgO, it seems that only Al³⁺ and Ca²⁺ have a different behaviour, relative to that established for MORB. Several questions arise: (a) what controls the initial loss of Al³⁺ relative to Ca²⁺ during Mg-addition, (b) does the reaction succeed in H⁺-production to perpetuate hydrolysis reactions, and (c) does Al₂O₃ take part in cation hydrolysis reactions later on like CaO does for MORB.

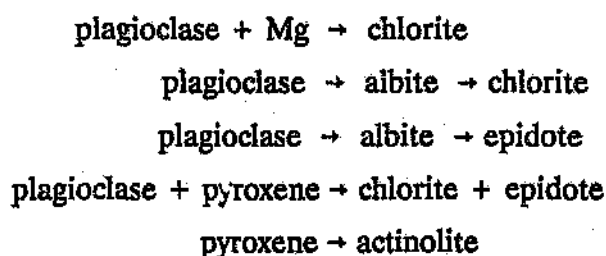
The study by Seyfried and Janecky (1985) shows that relatively small differences in basalt composition and crystallinity have significant effects on the outcome of the alteration products of their experiments. In view of this the control of bulk rock composition is explored.

The chemical differences between komatiitic and tholeiitic basalt have been assessed in Chapter 2 (Fig. 2.1) where in fact some of the study area pillow lavas can be classified as tholeiite according to the Jensen plot. At a glance, the differences in major elements between average komatiitic basalt and tholeiitic basalt (Table 2.5) are relatively small with only Al₂O₃ differing by a little more than 4.0 wt% in favour of tholeiite.

Crystallisation, however, of such tholeiitic and komatiitic melts will yield rocks with relatively large mineralogical differences (based on CIPW norm calculations). The

tholeiite will consist of approximately 50 wt% plagioclase (An_{55}) and 44 wt% pyroxene whereas the komatiitic basalt will comprise 60 wt% pyroxene and 34 wt% plagioclase (An_{55}). The hyalocrystallinity of pillowed rocks will tend to enhance the mineralogical differences even further. Plagioclase and pyroxene are the first major minerals to crystallise from tholeiitic (Gélinas and Brooks, 1973) and komatiitic basalt melts (Donaldson, 1982), respectively. Such initial crystallisation trends will favour the residual melts (destined to solidify as glassy mesostasis) to become progressively richer in the components of pyroxene and plagioclase for the tholeiitic and komatiitic basalts. Furthermore, as indicated by Donaldson (1982), and as is confirmed by the petrography of the komatiitic basalts, plagioclase is inhibited from crystallisation which implies that all plagioclase is mesostasic and, therefore, quite vulnerable to alteration. Thus, despite the relatively small chemical differences between the basalt types, it appears that relatively large phase (mineralogical) differences result during partial crystallisation which may provide a basis for the assessment of alteration products.

Mineral reactions occurring during basalt-seawater interaction at greenschist conditions according to several workers (Humphries and Thompson, 1978a,b; Rosenbauer and Bischoff, 1983), can be summarised as follows:



It is evident that plagioclase takes part in more reactions than pyroxene. Plagioclase for example, in a zone of Mg-addition, can easily be transformed into chlorite whereas pyroxene will change to actinolite. This is due to the fact that all the elements required for chlorite formation are in ready supply when plagioclase is altered in a hot Mg-rich aqueous environment, whereas in the case of pyroxene, Al_2O_3 is in short supply and therefore represents a limiting factor. The study area augite compositions clearly illustrate such an alumina deficiency (Table 2.4). Augite here is shown to have between 1 - 3 wt% Al_2O_3 , hardly enough to form actinolite but considerably below the amount

required for chlorite which is 23 wt% (Appendix: CHLORITE). From a mass balance point of view, it is easier to make chlorite from plagioclase (reaction 2) than to make it from pyroxene, which would require the introduction of large amounts of Al_2O_3 . The restricted formation of chlorite from pyroxene suggests that primary mineralogy initially exerts a strong control on the formation of secondary products. Because tholeiitic basalt contains more plagioclase than komatiitic basalt and the latter, in turn, contains more pyroxene than tholeiite it is suggested that the initial response of Mg-addition to komatiitic basalt is the formation of actinolite whereas the corresponding reaction for tholeiite leads to the formation of chlorite.

Thus, to answer the first question (posed above), it seems that primary mineralogy, which is controlled by bulk rock composition, may initially control the loss of Al_2O_3 from komatiitic basalt. As augite is preferentially replaced by alumina-poor actinolite the loss of alumina arises from the inability of the later mineral to accommodate the alumina released by the ongoing breakdown of plagioclasic glass (*cf* reaction 3). The reaction also shows that Na^+ is concomitantly lost with Al^{3+} and that H^+ is produced answering to the second part of the question, i.e. whether the reaction could succeed in perpetuating the hydrolysis reaction.

The third question, whether Al_2O_3 participates in cation hydrolysis reactions subsequent to being lost to solution, is petrographically corroborated by Figs. 2.6D, 3.6A, 3.6B, geochemically portrayed by the flux plot of Fig. 3.14a and represented by reactions 4 and 5. The incorporation of Al_2O_3 amongst others (e.g. SiO_2 and Na_2O) into rock results in the requisite formation of H^+ and interestingly the same end-products (i.e. albite $\pm$ epidote) as for Ca-fixation during MORB alteration. The only difference is that for MORB the starting material is thought to be the anorthite component in plagioclase (Seyfried and Janecky, 1985; Shanks and Seyfried, 1987; Seyfried *et al.*, 1988) whereas for komatiitic basalt it is amphibole.

Compared to MORB alteration, it is concluded that CaO and Al_2O_3 fluxes are different for komatiitic basalt because they have switched roles relative to that of MgO. All other fluxes are the same. The changed fluxes cannot be attributed to differences in composition of Archaean and modern seawater as both seawaters react similarly via

hydrolysis and Mg^{2+} -addition. The changed flux of the elements is attributed rather to the control bulk rock composition has on the crystallisation of pyroxene and plagioclase.

3.6.5 A Summary of Results

The alteration sequences, sulphide petrography, stable isotopes, alteration geochemistry and the volcanology discussed in Chapter 2 provide a basis for the reconstruction of the hydrothermal convection system which was operational in the study area. The relevant findings of this Chapter are:

- 1) The alteration types and associated textures of the study area rocks reveal that specific fluid-types have interacted with the exposed pillows resulting in essentially simple and gradational alteration styles. Typical MOR-types, with multiple fracture sets with crosscutting tendencies are conspicuously absent. These differences reflect on variations in (1) the tectonic setting of the extrusive lithologies, and (2) the general permeability of the rocks.
- 2) The alteration types of the study area indicate that fluids have essentially added either Mg^{2+} or Na^+ and Al^{3+} to the pillow rims and cooling crack selvages. The alteration types are inferred to have resulted from interaction with seawater- and rock-dominated hydrothermal fluids, respectively. These fluids are related to the intake and outlet zones of the hydrothermal convection system. The inference is based on MORB-seawater alteration studies that show down- and upwelling parts of convection systems to be associated respectively with zones of Mg- and Na and Ca fixation (e.g. Mottl, 1983; Seyfried and Janecky, 1985; Alt *et al.*, 1986; Seyfried *et al.*, 1988).
- 3) The alteration processes affecting the older komatiites and younger MORBs are essentially the same, but because of differing bulk rock compositions, the respective initial alteration parageneses are slightly different. However, once the fluid system of both rock types are sufficiently rock-dominated (ie. depleted in MgO) a convergence of the alteration products resulted as is evidenced by the common occurrence of albite, epidote, quartz and clinozoisite. Similar alteration products are especially evident for the komatiitic basalts and the high-temperature, SiO_2 -rich EPR basalt/seawater experiments of Seyfried and Janecky (1985).
- 4) For the komatiitic basalts only three out of fifteen chemical elements that have

taken part in seafloor metamorphism have significantly different distributions compared with MORB-alteration. The distribution of one of these elements, i.e. Cr_2O_3 , is attributed to the unique ultramafic country rock setting of the studied pillows whereas the other two (Al_2O_3 and CaO) appear to have changed fluxes relative to MgO as a result of bulk rock composition control.

- 5) The alteration patterns show that the sequences in which fluids reacted with each of the three pillows was different and unsystematic compared to those that affect MORB. Initial high-T assemblages overprinted by low-T assemblages, typical of MORB, is not evident for the Komati Formation samples. Instead, the Komati Formation pillows bear parageneses that remain well within greenschist facies conditions and tend to show increases in temperature as is evidenced by albite accompanied by clinozoisite and the prograde manner in which the sulphides have been precipitated. It appears that in the waning stages of seafloor metamorphism, temperatures of 400° to 430°C were experienced. The absence of CO_2 from the pillows suggests that low-T assemblages were never imprinted prior to burial metamorphism (that was to follow), and that the metamorphic path followed by the Komati Formation is different to that of typical MOR rock suites.
- 6) Fluid access and permeability of the study area seems to have been controlled only by an integrated network of volcanic structures such as thermal contraction cracks, interpillow cavities and flow contacts. In contrast, MORB undergoes multiple fracturing, suggesting that in addition to original volcanic porosity there is also a strong tectonic control on permeability. The last stage of fracture infilling to affect the study area, subsequent to the formation of cooling cracks, may or may not be related to the waning stages of seafloor hydrothermal metamorphism. The extent to which these fractures opened up the area for new fluids appears negligible.
- 7) Physicochemical conditions for the basal part of the Komati Formation probably were: (1) $P = 290 - 460$ bar; (2) $T = \sim 300^\circ - 430^\circ\text{C}$; (3) water/rock mass ratio = 0 - 2 for pillow cores and a ratio of ~ 30 for interpillow hyaloclastite; and (4) permeability = original volcanic fracture porosity. Hydrothermal metamorphism seems to have occurred in a thermally progressive environment and ceased to operate due to clogging of hydrothermal channelways by mineral precipitation and/or compaction by volcanic loading.

3.7 Discussion

3.7.1 Hydrothermal convection in the Komati Formation

It has been shown by Fehn *et al.* (1983) that the two most important controls for hydrothermal convective flow at spreading centres is (1) heat input and (2) permeability distribution. These criteria have been used below to model the convective flow for the Komati Formation along with other constraints such as volcanic structure, inferred style of emplacement and tectonic setting. The irregular paragenetic sequences observed in the Komati Formation (summarised in Table 3.10) relative to those of MOR-crust, are then explained in terms of the convective flow of the model. The model is schematically demonstrated in Fig. 3.18.

3.7.1.1 Heat input. In the previous chapter it is concluded that the bimodal maficity, volcanic structures, emplacement style and extrusive thickness of the Komati and Hooggenoeg Formations resemble an oceanic plateau rather than rocks formed at a mid-ocean ridge spreading centre. With regard to heat input, it is envisaged that the two tectonic settings differ markedly. As summarised earlier (Table 2.6) it seems that heat flow associated with oceanic plateaus is much greater than those of MOR-spreading centres (e.g. Tarduno *et al.*, 1991; Campbell and Griffiths, 1992). This is corroborated by higher extrusion rates which suggest larger melt volumes and therefore steeper geothermal gradients and higher heat flux. Oceanic plateaus are also closely associated with hotspots which are thought to arise from the activity of hot mantle plumes impinging on the base of the oceanic lithosphere (Campbell *et al.*, 1989; Richards *et al.*, 1989; Storey *et al.*, 1991). Such hotspots appear to be thermally perturbed domains of regional dimensions (Griffiths and Campbell, 1990). In the Komati Formation only two feeder dykes for the massive sheet flows have been identified (Williams and Furnell, 1979) suggesting that most of the heat was introduced by the lava flows and also by means of conductive heat flow from the underlying lithosphere (heated by the mantle plume). For the Komati Formation it is therefore conceived that heat input was mainly from above and below the hydrothermal layer. (The hydrothermal layer (Fig. 3.18) may be defined as the domain in which convective flow in the oceanic crust occurs. Its upper boundary is usually the seafloor through which free fluid-discharge and recharge occurs. The lower boundary is the limit of fluid flow penetration which is determined by rock permeability, cf Fehn *et al.*, 1983).

Heat input from the overlying hot lava flows is assumed to be less significant than the contribution of conductive heat from below due to the episodic nature of eruptions allowing for flows to cool partially between emplacements. The thickness or mass of individual lava flows is also small compared to that of the overlying ocean, suggesting that the heat would probably be dissipated efficiently by convective action into the large overlying heat sink. The hot lava flows, nevertheless, are assumed to act as layers of insulation preventing underlying crust, heated by relatively inefficient conduction, to cool as quickly as they would have without the overlying layers.

In contrast, the main heat input at mid-ocean ridges is dominated by the shallow magmatic activity along their axes (2 - 3 km deep, Seyfried *et al.*, 1988; Fehn *et al.*, 1983). This results in the formation of two main convection cells situated at either side of the axial ridge. The axial cells are locked into position and all new crust that is accreted at the axis passes through them as a result of lateral spreading. Secondary convection cells arise adjacent to the main spreading centre-driven cells. The secondary cells, however, are only driven by conductive heat from below, received from the cooling oceanic lithosphere. As a result the secondary cells migrate laterally with the oceanic plate in which convective activity decreases as the lithosphere cools (Fehn *et al.*, 1983). Heat input at MORs, for the high-T axial cells, is thus predominantly lateral (Fehn *et al.*, 1983) as opposed to vertical (from the base) of those envisaged for the Komati Formation (Fig. 3.18).

3.7.1.2 Permeability distribution. The permeability of the komatiitic rocks has been discussed earlier and is considered to differ from that of MORB regarding origin, degree and boundary conditions. The permeable structures in the komatiitic lavas are essentially of primary volcanic origin and are thought to result from an intricate network of thermal tension cracks which traverse all flows (massive and pillowed) at regular intervals (Figs. 2.2B, 2.4A) and other less abundant structures such as interpillow channel ways and flow contacts. They are envisaged to be more restricting than their counterpart structures typical found at MOR-spreading centres. These include much larger volumes of porous pillowed flows and associated pillow breccias, hyaloclastitic flows, collapse pit breccias in addition to their normal thermal contraction fractures and flow contacts. Subsequent to these features, permeability is additionally enhanced by regular rift-related tectonic

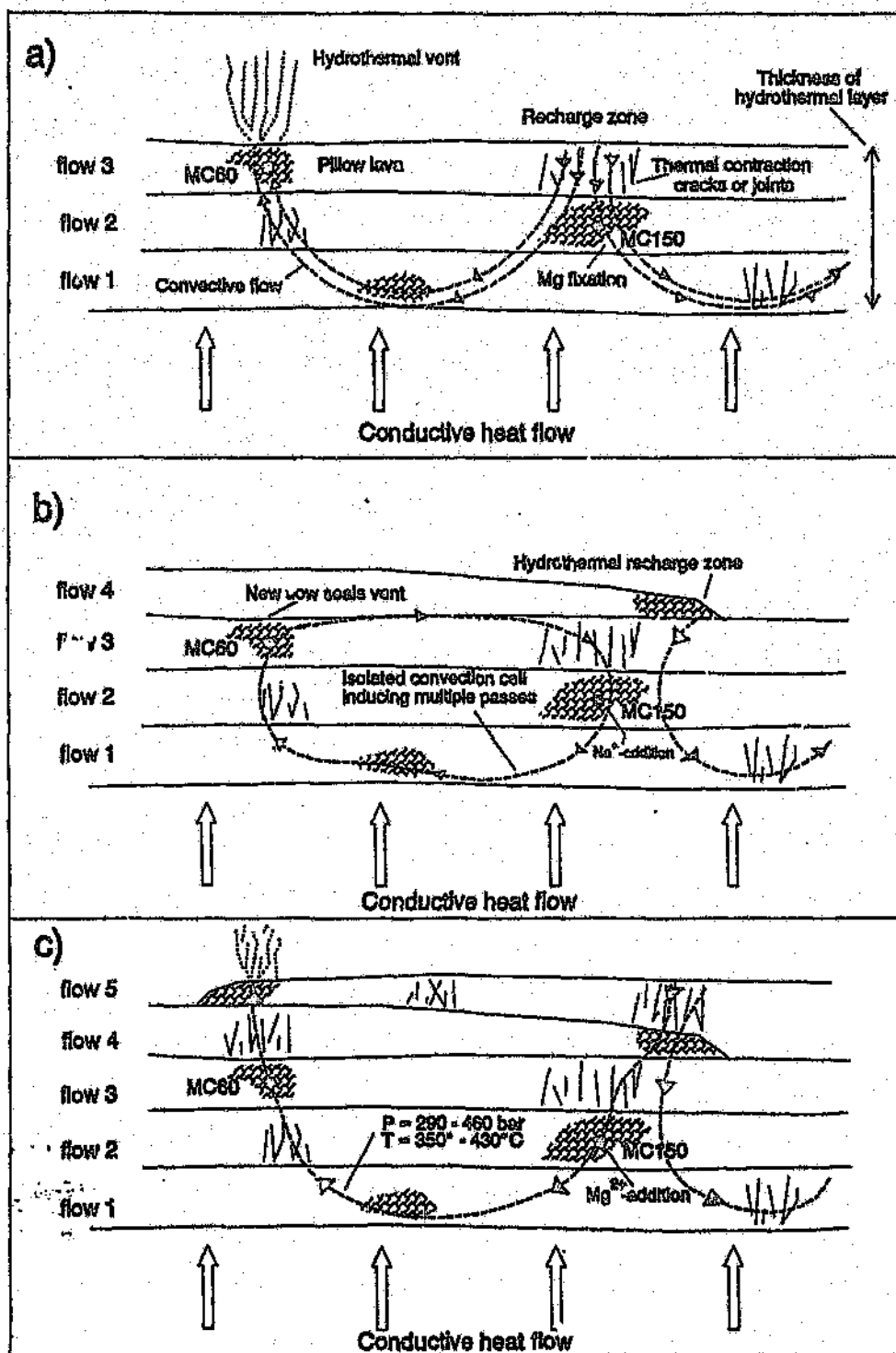


Fig. 3.18

A schematic model showing the development of high-temperature convective flow for the Komati Formation. Heat input is dominated by conduction from below and permeability is controlled largely by the original thermal-tension fracture porosity. Hydrothermal downwelling and upwelling is perceived to be mechanically controlled (21 times) by sheet flow emplacement. (See text for discussion).

fracturing keeping channelways open and creating new ones. MOR permeability, however, decreases rapidly with depth as the highly porous extrusive rocks are seldom more than 600 m thick (compared to 3.5 km thick of Komati Formation, Table 2.6). At fast-spreading centres the sheeted-dyke complexes are seldom thicker than 1 km and the magma chambers can be found at depths as shallow as 2 km. Here permeable fractures must form from relatively slow cooling, massive, coarse-grained rock by means of cracking fronts. The mechanism of crack front propagation essentially entails thermal contraction cracking associated with the encroachment of cooling (Sleep, 1983). Such cracks enable additional infiltration of water, further cooling and more (hairline) cracking. At equivalent depths in the thick volcanic sequence of the Komati Formation one would rather find well-dilated thermal contraction cracks (Fig. 2.2B). These cracks would have formed rapidly upon volcanic emplacement on the seabed. Burial of a pre-existing fracture network to equivalent depths of the magma chambers at fast-spreading ridges, may allow for relatively higher permeability and a deeper fluid penetration and therefore a deeper hydrothermal boundary.

Normally, the ocean floor is taken as the upper boundary of the hydrothermal layer of MOR-settings (Fehn *et al.*, 1983). It is assumed to allow free discharge and recharge (due to the highly porous nature of the extrusives that underlie it). However, for oceanic plateau-settings where extensive sheet-flooding is a common feature, the upper boundary may become sealed at irregular, volcanically controlled intervals. Sheet flows extending several kilometres at a time could therefore significantly disrupt the recharge and discharge activity of the underlying hydrothermal layer. Original permeability distributions may never be restored to exactly the same conditions once the overlying lava flow has cooled and undergone contraction cracking to rendered it permeable again. In such cases it is envisaged that convection cells may become temporarily isolated from the ocean waters, change flow directions as cells amalgamate or segregate and when isolated induce multiple fluid-passes (Fig. 3.18). Such fluids would become chemically evolved (rock-dominated) and increase in temperature due to inadequate heat removal. The temperature of the underlying flows (undergoing hydrothermal metamorphism) will also increase as they are exposed to successively higher geothermal gradients by way of continued burial. Continued hydrothermal activity and burial should eventually result in either mineral precipitates clogging the channelways and/or compaction thereof at which

stage no more flow occurs. Such parts of the system then become chemically closed, paving the way for isochemical burial metamorphism (to be discussed in the next chapter).

3.7.1.3 Model feasibility. The model for convective flow as presented in Fig. 3.18 and explained in principle above is entirely consistent with the volcanological findings of Chapter 2 and the alteration petrography and geochemistry discussed in this Chapter. The unsystematic way in which the alteration assemblages developed is explained, for example, by the volcanically controlled sealing (and opening) of the upper boundary of the hydrothermal layer. Convective flow is thus inadvertently manipulated by the mechanical closing and opening of recharge and discharge zones which may lead to the merging or splitting of convection cells and inevitably, changes in flow direction. For example, it was mentioned earlier (Table 3.10) that MC150 was mostly exposed to unevolved (Mg-rich) seawater excepting for a short and abrupt period of Na-metasomatism. Figures 3.18a,b,c, schematically show the development of the observed alteration sequence, depicted by Alteration Stages 1, 2 and 3 (Table 3.10). Exposure to recharge zone Mg-rich hydrothermal fluids is interrupted (controlled) by the emplacement of a massive flow (No 4, Fig. 3.18b). The flow seals-off the discharge (vent) zone of the fluids and induces the formation of an isolated convection cell. In doing so the evolved fluids for the first time interact with MC150 and albite the cooling crack selvages and pillow rim. Subsequent contraction cracking of flow 4 opens up the system and more or less restores the earlier convection pattern which continues with Mg-addition to MC150. Concomitantly, MC60 is only exposed to evolved hydrothermal fluids, depleted in Mg and enriched in SiO_2 , Na^+ and Ca^{2+} . The initial positioning of MC60 is such that isolation of the convection cell does not expose it to unevolved fluids resulting in the extensive albitisation of both cooling crack selvages and pillow rim. Such alteration sequences appear not to develop in rocks formed at MOR environments. This may be explained by the fact that the rocks spread through the main, high-temperature axial cells and then move off into convection cells with progressively lower temperature. This implies that the first formed high-temperature assemblages are overprinted by lower temperature assemblages. At these localities the rocks are usually enriched in carbonate and K-rich minerals (Staudigel *et al.*, 1981) a phenomenon which has not transpired for the Komati Formation.

The presence of pyrrhotite and pentlandite in the metasomatised hyaloclastite samples, paradoxically calls for the production of extensively altered material but also for their precipitation by very rock-dominated fluids. This may be modeled by having early alteration, by either seawater- or rock-dominated fluids prior to convection cell isolation. Isolation may result from either volcanically-controlled sealing or progressive burial and associated clogging and/or compaction blockages. In both instances, multiple passes by the fluid results in it becoming progressively more rock-dominated and temperature increases (as explained above) resulting in fluid temperature "moving" into the maximum solubility band observed by Seyfried and Janecky (1985). Continued burial and exposure to steeper geotherms cause fluid temperatures to rise above the 425°C solubility maximum for base metals which affects their precipitation under prograde conditions, as has been observed.

Such prograde hydrothermal conditions are contrary to what may be expected within the extrusive part of oceanic crust formed at a mid-ocean ridge spreading centre. In such tectonic settings it is generally observed that retrograde rather than prograde hydrothermal conditions arise with continued spreading (Alt *et al.*, 1986). General weathering and oxidation of the extrusive pile follows which would be bound to destroy pyrrhotite and pentlandite, especially at interpillow settings. Instead, in the Komati formation these sulphide assemblages have been preserved due to continued burial subsequent to seafloor metamorphism, driving the volcanic pile into a "chemically closed" environment, sealing them off from oxidising fluids.

3.7.2 Fluid circulation of the Hooggenoeg Formation.

Subsequent to the volcanic emplacement of the 3.5 km thick Komati Formation, a further 3.5 km of Hooggenoeg Formation extrusives follow. The conditions of convection of these rocks has been discussed earlier (section 2.6.2 of the previous chapter). It is shown that during Hooggenoeg Formation time seafloor alteration was characterised by fluctuating (0 - 700 m), shallow water to subaerial conditions and low temperatures (~60° - 150°C). The convection cells are thought to be relatively shallow-seated, having larger lateral than vertical dimensions (de Wit *et al.*, 1982) and resulted in the pervasive silicification of a wide range of lithologies. Heat input is inferred to be conductive and from below, similar to that of the Komati Formation but on a scaled-down size. Weaker

and shallower convection would result from a lower heat flow from below, in turn related to the waning of OOP volcanism and/or plate motion away from a relatively large but fixed intraplate hotspot.

3.8 Conclusions

The salient characteristics of Komati Formation hydrothermal metamorphism are:

- 1) High-temperature hydrothermal metamorphism has functioned chemically in a similar way to that established for modern basalt/seawater interaction. However, due largely to differing primary mineralogies, it seems evident from the element fluxes that, of Al and Ca have changed roles compared to that of MORB: the main alteration products being actinolitic amphibole and albite versus chlorite and albite for the komatiitic and tholeiitic rocks respectively.
- 2) Although similar reaction processes are involved, the resulting alteration sequences that develop are conflicting. The differences in alteration styles are explained by invoking largely mechanical mechanisms for the control of convective flow, which are consistent with the volcanic development of the rocks and their tectonic setting (i.e. oceanic plateau versus mid-ocean ridges).
- 3) The proposed model accounts for the observed irregular alteration sequences and especially the manner in which they reflect prograde and not retrograde metamorphic trends (as is observed for MORB). It also provides a convenient explanation for the onset of closed-system conditions leading to the beginning of burial metamorphism.

4. METAMORPHISM OF THE KOMATI AND HOOGENOEG FORMATIONS

4.1 Introduction

A comprehensive account of the metamorphism of the Komati and Hooggenoeg Formations based on metamorphic assemblages, mineral chemistry and geothermometry is not yet available. Apart from non-metamorphically orientated studies that make brief qualitative reference to some minerals and the greenschist-grade of the rocks (e.g. Viljoen and Viljoen, 1969 b,c,d; Williams and Furnell, 1979; Smith, 1980; López Martínez *et al.*, 1984, 1992; Hoffman 1985; de Wit *et al.*, 1987a) only three short communications addressing the metamorphism of the area have appeared (Cloete, 1990, 1991a, 1993).

The Komati and Hooggenoeg Formations on the western limb of the Onverwacht Anticline provide an unique opportunity to study the various forms and make-up of early Archaean greenstone belt metamorphism. In the Barberton Mountain Land (BML) two contrasting types of metamorphism, i.e. static- vs dynamic, are the main processes overprinting the greenstones after seafloor metamorphism and diagenesis. The phenomenon that Archaean greenstone belts often display a particular style of static- and dynamic metamorphism, was first reported for the eastern Yilgarn Block by Binns *et al.*, (1976). Worldwide, the origin of the superposed dynamic-style metamorphism is generally ascribed to the diapiric or tectonic emplacement of spatially associated granitoids (e.g. Binns *et al.*, 1976; Thurston and Breaks, 1978; Ayres, 1978; Anhaeusser, 1969, 1984; Bickle and Archibald, 1984; de Wit *et al.*, 1983; Collins, 1989). The cause of the regional static-style metamorphism, however, remote from the contact aureoles, is less clearly understood (Binns *et al.*, 1976; Thurston and Breaks, 1978; Ayres, 1978; Anhaeusser *et al.*, 1969; Anhaeusser, 1984; Bickle and Archibald, 1984).

Unlike Phanerozoic regional metamorphism where schistosity is pervasively developed on a regional scale, in the BML the interior parts of the belt are also metamorphosed on a regional scale but with schistosity only sparsely developed. In these areas, the rocks have retained much of the sedimentary and volcanic protolith structure and texture despite

reaching greenschist to lower-amphibolite facies at medium pressures (Cloete, 1990; 1993). It is only towards the granite/greenstone contacts and proximal to the major shear zones in the interior of the belt that tectonic fabric is generated. It appears, therefore, that no Phanerozoic equivalent of this style of metamorphism exists as obducted slivers of pillow-bearing ocean floor (without penetrative fabrics) typically show much lower metamorphic pressures i.e. ≤ 1 kb (e.g. Coleman, 1977; Liou and Ernst, 1979). The apparent lack of a Phanerozoic analogue of the metamorphism in the Barberton greenstone belt thus stifles its understanding, especially in view of plate tectonic-concepts.

An understanding of the metamorphic history of the Komati and Hooggenoeg Formations and an analysis of the early static-style of metamorphism should reveal the nature of the tectonic setting of the greenstones subsequent to seafloor metamorphism. Establishing the PT-conditions for static and dynamic metamorphism and knowing the conditions of seafloor metamorphism should show the passage of the metavolcanics through the early Archaean crust to the time of cratonisation and the subsequent exhumation.

The aim of this chapter is to describe the metamorphic history of the Komati and Hooggenoeg Formations subsequent to seafloor metamorphism. This entails demonstrating evidence for burial- and dynamic metamorphism. Evidence is based on rock textures, the spatial distribution of mineral assemblages, mineral chemistry, geothermometry and fluid-inclusion geobarometry. The conditions of metamorphism are discussed in the light of PTt-paths. The presence of a third type of metamorphism related to the physical displacement of the greenstone sequences during isoclinal folding and rotation to vertical is also considered.

