

**The Mavuradonha Layered Complex:
Neoproterozoic emplacement and
Pan-African granulite-facies metamorphism in the
Zambezi Allochthonous Terrane of the Mt. Darwin Area,
Zambezi belt, NE-Zimbabwe**

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All views and results presented in this thesis are those of the author, unless stated otherwise.

Ich versichere, dass ich die vorliegende Arbeit selbständig und nur unter Verwendung der angegebenen Quellen und Hilfsmittel verfasst habe.

Mühlhausen, den 30. Januar 2004

And I think to myself, what a wonderful world

L. Armstrong



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Abstract

An understanding of the geological evolution of the Zambezi belt is important for the geodynamic evolution of the Neoproterozoic network of orogenic belts in West-Gondwana. A detailed petrological, geochronological, geochemical and isotopic characterisation of the Mavuradonha Layered Complex in northern Zimbabwe has been carried out in order to contribute to a better understanding of the geological evolution of the NE-part of the Zambezi belt.

The Mavuradonha Layered Complex represents a lower crustal complex which was generated during an early Pan-African extensional event in the northeastern Zambezi belt. Emplacement of the complex was dated at 862 ± 4 Ma using SHRIMP and vapour transfer methods on zircons. This magmatic emplacement age is also supported by a Sm-Nd whole rock errorchron age of 899 ± 46 Ma. Additional age information was obtained on zircons of the Ocellar Gneiss, which structurally underlies the Mavuradonha Layered Complex. These zircons reveal a Kibaran protolith age of 1022 ± 7 Ma.

Within the Mavuradonha Layered Complex a multiphase magmatic differentiation is recorded in macro rhythmic units and small-scale layering. This multiphase differentiation led to the formation of magmatic sequences comprising pyroxenites, gabbro/norites, leuco-gabbros within the **layered gabbro series**. Additionally, large amounts of ferro-gabbros are associated with anorthosites in the **lower meta-anorthosite suite** and the **upper meta-anorthosite suite**.

It is also shown that there is evidence in the isotopic record for substantially depleted mantle values in the NE-Zambezi belt during the Neoproterozoic. Crustal contamination of the subalkaline, tholeiitic parental magma is indicated by variable ϵ_{Nd} values between + 0.3 and + 6.6, as well as abundant xenoliths and inherited zircons in the layered gabbro series of the Mavuradonha Layered Complex.

The Mavuradonha Layered Complex was overprinted by granulite-facies metamorphism with peak-conditions of 13 ± 2 kbar at 840 ± 30 °C during the Pan-African orogeny at about 554 ± 13 Ma. This age of metamorphism was estimated from metamorphic overgrowth on zircon using the SHRIMP method. After the granulite-facies peak, the complex was retrogressed under amphibolite-facies conditions at 11 ± 2 kbar and 680 ± 20 °C, and, locally further retrogression under lower amphibolite- to greenschist-facies conditions took place. The cooling event was accompanied by infiltration of highly enriched saline fluids as indicated by the abundance of late, chlorine-rich scapolites, amphiboles and corona-textures in the metagabbros. Retrogression occurred at about 546 ± 9 Ma and is constrained by Sm-Nd garnet-whole rock ages. Further cooling to greenschist-facies conditions occurred within the following 50 Ma as constrained by U-Pb

rutile ages ranging from 526 ± 7 to 501 ± 6 Ma. Taking into account the *PT* data, a *PT* path is suggested that resulted from crustal thickening related to plate convergence.

The parental magma of the metagabbroic bodies that occur as inliers in the Mt. Darwin area also have a subalkaline, continental tholeiitic character. These inlier metagabbros have a geochemical composition that is different from the Mavuradonha Layered Complex metagabbros. They are slightly enriched in incompatible and rare earth elements. The isotope characteristics of these metagabbros also imply a crustal setting for the inlier metagabbros, but probably with higher degrees of contamination by continental crust (ϵ_{Nd} ranging from + 2.4 to – 3.5) than is recognised for the Mavuradonha Layered Complex.

The data presented in this study suggest that the evolution of the Zambezi belt in NE-Zimbabwe began as a failed rift or an intracratonic basin. The results of the geochronological studies imply no genetic link between intrusion of the mafic magmas and high-temperature granulite-facies metamorphism. The age data show that granulite-facies regional metamorphism occurred at least 300 Ma after the emplacement of the mafic rocks. Therefore, emplacement of the Mavuradonha Layered Complex and the inlier gabbros represents an episode of intense plutonic activity during an early stage of rifting. The basin or failed rift was closed during the late Pan-African orogeny.

Zusammenfassung

Die paläotektonische Stellung und die geologische Entwicklung des Sambesi-Orogens als Bestandteil der neoproterozoischen Gebirgszüge in West-Gondwanaland ist für die Konsolidation von Gondwana während des späten Proterozoikums von besonderer Bedeutung. Trotz seiner zentralen Stellung in diesem kontinentüberspannenden Gebirgszug ist das Sambesi-Orogen weitgehend unerforscht. Die Korrelationen zu anderen Gebirgszügen im heutigen südlichen Afrika wie Lufilian Arc und Damara-Orogens im Westen und Mosambik-Orogens im Osten, werden daher weiterhin kontrovers diskutiert.

Im allgemeinen wird das Sambesi-Orogen als 830 Ma alter Falten- und Überschiebungsgürtel interpretiert. Im nordöstlichen Teil des Orogens ist eine komplexe tektono-metamorphe Entwicklung erkennbar, die aus Krustenverdickung und aus der Exhumierung von tief-krustalen Gesteinen sowie aus südgerichteten Überschiebungen in Richtung des Simbabwe-Kratons besteht. Im allgemeinen wird der nordöstliche Teil des Zambezi-Gürtels in drei tektono-stratigraphische Einheiten unterteilt:

- (1) das Migmatitic Gneiss Terrain
- (2) das Marginal Gneiss Terrain
- (3) das Zambezi Allochthonous Terrain

Das Zambezi Allochthonous Terrain wiederum setzt sich aus granitoiden Gneisen sowie in inliern aufgeschlossen Metagabbros und dem Mavuradonha Layered Complex, bestehend aus Amphiboliten, Metagabbros und Meta-Anorthositen, zusammen.

Gegenstand der vorliegenden Arbeit ist es, die paläotektonische Stellung und geologische Entwicklung des Sambesi-Orogens im NE Simbabwes herauszustellen. Dies erfolgt insbesondere durch eine detaillierte petrologische, geochemische und geochronologische Charakterisierung des Mavuradonha Layered Complex und von Metagabbros, die in Inliern aufgeschlossen sind. Der Mavuradonha Layered Complex repräsentiert eine mafische Intrusion in Unter- bis Mittelkrustenbereichen, deren Bildung im frühen Pan-African mit einem extensionalen Ereignis in Zusammenhang steht. Die heutige Stellung des Komplexes und der untersuchten Metagabbros als Deckenstapel auf granitoiden Gneisen wurde durch südwärts-gerichtete Überschiebungen bedingt, die in Verbindung zur Kompressionstektonik im späten Proterozoikum stehen.

Geochronologische Studien, die mit einer Kathodolumineszenzstudie kombiniert wurden, verdeutlichen die komplizierte Entwicklungsgeschichte innerhalb des Zambezi Allochthonous Terrain. Das Alter der Intrusion konnte mit der SHRIMP-Ionensonde auf 862 ± 4 Ma bestimmt werden und wurde an zwei Proben mit der U-Pb-Einzelzirkon-Datierungsmethode bestätigt. Dieses Alter wird ebenfalls durch ein Sm-Nd-Errorchronen-

Alter von 899 ± 46 Ma untermauert. Demgegenüber steht das Protolithalter des unterlagernden Ocellar Gneises von 1022 ± 7 Ma, das auf ein kibarisches Ereignis in diesem Teil des Sambesi-Orogens hinweist.

Innerhalb des Komplexes ist eine mehrphasige magmatische Differentiation, mit Klinopyroxen und Plagioklas als Hauptkumulatphasen erkennbar. Im Komplex sind neben einem makro-rhythmischen Lagenbau auch ein kleinmaßstäblicher Lagenbau im cm-Bereich erkennbar, die hauptsächlich in den Gesteinsabfolgen Pyroxenit, Gabbro/Norit, Leuco-Gabbro und Anorthosit der **layered gabbro series** auftreten. Ferro-Gabbros treten in enger Relation zu den Anorthositen der **lower meta-anorthosite suite** und der **upper meta-anorthosite suite** auf.

Ein weiteres bedeutendes Resultat aus dieser Studie für die geologische Entwicklung des Sambesi-Orogens ist der isotopisch verarmte Charakter der Mavuradonha Layered Complex- und Metagabbro-Proben. Eine Krustenkontamination der Magmen, die einen subalkalinen, tholeiitischen Charakter aufweisen, wird durch die variablen ϵ_{Nd} -Werte (+ 6.6 bis -3.5) angezeigt. Der isotopische Charakter der hochdifferenzierten Gesteine, das reichliche Auftreten von Xenolithen in der layered gabbro series sowie ererbte Zirkone sind eindeutige Beweise für eine Krustenkontamination und folglich für ein krustales Bildungsmilieu.

Der Mavuradonha Layered Complex wurde durch eine hochgradige, granulit-fazielle Metamorphose mit der Mineralparagenese Grt-Cpx-Pl-Qtz überprägt. Druck- und Temperaturberechnungen ergaben metamorphe Höchstbedingungen von 13 ± 2 kbar bei 840 ± 30 °C. Retrograde Bedingungen von 11 ± 2 kbar bei 680 ± 20 °C werden durch Amphibolite sowie durch retrograde Zonierungen im Granat und durch das Auftreten von retrograden Amphibolen dokumentiert. Die metamorphe Überprägung des Komplexes ereignete sich im späten Neoproterozoikum um 550 Ma. Für die metamorphe Überprägung wurde mit SHRIMP-Altersdaten an metamorphen Anwachs säumen um Zirkone ein Alter von 554 ± 13 Ma bestimmt. Die retrograde Metamorphose konnte mittels einer Granat-Gesamtgesteins-Isochrone auf ein Alter von 546 ± 9 Ma datiert werden. Eine weitere Abkühlung unter amphibolit- bis grünschieferfaziellen Bedingungen erfolgte in einem kurzen Zeitintervall von 50 Ma nach der Peak-Metamorphose (U-Pb Abkühlungsalter von 526 bis 501 Ma), bis die Schließungstemperatur des Rutil (420 $\pm$ 30 °C) erreicht wurde. Die Abkühlung wurde von der Infiltration von salz-angereicherten Fluidphasen begleitet, die durch das Auftreten von sekundären, chloridreichen Skapolithen und Amphibolen dokumentiert sind.

Die in dieser Arbeit präsentierten Ergebnisse lassen darauf schließen, dass die Metagabbros und der Mavuradonha Layered Complex tief- bis mittelkrustale Komplexe darstellen, die während eines frühen Pan-Afrikanischen Extensionsereignisses im

nordöstlichen Sambesi-Orogen intrudiert sind. Demgegenüber zeigt das Alter, das für die Metamorphose in den Mavuradonha-Bergen bestimmt wurde, eine Ursache, die in enger Verbindung zur endgültigen Konsolidation von Gondwana steht. Die Altersdatierungen beweisen aber auch, dass die granulitfazielle Regionalmetamorphose, die sich durch das massenhafte Auftreten von Granat in den Metagabbros und Meta-Anorthositen auszeichnet, mindestens 300 Ma nach der Intrusion des Mavuradonha Layered Complex erfolgte. Diese Resultate zeigen somit keinen genetischen Zusammenhang zwischen der Bildung der mafischen Magmen und der Hochtemperatur-Granulitfazies an. Daher kann eine *PT* Entwicklung für diesen Teil des Sambesi-Gürtels angenommen werden, wie sie vielfach für andere Gebiete mit Krustenverdickung an konvergierenden Plattengrenzen beschrieben wurden.

Aufgrund der Ergebnisse dieser Studie wird für das Sambesi-Orogen und die Entwicklung des Zambezi Allochthonous Terrain im Bereich von Mt. Darwin in NE-Simbabwe folgendes schematisches Drei-Stufen-Modell vorgeschlagen:

- (1) Um 1020 Ma erfolgte die Bildung von großen granitoiden Körpern.
- (2) In einer Phase intensiver plutonischer Aktivität während eines frühen Rift-Stadiums um 860 Ma erfolgte die Platznahme des Mavuradonha Layered Complex und die weiterer gabbroischer Komplexe in unter- bis mittelkrustalen Bereichen.
- (3) Im Verlauf der späteren Pan-Afrikanischen Orogenese und in enger Verbindung zur Bildung von Gondwana wurden dieses fehlgeschlagenen Rifts oder Becken in Folge von Komprimierungsprozessen um 550 Ma wieder geschlossen und unter granulitfaziellen Bedingungen überprägt.

1 Introduction

1.1 Geological evolution of Gondwana

The Gondwana supercontinent was finally assembled during the Neoproterozoic from East- and West-Gondwana along a system of Pan-African mobile belts, in particular the Mozambique belt or East African Orogen (Pinna et al., 1993; Stern, 1994; Shackleton, 1996). The suture between East-Gondwana, consisting of Antarctica, Australia and Madagascar, and West-Gondwana, which comprises South America and Africa, has been discussed by Shackleton (1996). An earlier proposal linked the suture between East and West-Gondwana in the southern Gondwana supercontinent to the transcontinental network of mobile belts consisting of the Mozambique belt with a continuation in the Zambezi belt, the Lufilian Arc and the Damara belt (Unrug, 1983; Hoffman, 1991). In contrast to this interpretation, this transcontinental orogenic belt has also been considered to reflect a chain of several intracratonic basins and rifts (Daly 1986; Porada 1989). These basins or rifts were initiated at about ~1 Ga in the Mesoproterozoic, and the closure is linked to the final assembly of Gondwana (Stern, 1994; Porada & Berhorst, 2000; John, 2001). The amalgamation of Gondwana is thus correlated with the collapse of the former supercontinent Rodinia. The final break-up of this supercontinent is marked by the formation of the so-called Mozambique Ocean at around 800-850 Ma (Hoffman, 1991). The assembly of crustal fragments to form Gondwana is linked to a fan-like closure of the Mozambique Ocean, with a pivot in South Africa (Hoffman, 1991), and culminating in a major collisional event around 550 Ma (Stern, 1994; Kröner et al., 1999; 2001). The consolidation of Gondwanaland is recorded in northeastern Africa by numerous ophiolites (Zimmer, 1985, Zimmer et al., 1995; Reischmann, 2000 and references therein) and by a large volume of juvenile crust (e.g. Reischmann, 1986; Kröner et al., 1991). The apparent lack of ophiolites and large mafic to ultramafic complexes in southern Gondwana indicates rifting of intracontinental basins, which did not open into large oceanic basins (Daly, 1986; Porada, 1989; Hoffman, 1991).

The geotectonic setting in the Neoproterozoic of southern and central Africa (Figure 1-1) is characterised by large, stable cratons including the Congo, Kalahari, Tanzania and Zimbabwe Cratons and their surrounding orogenic belts. These orogenic belts were formed during the Kibaran orogeny, such as the Irumide belt and the Namaqua-Natal belt. The cratons and the Palaeo- to Mesoproterozoic orogens were separated and partly influenced by a Pan-African transcontinental network of orogenic belts of varying ages. In this network of orogenic belts consisting of the West Congo belt, the Damara belt,

the Lufilian Arc, the Zambezi belt and the Mozambique belt (Kröner, 1977a; Pinna et al., 1993; Porada & Berhorst, 2000), mostly Pan-African magmatic and metamorphic events are recorded. They reflect magmatism and deposition, followed by major orogenic activity which occurred between 950 to 450 Ma (Kröner, 1984).

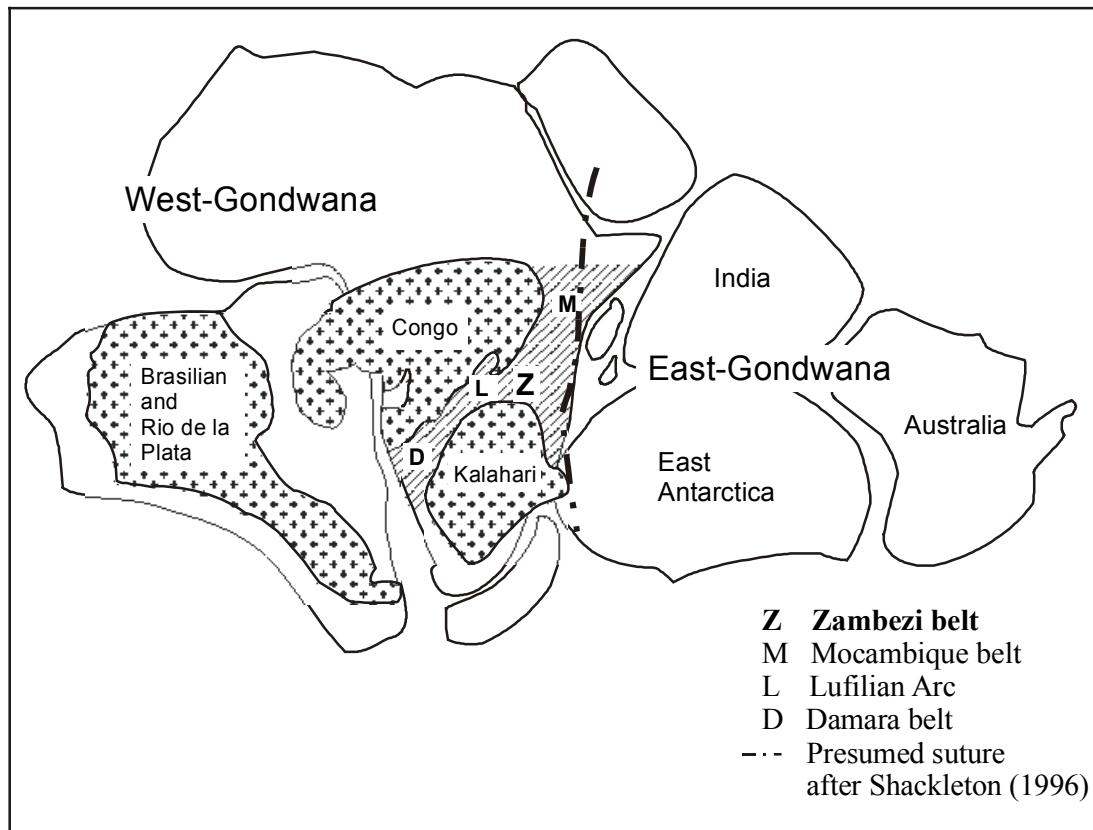


Figure 1-1. Geotectonic setting of Gondwana in the late Proterozoic. Modified after Unrug (1983) with one possible suture as discussed by Shackleton (1996).

1.2 The Zambezi belt

1.2.1 General geology

The Zambezi belt represents one of the least known parts within the Gondwana framework despite its central geological position in Gondwana. Therefore, the geological evolution of the Zambezi belt and the still controversially debated links to the Mozambique belt, the Lufilian Arc and the Damara belt are important for the Pan-African development of southern Africa.

The Zambezi belt (Figure 1-2) is continuous into the Mozambique belt in the East, where these two belts merge across a triple junction (Barton et al., 1991; 1993; Pinna et al.,

1993). In the West, the Mwembeshi Dislocation, a major sinistral transcurrent shear zone, separates the Zambezi belt from the Lufilian Arc (de Swardt et al., 1965; Hanson et al., 1993; Wilson et al., 1993). Across the Mwembeshi Dislocation the Lufilian Arc exhibits distinct differences in structural and metamorphic history (Hanson et al., 1993) and continues under the Phanerozoic cover in Zambia into the Damara belt (Hartnady et al., 1985). This continuation to the West and a synchronous evolution of the Zambezi belt with the other belts in the west is questioned (Hanson et al., 1988a) since it has been shown that the evolution of this intracontinental orogen was diachronous. Initial rifting occurred around 870 Ma in the Zambezi belt (Hanson et al., 1988a) but more than 100 Ma later at ~750 Ma in the Damara belt (Miller, 1983). Whereas thrusting in the Zambezi belt began around 820 Ma (Hanson et al., 1988a), the first thrusting in the Damara belt occurred around 550 Ma (Haack et al., 1983; Kröner, 1982; Hawkesworth et al., 1983).

Several geodynamic models emphasize the evolution of this major transcontinental network of mobile belts, but the position of the Zambezi belt within this framework remains unclear. Based on the interpretation of the Zambezi belt as an 830 Ma-old fold and thrust belt (Barton et al., 1991; Hanson et al., 1994; Wilson et al., 1993; 1997) two models comprise the following aspects:

- a) The evolution of the Zambezi belt was caused by a continent-continent collision due to the closure of an oceanic basin between the Congo and Kalahari Cratons (Coward and Daly, 1984; Daly, 1986), followed by the formation and exhumation of eclogites and white schists (Vrána et al., 1975; John et al., 2000; John, 2001).
- b) The Zambezi belt evolved as an intracratonic rift basin which was filled with shallow-water sediments and bimodal volcanic rocks at around 880 Ma (Hanson et al. 1988a; Wilson et al., 1993; Munyanyiwa et al., 1997). Closure of the basin occurred during a major orogenic event around 820 Ma and was dated on zircons of the syntectonic Ngoma Gneiss (Hanson et al., 1988a) and the Lusaka granite (846 ± 68 Ma; Barr et al., 1977).

The first model is questionable although locally high-pressure rocks with MORB geochemical signatures (Vrána, 1975; Dirks et al., 1997; Goscombe et al., 1997; 2000; Dirks & Sithole, 1999; John et al., 2000; John et al., 2001; John, 2001), an ophiolite (Oliver et al., 1998) and eclogites (Dirks & Sithole, 1999; John et al., 2001; John, 2001) are exposed in the western part of the belt. However, no clear evidence was presented for the existence of a major suture zone along the Zambezi belt-Lufilian Arc-Damara orogen.

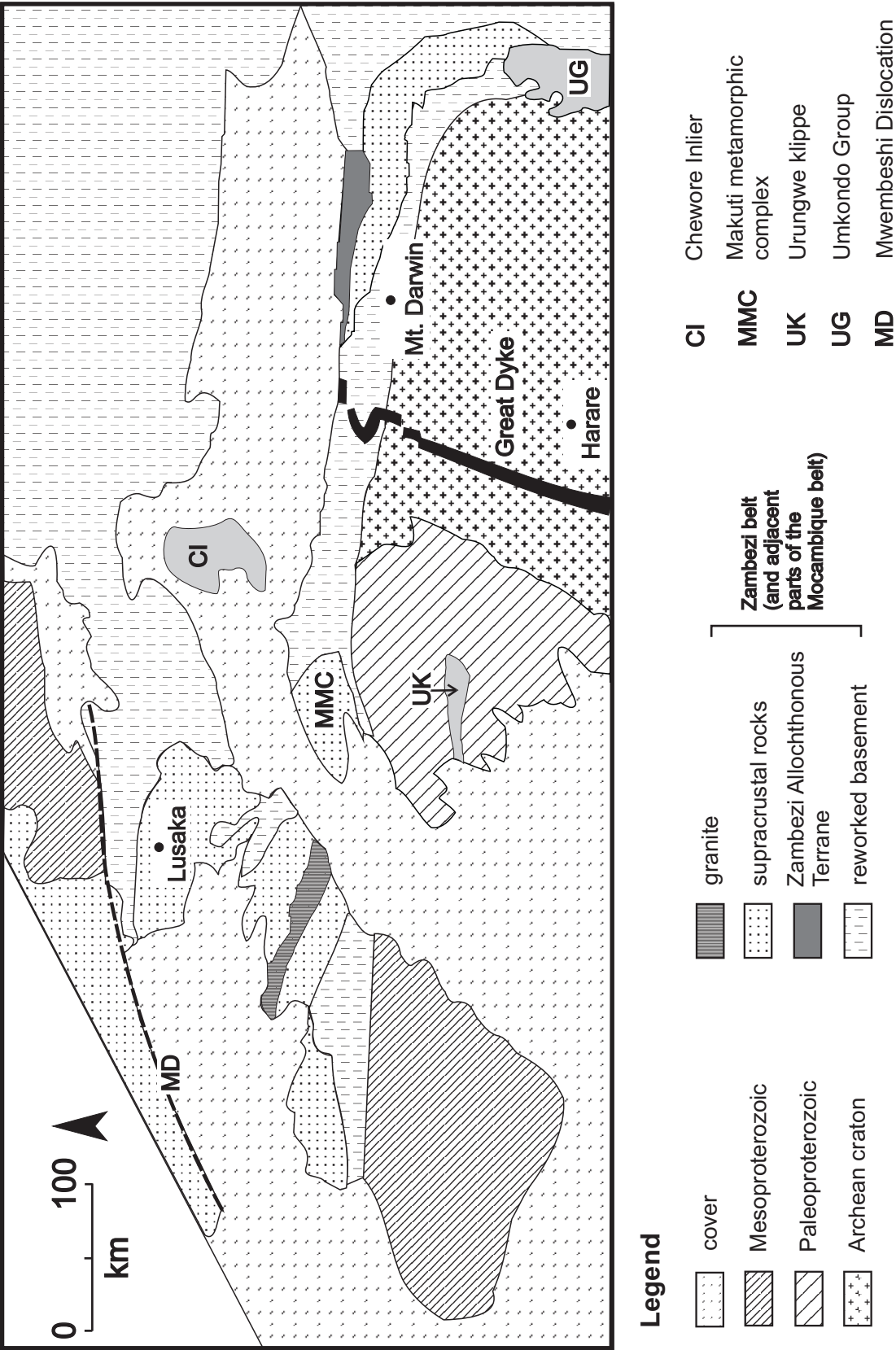


Fig. 1-2. The Zambezi belt of Zimbabwe and Zambia; modified after Hanson et al. (1994).

Moreover, the structural trends of the Irumide belt rather support the interpretation that the Zambezi belt evolved during the closure of an intracontinental basin (Hanson et al., 1988b; Hanson, 2003; Kröner, 1977b; 1983) than by closure of a wide ocean, since similar structures can be found on both sides of the Zambezi belt. In addition, the Zambezi belt contains extensive tracts of remobilised sialic basement which are structurally overlain by large masses of supracrustal rocks (Leitner & Phaup, 1974; Wilson et al., 1993; Hanson et al., 1993). However, it should be considered that the direct correlation between Irumide belt and Choma-Kalomo block was recently discounted on the basis of new SHRIMP zircon ages that indicate different histories of these older terranes (DeWaele et al., 2003). The Mwembeshi dislocation, often described as a major shear zone, did not have much influence on the formation of the transcontinental mobile belt (Katongo & Tembo, 1999; Porada & Berhorst, 2000). Moreover, the Mwembeshi dislocation is interpreted as a remobilised shear-zone predating the Pan-African orogeny. These arguments led to the suggestion that the Zambezi belt evolved as a narrow intracratonic basin which was filled with shallow-water sediments and bimodal volcanic rocks (Wilson et al., 1993; Munyanyiwa et al., 1997). These bimodal volcanic rocks (Leitner & Phaup, 1974; Wilson et al., 1993; Hanson et al., 1993; Munyanyiwa et al., 1997) reflect initial extension-related magmatism. Daly (1986) proposed a linked array of different aulacogens, pull-apart basins and shallow oceanic basins for the development of the Zambezi belt. These basins were closed at different times in the Neoproterozoic. As described in Porada & Berhorst (2000) several small-scale basins evolved into larger basins on the western flank of Rodinia, partly with an oceanic character as formed in the Lufilian Arc and partly remaining as intracratonic basins as in the Zambezi belt and in the Damara belt.

Because of the presence of ophiolites, eclogites and other high-pressure rocks, the model suggesting a Red Sea type setting is another possible explanation for the Pan-African geotectonic setting in southern Africa (Porada, 1989). In this model, two rifts evolved into an ocean, whereas the third branch remained as an intracratonic failed rift. This failed rift may be represented in the eastern part of the Zambezi belt in Zimbabwe, whereas in the Zambezi belt of Zambia, the Lufilian Arc and the Damara belt the rifts evolved into oceans.

Dirks et al. (1998) suggested a model for the evolution of the Zambezi belt, comprising two major tectono-thermal events based on structural analysis in NE-Zimbabwe. In their interpretation an extensional event occurred between 900 and 800 Ma, accompanied by extension-related magmatism. More than 300 Ma later compressional tectonics led to the formation of the Pan-African Zambezi belt at around 500 Ma. These two major events are corroborated by recently available age data (Hahn et al., 1991; Goscombe et al., 1997; 2000; Vinyu et al., 1997; 1999).

Generally, the Zambezi belt exhibits ductile deformation and metamorphism reaching upper amphibolite- to granulite-facies conditions (Treloar et al., 1990; Carney et al., 1991; Hargrove et al., 1998; 2003), and locally eclogite-facies mineral assemblages in gabbroic rocks (Dirks & Sithole, 1999; John, 2001), are preserved in the belt. Thrusting to the south onto the Zimbabwe Craton was accompanied by the exhumation of deep crustal rocks in the northeastern Zambezi belt (Treloar et al., 1990; Barton et al., 1991; Carney et al., 1991). Transcurrent shearing was parallel to the trend of the belt (Barton et al., 1985; 1991; 1993; Hanson et al., 1993), and back-folding and back-thrusting occurred during a compressional event (Barton et al., 1991; 1993).

1.2.2 The Zambezi belt in NE-Zimbabwe

The Zambezi belt in NE-Zimbabwe (Fig. 1-3) consists of metasedimentary sequences that were deposited on a reworked Archaean and Palaeoproterozoic sialic basement, as well as felsic and mafic intrusive rocks (Barton et al., 1991). The belt is subdivided into several major tectonostratigraphic units (terranes) which exhibit different degrees of tectono-thermal rejuvenation during the Pan-African orogeny. The following description is based on the compilations of Leitner & Phaup (1974), Bache et al. (1983; 1984) and Barton et al. (1985; 1991; 1993):

In the southern part the belt is composed of the Migmatitic Gneiss Terrain which is a zone of reworked Archaean basement that marks the contact between the Zambezi belt and the Zimbabwe Craton. It consists mainly of Archaean tonalitic gneisses, mica schists, and metasedimentary and metabasic inclusions that are infolded with the migmatitic gneisses. This terrain generally exhibits amphibolite-facies metamorphism and is complexly deformed. Garnet-orthopyroxene bearing mineral assemblages locally indicate granulite-facies conditions.

The Basal Rushinga Intrusive Complex, a tabular-shaped granitoid body, marks the interface between the Migmatitic Gneiss Terrain and the Marginal Gneiss Terrain. A whole-rock Rb-Sr errorchron age of 829 ± 78 Ma (Barton et al., 1991) and an U-Pb age on zircons of 805 ± 11 Ma (Vinyu et al., 1999) date this early Neoproterozoic magmatic event.

The Marginal Gneiss Terrain structurally overlies the Migmatitic Gneiss Terrain and consists of metasedimentary sequences of the Rushinga Metamorphic Suite and quartzo-feldspatic and biotite-rich gneisses of the Chimanda Metamorphic Suite.

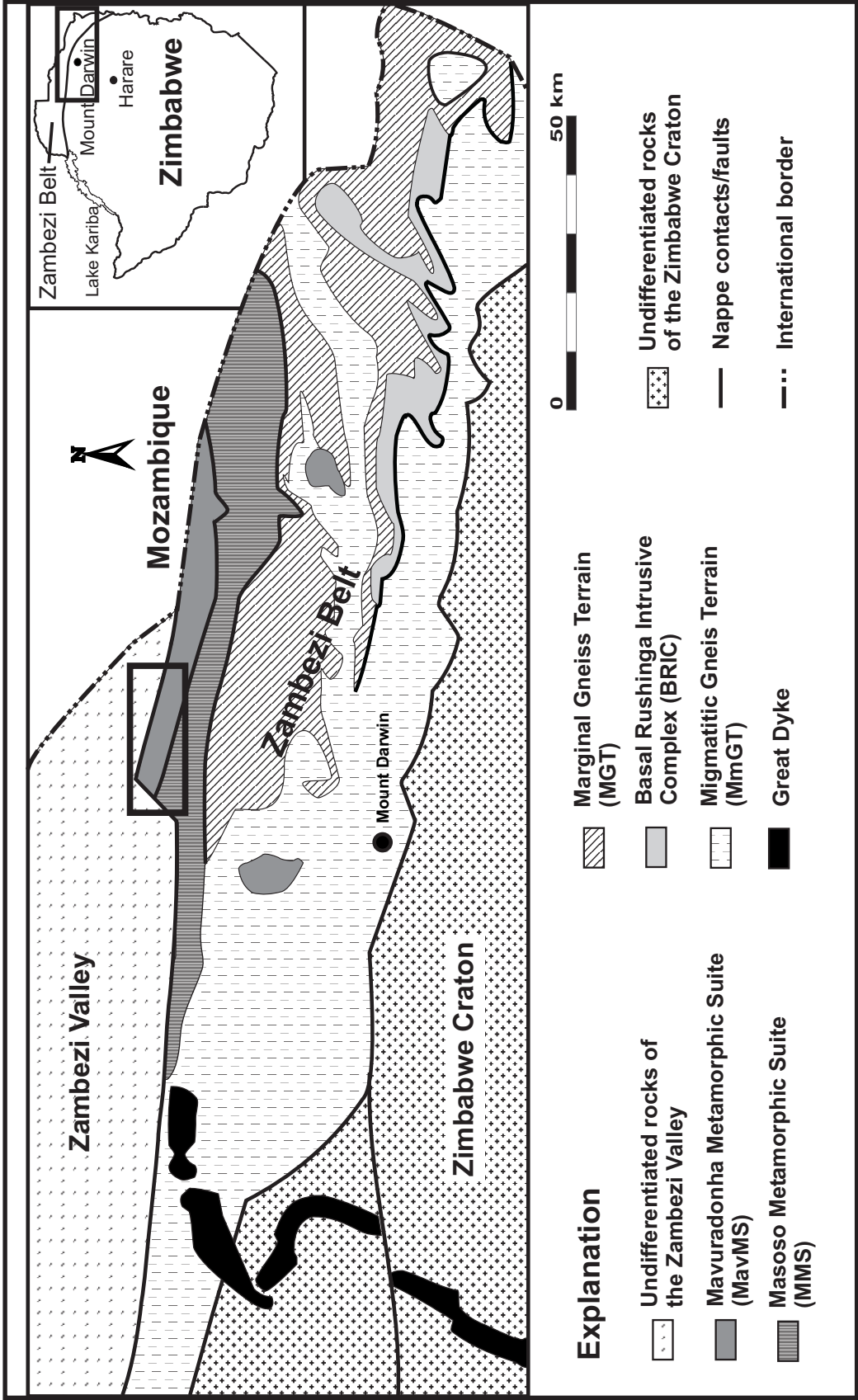


Figure 1-3. Geology of the Zambezi Belt in NE-Zimbabwe after Barton et al. (1991). The box on the map marks the study area in the Mavuradonha Mountains (Fig. 2-1). The inset shows the position of the Zambezi Belt in NE-Zimbabwe and the box marks that part of the belt shown in detail on the map.