4.2 Petrography and Metamorphic Zones

The metamorphic petrology of 14 pillowed metabasalts of the Komati- and Hooggenoeg Formations has been studied, all of which are komatiitic except MC140 which is tholeiitic in composition (Fig. 4.1). The range in mineral assemblages displayed by the mafic rocks in response to changing metamorphic grade is relatively small, yet distinct. Only two mineral

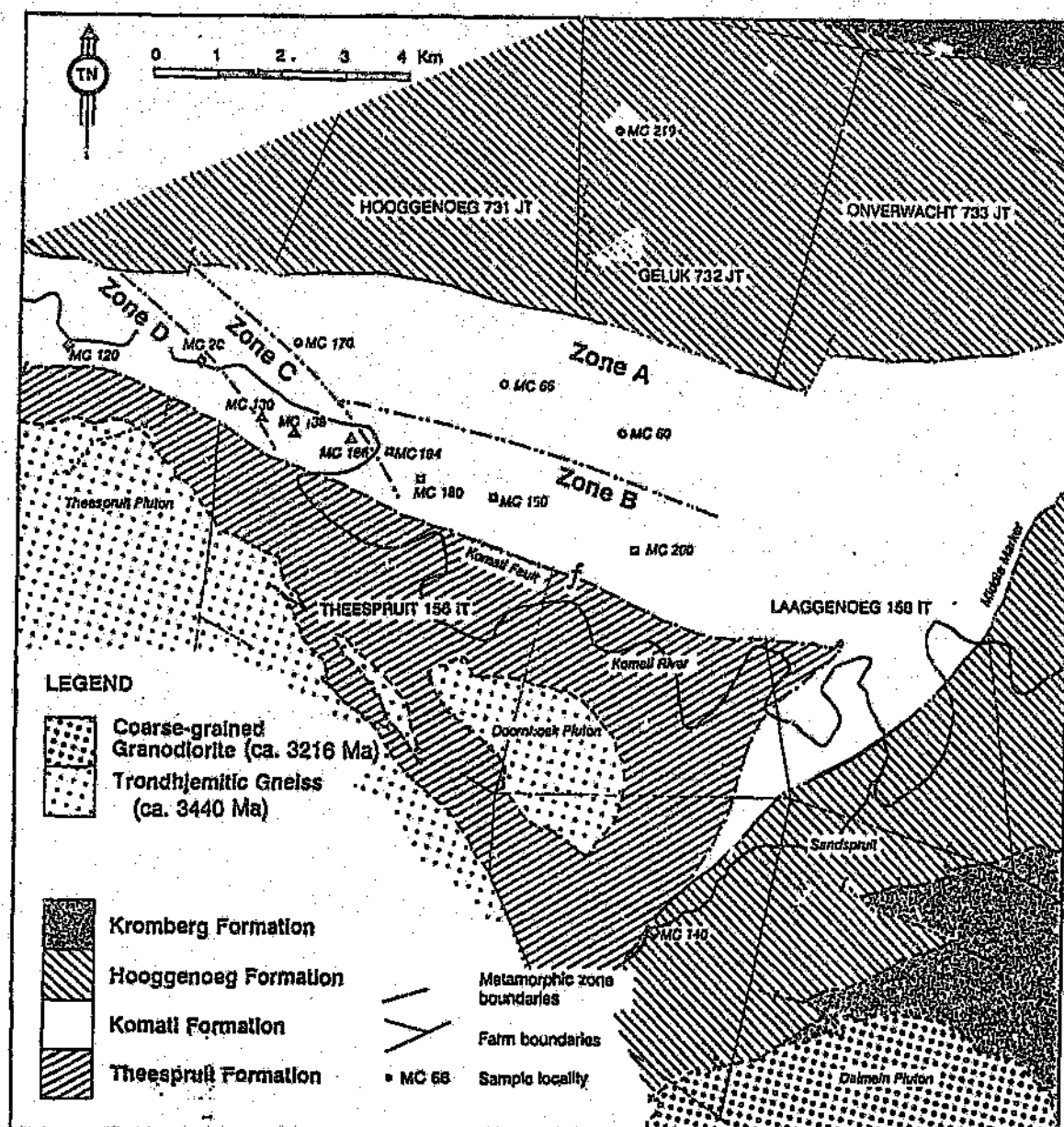


Fig. 4.1

Map of the study area indicating sample localities and metamorphic zones on the western limb of the Onverwacht Anticline. The boundaries of the metamorphic zones are inferred from the isopleth maps of the thermally sensitive minerals plagioclase and amphibole (Figs. 4.4 and 4.5, see text for discussion). The different sample locality symbols serve to accentuate the various metamorphic zones. ○ = statically metamorphosed at greenschist grade conditions; □ = statically metamorphosed at greenschist amphibolite transition zone grade; △ = statically metamorphosed at greenschist amphibolite transition zone conditions with retrograde overprint; ◇ = dynamically metamorphosed at greenschist grade conditions.

zones are found (in ascending order of metamorphic grade): albite + actinolite + chlorite + epidote/clinozoisite and oligoclase (An_{20-50}) $\pm$ albite + hornblende $\pm$ actinolite + chlorite + clinozoisite/epidote $\pm$ quartz. These mineral zones characteristically reflect metamorphic grades of the greenschist facies and the greenschist-amphibolite transition, respectively.

Apart from the mineral zones, three textural terrains are encountered along strike toward the west of the Komati Formation type-area. The type-area itself is characterised by static-style textures whereas the western-most part exhibits clear evidence of tectonic flattening (*cf de Wit et al., 1983*). In between the two, a transitional terrain is found which is marked by static-style textures with a retrogressive overprint.

As a result of the combinations between the above-mentioned mineral zones and textural terrains, it is possible to distinguish between four metamorphic areas:

- a) Zone A is characterised by statically recrystallised albite + actinolite + chlorite + epidote.
- b) Zone B is characterised by statically recrystallised plagioclase (An_{20-50}) $\pm$ albite + hornblende $\pm$ actinolite + chlorite + clinozoisite/epidote.
- c) Zone C is characterised by statically recrystallised rock with a retrogressive assemblage consisting of albite $\pm$ hornblende $\pm$ actinolite + chlorite + clinozoisite/epidote.
- d) Zone D is characterised by dynamically recrystallised albite + actinolite + chlorite + epidote/clinozoisite.

The static-style zones (A, B and C) have been determined by means of petrographic techniques as they are not easily recognised in the field. Smith (1980), however, noted from geochemical work that the western pillows (coinciding with MC130, MC138 and MC186 of this study) were distinctly more chemically-altered than the eastern pillows. The higher grade Zone B rocks have a dark grey colour whereas the lower grade Zone A and Zone C rocks are clearly green.

The difficulty in using mafic assemblages in the field to distinguish between metamorphic zones has been outlined by Laird and Albee (1981). They point out that the mafic assemblage is essentially constant in the greenschist, amphibolite and blueschist facies; i.e. amphibole + chlorite + epidote + plagioclase + quartz, but that mineral compositions and modes change continuously. Unlike in pelitic assemblages where metamorphic grade can be identified in the field by the recognition of index minerals, accurate mineral compositions must be known to do this for mafic assemblages. For this purpose mineral compositions from each of the zones have been determined using an electron microprobe. As a result mineral maps were compiled, based on ca. 700 microprobe analyses (see Appendix; AMPHIBOLE, PLAGIOCLASE, EPIDOTE and CHLORITE for listing of analyses), which show the compositional variation of amphibole, plagioclase, chlorite and epidote in the study area (Figs. 4.2, 4.3, 4.4, 4.5).

The isopleths (isocomposition lines) represent contours based on average compositions of the respective minerals from each pillow. The averages for amphibole and plagioclase represent analyses of grains that are buffered by the assemblage amphibole + plagioclase + titanite $\pm$ epidote $\pm$ chlorite $\pm$ quartz $\pm$ sulphide. The averages of chlorite and epidote represent analyses of grains buffered by the assemblage amphibole + titanite + sulphide $\pm$ plagioclase $\pm$ chlorite $\pm$ epidote $\pm$ quartz. The presence or absence of compositional gaps, i.e. peristerite¹, amphibole and pistacite², can be seen by the large or small standard deviations of the average compositions (cited in parentheses).

4.2.1 Zone A

The upper parts of the Komati- and the Hoogenoeg Formations belong to this zone (sample localities marked by dots, Fig. 4.1). The samples are characterised by superb

¹ Peristerite is an unmixed sodic plagioclase with a composition ranging between An₂ and An₂₄. The compositional gap between the unmixed phases is referred to as the peristerite gap.

² Pistacite is synonymous with epidote. The term is also used to indicate the degree of solid solution between the end member series of epidote and clinozoisite as will be discussed below. A compositional gap has been observed between epidote and clinozoisite by some workers and is referred to as the pistacite gap.

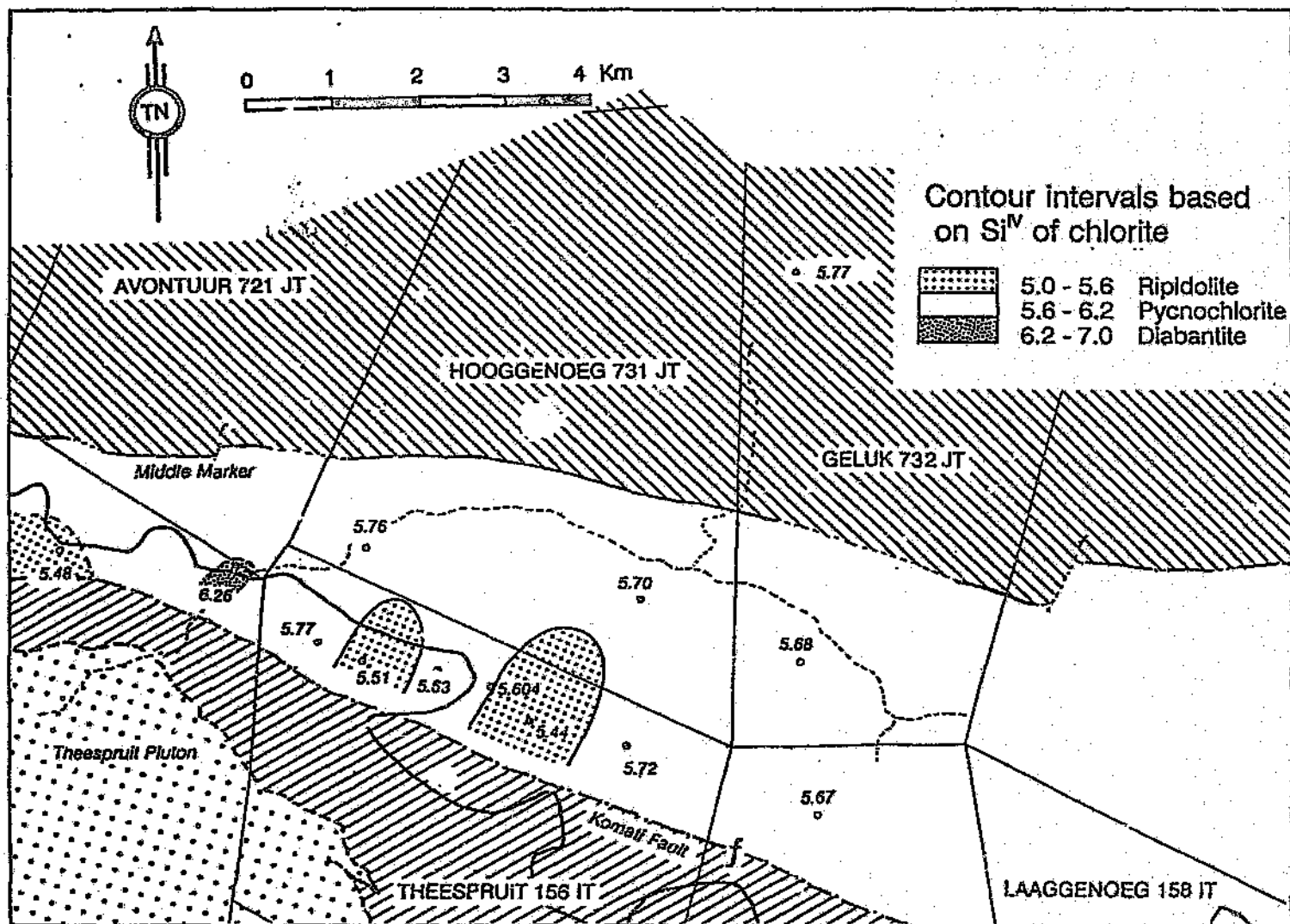


Fig. 4.2

Isopleth map for Si^{IV} in chlorite of the study area on the western limb of the Onverwacht Anticline. Chlorite nomenclature shown in legend is after Hey (1954). IMA classification of chlorites is discussed in the text and shown in Fig. 4.7. Number of chlorite analyses = 74.

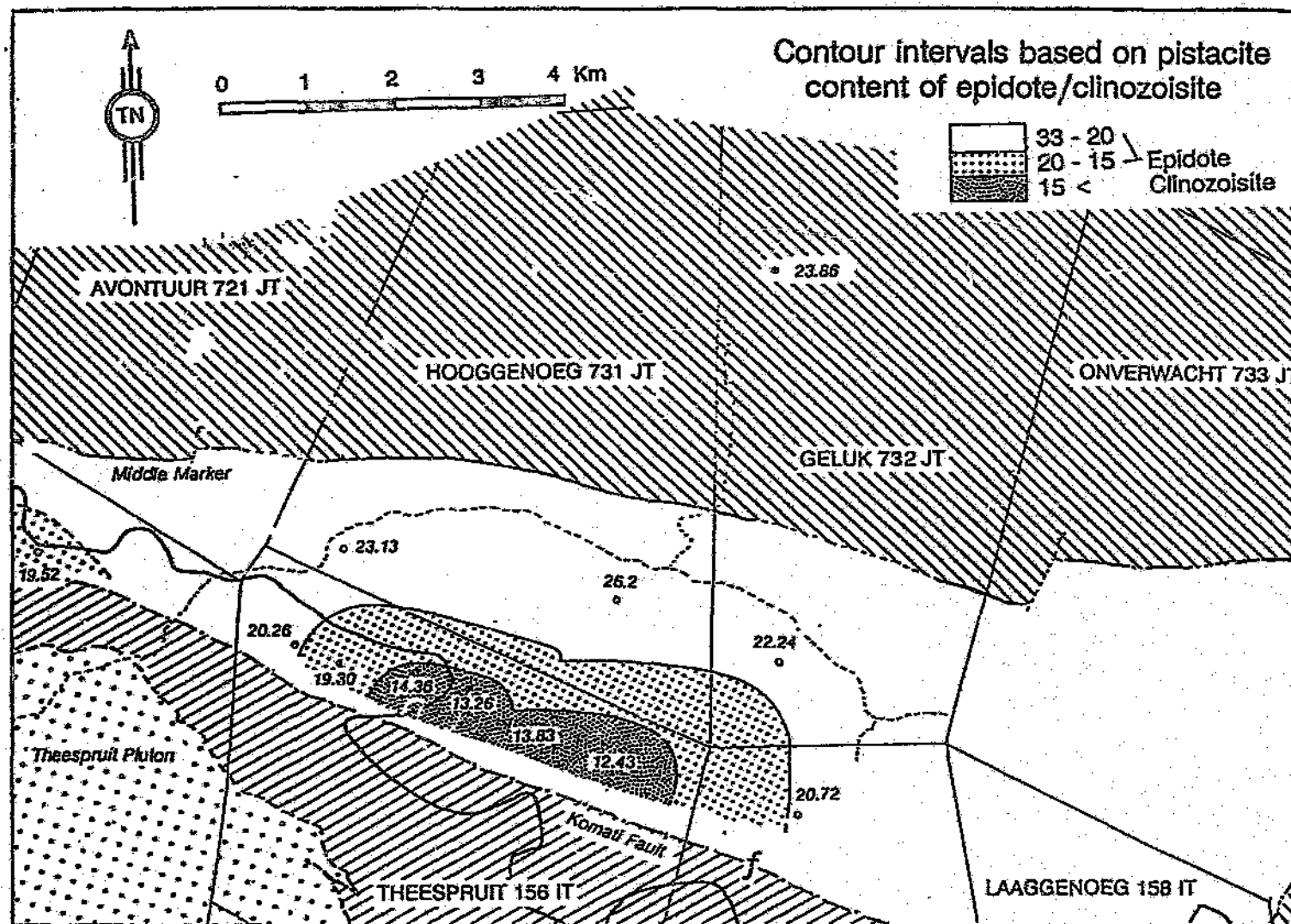


Fig. 43

Isopleth map for the pistacite ($= \text{Fe}^{3+} / [\text{Fe}^{3+} + \text{Al}] \times 100$) content of epidotes in the study area on the western limb of the Onverwacht Anticline. Contours are based on the average pistacite values of the respective pillows which involve the averaging of core and rim compositions of individual grains. Number of epidote analyses = 148.

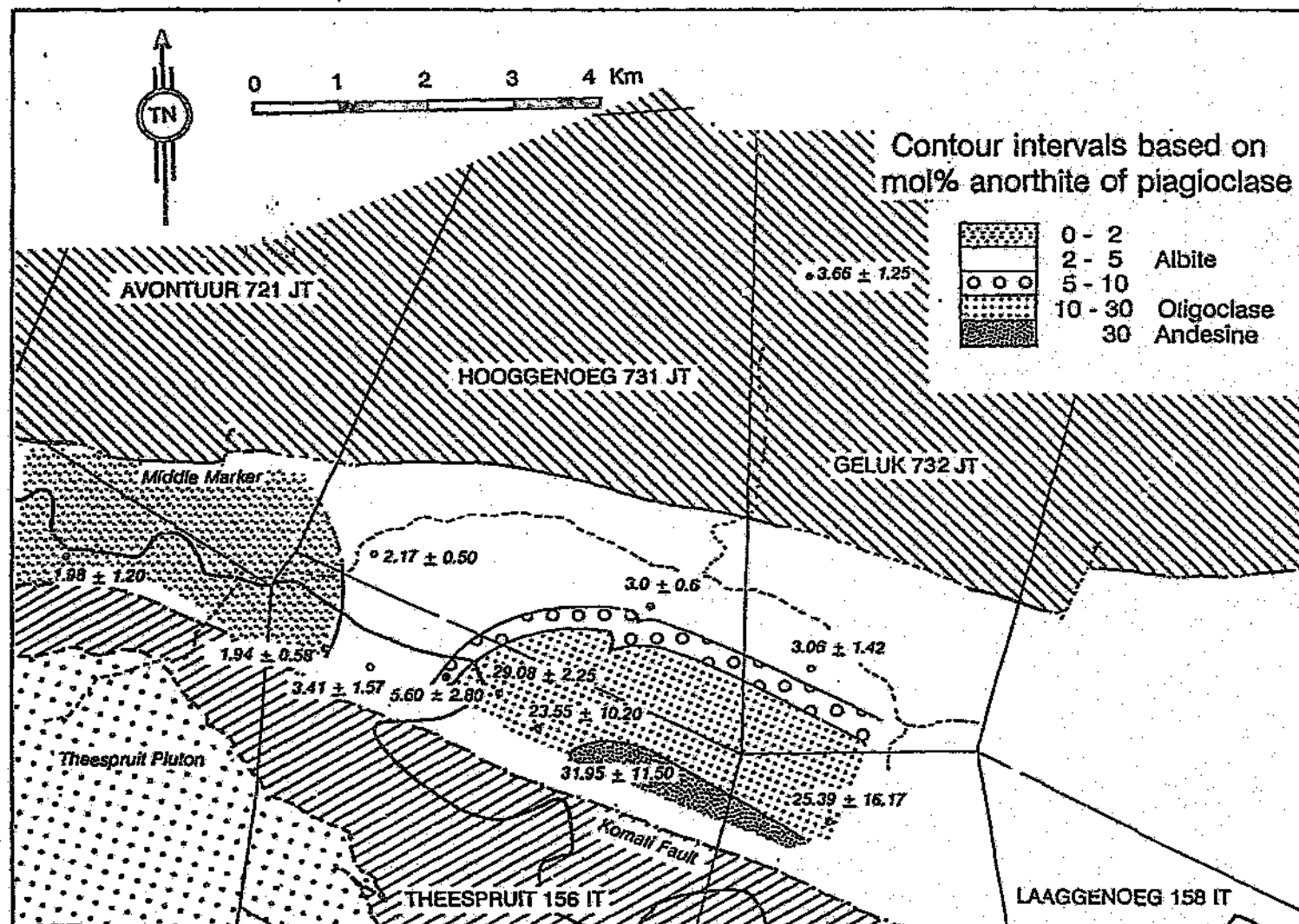


Fig. 4.4 Isopleth map for the anorthite content of plagioclase in the study area on the western limb of the Onverwacht Anticline. Contours are based on the average mol% anorthite content of plagioclase for the respective pillows. Note that large standard deviations in An-content imply the presence of peristerite pairs. Number of plagioclase analyses = 243.

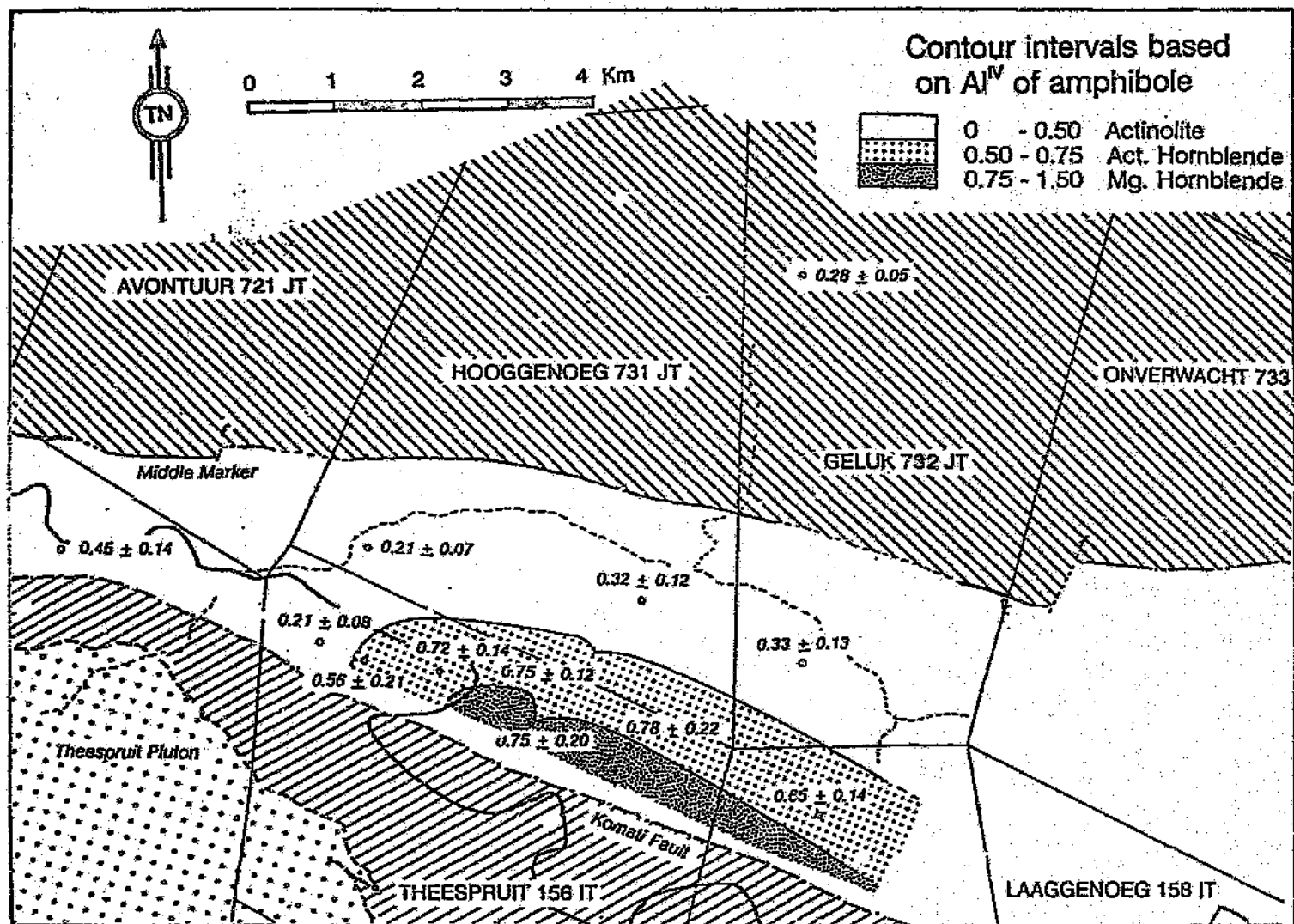


Fig. 4.5

Isopleth map for Al^{IV} of amphibole in the study area on the western limb of the Onverwacht Anticline. Nomenclature based on the average Al^{IV} -content of amphibole of the respective pillows (Leake, 1978). Note that large standard deviations imply the presence of amphibole pairs. Number of amphibole analyses = 204.

preservation of primary textures (described in Chapters 2 and 3), are free of any tectonic fabric and bear no post metamorphic-peak retrogression. The igneous minerals chromite and augite (discussed in Chapter 2) range from seeming'y unaltered to partially altered with coronas of actinolite around the augite. The dominant feature of the pillow cores is micro-spinifex-textured actinolite, pseudomorphous after clinopyroxene. Individual laths of the micro-spinifex habit do not comprise single actinolite grains but are filled with numerous, very small (< 0.05 mm) multi-orientated grains which impart a mottled appearance when viewed under crossed polars (Fig. 4.6C). The bulk of the pillow cores consist of the monotonous assemblage actinolite + titanite and where ocelli occur, albite is additional. The more extensively buffered assemblages diagnostic of Zone A, however, occur mostly within metadomains³ consisting of amygdales, phenocrystic sites, re-crystallised pillow cores adjacent to the hydrothermally-infilled thermal-tension cracks and in the lava pillow areas where metahyaloclastite is juxtaposed with infilled material. Mesostasis of the pillow cores consists of a murky brown groundmass dominated by sodic plagioclase, titanite and minor actinolite. Pillow rims, depending on the nature of earlier seafloor metasomatism, may consist of near-monomineralic albite intercrystallised with titanite dust (Fig. 3.4D, 3.11) and minor epidote or fine-grained felted to nematoblastic growths of clinocllore and actinolite delineated by the latter. The modal variability of the metamorphic minerals highlight the range of compositions of the precursor pillow structures. Bulk compositions underpinning the assemblages are, therefore, quite variable within individual pillows and even on thin-section scale.

The scattered but ubiquitous distribution of coexisting actinolite + albite + epidote + chlorite, in especially amygdales and phenocryst sites, is the diagnostic assemblage of Zone A indicating that the meta mafites belong to the greenschist facies. The average compositions of the main minerals of the assemblage that define this zone are displayed in Fig. 4.2, 4.3, 4.4, 4.5.

³ Metadomain = a macroscopically recognisable part of an altered rock that can be regarded as having a distinctive lithological or bulk chemical composition.

4.2.2 Zone B

This zone stratigraphically underlies Zone A and comprises the basal portions of the Komati Formation just above the Komati Fault (sample localities indicated by squares in Fig. 4.1). Similar to Zone A pillows, these are marked by equally well-preserved textures, have no trace of a tectonic fabric and the rocks show no signs of a retrograde overprint except for MC180. The sample shows patchy incipient sericitisation of plagioclase infilling the original cooling cracks. Most of the amygdales and recrystallised phenocrysts, however, are free of this seemingly late overprint. Retrogression of the sample may be related to its proximity to the Hooggenoeg Fault (Fig. 1.4). It is interesting to note that the retrogression appears to have affected the average compositions of all the minerals of the pillow which is manifest as a kink in the isopleths at this sample locality (i.e. MC180, Fig. 4.2, 4.3, 4.4 and 4.5).

The diagnostic assemblages of this metamorphic zone also occur sporadically throughout the pillows and are situated in amygdales (Fig. 4.6A), metamorphically recrystallised phenocrysts (hereinafter referred to as phenocrysts) and hyaloclastitic zones with favourable bulk compositions. Albite and oligoclase coexist as discrete, rounded, small grains especially in the amygdales and phenocrysts. In the hydrothermally infilled interpillow cavities away from the mafic minerals, albite is usually the only feldspar to coexist with quartz, suggesting bulk rock control of what would otherwise have been a metastable relationship. However, as the mafic minerals are approached grain size decreases and both oligoclase and albite may be detected.

Hyaloclastitic metadomains are usually dominated by nematoblastically textured assemblages defined by zoned tschermakite coexisting with clinocllore and clinozoisite. The zoned grains usually consist of a core of tschermakitic hornblende heavily dusted with titanite and clear titanite-free outer rims of tschermakite or occasionally pargasite. Different amphiboles often coexist in the absence of plagioclase in the metahyaloclastic domains and range from actinolite through to tschermakite, pargasite and cummingtonite. However, when plagioclase is present the buffered amphibole is usually magnesiohornblende. In only a few situations were two amphiboles (actinolite and magnesiohornblende) found to coexist in the presence

of plagioclase An 40 (e.g. sample MC150K14). Plagioclase coexisting with magnesiohornblende usually has straight contacts defined by the amphibole margins (Fig. 4.6C). The magnesiohornblendes, which range between needle- and spindle shaped, are usually only slightly zoned. In and around amygdales the outer, zoned margins of amphibole may overgrow the grain tips to form long wispy terminations, not to be confused with uralitisation.

The mineral assemblages of the Zone B pillows are:

- (1) actinolite + hornblende + oligoclase
- (2) actinolite + hornblende + albite + oligoclase + clinozoisite + quartz
- (3) hornblende + albite + oligoclase + quartz
- (4) hornblende + albite + oligoclase
- (5) hornblende + clinozoisite + chlorite
- (6) hornblende + oligoclase + clinozoisite
- (7) hornblende + oligoclase + chlorite + clinozoisite

Accessory phases which may be added to the listed assemblages are titanite and sulphide. No Fe-oxide phases (magnetite or hematite) were found in any of the samples. Newly formed ilmenite very rarely occurred as cores to large clusters of titanite. Microcline occasionally coexists with amphibole and plagioclase in the hyaloclastitic metadomains. Calcite is very rare in the pillows of this zone.

The metamorphic assemblages of the Zone B rocks contain both the peristerite pair and the actinolite-hornblende pair of which the pairs may or may not coexist with each other. Actinolite is not found with oligoclase to the exclusion of hornblende. Hornblende and oligoclase do not coexist to the exclusion of chlorite and/or clinozoisite-epidote. The full assemblage comprising actinolite + hornblende + albite + oligoclase + epidote + chlorite + quartz was not found to coexist as a single paragenesis (all in mutual contact) as defined by Winkler (1976). All these minerals, however, may be found within a thin section. The listed minerals are, nevertheless, characteristic of the transition zone between the greenschist

and amphibolite facies as described for naturally occurring assemblages (e.g. Turner, 1968, 1981; Heitanen, 1974; Tagiri, 1977; Maruyama *et al.*, 1983) and experimental studies (e.g. Liou *et al.*, 1974; Moody *et al.*, 1983; Apter and Liou, 1983). With regard to metamorphic grade and style, the Zone B rocks are accordingly attributed to static-style metamorphism in the transition zone between greenschist and amphibolite facies.

4.2.3 Zone C

The pillow samples of this zone (MC130, MC138 and MC186 marked by triangles in Fig. 4.1) are in several respects similar to those of Zone B. The similarities are: (a) the pillows occur along the faulted stratigraphic base of the Komati Formation (along strike immediately to the west of Zone B); (b) they are stratigraphically overlain by Zone A; (c) the pillows are not flattened nor do they have tectonic fabric; and (d) the significant mineral assemblages are also mostly situated in amygdales, phenocrystic sites and in zones where earlier hydrothermal alteration has prepared favourable bulk compositions.

The pillows differ, however, from the B Zone samples in that their assemblages have a distinct alteration overprint. Plagioclase grains are partly altered to sericite $\pm$ chlorite and hornblende to actinolite $\pm$ chlorite. Of the two, plagioclase is preferentially altered. The isopleth maps also confirm the preferential alteration of plagioclase. This relationship is schematically demonstrated, for example, by the An_{10} isopleth positioned to the "outside" of the actinolitic hornblende ($Al^IV = 0.5$) isopleth in Zones A and B where essentially none of the above mentioned alteration occurs (Fig. 4.4 and 4.5). However, as Zone C is approached (along strike to the west) the An_{10} isopleth crosses the actinolitic hornblende isopleth and thus becomes positioned on its "inside", at locality MC186 (view Figs. 4.4 and 4.5 in conjunction with Fig. 4.1). The crossing of isopleths suggests disequilibrium and thus confirms the petrographic observations of retrograde overprinting of the Zone C metamorphic assemblages. Such a conclusion is also consistent with the geochemically based observation of Smith (1980) that the "western pillows" are more altered than the "eastern pillows".

4.2.4 Zone D

The two westernmost pillows of the study area (MC120 and MC20), situated to the north of the Theespruit pluton and the easternmost sample on the eastern limb of the Onverwacht Anticline (MC140), positioned between the Doornhoek and Dalmein plutons, belong to Zone D (indicated by diamonds, Fig. 4.1). The pillows have flattened shapes but do not necessarily have a pervasive penetrative fabric. The southeastern sample (MC140 corresponds to sample V27 of Viljoen and Viljoen, 1969e) situated in the Sandspruit stream, is the most flattened pillow and has an isoclinally folded pillow margin with fold axes trending NE. Of the three deformed pillows this is the only one with a pervasive tectonic fabric. MC120 (Fig. 3.2) still contains some relics of protolith textures such as ocelli and amygdales. The microspinifex textures, that dominate the static-style pillows, however, have been replaced pervasively by massive actinolite with no preferred orientation. The ocelli bearing zones consist of xenoblastic actinolite and actinolitic hornblende with small (< 0.1 mm) clots of albite which seem to represent the original plagioclase-rich mesostasis of the protolith. Here, however, the murky brown albites are recrystallised to clear albite. Amygdale sites, which may include reshaped and rounded phenocrysts, are the zones that more readily contain the common assemblage. The common assemblage is also found more frequently in MC140 which may be attributable to its tholeiitic bulk composition. Compared with the other komatiitic assemblages it contains appreciably more albite and chlorite and less amphibole.

MC20, however, has carbonate in most of its assemblages and the minerals collectively impart a felted texture, free of preferred orientation despite the flattened pillow shapes. This relationship suggests that the CO_2 -rich assemblage is post-tectonic. The assemblage comprises stilpnomelane + calcite + chlorite + ilmenite $\pm$ sulphide.

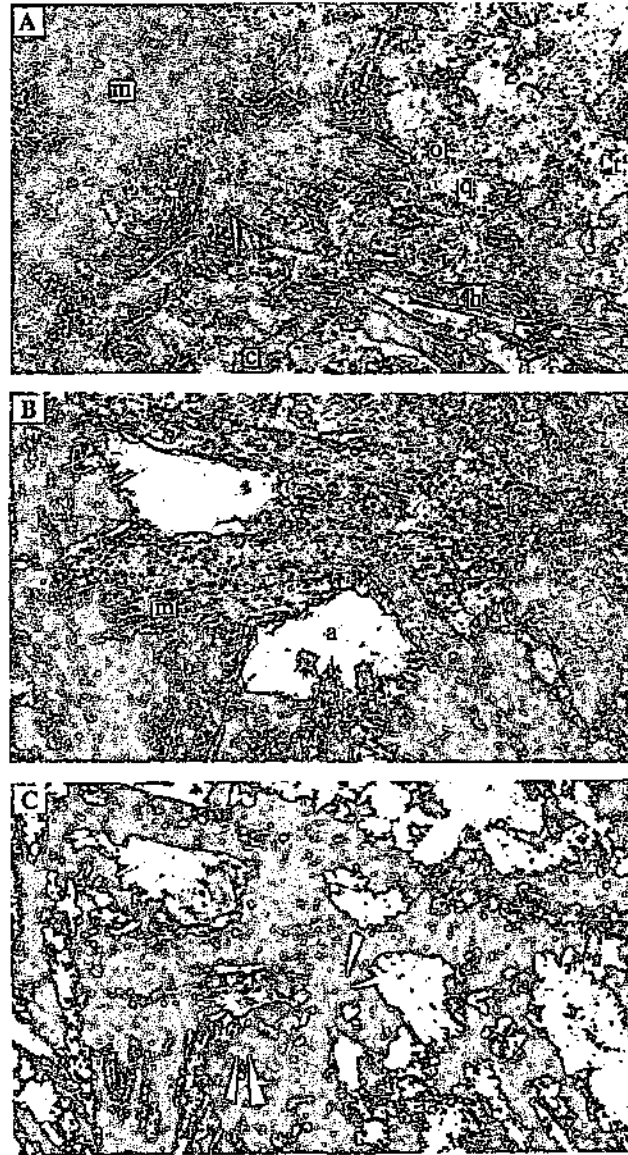
The metamorphic assemblages of the pillows in this zone attest to the greenschist facies of metamorphism (e.g. Winkler, 1976). Due to superposed deformation, especially as evidenced by flat pillow habits, the samples of this group are referred to as the *dynamic-style* pillows of Zone D.

4.2.5 Bulk Composition

From the petrographic descriptions thus far it is evident, especially for the static-style pillows, that modal proportions of minerals often vary over short distances which imply varying bulk composition. Obtaining bulk composition analyses from the small textural domains at which the assemblages occur would be virtually impossible. This poses a problem for the comparison of amphibole compositions as it has been pointed out that bulk rock composition has a great effect on the equilibrium compositions of amphibole and plagioclase (e.g. Leake, 1965; Miyashiro, 1968).

Laird and Albee (1981) refer to the mafic assemblage amphibole + chlorite + epidote + plagioclase + quartz + Ti-phase + Fe-oxide as the common assemblage. They also regard the compositions of the specific minerals to be buffered by the assemblage as a whole and therefore to be essentially a function of P and T, rather than bulk composition (differing bulk compositions, within the plagioclase, amphibole and chlorite field on a ACF-diagram will result in different modal proportions of the mentioned minerals but still yield the common assemblage). Hynes (1982), who compared amphiboles of medium- and low-pressure metabasites, found the buffering of mineral compositions within the common assemblage to be a broadly acceptable basis for comparison. He found for example that the Mg/Fe bulk-rock ratio does not affect amphibole compositions significantly (for comparative purposes) but accepted that the temperatures of reactions are influenced by the ratio. Brown (1977) found that Fe-oxide buffers the glaucophane content in amphiboles so that if iron oxide is absent, lesser than maximum amounts of glaucophane occur. The general absence in the study area of iron oxide phases coexisting with amphibole thus prohibits glaucophane-content comparisons to those of other amphiboles coexisting with Fe-oxides. As the study area rocks have not formed in a high-pressure environment (blue amphibole not present), glaucophane-content comparisons are not regarded as important. Fortunately, because of experimental studies, the phase relations between greenschist and amphibolite facies in the basaltic system at various oxygen fugacities are known (e.g. Liou *et al.*, 1974; Moody *et al.*, 1983; Apter and Liou, 1983), and allow for relative comparisons to be made.

Figure 4.6



- A Photomicrograph of a typical amygdale assemblage (lower right half) and komatiitic metabasalt host (upper left half). The host consists of titanite inclusion-rich magnesiohornblende (m) pseudomorphous after augite microlites (cf also Fig. 2.5B). The amygdale consists of clear magnesiohornblende (h), inclusion-rich oligoclase (o), rounded quartz (q), turbid, fine-grained clinzoisite (c) and clear titanite (t). Sample MC184e. Plane polarised light. Length of field of view = 0.5 mm.
- B Photomicrograph of microspinel-textured rock showing recrystallised mesostasis (a) interstitial to amphibole laths (m). The mesostatic plagioclase is andesine (An 35 - 40) and has a clear aspect compared with most other mesostasis of the same thin section which are distinctly turbid as shown in Fig. 2.5D. Amphibole (m) is magnesiohornblende and is pseudomorphous after augite and the associated high relief inclusions are titanite. Plane polarised light. Sample MC150c. Length of field of view = 0.5 mm.
- C Photomicrograph of the same field of view as shown in the previous figure (Fig. 4.6B) but with crossed nicols to emphasize the extent of recrystallisation. Note that magnesiohornblende (m) shows mottled extinction suggesting that its lath-like habit consists of several crystals and not only one. Andesine of the central mesostatic patch has granoblastic shapes and grain boundaries with adjacent amphibole are straight. When analysed with the microprobe, e.g. at arrow tips, consistent geothermometer temperatures are produced employing the method of Spear, 1980 (cf Fig. 4.16).

As the common assemblage is not always present in the study-area metabasites, comparison of mineral compositions is therefore not without complication. This problem is more apparent for the higher grade Zone B rocks where fewer phases are normally present than constitute the (full) common assemblage. Buffering of amphibole compositions with fewer phases is tested below (in the next section) and this shows that those that are buffered by plagioclase-bearing assemblages do not vary significantly from those buffered by the common assemblage (always less Fe-oxide), but are notably different from the plagioclase-absent ones.

4.2.6 Attainment of Equilibrium

Spry (1969) maintains that a mineral assemblage in a rock may be regarded to be in equilibrium if the mineral phases look to be stable and not inclining to be breaking down individually or to be reacting with each other. He also points out that no experimental, mineralogical or petrological criteria can tell unequivocally that an assemblage is stable. An indication, however, of the degree of equilibrium attained by metamorphic assemblages in a thin section, or rocks of the same area, may be found in the reproducibility of various geothermometer temperatures (Rollinson, 1982). Accordingly, the position taken here to demonstrate this supposes that on a scale of tens of microns, equilibrium was reached across mineral grain boundaries. Therefore, for the geothermometric and geobarometric calculations mineral analyses have been made at grain edges of adjacent pairs unless otherwise stated (*cf* Figs. 4.6B, 4.6C).

4.3 Mineral Chemistry

4.3.1 Chlorite

Chlorite is common in Zone A, B, C, and D samples and in accordance with the nomenclature of the IMA (International Mineralogical Association) (Bailey, 1980), it can be classified as clinochlore (Fig. 4.7; see Appendix CHLORITE for list of analyses). After the nomenclature of Hey (1954), however, of which the various species apparently have little structural significance (Brown and Bailey, 1962), the chlorites are all ripidolite and pycnochlorite, except for diabantite which occur in the Zone D carbonated rocks (MC20, Fig. 4.2). In accordance with

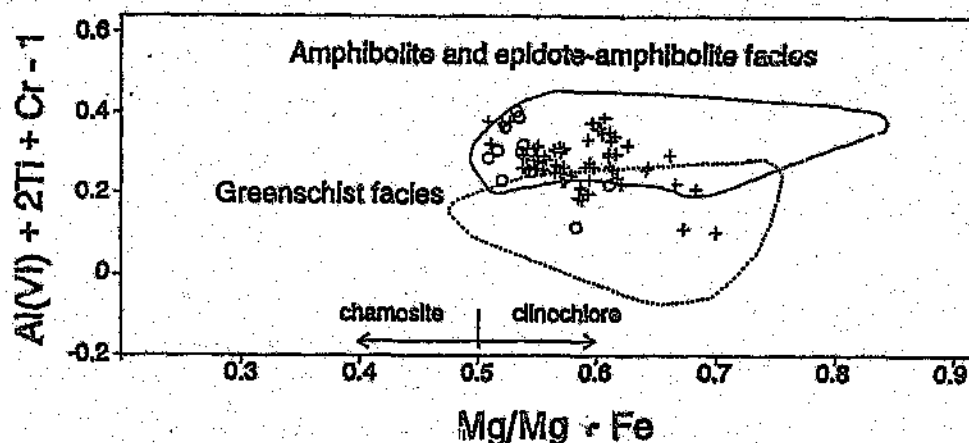


Fig. 4.7 Chlorite from komatiitic metabasalt in a plot of formula proportion $Al^{IV} + 2Ti + Cr - 1$ vs atom ratio $Mg/(Mg + Fe^{2+})$. The atom ratio $Mg/(Mg + Fe^{2+})$ indicates their clinochlore vs chamosite content. Envelopes distinguish greenschist facies from epidote-amphibolite plus amphibolite facies chlorites from worldwide occurrences (Laird, 1988). Crosses represent chlorite from Zones A, B and D (undifferentiated due to poor separation) and circles chlorite from Zone C.

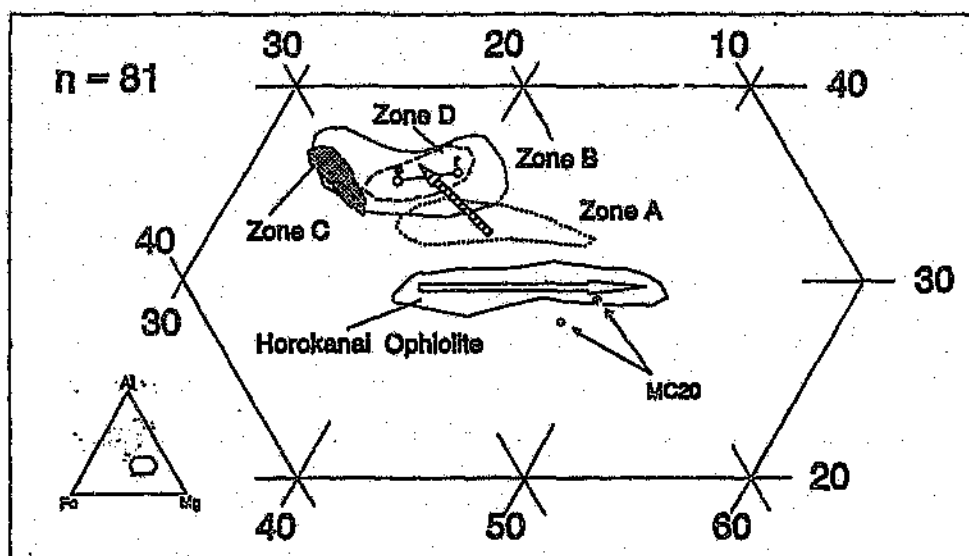


Fig. 4.8 Compositional variation of chlorite in terms of atomic Al, Fe^{2+} and Mg from the komatiitic metabasalts and from the Horokanai Ophiolite (Ishizuka, 1985). The hatched and clear arrows show trends of increasing grade for the Onverwacht and Horokanai metabasites, respectively.

the IMA nomenclature, specific chlorite compositions will be expressed in terms of mol percent clinochlore ($\text{Mg}/\text{Mg} + \text{Fe}^{2+} \times 100$). For example, the range of chlorite analysed in this study is $\text{Cln}_{(51-70)}$, where all iron is assumed to be ferrous (after Laird, 1988). In comparison to this method Figure 4.2 displays chlorites after the nomenclature of Hey (1954).

Chlorites of individual pillows have variable Fe/Mg ratios as a result of variable bulk compositions of the pillow domains (i.e. rims, core, amygdalae and hyaloclastite). Chlorite compositions within the individual domains, however, remain fairly constant and rims and cores of grains usually show little, if any variation. Variations, when encountered, especially in the progressively metamorphosed Zone B rocks, tend to show grain rims enriched in clinochlore. A core- (c) and rim (r) pair showing a particularly large deviation are joined with a tie line in Fig. 4.8. The chlorites of Zones A and B and to a lesser extent those of Zone D show a moderate spread with regard to their Mg/Fe content, the Zone C chlorites show a much reduced spread, characteristically favouring a iron-rich composition (Figs. 4.7 and 4.8).

The $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio of the chlorites decreases with increasing metamorphic grade as is demonstrated by the progressively metamorphosed rocks of Zone A and B (Figs. 4.8b, 4.8c). Similar compositional trends have been described from other metamorphic terrains (e.g. Maruyama *et al.*, 1983, Fig. 3; Laird, 1988, Fig. 19). In contrast, Ishizuka (1985) found the $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio of the Horokanai Ophiolite to remain constant with increasing metamorphic grade. The Horokanai rocks, which have a relatively fixed FeO/MgO bulk ratio, have chlorites of which the iron/magnesium ratio decreases with increasing metamorphic grade (Fig. 4.8, clear arrow symbol). It is interesting to note that the core and rim pairs, mentioned above (Fig. 4.8, c-r tie line), also show the same iron/magnesium trend, suggesting the presence of a subtrend within the larger one (hatched arrow). From experimental work it is known that the Mg-content of chlorite increases with increasing temperature of formation (Moody *et al.*, 1983).

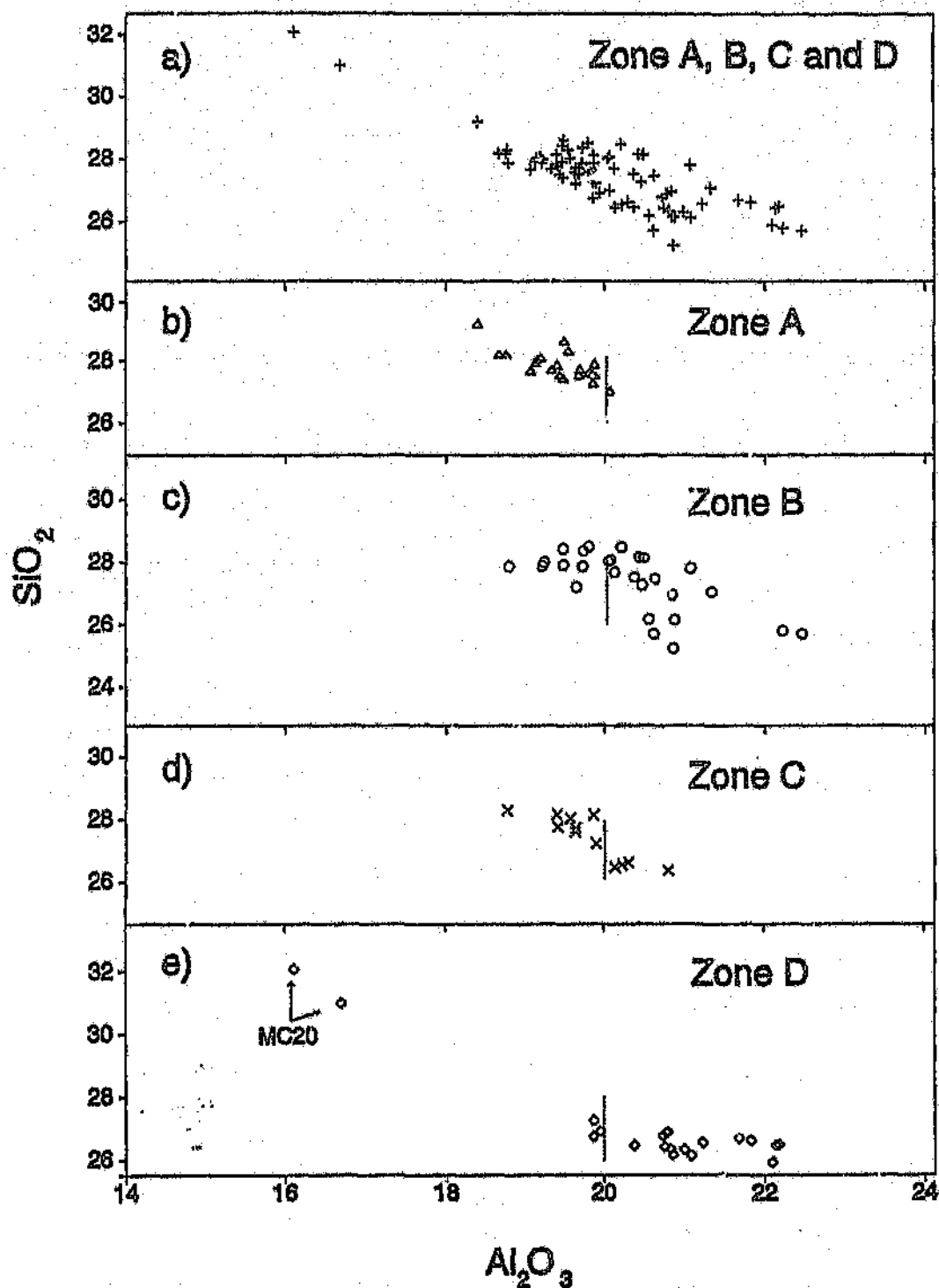


Fig. 4.9

SiO_2 wt% vs Al_2O_3 wt% plot for chlorite in komatiitic and tholeiitic metabasalt pillow lavas of the study area. The dashed line is intended to aid comparison. See text for discussion.

Both the retrogressively and dynamically metamorphosed rocks of Zones C and D belong to the greenschist facies but have chlorites which plot within the field of the higher-grade Zone B metabasites (Fig. 4.8). They notably do not plot within the field of the isofacial Zone A rocks. The chlorites of carbonated rocks (labelled MC20), which coexist with stilpnomelane, plot far removed. They do, however, share the same overall trend with the rest of the samples in Fig. 4.9 and plot at the low-grade end of the range according to the trend established by Maruyama *et al.*, (1983). These samples also plot antithetically with respect to the progressive trend (hatched arrow, Fig. 4.8) established for the study area chlorites, suggesting that carbonatisation of the dynamic-style pillows is retrograde in nature.

Laird (1988) reports that the TK exchange vector in chlorite (i.e. $Al^{VI} + Cr + Ti - 1 =$ Tschermack substitution) increases from greenschist to amphibolite facies. The exchange vector combined with $Mg/Mg + Fe$, for example, shows a positive relationship for worldwide occurrences of chlorite in mafic schist (i.e. offset of stippled- and solid envelopes in Fig. 4.7). The statically-metamorphosed greenschist-grade chlorites analysed here (not differentiated) have a TK-content marginally higher than the world-wide average, allowing for approximately half of them to plot in the amphibolite facies field (Fig. 4.7). The Zone C chlorites (differentiated as open circles) plot as the most distinct sub-population and tend to cluster on the chamosite side of the population as a whole. The iron-rich affinity of the chlorites may be linked to the retrograde overprint observed for these rocks as opposed to the prograde tendency to increase Mg. A general sensitivity for late fluids is also portrayed by the chlorite isopleth map which shows the increase of ripidolite coinciding with their proximity to faults (*cf* Fig. 1.4 for faults and Fig. 4.2 for chlorite distribution).

4.3.2 Epidote

Epidote is common in all metamorphic zones except for the carbonate-rich dynamic-style pillows (MC20) of Zone D (see Appendix EPIDOTE for list of analyses). The average pistacite content of epidote ($Ps = Fe^{3+}/(Fe^{3+} + Al) \times 100$) of individual pillows is shown on the isopleth map (Fig. 4.3) and its frequency distribution in the various metamorphic

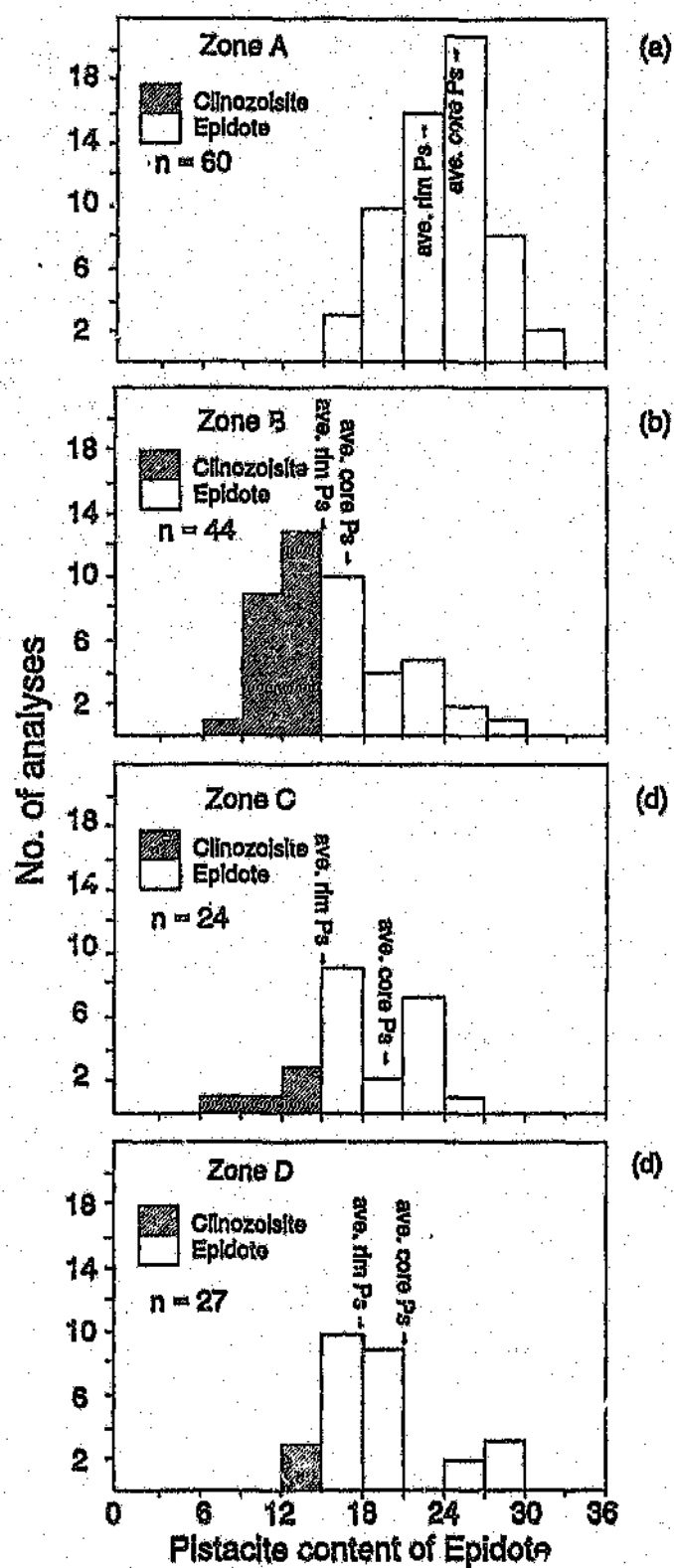


Fig. 4.10 Frequency histogram of epidote compositions from the metamorphic zones as indicated.

zones is shown in Fig. 4.10. The only opaque phases to coexist with the pistacites are sulphides (chalcopyrite, pyrite $\pm$ pyrrhotite). The epidotes of the static-style area show a progressive decrease in pistacite content from Zone A to Zone B, shown by abundance maxima shifts from Ps₂₅ to Ps₁₅, respectively. The corresponding ranges shift from Ps₃₃₋₁₅ to Ps₃₀₋₆, respectively (Fig. 4.10). It is evident from the figure that Zone A only has epidote whereas Zone B has both epidote and clinozoisite.

The Zone A epidotes are only occasionally prominently zoned with the cores being more iron-rich (ave. Ps = 25) than the attendant rims (ave. Ps = 22.7). Despite the zoning their compositions display a normal distribution and no signs of a compositional gap can be observed, which according to some workers, is present around Ps₁₇ (cf Deer *et al.*, 1986).

Zone B has zoned epidotes which may also be mantled by clinozoisite and relatively unzoned grains consisting only of clinozoisite which often poikiloblastically enclose sulphide. The cores and rims of the epidotes and clinozoisites average at Ps_{16.6} and Ps_{15.1}, showing a tendency to be less zoned compared with those of Zone A. Both the shift in frequency peaks and compositional ranges of the analysed epidotes in Zone A and B suggest that the clinozoisite component of epidote increases with increasing metamorphic grade, a phenomena that is often reported from progressively metamorphosed environments (e.g. Nakajima *et al.*, 1977; Maruyama *et al.*, 1983; Ishizuka, 1985). The Zone B averages are decreased somewhat by the relatively high values of the isofacial sample MC200. The higher pistacite values of this sample is clearly displayed by the epidote isopleth map suggesting sharply changed fO_2 conditions for the eastern part of the study area (Fig. 4.5). It is long known that the pistacite content of epidote/clinozoisite is not only affected by temperature but also by oxygen fugacity (e.g. Holdaway, 1972; Liou, 1973; Fiala *et al.*, 1980) and by the partial pressure of CO₂ (Sivaprakash, 1982). The epidote isopleths are, therefore, not considered to be linked to or diagnostic of metamorphic grade.

The Zone C and D epidote-clinozoisites tend to have bimodal pistacite distributions. Despite the greenschist grade assemblages of the zones, they notably have pistacite values

which are more akin to the higher-grade Zone B rocks. The core and rim averages for Zones C and D are $Ps_{19.5}$ vs $Ps_{14.9}$ and $Ps_{21.2}$ vs $Ps_{17.7}$ respectively (Fig. 4.10c, d). The greater divergence between average core and rim compositions of these rocks compared to the static-style zones reflect a changed metamorphic- and f_{O_2} environment. The possible origins of which will be discussed in section 4.6.

4.3.3 Plagioclase

Plagioclase occurs in all metamorphic zones except in the carbonated rocks of Zone D (see Appendix PLAGIOCLASE for analyses). The anorthite content for individual pillows is illustrated by the plagioclase isopleth map (Fig. 4.4) and the compositional variation of the various metamorphic zones is shown in the frequency vs anorthite-content plots of Figure 4.11.

Zone A plagioclase is albite and has a narrow compositional range and an average An-content of 2.85 mol% (Fig. 4.11a).

Zone B plagioclase is characterised by a wide compositional range, i.e. from albite through oligoclase to andesine (An_{0-45} , Fig. 4.11d-g). Individual metadomains either contain albite and oligoclase as peristerite pairs or only one plagioclase (i.e. $An_{\pm 10-45}$) which coexist with one or two amphiboles. Wide compositional ranges for plagioclase are relatively unique as they appear to form just above the consolute temperature of the peristerite solvus, which is narrowed at slightly higher grades (Maruyama *et al.*, 1983). Samples MC200 and MC150 of Zone B show the presence and the closure of the peristerite solvus respectively. The presence of peristerite pairs and plagioclase with consolute compositions within the same thin section and along strike may be attributed to localised variations in bulk composition, kinetics, localised equilibrium and other parameters such as f_{O_2} (cf Maruyama *et al.*, 1983). Plagioclase with peristerite and consolute compositions, however, occurs in a narrow temperature interval in P-T-space and thus constrains the conditions of metamorphism for Zone B (discussed in section 4.6 "Conditions of Metamorphism").

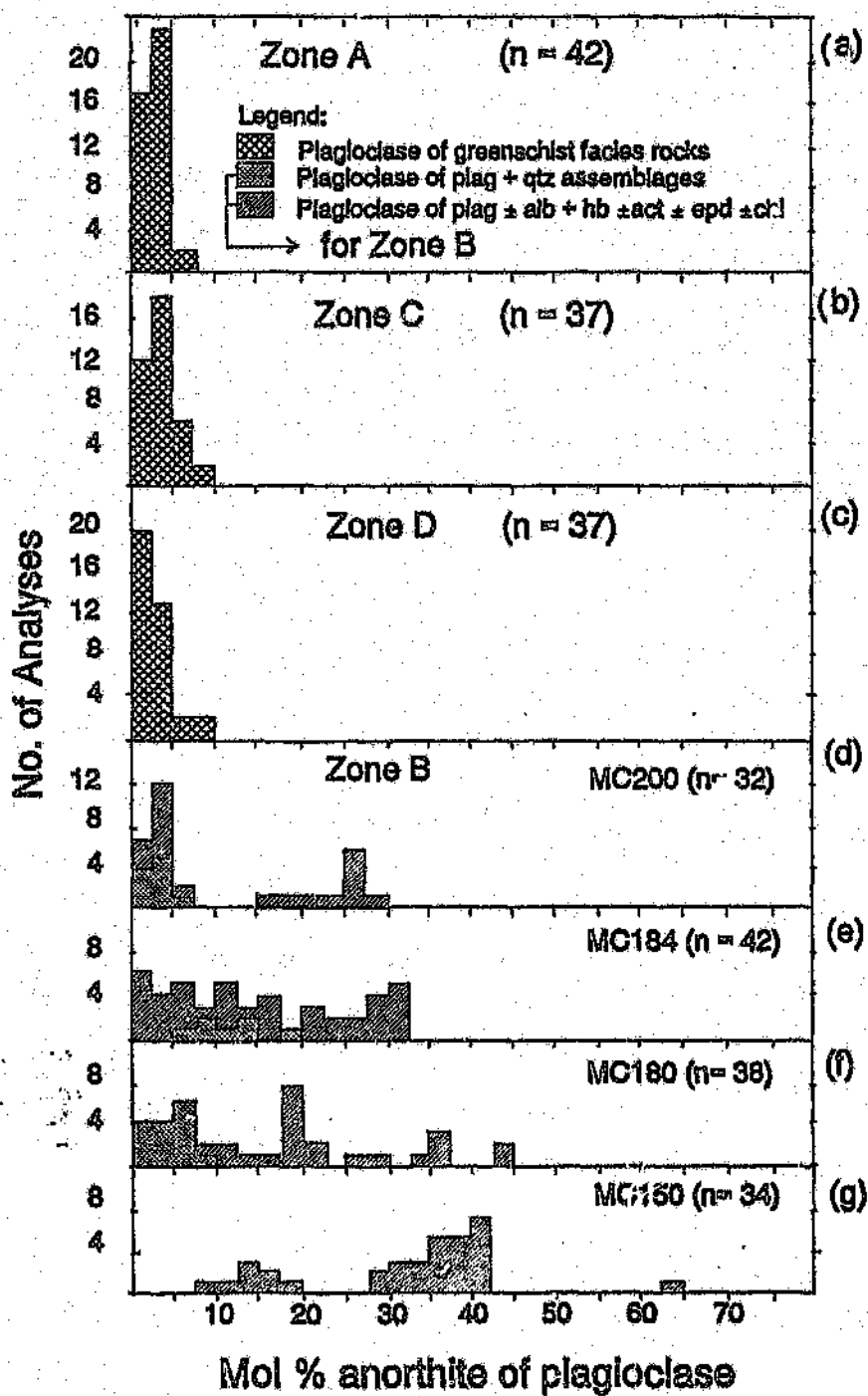


Fig. 4.11 Frequency histogram of plagioclase compositions from the metamorphic zones as indicated

A plagioclase grain with an unusually high anorthite content (i.e. rim = An_{63} and core = An_{35} , Fig. 4.11g and Appendix: PLAGIOCLASE, sample MC150G # 08p1 and -2) was found juxtaposed to tschermakite (Appendix: AMPHIBOLE, sample MC150G, 08a1 and -2, also marked sample x in Fig. 4.15) and chlorite (Appendix: CHLORITE, sample MC150G 08c2, $Cl_{in} = 60$). Tschermakite "coexisting" with plagioclase, on the other hand, was found to be rare as only two other cases were encountered. Figure 4.15a shows the bulk of amphibole coexisting with plagioclase, to be magnesiohornblende (i.e. ≥ 7.25 Si per formula unit). The anomalous compositions of both phases appear to be due to their exceptionally high Al_2O_3 -content whereas that of the adjacent chlorite is normal. An indication of the degree of equilibration of the Al-rich plagioclase-amphibole pair was obtained by applying the geothermometer of Spear (1980). It yielded a temperature of $730 \pm 25^\circ C$. Eight other pairs from the same pillow (MC150) yielded a narrow temperature range of between $500 - 530 \pm 20^\circ C$ (Fig. 4.16, solid triangles). The spurious temperature, in excess of $200^\circ C$ of the average obtained for the pillow, which is also in disagreement with the ambient mineral assemblages, suggest that the plagioclase and tschermakite grains did not attain equilibrium. Disequilibrium is also corroborated by the observations of Maruyama *et al.*, (1983), namely that labradorite is uncharacteristic of the greenschist-amphibolite facies transition.