The Rushinga Metamorphic Suite is interpreted as a sequence of shallow water sediments representing sedimentation along the margin of the Zimbabwe Craton. This strongly deformed sequence shows different metamorphic grades ranging from amphibolite-facies to granulite-facies. The Chimanda Metamorphic Suite consists of reworked basement and supracrustal rocks.

The Zambezi Allochthonous Terrain is the structurally highest unit in the north of the Zambezi belt. This terrain is bordered in the north by the Zambezi Escarpment and to the south by the Masoso thrust of the Rushinga Metamorphic Suite. The Zambezi Allochthonous Terrain is interpreted as a thrust pile composed of felsic and mafic intrusive rocks of the Masoso Metamorphic Suite and Mavuradonha Metamorphic Suite which were thrust onto the Marginal Gneiss Terrain. The Zambezi Allochthonous Terrain contains rocks of the highest grade of metamorphism that were found in this part of the Zambezi belt and consists of garnet- and orthopyroxene-bearing granulites.

The Masoso Metamorphic Suite forms the basal unit of the Zambezi Allochthonous Terrain and consists of a bimodal suite of leucomigmatites and striped mafic gneisses and metaplutonic rocks of amphibolite-facies grade. Locally, within mafic horizons garnet-bearing granulites are observed. Some of the Masoso rocks show geochemical affinities to the Basal Rushinga Intrusive Complex.

The Mavuradonha Metamorphic Suite is the structurally highest unit in the northeastern part of Zambezi belt and was thrust onto the Masoso Metamorphic Suite along the Mavuradonha thrust. The Mavuradonha Metamorphic Suite was overprinted under granulite-grade peak metamorphic conditions, indicating the highest metamorphic conditions in this part of the Zambezi belt. At the base it is composed of the Nyamasoto gneiss (Barton et al., 1991) and the Ocellar gneiss (Bache et al., 1983). Granulite-facies metagabbros, which are retrogressed under amphibolite-facies conditions, rest structurally on top of the gneisses. Gabbroic to anorthositic garnet- and clinopyroxene-bearing granulites are exposed next to the Zambezi Escarpment. The base of the Mavuradonha Metamorphic Suite exhibits large bodies of garnet-free amphibolites and garnet-bearing amphibolites that locally preserve gabbroic textures.

Three major deformation events were observed in the Zambezi belt of NE Zimbabwe, which are referred to as the DZM 1 to DZM 3 (**D**eformation episodes **Z**ambezi-**M**ozambique) by Barton et al. (1991; 1993).

DZM 1 refers to the planar compositional layering and is presumed to be the result of deep crustal extension of thickened crust due to underplating. This event is restricted to the Masoso Metamorphic Suite and the Mavuradonha Metamorphic Suite. Emplacement of allochthonous terranes of the Zambezi Allochthonous Terrain onto the craton is correlated with the DZM 2 episode. During this event, the Marginal Gneiss Terrain was

metamorphosed under high-pressure granulite-facies conditions of ~ 800 °C and 12 kbar (Treloar et al., 1990; Carney et al., 1991). This event was accompanied by exhumation of deep crustal granulites. DZM 3 indicates infolding and north-directed overthrusting which is related to compressional tectonics and the closure of a basin.

These deformational episodes have recently been revised by Dirks et al. (1998). The authors divided the deformational episodes into three groups. The D1 event proposed by Dirks et al. (1998) represents structures of Archaean age which are restricted to the Archaean basement. D2 deformation is supposed to be Pan-African structures indicating E- to NNE-directed shearing that resulted in the formation of isoclinal, recumbent and upright fold geometries that affected the entire Zambezi belt. Structures related to D2 are inferred to be coeval with the emplacement of the Basal Rushinga Intrusive Complex, and mafic dykes which are suggested to be related to extensional processes. D3 structures are restricted to shear bands that truncate the D2 fabric.

1.3 Previous work

The first mapping was undertaken by Leitner & Phaup (1974) who named the units in the Mt. Darwin area and subdivided the Rushinga Group into three units: the lower, the middle and the Upper Rushinga Group. They named the metagabbros west of the village of Dotito the Nyamhanda Inlier and interpreted these rocks as complexly infolded with the underlying gneisses. The upper Rushinga Group was re-named by Bache et al. (1983) who also mapped the Mavuradonha Mountains and several other gabbroic bodies in the Mt. Darwin and Centenary areas. They interpreted the metagabbros in the Mavuradonha Mountains as a thick sill which intruded into the underlying amphibolites. Other metagabbros were interpreted as outliers which were thrust onto the felsic gneisses. More recently, Barton et al. (1991) mapped the Rushinga area and neighbouring terrains and interpreted the metagabbro-leuco-metagabbro-amphibolite-association in the Mavuradonha Mountains as highly fractionated rocks. Published radiometric whole-rock Rb/Sr age data fall into two groups: the first group with ages of 830 ± 30 Ma reflects the metamorphic and plutonic climax. The second group shows ages between ~ 660 and ~ 470 Ma and was interpreted as cooling ages. Treloar et al. (1990) and Carney et al. (1991) reported metamorphic *PT* conditions of ~ 800 °C and 12 kbar for the high-grade regional metamorphism corresponding to DZM 1, and 625 to 700 °C and 7 ± 2 kbar for metamorphism related to the DZM 2 in the neighbouring Rushinga area. Collectively, these data suggest an anticlockwise *PT* path. Makanza (1993) and Muzuwa (1993) mapped the area next to the Zambezi Escarpment north of the main ridge of the Mavuradonha Mountains and interpreted the metagabbro-anorthosite-amphibolite association as layered sequences that were metamorphosed under amphibolite- to granulite-facies conditions.

Hargrove (pers. comm., 2000) reported geochemical, petrological and geochronological data of the units in the Mt. Darwin area. He interpreted the metagabbro-anorthosite association north of the Mavuradonha Mission as a 1.8 Ga-old complex which was metamorphosed under granulite-facies conditions at 877 ± 19 °C and 11 kbars (Hargrove et al., 1998; 2003). Additional age dating was carried out by these authors on the Ocellar Gneiss which was dated at ~ 1.15 Ga (Hargrove et al., 1998) and 1051.7 ± 1.3 Ma (Hargrove et al., 2003). Another protolith age for the Ocellar Gneiss was estimated at 1050 ± 4 Ma (Hargrove pers. comm., 2000). An amphibolite-facies metamorphic event represented by a late syntectonic pegmatite occurred at 550 to 530 Ma. Hanson et al. (1998) suggested a maximum age of 900 Ma for the mafic complex in the Mavuradonha Mountains. Vinyu et al. (1997; 1999) presented age data on migmatites and a granite vein in the Rushinga area. These ages were interpreted as the age of regional metamorphism. Migmatisation was found in autochthonous granulites. Here, the ages range from 870 to 850 Ma and were interpreted as the age of high-grade metamorphism. Retrogression under amphibolite-facies metamorphic conditions was suggested to have taken place at ~ 535 Ma, dated on a synmetamorphic granite vein (Vinyu et al., 1997; 1999).

1.4 Aim of this thesis

The aim of this study is the petrological, geochemical and isotopic characterisation of a layered mafic complex in the Zambezi Allochthonous Terrain in order to clarify the evolution of this complex within the framework of the Zambezi belt. There are no recent studies dealing with its petrogenesis, and precise age data for the magmatic emplacement and the granulite-grade event are lacking. Geochronological work was thus carried out on zircons to obtain absolute ages for different geological events in the Zambezi Allochthonous Terrain. Additional studies were performed on tectonically underlying gneisses and other mafic bodies in the Mt. Darwin area.

This study was initiated because no complete data set for this part of the Zambezi belt was available, except for some results from mapping projects and unpublished reports. Fieldwork in the Mavuradonha Mountain range and the Mt. Darwin area was carried out during three campaigns in April 1998, and in September and October 1998 and 1999. A total of about 400 samples were taken along sections across the Mavuradonha Mountain range and the inliers. Field work was carried out on the basis of the available geological map of Bache et al. (1983) at a scale of 1:100 000 and topographical maps at a scale of 1:25 000.

Since the rocks were metamorphosed under granulite-facies conditions, radiometric age determinations using the U-Pb and Pb-Pb isotopic systems were used to date magmatic and metamorphic events in these rocks. Furthermore, age determinations were carried out

on minerals with different closure temperatures (T_c) in order to estimate the time of metamorphism. The Sm-Nd system was utilized to deduce age constraints related to the near-peak to retrograde path of regional metamorphism, using the major rock-forming mineral garnet together with the whole rock. U-Pb isotope data were obtained on rutile which has a lower closure temperature for this system than zircon, and may, therefore, provide age constraints related to the cooling path. Therefore, rutile and garnet were used to date the retrograde part of the *PT*-path.

Pressure/temperature-paths of granulites provide important information on the tectonic setting of high-grade terrains. High-temperature granulite-facies terrains commonly exhibit one of two distinct *PT*-paths that reflect their tectonic evolution. Clockwise *PT* paths are generated by rapid crustal thickening and are often related to plate convergence. Here the thermal peak is reached after peak pressure and main deformation. Commonly, these *PT* paths show near-isobaric cooling after the peak of metamorphism (Ellis, 1987; Bohlen, 1987; Harley, 1989). In contrast, anticlockwise *PT* paths are thought to be related to crustal thickening mainly through intrusion and crystallisation of mafic magmas into, or at the base of, the lower crust (Bohlen, 1987; Bohlen & Mezger, 1989; Jung, 2000). These *PT* paths also exhibit isobaric cooling, but probably at higher pressures. Pressure and temperature estimates were deduced from thermobarometric calculations using mineral assemblages which are interpreted to reflect peak or retrograde conditions of metamorphism. Petrological investigations on reaction textures provide further information about the cooling path that occurred after peak metamorphic conditions. The mineral assemblage garnet + clinopyroxene + plagioclase was used for estimation of peak metamorphic conditions, whereas the assemblage garnet + hornblende + plagioclase and inclusions of relict minerals and reaction textures were taken to define the cooling part on the *PT*-path after peak metamorphism.

Pre-metamorphic information about the source rocks of the Mavuradonha Layered Complex samples is recorded in the isotopic composition. The combination of the different isotope systems (Rb-Sr, Sm-Nd and Pb-Pb) may not only provide information on the source, but also on the subsequent evolution of the complex.

Finally, the results of this study were combined into an evolutionary model for the Zambezi belt of the Zambezi Allochthonous Terrain which provides some insight into the Neoproterozoic evolution of this part of the Zambezi belt in the Gondwana framework.

2 Geology of the study areas

The Zambezi belt of NE-Zimbabwe contains several large mafic masses that consist of amphibolite- to granulite-facies metagabbroic rocks (Leitner & Phaup, 1974; Bache et al., 1983, 1984; Barton et al., 1991, 1993; Carney et al., 1991; Hargrove et al. 1998, 2003; Mariga et al., 1998). The largest block, with an extent of ~60 km, occurs in the Mavuradonha Mountain range, where a layered mafic complex is exposed. Another mafic block is exposed near the township of Dotito, some 20 km to the NE of Mt. Darwin in the Nyamhanda Inlier. The metagabbros in the Nyamhanda Inlier are exposed in a circular structure, surrounded by quartzo-feldspathic gneisses. A smaller metagabbro crops out in the Chimwaya Hill Inlier, which is part of the Mavuradonha Mountain range farther to the west. The Chimwaya Hill Inlier is located 10 km to the N of the Nyamhanda Inlier. The geology of the inliers is only briefly summarized, detailed descriptions are given in Leitner & Phaup (1974) and Bache et al. (1983).

2.1 Geology of the Mavuradonha Mountain range

The Mavuradonha Mts. are bounded to the north by the Zambezi Escarpment and form WNW trending ridges. The highest peak of 1511 m above sea level rises some 600 m (in the S) above the surrounding plains and up to 900 m (in the N) higher than the Zambezi Valley (see cross-section in Appendix I). The most common rock types in the Mavuradonha Mts. (Fig. 2-1) are garnet-bearing amphibolites and metagabbros. Metapyroxenites, meta-anorthosites and garnet-bearing metagabbros occur to a minor extent. The granitoid Ocellar Gneiss crops out at the base of the mountains in the south and is also exposed west of the Mavuradonha Mission. In the valley west of the Mavuradonha Mission the gneisses taper off into a band which is ~100 m wide (see Fig. 2-1 and cross-section in Appendix I).

The formation of steep cliffs in meta-anorthosites and corona-textured metagabbros, low hills in amphibolites, and plains in the Ocellar Gneiss (Plate 1-1) are common characteristic features on the southern side of the Mavuradonha Mts. An increase in metamorphic grade was observed within the mountain range. In the southern part, the Ocellar Gneiss, the metagabbroic rocks and amphibolites are exposed under mid- to upper amphibolite-facies conditions, whereas in the northern part of the study area, near the Zambezi Escarpment, the rocks preserve the assemblage garnet, clinopyroxene and plagioclase which is interpreted as a granulite-facies assemblage (Bache et al. 1983; Makanza, 1993; Hargrove et al. 1998, 2003).

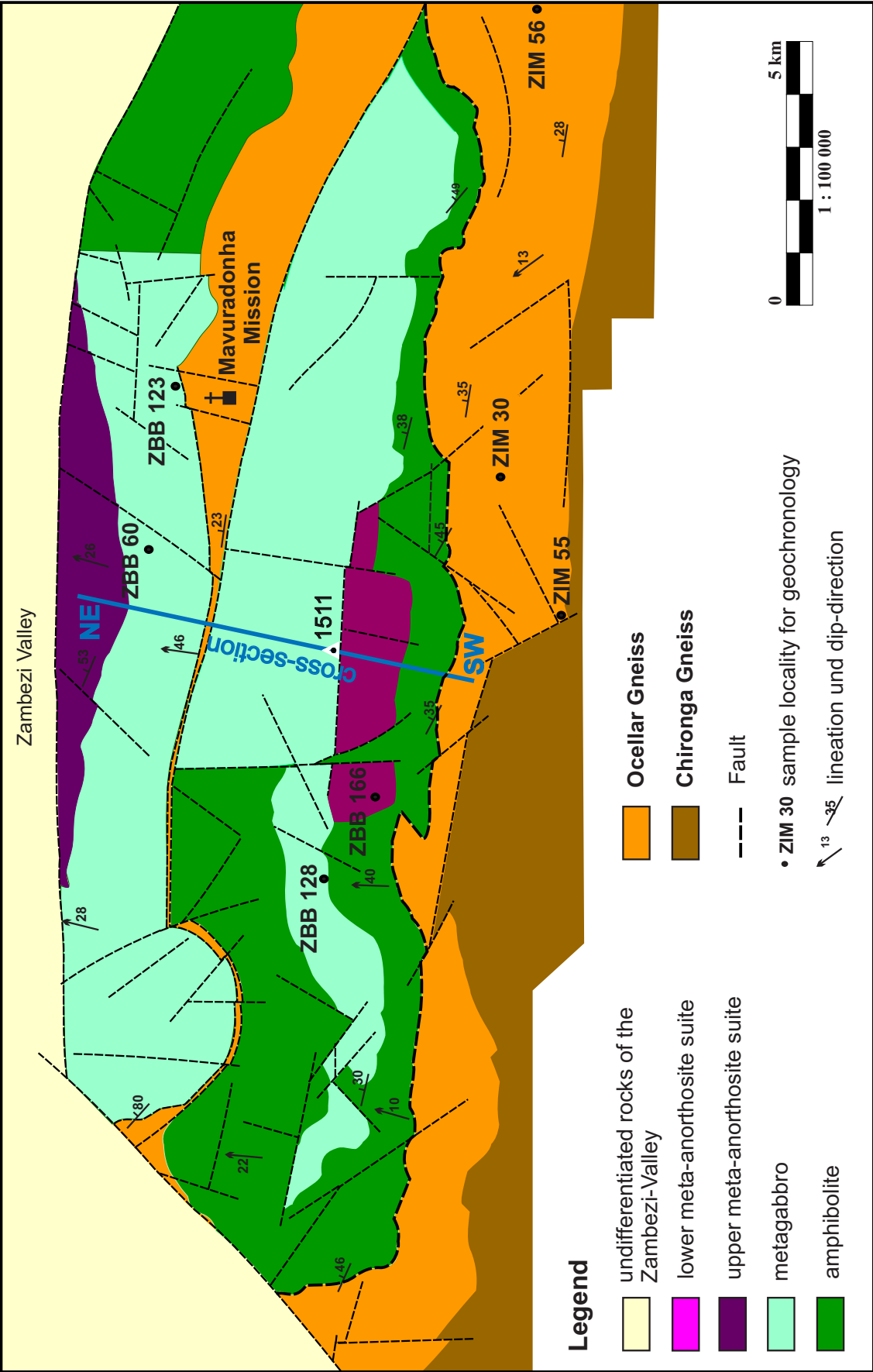


Fig. 2-1. Geology of the Mavuradonha Layered Complex (modified after Bache et al. 1983).

The Mavuradonha Mts. consist of two meta-anorthositic suites and several layered metagabbroic sequences. Amphibolites, comprising garnet-free and garnet-bearing mineral assemblages, are exposed at the base of the layered metagabbroic sequences and within the southern meta-anorthosite suite.

Within the complex a marked rhythmic layering of inferred igneous origin (Plate 1-2) can be observed. Transitional layering is developed mainly in the metagabbroic to metanoritic sequences and in the meta-anorthosite suite in the south. The layering occurs on microscale (mm-cm) and on macroscale (dm-m) with plagioclase and green clinopyroxene as the major primary cumulus minerals. Gravitational accumulation is observed as the major differentiation process in the layered sequences of the complex. In addition to the accumulation of plagioclase and the gravitational segregation, a coarsening in grain size from metapyroxenite to leuco-metagabbro is visible. A repetition of these layered sequences is visible in the entire complex. This repetition is always defined by a sharp contact between the leucocratic layers and the mafic bands.

The base of every layered sequence is formed by a metapyroxenite or fine-grained metagabbro cumulate. The layering grades into coarse-grained metagabbro due to the accumulation of plagioclase. These metagabbros contain coarse corona-textured clinopyroxene and/or orthopyroxene. Enrichment of plagioclase culminates in the formation of leuco-metagabbros. The layered sequence southeast of the highest point in the mountains consists of numerous metapyroxenites, fine-grained metagabbros and corona-textured metagabbros. Additionally, a serpentinite is exposed, suggesting that this sequence represents the most mafic part in the complex.

Concerning the different rock types occurring in the metagabbroic to metanoritic sequences, several characteristic features can be observed. Metapyroxenites and fine-grained metagabbros often reach a maximum thickness of 1.5 m. In contrast, corona-textured metagabbros often exhibit an average thickness of several metres. In external parts of the mountain range several hundred metres of corona-textured metagabbros are exposed. Additionally, within these coarse-grained metagabbros an internal small-scale layering can be recognised. Numerous xenoliths are limited to the layered sequences and are commonly exposed in the coarse-grained metagabbros. These xenoliths consist mainly of fine-grained clinopyroxene, feldspar and scapolite and probably represent wall rocks of the emplacement setting of the Mavuradonha Layered Complex. In the following these layered sequences are termed **layered gabbro series**.

The layering in amphibolites is different from the layering in the metagabbros. In most cases sharp contacts (Plate 1-3) exist between amphibole-rich and plagioclase-rich layers. Garnets are mainly restricted to mafic bands but also occur in intermediate and felsic layers.

The two meta-anorthosite suites differ from each other in their field occurrence and their metamorphic grade. The suite exposed on the southern side of the mountain range exhibits amphibolite-facies grade. The meta-anorthositic layers of this locality are massive, and white to bluish grey in colour, and a common feature of these meta-anorthosites is the mottled texture (Plate 1-4). Amphibole is the most common mafic mineral. Within the meta-anorthosite suite a decrease in abundance of amphiboles from the base to the top is observed. Garnet-bearing amphibolites form the associated mafic layers. Contacts between these meta-anorthosites and garnet-amphibolites are sharp, and no transition is developed. In the following chapters this meta-anorthosite suite is termed **lower meta-anorthosite suite**.

In contrast, the white-coloured meta-anorthosite suite, next to the Zambezi Escarpment, exhibits the mineral assemblage plagioclase, clinopyroxene and garnet. A characteristic feature is the concentration of garnet and pyroxene in bands of variable scale. Garnet-bearing metagabbros form the associated mafic layers with sharp contacts to the meta-anorthosites (Plate 1-5). Both garnet-bearing mafic layers and meta-anorthosites range from 20 cm to several metres in thickness. In the following chapters this suite is termed **upper meta-anorthosite suite**.

Considering the observations described above, the layered sequences and the meta-anorthosite suites in the Mavuradonha Mts. are termed **Mavuradonha Layered Complex** in the following chapters.

In the NW of the layered complex, in the area of Nanuta, a circular, caldera-like structure of metagabbro is exposed. The central part of Nanuta forms a depressed plateau, which is enclosed by steep hills. The metagabbros are often associated with monomineralic green-coloured metapyroxenites. On the northeastern slope, the metagabbros grade into a leuco-metagabbro due to the occurrence of plagioclase. The metagabbros in the central and southern part of Nanuta differ from the metagabbros exposed in the Mavuradonha Layered Complex. The plagioclase within these metagabbros is green-coloured, and the pyroxenes have no hornblende corona. The southeastern slope of the circular structure consists of pure Fe-rich opaques and is described as a magnetite body (Makanza, 1993; Muzuwa, 1993).

The quartzo-feldspathic Ocellar Gneiss is often strongly weathered. Outcrops are found in streambeds and along the road to Mukumbura. The gneiss has a heterogeneous field appearance, and most of the gneiss is ochre to grey coloured, and strongly foliated. Aligned hornblende and biotite define the foliation. K-feldspar phenocrysts occur as augen in the gneisses. Other varieties of the Ocellar Gneiss are enriched in white mica. At some localities a metamorphic layering is exposed consisting of leucocratic feldspar-bearing beds as well as biotite and/or hornblende-bearing mafic beds. The leucocratic layers are

medium-grained and also contain garnet. A common feature is the occurrence of these leucocratic layers as schlieren. Pegmatites and quartz veins are folded and sheared in the Ocellar Gneiss.

Intrusive relationships between the granitoid gneisses and the mafic complex have not been observed. The Ocellar Gneiss and the mafic rocks display the same penetrative foliation, and the granitoid gneisses and the mafic complex have parallel tectonic contacts to each other. Field evidence points to thrusting of the mafic complex onto the Ocellar Gneiss (Bache et al., 1983; Barton et al., 1991; Hargrove et al., 1998). Tectonic movement of the layered complex to the south is recorded by shear-sense indicators in the metagabbros (pyroxenes as σ -clasts and s-c fabrics). Shear-sense indicators in the Ocellar Gneiss show mainly top-to-the-North movement (Plate 1-6). This opposite direction of movement is probably the result of intense folding after overthrusting that has affected the gneiss after the development of its foliation, lineation and the shear sense indicators. The nappe contact is marked by a depression and an increase in deformation within the Ocellar Gneiss towards the mafic complex. The occurrence of mylonitic rocks at the contact between the Ocellar Gneiss and amphibolites give further evidence for a tectonic contact between the two units.

Pegmatites, calc-silicates and marbles are found in the Mavuradonha Layered Complex as well as in the Ocellar Gneiss. Massive quartzites are restricted to the Ocellar Gneiss on its southern side near the base of the Mavuradonha Layered Complex. Marbles and calc-silicates are found only north of the main ridges and occur as lenticular bodies. In one of the marbles next to the Zambezi Escarpment, fragments of the surrounding host rocks are found. Pegmatites occur as medium to coarse-grained dykes with varying width. Commonly these pegmatites are perpendicular to the foliation of the host rocks or crosscut the foliation or layering of their country rock. One pegmatite sampled in the north of the Mavuradonha Mountains (sample ZZB 60 in Fig. 2-1) intruded during the last tectono-thermal event. This pegmatite is strongly sheared, and the foliation developed in the metagabbros is dragged into the crosscutting pegmatite (plate 1-8). The pegmatite is interpreted as syntectonic-synmetamorphic with respect to the last major tectono-thermal event.

The observed layering in the Mavuradonha Layered Complex is inclined to the N with a predominant dip of 40°. The granulites and amphibolites, as well as the Ocellar Gneiss, show a marked foliation with a NE strike and a dip between 40° to 60° (Fig. 2-1). This foliation can most likely be correlated with a first deformation event in these rocks. In addition to the foliation, a lineation is developed. A hornblende mineral lineation in the case of the amphibolites dips around 40° to the NE. The lineation in the Ocellar Gneiss dips 10° to 30° N to NW (Fig. 2-1). Recumbent and isoclinal folds mark a second

Plate I-1: View at the landscape formed by the lower meta-anorthosite suite in the central part of the Mavuradonha Layered Complex. Note the steep cliff in the background which is built of nearly pure meta-anorthosite and the flat hills formed by amphibolites. In contrast, the front of the photograph is located in the Ocellar Gneis and forms flat plains.

Plate I-2: Rhythmic layering in the layered gabbro series in the central part of the Mavuradonha Layered Complex (appendix I, Fig. I-3). The photograph shows transitional layering with meta-pyroxenites at the base of every sequence grading into corona-textured metagabbros and leuco-metagabbros.



Plate I-3: Layering in the ferro-metagabbros in the south of the Mavuradonha Layered Complex (appendix I, Fig. I-3). Note sharp contacts between the ferro-metagabbros (here garnet-amphibolites) and the meta-anorthosites, whereas a transitional layering also occurs between garnet-free amphibolites and meta-anorthosites.

Plate I-4: Mottled appearance of meta-anorthosites with oikocrysts of retrograde amphiboles after clinopyroxene in plagioclase from the lower meta-anorthosite suite of the Mavuradonha Layered Complex (appendix I, Fig. I-3).



Plate I-5: Compositional layering in the upper meta-anorthosite suite of the Mavuradonha Layered Complex (appendix I, Fig. I-3). Note sharp contacts between the ferro-metagabbros and the meta-anorthosites.

Plate I-6: Shear-sense indicator within the Ocellar Gneiss (appendix I, Fig. I-3) showing top-to-the-north directed shear movement. The picture shows deformed feldspar lenses in hornblende-rich mafic layers.



Plate I-7: Shearzone within the corona-textured metagabbros of the layered gabbro series (appendix I, Fig. I-3). Note characteristic feature of pinkish garnet-coronas around clinopyroxene in these shearzones.

Plate I-8: Pegmatite cruss-cutting the internal layering of a ferro-metagabbro at the base of the upper meta-anorthosite suite (appendix I, Fig. I-3). Note that the layering is dragged into the pegmatite.



deformational event. This event occurred in the amphibolites and the Ocellar Gneiss, resulting in folding of the existing foliation. Fold axes plunge gently to the SW and NE in amphibolites and the Ocellar Gneiss respectively. High-grade shear zones are a common feature in the Mavuradonha Layered Complex. These shear zones are assumed to be the result of the third deformation event which affected mainly the central part of the Mavuradonha Mts. Locally these shear zones affected the foliation and folds within the metagabbros. Intense shearing of metagabbros is commonly associated with the formation of garnet between plagioclase and clinopyroxene (Plate 1-7). In amphibolites and metagabbros, low strain rigid blocks are often observed in association with the shear zones. These blocks still preserve their magmatic gabbroic textures. Shearing, which is probably related to the third deformation event in the anorthosites of the upper meta-anorthosite suite, is visible by the spotted texture and the development of pressure shadows consisting of clinopyroxene and/or amphibole around garnet.

Most of the major faults in the main part of the Mavuradonha Layered Complex are parallel to the Zambezi Escarpment. Therefore, it is argued that the mountain range was significantly affected by the formation of the Zambezi Valley. These effects are shown by the formation of step faults. In some outcrops brittle deformation on the fault planes is observed. Consequently, reactivation of older shearzones by brittle faults is suggested, e.g., there was brittle deformation at the base of the amphibolites instead of the nappe contact between Ocellar Gneiss and the Mavuradonha Layered Complex.

2.2 Geology of the inlier

2.2.1 Nyamhanda Inlier

In the Nyamhanda Inlier (Fig. 2-2) metagabbroic rocks and migmatitic gneisses occur in an oval synclinal structure that was thrust onto the Ruya Granitic Gneiss (Bache et al., 1983). The metagabbroic rocks are complexly infolded with the surrounding migmatitic Chironga Gneiss. The highest point of 1265 m above sea level is located in the central part of the inlier. Plains are developed due to intense weathering of the migmatitic gneisses, which show a heterogeneous nature. The Chironga Gneiss consists of leucocratic, feldspar- and quartz-rich and mafic, hornblende- and biotite-rich layers. The muscovite-bearing and garnet-rich Wadze Gneiss occupies the outer part of this inlier.

Amphibolites and metagabbros are the common mafic rock types. The metagabbros range from dark-coloured, fine-grained varieties to leucogabbroic varieties. In the NW the gabbros show a layered appearance. The base of the metagabbros is always formed by amphibolites.

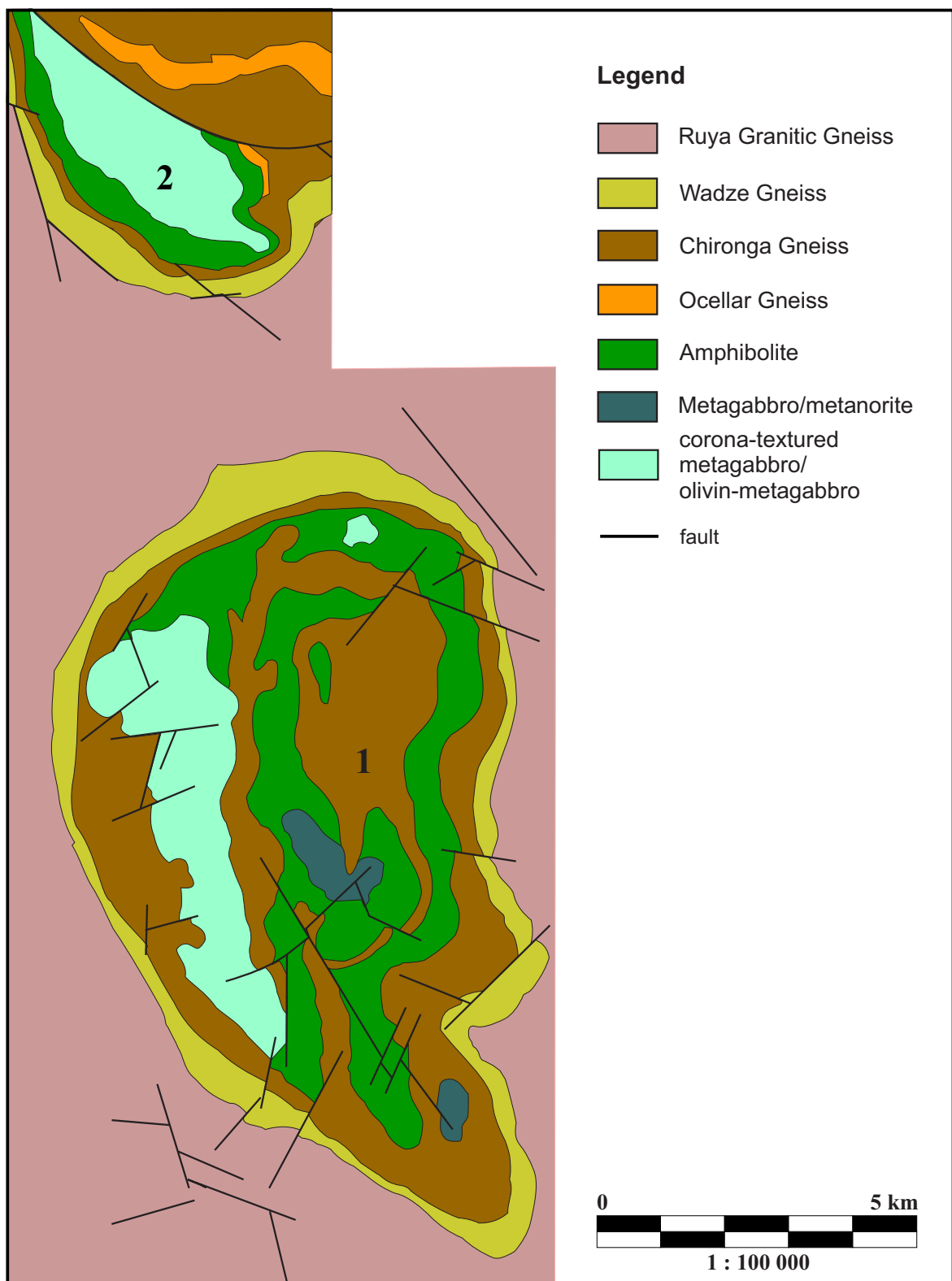


Fig. 2-2. Geology of the Nyamhanda Inlier (1) and the Chimwaya Hill Inlier (2) (modified after Bache et al. 1983).