The reason for the anomalous compositions may be the thermal decomposition of chlorite. It has been shown experimentally that chlorite decomposition, which is fO_2 dependent, commences at ca. $475^\circ C$ and is completed at ca. $575^\circ C$ (Liou *et al.*, 1974; Apter and Liou, 1983; Moody *et al.*, 1983). Sharp increases in the Al-content of amphibole and plagioclase coexisting with thermally decomposing chlorite have been noted by Moody *et al.* (1983). The proposed disequilibrium relationship between the plagioclase- and tschermakite pair may thus have resulted from a "flood" of chlorite-bound alumina inundating the juxtaposed magnesiohornblende and andesine (as indicated by their respective core compositions). The spurious equilibration temperature thus points to disequilibrium caused by the thermal breakdown of chlorite which, in itself, proves the geothermometer temperature of the other eight pairs from the same sample to be right. Only two other tschermakite-plagioclase pairs were found, suggesting similar situations. The relative scarcity of such pairs suggests that the

process of chlorite breakdown (in the presence of plagioclase) had just started for Zone B when the conditions of metamorphism started to recede, i.e. that maximum PT conditions of the rocks are within the thermal breakdown range of chlorite but below its maximum.

Plagioclase from the hydrothermally infilled pillow cavities and interstices, consisting only of albite + quartz, has also been analysed to see to which extent the compositionally "sterile"⁴ metadomains have become adjusted to metamorphism. The An-content of several plagioclase grains of the inert metadomains are compared to their counterparts of the more "fertile" (basic) metadomains (stippled vs hatched areas, respectively, Figs. 4.11d, 11e, 11f). In sample MC200, in which the peristerite solvus is present, plagioclase compositions from the "sterile" domains show an frequency distribution pattern ($An = 0 - \leq 7.5 \text{ mol\%}$) identical to those on the albite side of the gap from the "fertile" metadomains. No compositions were recorded plotting on the anorthite side of the gap. A similar scenario is found for MC180, whereas the sterile-domain plagioclase of MC184 only show consolute compositions similar to those from the fertile parts, suggesting that the $a_{Ca^{2+}}$ of intergranular fluids in some instances was sufficient to convert albite to plagioclase and thus modify bulk composition non-isochemically. This observation has an important bearing on the composition of the metamorphic waters of Zone B and will be discussed below in conjunction with the fluid inclusion data.

Zone C and D plagioclase is albite and in both zones the anorthite content ranges from 1 - 10 mol.% with a maximum abundance of 4.1 and 2.5 mol.% respectively. Metamorphic albites of the greenschist facies with compositions of $An = 8 - 10$ are unusual as they plot within the peristerite solvus. Two sources for the origin of these unusual compositions may be considered: (1) that they represent artifacts of analyses or, (2) that they have inherent geological significance. Based on the oxide totals and the stoichiometry of the analyses it is

⁴ "sterile metadomains" are here as consisting of assemblages which should not reflect any change with increasing metamorphic grade. Quartz and albite will be stable over a large range of PT conditions unless other components are metasomatically introduced. Bulk compositions, on the other hand, that do yield assemblages reflecting changes in metamorphic grade result from "fertile" metadomains.

doubtful that the albites were incorrectly analysed. The plagioclase isopleth map (Fig. 4.4) also shows that MC170 and MC66 have very restricted albite compositions and small standard deviations suggesting that good accuracy has been achieved. Therefore, the second alternative is implied. In this regard it is notable that most of the $An_{7.5-10}$ albites occur in two samples, one from each of the metamorphic Zones C and D, i.e. MC186 and MC140, respectively (cf Appendix: PLAGIOCLASE). From the plagioclase isopleth map (Fig. 4.4 and 4.1), it is evident that MC186 is situated close to the transition with the higher-grade Zone B. The close proximity to Zone B and evidence of retrograde tendencies of the other minerals in Zone C is consistent with the interpretation that they could also be of retrograde origin, i.e. incompletely retrograded oligoclase. The same would seem to apply for the dynamically overprinted pillow MC140. Here the greenschist facies retrograde overprint is not petrographically obvious, as for example the alteration fabric is for Zone C, but evident only from the other mineral compositions (i.e. chlorite, epidote and amphibole). The anomalous compositions of the Zone C and D albites corroborate the retromorphic tendencies of the other phases with which they coexist, and are thus seen to represent relicts of incompletely retrograded oligoclase or andesine.

4.3.4 Amphibole

Amphibole occurs in all metamorphic zones (A, B, C, and D) but the strongly carbonated pillows (MC20) of Zone D (see Appendix: AMPHIBOLE for list of analyses). The amphiboles of the metamorphic zones show a wide range in composition and belong to both the iron-magnesium and calcic amphibole groups. Cummingtonite of the first group is relatively scarce (≤ 1 vol.%), it coexists with other amphiboles and is found only in the Mg- and Fe-enriched rims and hyaloclastites of some pillows (i.e. MC150 and MC184). The analysed calcic amphiboles are plotted in Fig. 4.12 on the appropriate classification diagrams after Leake (1978). From the diagrams it is evident that the amphiboles range from actinolite through actinolitic hornblende, magnesio-hornblende, tschermakitic hornblende to tschermakite. The bulk of the amphiboles have a low (≤ 0.5) A-site [= $(Na + K)_A$] occupancy which may range from 0 - 1 for the group as a whole. A small proportion of amphiboles, on the other hand, have more than half of the A-site occupied, i.e. $(Na + K)_A$

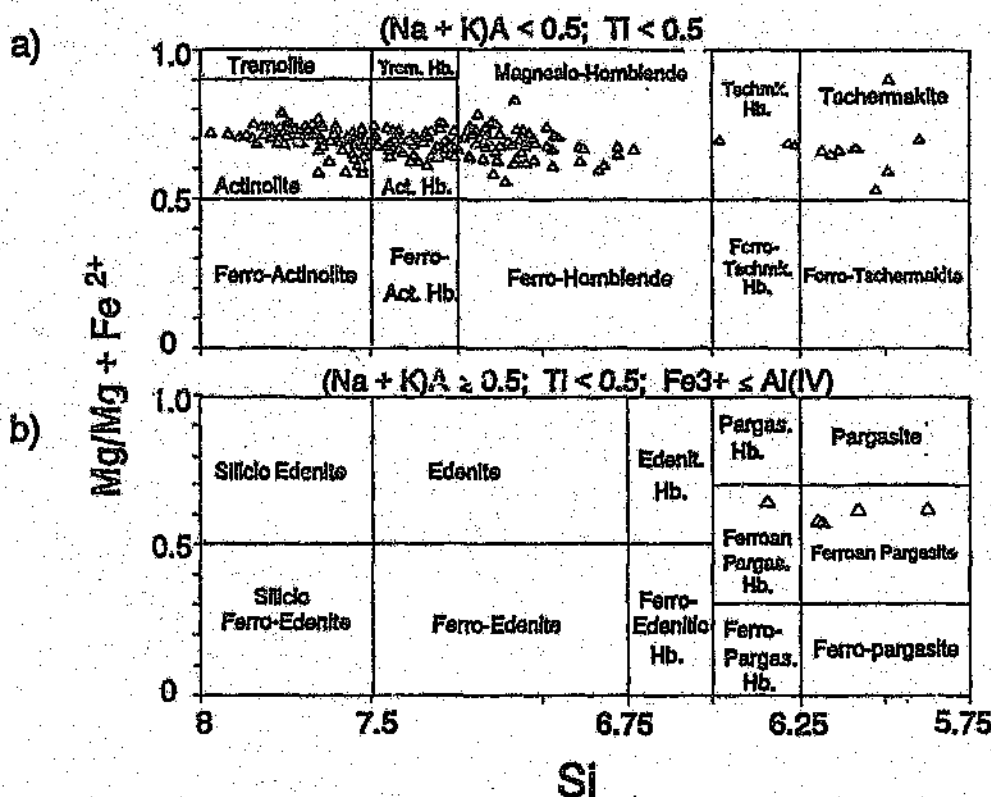


Fig. 4.12 Classification of calcic amphibole [in which $(Ca + Na)_B \geq 1.34$, $Na_B < 0.67$] of the study area according to Leake (1978).

≥ 0.5 , classing them as ferroan pargasitic hornblende and ferroan pargasite (Fig. 4.12b). From a three-dimensional perspective, Fig. 4.12a can be visualised as filling the space immediately above Fig. 4.12a with $(Na + K)_A$ representing the (z) axis perpendicular to the plain of the diagram. Such a stacked relationship would demonstrate the close chemical relationship between the tschermakites and ferroan pargasites and that there is a progressive rather than a discontinuous relationship between them.

Figs. 4.13i,ii,iii,iv shows the range of amphibole compositions encountered in Zones A, B, C and D, respectively. The isopleth map for amphibole (Fig. 4.5) delineates the average amphibole compositions for individual pillows.

Zone A is essentially the only metamorphic zone that contains only actinolite (Fig. 4.13i) coexisting with the other phases common to this zone.

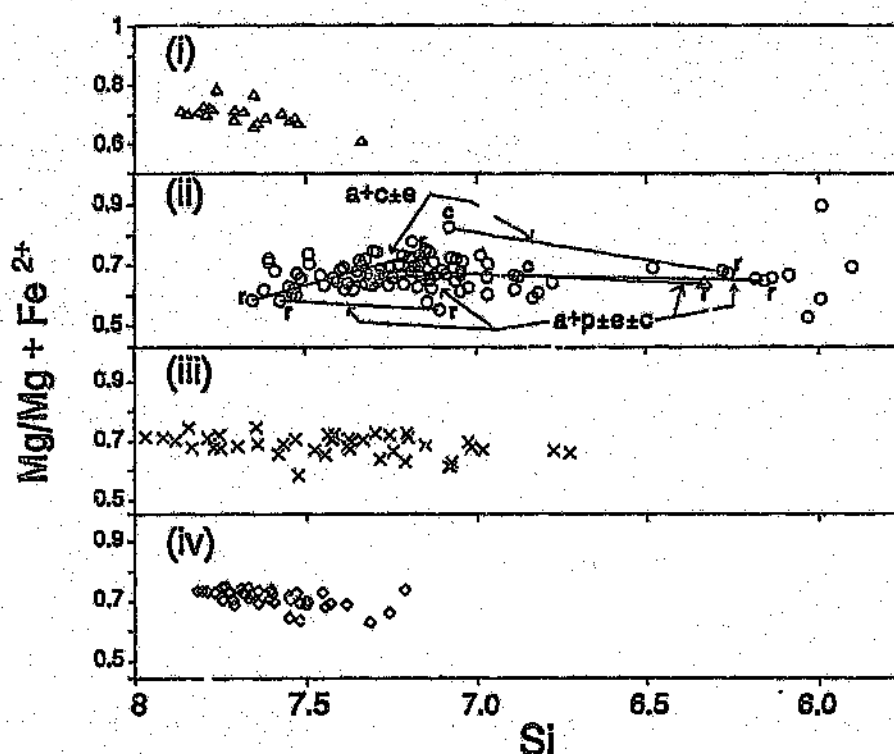


Fig. 4.13 Mg/Mg + Fe²⁺ vs Si plot for compositions of amphibole of the metamorphic zones: (i) Δ = Zone A; (ii) $\circ$ = Zone B; (iii) $\times$ = Zone C; (iv) $\diamond$ = Zone D. For Zone B several tie lines are indicated; r-r = rim compositions of coexisting amphiboles, c-r = core and rim compositions of individual grains. Arrows indicate phases coexisting with the tie-line mineral, a = amphibole, c = chlorite, e = epidote, p = plagioclase.

Zone B has amphibole with the widest compositional range and is portrayed by the amount of tetrahedrally (T-site) coordinated Si (Fig. 4.13ii). Si decreases at the T-sites with a concomitant increase of Al^{IV} to make up a total = 8. However, Al^{IV} is generally seen to increase with increasing metamorphic grade (i.e. decreasing Si). Only in a few instances was actinolite found to coexist with magnesio-hornblende (indicated by solid tie lines, Fig. 4.13ii). The relationship manifests the compositional break between actinolite and hornblende (e.g.

Maruyama *et al.*, 1983;). The amphibole plot of $Mg/(Mg + Fe_{2+})$ vs Si (Fig. 4.13ii), however, does not show a hiatus in the data to support the amphibole compositional gap (the hiatus position commonly ranges from about 7.6 to 7.2 Si). The absence of the compositional gap may be explained by the following:

- 1) Bulk compositions within the pillows vary from domain to domain with the result that the amphiboles are not all buffered by the same assemblages.
- 2) The exact positioning of the compositional gap is f_{O_2} dependent (Maruyama *et al.*, 1983).
- 3) The displayed analyses of Fig. 4.13ii represent both rim and core compositions of grains which have different relationships with respect to the gap. Many amphiboles, for example, have cores that have not bridged the gap whereas the rims have. The indiscriminate plotting of such data clearly would mask the compositional break (explanation to follow).

An example of the latter, where crest⁵ rim-compositions envelope internal zones with gap compositions, is shown in Fig. 4.14. It is assumed that the amphiboles grew by accreting crystalline material to their outer surfaces, which more or less shielded inner zonations. The observed relationships support the prior existence of the compositional gap. Figure 4.14 shows the Al^{IV} -content, from core to rim, of two faintly zoned amphiboles buffered by plagioclase, epidote and titanite coexisting within an amygdale. The innermost core composition of both grains, at core 1, are just above the maximum Al^{IV} -content for actinolite (i.e. between 4.48 and 4.88 wt% Al_2O_3 , cf also Maruyama *et al.*, 1983 and Ishizuka, 1985). At core 2, the grain of traverse 1 - 4 becomes poorer in Al^{IV} and in fact becomes a true actinolite whereas the grain marked by traverse 9 - 12 develops into a true magnesio-hornblende. One position closer toward the rim, at core 3, actinolite of traverse 1-4 is zoned by actinolitic hornblende whereas the coexisting grain of magnesio-hornblende is zone 1 by similar amphibole only marginally enriched with Al^{IV} , ultimately allowing the compositional hiatus to recede. At the grain rims the compositions converge to a composition close to one

⁵ The crest composition is attained when the amphibole gap is bridged, used by Maruyama *et al.*, 1983.

another containing roughly one tetrahedrally coordinated Al atom per unit cell (corresponding to between 8.60 and 9.04 wt% Al_2O_3). The compositions at which the amphibole pair started and ceased to exist (i.e. the range of the gap) are similar to those reported by Maruyama *et al.* (1983) and Ishizuka (1985). They report the gaps initial emergence at 3.7 and 4.82 wt% Al_2O_3 , respectively and its closure at ca. 2.0 and 7.0 wt%, respectively.

The degree of correspondence of the amphiboles shown in Fig. 4.14 and those of the mentioned workers strongly support the prior coexistence of actinolite and hornblende and that the compositional gap of the amphiboles was bridged only during the latter part of their growth history. On the other hand, the presence of amphibole pairs with rim compositions manifesting the gap (see rim 4 in Fig. 4.13ii), actually suggest that the critical curve for coexisting amphibole was initially overstepped with respect to P and T requirements of the reaction. It is thus suggested that PT_{max} of Zone B is also constrained by this reaction line in PT-space.

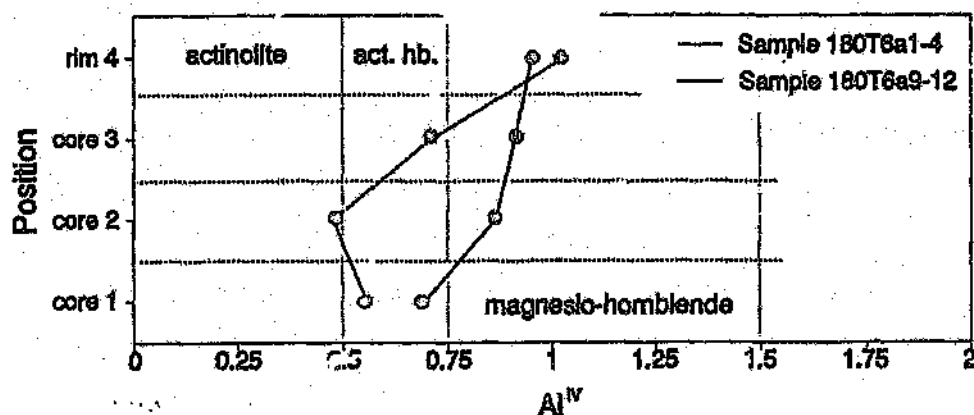


Fig. 4.14 Compositional variation of two coexisting amphibole grains with essentially the same rim composition. The divergence of core compositions however, suggest the earlier existence of the amphibole gap.

A second more obvious "compositional gap" exists at higher Al^{IV} -values, i.e. between 6.75 - 6.5 Si (Fig. 4.12a and 4.13ii). This hiatus, however, is not to be confused with the amphibole gap commonly referenced in the literature and the one discussed above. The one observed

here appears to be related to the unique conditions of metamorphism and bulk rock composition. The amphiboles are distinctive by their low T-site Si-content, or conversely, high Al^{IV} -content. Accordingly, they can be classified as tschermakites and ferroan pargasites. The tschermakites usually occur as overgrowths around magnesiohornblende cores (dashed c-r tie lines, Fig. 4.13ii) or as the cores of grains which are rimmed by ferroan pargasite. Their most common association is with chlorite $\pm$ clinozoisite. These assemblages are most frequently found at the Mg and Fe-rich margins and hyaloclastites of pillows, initially enriched in these elements by seafloor metasomatism. Tschermakite and pargasite, however, are not found in the presence of plagioclase, excepting for the anomaly mentioned earlier (e.g. sample x, Fig. 4.15). In the hyaloclastite of MC180, metadomains of plagioclase coexisting with magnesiohornblende are found enclosed by tschermakite with the latter only in contact with the mentioned amphibole, chlorite and clinozoisite. By means of elimination it appears that chlorite is the only phase that is always present with tschermakite and ferroan pargasite. As suggested earlier, it is notable that these very Al-rich amphiboles always occur with chlorite, implying a possible genetic link, namely the thermal breakdown of chlorite.

Petrographic examination showed that the common assemblage (as defined earlier) was found less often in the higher-grade Zone B rocks than in any of the other metamorphic zones despite a similar range of bulk compositions ($\sim$ komatiitic basalt). This phenomena hinders the previously defined basis (i.e. the common assemblage) for the comparison of amphibole compositions of the higher- and lower-grade areas and those of other studies. In order to find a next-best basis for such comparisons, amphibole of variously buffered assemblages has been tested. In Figure 4.15, amphibole buffered by plagioclase-bearing assemblages but deficient with respect to the common assemblage, non-plagioclase bearing assemblages and those buffered by the common assemblage are plotted for comparison. The dotted reference line of each of the figures (a,b,c) has the same position and slope relative to the X- and Y-axes. In Figure 4.15a the reference line bisects the amphibole population coexisting with $plag \pm epd \pm chl \pm qtz$. In Figure 4.15b amphibole coexisting with plagioclase-absent assemblages has a marginal but distinct offset to higher magnesium

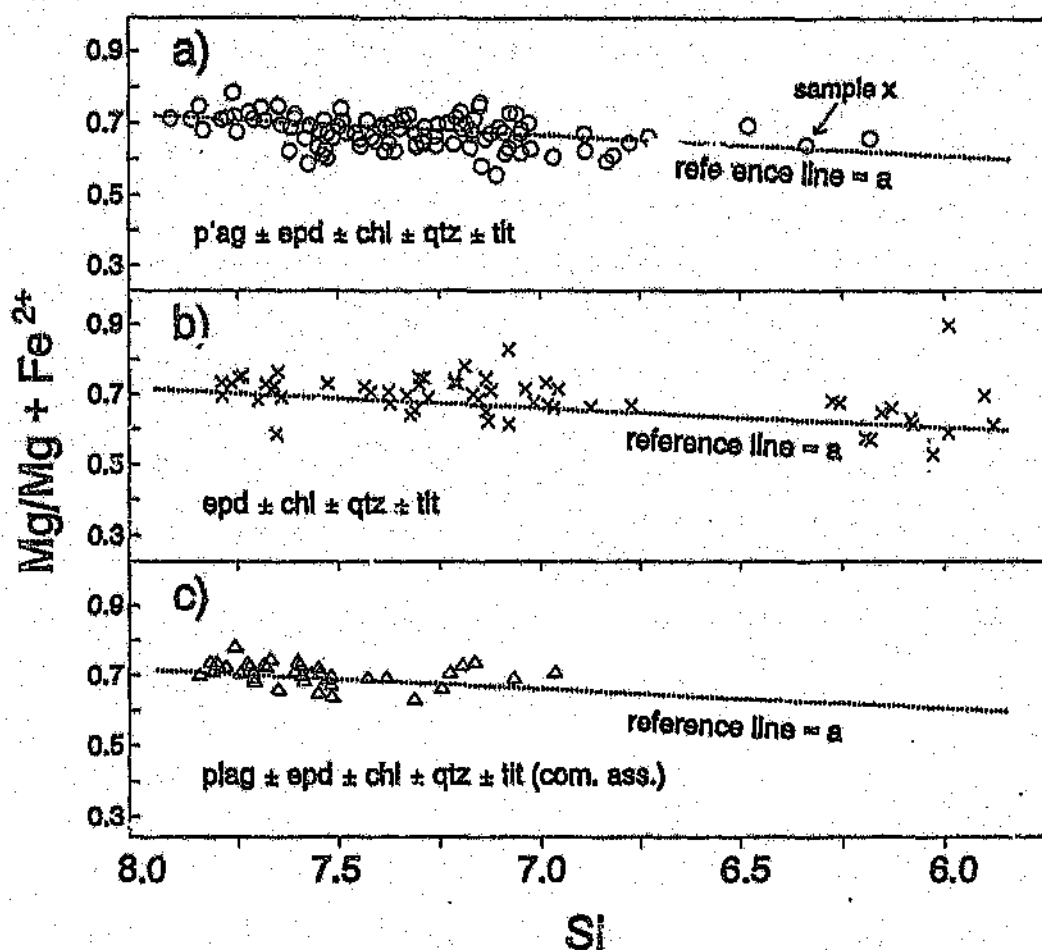


Fig. 4.15 Compositional variation of amphibole in terms of $Mg/Mg + Fe^{2+}$ vs Si buffered by the indicated mineral assemblages: a) plagioclase-present assemblages = $\circ$, b) plagioclase-absent assemblages = $\times$, and c) the common assemblage = Δ . Reference line = a is similar for all plots and intended to aid comparison (see text for discussion).

values. In Figure 4.15c the reference line passes approximately along the long-axis of the common assemblage-buffered amphibole population. The good correspondence between the amphibole population buffered by the (deficient) plagioclase-bearing assemblages and those of the common assemblage, relative to reference line "a", show that sufficient grounds exist for general amphibole comparisons. Figure 4.15c gives a perspective on the relative scarcity of the common assemblage in the higher-grade rocks (i.e. note the small number of data points with ≤ 7.5 Si pfu compared to those in Figs. 4.15a,b).

The Zone C amphiboles have uncommon Si and $\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ compositions which may be attributed to a retrograde origin. The amphiboles have a wide range in Si-content which comprises some very Si-rich actinolites and Si-depleted magnesiohornblendes. The range is unusual as the actinolites suggest formation at relatively low metamorphic grades whereas the magnesiohornblendes suggest formation at higher grades. Actinolites of prograde trends that coexist with hornblendes usually have around 7.5 Si pfu and not close to 8, as is the case here, suggesting that they do not represent an equilibrium range. The amphiboles also have a very narrow $\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ distribution range which results in a near-horizontal trend. The flat slope contrasts with those typical of the prograde amphiboles, for example, the trends of Fig. 4.12a or Fig. 4.15a which decrease in $\text{Mg}/\text{Mg} + \text{Fe}^{2+}$ with decreasing Si (cf also for prograde trends of Hynes, 1982, Fig. 2). The difference in mineral chemistry of these amphiboles compared to those showing prograde trends is thus attributed to the retrograde metamorphism of Zone C which is also petrographically evident.

Zone D amphiboles display a range of compositions between actinolite and actinolitic hornblende (Fig. 4.13iv). No coexisting amphibole pairs or grains with zonations suggesting the prior existence of the amphibole hiatus were found. The actinolitic hornblendes, however, plot in the field where the compositional gap normally occurs for rocks with a similar bulk FeO/MgO composition. For example, pillow MC120 has a bulk $\text{FeO}/\text{MgO} = 1.18$ and the amphiboles analysed from it plot within the compositional gap of amphiboles from a metabasite with similar bulk FeO/MgO value demonstrated by Maruyama *et al.*, (1983, Fig. 5). It is also interesting that ca. 60% of the 14 core-rim pairs analysed have rim compositions containing less Al^{IV} (i.e. more Si) than the corresponding cores, implying a retrograde tendency. The amphibole rims, however, are typical of the greenschist facies. The unusual gap compositions can, therefore, be explained by invoking dynamic metamorphism accompanied by greenschist grade temperatures, and that the precursor rocks consisted of higher grade amphibole, probably similar to those of Zone B. It is envisaged that partial chemical re-equilibration and synchronous flattening of the rocks at greenschist grade conditions could have effected a compositional shift of the Al-richer precursor amphiboles into the observed compositional-gap region. Such an origin is also consistent

with the retrograde trends shown by the other phases discussed above.

4.4 Geothermometry and Geobarometry

4.4.1 Geothermometry

The empirical hornblende-plagioclase geothermometer of Spear (1980) is applied to the komatiitic metabasalts of Zones A, B and D (Fig. 4.16). The Zone A samples (open symbols) plot in the temperature range 320 - 413°C. The stratigraphically highest sample (open square symbol, from the Hooggenoeg Formation) has the lowest temperature of $320 \pm 25^\circ\text{C}$ and the others which are on the same stratigraphic height in the underlying Komati Formation range between 350 - 413°C.

The Zone B samples (filled symbols) yield temperatures between 490 - 530°C. There are, however, two outliers, one just above 725°C and the other at $\pm 430^\circ\text{C}$. The high temperature one has been discussed previously where it was concluded that equilibrium of the plagioclase-hornblende pair had been disrupted by the breakdown of chlorite. The other outlier, was obtained from an amygdale plagioclase-amphibole pair, 2 cm away from a similar amygdale pair which yielded 530°C (Fig. 4.16, other filled circle). The low temperature may be ascribed to kinetic inertia and shows that a small proportion of assemblages have not attained equilibrium. The majority (86%) of the Zone B amphibole-plagioclase pairs, nevertheless, produce a cluster of temperatures between 490 - 530°C. This temperature interval may be accepted as the thermal maximum for the Zone B rocks situated at the fault-bound stratigraphic base of the Komati Formation.

The Zone D rocks display the widest range of temperatures, i.e. from 280 - 415°C (Fig. 4.16, cross symbols). It is interesting to note that the amphibole-plagioclase pairs of the cores of both pillows are thermally blocked at higher temperatures than those of the corresponding pillow rims (indicated by dotted crosses). This is particularly evident for the two pillow rim temperatures of MC140 recorded at $280 \pm 25^\circ\text{C}$. Such a decreasing trend for the pillows seems to be consistent with the petrographically established retrograde tendencies mentioned earlier if the interpillow regions are viewed as preferentially hosting late-fluids

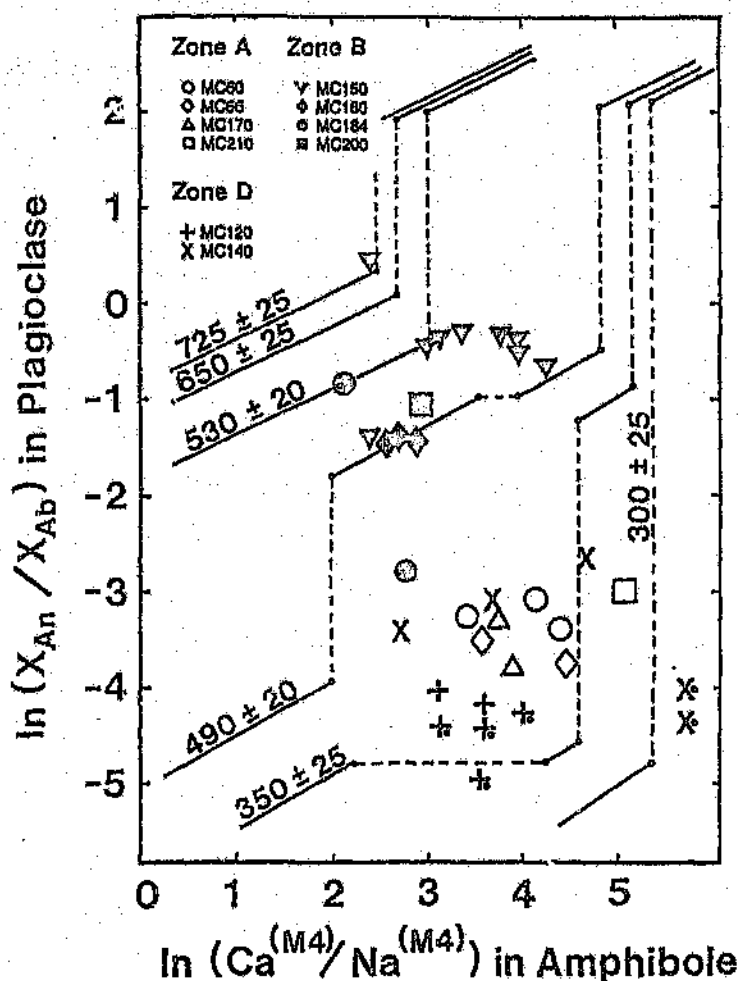


Fig. 4.16 $\ln(X_{An}/X_{Ab})$ in plagioclase vs $\ln(Ca_{M4}/Na_{M4})$ in amphibole, an empirical geothermometer proposed by Spear (1980), based on $NaSi-Ca_1Al_1$ exchange between plagioclase and hornblende. The isotherm contours consist of vertical and horizontal dashed lines representing plagioclase and amphibole immiscibility, respectively, whereas the solid portions reveal the (non-ideal) partitioning of the mentioned exchange vector between plagioclase and amphibole. Amphibole recalculation procedure is similar to the method used by Spear (1980). Amphibole origin with respect to the various metamorphic zones is indicated on the graph legend. Dotted crosses represent plagioclase-amphibole pairs from the rims of pillows whereas the others are from the cores of pillows.

relative to the more massive pillow cores.

4.4.2 Pressure Plots

A plot of total Ti vs total Al shows that the amphiboles of all the metamorphic zones have relatively low Ti-contents although they are buffered by titanite. Hynes (1982) shows that similarly buffered amphiboles from low- and medium pressure environments plot above and below the reference line in Fig. 4.17a, respectively. Medium pressures of about 2 - 5 kb (cf Winkler 1979) are thus semi-quantitatively indicated for the study-area amphiboles.

A medium pressure origin is also supported by the pressure estimates of Raase (Fig. 4.17b) and Laird and Albee (Fig. 4.17c). In the Raase pressure diagram, based on the Al^{VI} vs Si values of amphiboles (i.e. hornblendes only) the data plot on a trend of which the upper portions straddle the ± 5 kb line. The diagram only discriminates between hornblendes of high and low-pressure. It is apparent, nevertheless, that the amphiboles plot on the high pressure part of the low-pressure field, corroborating the medium pressure indicated above. Similarly, the $Al^{VI} + Fe^{3+} + Ti$ vs Al^{IV} diagram of Laird and Albee (1981) shows that the study area amphiboles plot in a field which coincides with the medium pressure amphiboles from worldwide occurrences of mafic schist. The fields of the medium pressure amphiboles are contrasted by those from the high-pressure Sanbagawa & Franciscan terranes and the low-pressure Abukuma terrane (Laird and Albee, 1981, Fig. 8B and 12B).

4.4.3 Geothermobarometry

Application of the empirically-based (Na-Ca) amphibole-albite-chlorite-epidote-quartz geothermobarometer of Triboulet (1992) has been attempted. The application of the geothermobarometer is limited to Ca-free plagioclase ($An \leq 10$) assemblages and allows for the construction of complete P-T paths in the low- to high P and low- to medium T fields (i.e. ca. 1 - 10 kb and 300 - 650°C). The results of the study-area amphiboles are briefly discussed below.

Due to the feldspar compositional requirements of the geothermobarometer, only the lower-

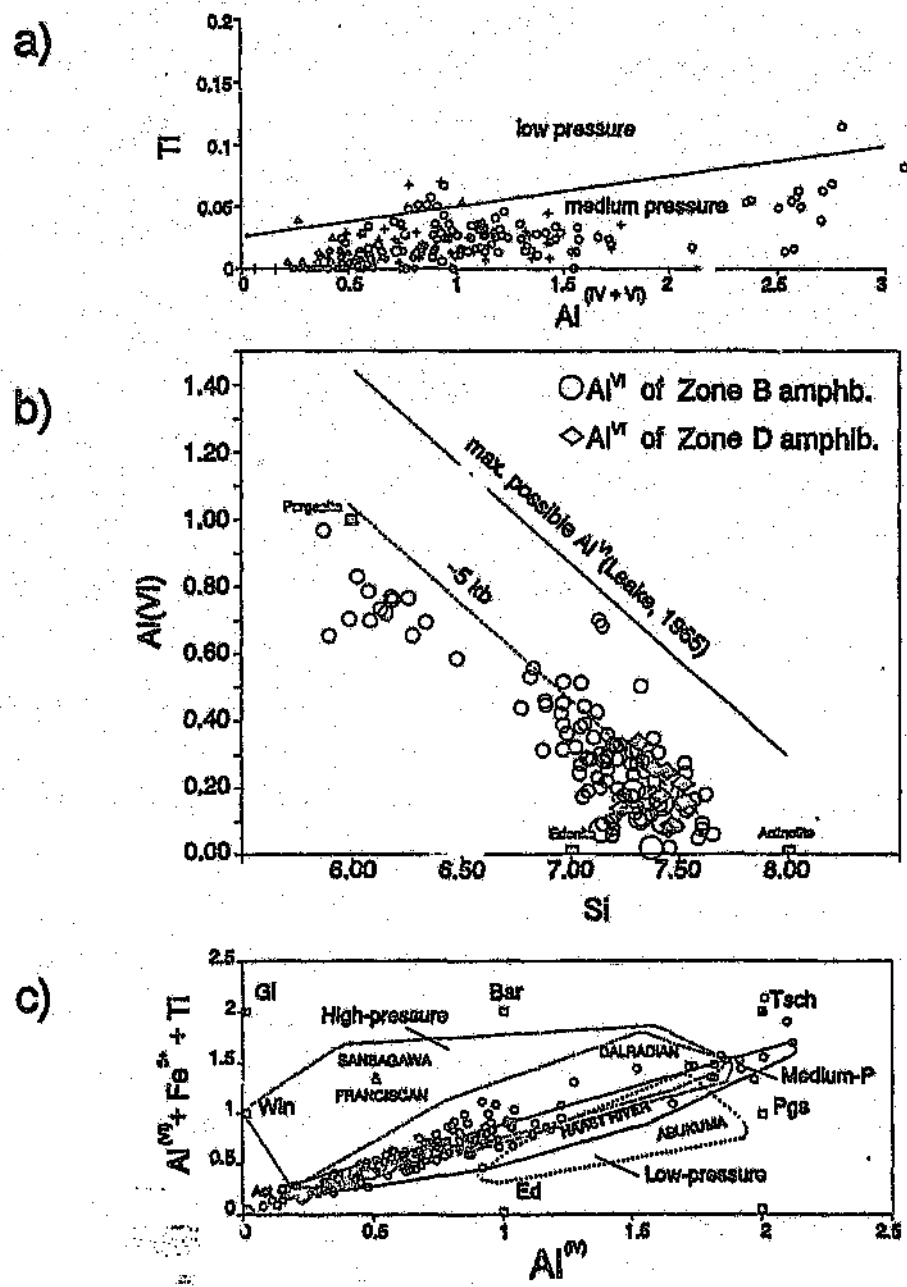


Fig. 4.17

Amphibole pressure plots: (a) total Ti versus total Al after Hynes, (1982); (b) formula proportion Al^{VI} vs Si of hornblendes after Raase, (1974), and (c) formula proportion $Al^{VI} + Fe^{2+} + Ti$ vs Al^{IV} of calcic amphibole after Laird and Albee, (1981). Symbols of (a) and (b) are the same as for Fig. 4.13. Note that the Zone A amphiboles (actinolites) are excluded from plot (b) as it is only intended for true hornblendes ($Si \leq 7.5$), c/ Fig. 4.12 and 4.13i. Amphibole abbreviations: Gl = glaucofane; Bar = barroisite; Tsch = tschermakite; Win = winchite; Pgs = pargasite; Act = actinolite; Ed = edenite.

grade Zone A and D amphiboles qualify for application. Of these, the retrograded dynamic-style samples show the most coherent range of PT-conditions, varying between 2 - 5 kb and 310 - 490°C. Most amphibole pairs show that rim compositions have formed at lower P and T conditions than their respective cores, confirming the mineral chemistry trends. There are some pairs, however, that show the opposite tendency. The core-rim pair vectors are not sufficiently restricted to confidently construct an unambiguous path in PT-space. The other static-style samples of MC60 and MC66 plot in a somewhat larger pressure range but with a narrower temperature range, i.e. 2 - 8 kb and 300 - 400°C. The somewhat unrealistically large pressure range obtained for the samples may be due to the lack of quartz in some of the analysed assemblages. The presence of oligoclase and andesine ($An = 10 - 50$) in Zone B assemblages automatically exclude these amphiboles from this application. It is interesting, nevertheless, that the application of the geothermobarometer yielded a range of temperatures comparable to the temperatures obtained from the Spear geothermometer (Fig. 4.16, i.e. 400 - 550°C). Pressures of 2 to 8 kb were recorded of which sequentially analysed grains (from cores to rims) produced in some cases progressive PT trends whereas others showed the opposite. Triboulet (1992), however, clearly states that the use of amphiboles from assemblages containing plagioclase with $An \geq 10$, will result in excessive pressures. The temperatures obtained by means of the geothermobarometer correspond surprisingly well with those in Fig. 4.16 whereas the pressures, as expected, are too high. However, the opposing PT-trends for the sequentially analysed grains was not expected. These tendencies are suspected to be related to the cryptical presence of the amphibole gap, as discussed earlier (cf Fig. 4.14), and probably indicate that the geothermobarometer is not suited for coexisting amphiboles, an aspect not mentioned by Triboulet (1992).

The calcic amphibole geothermometer of Blundy and Holland (1990), which is best suited for rocks that have equilibrated between 500 - 1100°C was also tested. Although the authors claim it can be satisfactorily used in the greenschist to lower amphibolite range, they also admit that it is susceptible to larger errors below 600°C, possibly due to ordering of plagioclase and amphibole solid solutions. For this study, nevertheless, it was found to be unsatisfactory as a P input is also required. Despite constant pressure inputs, relatively large

temperature variations for individual samples were obtained.

Due to a scarcity of quantitative geobarometers for greenschist- and lower-amphibolite facies metabasites, constraints on pressure of metamorphism have been sought via fluid inclusion microthermometry. The aim of the fluid inclusion study was essentially to establish the sequence of entrapment and the relevant chemical evolution of the metamorphic fluids in response to metamorphic grade and style. Pressures are constrained mainly by means of isochore and geothermometer-temperature intersections. Microthermometry for all metamorphic zones, except Zone C, were done. Zone C is excluded due to the alteration overprint of the metamorphic assemblages which invalidates geothermometer applications and thereby the necessity to obtain isochores for the zone. Pressure and temperature estimates, however, are not considered crucial for Zone C as it is transitional between Zones B and D.

4.5 Fluid Inclusions

4.5.1 Nature and Occurrence

Fluid inclusions were examined in 14 doubly-polished wafers from 6 sample localities. The samples, two from each of the metamorphic zones A, B, and D, cover a vertical, essentially litho-stratigraphic sequence of about 6 km in the statically metamorphosed terrain and along strike the superposed effects of dynamic metamorphism (Fig. 4.1). The fluid inclusions examined are hosted in quartz related to the interpillow- and drainage cavities of the metapillows (Fig. 2.4B, 2.4C). Only two inclusion types have been identified. They are, in decreasing order of abundance, aqueous inclusions and $\text{H}_2\text{O}-\text{CO}_2$ inclusions.

Only aqueous inclusions were found in the statically metamorphosed rocks of Zone A and B. In Zone D, the least deformed sample also contained only aqueous inclusions, whereas the most deformed pillow contained both aqueous and $\text{H}_2\text{O}-\text{CO}_2$ inclusions. The $\text{H}_2\text{O}-\text{CO}_2$ inclusions are not abundant and comprise $\leq 1\%$ of the inclusions in the sample. A petrographic scan of a quartz vein sample, taken about 5m to the southeast of the pillow outcrop, showed the presence of a much larger proportion of $\text{H}_2\text{O}-\text{CO}_2$ inclusions.

Nevertheless, all fluid inclusions are small with sizes ranging from 1-30 μm and averaging about 5 to 10 μm . Most inclusions have irregular to rounded shapes and only seldom negative crystal habits.

4.5.2 Fluid inclusion petrography

Fluid inclusions occur in two different ways, namely in large ($\geq 200 \times 200 \times 200 \mu\text{m}$), high population-density, three-dimensional clusters or clouds and as trails along healed fractures. (The term cloud is preferred to cluster or group as these generally refer to smaller three-dimensional inclusion assemblies). The high population-density fluid inclusion clouds are subdivided into three categories, of decreasing certainty of primary origin (*cf* also Roedder 1967, 1981; Hollister *et al.*, 1981). For ease of description the high population-density cloud inclusions have been termed growth zone inclusions, orientated cloud-inclusions and random cloud-inclusions and are defined as follows:

- a) Growth zone inclusions occur in inclusion clouds that accentuate the growth zones of the quartz host and the individual inclusions may or may not be crystallographically orientated (Figs. 4.18A, 4.18B).
- b) Orientated cloud-inclusions occur in inclusion clouds that are contained within individual grains in which population densities do not accentuate growth zones but of which the longest dimensions of individual inclusions follow crystallographic orientations (Figs. 4.19A, 4.19B).
- c) Random cloud-inclusions occur in inclusion clouds filling whole grains or portions of grains with no growth zone accentuation and of which individual inclusions have no specific orientation (i.e. similar to those of Fig. 4.19B but without the preferred orientation, also comprising the most abundant category).

The inclusions that occur as trails along healed fractures are often present and qualify either as secondary or pseudosecondary according to whether they crosscut grain boundaries or not (Roedder, 1981). However, distinction of trail inclusions within inclusion clouds is not always possible as they are usually obscured by the high background population of the latter, translocation of the planar-bound inclusions and the propensity for filling ratio's of both

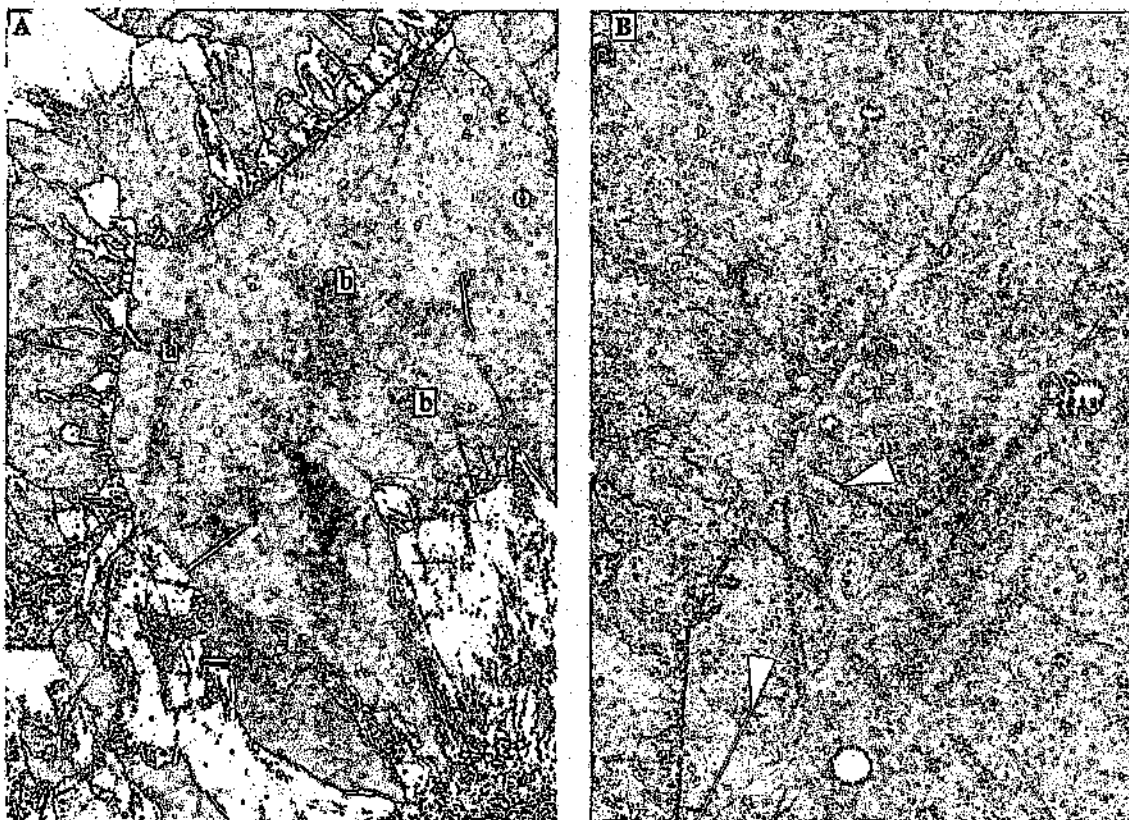


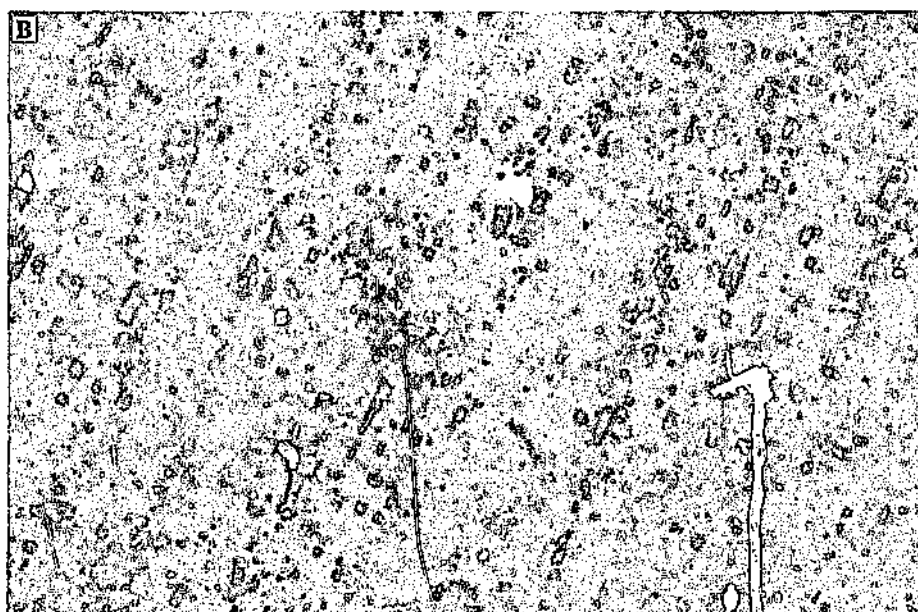
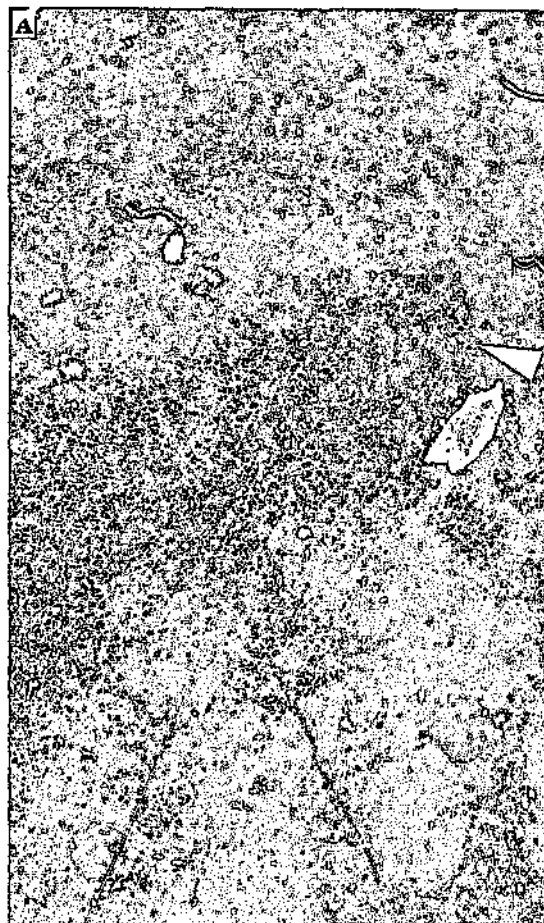
Fig. 4.18

- A** Photomicrograph of columnar quartz in partially-polarised light showing successive growth stages. The sample serves to demonstrate the general petrographic criteria used for identifying primary (cloud) inclusions. The dark patches and zones within the two central grains comprise high densities of fluid inclusions, some of which are zoned. Due to the inclusions occurring in three dimensional clouds which are growth zone related [i.e. crystallographically orientated and parallel to idiomorphic grain boundaries = (a), and of irregular shape that occasionally merge with the growth zone type and which do not transgress grain boundaries = (b)], suggest they are of primary origin. Sequences of fluid entrapment are clearly implied by the successive growth zones and by the succession of quartz formation. Inclusion trails, which occur subperpendicular to the growth zone clouds and stop at boundaries of the host grains, are also of primary origin. For more detail of the pseudosecondary inclusions see Fig. 4.18B. Height of image = 10.5 mm.
- B** An enlargement of the top part of the columnar quartz grain (left of centre) in Fig. 4.18A. The larger scale of the micrograph shows that ~9 inclusion trails, some of which are shown by arrows, terminate at the same grain boundary implying pseudosecondary status. It is in fact difficult to find true secondary inclusions in the entire doubly polished wafer. In the top left hand corner, overlaying a large black cement bubble, a smallish irregular shaped inclusion cloud is evident of which the individual inclusions can faintly be seen to be orientated. Fig. 4.19A and Fig 4.19B offer better detail of this type of cloud inclusion, i.e. orientated cloud inclusions. Height of image = 3.25 mm.

Fig. 4.19

- A Photomicrograph of orientated cloud inclusions in quartz. It is noticeable that the inclusions of the central grain occur as a large grouping of aligned inclusions stretching over a substantial part of the host quartz and that the fabric of the cloud is terminated at the grain boundary (arrow). The inclusions of the grain to the right of the triple junction also appear to be orientated, in this instance, sub-perpendicular to those of the central grain. Height of image = 0.75 mm.

- B Enlargement of a part of the central grain in Fig. 4.19A. It is evident that the inclusions are aligned and that the long axes of the fluid inclusions are orientated. Width of image = 0.15mm.



inclusion types to be similar. The latter feature suggests that fluid densities of the primary cloud-inclusions are not significantly different from those of the secondary inclusions.

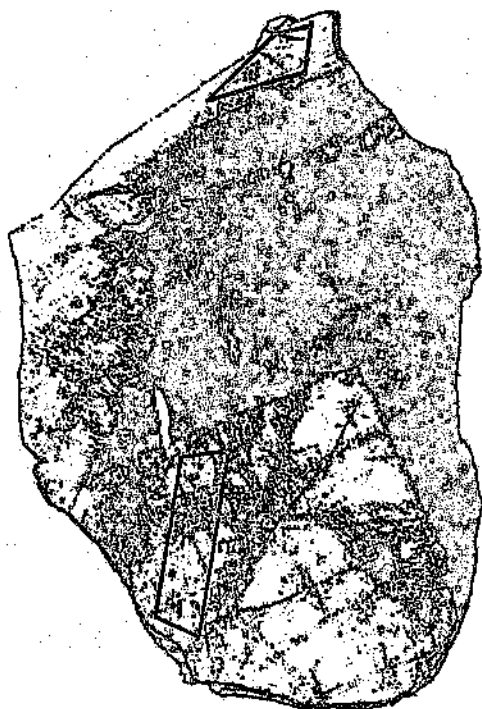
4.5.2.1 Fluid inclusions of Zone A metabasites. These fluid inclusions are uncomplicated and texturally least disturbed by secondary fracturing and annealing and the inclusions accordingly serve as a basis for interpretation of the more complicated inclusions of the other metamorphic zones. Both samples, MC210 and MC60, comprise milky quartz from interpillow-cavity infills. Both samples selected for analysis show growth zone inclusions. Healed inclusion trails are noticeable and are mostly contained within grain boundaries. Microthermometric traverses across the growth zones, from grain cores to -rims have been made and are shown in Figs. 4.20A, 4.20B. Inner and outer growth zones are seen to have trapped relatively early and late metamorphic fluids, respectively.

4.5.2.2 Fluid inclusions of Zone B metabasites. The fluid inclusions from the stratigraphically-lower Zone B rocks originate from a quartz-filled drainage cavity (MC150, similar to Fig. 2.4B) and an infilled interpillow cavity (MC180, similar to Fig. 2.4C).

Both samples comprise coarse-grained columnar quartz (Figs. 4.21A, 4.21B). The drainage cavity quartz columns, however, emanate from fine-grained granular quartz (Fig. 4.21A). The fine-grained quartz, texturally represents the initial encrustation of the cavity walls upon which the columnar quartz nucleated and grew during infilling. The grains of the quartz lining are clear and contain few inclusions except for some obvious late trails. Away from the trails the isolated inclusions have bubbles that make up 10 - 15% of the inclusion volumes (assessed with the aid of Hollister *et al.*, 1981, Fig. 12.1).

The inclusions of the columnar grains of both samples are quite similar. Two types are present. They are: (1) random cloud-inclusions which preferentially populate the upper halves of the quartz columns. Oriented cloud-inclusions occurred in isolated patches of MC180; and (2) secondary inclusions occurring in healed microfractures which crosscut several grains at a time. Fracture-bound inclusions occur throughout the columnar grains but

A



B

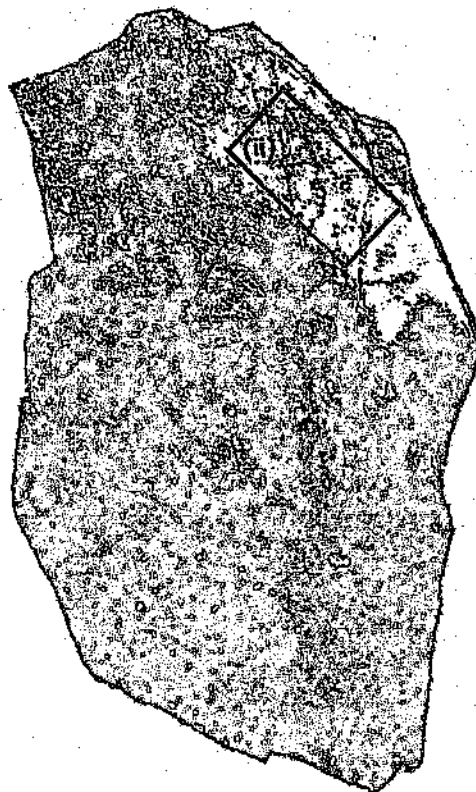


Fig. 4.20

- A** Photomicrograph of a doubly polished quartz grain MC210 exhibiting alternating crystallographical zones of relative clarity and extreme turbidity. Turbidity is the result of high fluid inclusion density. As the fluid inclusions are growth-zone related they are considered to be primary. The trail inclusions are pseudosecondary as they don't transect the grain boundaries (of which a part coincides with the upper confines of the chip). The analysed portions form a traverse from core to rim and occur in the marked area on the left zoned portion of the grain, i.e. (i) clear central core, (ii) dark outer growth zone, and (iii) clear peripheral growth zone. Vertical dimension of chip = 7.5 mm.
- B** Photomicrograph of a doubly polished quartz chip MC60 exhibiting a clouded central- and a clearer upper zone. The contact of the clouded inner- and clearer outer zones is marked by a silhouette of rhombohedral terminations characteristic of quartz. The fluid inclusions which cause the turbidity of the quartz are thus growth-zone related and therefore primary. The sequence of fluid trapping is clearly implicated; initially the core inclusions followed by those of the outer zone. The differences in the fluid properties of the marked (i) clouded inner- and (ii) outer core zones are shown in Fig. 4.25a. Average width of sample = 4 mm.

are most distinct at their basal portions because at the upper parts they tend to be obscured by the high population densities of the cloud inclusions (Fig. 4.21B). The secondary inclusions are usually slightly transposed and rarely form distinct planar arrays.

The filling ratios of the secondary inclusions are indistinguishable from those of the cloud inclusions and both have bubbles making up between 3 - 6% of the inclusion volumes. These aspects of the inclusions, i.e. the tendency of the secondaries to be transposed and have similar filling ratios to the primary-cloud inclusions, makes distinction between them difficult. As a result the measured population of cloud inclusions may contain both primary and secondary inclusions. The population marked as secondary inclusions for microthermometry, however, does not contain primaries as they have been sampled along the clear basal portions of the grains where primary cloud inclusions do not commonly occur (Fig. 4.21B).

The frequency of healed-fracture plains spanning grain boundaries is distinctly more noticeable for Zone B than for Zone A. Both the Zone A and B samples have sequentially entrapped primary fluids during quartz growth but subsequent to this only the Zone B samples entrained fluids in healed secondary microfractures. This petrographic phenomenon is considered to be the main difference between the Zone B and A samples and it is seen to have an important bearing on the evolution of the metamorphic fluids. (It will be demonstrated below that the secondary micro-fractures are intimately related to the higher metamorphic grade of Zone B, and therefore a primary feature of progressive metamorphism.

4.5.2.3 Fluid inclusions of Zone D metabasites. Sample MC120 exhibits growth zone inclusions and trails of which few crosscut grain boundaries (Fig. 4.22A). An interesting aspect of the grain is the absence of undulatory extinction which characterises most of the other quartz grains elsewhere in the interpillow area. The studied grain, however, is twinned along the c-axis direction as indicated by stippled lines (see Fig. 4.22A). Of the numerous twinning laws reported for natural quartz, the most common types are the interpenetrant,

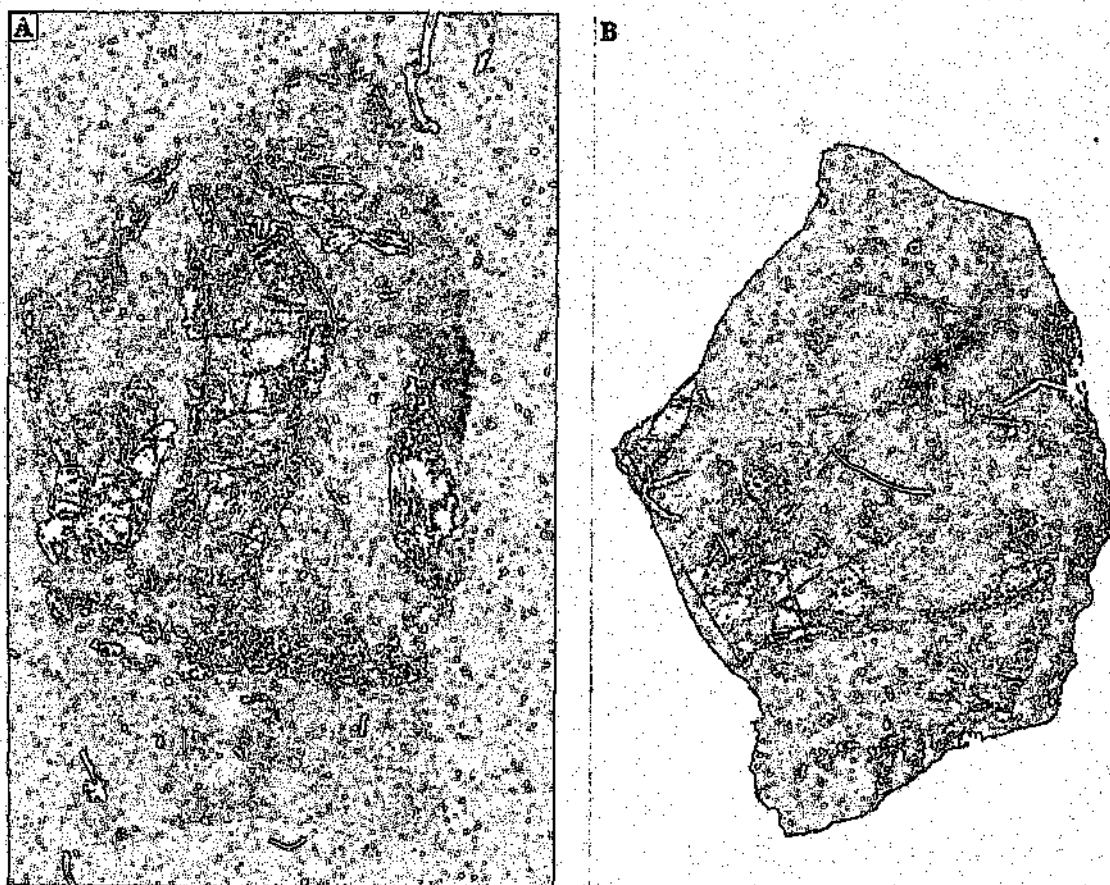


Fig. 4.21

- A** Photomicrograph of a doubly polished wafer MC150 in partially-polarised light showing a base of opaque metalava (drainage cavity wallrock) encrusted with a thin layer of fine-grained quartz supporting large subperpendicular grains of columnar quartz. The columnar grains bear both cloud inclusions and numerous arrays of crosscutting secondary inclusions. Vertical dimension of chip = 7 mm.
- B** Photomicrograph of a doubly polished wafer MC180 in partially polarised light (to enhance grain-shapes, i.e. columns radiate approximately from the lower left margin of the photograph). Cloud inclusions are evident as a distinct even- or patchy graininess of the quartz columns. Secondary inclusion trails clearly crosscut the grain bases of which some tend to merge or become less distinctive toward the cloud-inclusion rich upper parts (right of centre). Vertical dimension of chip = 6 mm.