Two smaller gabbroic bodies are exposed in the NE and SE and form distinct hills. The metagabbro in the NE is surrounded by amphibolites, whereas gneisses enclose the metagabbro in the SE.

Amphibolites occur as both garnet-free and garnet-bearing varieties. Some of the amphibolites are enriched in plagioclase, whereas other types are mafic, hornblende-rich varieties. The latter are interpreted to be of metamorphic origin. The amphibolites are always in contact with the underlying gneisses. Intrusive relationships between the gneisses and the metagabbroic rocks were not observed.

The metagabbro/amphibolite associations in the Nyamhanda Inlier form NW trending ridges. This trend is also observed in the gneisses and probably marks the orientation of the fold axes. The observed layering and foliation in the amphibolites and the gneisses are tilted to the W and E, depending on the occurrence in the fold limbs, with a predominant dip of 60°-80°. This foliation most likely represents the first deformation event in these rocks of the Nyamhanda Inlier. In the gneisses a lineation is developed in addition to the foliation, which plunges 65° to the SE. The second deformational event marks the folding of the gabbroic rocks and the gneisses. A further folding and refolding of pre-existing folds is observed in the gneisses. High-grade shearzones showing ductile deformation are a common feature in the inlier. These shearzones are assumed to be the result of the third deformation event because all units are affected by these shearzones.

2.2.2 Chimwaya Hill Inlier

A small metagabbro (Fig. 2-2) crops out at Chimwaya Hill (1296 m), which is located in the Mavuradonha Mountain range. The Chimwaya Hill Inlier shows the same field appearances as the metagabbros in the Mavuradonha Layered Complex and Nyamhanda Inlier. The contact between gneisses at the base and the metagabbros is marked by amphibolites. Intrusive relationships between the gneisses and the mafic complex are not preserved.

Metagabbros and amphibolites display an inherited igneous layering, grading from fine-grained dark coloured to coarse-grained leucocratic varieties. Two types of gneisses are exposed surrounding the metagabbros. The migmatitic Chironga Gneiss, also exposed in the Nyamhanda Inlier, occurs in direct neighbourhood to the metagabbroic rocks, but the contact is not exposed. The second gneiss is the muscovite-bearing and garnet-rich Wadze Gneiss, which is exposed in the outer part of this inlier. Characteristic for this gneiss is the formation of a deep depression due to intense shearing and weathering.

Shearing is marked by the formation of a distinct foliation and elongated clinopyroxenes in the metagabbros. The foliation dips around 60° to the NE. These sheared metagabbros are

mainly exposed in the central part at the base of Chimwaya Hill. Shearing is also observed in the gneisses documented as pressure shadows formed on garnets. These garnets show a shear-sense of top-to-the-North.

3 Analytical methods

More than 200 representative samples were collected for geochemistry during three field trips in 1998 and 1999, each sample weighing between 3 and 5 kg. Two hundred additional hand specimens were collected for microprobe and microscopic investigations. Rock crushing (jaw crusher only) of 100 samples and sawing of all rock slices were performed at the Geology Department, University of Zimbabwe, Harare. A homogeneous split of 500 g from the crushed samples was used for geochemistry and isotope chemistry.

Mineral separation of samples from the Ocellar granitoid gneiss was undertaken at the Geology Department, University of Zimbabwe in Harare. Since zircons were observed in several thin sections, the remaining splits of the metagabbros were reduced to a grain size $< 400 \mu\text{m}$ in a roller mill in order to obtain material for mineral separation. Sixty-five samples were chosen for geochemistry, and a homogeneous split of each sample was ground in an agate mill in order to obtain a fine powder. The powders were dried overnight at 110°C before powder and glass tablets were prepared. Loss on ignition (LOI) was determined gravimetrically at 1100°C with modifications after Lechler & Desilets (1987). Major elements (except FeO) and trace elements were analysed by XRF on glass tablets and powder tablets respectively, using a Phillips SRS 200 fluorescence spectrometer. The calibration was based on international and laboratory whole-rock standards. FeO was separately determined by potentiometric titration with a Cerium solution on a semi-automatic Methrom-titriprocessor. Between 100 and 200 mg whole-rock powder was dissolved in a mixture of H_2SO_4 -HF and subsequently heated in a platinum crucible. The solution was added to a mixture of saturated H_3BO_3 - H_3PO_4 solution and titrated automatically. Concentrations of Fe_2O_3 were calculated using values of Fe_2O_3 determined by XRF and results obtained by titration for FeO. The Mg-numbers have been calculated using the following definition: $\text{Mg\#} = \text{MgO}/(\text{MgO} + \text{FeO})$, in which FeO is the value analysed by titration.

Accuracy was controlled by repeated measurements against several international and in-house standards, and the results (Appendix II, Table 1) are in good agreement with the recommended values. Precision for trace element analysis was verified by preparation and analysis of two powder tablets per sample. The variation for trace element concentrations measured in both tablets was 1 ppm for concentrations between 1 and 15 ppm and less than 2 ppm for element concentrations lower than 100 ppm. Variation was less than 5 ppm for concentrations of elements higher than 100 ppm.

For Sr and Nd isotopic analyses, sample preparation followed the procedure of White and Patchett (1984). Whole-rock powders (35 to 500 μg) were spiked with a mixed $^{85}\text{Rb}/^{84}\text{Sr}$ and $^{149}\text{Sm}/^{150}\text{Nd}$ tracer. The spiked samples were dissolved in Teflon[®] beakers

for several hours on a hotplate using a HF-HNO₃ mixture. Accessory mineral dissolution was achieved inside a Parr[®] bomb at 200 °C after 1-2 days.

Whole-rock samples were dried, re-dissolved in HNO₃ and dried again. 6 N HCl was added three times, and the solution was subsequently dried. Before separation of Rb, Sr and REE the samples were re-dissolved in 2.5 N HCl and centrifuged. Rb, Sr and REE were separated using 5 ml and 10 ml ion exchange columns filled with ion exchange resin (AG 50 W-X-12). The elements Sm and Nd were separated from the rest of the REE on 2 ml Teflon[®] columns filled with di-2-ethylhexyl orthophosphoric acid coated Teflon[®] powder.

The elements Rb, Sm and Nd were loaded with 2.5 N HCl on double Re-filaments. For Sm and Nd, 1 µl H₃PO₄ was added onto the filament. Strontium was loaded with 2 µl TaF₅ and 2.5 N HCl on single W filaments. Rb, Sr, Sm, and Nd were measured in static, multicollector mode on a Finnigan MAT 261 mass spectrometer at the Max-Planck-Institut für Chemie in Mainz. Fractionation corrections were made by normalising Sr measurements to ⁸⁶Sr/⁸⁸Sr = 0.1194 and Nd to ¹⁴⁶Nd/¹⁴⁴Nd = 0.7219. The mean values of all measured standards yielded a reproducibility for NBS 987 of ⁸⁷Sr/⁸⁶Sr = 0.710207 ± 0.000010 (2 σ, n = 11) and for La Jolla of ¹⁴³Nd/¹⁴⁴Nd = 0.511853 ± 0.000010 (2 σ, n = 12). Values for Nd < 40 pg and for Sr < 200 pg were measured as blanks. The initial ratios of Sr and Nd were calculated using the decay constants for Rb of Steiger and Jäger (1977) and for Sm of Lugmair and Marti (1978). The initial ε_{Nd} values were calculated using present-day ratios for ¹⁴³Nd/¹⁴⁴Nd and ¹⁴⁷Sm/¹⁴⁴Nd following the procedure of Jacobsen & Wasserburg (1980). The computer program Isoplot[®] of Ludwig (1994) was used to calculate best-fit lines for Rb/Sr and Sm/Nd isotope systems. A minimum error of 0.003 % was assumed for ¹⁴³Nd/¹⁴⁴Nd based on the reproducibility of the La Jolla standard, except for sample MAV 1. The error for this sample is 0.005 %. Errors for the ¹⁴⁷Sm/¹⁴⁴Nd ratios are less than 0.25 %. Lead isotopic compositions were analysed on plagioclase separates. Thirty to fifty mg of plagioclase were handpicked from the non-magnetic fraction. The samples were leached two to three times in a mixture of HF-H₂O for several minutes on a hot plate. After each leaching step the separates were washed three times in Millipore[®] water. Plagioclase was dissolved in HF overnight on a hot plate. Lead was extracted from the re-dissolved and dried samples using the HCl-HBr method on 0.5 ml Teflon[®] columns filled with DOWES AG 1X8 anion exchange resin. Lead was loaded on single Re-filaments with silica gel and measured in static, multicollector mode.

Mineral separation

After grinding, the samples were washed in distilled H₂O and dried. The dried samples were sieved to fractions of 500-350 µm, 350-125 µm and < 125 µm. The fraction of 350-125 µm was run through a magnetic separator in order to obtain the non-magnetic fraction. Zircons were separated from this fraction using heavy liquids (methylene iodide) and were finally handpicked. Most of the zircons from all samples show a younger rim, which is most likely the result of a second period of zircon growth. Consequently, all zircons analysed in this study were air-abraded (Krogh 1982) and finally checked for contrast differences using heavy liquids. In order to obtain cathodoluminescence images, zircons of similar sizes were handpicked and mounted together in epoxy resin. The mount was polished with diamond and aluminium-oxide polishing paste on a Teflon[®] disk until the centres of the zircons were exposed. After polishing, the mounts were dried overnight at 75 °C and subsequently coated with carbon. For CL images, a JEOL JXA 8900 RL electron microprobe at the Institut für Geowissenschaften, Mainz was used. Operating conditions were 15 kV accelerating voltage and 12 nA probe current.

Garnet was separated with a magnetic separator. Fractions between 0.8 and 1.3 g by weight were handpicked in order to obtain inclusion-free separates. The garnets were ground using an agate mortar, and the resulting powder was leached in a three-step boiling procedure for one hour in (1) 6 N HCl, (2) 10.8 N HCl and (3) aqua regia to remove remaining inclusions (Jung & Mezger, 2001). After leaching, the samples were spiked with a Sm-Nd tracer and dissolved on a hot plate in a HF-HNO₃ mixture for 2 days at 180 °C. REE were separated and measured in the same manner as described above for the whole-rock samples. The computer program Isoplot[®] of Ludwig (1994) was used to calculate best-fit lines for Sm/Nd isotope system.

SHRIMP method

U-Th-Pb analyses were performed by A. Kröner on the SHRIMP II ion microprobe of the Perth Consortium at Curtin University of Technology, Perth, Australia. A detailed description of the analytical procedure is given in Compston et al. (1984, 1992), Nelson (1997) and Williams and Claesson (1987). Data acquisition for each analysed spot was based on seven mass scans. Mass fractionation corrections were made by normalisation of the measured ratios to the ratios obtained from standard zircon CZ 3 (Nelson, 1997). Common lead was corrected using the measured non-radiogenic lead isotope ²⁰⁴Pb, since common Pb was assumed to be surface related (Kinny et al. 1989). In the case of low common Pb concentrations, similar to those measured on the CZ3 standard, the common Pb correction followed the method of Compston et al. (1984). The linear model of Cummings & Richards (1975) was assumed for unknown zircons in the case of six times

higher ^{204}Pb counts than the CZ3 ^{204}Pb counts (Nelson, 1997). Data regression followed the method described in Nelson (1997). Reported ages for single spot analyses are given at 1σ error. The errors are based on counting statistics. Concordia intercept calculations were prepared using the computer program of Ludwig (1994).

Single zircon evaporation method

Zircon evaporation was carried out after Kober (1986, 1987) with minor modifications after Kröner & Todt (1988) and Kröner & Hegner (1998). Zircons were embedded into Re-filaments and measured in double filament arrangements. The evaporation method involves repeated measurements of untreated zircons with gradually increased temperature for evaporation and ionisation. Deposition of the evaporated Pb isotopes takes place on a second filament from which the Pb isotopes are ionized. In this study only data without changes in isotopic ratios were considered for age determinations. The ages and their 2σ mean error were integrated from all measured $^{207}\text{Pb}/^{206}\text{Pb}$ ratios; the distribution of these ratios is shown in histograms. Pb mass fractionation is estimated around 0.3 ‰ per atomic mass unit (W. Todt, pers. comm. 1999). This fractionation is significantly less than the relative standard deviation of the measured $^{207}\text{Pb}/^{206}\text{Pb}$ ratios and therefore no corrections were made for mass fractionation. The common lead correction was performed following the two-stage lead evolution model (Stacey & Kramers, 1975).

Single zircon U-Pb vapour transfer method

Single zircons were dissolved in a Teflon[®] capsule with six separate holes following the procedure described in Parrish (1987) with modifications after Wendt and Todt (1991). The zircons were picked and ultrasonically washed in a Savilex[®] beaker in a two-step procedure. The first step involved washing the zircons for 30 min with 7 N HCl, and the second step included washing for 5 min with 3 N HNO₃. After each washing step the zircons were rinsed three to four times with Millipore[®] water. Single or up to three (only for the metagabbros) zircons were handpicked and transferred into each hole. An enriched mixed ^{265}U - ^{205}Pb tracer was added prior to dissolution of the zircons with HF for up to 5 days at 200 °C. The samples were dried, and 3 µl 6 N HCl was added, and the sealed capsule was again heated for 12 hours. Lead and U of some zircons were separated using 0.5 ml Teflon[®] columns with the HCl-HBr method using DOWES AG 1X8 anion exchange resin. The separation minimized the matrix effects on Pb and U, which led to a stable and strong signal for those isotopes during measurement. Zircon analyses, in which U and Pb were separated, are marked with a 'c' in Table V-1 and V-3, appendix V. Concordant data points are given as $^{206}\text{Pb}/^{238}\text{U}$ ages because of the more precisely determined isotopic ratios.

U-Pb rutile analyses

Three fractions of 4-8 mg rutile were handpicked from sample ZZB 150 in order to obtain optically inclusion-free separates. The rutile separates were washed in a mixture of 0.5 N HF/1N HCl or 6 N HCl respectively. The fractions were spiked with a $^{233}\text{U}/^{205}\text{Pb}$ mixed tracer and were dissolved in Teflon[®] beakers using a H_2SO_4 -HF mixture for four days in a Parr[®] bomb.

Subsequently, the rutile samples were dried on a hot plate at 200 °C for at least three weeks to remove H_2SO_4 . Uranium and lead were separated using 0.5 ml Teflon[®] columns. Lead was separated using the HCl-HBr method and using DOWES AG 1X8 anion exchange resin. Uranium was separated on UTEVA exchange resin (100-150 µm) using the 2.0 N HNO_3 – 0.02 N HNO_3 .

For U and Pb measurements of zircon and rutile the samples were taken up with three ml silica gel and loaded on single Re-filaments. Lead and U were ionised at different temperatures (Pb: 1180-1250 °C; U: 1350-1450 °C) and measured on a Finnigan MAT 261 mass spectrometer in peak jumping mode with a secondary electron multiplier (zircons) and in static, multicollector mode (rutile) at the Max Planck Institut für Chemie in Mainz. Blank values were 3.5 pg for lead in which the lead isotopic composition was $^{206}\text{Pb}/^{204}\text{Pb}$: 18.210, $^{207}\text{Pb}/^{204}\text{Pb}$: 15.162 and $^{208}\text{Pb}/^{204}\text{Pb}$: 36.890. Common lead corrections for the zircons and rutile of the Mavuradonha Layered Complex were performed with measured Pb isotope compositions from coexisting plagioclase. For zircons of the Ocellar Gneiss, common Pb corrections were undertaken for an age of 1000 Ma, using the value of Stacey and Kramers (1975). Lead fractionation (2.9 to 3.2 ‰ per amu during this study) was corrected each time by measuring a standard NBS 981 under the same conditions. All isotopic results were corrected for spike, common lead, blank and Pb fractionation. Regression lines were calculated after York (1969). Since the zircons were not weighed, no U and Pb concentrations could be calculated. However, the use of a mixed spike allows the calculation of a total U over radiogenic lead ratio.

Mineral chemistry

Electron microprobe analyses were performed using a JEOL JXA 8900 RL electron microprobe (EPMA) at the Institut für Geowissenschaften, Mainz. This was to identify mineral chemical characteristics within different layered sequences in the Mavuradonha Layered Complex and for the purpose of geothermobarometry. Operating conditions for SilUMainz, used for all minerals except garnet, were 15 kV accelerating voltage and 12 nA probe current. Chlorine was measured as an additional element for scapolite and amphibole analysis. For garnet analyses the method GranUMainz with 20 kV accelerating voltage and 20 nA probe current was used. All analyses were performed in

wavelength dispersive mode with a focused beam of about 2 μm . Matrix corrections were made with the EPMA internal $\phi\rho z$ correction program.

Structural formulae and endmembers of minerals were calculated following the procedure described in Deer et al. (1986) for plagioclase, in Morimoto et al. (1988) for pyroxene, in Spear (1993) for garnet, and in Leake et al. (1997) for amphibole. Structural formulae of scapolites are calculated following the recommendations of Shaw (1960) and Teertstra & Sherriff (1997), whereas endmembers are defined as suggested by Shaw (1960) and Ellis (1978). Fe^{3+} corrections in pyroxene, amphibole and garnet formulae followed the procedure of Droop (1987). Mineral abbreviations follow the suggestions of Kretz (1983).

Thermobarometry

Temperature calculations were performed using the garnet-pyroxene geothermometer of Ellis & Green (1979) and Krogh (1988) and the garnet-hornblende geothermometer of Graham & Powell (1984). Pressures were calculated applying the geobarometers of Newton & Perkins (1982), Powell & Holland (1985; 1988), Holland & Powell (1985) and Eckert et al. (1991) and using the mineral assemblage garnet-clinopyroxene-plagioclase-quartz. The geobarometer of Kohn and Spear (1989) was used for pressure calculations of the assemblage garnet-hornblende-plagioclase-quartz.

Total iron was assumed to be ferrous iron for *PT*-estimates following the suggestions of Schuhmacher (1991) and Canil & O'Neill (1996). This is because of the still unsolved problem of ferric iron estimation in many ferromagnesian minerals as mentioned by several authors (e.g., Bucher & Frey, 1994; Krogh Ravna, 2000; Schumacher, 1991). Comparative studies (Canil & O'Neill, 1996; McCammon et al., 1998) between Mössbauer spectrometry and the method of Droop (1987) have shown that ferric iron calculations for pyroxenes which are based on stoichiometry lead to imprecise ferric iron estimations caused by higher SiO_2 levels and lower total iron contents compared with spinel or garnet. Ferric iron estimates based on the method described by Droop (1987) may lead to erroneous results in thermobarometry, due to the extreme sensitivity to microprobe errors. Most of the geothermobarometers used in this study are calibrated using experimental materials and/or natural assemblages with poorly constrained amounts of ferric iron.

To obtain information about the *PT* evolution, quantitative *PT*-conditions have been evaluated applying the TWEEQ/TWQ (Thermobarometry With Estimation of EQUilibration state) computer application of Berman (1991) versions 1.02b (for garnet-amphibole-plagioclase calculations) and 2.02 (garnet-clinopyroxene-plagioclase). This enables the combination of the *PT* estimates with mineral reactions deduced from

microscopic investigations. The calculations were processed on the internally consistent thermodynamic data set of Berman (1988, 1990). These versions use the activity model for plagioclase of Fuhrman & Lindsley (1988), the model of Berman et al. (1995) and Aranovich & Berman (1996) for pyroxene, and for garnet the models of Berman (1990) and Berman & Aranovich (1996). Margules parameters, defined in Berman and Brown (1984), have been used to calculate the activity for amphiboles. Considered elements for calculation of the PT data were Si, Al, Ca, Na, K, Mg, Fe, Ti, O and H. Endmembers were for garnet almandine, pyrope and grossular, for clinopyroxene hedenbergite and diopside, for plagioclase albite and anorthite as well as for amphiboles tschermakite, pargasite, tremolite and the corresponding iron-rich endmembers.

4 Petrology and metamorphic evolution

This chapter describes the petrology and metamorphic evolution of the Zambezi Allochthonous Terrain, the structurally highest terrain in the Zambezi belt (Barton et al. 1991; 1993). As mentioned in chapter 1, previous studies assumed that this part of the Zambezi belt was overprinted under granulite-facies conditions (Carney et al. 1991). The *PT* calculations in this chapter were carried out in order to unravel the evolution of the Mavuradonha Metamorphic Suite (Barton, 1991; 1993), which forms the structurally highest part of the Zambezi Allochthonous Terrain.

Additional work was undertaken to describe the different layered sequences of the Mavuradonha Layered Complex. The terminology for the description of cumulate rocks used in this study follows terminology of Irvine (1982). This author redefined the terminology of Wager & Brown (1968), which is now available without relations to processes that lead to the formation of the cumulate rocks.

Additionally a brief petrographic description of the metagabbroic rocks in the Nyamhanda and Chimwaya Hill Inliers and the Ocellar Gneisses is given.

4.1 Petrography - mineral assemblages and textures

4.1.1 Mavuradonha Layered Complex

The first part of this chapter provides a description of the petrology of the different rock types and mineral phases occurring in the Mavuradonha Layered Complex in order to distinguish these rocks by their petrography. The mineral chemistry for garnet-bearing samples of the upper meta-anorthosite suite (e.g. metagabbros and meta-anorthosites), which were taken for thermobarometry, is described in greater detail. Special attention is given to corona textures associated with pyroxenes in the metagabbros. These textures are used to derive *PT* estimates for the retrograde part of the *PT*-path. The second part discusses the results of the thermobarometric estimates of each investigated sample. The results, in combination with the observed textures, were taken to deduce a *PT*-path.

Samples of the different rock types exposed in the Mavuradonha Layered Complex were taken along sections across the mountain range, whenever a marked change in the mineral assemblage was observed. A complete description of the mineral assemblages of the investigated samples is given in Table III-1 in Appendix III. Photographs of thin sections are presented in Plate II.

The samples can be subdivided into eight groups according to their mineral assemblage. The following types can be distinguished:

- serpentinite
- metapyroxenites
- garnet-bearing and garnet-free amphibolites
- fine-grained metagabbros with variable amounts of hornblende and plagioclase
- coarse-grained metagabbros
- garnet-bearing metagabbros
- leuco-metagabbros and leuco-amphibolites
- meta-anorthosites, with variable amounts of garnet, clinopyroxene and hornblende
- pegmatites

The most common rock types in the Mavuradonha Layered Complex are garnet-bearing amphibolites at the base of the complex and coarse-grained metagabbros. The mineral assemblage plagioclase + clinopyroxene + garnet is restricted to the northern side of the Mavuradonha Mountain range and appears in the upper meta-anorthosite suite.

Meta-anorthosites are restricted to two large suites but also occur in small layers within the layering of the metagabbro sequences. The term meta-anorthosite is used only to describe those rocks which occur within the two suites. Those that appear within the layered sequences of variable scale are termed 'leuco', because of the larger amounts of pyroxene pods and the limited scale of their occurrence.

4.1.1.1 Serpentinite

A serpentinite (ZZB 55) is exposed on the southern side of the Mavuradonha Mountain range within the sequence of the main metagabbro. The serpentinite consists of equigranular serpentine associated with opaque minerals and small amphiboles.

4.1.1.2 Metapyroxenites

Typical metapyroxenites are restricted to the circular structure of Nanuta and to the small hills north of Mavuradonha Mission. Metapyroxenites are always associated with coarse-grained metagabbros and show an equigranular interlobate texture. The metapyroxenites are monomineralic rocks consisting almost exclusively of clinopyroxene; rutile and ilmenite are the only accessory mineral phases. They often occur as exsolved grids within clinopyroxene. Clinopyroxene is partly replaced by green and brown

hornblende. In some samples calcite was found.

In the main metagabbro sequence, a metapyroxenite consisting of amphibole, plagioclase and pyrite, as the major opaque phase, is exposed. Amphibole and plagioclase are also present as symplectitic intergrowth.

4.1.1.3 Amphibolites

Two types of amphibolites are exposed in the Mavuradonha Layered Complex. The most common are garnet-bearing amphibolites. Garnet-free amphibolites occur to a minor extent in the southern part of the Mavuradonha Mountains at the base of the Mavuradonha Layered Complex.

4.1.1.3.1 Garnet-bearing amphibolites

The garnet-bearing amphibolites show a layering of mafic and felsic bands. Garnet shows a preferred affinity to the mafic bands, although it also occurs in the felsic units. The grain size of the porphyroblastic garnet in these amphibolites ranges from several mm up to 5 cm. In some samples, garnet occurs in two different populations. One population forms poikilitic grains (Plate II-1) with inclusions of hornblende, plagioclase, quartz, titanite and scapolite. This type of garnet occurs in all samples. A second garnet population is identified in some samples by an inclusion-free rim around a poikilitic core. Both stages of garnet growth are in coexistence with hornblende, plagioclase, scapolite, quartz, titanite, apatite, zircon and ilmenite. Clinopyroxene is also observed as relics (Plate II-1) and as inclusions in garnet in some samples. Garnet-bearing amphibolites have a seriate, polygonal to interlobate texture. Garnet is locally separated from other ferromagnesian phases by a plagioclase rim of variable size.

4.1.1.3.2 Garnet-free amphibolites

Garnet-free amphibolites are fine-grained and contain the mineral assemblage green amphibole, plagioclase, rutile, quartz, titanite, scapolite, apatite and epidote. Some of the amphibolites show a decussate texture and are well foliated due to the concentration of amphiboles and plagioclase alternating in thin bands. Light green amphiboles often contain numerous quartz inclusions. Titanite partly forms rims around rutile or amphibole. Another textural feature is the formation of titanite grids in amphiboles. Plagioclase occurs as primary twinned remnant grains or as recrystallised grains. Plagioclase is often replaced by scapolite and/or by needle-shaped or skeletal epidote.

4.1.1.4 Fine-grained metagabbros

Fine-grained metagabbros (Plate II-2) are exposed as mafic layers with variable amounts of plagioclase and amphibole. They occur at the base of every layered sequence in association with coarse-grained metagabbros and leuco-metagabbros, and they are often strongly sheared and recrystallised. The fine-grained metagabbros are well foliated, in which plagioclase forms lensoid, fine-grained aggregates. Nevertheless, some of the fine-grained metagabbros still partially preserve their original granoblastic texture (Plate II-2). The characteristic mineral assemblage consists of green amphibole, clinopyroxene, plagioclase, rutile or titanite, scapolite, garnet, apatite, ilmenite, epidote and opaques. The clinopyroxenes are overgrown by a corona of green amphibole. Plagioclase occurs as recrystallised fine-grained sodium-rich plagioclase aggregates. Skeletal or acicular epidote partly replaces coarse-grained plagioclase. Primary scapolite is often related to green amphiboles, whereas secondary scapolite grows at the expense of plagioclase. Ilmenite is often associated with remnants of clinopyroxene. Common features of the fine-grained metagabbros are titanite coronas which partially rim rutile and ilmenite. In some cases rutile is completely replaced by titanite which occurs as euhedral crystals or as aggregates. In samples where the metagabbro contains garnet, a symplectitic corona of plagioclase and green to colourless amphibole is formed around the garnets.

4.1.1.5 Coarse-grained metagabbros

In comparison with other gabbroic rocks, the coarse-grained metagabbros in the main part of the Mavuradonha Mountain range contain relatively large (up to several cm large) green pyroxene crystals. These pyroxenes are surrounded by a corona consisting of black amphibole and occur within a matrix of coarse-grained plagioclase. The primary mineral assemblage is clinopyroxene, orthopyroxene and plagioclase. The distribution of these phases is variable and depends on the sequence in which they are exposed within the complex. Metagabbros in the central part often contain orthopyroxene, whereas metagabbros in the outer parts consist solely of clinopyroxene and plagioclase. The composition of the coarse-grained metagabbros ranges from noritic, with nearly equal proportions of orthopyroxene and clinopyroxene, to gabbroic, with clinopyroxene as the single mafic mineral. Additional minerals are rutile, ilmenite, spinel, zircon, scapolite, epidote, garnet and quartz. Plagioclase often forms coarse crystals, which are partly recrystallised at the rims, forming a mortar texture. Another common textural feature of some of the coarse-grained metagabbros is the occurrence of small orthopyroxene crystals within large clinopyroxene crystals, locally associated with a corona of colourless amphibole or biotite. Typical for this rock type are the different corona textures around pyroxene and plagioclase. A common corona type is the symplectitic intergrowth of

colourless amphibole and quartz. This first corona is surrounded by a second corona of symplectitic intergrowth of two different types of green amphibole (Plate II-3). A third corona, which is limited only to a few samples, is a scapolite corona between plagioclase and the corona of green amphiboles (Plate II-4). Despite the above described corona textures, all samples show a polygonal granoblastic texture.

A fourth corona texture is the formation of garnet around either a symplectitic amphibole-quartz corona or between clinopyroxene and plagioclase. This corona type is often observed in sheared portions of the coarse-grained metagabbros. Inclusions of bluish-green amphibole, quartz and scapolite within the garnet corona are also observed. Further significant textural features are signs of intense deformation, indicated by deformed plagioclase twins, rotated pyroxene and bent exsolution laminae within the pyroxene. These exsolution laminae in clinopyroxene consist of orthopyroxene. In some samples the exsolution laminae in clinopyroxene consist of amphibole.

4.1.1.6 Garnet-bearing metagabbros

The occurrence of garnet-bearing metagabbros is limited to the area north of the Mavuradonha Mission. The garnet-bearing metagabbros (Plate II-5) are associated with meta-anorthosites in the upper meta-anorthosite suite. The major mineral assemblage consists of garnet, clinopyroxene, plagioclase, scapolite, quartz, rutile and ilmenite. Garnet is either nearly euhedral, with inclusions of pyroxene, plagioclase, rutile and ilmenite, or anhedral to subhedral with fewer or no inclusions. Both co-exist with the above described mineral assemblage. In some metagabbroic samples a fine-grained pyroxene rim, recrystallised between garnet and earlier larger pyroxenes, is assigned to belong to a second pyroxene generation. All metagabbros are inequigranular, with polygonal to interlobate textures. Green hornblende occurs as small rims with an extent of <20 μm around pyroxene or replaces pyroxene. Epidote and secondary scapolite replace plagioclase. In samples where a second generation of pyroxene has grown, ilmenite forms a corona around rutile. Ilmenite is rimmed by green amphibole and plagioclase, and titanite replaces rutile and ilmenite.

4.1.1.7 Leuco-metagabbros and leuco-amphibolites

The leuco-metagabbro and leuco-amphibolites form the final stage of every layered sequence. These layers occur at variable scale in metagabbroic and in amphibolitic sequences. The leuco-metagabbros contain very coarse-grained plagioclase, which is partly recrystallised at the rims. This coarse-grained plagioclase is replaced by acicular epidote and scapolite. Leuco-metagabbros contain large clinopyroxene crystals, which display

similar corona textures as the coarse-grained metagabbros. Clinopyroxene is completely uraltitized in some samples. The former granoblastic equigranular texture is still recognisable, despite the recrystallisation and the corona textures.

Layers of leucocratic amphibolites consist mainly of fine- to medium-grained plagioclase and amphibole, together with euhedral garnet, titanite, apatite and zircon in some samples. The leuco-amphibolites exhibit an inequigranular, partly interlobate texture. Scapolite and epidote occur as retrograde phases. Plagioclase is partly recrystallised at the rims.

4.1.1.8 Meta-anorthosites

Meta-anorthosites occur at two localities in the Mavuradonha Mountains. The upper meta-anorthosite suite occurs next to the Zambezi escarpment and exhibits the mineral assemblage plagioclase + clinopyroxene + garnet $\pm$ quartz. In contrast, the lower meta-anorthosite suite in the central mountain range shows the assemblage plagioclase + epidote + hornblende.

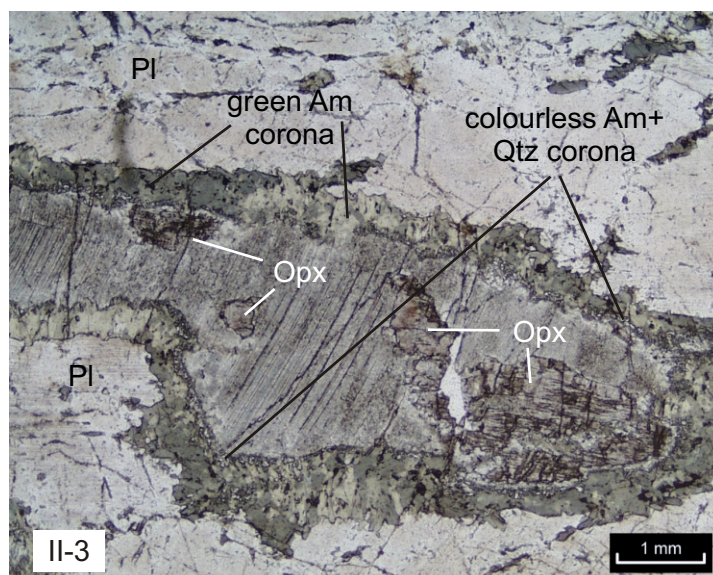
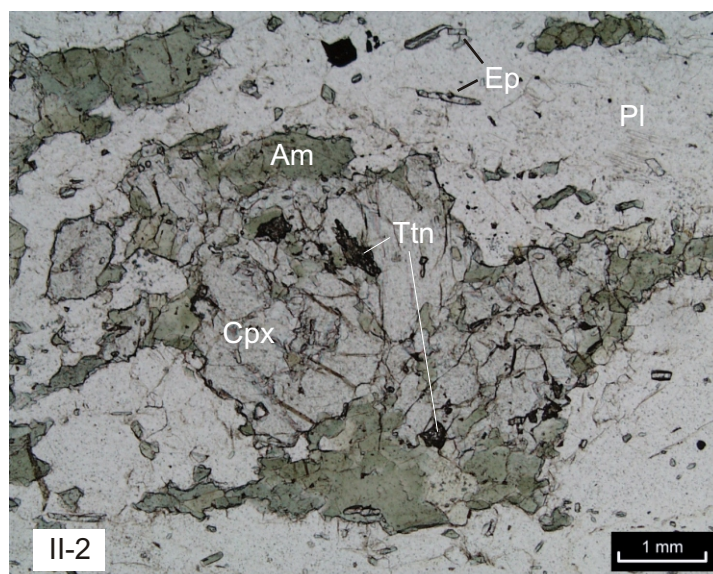
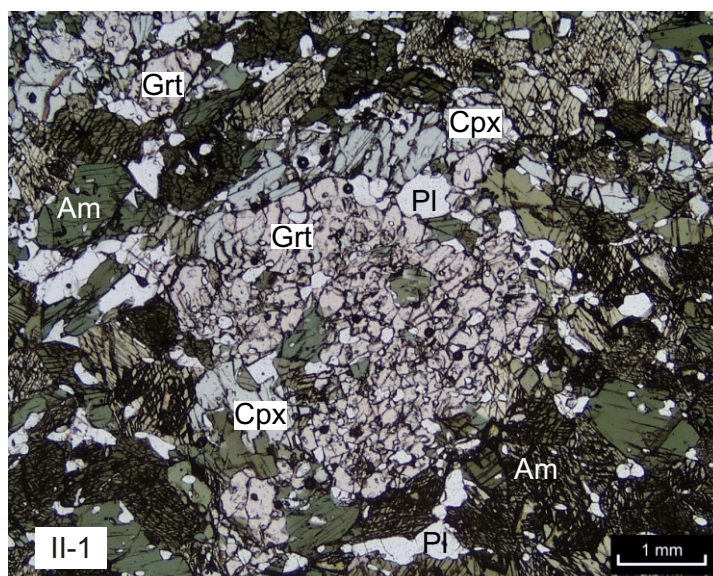
In the upper meta-anorthosite suite garnet and green clinopyroxene are concentrated and orientated in small bands within the plagioclase-dominated matrix displaying a 'spotted' texture (Plate II-6). Garnet and clinopyroxene often form 'augen' as a result of intense shearing. Commonly, the meta-anorthosites display a granoblastic texture, however, sheared samples are medium- to fine-grained. Garnet is nearly euhedral, containing inclusions of clinopyroxene, plagioclase and rutile. It coexists with clinopyroxene, plagioclase, scapolite, rutile, quartz and ilmenite. A common texture is the growth of small amphibole coronas between garnet and clinopyroxene. Additionally, clinopyroxene is sometimes replaced by green amphiboles. Further retrograde reactions are the replacement of plagioclase by acicular epidote and scapolite.

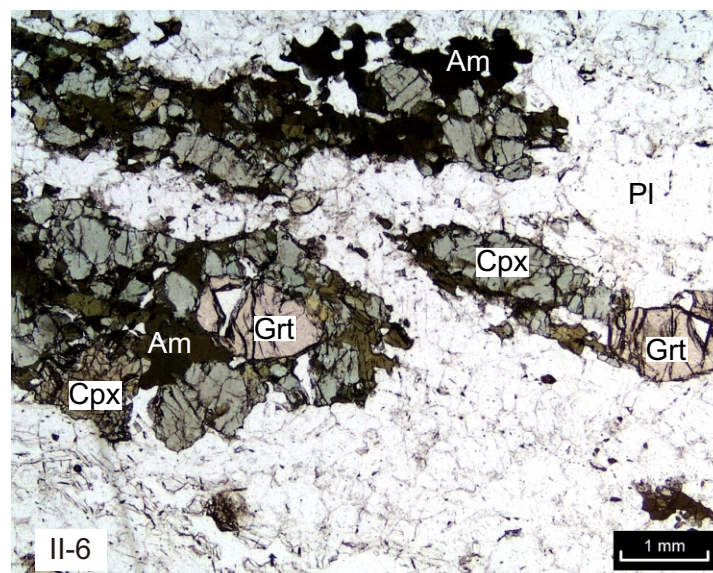
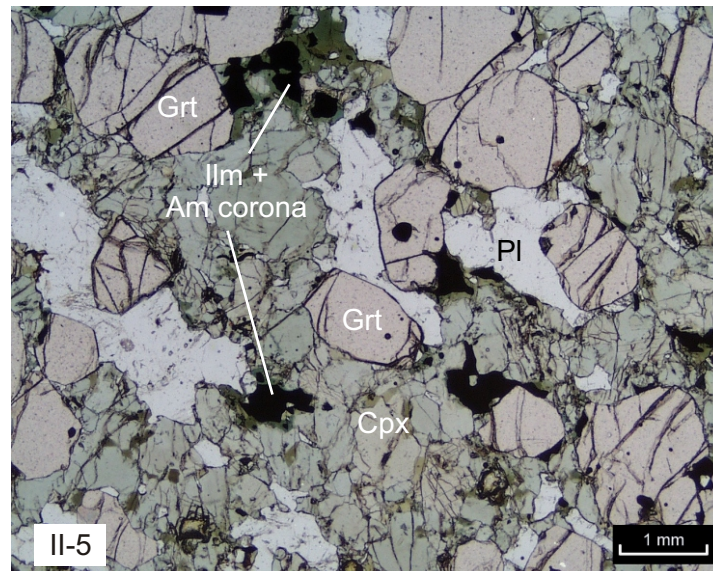
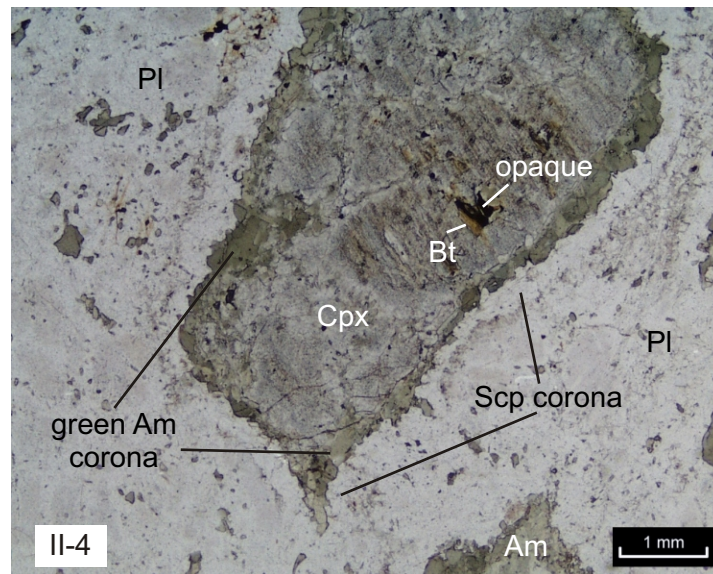
In the lower meta-anorthosite suite the meta-anorthosites are medium to fine-grained. Plagioclase coexists with hornblende, epidote, scapolite and titanite. Epidote forms euhedral crystals. The meta-anorthosites show an interlobate, inequigranular texture. A common textural relationship between mafic minerals and plagioclase is the 'mottled' appearance in which the mafic minerals occur as oikocrysts. Replacement of plagioclase by scapolite and the growth of small biotite and muscovite are commonly observed.

4.1.1.9 Pegmatites

The occurrence of pegmatites is widespread in the Mavuradonha Mountain range. The pegmatites are exposed in the Mavuradonha Layered Complex and in the underlying Ocellar Gneisses, crosscutting the primary igneous layering and truncate the foliation.

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- Plate II-1: Typical occurrence of a poikilitic garnet in a garnet-amphibolite with inclusions of plagioclase, green amphibole, quartz, epidote and scapolite. The photograph shows a garnet of garnet-amphibolite sample ZZB 41 in which relict clinopyroxene occurs in addition to green amphiboles and plagioclase. Thin section is shown in plain polarised light.
- Plate II-2: Slightly retrogressed clinopyroxene of fine-grained metagabbro sample ZZB 18 which clinopyroxene is replaced by green hornblende. Typical for this kind of rocks is the formation of titanite agglomerates replacing rutile and skeletal epidote in plagioclase. Thin section is shown in plain polarised light.
- Plate II-3: Corona texture around clinopyroxene in a coarse-grained metagabbro (sample ZZB 4). An inner corona of colourless, chlorine-free amphibole, quartz and an outer corona of chlorine-rich green amphibole are formed. Note also the orthopyroxene which occurs as variable sized crystals within the clinopyroxene. Thin section is shown in plain polarised light.
- Plate II-4: Coronas around clinopyroxene in a coarse-grained metagabbro (sample ZZB 15). An inner corona of chlorine-rich green amphibole and an outer corona of chlorine-rich scapolite is formed. Note also the formation of biotite next to the opaques. Thin section is shown in plain polarised light.
- Plate II-5: Typical occurrence of a garnet-bearing metagabbro (for example ZZB 175), in which clinopyroxene, garnet, plagioclase and opaques occur as rockforming minerals. Clinopyroxene is slightly replaced by green amphibole. Note the green amphibole corona textures around opaques. Thin section is shown in plain polarised light.
- Plate II-6: Common occurrence of a meta-anorthosite (ZZB 172) of the upper meta-anorthosite suite, with plagioclase, clinopyroxene and garnet as the major constituents. The concentration of clinopyroxene and garnet in thin bands leads to the spotted occurrence of the meta-anorthosites. Clinopyroxene is slightly retrogressed and rimmed by green amphibole. Plagioclase is often replaced by scapolite and/or epidote. Thin section is shown in plain polarised light.





The pegmatites usually exhibit amphibolite-facies mineral assemblages, and plagioclase, K-feldspar, quartz, amphibole, apatite, and zircon as major constituent can be observed. Additionally there is epidote, bluish hornblende, biotite, sericite and chlorite. Most of these minerals show a granoblastic, inequigranular texture.

Some pegmatites are intensively sheared and partly recrystallised. Hornblende forms slightly rotated porphyroblasts along shear-bands. A common texture is the replacement of apatite by allanite and epidote.

4.1.2 Nyamhanda Inlier and Chimwaya Hill Inlier

This part of the chapter provides a petrographic description of the three mafic rock types in the inliers. The samples were taken along cross-sections or when a marked change in the mineral assemblage was observed. Tables III-2 and III-3 in Appendix III give a complete mineral description of the investigated samples.

4.1.2.1 Amphibolites

Amphibolites occur in the Nyamhanda Inlier and in the Chimwaya Hill Inlier at the base of the metagabbros. On top of Nyamhandi Hill, a sheared contact between metagabbros and amphibolites is exposed. Both garnet-bearing amphibolites and garnet-free amphibolites occur in the inlier.

4.1.2.1.1 Garnet-bearing amphibolites

Garnet-bearing amphibolites in the Nyamhanda Inlier display a distinct layering of mafic to intermediate bands (plate III-1), whereas the rocks are not layered in the Chimwaya Hill Inlier. These garnet-amphibolites are exposed without these layered associations. Garnet porphyroblasts, ranging in grain size from several mm to 1 cm in diameter, are a common feature in garnet-bearing amphibolites (plate III-2). Garnets are poikilitic with inclusions of hornblende, plagioclase, quartz, titanite and scapolite. In most cases the inclusions are concentrated in the garnet core, whereas the rims are inclusion-free. Inclusions in these garnets show an orientation perpendicular to the foliation formed by small amphibole grains and plagioclase bands. In some samples garnets are slightly rotated. Garnets coexist with hornblende, plagioclase, scapolite, quartz, titanite, apatite and ilmenite. In one garnet-bearing amphibolite in the SW of the Nyamhanda Inlier zircon is present as an additional accessory mineral. Garnet-bearing amphibolites have seriate, polygonal to interlobate textures. Other textural relationships are shown by the growth of a titanite corona around ilmenite and by the separation of garnet from other ferromagnesian

phases by a plagioclase rim of variable size.

4.1.2.1.2 Garnet-free amphibolites

Garnet-free amphibolites are fine-grained and contain the typical mineral assemblage of green amphibole, plagioclase, quartz, titanite, zircon, scapolite, apatite and epidote. In both inliers the amphibolites show a pronounced layering. Some of the amphibolites show a decussate texture and are well foliated. In these samples remnants of rotated clinopyroxene with pressure shadows are observed. Unsheared amphibolites exhibit an inequigranular, polygonal texture. Light green amphiboles often contain numerous quartz inclusions. Another texture in the inlier amphibolites is the formation of tiny titanite bands associated with amphibole bands. Coarse-grained plagioclase occurs in felsic bands showing twinning and also as recrystallised grains. Plagioclase is often replaced by scapolite or overgrown by needle-shaped or skeletal epidote.

4.1.2.2 Metagabbros

The metagabbros in the inlier show distinct characteristics. Three distinct types of metagabbro are exposed in the Nyamhanda Inlier. In contrast to the Nyamhanda Inlier the metagabbros in the Chimwaya Hill Inlier have strong affinities to the Mavuradonha Layered Complex. These types of metagabbros in the inlier differ from each other by their mineral assemblage and occurrence in the field.

4.1.2.2.1 Chimwaya Hill Inlier

The major characteristic of the metagabbros in the Chimwaya Hill Inlier is the occurrence of green clinopyroxene within a matrix of medium-grained plagioclase. A common feature is the similar field occurrence as in the Mavuradonha Layered Complex because similar amphibole-corona textures are formed around the clinopyroxenes. The primary mineral assemblage of the metagabbros from the Chimwaya Hill Inlier is clinopyroxene and plagioclase. Plagioclase often forms large crystals, which are partly recrystallised at the rims. In one sample scapolite grows at the expense of plagioclase separating it from clinopyroxene and/or hornblende. Additional minerals are rutile, ilmenite, spinel, scapolite, epidote, garnet and quartz. The major difference between some Chimwaya Hill Inlier metagabbros and those of the Mavuradonha Layered Complex and Nyamhanda Inlier is the abundance of olivine (plate III-3).

Despite the evolved corona textures, the metagabbros still show a polygonal granoblastic texture. Commonly two corona types are formed. The first corona type is a

symplectitic intergrowth of colourless amphibole and quartz. The second is related to the formation of garnet between plagioclase and clinopyroxene and/or hornblende.

Evidence for intense deformation is provided by deformed plagioclase twins, rotated pyroxenes as well as bent exsolution laminae within pyroxene consisting of retrograde amphibole.

4.1.2.2.2 Nyamhanda Inlier

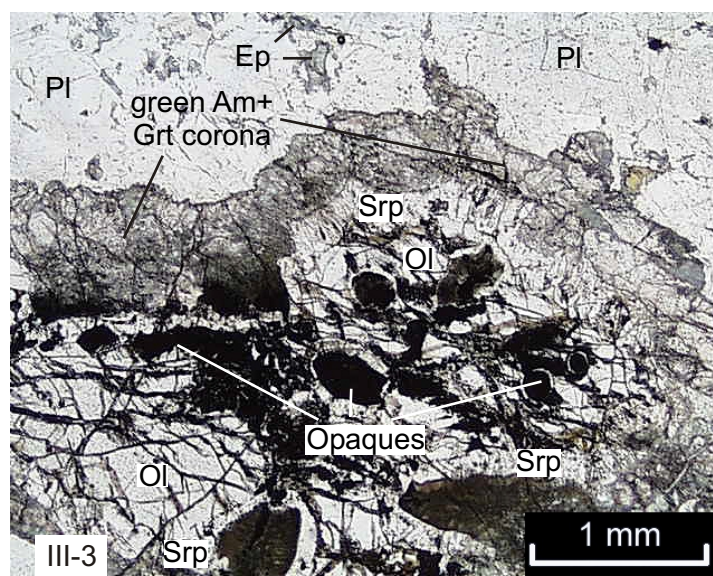
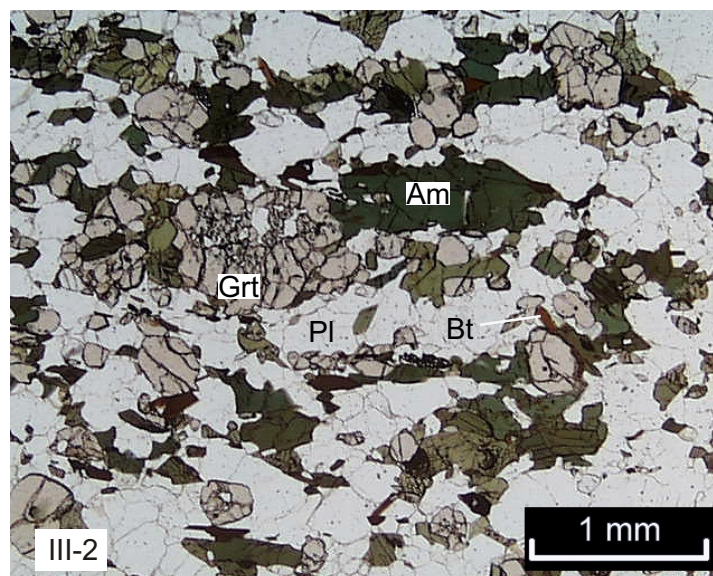
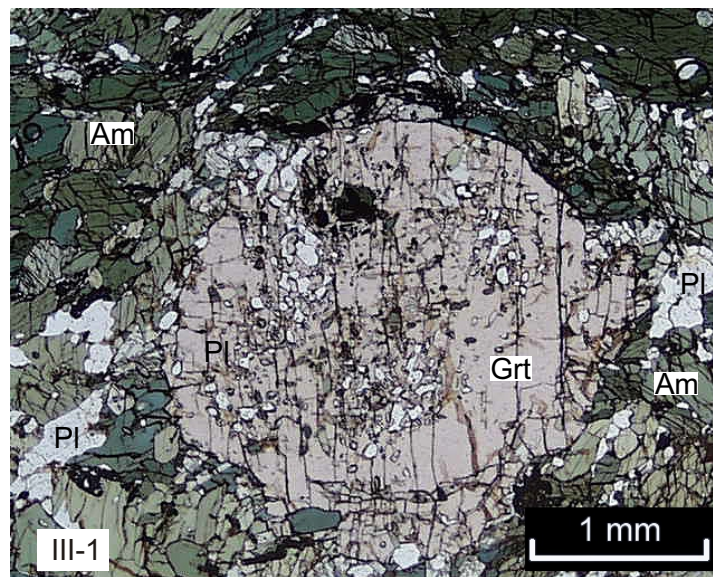
The metagabbros in the Nyamhanda Inlier show different characteristics in the field relative to the previously described metagabbros. Within the inlier different metagabbro types can be distinguished: One type is exposed in the NE where it forms a small hill within amphibolites (plate III-4 and III-5). The primary mineral assemblage is clinopyroxene, orthopyroxene and plagioclase. Olivine is reported (pers. comm. H. Jelsma, 1998) but was not found in thin section, because of strong retrogression of the sampled metagabbros. Chlorite and actinolite are often the major rock-forming minerals observed in outcrop. Further accessory minerals are rutile, ilmenite, spinel, scapolite, epidote, garnet and quartz. Garnet forms corona-like structures between plagioclase and pyroxene. Pyroxenes are often completely replaced by amphiboles and are uralitized. Cores of plagioclase are often almost completely transformed to sericite and epidote. Another common texture for this metagabbro type is the replacement of rutile by opaque minerals.

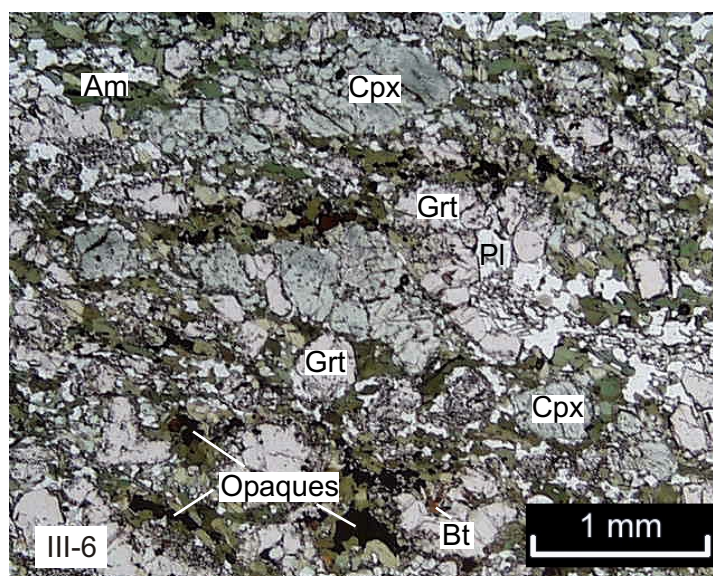
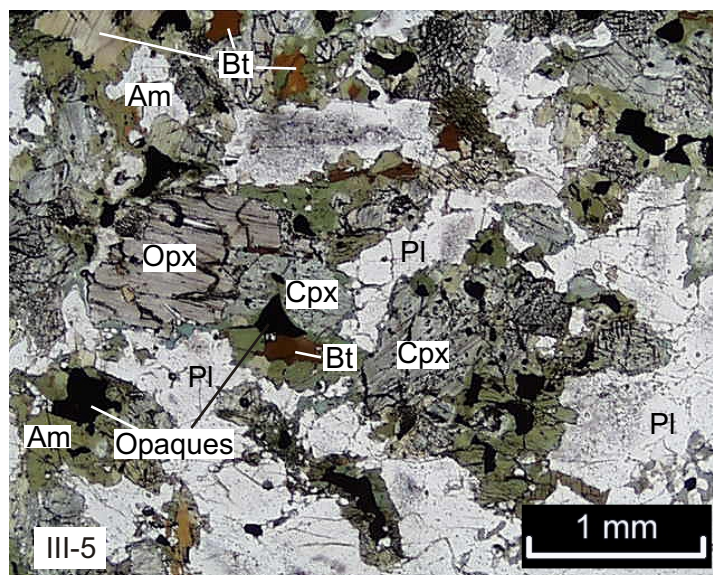
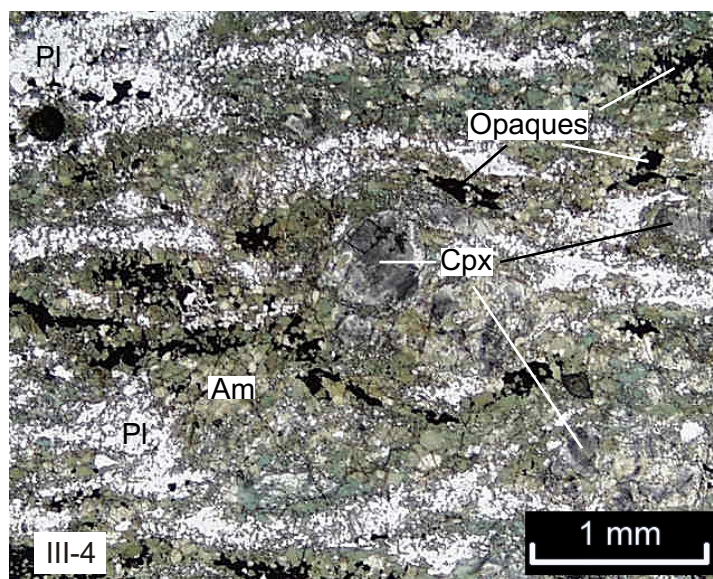
The second metagabbroic unit occurs in the central part and in the NW of the inlier. The primary mineral assemblage of clinopyroxene, orthopyroxene, plagioclase and opaques is only preserved in relics (plate III-6). Some metagabbros also show a subophitic texture (plate III-7). Another common texture for the metagabbro samples is the occurrence of small orthopyroxene crystals, included within large clinopyroxene. Typical for the clinopyroxene is a grid formed by rutile. Clinopyroxene and opaques are rimmed and replaced by brown amphibole or biotite (plate III-7). Also common is a corona of bluish-green amphibole and quartz at the expense of clinopyroxene. These amphiboles are often replaced by biotite. Garnet around amphibole and quartz forms a third corona separating the former pyroxene from plagioclase (plate III-8). Additionally, euhedral garnet is formed in plagioclase. Plagioclase is recrystallised at the rims and forms a mortar texture. In most samples plagioclase is replaced by scapolite, needle-shaped epidote and sericite. These metagabbros display an earlier polygonal granoblastic texture, which is in part preserved despite the occurrence of corona textures, recrystallisation and strong local retrogression.

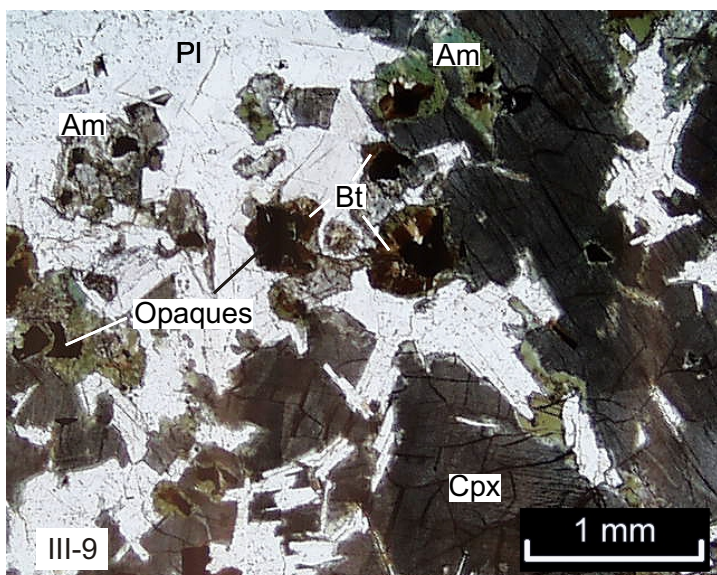
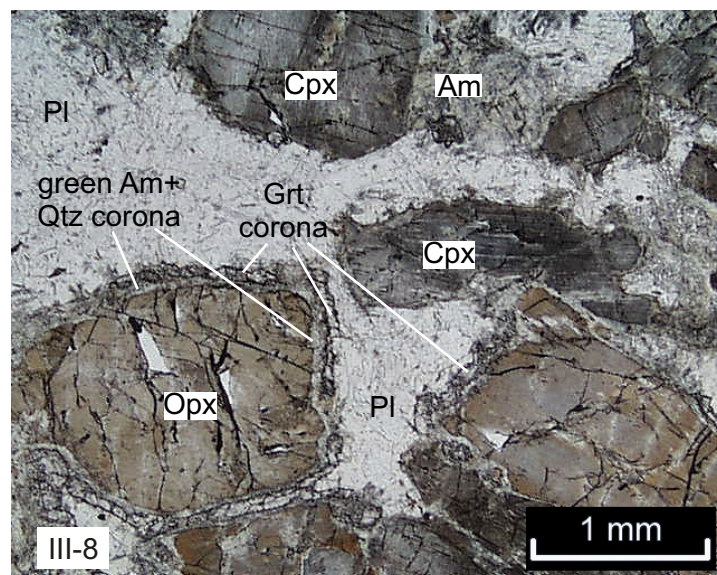
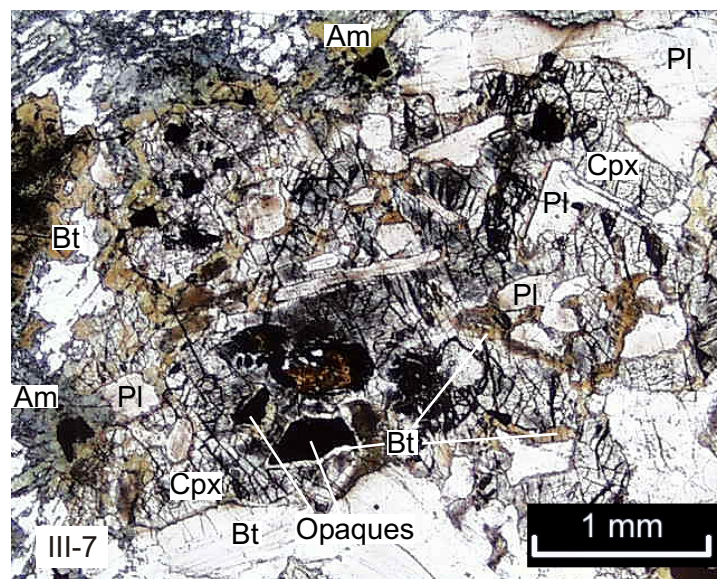
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- Plate III-1: Poikilitic garnet in coexistence with green amphibole and plagioclase in garnet-amphibolite sample ZCW 14 (Chimwaya Hill Inlier; appendix I, Fig. 1-5). Plagioclase, green amphibole, quartz, epidote and scapolite occur as inclusions in the garnet. Note also the orientation of the amphiboles enclosing the garnet in the upper part of the image. Thin section is shown in plain polarised light.
- Plate III-2: Garnet, together with plagioclase and green amphibole, occurs as major rockforming- mineral in garnet-amphibolite in the Nyamhanda Inlier (sample DKZ 53; appendix I, Fig. 1-5). Typical for this kind of rock is the occurrence of Fe-Mg-silicates in small bands which leads to the foliation in the garnet-amphibolites. Thin section is shown in plain polarised light.
- Plate III-3: Corona around serpentinised olivine in an olivine-bearing metagabbro from the Chimwaya Hill Inlier (sample ZCW 6; appendix I, Fig. 1-5). A corona of green amphibole and garnet is formed around between olivine and plagioclase. Note also the opaques within olivine that are surrounded by a corona of serpentine. Plagioclase is partly replaced by needle-shaped epidote. Thin section is shown in plain polarised light.
- Plate III-4: Relic clinopyroxenes in retrogressed metagabbro in the Nyamhanda Inlier (sample DKZ 1; appendix I, Fig. 1-5). Amphibole and opaques are concentrated in small bands, and their orientation defines a foliation in these amphibolite- to greenschist-facies metagabbros. The orientation of the minerals also mimics the original shape of the former clinopyroxene. Thin section is shown in plain polarised light.
- Plate III-5: Noritic metagabbro of the Nyamhanda Inlier (sample DKZ 4; appendix I, Fig. 1-5), in which clinopyroxene, orthopyroxene, plagioclase and opaques occur as rock-forming minerals. Clinopyroxene and orthopyroxene are overgrown by green amphibole and/or biotite. Note green amphibole corona textures around opaques and replacement of biotite by green amphibole. Thin section is shown in plain polarised light.
- Plate III-6: Strongly retrogressed garnet-bearing metagabbro (Nyamhanda Inlier, sample DKZ 16; appendix I, Fig. 1-5) with the primary mineral assemblage clinopyroxene, garnet and plagioclase. Clinopyroxene is replaced by secondary green amphibole, and opaques are surrounded by a corona of green amphibole. Plagioclase is often replaced by scapolite and/or epidote. Thin section is shown in plain polarised light.
- Plate III-7: Subophitic texture in metagabbro of the Nyamhanda Inlier garnet-bearing metagabbro (sample DKZ 12; appendix I, Fig. 1-5) consisting of clinopyroxene, plagioclase and opaques as major mineral phases. Clinopyroxene is replaced by green amphibole, and opaques are surrounded by corona of biotite or replaced by biotite. Thin section is shown in plain polarised light.

Plate III-8: Corona textures around clinopyroxene and orthopyroxene in a coarse-grained metagabbro (sample DKZ 22; appendix I, Fig. 1-5). In the case of orthopyroxene an inner corona of green amphibole and quartz and an outer corona of garnet are formed. Around clinopyroxene a corona of green amphibole and quartz is grown. Clinopyroxene is in some parts completely uralitised as shown by the amphiboles in the image. Thin section is shown in plain polarised light.

Plate III-9: Subophitic texture in a metagabbro in the Nyamhanda Inlier (sample DKZ 57; appendix I, Fig. 1-5). Clinopyroxene and plagioclase occur as rock-forming minerals. Clinopyroxene is partly replaced by green amphibole, which is also occurs as a corona around opaques. Biotite partly grew at the expense of opaques. Thin section is shown in plain polarised light.







The third metagabbro variety in the inlier shows a subophitic texture and cloudy plagioclase laths. This variety is preserved in the SE where it forms a smaller hill (plate III-9). Remnants of pyroxene are preserved, and clinopyroxene is often replaced by bluish-green amphibole. These amphiboles are frequently associated with quartz, titanite and biotite. A common texture around acicular opaques is the formation of an inner hornblende corona and an outer biotite corona.