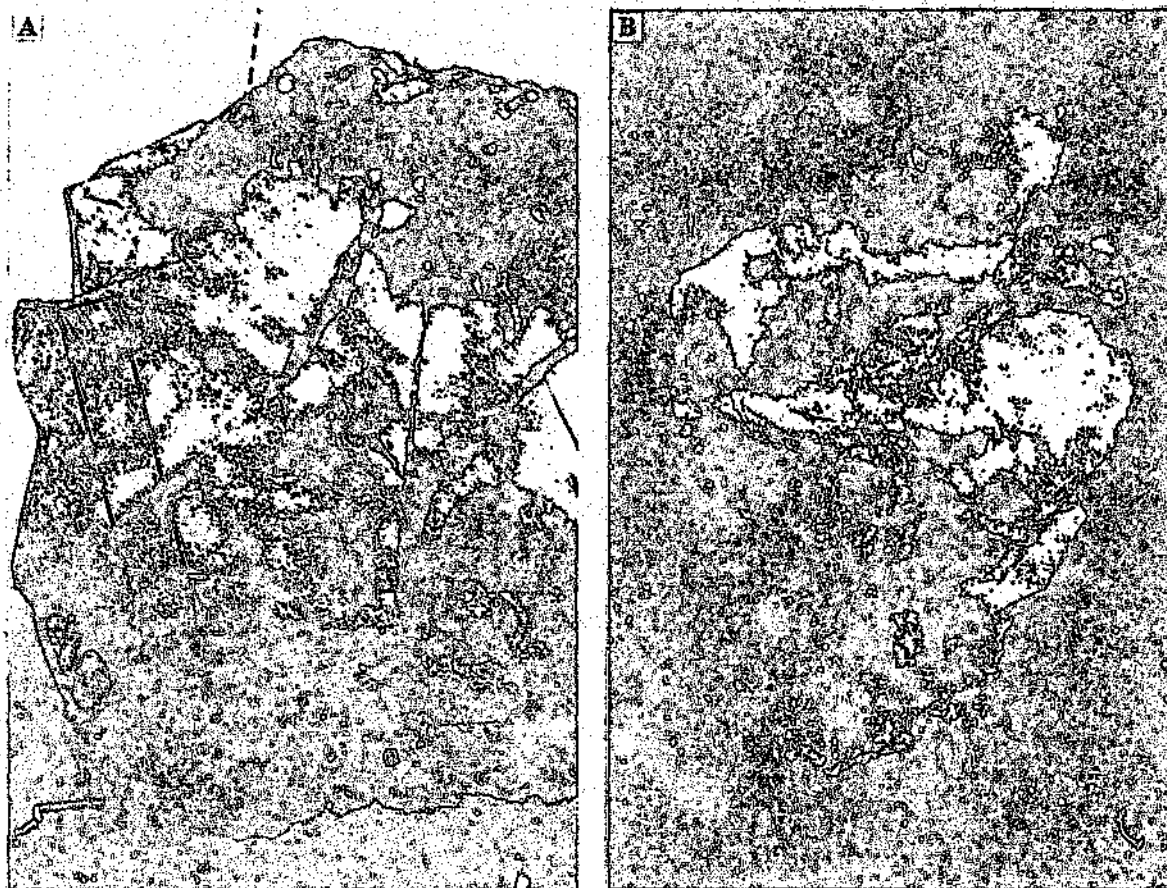


Fig. 4.22

- A** Photomicrograph of a doubly polished chip MC120 consisting of twinned quartz showing growth zonations highlighted by fluid inclusion distribution. Late calcite occurs at the upper- and lower-left edges of the chip. The position of the twin axis (visible only in polarised light) is marked by the stippled lines above and below the chip. The traverse of analysed fluid inclusions lies in the left twin, at approximately right angles to the zonations marked by the superposed box. Vertical dimension of chip = 9 mm.
- B** Photomicrograph of a doubly polished quartz wafer MC140 with incipient mortar texture highlighted in partially polarised light. The largest part of the sample consists of quartz relicts in optical continuity (grey) with slight undulose extinction and interstitially, fine-grained recrystallised quartz (whitish hues). Fluid inclusions of both generations of quartz have been analysed and is presented in Fig. 4.23, 4.30).

parallel-axis twins of the Dauphiné and Brazil laws. Of these the Brazil twins are rare in deformed quartz whereas the Dauphiné twins are common (Blacic and Christie, 1984), suggesting that those of Fig. 4.22A may be of this type. The above-mentioned authors also maintain that the mechanism for such twinning involves small displacive motions of the atoms which does not contribute to any strain of the crystals. The origin of the growth zone fluid-inclusions and their petrographic relationship to the twinned host remains unclear.

The Zone D sample, MC140, represents the most deformed end-member of the suite of samples from the study area. The analysed wafer consists of large grains of quartz showing varying degrees of undulatory extinction set in a finely recrystallised matrix of very clear polygonal quartz, imparting as a whole, a typical mortar texture (Fig. 4.22B). The fluid inclusions in both quartz types were analysed due to the clear chronological and genetic relationships. The early coarse-grained quartz contains random cloud inclusions of the two-phase aqueous type. These inclusions are interspersed with small three-dimensional clusters of H_2O-CO_2 inclusions. The H_2O-CO_2 inclusions contain three phases at room temperature, have constant filling ratio's and are quite scarce compared to the aqueous ones (≤ 2 area % of wafer). Both inclusion-types reveal no particular primary or secondary textural relationships. The clear recrystallised quartz matrix, however, contains only two-phase aqueous inclusions (bubble = 1-2 vol%) which often do not nucleate a bubble after freezing and heating runs, indicating the presence of high density fluids (Reynolds, 1991)

4.5.3 Fluid Inclusion Microthermometry

All fluid inclusions for which freezing data was obtained were relocated via sketch maps for heating measurements. All measurements were duplicated to check for leakages and erroneous observations. Thermometric data were acquired using a Chaixmeca stage (Coty *et al.*, 1976; Hollister *et al.*, 1981). The accuracy of the measurements was determined to be better than $\pm 0.3^\circ C$ for the temperature range below $100^\circ C$ and $\pm 2^\circ C$ for the higher range up to $400^\circ C$.

4.5.3.1 Aqueous Inclusions. Microthermometric data for the aqueous inclusions are

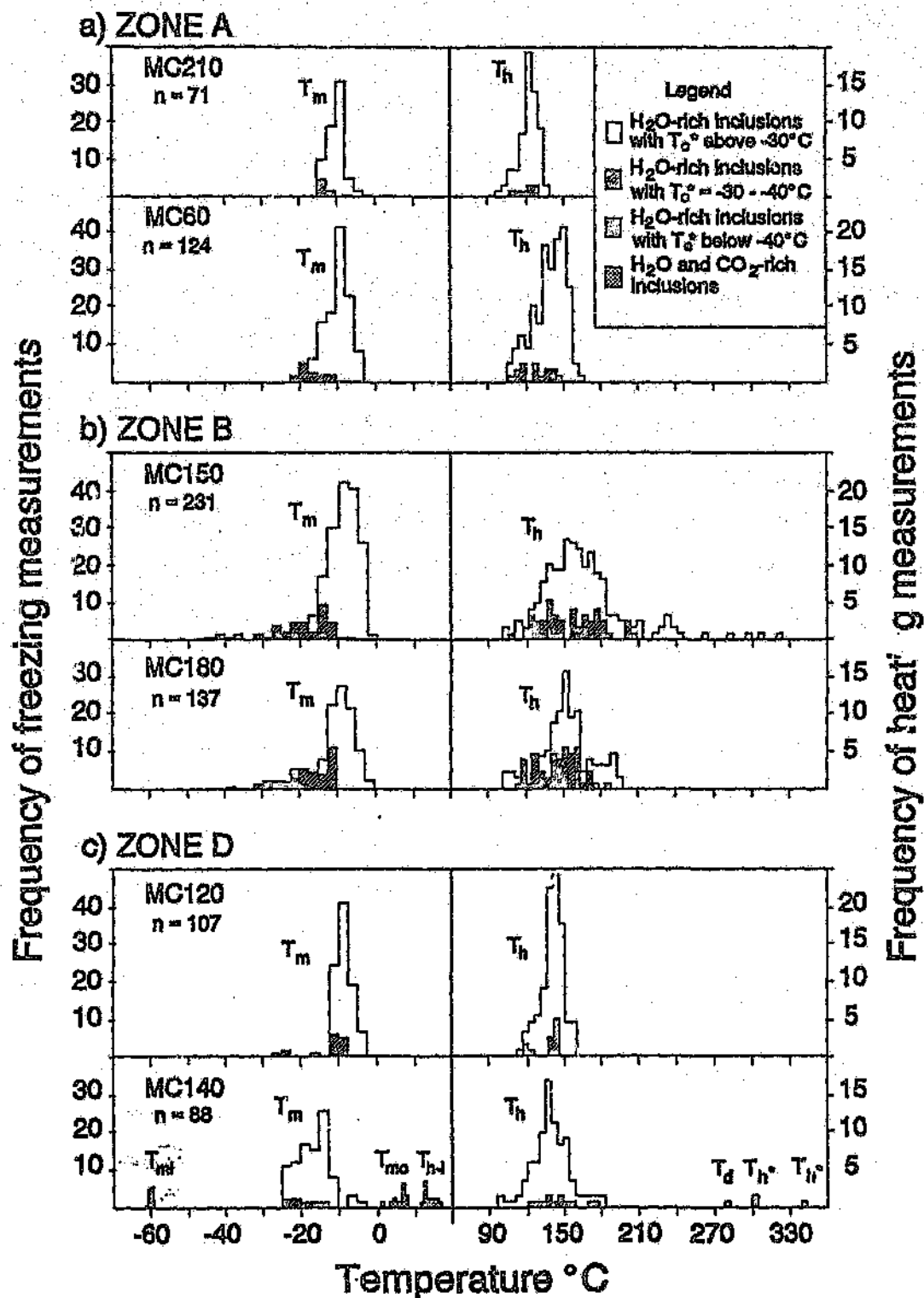


Fig. 4.23

Histograms of freezing and heating measurements of the fluid inclusions of the main metamorphic zones as indicated. The measurements indicated by the clear, hatched and dark symbols of the legend represent temperatures of final melting (T_m) and temperature of homogenisation (T_h). The stippled bars in the histogram are qualified by: T_{m1} = T of final melt of carbonic ice; T_{m2} = T of final melt of clathrate; T_{h1} = T of homogenisation to liquid of the carbonic phase; T_d = T of decrepitation; T_{h*} = T of bulk homogenisation.

summarized in histograms showing temperatures of homogenisation (T_h) and final melting (T_m , Fig. 4.23) and also in XY-scatter plots of T_h vs T_m (Fig. 4.24 to 4.34). Both plots additionally show initial melting data, i.e. temperature ranges of apparent eutectic melting $= T_e^*$. The intention of indicating apparent eutectic temperatures is not to claim the precise temperatures of first melting but rather temperatures at which first melting was clearly observed, i.e. a graininess of the ice accompanied with the appearance or movement of a vapour phase. The reported apparent eutectica probably represent temperatures well above the true eutectic temperatures. Their documentation is nevertheless justified as they still lie well below the minimum equilibrium and/or metastable eutectic temperatures of the most common ternary system $\text{NaCl} + \text{KCl} + \text{H}_2\text{O}$, i.e. at -22.9°C (metastable melting of $\text{NaCl} + \text{H}_2\text{O}$ occurs at -28°C , Crawford, 1981). Groupings of apparent eutectic melting for the following plots are: $\geq -30^\circ\text{C}$, $-30 - -40^\circ\text{C}$ and $\leq -40^\circ\text{C}$. The intervals were chosen arbitrarily and assist in the demonstration of entrapment of fluids with progressively lower apparent eutectics.

Zone A. At room temperature none of the Zone A fluid inclusions are halite-bearing. The freezing behaviour of the fluid inclusions is relatively simple as all final melting is that of ice. Upon warming, after freezing, apparent initial melting of most inclusions (in the presence of a vapour phase) occurs in the range of -23 to -20°C suggesting that the fluids contain mostly NaCl with or without KCl (Crawford, 1981a). A few inclusions, however, showed initial melting below -30°C , suggesting the presence of an additional eutectic-depressing species (Fig. 4.23a). Several inclusions, however, especially those in the outer growth zone of MC210 (Fig. 4.20A), exhibit the tendency of erratic bubble nucleation and therefore disequilibrium melting. Temperatures of final melting for the inclusions range from -22° to -3°C , of which the stratigraphically higher sample (MC210) has a slightly narrower spread (-15° to -4°C). The final melting temperatures of both samples are similar and peak between -10° and -7.5°C .

The homogenisation temperatures of both Zone A samples range from 95° to 165°C (Fig. 4.23). The stratigraphically lower Komati Formation sample MC60 has a slightly higher

T_h vs. T_{fm} of H_2O -rich inclusions of MC210 (Zone A)

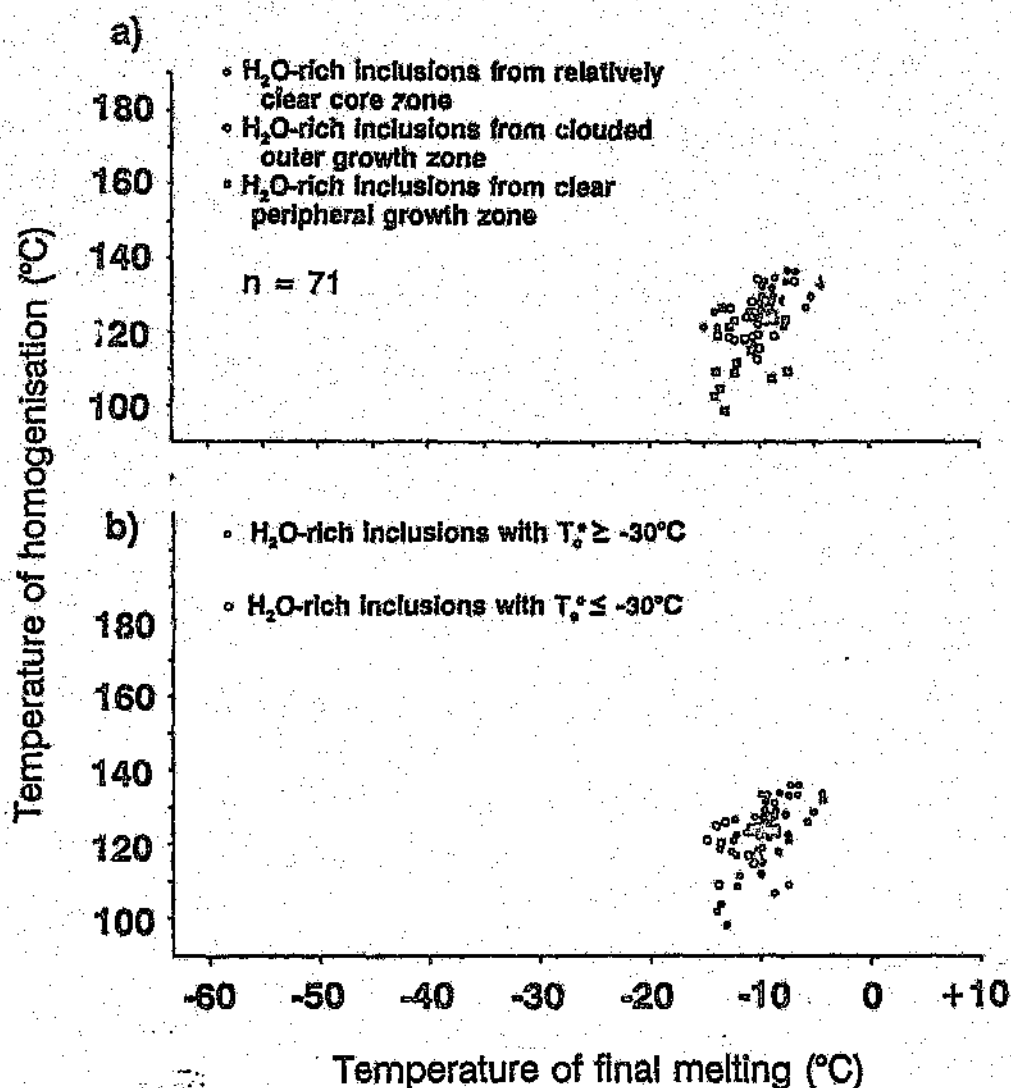


Fig. 4.24

Variation of T of homogenisation vs T of final melting for MC210 fluid inclusions, (a) groupings are according to the origin of inclusions with respect to crystal growth zones, and (b) grouping of data is based on T of apparent eutectics (T_e^*). The growth zones referred to in legend of (a) are from Fig. 4.20A.

T_h vs. T_{fm} for fluid inclusions of MC60 (Zone A)

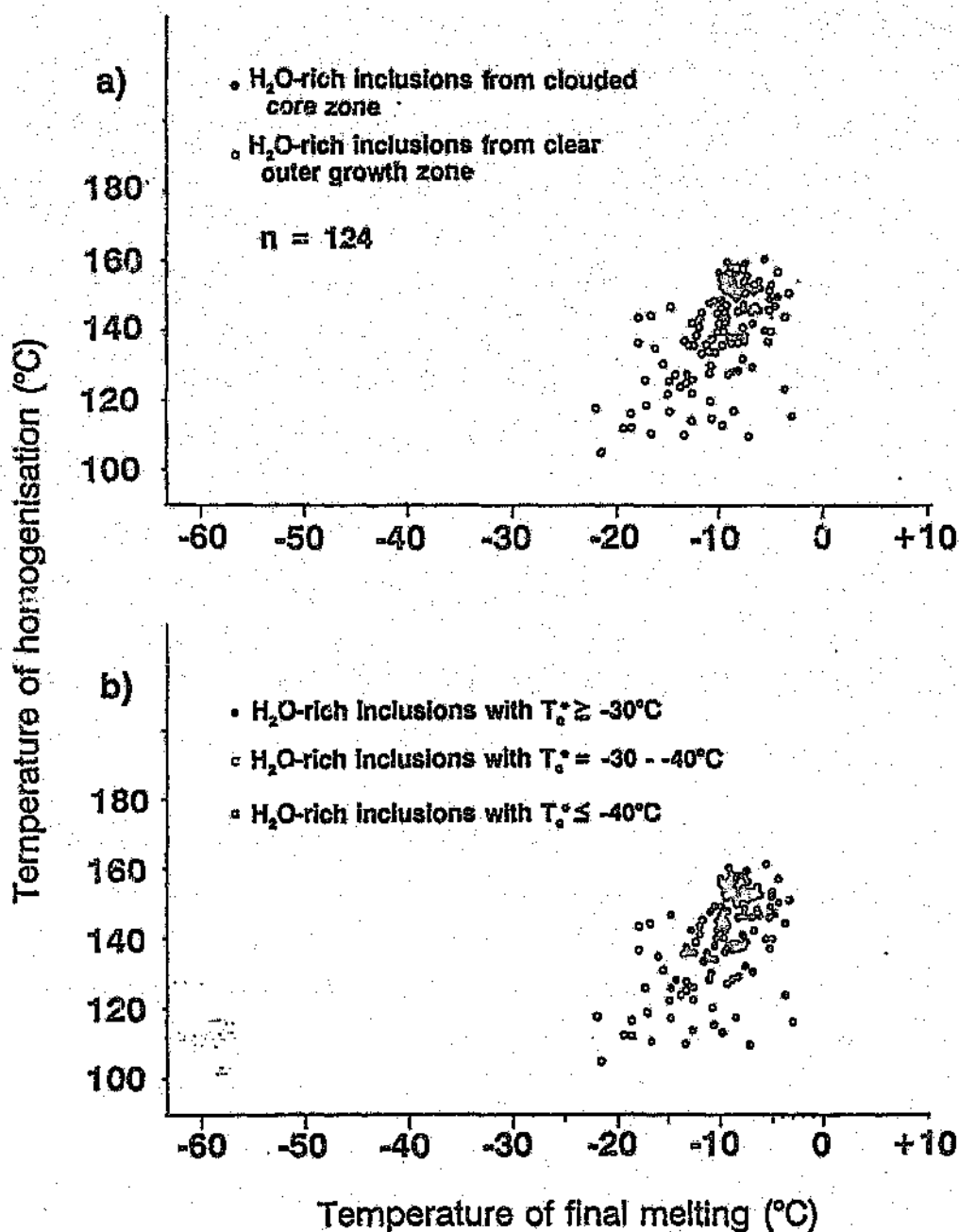


Fig. 4.25

Variation of T of homogenisation vs T of final melting for MC60 fluid inclusions, (a) groupings are according to the origin of inclusions with respect to crystal growth zones, and (b) groupings of data is based on T of apparent eutectica (T_e). The growth zones referred to in legend of (a) are from Fig. 4.20B.

range (105°-165°C) than the overlying Hooggenoeg Formation sample (MC210, 95-138°C). The two samples show homogenisation temperature maxima at 125° and 150°C for the stratigraphically upper and lower samples, respectively.

Despite the relatively narrow ranges displayed by the microthermometric data, both samples also show a progressive variation of fluid properties on a growth-zone scale. The fluids of growth zone inclusions tend to become more saline (lower T_m) and denser (lower T_h) towards the edges of the grains (Figs. 4.20A, 4.20B). For sample MC210 the averages of the analysed growth zones from core to rim are: inner clear zone ($n = 26$), $T_h = 127.3^\circ$ and $T_m = -8.8^\circ\text{C}$; outer dark zone ($n = 18$), $T_h = 123.7^\circ$ and $T_m = -10.6^\circ\text{C}$; peripheral clear zone ($n = 17$), $T_h = 114.9^\circ$ and $T_m = -11.2^\circ\text{C}$. For sample MC60 the averages of the growth zones from core to rim are: dark core zone ($n = 54$), $T_h = 148.7^\circ$ and $T_m = -8.0^\circ\text{C}$; clear outer zone ($n = 70$), $T_h = 134.5^\circ$ and $T_m = -11.6^\circ\text{C}$.

Zone B. The freezing behaviour of the higher-grade Zone B fluid inclusions is similar to that of the Zone A inclusions. However, the proportion of inclusions that have apparent eutectica below -30°C is noticeably larger and the proportion with apparent eutectica below -40°C , are significantly larger (an increase from 1 to 45, Fig. 4.23b). The range of apparent eutectic melting increases from $-37^\circ - -20^\circ\text{C}$ in Zone A to $-62^\circ - -20^\circ\text{C}$ in Zone B. Concomitant with this increase is an expansion in the range of final melting temperature from ca. -22° to -2.5°C for Zone A to $-44^\circ - -2.5^\circ\text{C}$ for Zone B. The final melting maxima of the two Zone B samples are identical and compare favourably with the Zone A samples. The possibility that some of the recorded final-melting T s actually represent that of salt hydrate (hydrohalite) as opposed to that of ice is discounted on the basis of large degrees of bubble shrinkage on the freezing of the inclusions. Large shrinkages only occur in low salinity fluids (Crawford, 1981a).

Solid phases were found in a small proportion ($\leq 20\%$) of the fluid inclusions. The trapped solids are all translucent and have regular to rounded shapes of which a few are isometric and are thought to be halite. The halite crystals preferentially occur in the inclusions with

T_h vs. T_m for fluid inclusions of MC150 (Zone B)

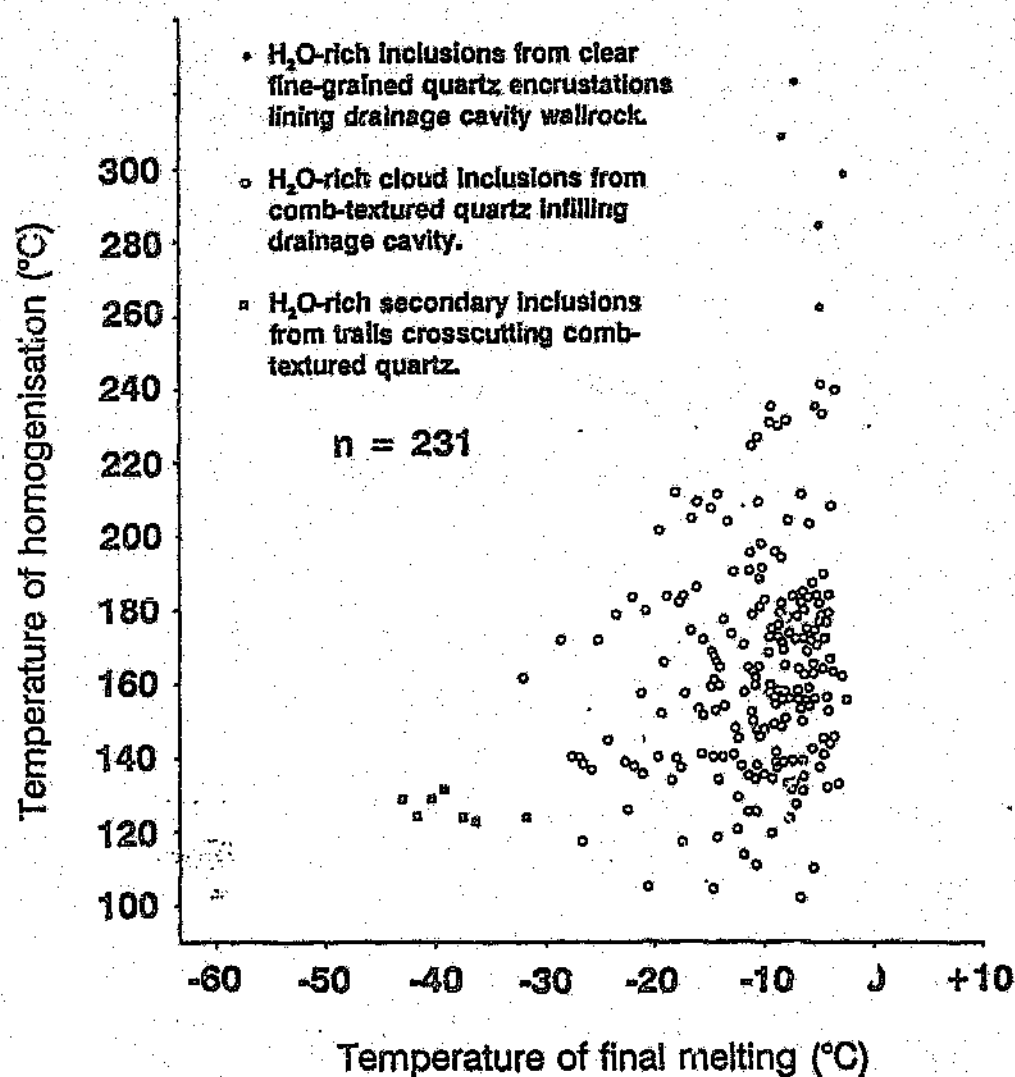


Fig. 4.26

Variation of T of homogenisation vs T of final melting for MC150. Grouping of data is according to fluid inclusion origin based on the quartz petrography referred to in Fig. 4.21A.

T_h vs. T_{fm} for fluid inclusions of MC150 (Zone B)

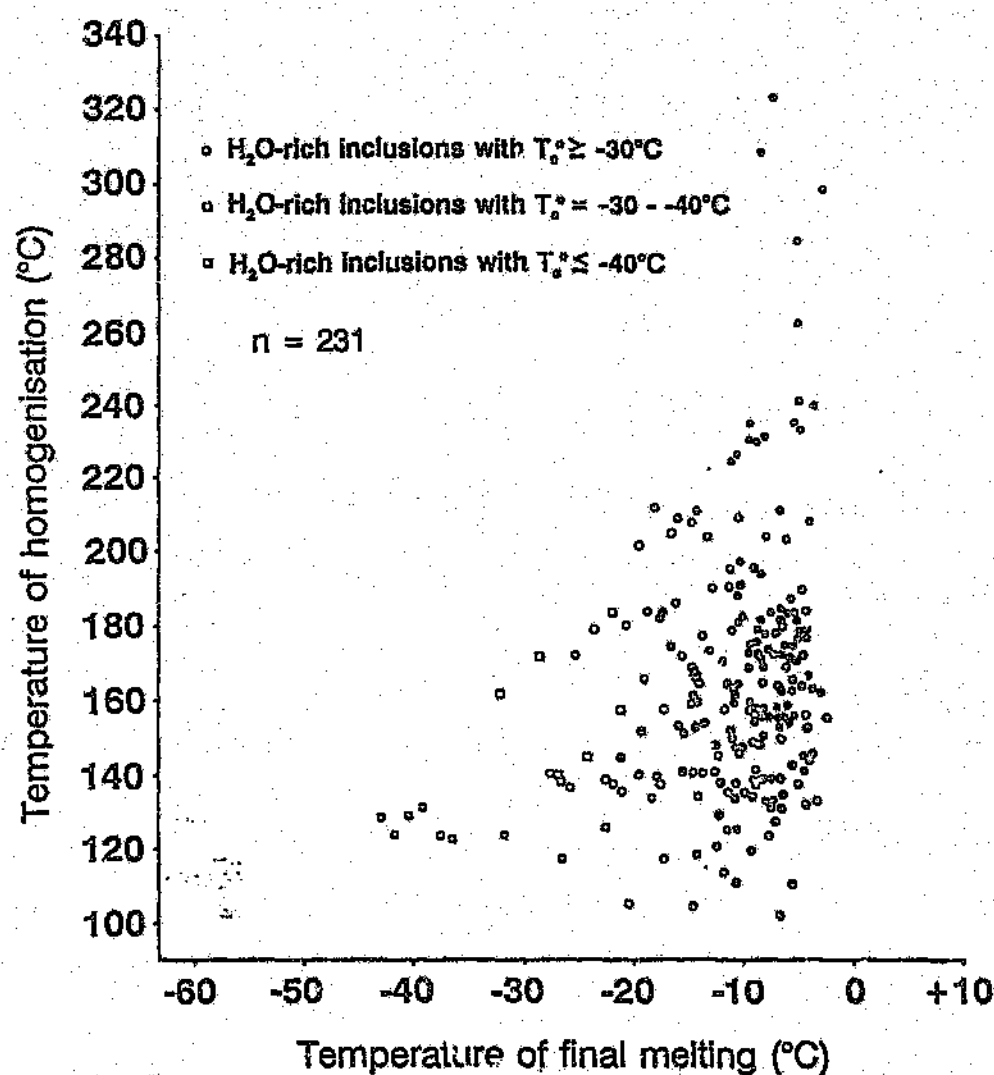


Fig. 4.27

Variation of T of homogenisation vs T of final melting for MC150. Grouping is according to the apparent eutectica of the inclusion fluids.