4.1.3 Ocellar Gneiss

In thin section the grey to ochre coloured Ocellar Gneiss shows the same heterogeneity as in its field occurrence. A common mineral assemblage of the Ocellar Gneiss is biotite, bluish-green amphibole, garnet, plagioclase, potassium feldspar, muscovite, quartz, rutile, titanite, zircon, apatite and opaques. Biotite, epidote, allanite and chlorite occur as secondary phases. In thin section the Ocellar Gneiss generally exhibits a granoblastic interlobate texture except in samples near the contact with the tectonically overlying Mavuradonha Layered Complex. In these samples the Ocellar Gneiss is fine-grained to mylonitic. Further textural relations are given by the replacement of amphibole by biotite, by the formation of allanite and epidote coronas around apatite and by the formation of a titanite corona around rutile.

Bluish-green hornblende and plagioclase are the predominant phases in samples collected close to the Mavuradonha Layered Complex, whereas with increasing distance biotite and potassium feldspar become predominant. Commonly the mafic phases are well aligned and define a pervasive foliation in the Ocellar Gneiss.

Whereas in the southern part of the Mavuradonha Mountain range the Ocellar Gneiss exhibits amphibolite-facies characteristics, the small intercalated gneiss band west of the Mavuradonha Mission exhibits in parts high-grade relicts by the appearance of garnet and clinopyroxene.

4.2 Mineral chemistry of samples from the Mavuradonha Layered Complex

The major mineral phases of the different rock types of the Mavuradonha Layered Complex were taken to identify the mineral chemical composition. Garnet and clinopyroxene are studied in more detail, whereas plagioclase, amphibole and scapolite are only briefly discussed due to the lack of significant chemical variation. Representative mineral compositions of all analysed phases are given in Appendix III, Tables III-4 to III-10.

4.2.1 Garnet

Three different types of garnet are found in the Mavuradonha Layered Complex. The first garnet population occurs in the upper meta-anorthosite suite. Inclusions in these garnets are rare and consist of clinopyroxene, plagioclase and rutile. In contrast to the meta-anorthosites, where the garnets are concentrated in small bands, the associated garnet-metagabbros contain larger amounts of garnets. Garnet of the second population occurs as poikilitic grains in the garnet-amphibolites, containing clinopyroxene, green hornblende, plagioclase, scapolite, titanite and rutile as inclusions.

Garnet coronas around clinopyroxene or between clinopyroxene and plagioclase are assigned to be of the third population due to bluish-green hornblende, plagioclase and titanite as inclusions and growth of garnet-coronas often in association with shear-zones. Only garnet of the first two groups occurs as major rock-forming mineral.

Garnets from meta-anorthosites from the upper meta-anorthosite suite exhibit a compositional range from $\text{Alm}_{46-49}\text{Prp}_{20-24}\text{Gross}_{21-28}$ in sample ZZB 172 to $\text{Alm}_{30-32}\text{Prp}_{45-49}\text{Gross}_{14-19}$ in sample ZZB 176. Garnets from these samples are almandine-pyrope solid solutions with a minor spessartine component (Fig. 4-1). The grossular component shows only minor variation relative to its occurrence in the upper meta-anorthosite suite.

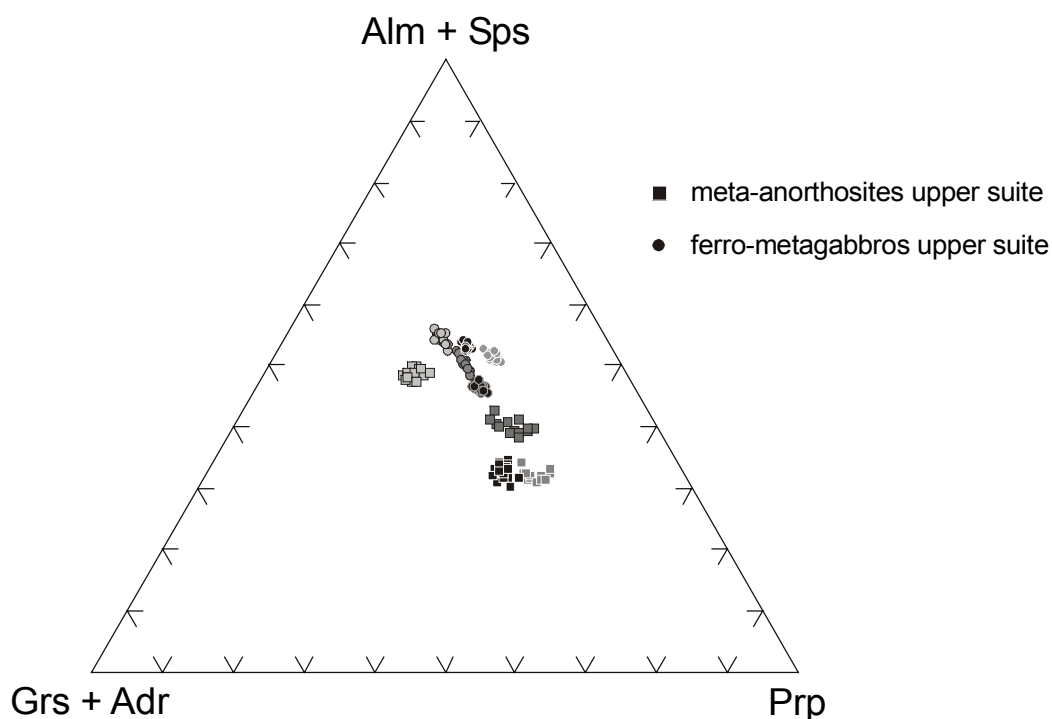


Figure 4-1. Chemical composition of garnets of the upper meta-anorthosite suite in a ternary plot of garnet endmembers.

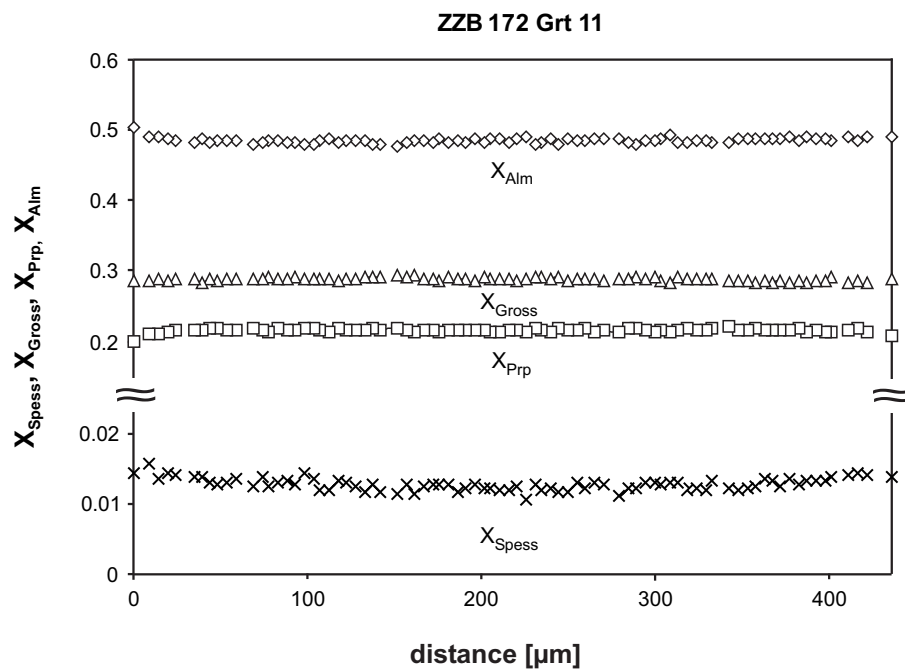
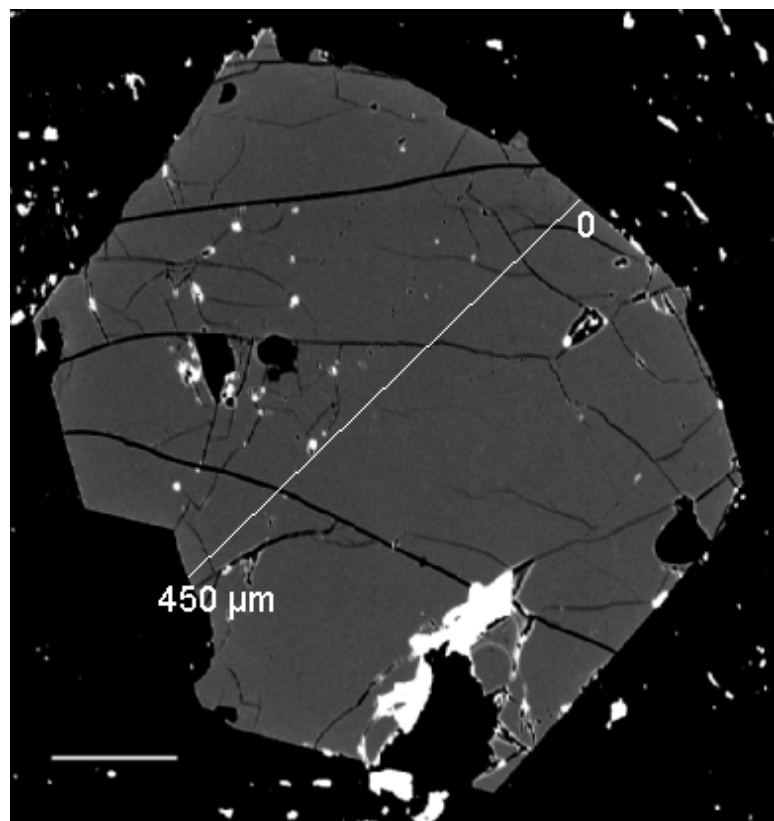


Figure 4-2 a. BSE image and chemical zonation of a garnet from meta-anorthosite sample ZZB 172 of the upper meta-anorthosite suite. Both are showing a nearly homogenous distribution of almandine, pyrope, grossular and spessartine components between core and rim. The chemical profile is indicated by the traverse in the BSE image across the garnet. Scale bar in the image is 100 μm .

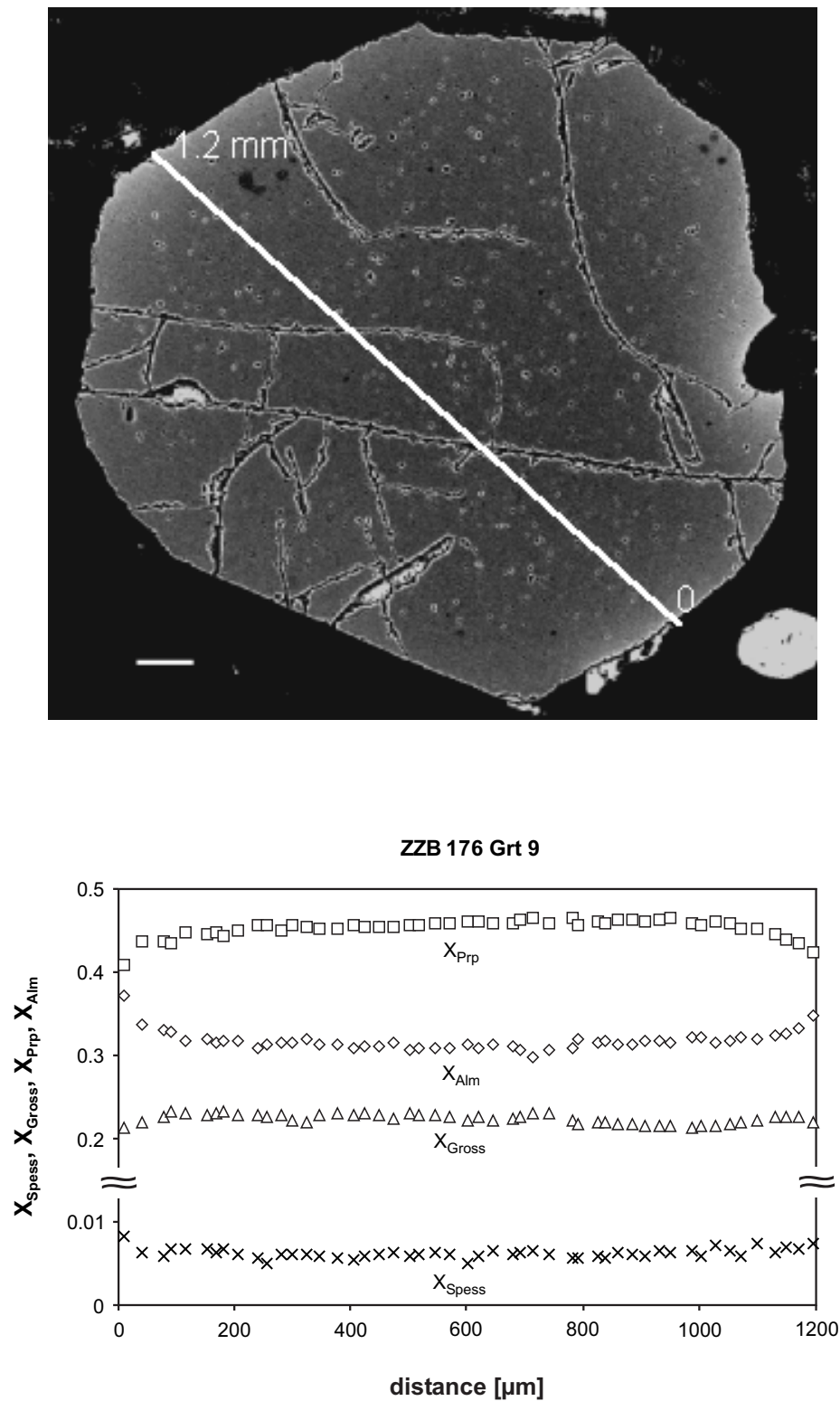


Figure 4-2 b. BSE image and chemical zonation of a garnet from meta-anorthosite sample ZZB 176 of the upper meta-anorthosite suite. Both are showing enrichment in the almandine component and depletion in the pyrope component in outer parts of the garnet. Note the compositional change compared to the previously shown garnets from meta-anorthosites. The chemical profile is indicated by the traverse in the BSE image across the garnet. Scale bar in the image is 100 μm.

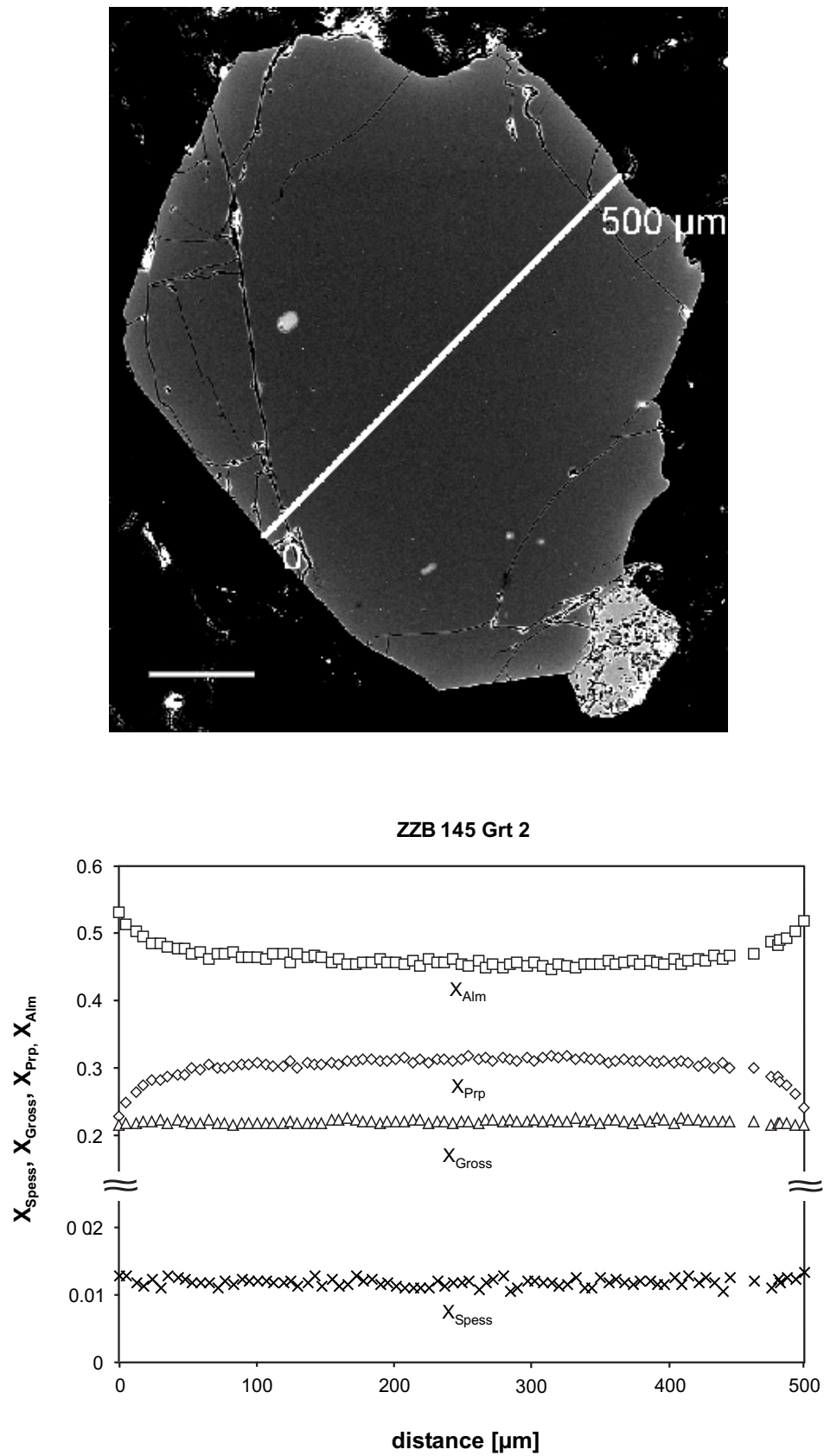


Figure 4-3 a. BSE image and chemical zonation of a garnet from ferro-metagabbro sample ZZB 145 of the upper meta-anorthosite suite. Both are showing enrichment in the almandine component and depletion in the pyrope component in outer parts of the garnet. The chemical profile is indicated by the traverse in the BSE image across the garnet. Scale bar in the image is 100 μm .

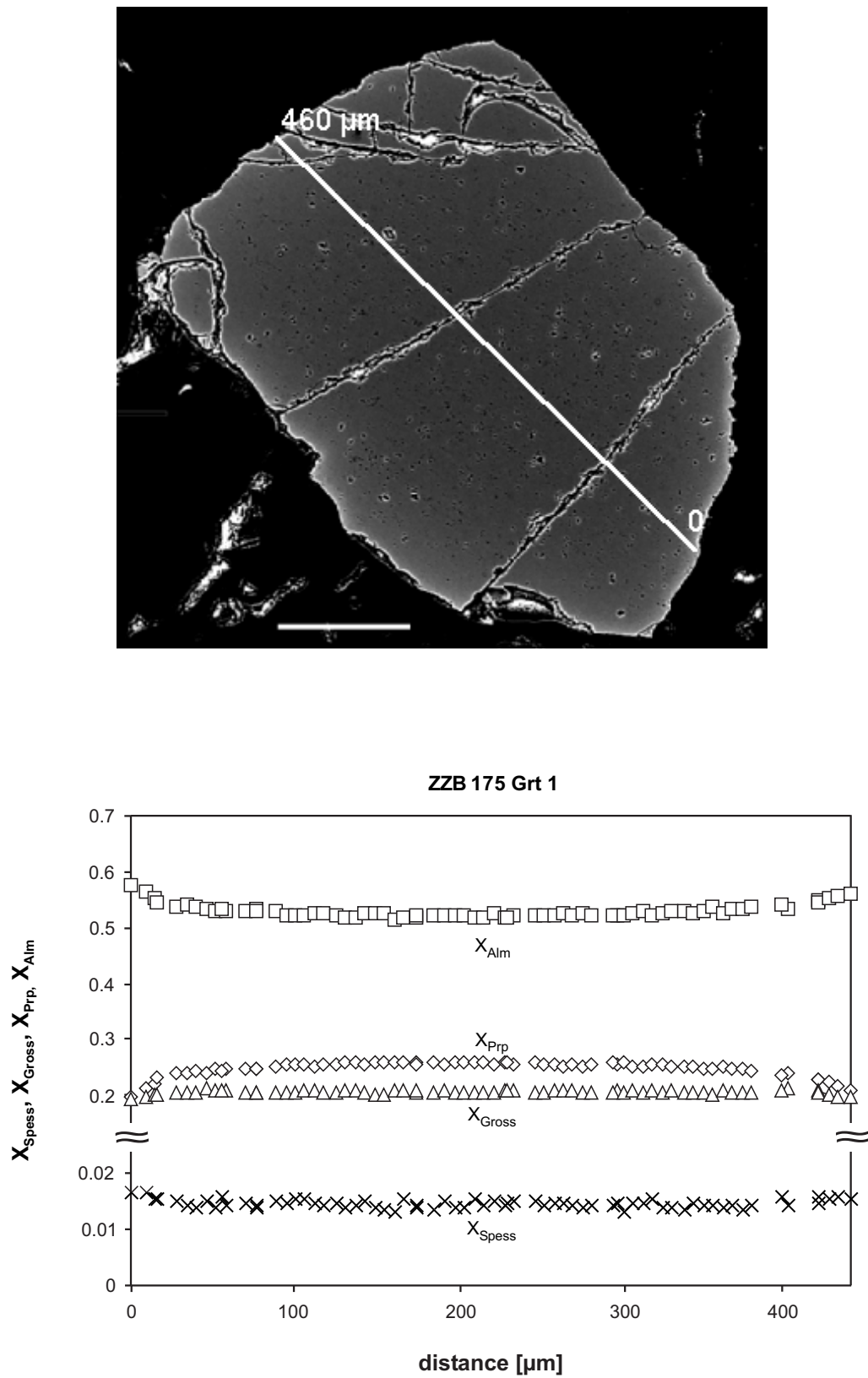


Figure 4-3 b. BSE image and chemical zonation of a garnet from ferro-metagabbro sample ZZB 175 of the upper meta-anorthosite suite. Both are showing enrichment in the almandine component and depletion in the pyrope component in outer parts of the garnet. The chemical profile is indicated by the traverse in the BSE image across the garnet. Scale bar in the image is 100 μm .

Zonation profiles (Fig. 4-2 a + b) of the garnets lack any prograde chemical growth zoning. In nearly all investigated profiles, with the exception of garnet in sample ZZB 172, a retrograde Fe-Mg exchange with adjacent amphiboles or clinopyroxene can be observed. This retrograde zonation rim extends up to 50 μm . Compared with the associated metagabbros the zonation profiles reflect larger variations in X_{Alm} and X_{Prp} . This heterogeneous appearance of the garnets probably reflects the abundance of other ferromagnesian minerals and, therefore, a local change of the bulk chemistry.

Garnets of the metagabbros are more homogeneous than those of the respective meta-anorthosite layer. The garnets exhibit a compositional range of $\text{Alm}_{44-46}\text{Prp}_{31-33}\text{Gross}_{16-19}$ (in ZZB 145) to $\text{Alm}_{51-56}\text{Prp}_{20-24}\text{Gross}_{16-24}$ (ZZB 153; Fig. 4-1). They are solid solutions of almandine-pyrope with minor grossular-spessartine components. The zonation profiles (Fig. 4-3 a + b) show a similar characteristic: the garnets exhibit high X_{Alm} values, the X_{Grs} component is always lower than the X_{Prp} . No prograde chemical zoning is visible within the zonation profiles. An iron-magnesia exchange with surrounding phases is visible by the retrograde zoning towards the rim. The extent of the retrograde zoning rims of ~ 100 μm are somewhat larger than those of the meta-anorthosites.

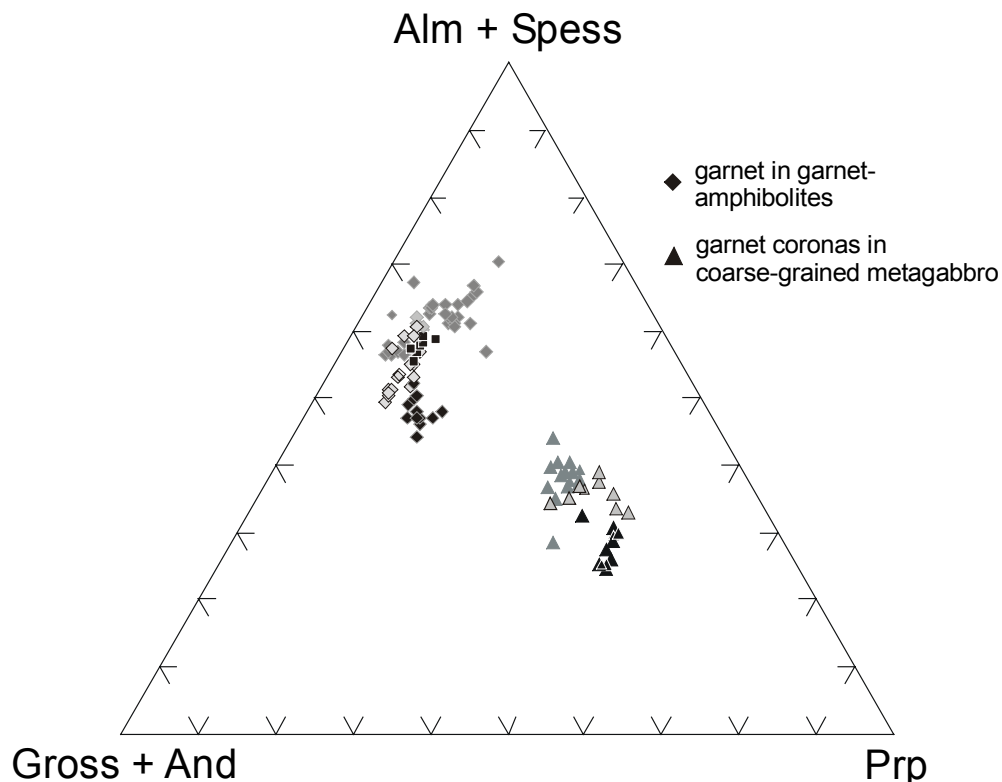


Figure 4-4. Chemical composition of garnets of the lower meta-anorthosite suite and in corona textures in a ternary plot of garnet endmembers.

Garnets from the amphibolites represent almandine-grossular solid solutions with a compositional range ranging from Alm₄₇₋₅₃Gross₂₉₋₃₄Prp₈₋₁₂ in sample ZZB 7 to Alm₅₄₋₆₃Gross₂₁₋₃₂Prp₄₋₁₁ in sample ZZB 28. Figure 4-4 shows the compositional variation of these garnets. The garnets contain only a minor pyrope and spessartine component.

The third group comprises garnets which form corona-textures in coarse-grained metagabbros. They are pyrope-almandine solid solutions with a minor grossular and spessartine component (Fig. 4-4), in which the pyrope-component is enriched relative to other garnet compositions. The composition ranges from Alm₂₃₋₃₂Pyr₄₃₋₅₀Gross₁₈₋₂₄ in sample ZZB 122 to Alm₃₄₋₃₉Gross₁₆₋₂₃Pyr₃₆₋₄₀ in sample ZZB 12.

4.2.2 Pyroxene

Both clinopyroxene and orthopyroxene occur in the Mavuradonha Layered Complex. Whereas the distribution of clinopyroxene is widespread, the occurrence of orthopyroxene is restricted to coarse-grained metagabbros.

4.2.2.1 Clinopyroxene

Clinopyroxene occurs in coarse-grained metagabbros, in metapyroxenites, in fine-grained metagabbros, in garnet-bearing metagabbros and meta-anorthosites in the upper meta-anorthosite suite of the Mavuradonha Layered Complex. Clinopyroxenes are diopside-hedenbergite solid solutions (Morimoto et al., 1988). Some of them show the presence of augitic components. In one garnet-amphibolite relic clinopyroxene has the same composition as the clinopyroxene in the garnet-bearing metagabbros. The clinopyroxenes show an increase in the hedenbergite component from metagabbros of the southern part of the Mavuradonha Layered Complex to the metagabbros and metapyroxenites of the north, next to the Zambezi Escarpment. High sodium and alumina contents are a common feature in the chemical composition of clinopyroxene of the upper meta-anorthosite suite. This feature is shown in the diagrams Al_{IV} + Al_{VI} vs. Mg + Si (Fig. 4-5 a) and Na vs. Ca + Mg (Fig. 4-5 b), and is related to the exchange of Tschermak- and jadeite-components in the calcic pyroxenes. In both exchange diagrams an increase of Al_{IV}Al_{VI} and Na is evident. These elements increase from low values in metagabbros of the layered gabbro series in the southern part of the mountain range to high values up to 0.4 Al_{tot} ions p.f.u. and 0.12 Na ions p.f.u. in the upper meta-anorthosite suite and metagabbros which are exposed northwest of the Mavuradonha Mission.

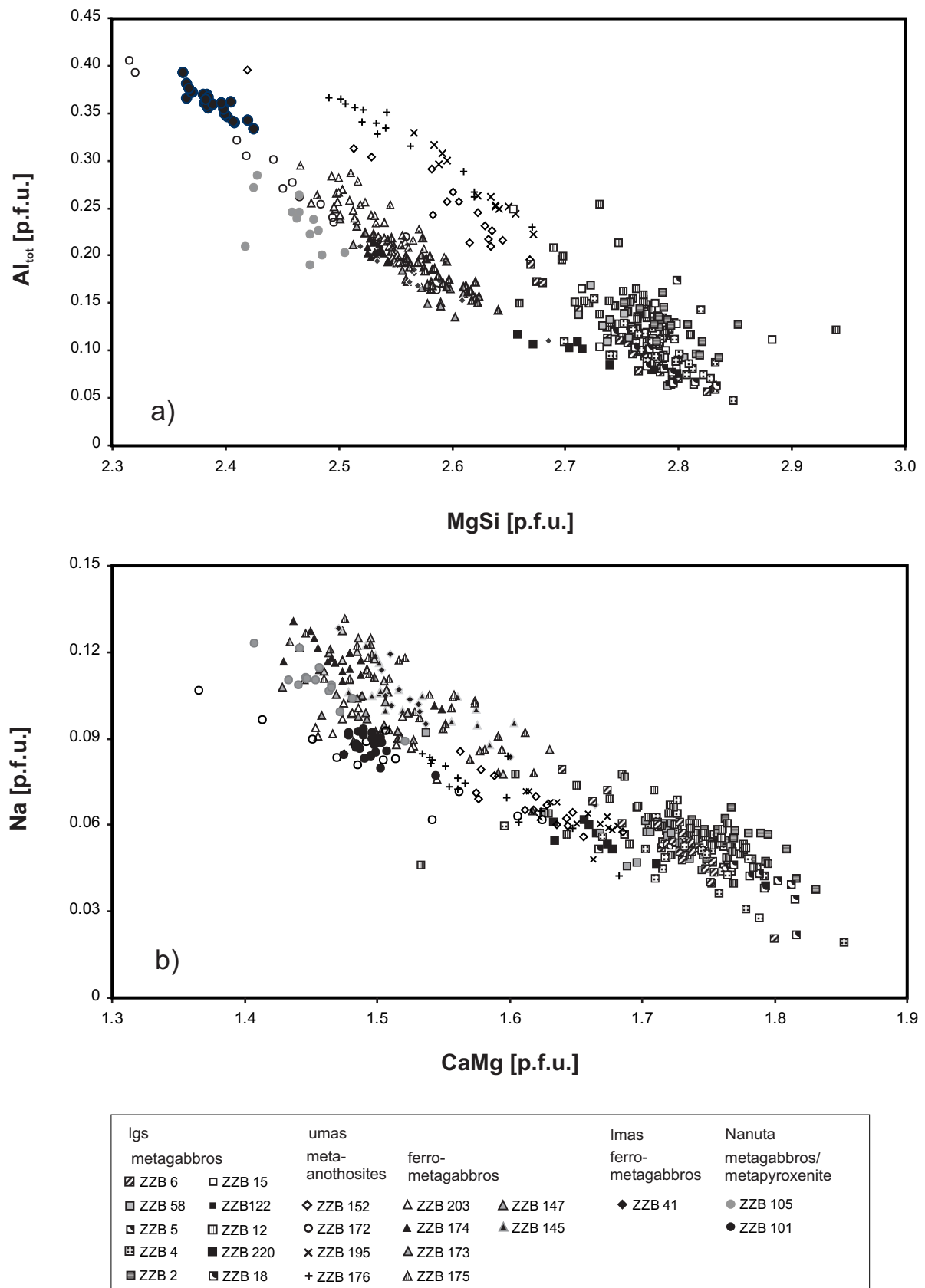


Figure 4-5 a and b. Tschermakite and jadeite exchange trends in clinopyroxene of the Mavuradonha Layered Complex. Lgs: layered gabbro series, umas: upper meta-anorthosite suite, lmas: lower

The zonation profiles for clinopyroxenes of all samples lack any growth zonation pattern (Fig. 4-6). A variable retrograde Fe-Mg exchange with surrounding ferromagnesian phases is visible in all investigated samples.