T_h vs. T_{fm} for fluid inclusions of MC180 (Zone B)

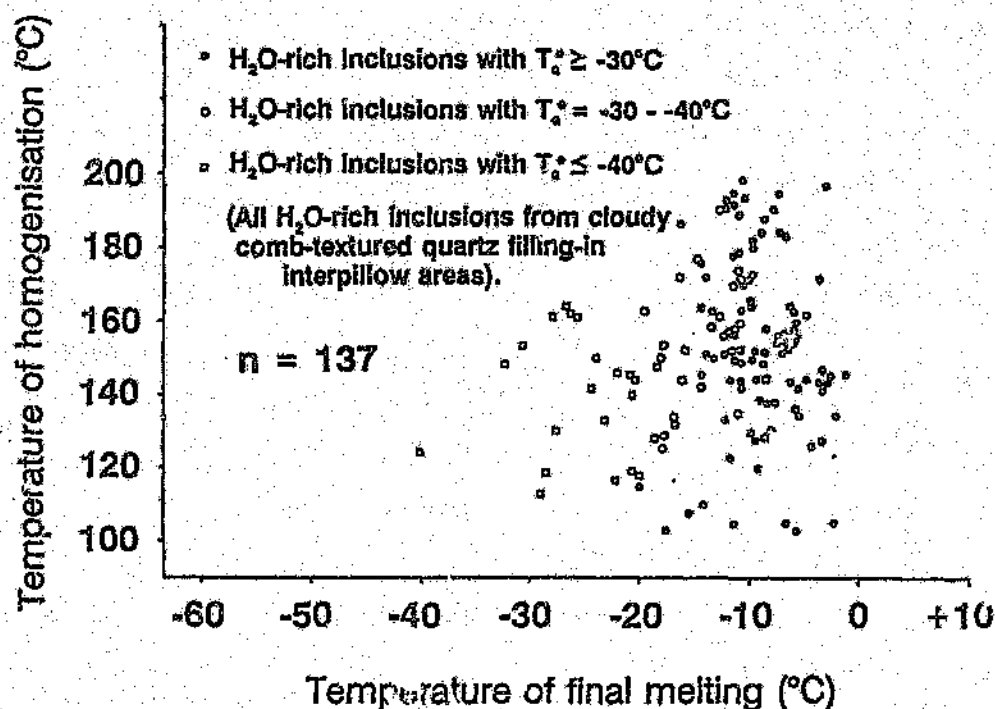


Fig. 4.28

Variation of T of homogenisation vs T of final melting for MC180. Grouping of data according to the apparent eutectica of the inclusion fluids based on the quartz petrography referred to in Fig. 4.21B.

very low apparent eutectica (i.e. those $\leq -40^\circ C$). Of the 45 inclusions that fall within this category only 6 ($\leq 14\%$) contain a possible halite daughter mineral. However, like the halite inclusions, the other trapped solids did not significantly change shape or size upon heating up to the temperatures of vapour/liquid homogenisation (T_h) or subsequent decrepitation. All solid inclusions are therefore thought to represent foreign inclusions and not true daughter minerals. The inclusions that have trapped the solids are thus thought to have (accidentally) trapped heterogeneous fluids (cf Crawford, 1981a).

All Zone B inclusions homogenise into the liquid phase. The bulk of the homogenisation temperatures of the two Zone B samples are essentially similar as they range between 105° - 210°C and both have maxima at ca. 150° - 155°C. Sample MC150, however, has several inclusions which homogenise outside this range. These inclusions originate from the fine-grained quartz encrustation and homogenise in the range 230° to 320°C (Fig. 4.23b). The inclusions which entrained solids also homogenise within the range of the non-solid bearing inclusions, indicating that fluid properties are not significantly different.

Compared with the Zone A samples, the homogenisation maxima of the Zone B samples are only marginally higher (5-25°C). The range of the bulk populations have similar lower limits (ca. 100°C) whereas the upper limits are marginally higher (ca. 40°C).

The T_h vs T_m relationship for the drainage-cavity quartz inclusions of MC150, for which the relative timing of entrapment has been petrographically established (Fig. 4.21A), shows that with a decrease in T_h there is a general decrease of T_m and T_e^* (Fig. 4.26, 4.27). The microthermometric data combined with the petrographic relationships suggest that the encrustation quartz initially trapped fluids with relatively low density and low salinity (i.e. high T_h and high T_m). Subsequently, during growth of the columnar grains, moderately dense and low to moderately saline fluids were incorporated to form the cloud-inclusions. Microfracturing and annealing of the columnar grains was the final mechanism of fluid entrapment leading to the formation of the secondary inclusions which became increasingly more saline and dense (i.e. low T_h and low T_m). At the same time, as evidenced by the progressive lowering of apparent eutectica (T_e^*), greater proportions of an eutectic-depressant(s) were being incorporated (Fig. 4.27). The early encrustation- and cloud inclusion fluids have a clear propensity to have apparent eutectica above -30°C and -40°C respectively, whereas the late secondary inclusions show a strong tendency to be below -40°C suggesting a progressive change in the proportion of salt species with time.

The T_h vs T_m relationships of MC180 is similar to that of the MC150 data (Fig. 4.28). The correspondence suggests that despite differences in the origin of the host quartz and the

more than 1 km strike difference between the samples, fluids with essentially similar physical properties were trapped over a large part, if not the whole of Zone B. This phenomenon points to a regional control of fluid composition and density. (The regional control of the two Zone A samples is also eminent despite the large (4 km) stratigraphic height differences and interformational differences, namely Komati vs Hooggenoeg Formations).

The most prominent compositional difference between Zone A and -B samples is the presence of fluids with T_c^* below -40°C and their tendency to occur in secondary healed fractures. The microthermometric similarities of the cloud-inclusions of both zones also represent an important constraint on the evolution of the fluids. The regional differences in fluid properties between the metamorphic zones, occurring in rocks that are essentially similar in composition, except for their metamorphic grade, seems to suggest that the physical properties of the fluid systems were metamorphically controlled.

Zone D. The fluid inclusion freezing behaviour of the two Zone D samples (MC140 and MC120) is dissimilar. They have contrasting ranges of final melting with very little overlap i.e. -24.5° to -10°C vs -12.5° to -2.5°C , respectively. Less than 5% of the inclusion data scatter outside the mentioned ranges (Fig. 4.23c). The two populations have melting maxima between -15° - -12.5°C and -10° - -7.5°C , respectively, which involves the tectonically more deformed sample MC140 trapping more saline fluids than the less deformed sample MC120. A distinct shoulder at the low temperature end of the histogram plot of MC140 shows that there is a sharp cutoff for final melting below -24.2°C (Fig. 4.23c).

The ranges of the temperature of homogenisation for both samples are essentially the same, with that of MC120 being slightly narrower (115° - 160°C) than that of MC140 (115° to 185°C , Fig. 4.23c). The homogenisation maxima are also similar, i.e. 135° - 140°C and 140° - 145°C for MC140 and MC120, respectively.

The final melting T ranges and maxima of MC120 appear to correspond more closely to the static-style Zone A and B samples than with that of its more deformed counterpart MC140

T_h vs. T_m for fluid inclusions of MC120 (Zone D)

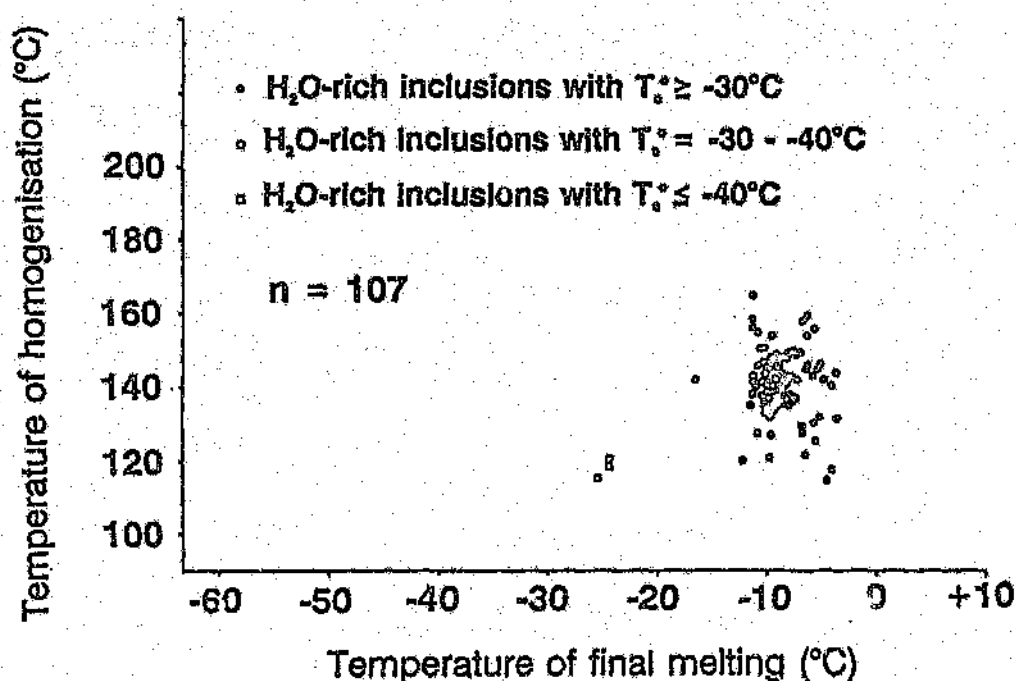


Fig. 4.29 Variation of T of homogenisation vs T of final melting for MC120. Grouping of data according to the apparent eutectica of the inclusion fluids analysed on the traverse shown in Fig. 4.22A.

(Fig. 4.27c). The abundance, however, of low-eutectic fluid data for MC120 is noticeably less than for the statically metamorphosed samples. These low eutectic inclusions were encountered as part of a three-dimensional cluster close to the core of the quartz grain. It was not clear whether they belonged to a healed fracture or not.

The inclusions in the twinned quartz grain from the intermediately deformed pillows (MC120) show a tight cluster of data points (Fig. 4.29) as opposed to two less restricted,

T_h vs. T_{fm} of H_2O -rich inclusions of MC140 (Zone D)

(Data for $H_2O + CO_2$ inclusions elsewhere - see text)

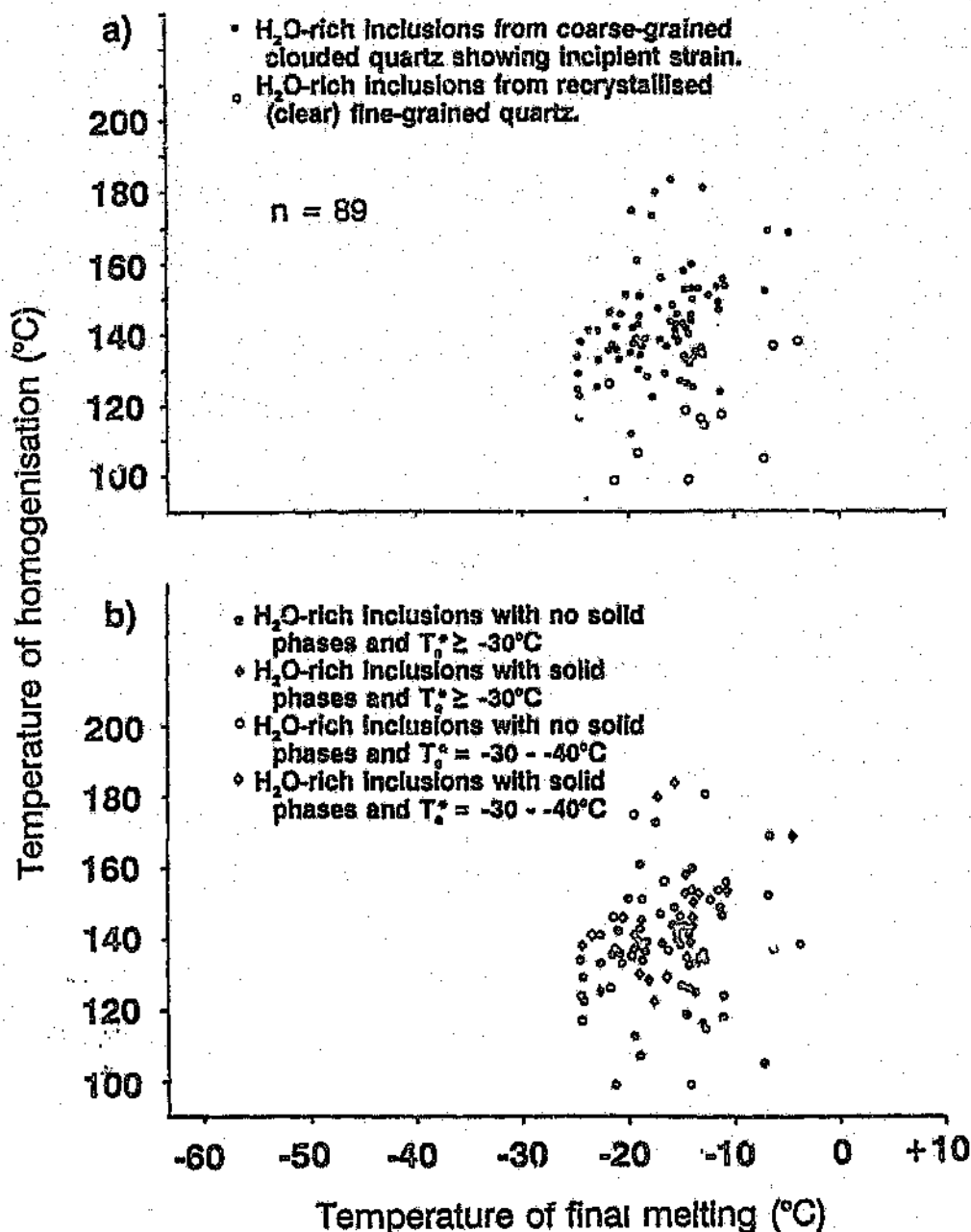


Fig. 4.30

Variation of T of homogenisation vs T of final melting for MC140 fluid inclusions, (a) groupings are according to the origin of inclusions with respect to quartz generation (Fig. 4.22B), and (b) grouping of data is based on T of apparent eutectica and the presence or absence of solid inclusions.

partially overlapping data clusters for MC140 (Fig. 4.30). Three outliers with eutectica below -40°C plot away from the main data group. Except for these inclusions, no other variation in fluid properties across the growth zones were found. This contrasts with that found for MC140 which displays two distinct clusters obtained for the pre- to syn- and post tectonic quartz, respectively (Fig. 4.30).

The pre- to syn-deformational coarse-grained quartz contains the bulk of the analysed aqueous inclusions as well as the $\text{H}_2\text{O}-\text{CO}_2$ inclusions. The post-deformational, finely recrystallised quartz matrix, on the other hand, has trapped fluids that are more dense (lower T_h) and marginally less saline (higher T_m). The fluids trapped in the quartz matrix seem to have a similar composition to that of the precursor quartz as the ranges of final melting demonstrated by the two populations are comparable. The tendency of the late fluids to be of similar salinity contrasts with the static-style fluids which tend to become progressively more saline with time.

Trapped solids occur in approximately 20% of the analysed inclusions of which $\leq 5\%$ contain small isometric daughter minerals which may be halite. The inclusion solids, apart from having an erratic mode of occurrence, are not uniform in shape or size. At room temperature, the solids are mostly anisotropic and, together with those that could be halite, do not dissolve upon heating before the bubbles disappear or subsequent decrepitation. The presence of trapped solids in the inclusions does not appear to influence homogenisation temperatures, apparent eutectica or final melting as is demonstrated by Fig. 4.30b. The inclusions that have trapped such solids (including those believed to be halite), therefore, are thought to have derived from non-homogenous fluids (Crawford, 1981), implying that the crystals were present in the fluid prior to being trapped.

4.5.3.2 $\text{H}_2\text{O}-\text{CO}_2$ inclusions. The $\text{H}_2\text{O}-\text{CO}_2$ inclusions have clathrate final-melting temperatures in the range of 1° to 9°C and carbonic phase-melting temperatures (T_{mc}) below -58° (Fig. 4.23c) implying the presence of other gas species. Because of the presence of these species the clathrate-melting temperatures cannot be used to estimate salinity. Upon heating

most of the inclusions decrepitated before bulk homogenisation. Only three were seen to homogenise between 302° - 342°C. These homogenisations, however, can be used as a rough minimum temperature of entrapment.

4.5.4 Salinity Estimates

The salinities of the fluid inclusions are listed as averages with one sigma standard deviation in Table 4.1 together with other relevant data such as temperatures of final melting, temperatures of homogenisation and fluid densities. The averages of final melting and homogenisation are obtained from the data plotted in Fig. 4.23 to 4.30. The salinities for the various aqueous inclusion types have been derived by modelling the phase behaviour in the system NaCl-CaCl₂-H₂O using the final melting of ice (T_m) in accordance with the temperature contoured ternary phase diagram of the system (Fig. 4.31). Reasons for the choice of the system is discussed below. Isochores, lines of constant density in PT-space, have also been calculated from the T_h of the inclusion-types with the aid of the computer program GEO-HALWAT (Niccols and Crawford, 1985). The fluid properties of the various samples are displayed according to metamorphic zones with the exception of MC180 which is very similar to MC150 and is hereafter represented by the latter.

The determination of fluid inclusion salinities has been extensively discussed in the literature (e.g. Crawford, 1981a; Roedder, 1984) and will not be repeated here. Such determinations are usually done by measuring the final melting point of ice in inclusion-fluids, that is if their composition is less saline than the eutectic composition of the system. If their compositions are more saline, the final melting or dissolution of salt hydrate or salt daughter crystals is determined. In both instances the final melting temperatures can be converted to weight percent NaCl equivalent.

Identification of the fluid systems in the practice of microthermometry is usually done by the identification of their eutectic temperatures (Crawford, 1981a). Generally aqueous fluid inclusions contain monovalent cationic salts of which NaCl followed by KCl are the most common. The fluid systems of inclusions containing the monovalent cationic salts can either

Table 4.1 A summary of the fluid inclusion properties of the studied samples emphasising sequence of entrapment vs metamorphism.

Sample No	Metamorphic Zone	Fluid properties	Paragenetic sequence of fluid inclusion-bearing quartz					
			Static-style samples			Dynamic-style samples		
			Inclusions of d. cavity lining	Cloud-incls. of columnar grains	Inclusions of healed fractures	Twinned quartz	Mortar-textured quartz	
							Relict grains	Recryst. matrix
MC210	Zone A	T _m (°C) Salinity (wt% aNaCl) T _h (°C) Density (g/cc)		-10.0 ± 2.4 14 ± 2.5 122 ± 8.5 1.043				
MC60		T _m (°C) Salinity (wt% aNaCl) T _h (°C) Density (g/cc)		-10.2 ± 4.0 14.5 ± 4.5 141 ± 13.6 1.022				
MC150	Zone B	T _m (°C) Salinity (wt% aNaCl) T _h (°C) Density (g/cc)	-8.0 ± 1.9 8.0 ± 3.0 287 ± 27.3 0.831	-9.0 ± 3.9 13.0 ± 4.5 163 ± 30.6 0.999	-20.9 ± 9.0 22.0 ± 7.0 153 ± 24.1 1.0812			
MC180		T _m (°C) Salinity (wt% aNaCl) T _h (°C) Density (g/cc)		-8.6 ± 3.5 12.5 ± 4.0 151 ± 21.9 1.002	-18.9 ± 6.6 20.5 ± 6.0 147 ± 17.2 1.063			
MC120	Zone D	T _m (°C) Salinity (wt% aNaCl) T _h (°C) Density (g/cc)				-9.1 ± 3.6 13.0 ± 4.0 139 ± 9.6 1.020		
MC140		T _m (°C) Salinity (wt% aNaCl) T _h (°C) Density (g/cc)					-16.8 ± 4.4 19.5 ± 3.0 151 ± 14.1 1.061	-13.1 ± 5.4 16.5 ± 5.0 122 ± 13.8 1.042

be represented by binary or ternary phase diagrams, of which the eutectica are all above -22.9°C (*cf* Table 4.2). The eutectica of any of these systems will be significantly depressed, to between -56° and -35°C , by the presence of any one of the more common divalent cationic salts such as CaCl_2 and MgCl_2 .

The fact that apparent eutectica below -30°C have been observed for a significant portion of the analysed samples, suggests that dissolved species other than Na, K and Cl are present. CaCl_2 , rather than MgCl_2 , has been shown to occur commonly in oceanic hydrothermal systems (Hardie, 1983; Vanko *et al.*, 1988), Archaean gold deposits (Robert and Kelly, 1987) and low-grade metamorphic environments (Crawford *et al.*, 1984). CaCl_2 can depress the eutectic melting of aqueous systems to -52°C in the system $\text{NaCl}-\text{CaCl}_2-\text{H}_2\text{O}$ (Yanatieva,

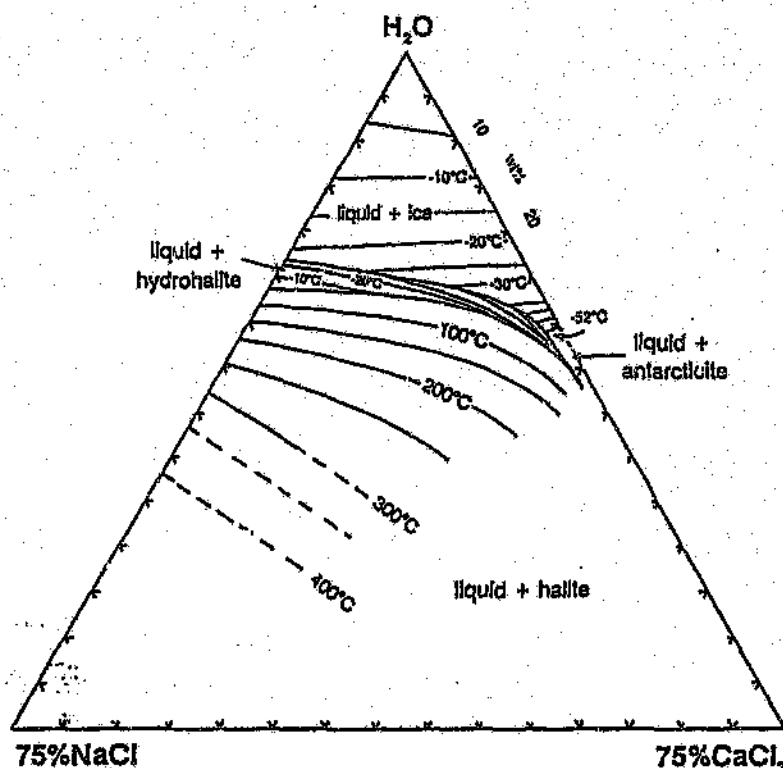


Fig. 4.31

Phase diagram for the ternary system $\text{H}_2\text{O}+\text{NaCl}+\text{CaCl}_2$ with temperature contours. The data used to construct the phase boundaries and subzero isotherms above -35°C are from Oakes *et al.* (1990) and those below -35°C as well as the isotherms for the halite-saturation surface are from Williams-Jones and Samson (1990).

Table 4.2 Selected phase data for aqueous solutions of chloride species commonly found in fluid inclusions (after Crawford 1981, Table 4.1).

Dissolved species	Eutectic temperature (°C)	Eutectic composition (wt. %)	Solid phases			Solid melting relations
			H ₂ O	ice	hexagonal colorless RI = 1.313 ω 1.309	0°C, congruent
NaCl	-20.3	23.3% NaCl	NaCl · 2H ₂ O	hydrohalite	monoclinic colorless RI 1.416	+0.1°C, incongruent
			NaCl	halite	cubic colorless RI 1.544	
KCl	-10.6	19.7% KCl	KCl	sylvite	cubic colorless yellowish RI 1.490	
CaCl ₂	-49.8	30.2% CaCl ₂	CaCl ₂ · 6H ₂ O	antarcticite	hexagonal colorless RI = 1.393 ω 1.417	+30.08°C, incongruent
MgCl ₂	-33.6	21.0% MgCl ₂	MgCl ₂ · 12H ₂ O			-16.4°C, congruent
NaCl-KCl	-22.9	20.17% NaCl 5.81% KCl				
NaCl-CaCl ₂	-52.0	1.8% NaCl 29.4% CaCl ₂				
NaCl-MgCl ₂	-35	1.56% NaCl 22.75% MgCl ₂				

1946; reported in Vanko *et al.*, 1988) and to -56°C in the system $\text{NaCl-KCl-CaCl}_2\text{-H}_2\text{O}$ (Konnerup-Madsen 1979). However, a clear apparent eutectic of -62°C observed here for some of the fluids requires the presence of yet an additional component/s. The exact salinity of such fluids can only be determined if the principal dissolved components are known and if the relevant phase diagrams exist. However, in the absence of knowing the main dissolved species and/or having the thermochemical data for complex systems, the compositions of the included fluids here are determined by modelling their phase behaviour in the system $\text{NaCl-CaCl}_2\text{-H}_2\text{O}$ using the $T_{\text{m(ice)}}$ in conjunction with the temperature-contoured ternary phase diagrams for this system (Williams-Jones 1990, for temperatures below -35°C ; Oakes *et al.*, 1990, for temperatures above -35°C).

The grouping of apparent eutectica into various temperature ranges have been arbitrarily chosen and serve to graphically highlight the progressive changes of observed initial melting in the successive metamorphic zones. Final melting below and above -28°C separates the entrapped fluids into CaCl_2 - and NaCl dominated fluids, i.e. $X_{\text{NaCl}} \leq$ and ≥ 0.50 , respectively. The tendency of the encrustation quartz-inclusions and most of the cloud inclusions of Zone B to have apparent eutectica above -30°C suggest that they have trapped NaCl -dominated fluids. In contrast, the paragenetically late secondary inclusions of Zone B, which have inclusions with apparent eutectica mostly below -30°C , suggest that the late fluids progressively became richer in- and finally CaCl_2 -dominated.

4.6 Conditions of Metamorphism

4.6.1 Mineral Equilibria

Data for the metamorphic assemblages, geothermometry and barometry of the metamorphic zones are summarized in a petrogenetic grid in which the conditions of metamorphism are bracketed (Fig. 4.32).

The main feature of Fig. 4.32 is the presence of three reaction zones within a narrow P-T band situated between the greenschist- and amphibolite fields, referred to commonly as the greenschist-amphibolite transition zone. The P-T position of the transition zone used here,

stems from a comprehensive study by Maruyama *et al.* (1983). It shows the transition zone to comprise a complex region of reactions, marked by the curve for coexisting actinolite and hornblende, the zone for coexisting oligoclase and albite and the chlorite- or epidote-out curve. In the metabasalt system the reactions are interlinked and characterised by the partial dehydration reactions:

- (1) Clinzoisite + tremolite + chlorite = hornblende + H_2O
- (2) Clinzoisite + chlorite = oligoclase + hornblende + H_2O
- (3) Oligoclase + chlorite + tremolite = hornblende + H_2O

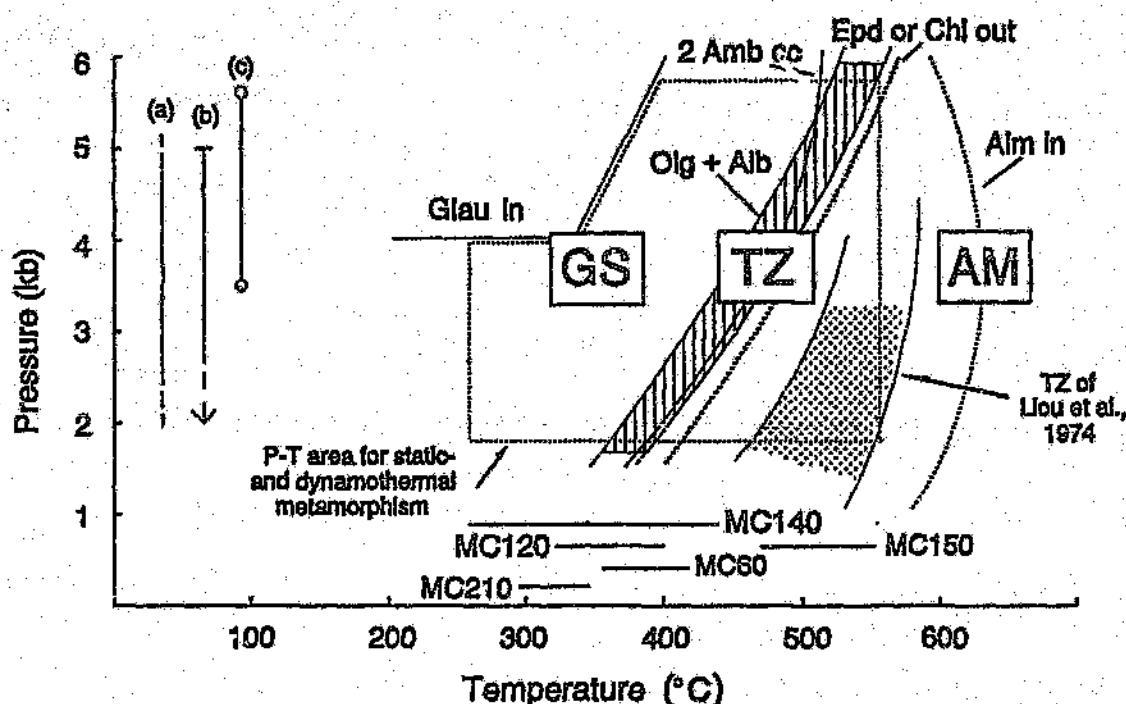
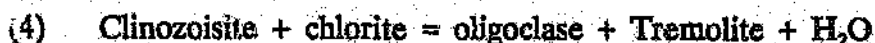


Fig. 4.32

A petrogenetic grid illustrating the phase relations of the greenschist (GS) to amphibolite (AM) transition deduced from natural parageneses (after Maruyama *et al.*, 1983; 2 Amb cc = two amphibole critical curve; Olg + Alb = peristerite field; Epd or Chl out = start of amphibolite field). Also superposed are the experimentally determined position of the transition zone (TZ) of Liou *et al.* (1974); Glauc in = glaucophane in field of Maresch (1977); Alm in = almandine in field of Thiéblemont *et al.* (1988); horizontal bars = geothermometer temperatures for the studied samples from Fig. 4.16; vertical bars from Fig. 4.17: (a) medium pressures indicated by the Hynes plot, (b) $\sim \pm 5$ kb indicated by the Raase plot, (c) medium pressures cited by Laird and Albee (1981). Medium pressures are taken to be = 2 - 5 kb at $\pm 500^\circ\text{C}$ according to Winkler (1979, Fig. 1.1).



The reactions lead to several assemblages, namely two Ca-amphiboles + two sodic plagioclases, one Ca-amphibole + two sodic plagioclases and hornblende + plagioclase together with epidote, chlorite and quartz (distinction between the first two assemblages depends on the host rock FeO/MgO ratio). The high-pressure transformation of the zone is characterised by the assemblage albite + hornblende, termed the albite-epidote amphibolite facies by Miyashiro (1973) whereas the low pressure one is defined by actinolite + calcic plagioclase.

The greenschist-amphibolite transition has also been experimentally studied by Liou *et al.* (1974) and Moody *et al.* (1983) between 2 - 5 and 2 - 4 kb, respectively, and by Apter and Liou (1983) between 5 - 7 kb. The results of the studies are in general agreement with the transition zone studies based on natural assemblages (e.g. Harte and Graham, 1975; Liou and Ernst, 1979; Laird 1980; Laird and Albee, 1981; Maruyama *et al.*, 1983). Available experimental results for the position of the transition zone are almost consistent and are marked by chlorite-decreasing and chlorite-out reactions (e.g. Liou *et al.*, 1974, Moody *et al.*, 1983). Compared to the transition zone position derived from natural assemblages (Maruyama *et al.*, 1983), the experimental data, however, define a range and occur at slightly higher temperatures (Fig. 4.32). Maruyama *et al.* (1983) point out that a problem of the experimental studies may arise from the fine grain size in the run products which inhibits detection of the nature of coexisting plagioclases and/or amphiboles. An advantage of these studies is, however, that they have been run over a wide range of oxygen buffers, notably by Moody *et al.* (1983). They report that the transition zone is strongly f_{O_2} dependent and it may be constricted to a T range as little as 30°C in more oxidizing environments and broaden to a range of as much as 150°C in the more reducing environments.

The Zone A and -D rocks comprise the assemblage actinolite + albite + epidote + chlorite characteristic of the greenschist field (Fig. 4.32). The Zone B samples consist of assemblages in which the breakdown of chlorite is taking place, peristerite or consolute oligoclase and andesine occur to the exclusion of each other, and magnesiohornblende occurs with or

without actinolite. Hornblende and andesine coexist but not without chlorite or epidote. The Zone B samples comprise the phase assemblages of the transition zone (discussed above) and thus, the conditions of metamorphism can be ascribed to this narrow T range. Assemblages typical of amphibolite grade were not encountered.

Accurate geobarometers for medium pressure metabasic rocks do not exist (cf also Ishizuka, 1985; Donato, 1989). However, a consistent, semi-quantitative estimation of the pressure is given by the three amphibole pressure plots of Fig. 4.17. The Hynes-plot indicates medium pressure facies series, which according to Winkler (1979, Fig. 1.1) amounts to ca. 2-5 kb at around 500°C. The Raase-plot indicates pressures just below 5 kb which is in good agreement. On the Laird and Albee plot the amphiboles of the study area correspond best to the worldwide amphibole occurrences of medium-pressure facies series origin. The Vermont medium-pressure amphiboles, for example, plot between 3.5 - 5.7 kb (vertical bars, Fig. 4.32).

The empirical geothermometer of Spear (1981) has produced a range of temperatures for individual samples which are plotted as horizontal bars along the temperature axis of Fig. 4.32. The thermometer temperatures and the ranges of the pressure-plots demarcate a P-T area in which static- and dynamic metamorphism of the study area seem to have occurred. Although the narrow P-T band of the transition zone traverses the central portion of the P-T box, it is nevertheless unsatisfactorily large and too unspecific to resolve any details of metamorphism.

4.6.2 Fluid Inclusion Thermobarometry

To constrain the pressures of metamorphism more precisely fluid inclusion thermobarometry is utilised. The method involves the microthermometric properties of fluid inclusions to construct isochores which are equal density curves and provide a sliding scale for the conditions at which the inclusions were trapped. If either temperature or pressure of entrapment can be determined separately the isochore intersections serve either as a barometer or thermometer (e.g. Hollister *et al.*, 1979; Hollister and Crawford, 1982;

Silverstone and Spear, 1985; Hames *et al.*, 1989).

For the study area, the fluid inclusion data summarised in Table 4.1 enable the construction of isochore ranges for the individual samples. The Spear geothermometer, applied to plagioclase and amphibole pairs, independently constrains the thermal ranges of the samples (Fig. 4.16). As the fluid trapping sequences of the rocks are also known with respect to their thermal evolution, the appropriate pressures of metamorphism can be obtained. Isochores representing the minimum and maximum densities of the final-stage metamorphic fluids are

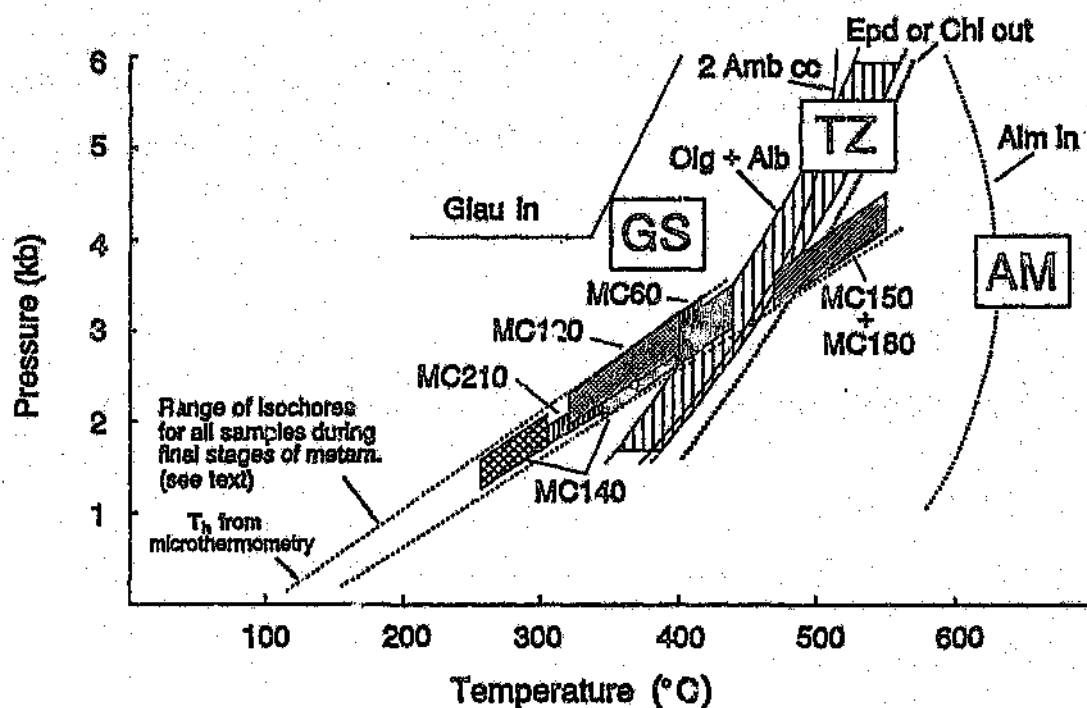


Fig. 4.33

Progenetic grid showing the same natural metamorphic mineral equilibria as Fig 4.32 with the P-T boxes (maxima) for the individual samples. The P-T boxes are situated at the isochore and geothermometer temperature intersections for the respective samples. Note that an independent geothermometer is utilised for determining the temperature of formation (i.e. the trapping T of the fluids). This precludes the need for a "pressure correction" to be made (*c/* Roedder, 1984). Note also that virtually all the P-T boxes fall within the large (dash) outlined PT area indicated in Fig. 4.32. Samples represent the following metamorphic zonations: Zone A = MC210 and MC60; Zone B = MC150 and MC180; Zone D = MC120 and MC140.

depicted by the slopes of the upper and lower margins of each of the shaded P-T boxes. The entire isochore range is schematically shown by the (thin) stippled lines of Fig. 4.33. The origins of the plotted isochores mark the homogenisation maxima of the fluid inclusions (summarised in Fig. 4.23 and Table 4.1). The vertical terminations of the boxes are determined by the minimum and maximum geothermometer ranges.

For the P-T boxes to be valid, two major assumptions are made: (1) that fluid entrapment occurred synchronously with the thermal (thermometer) maxima of the samples; and (2), that both the volume and the mass of the inclusions have remained constant subsequent to trapping. Data in support of the assumptions are presented below. The individual P-T boxes shown in Figure 4.33 are in good agreement with the broad field of metamorphism indicated above (PT-box with stippled outline, Fig. 4.32) and they plot across the transition zone as defined by Maruyama *et al.* (1983). The validity of the method employed here will be evaluated further, subsequent to the presentation of more data on the conditions of metamorphism.

Zone A. The P-T boxes of MC210 and MC60 clearly fall within the greenschist facies corroborating the metamorphic mineral assemblage data. The designated P-T area of the stratigraphically higher sample, MC 210, of the Hooggenoeg Formation has an average pressure of 1.9 ± 0.3 kb whereas that of MC60 of the underlying Komati Formation is at 2.9 ± 0.5 kb. Their temperatures are $320 \pm 25^\circ\text{C}$ and $384 \pm 30^\circ\text{C}$, respectively.

Zone B. The P-T boxes representative of this zone, i.e. MC150 and MC180, overlap and plot within a single field with $P = 3.9 \pm 0.6$ kb and $T = 510 \pm 30^\circ\text{C}$, essentially confirming the assemblage- and pressure plot data (Fig. 4.32). The lower portion of the P-T box intersects the transition zone described by Maruyama *et al.* (1983), with the upper portion reaching into the amphibolite facies field. The lower portion, however, intersects the three main reaction zones of which the plagioclase and amphibole ones seem to be in good agreement with the observed assemblage data. The third, the chlorite or epidote out curve seems disproportionately overstepped by the P-T box compared to the relatively few occurrences

of chlorite-breakdown. This may imply: 1) that the thermometer temperature average is 15 - 25°C too low; 2) or that the width of the transition zone is somewhat wider than indicated by Maruyama and co-workers. From experiments, it is well known that the transition zone width is quite variable as a result of a_{FeO} , f_{O_2} and $a_{\text{H}_2\text{O}}$. The activities of these species will significantly influence equilibrium temperatures of reactions in the metabasite system and therefore the position of the transition zone (e.g. Harte and Graham, 1975; Maruyama *et al.*, 1983; Moody *et al.*, 1983). For example, Al-substitution into amphiboles (the actinolite-hornblende conversion) takes place at higher temperatures for rocks with lower Mg/Fe. The fact that the chlorite-out curves of Liou *et al.* (1974) and Moody *et al.* (1983, not shown) are not exceeded by the P-T box position (Fig. 4.32) seems to support the latter possibility. The fact that the assemblages studied here have formed under relatively low f_{O_2} conditions, as suggested by the absence of Fe-oxide minerals, would also support a wider transition zone (Moody *et al.*, 1983). Isochore slope variations would also affect the relative position of the P-T box influencing the apparent degree of over- or understepping. As the main reactions of the transition zone are all intersected by the independent technique, it would appear that application of the isochore-geobarometer is justified and that reasonable results have been produced with respect to the Zone B samples.

Zone D. MC120 and MC140 plot within the greenschist field, verifying the phase assemblage- and pressure-plot data. MC120 has an average $P = 2.6 \pm 0.6$ kb and $T = 360 \pm 50^\circ\text{C}$. The wide temperature range is due to the difference in thermal blocking of the pillow core and -rim assemblages. The thermal range for MC140 was found to be wider still with the pillow core bracketed between 440 - 345°C and pillow rim at a distinctly lower range of $280 \pm 25^\circ\text{C}$. The pillow interstices, where the latter temperatures were recorded also contained the cataclastic quartz with the two generations of fluid inclusions described earlier. As a result of the close spatial association of the cataclastic quartz and the low-temperature thermometer pairs the timing of fluid entrapment of the older quartz is equated with the higher temperature span of the pillow cores whereas entrainment of the younger fluids is assigned to the lower temperature event of the pillow rim. The isochore/thermometer intersections accordingly produce two P-T boxes for MC140, the one

with the higher grade box spanning from just within the peristerite zone to well into the greenschist field (i.e. $P = 2.7 \pm 0.7$ kb; $T = 390 \pm 50^\circ\text{C}$) and the later event occurring in the very-low-grade range (i.e. $P = 1.65 \pm 0.3$ kb; $T = 280 \pm 25^\circ\text{C}$). The displacement of the two P-T boxes suggest an uplift trajectory which corroborates the assemblage and pressure-plot data and also the mineral chemistry data which show clear signs of retrogression.

Similar signs of chemical retrogression have also been identified for the other Zone D rocks on the western limb of the Onverwacht Anticline:

- a) Actinolite from MC120 is inversely zoned and the pistacite content of epidote is unusually high for greenschist-grades. It has a bimodal distribution unlike the progressively metamorphosed rocks of similar grade and chlorite has an alumina content more characteristic of the greenschist-amphibolite transition.
- b) In MC20, very-low-grade metamorphism is indicated by the presence of stilpnomelane (Winkler, 1979). The very-low-grade is corroborated by coexisting SiO_2 -rich chlorite (Fig. 4.8a,e). The upper thermal limit of very-low-grade metamorphism appears to be ca. $320 - 330^\circ\text{C}$ at pressures ≤ 2 kb (Winkler, 1979, Fig. 1.1). As the flattening fabric of the pillow is overprinted by the stilpnomelane and carbonate assemblage, it is evident that MC20 was exposed subsequent to dynamism to yet a second stage of very-low-grade metamorphism (or metasomatism as large amounts of CO_2 and K_2O have been involved, cf Fig. 3.9 and Viljoen and Viljoen, 1969a, Fig. 5a). The trend of decreasing PT-conditions for the dynamically influenced rocks on the western limb is analogous that found on the eastern limb of the Onverwacht Anticline.

4.6.3 Relationship between Fluid Inclusions and Metamorphism.

4.6.3.1 *Static metamorphism (Zones A and B).* The trapping sequences and the fluid properties of the static-style fluid inclusions show a distinct progression sympathetic to metamorphic grade (summarised in Table 4.1). The drainage-cavity infill of MC150, however, holds the petrographic key to interpreting the fluid trapping history of the statically

metamorphosed rocks. The progressive development of the encrustation- and comb-textured quartz hosts, described earlier, constrain the relative ages of the primary inclusions associated with each quartz type. Inclusion trails crosscutting the grain-boundaries of both quartz generations clearly suggest a late entrapment for the low-eutectic secondary inclusions.

The fact that encrustation quartz was only found in drainage cavity of MC150 suggests a mechanism of preferential preservation. Preservation of the thin quartz lining may have come about by virtue of the drainage cavities being contained within pillows as opposed to the interpillow cavities being externally situated (i.e. between pillows). It is conceivable that the internally situated encrustation sites may have been protected by the drainage cavity wallrock from the harsher hydrothermal environment of the interpillow channelways. It is thus envisaged that the encrustation quartz inclusions represent fluids trapped at an early stage in the metamorphic history of the pillows. From between initial emplacement of the lavas, through early up to late seafloor metamorphism and/or early burial. Complete infilling of the pillow cavities, which had to follow the encrustation quartz, for example, could not have been early seafloor metamorphism or else the observed hydrothermal sequences (described in Chapter 3) would not have developed.

The encrustation inclusions have fluid densities of about 0.8 g/cc. According to the isochore data of Nicholls and Crawford (1985), and the high-temperature seafloor alteration assemblages of the pillows (i.e. 350° - 430°C described in Chapter 3), it is evident that relatively low pressure conditions (≤ 1 kb) are required for trapping fluids of similar density (*cf* Le Bel and Oudin, 1982; Peter and Scott, 1988; Nehlig and Juteau, 1988). Such low pressures are consistent with the interpretation that the encrustation inclusions were formed at an early metamorphic stage.

The inclusion clouds of the columnar grains are common to both Zone A and B and have several features in common: (1) the inclusions are all considered to be primary; (2) they all have similar filling-ratios, and similar densities; (3) they have similar temperatures of final

melting, and thus salinities (Table 4.1); and (iv) they are hosted by columnar quartz. These affinities suggest similar P-T conditions of cavity-filling for both metamorphic zones.

The secondary inclusions have only been developed in the Zone B columnar quartz samples. The healed microfractures have trapped fluids marginally more dense than those of the primary cloud inclusions (filling-ratio's petrographically indistinguishable), but with average salinities being markedly higher, i.e. ca. 13 wt% vs 21 wt% NaCl. The fluids also show on average much reduced apparent eutectica suggesting that the salt species are dominated by divalent cations, such as Ca^{2+} , to the cloud inclusions which in content appear to be NaCl-dominated.

The reason as to why Zone B quartz has both primary and secondary fluid inclusions whereas Zone A essentially only has primaries, appears to be related to the metamorphic grade. Several lines of evidence link the origin of the secondary inclusions to the onset of transition zone grade metamorphism, i.e.:

- 1) The transition zone reactions, outlined above (section 4.6.1), are partial dehydration reactions which would lead to an increased density of the host and volume reductions (cf also Table 4.4 for rock densities). Volume reductions would easily bring about differential strain, effecting the minuscule randomly orientated microfractures of the columnar quartz grains. Some of the expelled fluids could conceivably have occupied the newly formed microfractures after which annealment would lead to secondary entrapment.
- 2) It is only the metamorphic waters of the higher grade transition zone that appear to have had high a_{Ca} as is indicated by the conversion of quartz + albite of the infill domains to oligoclase + quartz (Fig. 4.11). Such compositional conversions (localised metasomatism) suggest high abundances of Ca in the metamorphic fluids which corroborate the low fluid-eutectica of the late secondary inclusions.

If the encrustation quartz inclusions have trapped the high-T, open-system fluids that have affected seafloor metamorphism and the secondary inclusions have resulted from transition

zone metamorphism, where then were the cloud inclusions of the columnar quartz grains formed? Their paragenetic position between the two inclusion types suggest that they must have formed between seafloor- and transition zone metamorphism, i.e. that quartz infilling occurred at greenschist grades and moderate load conditions. Such conditions of entrapment ideally explain the common origin discussed earlier. Their common origin is thus thought to be related to the PT path of Zone B and then of Zone A through normal greenschist grade conditions at medium pressures. By the time that Zone A reached these conditions the underlying Zone B became more deeply buried and was exposed to transition zone conditions where the secondary inclusions were superposed.

The sequence of fluid entrainment in MC150 shows progressive changes with regard to salinity, density and salt species (Table 4.1), which is concordant with the progressive style of static metamorphism. The secondary fluid inclusions represent the end-member of fluid evolution while the transition zone grade of Zone B represents the end-member of progressive static metamorphism.

The parageneses of fluid trapping with respect to metamorphism can be summarised as follows: (1) encrustation quartz inclusions form at low pressure seafloor metamorphism and at greenschist temperatures; (2) columnar-quartz inclusions form at mid- to late burial at medium pressures and greenschist grades; and (3) annealed microfracture inclusions form during late burial at medium pressures and greenschist-amphibolite transition grades.

4.6.3.2 Dynamic metamorphism (Zone D). The dynamic-style metabasites contain fluid inclusions that are petrographically and compositionally different in character to those of the static-style samples. This seems to imply that the precursor static-style fluid signatures have been largely overprinted with the fluid characteristics of the dynamic event(s). Good petrographic control on fluid trapping, especially for MC140, provides the key to interpreting the P-T evolution of dynamic metamorphism (Fig. 4.33). Fluid inclusion data of the dynamic-style rocks are summarised in Table 4.1.

Two generations of fluid inclusion-bearing quartz have been described for MC140, the one clearly a recrystallised product of the other (Fig. 4.22B). The older relict-phase inclusions contain moderate salinity, high-density aqueous- and $\text{H}_2\text{O}-\text{CO}_2$ fluids. The recrystallised matrix appears to have trapped only aqueous fluids with compositions similar to or slightly less saline than those of the relict quartz inclusions. The similarities in composition may have resulted from the re-entrapment of the precursor fluids. However, the recrystallised matrix entrained fluids with marginally lower densities, implying lower pressures (Table 4.1).

Justification for modelling the isochores of the relict quartz- and recrystallised quartz inclusion fluids with the high- and low thermometer temperature-brackets, respectively, is as follows:

- 1) The bulk homogenisation temperatures of the $\text{H}_2\text{O}-\text{CO}_2$ inclusions at ca. 300 - 340°C, which occur only in the older relict quartz, indicate a minimum filling-range and thereby a minimum temperature of formation of the quartz host. The isochores of the cohabiting primary (cloud) aqueous inclusions can therefore only be linked with the higher temperature assemblage of the pillow core and not with those of the pillow rim that have recorded the low thermal event of $280 \pm 25^\circ\text{C}$. The low thermal event is correlated by means of default with the younger recrystallisation quartz inclusions.
- 2) Two P-T boxes with a retrogressive (uplift) relationship are developed (Fig. 4.33). The retrograde relationship is verified independently by the chemistry of the main mineral phases; i.e. chlorite, plagioclase, amphibole and epidote which reveal compositional features reflective of earlier higher grade history.

The fluid inclusions of the twinned quartz of MC120 produced a very restricted cluster of homogenisation readings suggesting fluid entrapment in a physicochemically restricted environment (Fig. 4.29). The thermometer temperature of the pillow is marginally more restrained than that of MC140 and the chemistry of the main mineral phases similarly favour a retrogressive origin. Justification for modelling the thermometer assemblage temperatures with the twinned quartz fluids is summarised in the following:

- 1) Both the geothermometer assemblage phases and the twinned quartz texturally seem to be products of dynamic metamorphism, during which the temperatures and fluids could have been imprinted and trapped, respectively.
- 2) The petrographic features of dynamic metamorphism are not overprinted suggesting that the observed geothermometer and fluid inclusion relationships cannot be incorrectly modelled with those of other metamorphic phenomena.

It is interesting to note that the PT-box of MC120 has a retrograde (uplift) relationship relative to that of MC150 which represents the statically metamorphosed samples along strike to the east (Fig. 4.33). This relationship also corroborates the retrograde mineralogical tendencies determined for the MC120. The retrograde path that is thus formed closely corresponds to that of MC140 on the eastern limb of the Onverwacht Anticline.

4.6.4 Fluid Inclusion Integrity

Fluid inclusion leakages subsequent to trapping and/or changes in the size will alter either the mass and/or volume of the inclusions resulting in erroneous density and isochore determinations. It is well known that geologically and experimentally-induced leakages to fluid inclusions (autodecrepitation) and/or stretching result when the internal pressure of inclusions sufficiently exceeds the confining strength of the host. Such conditions result when the rocks move from high- to low-pressure environments with the condition that the PT path of the rock crosses the isochores from high- to low densities (e.g. Crawford, 1981b; Sterner and Bodnar, 1989). On the other hand, if the confining pressure is much greater than the internal pressure of an inclusion, the host mineral may deform to decrease the volume of the inclusion (Olsen *et al.*, 1988; Sterner and Bodnar, 1989). Experiments of high confining- and low internal pressures show the widespread collapse of inclusions with three dimensional haloes of minute inclusions around each of the disintegrated features. Olsen *et al.*, (1988) shows that high confining pressures also bring about high density inclusions which are associated with "micro-shear" zones and "normal" density ones occur between the shears.

There is one condition, however, that allows for fluid inclusions to preserve their volumetric integrity and that is when they cool isochorically (Sterner and Bodnar, 1989). Such

conditions, fortunately, seem to have been the case for most of the fluid inclusions studied here. The uplift paths of the dynamic-style rocks project toward the origin of the graph which results in a very small angle between them and the isochores. The uplift paths of the statically metamorphosed rocks have not been mineralogically recorded but are directly linked to the uplift path of MC120, as will be elaborated on below. The restricted isochore range of the P_T maxima of all samples (Fig. 4.33) concurs with the uplift paths of the samples, implying that cooling and exhumation occurred isochorically. If a slightly convex path of exhumation is envisaged, as what normally is the case (England and Thompson, 1984), concurrence would be close 100%.

The early inclusions, however, for example those of the drainage-cavity lining, would have been exposed to relatively large (1 - 3 kb) negative internal pressures at peak metamorphic conditions. The relative negative pressures would have resulted from the difference between the low confining pressure (< 1 kbar) of the shallowly trapped encrustation quartz fluids and the high confining pressures (3.9 kb) of the secondary inclusions trapped toward the end of burial metamorphism. None of the typical compression textures, i.e. "micro-shears" or "three dimensional collapse halos" described by Olsen *et al.* (1988) and Sterner and Bodnar (1989) respectively, indicative of excessive confining pressures, were found. The determined density range for the encrustation quartz inclusions is marginally higher than those of high temperature fluids presently convecting in newly formed oceanic-crust. (Peter and Scott, 1988; Nehlig and Juteau, 1988; Vanko, 1988). If corrections were to be made to compensate for any possible compression of the inclusions, they would effectively have to be adjusted toward lower densities implying higher temperatures and lower pressures of entrapment. Modelling toward such environments, however, is soon met with the two-phase field of boiling (Rosenbauer and Bischoff, 1987). Evidence for boiling, however, was not found. Nevertheless, the implications of the early trapped fluids are not crucial to any $P_{T_{max}}$ determinations.

4.6.5 Relationships between Stratigraphic Load and Calculated Pressure

To demonstrate stratigraphic load control on the pressure of metamorphism a test has been

devised in which the petrographic geobarometer and empirically determined pressures can be compared. The minimum, maximum and average pressure ranges of the P-T boxes in Fig. 4.33 are listed according to sample number in the first four columns of the Table 4.3. To limit uncertainties regarding overlying lithologies and to confine calculations to the rocks of the study area, the stratigraphically upper most sample MC210 has been selected as the stratigraphic- and pressure datum (i.e. = 0 kbar). The stratigraphic load pressures of the samples have been calculated from their respective stratigraphic elevations below the datum (second column from the right, Table 4.3). The differences in stratigraphic height between the samples are shown in the right hand column. Due to the subvertical dips of the Komati and Hooggenoeg Formations, the values are obtained by measuring perpendicular, γ across strike to corresponding stratigraphic elevations (cf Figs. 1.4, 4.1). To calculate stratigraphic load, average densities of 2.9 and 3.0 g/cc have been selected for the upper and lower Komati Formation, respectively. These values are based on density measurements obtained from the main rock types of the area (Table 4.4). The metamorphic grade is also listed. It is interesting to note that the higher grade rocks have densities close to 3.0 g/cc whereas the greenschists are just above 2.9 g/cc.

Table 4.3 A comparison of fluid inclusion geobarometer pressures versus pressures determined from stratigraphic elevations.

Sample	P(fluid inclusion geobarometry)	Pressure difference between samples in the left column (datum = MC210 = 0 kb)	Pressure difference according to stratigraphic elevation (datum = MC210; $\rho = 2.9 \text{ g/cm}^3$; $\rho = 3.0 \text{ g/cm}^3$)	Difference in stratigraphic elevation of the samples (datum = MC210)
MC210	$1.9 \pm 0.3 \text{ kb}$	0 kb	0 kb	0 m
MC60	$2.9 \pm 0.5 \text{ kb}$	$1.0 \pm 0.5 \text{ kb}$	1.25 kb	4300 m
MC150	$3.9 \pm 0.6 \text{ kb}$	$2.0 \pm 0.6 \text{ kb}$	1.72 kb (ρ)	5900 m

The differences between the isochore/geobarometer- and stratigraphically (empirically) calculated pressures are evident from Table 4.3. The stratigraphically determined pressures fall well within the indicated error margins of the isochore-geobarometer.

Table 4.4 A list of measured densities of some typical metamaftes of the study area.

Sample	Rock type	Metamorphic grade	Density
MC40	pillowed tholeiitic basalt	greenschist facies	2.93 g/cc
BF1	komatilita with pyx. spinifex	greenschist facies	2.92 g/cc
MC150	pillowed komatilitic basalt	lower-amphibolite facies	3.02 g/cc
SCH1	komatilita with ol. spinifex	lower-amphibolite facies	2.99 g/cc

4.6.6 Relationships between Stratigraphy and Temperature

Stratigraphic control on the thermal structure of the static-style area is revealed by the isopleths of the thermally sensitive minerals plagioclase and amphibole (Fig. 4.4, 4.5). The isopleths tend to lie parallel to the stratigraphy and increase in value with increasing stratigraphic depth, implying increasing grade with depth. Deviations from the parallel relationship in the static-style area are due to retrogression associated with crosscutting faults (*cf* Fig. 1.4). Retromorphism is indicated by isopleths of low value cutting into areas of higher value, i.e. by lower grade phases replacing higher-grade phases. In the retrogressed area further west (i.e. Zones C and D) the isopleths curl southward towards the Komati Fault. From this it is evident that lower grade assemblages developed towards the south more extensively in the western parts than in eastern parts of the area. As the margin of the Theespruit pluton is approached the isopleths of plagioclase and epidote tend to reflect the contact whereas chlorite and amphibole do not.

The relationships between stratigraphy and P and T of metamorphism of the statically metamorphosed area support the following conclusions:

- 1) The pressure of metamorphism is controlled by stratigraphic load (opposed to directed or tectonic pressure).
- 2) The good correspondence between the isochore-barometer and stratigraphically determined pressures suggest that the P-T boxes of Fig. 4.33 and the method itself are credible.
- 3) The (total) pressures recorded by the samples close to the fault-bound stratigraphic base of the Komati Formation (3.9 ± 0.6 kb) equate with a lithological thickness of

ca. 14 ± 2 km, which corresponds to the total thickness of the Swaziland Supergroup (above the Komati Fault) as presented by Lowe *et al.* (1985). However, Lowe (1991) presents a revised thickness of ca. 10 km for the same column. It is not clear whether or not the reduced thickness compensates for the known structural duplication of strata between the upper Onverwacht- and lower Fig Tree Groups (cf de Wit *et al.*, 1983; Lowe *et al.*, 1985; Lamb and Paris, 1988). The metamorphically recorded pressures, nevertheless, include (syndimentary) tectonic loading as will be discussed in the following chapter.

- 4) Importantly, the results show that the tectono-stratigraphic column was metamorphosed prior to being rotated to its present flat-lying position. This is deduced from the fact that both thermal and pressure gradients, which increase with increasing stratigraphic depth are transposed to the horizontal, (i.e. increase toward the south).
- 5) The minimal resetting of the transposed assemblages of the static-style area constrains the intensity of compressional forces responsible for isoclinal folding. The intensity of the tectonic forces must have been substantially less than those responsible for isoclinal folding in Phanerozoic orogenic belts where pervasive penetrative fabrics occur. In the study area such fabrics are only locally developed as is the case for Zone D. The fact that the static-style area attained pressures of ≤ 4 kb implies that it would have deformed plastically had significant compressional forces been present.

4.6.7 Burial Metamorphism

Several features of the statically-metamorphosed parts of the Komati and Hooggenoeg Formations reveal that post-seafloor metamorphism was followed by an extensive period of burial metamorphism dominated by stratigraphic loading of which a significant portion may also have been contributed by syndepositional tectonic imbrication. The following features strongly support the widespread presence of static burial metamorphism for the southwestern parts of the Barberton greenstone belt:

- 1) Isotherms trend parallel to the stratigraphy.

- 2) Metamorphic grade increases with stratigraphic depth.
- 3) Fluid inclusion data show a gradual progression from very-low to medium-pressure conditions for fluid entrapment.
- 4) Pressures of metamorphism equate with the tectonostratigraphic load of the area.
- 5) The excellent preservation of protolith structure and texture suggests a non-directional or hydrostatic stress environment.

4.6.8 P-T Paths

It has long been known that maximum PT conditions obtained from mineral assemblages are of restricted value and cannot directly be used, for example, to produce the ambient geothermal gradient of metamorphism (e.g. England and Thompson, 1984). This has led to an advancement in metamorphic petrology from merely calculating 'peak' metamorphic conditions towards deriving PTt paths from rocks and relating metamorphic conditions to tectonic processes (Selverstone and Spear, 1985; Hodges, 1991).

The lack of accurate mineral geothermobarometers for metabasites of medium- to low pressures precludes the determination of complete P-T paths for the study area. The method devised here for constructing the burial paths of the metabasites makes use of the volcanological environments of emplacement (discussed in Chapter 2) to constrain the starting points of the paths. The path starting points are then connected to the determined maximum PT conditions by tie-lines which represent the burial vectors of the samples. The actual burial paths, however, are more likely to be slightly convexly curved toward the T-axis than straight (Robinson, 1987). Such (downwardly curved) paths are indicative of thermal relaxation and decreasing rates of burial as will be shown below.

4.6.8.1 *Statically metamorphosed rocks.* The P-T boxes of MC210 and MC60 (Zone A) are plotted in Figures 4.34a,b together with the relevant isochore and geothermometer temperature data from which they were derived (initially omitted from Fig. 4.33 for clarity). The conditions of emplacement for MC210 (as discussed in Chapter 2) ranged between 0 - 70 bar and 60 - 150°C.