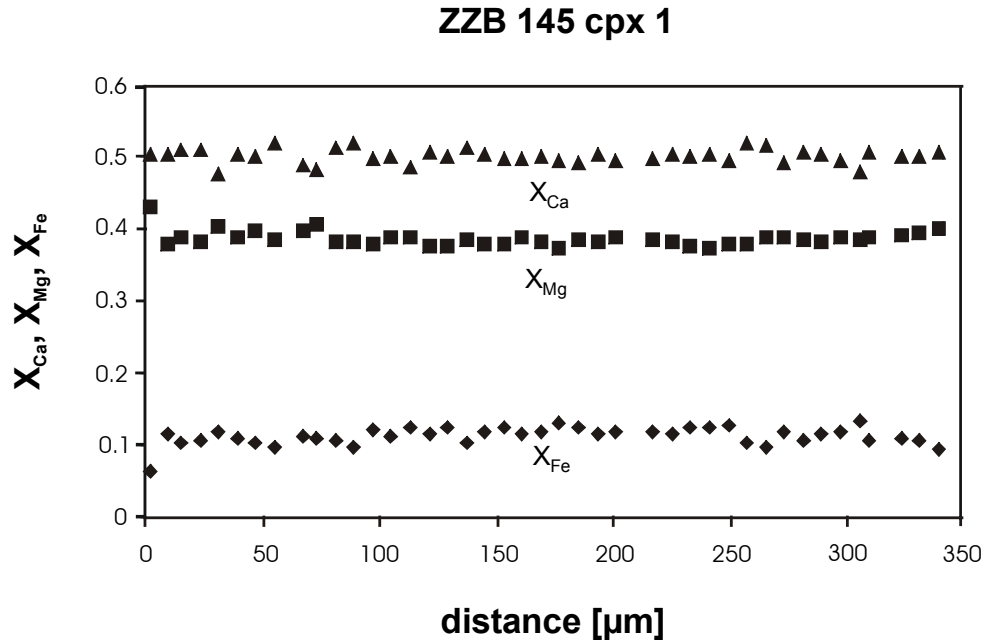


Figure 4-6. Chemical zonation profile of a clinopyroxene from garnet-bearing metagabbro sample ZZB 145 from the upper meta-anorthosites suite. Note the increase of X_{Mg} and the depletion of X_{Fe} in outer parts of the clinopyroxene, as well as the homogeneous distribution of X_{Ca} .

4.2.2.2 Orthopyroxene

Orthopyroxene is enstatite according to the classification of Morimoto et al. (1988). Orthopyroxene of two samples are distinguishable from each other and display their occurrence in the Mavuradonha Layered Complex. Sample ZZB 4 was sampled in the south and, therefore, in stratigraphical lower parts of the complex, the sampling location of sample ZZB 220 lies in the north close to the Mavuradonha Mission. This area belongs to higher a stratigraphic level. Whereas the orthopyroxenes of sample ZZB 4 are enriched in the enstatite component, the orthopyroxenes of ZZB 220 are enriched in the ferrosilite component.

4.2.2.3 Pigeonite

Pigeonite occurs either as exsolution lamellae in augitic clinopyroxene or as newly grown pyroxene between clinopyroxene and plagioclase.

4.2.3 Plagioclase

The chemical composition of plagioclase of the upper meta-anorthosite suite ranges from oligoclase to andesine (Fig. 4-7 a). The plagioclases in meta-anorthosites are predominantly andesine (An_{41} in ZZB 172 to An_{45} in ZZB 176). Another meta-anorthosite sample (ZZB 195) contains the most calcic plagioclase with an anorthite content of An_{69} .

Plagioclases of the associated metagabbros are transitional with An-contents ranging from An_{24} (oligoclase) in ZZB 167 to An_{35} (andesine) in ZZB 145.

The plagioclases of the lower meta-anorthosite suite (Fig. 4-7 a) show a larger scatter in their chemical composition than the plagioclase of the upper meta-anorthosite suite. The meta-anorthositic plagioclases in ZZB 160 and ZZB 162 have An-contents ranging from labradorite to bytownite (An_{54} to An_{87}). Plagioclase of the associated garnet-amphibolites shows a composition of An_{26} (oligoclase; Fig. 4-7 a).

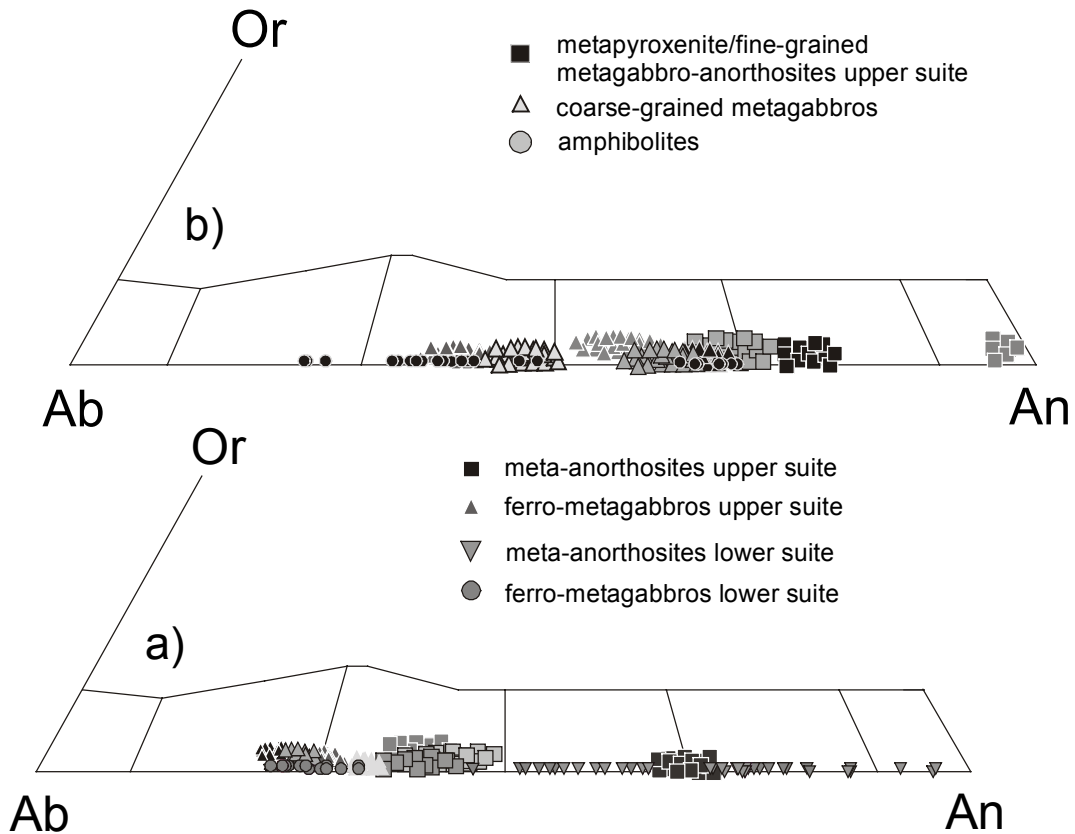


Figure 4-7 a and b. Chemical composition of plagioclase of the Mavuradonha Layered Complex in ternary plots of feldspar endmembers.

Plagioclases of other garnet-amphibolites are oligoclases and have similar An-contents ranging from An_{22} in ZZB 28 to An_{29} in ZZB 7. The variation in anorthite

component in amphibolites (Fig. 4-7 b) is somewhat larger, ranging from An₃₉ in ZZB 9 to An₆₄ in ZZB 8.

Nearly pure anorthite (An₉₇) was encountered in one metapyroxenite on the southern side of the Mavuradonha Mountain range. The plagioclase composition in fine-grained metagabbros is somewhat more sodic and is classified as bytownite (An₇₂₋₇₆). Plagioclase has a labradoritic composition in coarse-grained metagabbros with An-contents in the range of An₄₁ in ZZB 5 to An₆₇ in ZZB 4 (Fig. 4-7 b).

The compositional range of plagioclase in the Mavuradonha Layered Complex monitors the differences between the rock types and the occurrence within the layered sequence. Common to all samples is the relatively low orthoclase component of <2.4 %. Only minor variations in An-content are shown in the garnet-metagabbros and meta-anorthosites of the upper meta-anorthosite suite. Plagioclases of samples that re-equilibrated under amphibolite-facies conditions and in sheared rocks show a larger compositional range, e.g., the meta-anorthosites of the lower meta-anorthosite suite. The recrystallised plagioclases are more calcic in composition than the primary plagioclase.

4.2.4 Amphibole

Most amphiboles replace pyroxene. The green and brown amphiboles have a calcic amphibole chemistry according to Leake et al. (1997). The amphibole compositions in garnet-bearing amphibolites range from tschermakite to pargasite. Magnesiohornblende occurs in garnet-free amphibolites. Commonly these amphiboles are chlorine-free or contain only minor chlorine concentrations up to 0.5 %.

The chemical composition of amphiboles in corona-textures and as retrograde phases replacing clinopyroxene is more variable. Magnesiohornblende to tremolite occur only as inner coronas around pyroxene and always in association with quartz. The second amphibole group is widespread and classifies as ferroan-pargasites and replaces pyroxene or plagioclase. Amphiboles are mostly present as an outer corona surrounding pyroxene or occur in plagioclase in close relation to clinopyroxene. A general observation for these pargasites of the second corona and in plagioclase is that they possess chlorine contents up to 4.5 % whereas the chlorine concentrations in magnesiohornblende and tremolite (first, inner corona) reach values up to 0.6 %.

4.2.5 Scapolite

Scapolite is present in nearly all rock types, except metapyroxenites and the serpentinite. It occurs either as large grains or as coronas at the margin of plagioclase, and both types exhibit a homogenous composition in all samples. Common features of the chemical composition of the scapolites are high chlorine concentrations of up to 3 % as well as the presence of sulphate. High chlorine concentrations of up to 3 % in scapolite as well as the presence of sulphate lead to the stabilization of scapolites even under granulite-facies conditions (Newton & Goldsmith, 1975; Orville, 1975; Goldsmith, 1976; Goldsmith & Newton, 1977). Sulphate is a typical characteristic for scapolites of lower crustal, amphibolite- and granulite-grade regimes (Moecher & Essene, 1991). On the other hand, sodium- and chlorine-rich scapolites also indicate late phase tectonic activity in high-grade rocks (Moecher & Essene, 1991). The scapolites are classified according to their meionite content and equivalent anorthite content (Eq An, after Ellis 1978) as chlorine-rich marialites with a composition ranging between Eq An₂₀ to Eq An₃₆. This mineral composition is within the range of marialites reported by Kwak (1977). Considering the chemical composition of the scapolites no systematic variation can be observed.

4.3 PT evolution of the Mavuradonha Layered Complex

Chemical analyses of minerals which are interpreted to be in equilibrium with each other (Fig. 4-8) have been used in conjunction with geobarometers and geothermometers to estimate pressures (P) and temperatures (T) for metamorphic conditions.

In the following *PT*-estimates total iron is assumed to be ferrous iron as described in the appropriate part of chapter three which has the effect to get a consistent *PT* data set. Representative *PT* conditions of all analysed samples are given in Appendix III, Tables III-11 and III-12.

4.3.1 Granulite-facies *PT* conditions

Four ferro-metagabbros and one meta-anorthosite of the upper meta-anorthosite suite are taken for an estimate of granulite-facies *PT* conditions. These samples were chosen in order to get information about the *PT* evolution of different units of the upper meta-anorthosite suite. The mineral assemblages garnet-clinopyroxene-plagioclase-quartz and garnet-clinopyroxene were used for the calculation of *PT* conditions using core-core compositions of coexisting minerals.

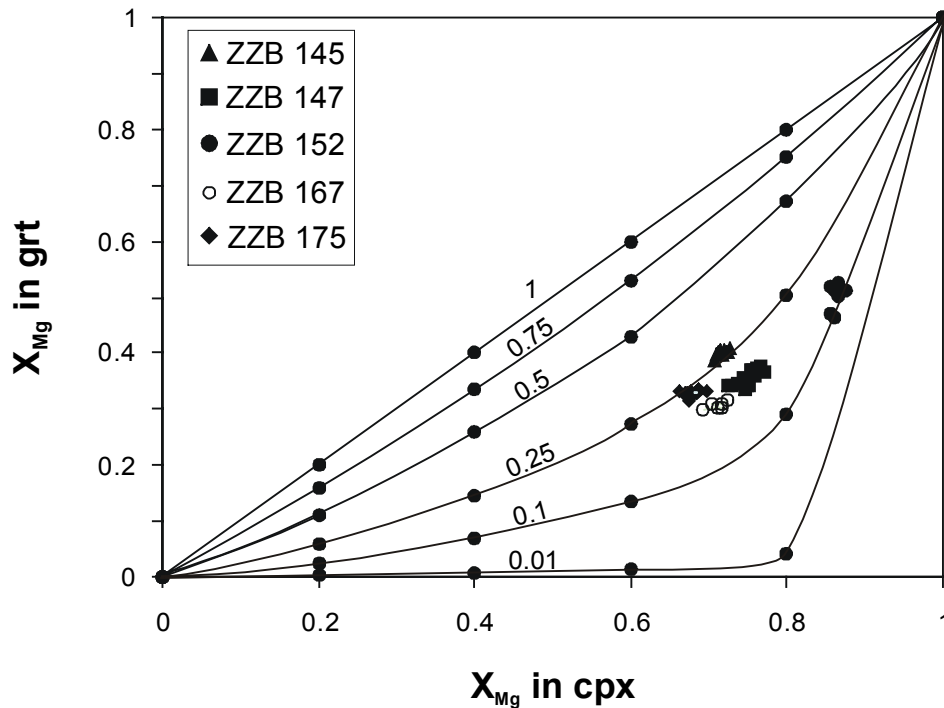


Figure 4-8. K_D -plot for meta-anorthosite and ferro-gabbro samples which are used for the calculation of peak granulite-facies conditions (after Spear, 1993). The slight variations in the spread of the data points are probably the result of an error propagation of analytical and calculation errors. The numbers on the K_D curves represent the K_D value and reflect the partitioning of Fe and Mg between garnet and clinopyroxene.

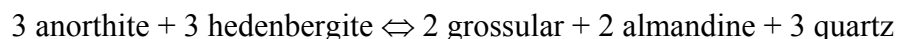
The compositions of the cores are assumed to represent peak-metamorphic conditions. Rim-rim pairs were not used in order to avoid retrograde diffusion effects occurring between rims of garnet and clinopyroxene because the observed retrograde Fe-Mg exchange is restricted to immediate contact zones between the mineral pairs as shown by the zonation profiles. It is unlikely that, within these zones, chemical equilibrium by net-transfer reaction is achieved. Secondly, tiny amphibole coronas around clinopyroxene are further signs for disequilibrium. Therefore, pressures and temperatures obtained from rim compositions are likely to be erroneous and, hence, the derivation of reliable data points on the retrograde PT -path is not possible (Selverstone & Chamberlain, 1990).

Garnet-clinopyroxene-plagioclase-quartz barometry (GADS)

To obtain pressures characterising peak-metamorphic conditions of the Mavuradonha Layered Complex, three geobarometers were applied to the samples. The calibrations of Newton & Perkins (1982) and Eckert et al. (1991) are based on the net-transfer reaction:

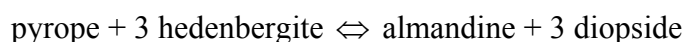


whereas the calibration of Powell & Holland (1988) uses a second net-transfer reaction:



Garnet-clinopyroxene thermometry

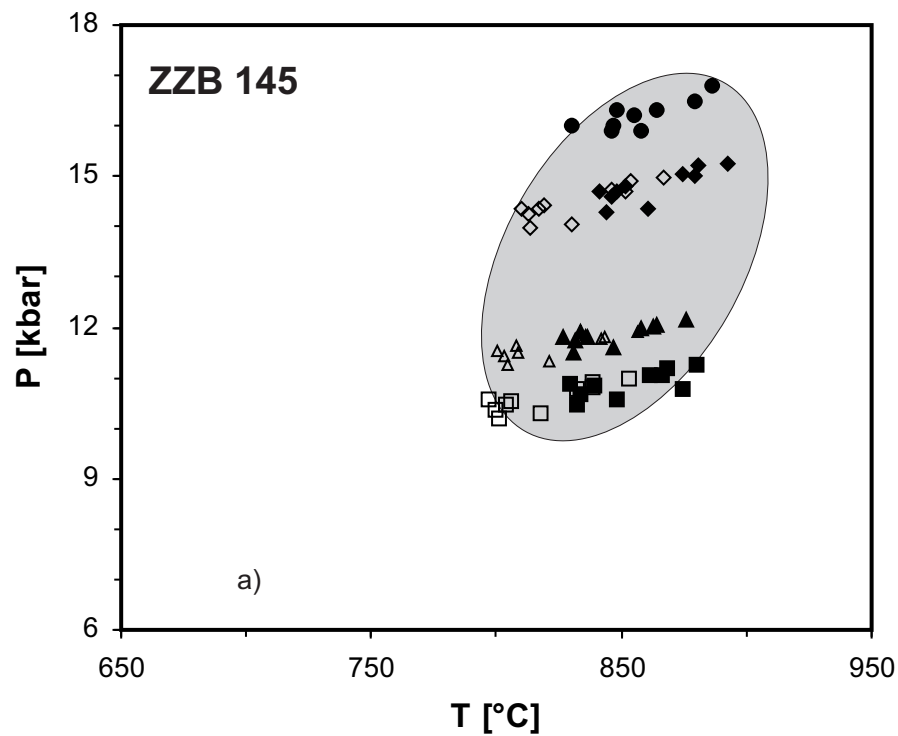
Garnet-clinopyroxene geothermometers are based on the exchange reaction of Fe^{2+} and Mg between both ferromagnesian mineral phases. Both calibrations applied in this study (Ellis & Green, 1979; Krogh, 1988) considered the dependence of the grossular content in their formula on the Fe^{2+} -Mg exchange reaction. This exchange reaction can be written as:



Summary of granulite-facies geothermobarometry:

In each of the five thin sections, cores of 5 to 20 pairs of adjacent garnet and clinopyroxene grains were analysed for the determination of maximum *PT* conditions. Calculated peak temperatures were in the range of 780 to 860 °C, and most of the results cluster within a range of 50 °C. The calibrated thermometers of Ellis & Green (1979) and Krogh (1988) work well in an interval of about 40 °C. Estimated pressures from core analyses range from 10 to 15.5 kbar for the temperature ranges estimated using the thermometers of Ellis & Green (1979) and Krogh (1988). Pressures deduced using the calibration of Eckert et al. (1991) vary around 15 ± 1 kbar, which are systematically higher than the two other calibrations. Pressures obtained using the calibrations of Powell & Holland (1988) and Newton & Perkins (1982) are in a range of 10 to 13 kbar and differ by 1.5 kbar in each section. Taking into account the error of each geobarometer calibration the results of these barometers were taken for derivation of peak pressures.

The highest peak metamorphic conditions were obtained for two ferro-metagabbros (ZZB 145 and ZZB 147). The peak conditions for these samples are 13 ± 2 kbar at 840 ± 30 °C. Estimated temperature conditions for further three samples (two garnet-metagabbros and one meta-anorthosite) are somewhat lower than those of the above samples and vary between 810 and 830 °C at pressures of 12 and 13 kbars. The calculated *PT*-conditions are displayed in Figures 4-9 a-e.



- ◆ Ellis & Green (1979) and Eckert et al. (1991)
- Ellis & Green (1979) and Newton & Perkins (1982)
- ▲ Ellis & Green (1979) and Powell & Holland (1988)
- Berman (1991) TWQ 2.01
- ◇ Krogh (1988) and Eckert et al. (1991)
- Krogh (1988) and Newton & Perkins (1982)
- △ Krogh (1988) and Powell & Holland (1988)

Figure 4-9 a-e. *PT* conditions of granulite-facies metamorphism for ferro-metagabbros and a meta-anorthosite, derived from thermobarometry on the assemblage garnet, clinopyroxene and plagioclase. Filled circles show the results of *PT* calculations using TWQ.

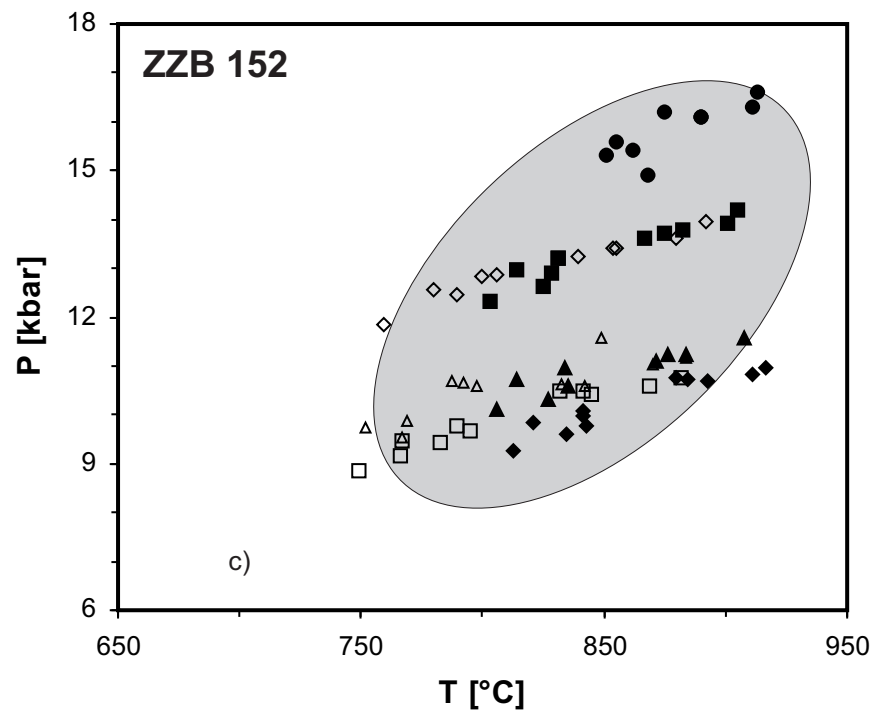
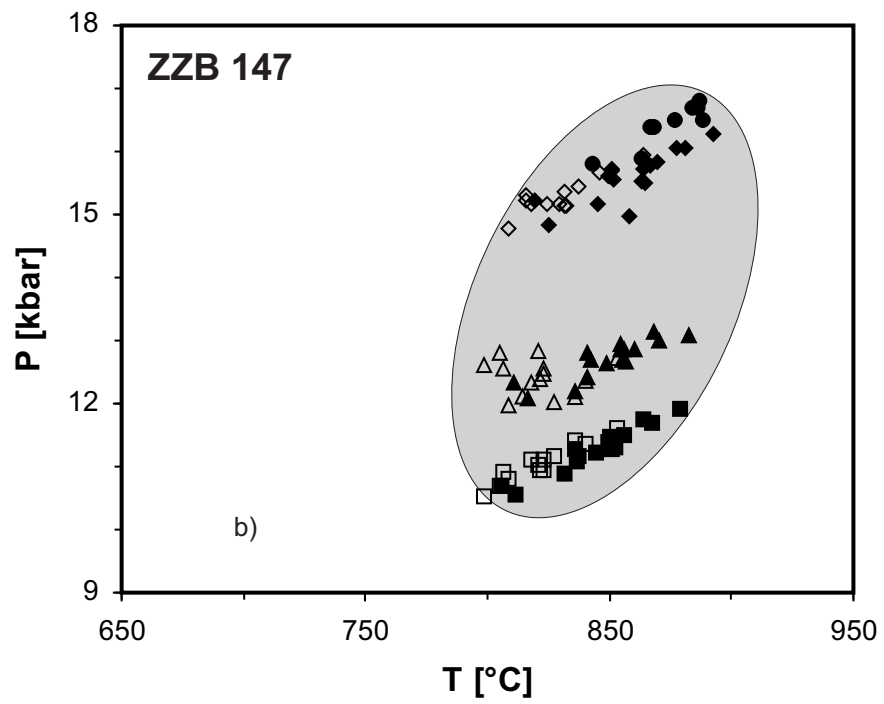


Figure 4-9 a-e. Continued.

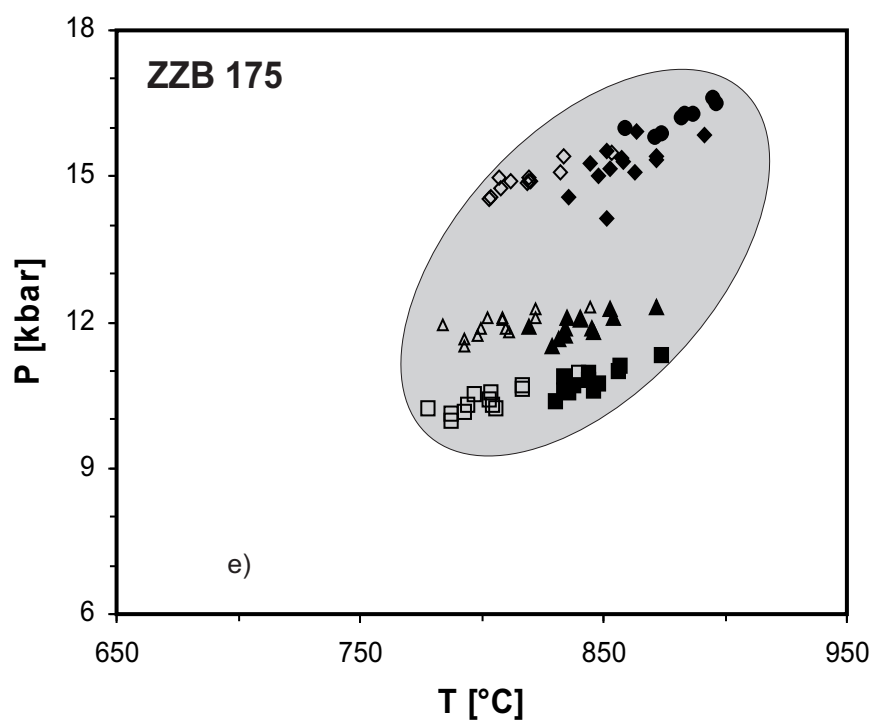
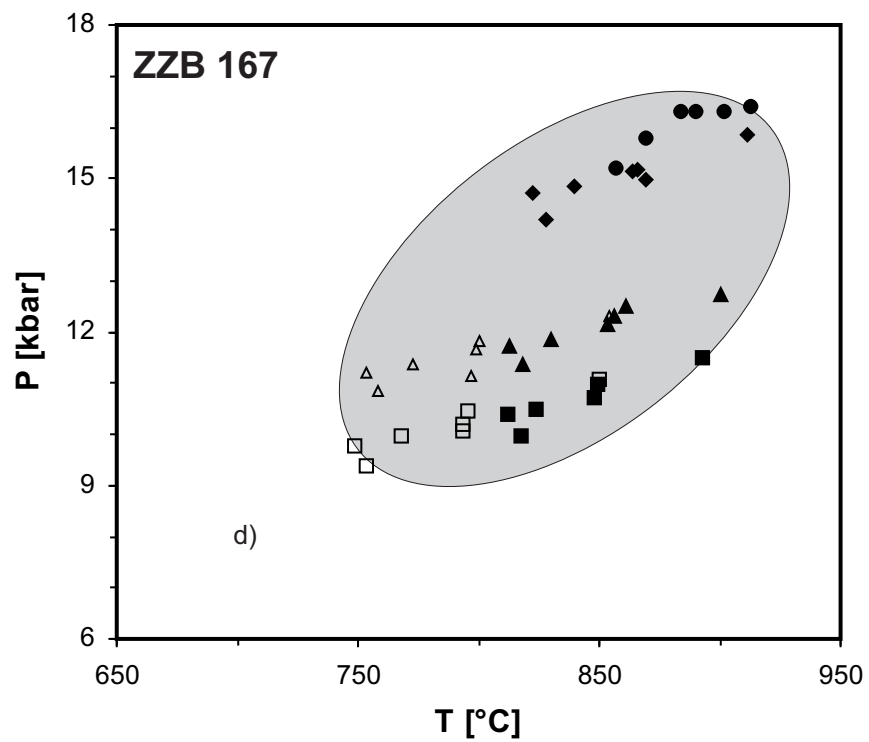


Figure 4-9 a-e. Continued.

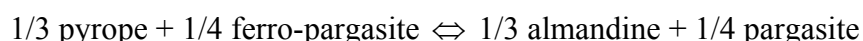
The use of $\text{Fe}_{\text{tot}} = \text{Fe}^{2+}$ yielded a consistent data set for all investigated samples. The calculation of ferrous iron and the use in thermobarometry have provided inconsistencies in the temperature data with major differences of 200 °C. These differences were probably the result of analytical errors (Canil & O'Neill, 1996; McCammon et al., 1998), which lead to underestimation and unreliable temperature conditions in most cases. Contrary to this, the temperature data estimated using Fe^{2+} only are more reliable and fall in the range of the *PT* data observed by Hargrove et al. (1998; 2003).

4.3.2 Retrograde *PT* evolution

Retrograde *PT* conditions were calculated for three garnet-amphibolites using core compositions of garnet-hornblende-plagioclase-quartz assemblages. These three garnet amphibolites were sampled in the southern part of the mountain range, samples ZZB 28 and ZZB 166 are part of the lower meta-anorthosite suite, ZZB 7 belongs to a garnet amphibolite - amphibolite association in the eastern mountain range. These garnet amphibolites were taken in order to obtain information on the retrograde *PT* evolution in different parts of the complex. In addition to the mineral pairs, amphibole and plagioclase inclusions within the garnets were used for *PT*-estimates.

Garnet-hornblende thermometry

The exchange reaction of the garnet-hornblende geothermometer of Graham & Powell (1984) is based on the exchange reaction of Fe^{2+} and Mg between both ferromagnesian mineral phases and can be written as:

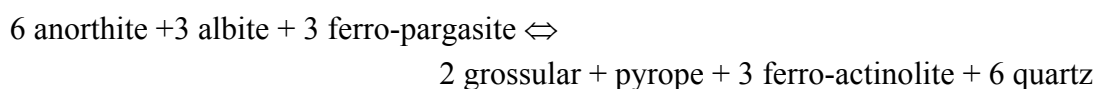


Garnet-hornblende-plagioclase-quartz barometry

The following net transfer reactions serve as the basis for the geobarometer which was empirically calibrated by Kohn & Spear (1989):



and



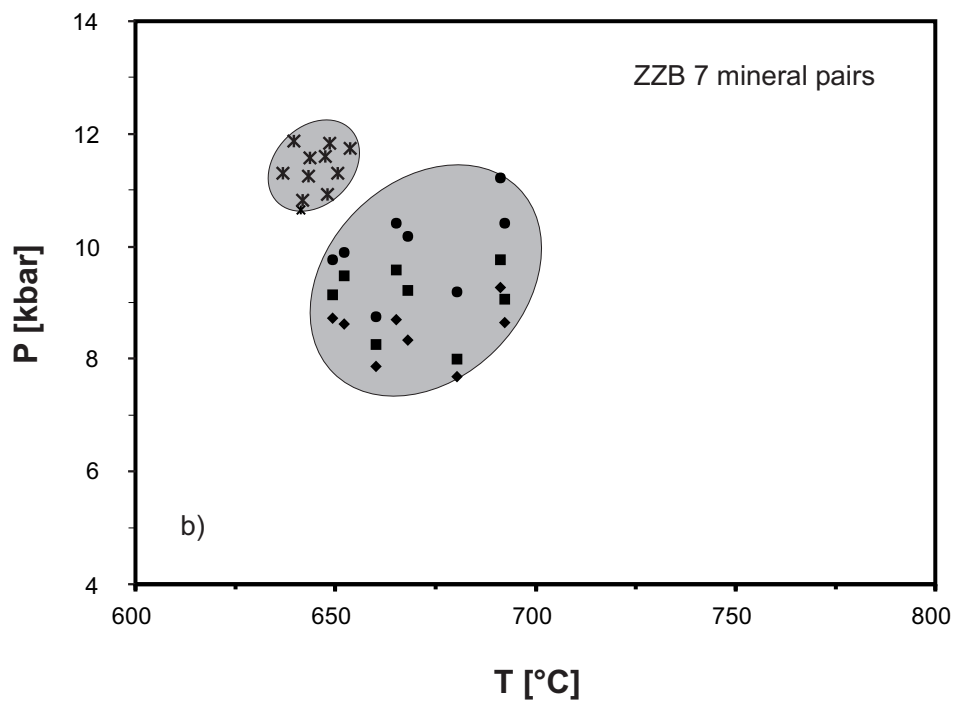
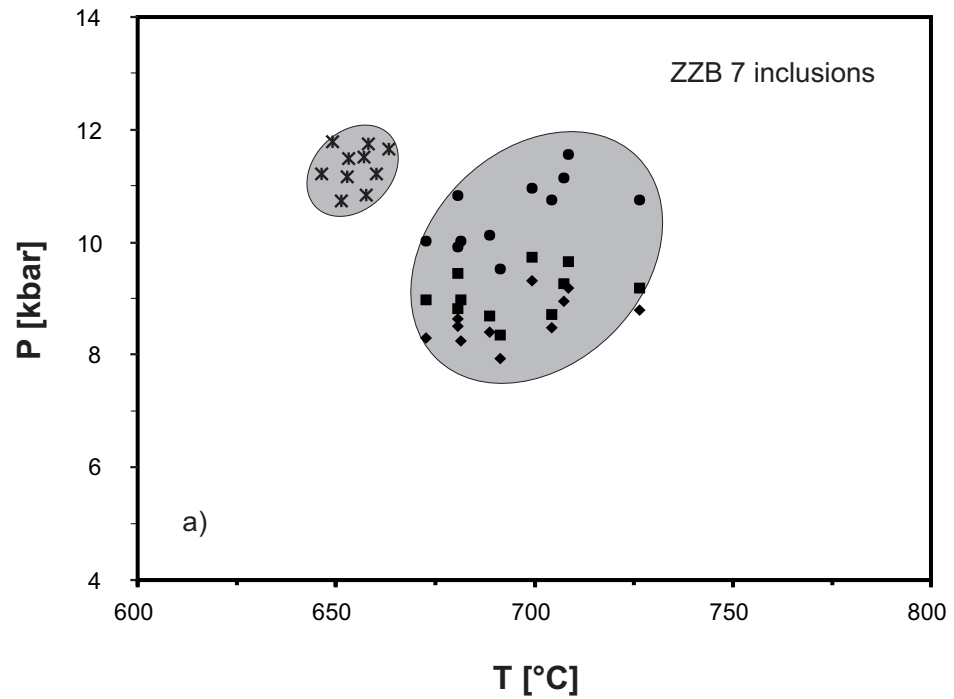
Summary of amphibolite-facies geothermobarometry

The geobarometer of Kohn & Spear (1989) yielded pressures of 8 to 13 kbar at given temperatures which were calculated using the Graham & Powell geothermometer (Figs. 1 4-10 a - d). The calibration of Graham & Powell (1984) yielded temperatures in the range of 660° to 730 °C. Temperatures and pressures of garnet amphibolite samples of the lower meta-anorthosite suite (ZZB 28 and ZZB 166; mean of 710 °C at 11 and 12 kbar) were slightly higher than those calculated for garnet-amphibolite sample ZZB 7 (~680 °C at 9.5 kbar). Inclusions of plagioclase and amphibole within garnet in sample ZZB 7 yielded temperatures and pressures which are indistinguishable from those estimated using touching mineral pairs. Considering the errors assigned to the calibrations, the estimated temperatures and pressures are similar and, therefore, a mean of 690 ± 20 °C at 10 ± 2 kbar for the retrogression under amphibolite-facies condition is estimated.

The use of total Fe^{2+} instead of ferrous iron yielded T-conditions which were ~50 °C higher than the ones with ferrous iron. Considering the errors of each thermometer, these T-conditions were indistinguishable. Therefore, it seems that the ferrous iron content did not have much influence on the calculated temperatures as has been shown for the garnet-clinopyroxene thermometers.

4.3.3 PT conditions calculated using TWQ

In addition to the previously described geothermobarometry, *PT* conditions were estimated using TWQ 1.02b and 2.02 (Berman, 1991) for garnet-clinopyroxene-plagioclase-quartz and garnet-hornblende-plagioclase-quartz mineral assemblages. The advantage of applying TWQ in addition to the independent calibration lies in the thermodynamic internally consistent data set. Two major differences are visible in all samples: The use of TWQ yielded consistent *PT*-estimates for all samples scattering in a range of ± 2 kbar and ± 50 °C. These *PT* conditions are slightly higher than the ones estimated using conventional geothermobarometry. The peak metamorphic *PT*-conditions estimated for all samples are 16 ± 1 kbar and 873 ± 20 °C. Calculation of retrograde *PT*-conditions for the garnet-amphibolites yielded a *PT*-range of 11 ± 2 kbar and 680 ± 25 °C. The *PT* conditions calculated by this computer program for sample ZZB 7 are outside the range for the *PT* conditions using conventional geothermometry, but if the errors in both methods are taken into account the conditions overlap within the error.



- | | |
|----------------------------------|----------------------------------|
| ◆ Kohn & Spear (1989) Mg model 1 | ▲ Kohn & Spear (1989) Mg model 2 |
| ■ Kohn & Spear (1989) Fe model 1 | * Berman (1991) TWQ 1.02 |
| ● Kohn & Spear (1989) Fe model 2 | |

Figure 4-10 a-d. *PT* conditions of retrograde metamorphism for garnet amphibolites, derived from thermobarometry of the assemblage garnet, hornblende and plagioclase. Temperatures were estimated using Graham & Powell (1984). The models given above refer to the barometric equations published in Kohn & Spear (1989).

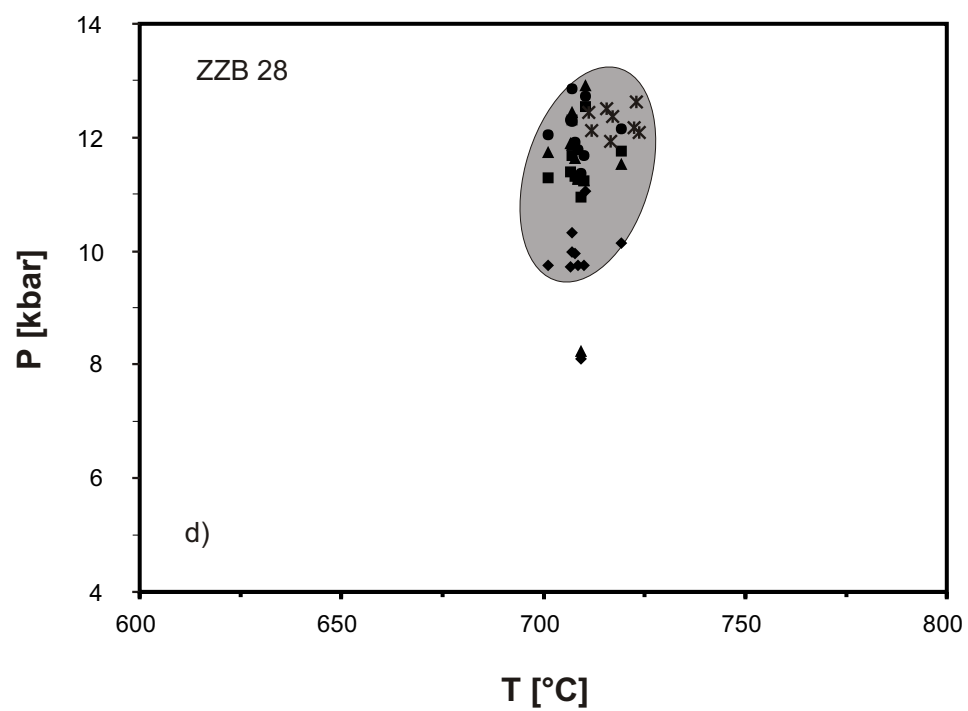
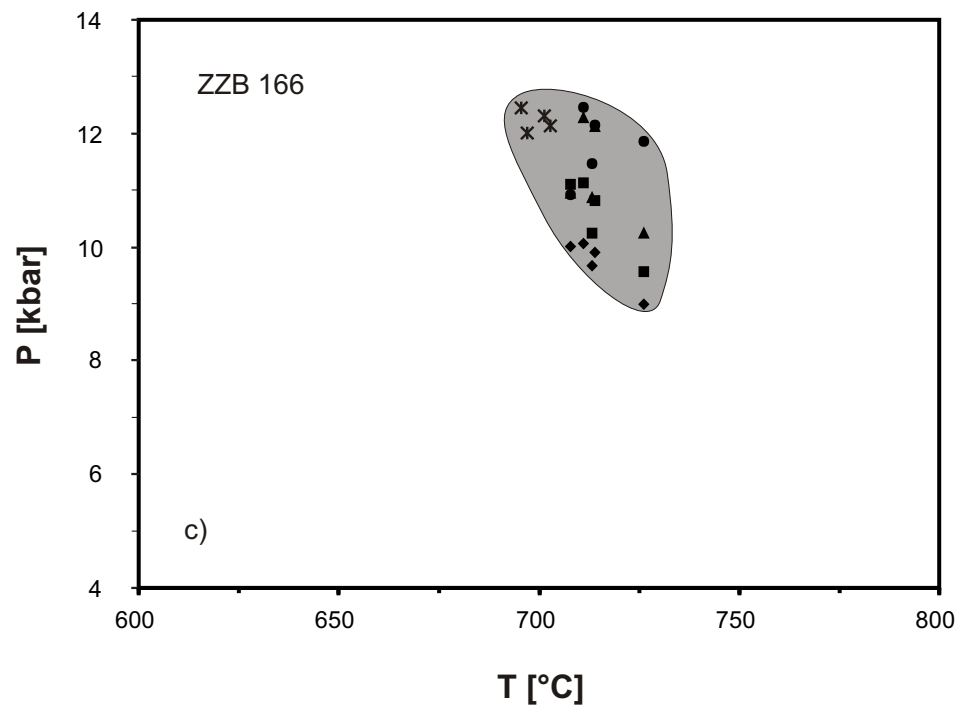


Figure 4-10 a-d. Continued.

4.4 Summary and interpretation of textures and *PT*-evolution of the Mavuradonha Layered Complex

The previously described small amphiboles within the serpentinite are interpreted to have replaced primary pyroxenes. These amphiboles (former pyroxenes) and also the plagioclase that occurs in some of the metapyroxenites are suggested to represent intercumulus phases because they are always present in contact areas of cumulus phases.

Interpretation of the occurrence and chemical variations of garnet and plagioclase are dependent on the *PT* conditions. Textural relations of garnets in metagabbros indicate that they have grown during a high-grade thermal event. In contrast, inclusions in garnets from garnet-amphibolites indicate growth of these garnets during retrogression following peak metamorphism. Garnet coronas around clinopyroxene in metagabbros can also be interpreted to be the result of retrogression. They also seem to be the youngest garnet type, because the inclusions in these garnets are found as a secondary assemblage in metagabbros and as primary minerals in the amphibolites. Observed compositional differences in plagioclase from the Mavuradonha Layered Complex monitors differences in *PT* conditions. Whereas only minor variations in An-content are shown in sample that are not re-equilibrated under retrograde metamorphic conditions, plagioclases from samples that re-equilibrated under amphibolite-facies conditions show a larger compositional range. Indicative for re-equilibration is also the enrichment in albite component in recrystallised plagioclases.

The results of geothermobarometry are summarized in Figure 4-11. Within the complex, a thermal gradient was observed from the amphibolites in the S to the granulites in the N which is mainly the effect of intense shearing and fluid infiltration in combination with retrogression in the S. Granulite-facies *PT* conditions reached pressures of 13 ± 2 kbar and temperatures of 840 ± 30 °C. After high-temperature metamorphism under high-pressure granulite-facies conditions the complex experienced a phase of cooling. During cooling at amphibolite-facies conditions pressures of 11 kbar at ~ 680 °C were attained. This retrograde amphibolite-facies event occurred under static conditions that gave rise to the observed corona textures of hornblende-quartz or pargasite around pyroxene. The formation of green amphibole coronas around clinopyroxenes in metapyroxenites can be interpreted as a fluid-induced cooling texture (Indares, 1993). Despite the described corona textures in coarse- and fine-grained metagabbros, all samples, however, show a polygonal granoblastic texture, which suggests static annealing under high T-conditions for a prolonged period of time. The common textures in amphibolites where green amphiboles often contain numerous quartz inclusions seem to reflect the formation of amphibole and

quartz at the expense of former clinopyroxene. The clinopyroxenes that were found in some garnet-amphibolites (e.g. ZZB 41) are interpreted as remnants of a previous paragenesis.

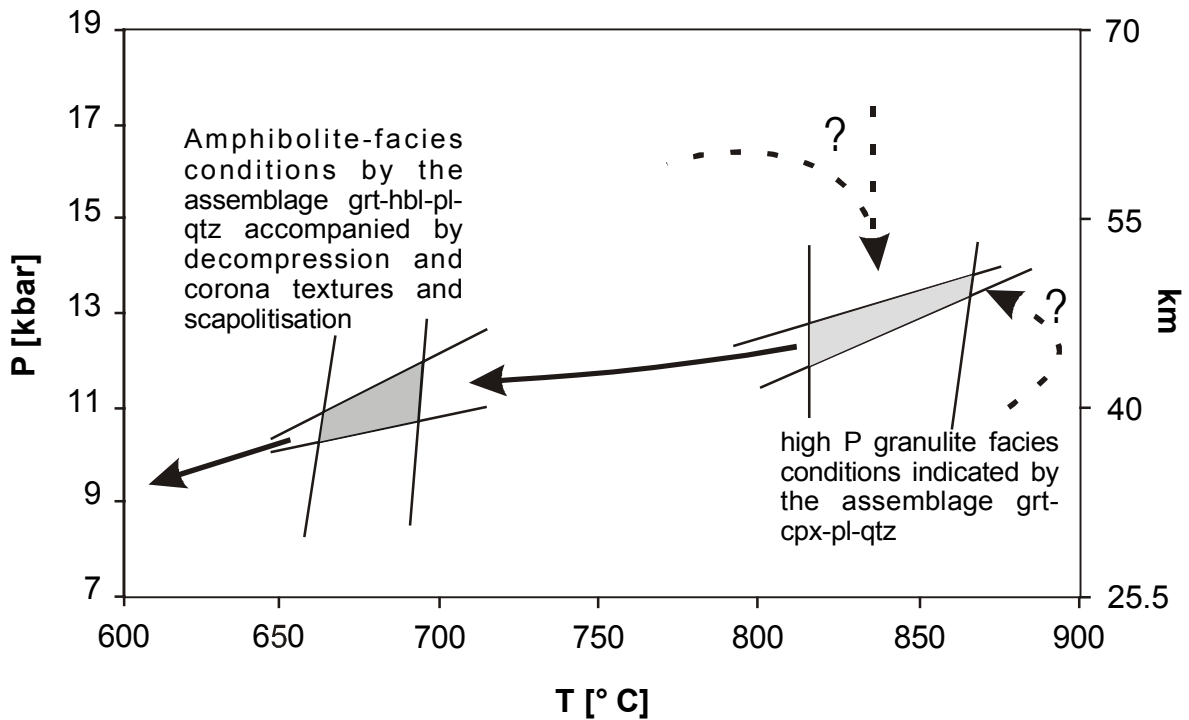


Figure 4-11. Summary of estimated PT conditions for the Mavuradonha Layered Complex through time, indicating a slight decrease in P and T . The decompression and subsequent cooling are consistent with petrographic observations.

Amphibolite-facies retrogression was accompanied by fluid infiltration. By combining textural and chemical evidence this can be divided into two stages of fluid-rock interaction: At first, the formation of chlorine-free amphiboles and quartz is indicative for the initial stage of the fluid-rock interactions with low chlorine concentrations within the fluid. Secondly, during the final stage the growth of chlorine-rich amphiboles and scapolite can be observed. High chlorine concentrations of up to 3 % in scapolites and up to 4.5 % in amphiboles of outer coronas around clinopyroxene and replacing plagioclase are interpreted to reflect ongoing partitioning and enrichment of chlorine in the fluid phase (Kullerud, 1996; Kullerud & Erambert, 1999; Kühn & Austrheim, 2002). The formation of these late stage minerals subsequently occurred in equilibrium with the fluid phase, necessary constituents were mobilized by the fluid-rock interactions and provided by the fluid phase (Kullerud, 1996; Kullerud & Erambert, 1999). However, due to the lack of systematic variations in the scapolites the scapolitisation is interpreted as a single event.

Decompression textures in garnet-amphibolites, e.g. plagioclase coronas around garnet, indicate cooling and exhumation. Changes in PT-conditions are also indicated by the formation of titanite exsolution grids in amphiboles which are interpreted to indicate decompression (Schuhmacher, R., 1991). Yet, another textural feature supporting decompression is the formation of titanite rims around rutile or amphibole in garnet-free amphibolites.

5 Geochemistry

The first two chapters have shown some geological and petrographical similarities for several rock types of the Mavuradonha Layered Complex, the Nyamhanda Inlier and the Chimwaya Hill Inlier, such as gabbroic textures in outcrops and the close relation of garnet-bearing metagabbroic rocks, such as garnet-amphibolites, and corona-textured, garnet-free metagabbros. Based on these similarities in petrography and appearance in the field rock types were grouped for the following descriptions and discussions. These groups are (1) **ferro-metagabbros** comprising garnet-bearing gabbroic rocks, (2) **metagabbros** are represented by garnet-free gabbroic rocks, (3) **metapyroxenites** that comprise the most mafic rocks, whereas the fourth group (**meta-anorthosites**) is composed of the felsic rock types meta-anorthosites and leuco-metagabbros. However, many differences are evident when comparing the Mavuradonha Layered Complex and the rocks of the Nyamhanda and Chimwaya Hill Inlier, such as the lack of common layering in the Inlier metagabbros, different textures in thin section and the occurrence of olivine.

In the following, the geochemistry of the metagabbroic rocks of the three complexes/Inliers is described, and major and trace elements are used in order to substantiate genetic links between the Mavuradonha Layered Complex rock types and also to work out whether genetic correlations between the metagabbroic rocks of the Mavuradonha Layered Complex and the Inlier metagabbros exist. Therefore, a large number of samples of the Mavuradonha Layered Complex and the Inliers were analysed for major and trace element composition in order to cover a large compositional spectrum, and in order to determine geochemical correlations between the Mavuradonha Layered Complex and the inliers. The CIPW normative mineral composition and the whole-rock geochemistry data are given in Appendix IV, Table IV-1. The samples were chosen according to their petrography and their degree of alteration. Samples showing an internal, micro scale layering, samples with extreme alteration and samples with large proportions of vein material were excluded.

The first part of this chapter discusses the geochemical classification of the complete set of analysed metagabbroic rocks, the second part describes the major and trace element geochemistry of the Mavuradonha Layered Complex and the inliers, where also selected variation diagrams are discussed.

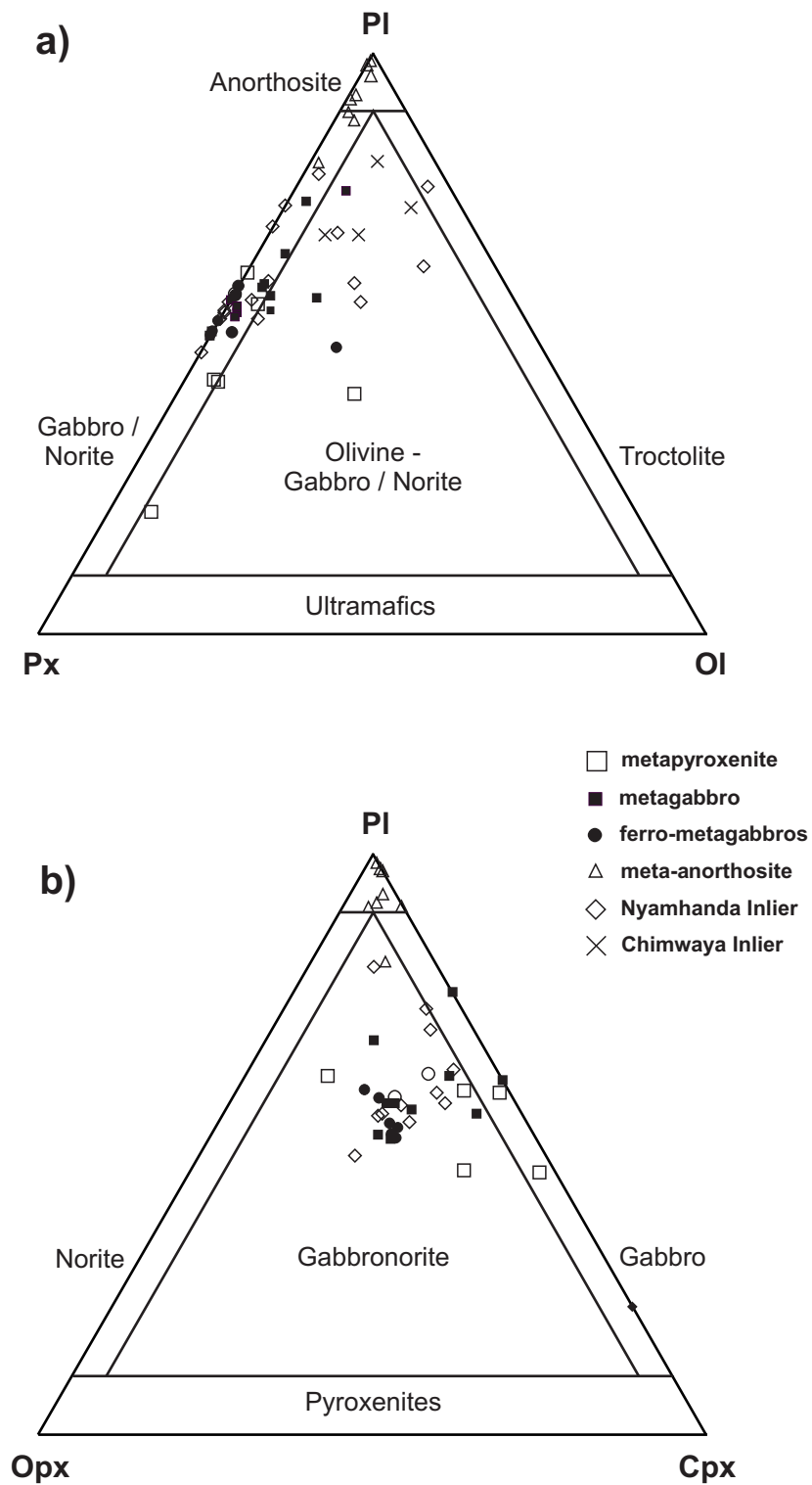


Figure 5-1 a and b. Classification of Mavuradonha Layered Complex, Nyamhanda Inlier and Chimwaya Hill Inlier samples using normative mineral compositions. Classification diagrams according to the IUGS suggestions after Streckeisen (1976; 1980).

5.1 Classification of the metagabbros

In order to characterise the metagabbroic to meta-anorthositic rocks by their geochemistry the CIPW norm is used because the primary modal mineral composition was significantly modified due the high-grade metamorphic overprint.

As shown in Figure 5-1 a) and b) the investigated samples display a gabbroic/noritic to anorthositic character. All samples, except the metapyroxenites, plot in the upper half of both diagrams, indicating that plagioclase was a prominent constituent, whereas pyroxene and/or olivine played an important role in the metapyroxenites.

More than 20 % of the samples can be classified as leuco-gabbros/norites, because in both diagrams the analysed felsic rocks plot in or slightly below the field of anorthosites. The distribution of the data points in both diagrams corroborate that clinopyroxene and plagioclase are the major cumulus phases.

A tholeiitic character of the rocks from the Mavuradonha Layered Complex is displayed in the AFM-diagram (Fig. 5-2). An iron enrichment is also evident in Figure 5-2 which mainly effects a single rock type of the Mavuradonha Layered Complex, namely the ferro-metagabbros. The diagram also displays a trend of iron enrichment which progressively increases from metapyroxenite to ferro-metagabbro. Although the meta-anorthosites plot in the alkaline field, the diagram rather reflects the differentiation of the parental magma and the decreasing content of ferromagnesian phases than a primary, more alkaline composition. A compositionally similar array is shown for the Malagasy anorthosites (Ashwal et al., 1998), demonstrating an enrichment in alkaline component. This enrichment is interpreted by these authors to reflect the variable amount of ferromagnesian phases. Taking both trends into account, iron enrichment and the enrichment in alkaline composition in the anorthosites, a comagmatic evolution for the rocks of the Mavuradonha Layered Complex is suggested. Magnesium-enriched cumulates such as pyroxenites and gabbros/norites originated first and show a wide compositional range. With higher degrees of differentiation, the metagabbros merge into ferro-metagabbros and meta-anorthosites.

The Nyamhanda Inlier metagabbros show a strong iron enrichment which is typical of tholeiites. Most of these gabbroic rocks, except the olivine-bearing rocks, turn out to be much richer in iron than the Chimwaya Hill Inlier metagabbros and the Mavuradonha Layered Complex metagabbroic rocks. In Fig. 5-2 the metagabbros of both inliers show affinities to rocks of the Mavuradonha Layered Complex. Whereas most of the Nyamhanda Inlier metagabbros show strong affinities to the ferro-metagabbroic rocks of the Mavuradonha Layered Complex, the olivine-bearing metagabbros and most of the

Chimwaya Hill Inlier metagabbros may be related to the metagabbros of the Mavuradonha Layered Complex.

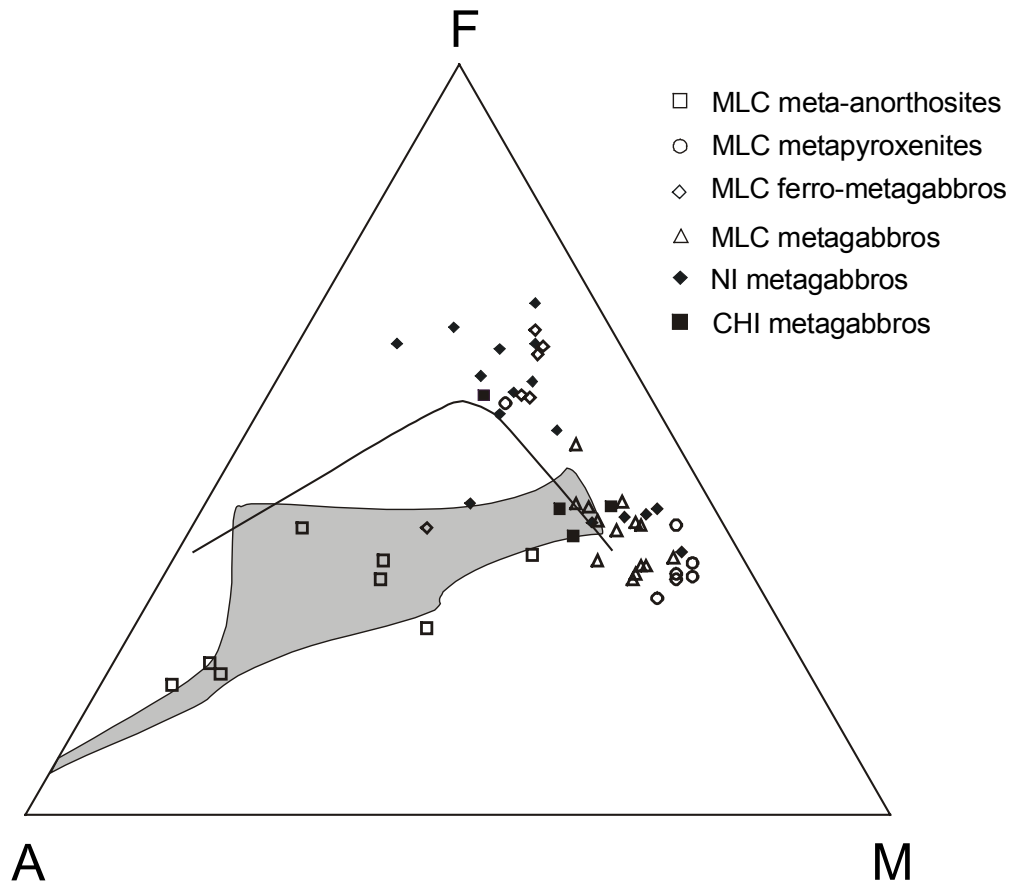


Figure 5-2. Chemical classification of metagabbroic rocks according to the AFM-diagram (Irvine & Baragar, 1971). The line marks the boundary between the calc-alkaline and the tholeiitic field of Irvine & Baragar (1971). Shaded area marks the field of compositions for the Malagasy anorthosites, Madagascar (data from Ashwal et al., 1998).

5.2 Major and trace element variations

5.2.1 Mavuradonha Layered Complex

The geology and petrology of the various rock types from the Mavuradonha Layered Complex show a close relationship between single rock types, although they were collected at several localities which are separated by great distances (up to several km). As is obvious from the previous descriptions the different rock types display similarities in field occurrence such as gabbroic textures in outcrops, similar microscopic textures and petrological/chemical similarities. The analysed rocks have MgO concentrations ranging

from 0.28 to 15.18 wt. % and Mg# between 0.31 and 0.86. All rock types are generally depleted in K_2O , MnO and P_2O_5 relative to other metagabbros in the Zambezi belt and the Lufilian Arc (Barton et al. 1991; John, 2001; Tembo, 1994). The metapyroxenites, metagabbros and meta-anorthosites contain low concentrations of TiO_2 . In contrast, the ferro-metagabbros are enriched in TiO_2 showing concentrations up to 2.1 wt. %. SiO_2 shows a variation ranging from 46.77 wt. % in the metapyroxenites up to 52.21 wt. % in metagabbros, concentrations of Al_2O_3 range from 12.20 to 32.03 wt. %.

A large scatter is shown by the Fe_2O_3 contents, ranging between 0.17 and 12.17 wt. %. Larger variations are also observed for CaO (9.25 to 16.90 wt. %) and Na_2O (1.24 to 5.91 wt. %).

In order to figure out some petrogenetic aspects for the within the complex from whole-rock geochemistry MgO [wt. %] was selected as the index oxide, since it indicates the differences between the various groups most clearly. If individual elements in selected variation diagrams (Fig. 5-3) are plotted versus MgO , several distinct trends are apparent. These trends clearly reflect the petrology of the rock units, particularly the variable amounts of plagioclase, clinopyroxene, and Fe-Ti-Oxides. In the case of Al_2O_3 two different trends are evident, the first indicates the accumulation of plagioclase as an intercumulus phase in some metapyroxenites up to a major cumulus phase in meta-anorthosites. The second trend displays enrichment of Al_2O_3 in the metapyroxenites with a depletion trend towards the ferro-metagabbros. In the diagram Na_2O vs. MgO the trend reflects the enrichment of the residual magma in NaO_2 and the accumulation of sodium-rich plagioclase and clinopyroxene in meta-anorthosites. Fe_2O_3 and TiO_2 correlate negatively with MgO and show a marked increase in the two elements, which is restricted to ferro-metagabbros. Both elements correlate well as can be seen in Figure 5-3. Whereas TiO_2 and FeO contents are generally low in the metagabbros, metapyroxenites and meta-anorthosites, there is a marked increase in both elements if MgO is lower than 8 %. This increase in FeO and TiO_2 concentrations reflects the crystallisation of Fe-Ti-Oxides (titanomagnetite or ilmenite) as well as clinopyroxene enriched in hedenbergite component from the magma. A variation diagram of CaO/Al_2O_3 vs. MgO (Fig. 5-3) shows progressively decreasing CaO/Al_2O_3 ratios with decreasing MgO .

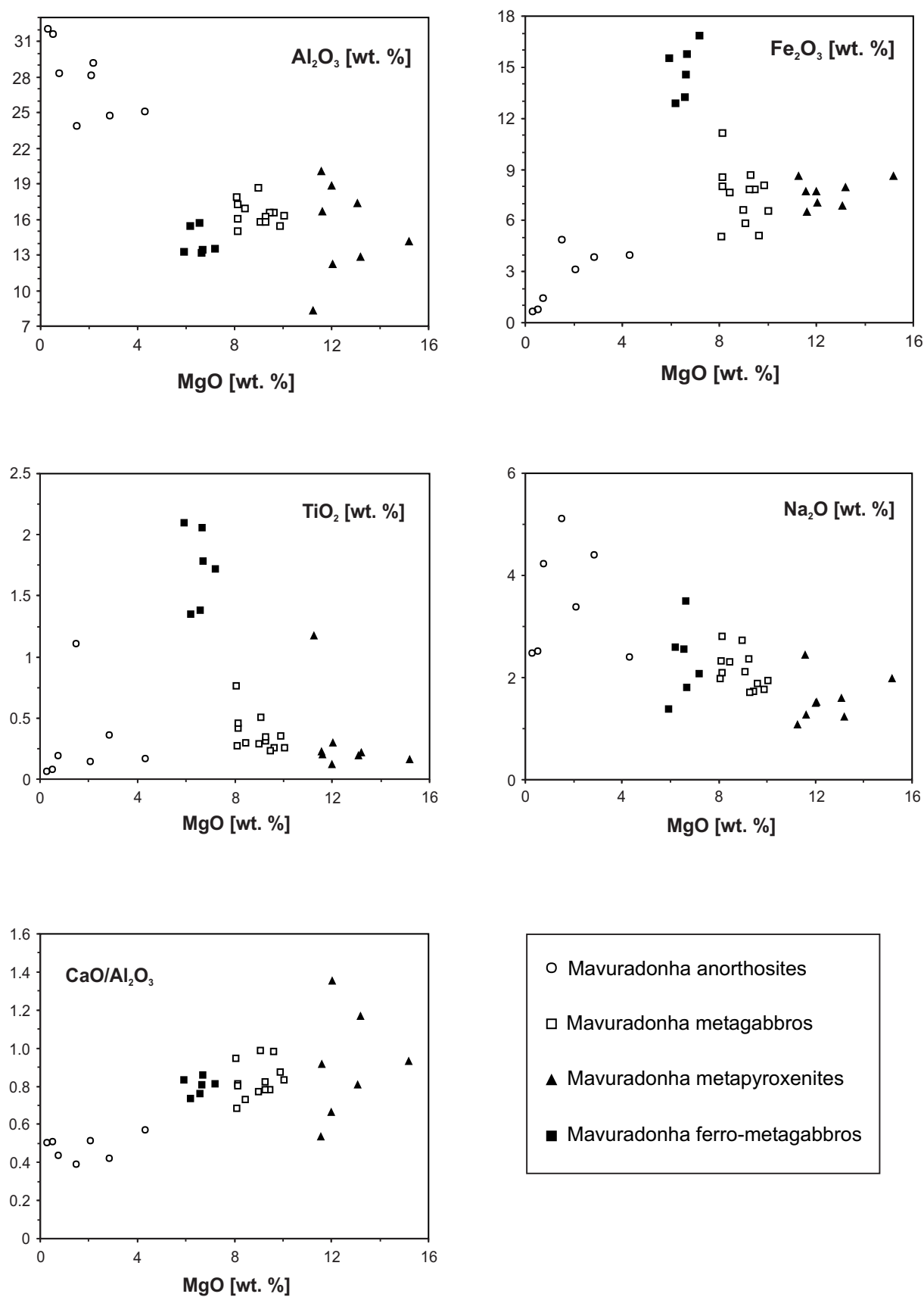


Figure 5-3. Selected major and trace element variations versus MgO for metagabbroic rocks of the Mavuradonha Layered Complex.

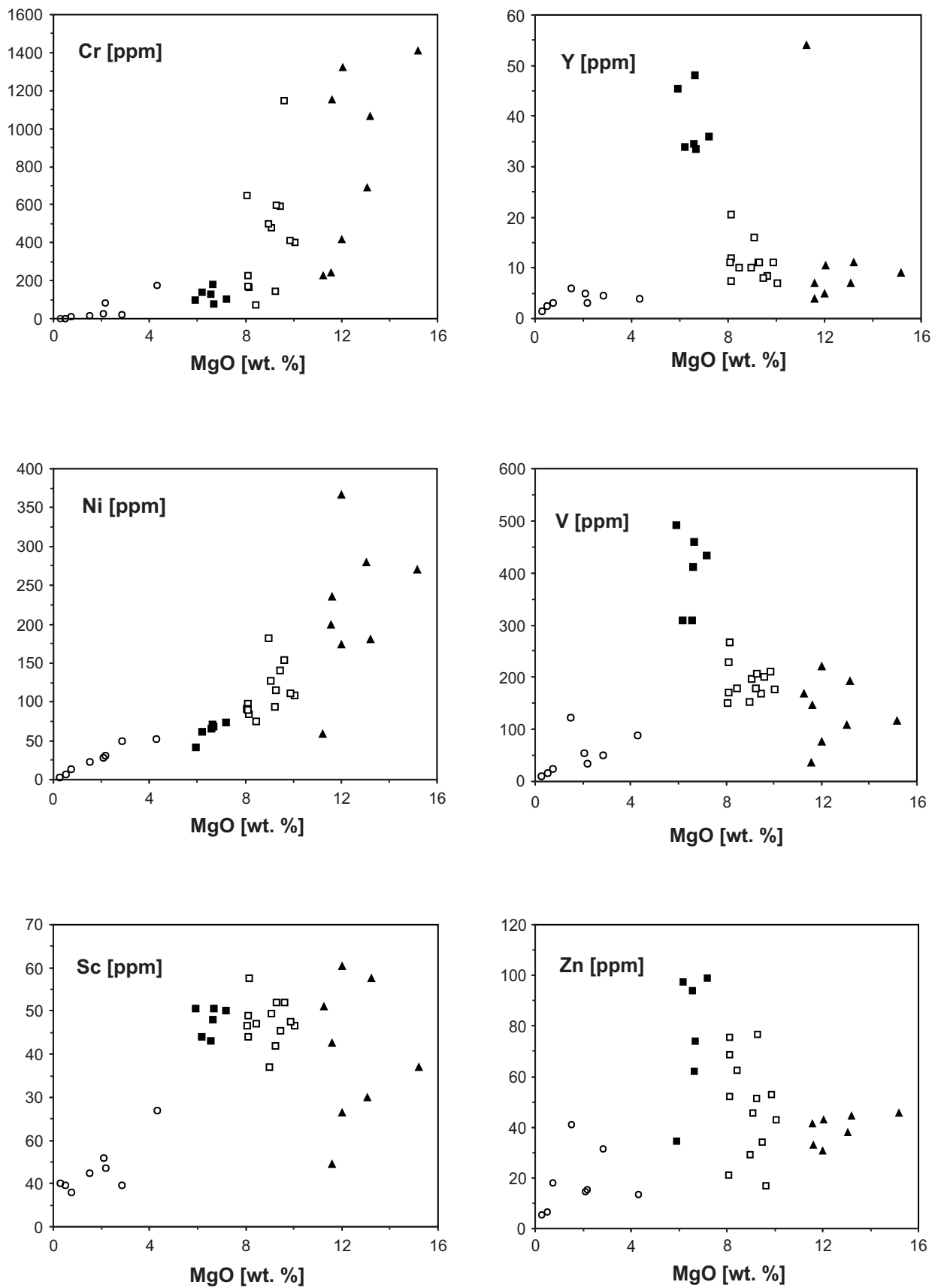


Figure 5-3. Continued.

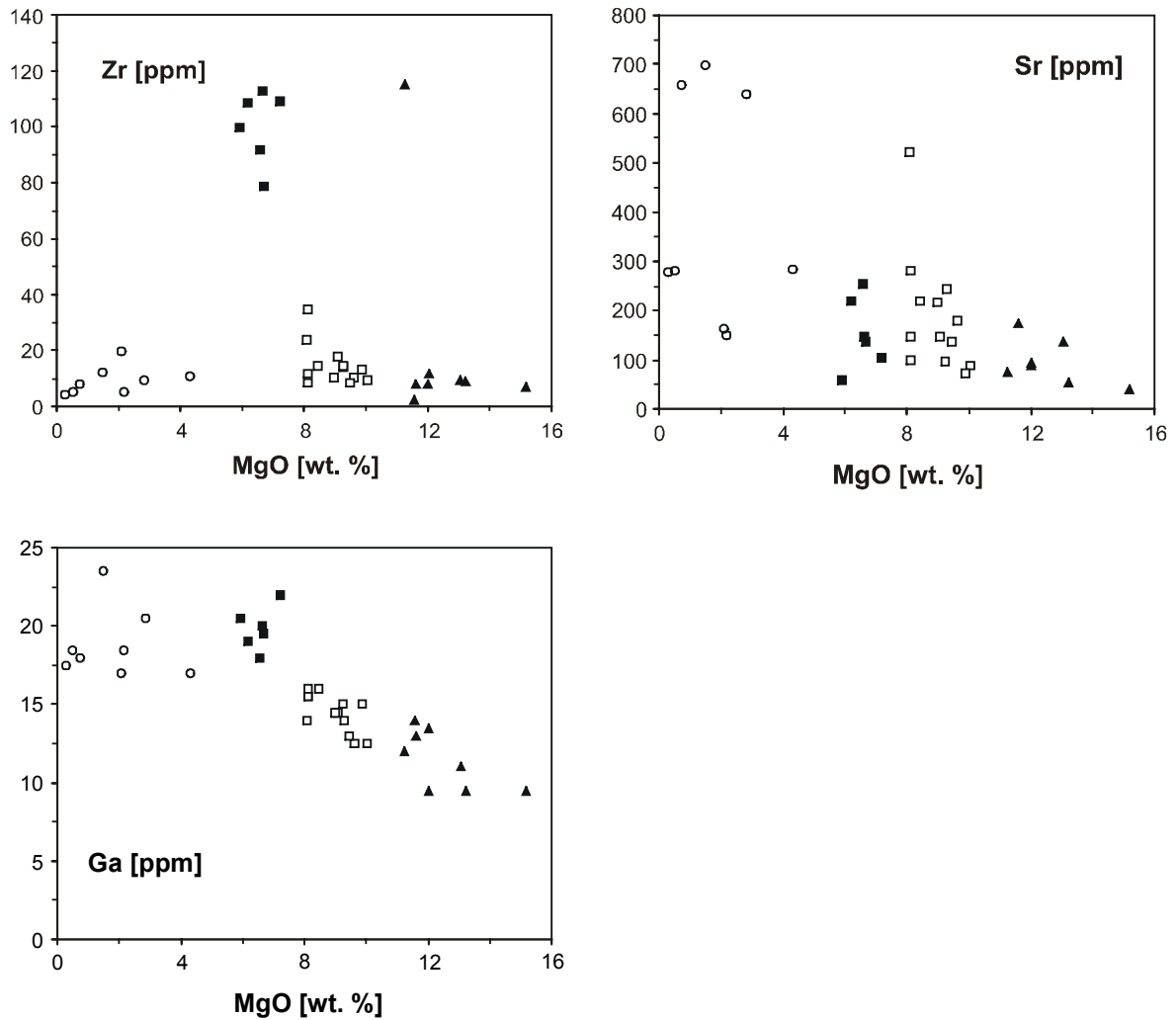


Figure 5-3. Continued.

Samples from the Mavuradonha Layered Complex display two characteristic features in their trace element composition:

- First, the meta-anorthositic rocks show a different behaviour from the other three groups. Generally, the trace elements in the meta-anorthosites are depleted relative to the other groups, except for Sr and Ga.
- Additionally, some elements such as Y, V, Zr, Nb and Zn show a behaviour which can be described as ferro-titanophile, correlating well with FeO and TiO₂.

The trace elements Cr and Ni correlate well with MgO within the different groups of the Mavuradonha Layered Complex. These compatible elements can be used as

indicators for early differentiation processes within the magma chamber because both elements decrease with decreasing MgO concentrations. Scandium displays a general depletion trend with decreasing MgO from relatively high concentrations in metapyroxenites to low concentrations in meta-anorthosites. These trends imply crystallisation of mineral phases such as pyroxene or olivine and opaques from the magma which can incorporate these elements in their crystal lattice.

Yttrium, V, Nb, Zn, Sr and Ga show negative correlations with MgO. These negative correlations are restricted to the metagabbros and metapyroxenites. These elements, except Sr and Ga, are generally depleted in the meta-anorthosites. The negative correlation of these elements indicates enrichment in the residual magma. This enrichment of Y, V, Nb, Zn implies the occurrence of Fe-Ti oxides and apatite as additional cumulate phases. The negative Sr and Ga correlation reflects accumulation of plagioclase as the major Sr and Ga-bearing phase (McBirney, 1998).

5.2.2 Nyamhanda Inlier and Chimwaya Hill Inlier

Field observations and investigations in thin sections lead to the supposition for a genetic relation between the Chimwaya Hill Inlier and the Mavuradonha Layered Complex rocks. Contrary to this, the Nyamhanda Inlier rocks show a different occurrence in the field, and petrographic investigations demonstrated major differences between the Nyamhanda Inlier and the Mavuradonha Layered Complex metagabbroic rocks, such as the appearance of different corona textures around the mafic phases and the clearly subophitic texture in most of the samples. The results of major and trace element investigations for each rock type which was distinguished in the metagabbros of both inliers during field work are shown in Figure 5-4 and are listed in Appendix IV, Table IV-2.

MgO in the metagabbroic samples of the inliers ranges from 2.10 to 13.90 wt. % and SiO₂ ranges from 43.76 to 54.26 wt. %. Al₂O₃ ranges between 13.15 and 24.81 wt. % and Fe₂O₃ is ranging from 4.39 to 15.61 wt. %. K₂O, MnO and P₂O₅ concentrations are generally higher than those reported from the Mavuradonha Layered Complex. These major elements are enriched in samples with low MgO but high FeO concentrations. Major element trends (Fig. 5-4) are restricted to CaO, TiO₂ and Na₂O. CaO (8.08 – 14.27 wt. %) correlates well with MgO and increases with increasing MgO. TiO₂ (2.27 – 3.97 wt. %) and Na₂O (0.75 - 4.56 wt. %) are negatively correlated with MgO, whereas CaO/Al₂O₃ ratios are fairly constant at decreasing MgO.

The inlier samples display common trace element characteristics, in which compatible elements are enriched in samples with high MgO concentrations. Negative correlations are visible in HFS and incompatible elements.

The trace elements Cr, Ni and Co correlate well with MgO. These three elements decrease with decreasing MgO content, which implies crystallisation of early phases from the mafic magma, e.g. pyroxene and/or olivine as well as opaques. These elements are sensitive indicators of fractionation of these early crystallising minerals.

Strontium, Rb, Ba and Ga yield negative correlations with MgO, and all four elements are incompatible elements. They increase with decreasing MgO. This behaviour reflects crystallisation and removal of feldspars from the magma. Y, V, Zr, Nb and Zn show negative correlations with MgO. These elements are enriched mainly in Fe-rich metagabbros and metagabbros containing high Fe_2O_3 concentrations indicating the presence of Fe-Ti oxides and apatite.

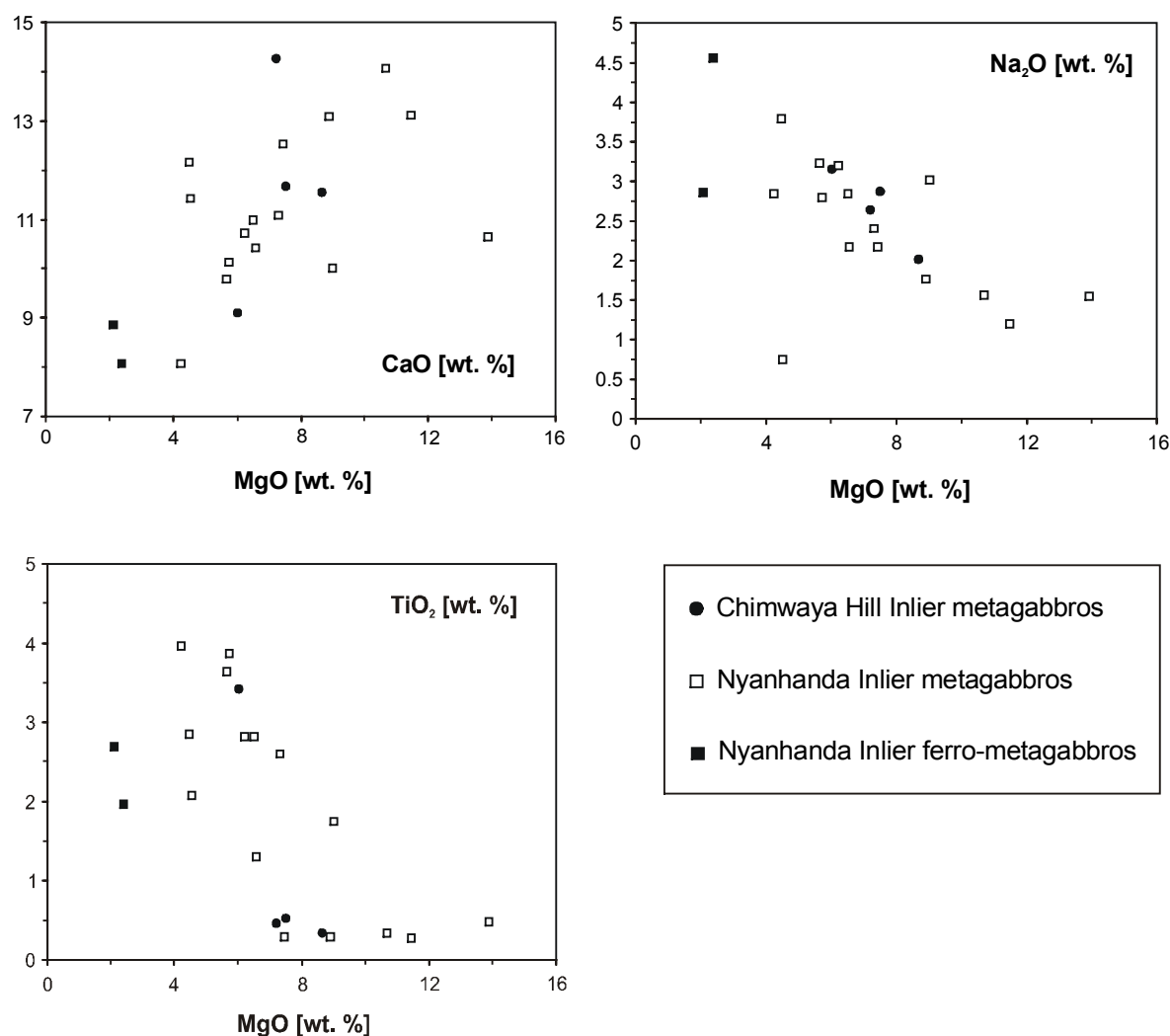


Figure 5-4. Selected major and trace element variations versus MgO for metagabbroic rocks of the Nyanhanda Inlier and Chimwaya Hill Inlier.

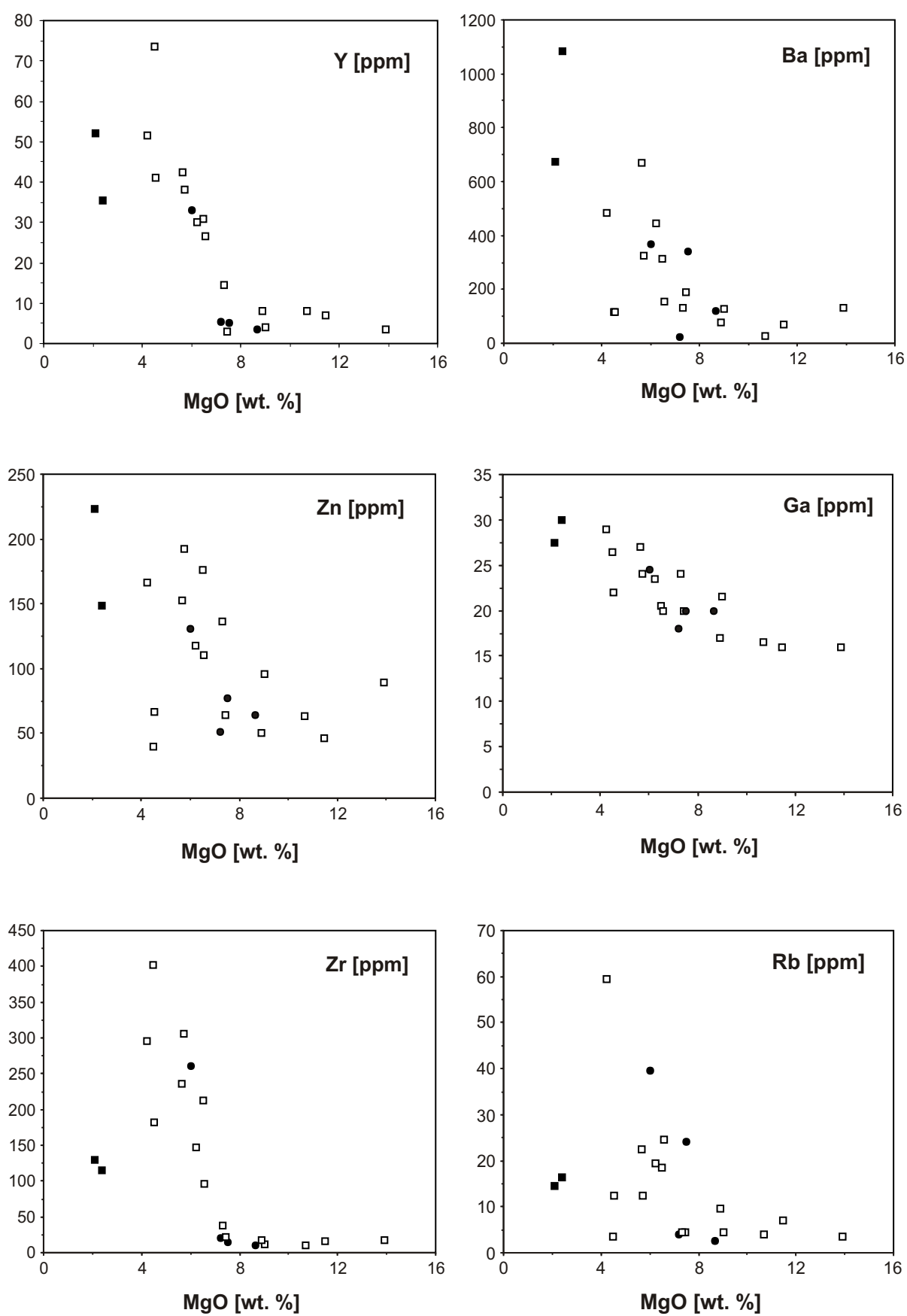


Figure 5-4. Continued.

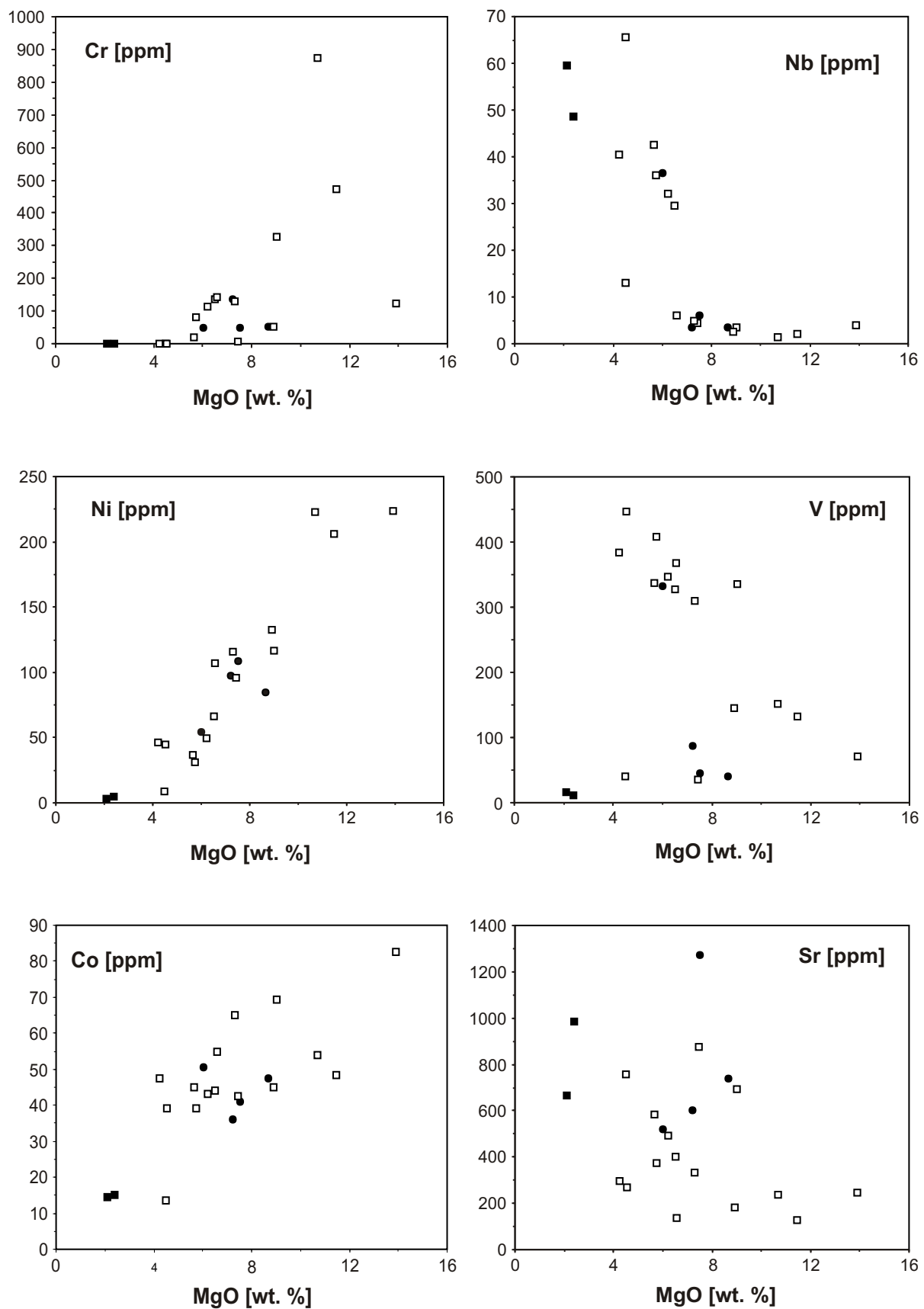


Figure 5-4. Continued.

5.3 TiO₂, LIL and HFS element correlations

Additional aspects concerning the composition of the metagabbroic rocks can be obtained from the trace element distribution because some elements such as Y, V, Zr, Nb and Zn, correlate well with Fe₂O₃ and TiO₂. Variation diagrams of LIL and HFS elements versus TiO₂ (Fig. 5-5) provide some genetic relations for the evolution of the parental magma to the metagabbros of the Mavuradonha Layered Complex, Nyamhanda Inlier and Chimwaya Hill Inlier.

First, considering the affinity of LIL and HFS elements as highly incompatible, the diagrams indicate that these are enriched in the residual magma with higher degrees of differentiation. The concentrations of these elements are fairly constant or show a progressive increase at low stages of differentiation. Second, the evolution of the magma resulted in accumulation of mineral phases enriched in FeO, TiO₂, LIL and HFS elements, mainly in the garnet-bearing ferro-metagabbros. Once a TiO₂ concentration of 1.2% wt. % in the differentiated residual magma was reached, a marked increase in the concentrations of LIL and HFS elements above this TiO₂-content is visible. In contrast to the metagabbros and metapyroxenites, an increase in these elements indicates that the ferro-metagabbros reflect a higher degree of differentiation and, therefore, have accumulated mineral phases carrying those elements (Zr, Nb and V).

In addition to these aspects it is obvious from the diagrams that some of the metagabbros and ferro-metagabbros of the Nyamhanda Inlier are much richer in TiO₂, Zr, Zn and Nb than the Mavuradonha Layered Complex and Chimwaya Hill Inlier metagabbroic rocks, whereas other elements have similar concentrations.

5.4 Summary and interpretation

5.4.1 Mavuradonha Layered Complex

The chapters geology and petrology have summarised some typical features of a layered intrusion, although these rocks were overprinted by high-grade metamorphic conditions. These features are:

- a general differentiation sequence from mafic/ultramafic to anorthositic rocks
- the gradual layering at variable scales
- typical textures such as mortar and spotted texture of anorthosites
- mineral chemical fractionation trends as observed for clinopyroxene and plagioclase.

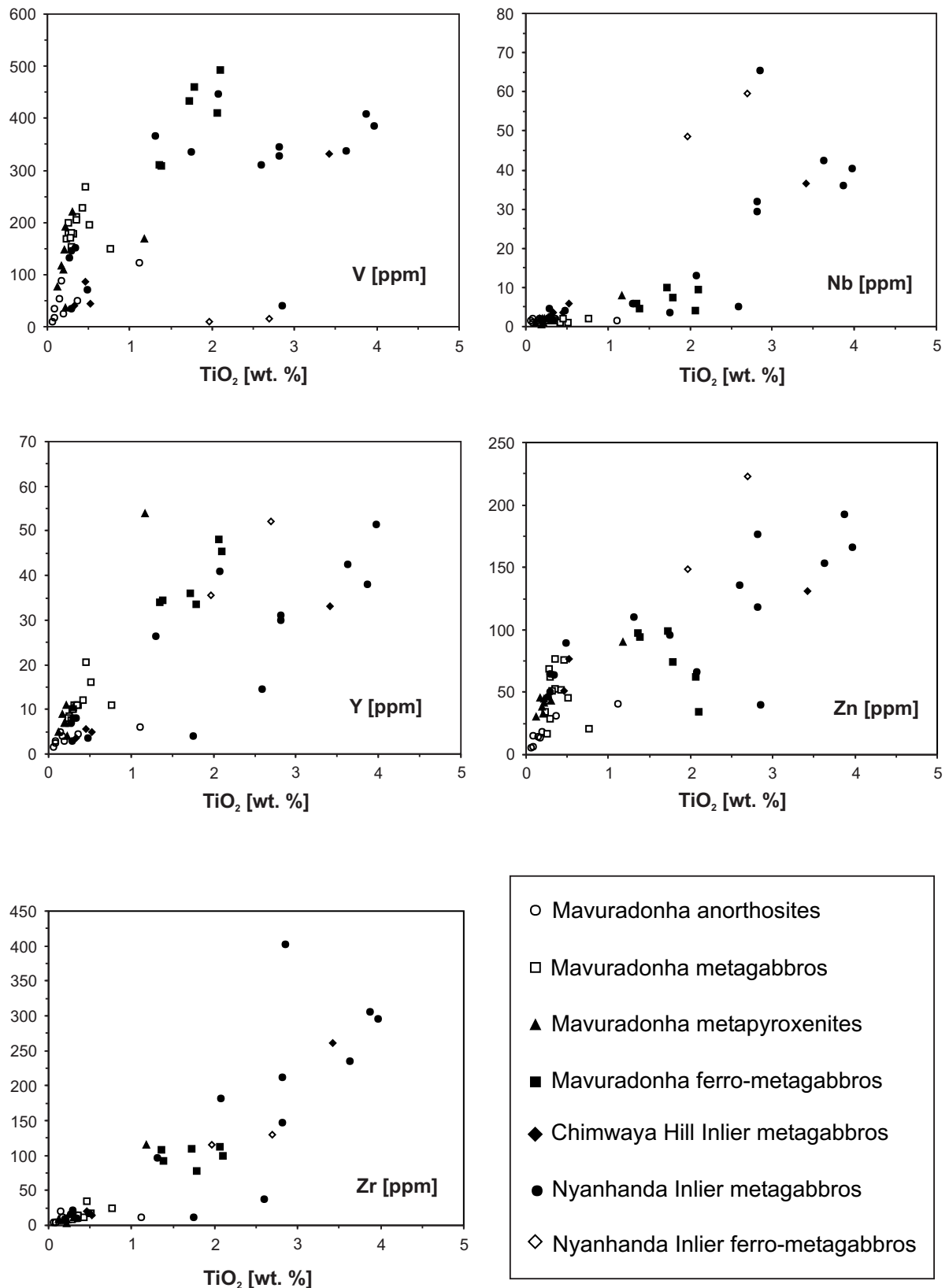


Figure 5-5. Variation diagrams of HFS and LIL elements versus TiO_2 . Note the marked increase of the HFS and LIL elements above a TiO_2 concentration of 1.2 wt. %.

The results of geochemical investigations corroborate the interpretation of the Mavuradonha Layered Complex as a layered intrusion.

Concerning the results of this chapter the investigated rocks of the Mavuradonha Layered Complex which are now exposed in individual sequences reflect accumulation of the major mineral phases clinopyroxene and plagioclase. With higher degrees of differentiation of the residual magma additional phases such as Fe-Ti oxides and apatite were accumulated. In turn, these major cumulate phases control the whole-rock geochemistry which also reflects the variations of elements as favoured by these major mineral phases, as has been recognised for many other layered intrusions and anorthosites (e.g., Ashwal, 1993; McBirney, 1995; Haskin & Salpas, 1992; Markl & Frost, 1999).

So far the following geochemical aspects have been shown: The rocks of the Mavuradonha Layered Complex have a tholeiitic character (Fig. 5-2). The trends shown in the variation diagrams (Fig. 5-3) can be described as strong iron enrichment, decreasing CaO/Al₂O₃ ratios, and an increase of TiO, Na₂O, LIL and HFS elements with increasing differentiation, whereas other trace elements are generally compatible with the major element variations. Although major and trace element variations exist within each group, a general differentiation trend is evident in the Mavuradonha Layered Complex:

First, Mg-rich cumulates crystallised, whereas the residual magma was enriched in iron and incompatible elements. The metapyroxenites, which were accumulated first, reflect the most primitive geochemical composition in the Mavuradonha Layered Complex, because these samples are enriched in MgO, CaO, Cr and Ni and show depletion in TiO₂, Na₂O, Ga, LIL and HFS elements. This suggests a relatively early differentiation stage, which is also supported by petrological observations (Chapter 4).

During the second and third evolutionary stages metagabbros and ferro-metagabbros were accumulated. A slight depletion in CaO, Al₂O₃, Cr, Ni and Sc, an enrichment in TiO₂, Na₂O, Ga, LIL and HFS elements are features of the metagabbros. With ongoing differentiation the samples are dominantly garnet-bearing and a marked iron enrichment is present in garnet-bearing ferro-metagabbros. They also contain less aluminium and calcium but more titanium and sodium as compared to the metagabbros. Ferro-metagabbros are also depleted in Ni and Cr and other compatible elements but contain higher concentrations of incompatible elements such as Sr, Ga, LIL and HFS elements. Considering this, the formation of garnet seems to be controlled by the composition of the rocks rather than by an increase in high-grade metamorphic conditions. Ferro-metagabbros have compositional affinities to ferro-gabbros and ferro-diorites as described from massive-type anorthosites (e.g. Kolker et al., 1990; Olson & Morse, 1990; Ashwal, 1993 and references therein; Mitchell et al., 1995; 1996; Bhattacharya et al., 1998; Markl & Frost, 1999). The ferro-metagabbros would therefore be in line with the model

proposed for gabbros to be the residual magma of the anorthosites (McLelland et al., 1994; Mitchell et al., 1995; 1996). These authors proposed aluminium and iron enrichment of a basaltic magma due to the crystallisation and accumulation of ferromagnesian phases in cumulate sequences. This enriched magma later behaved as the parental magma from which the high-Al gabbros and the associated anorthosites were extracted.

During a final stage, and through the decrease of mafic components, anorthosites crystallised. Generally, meta-anorthosites display a different geochemical characteristic than the metapyroxenites and metagabbroic rocks. Additionally, the meta-anorthosites occupy a special position within most of the differentiation trends that are displayed in the variation diagrams. The meta-anorthosites are enriched in Al_2O_3 , Na_2O , Ga and Sr, but depleted in other major elements, compatible trace, LIL and HFS elements.

Anorthosites are common in layered intrusions (e.g. anorthosites of the Critical Zone and the Main Zone, Bushveld Complex, Eales et al. 1986; AN-I and AN-II of the Stillwater Complex; Haskin & Salpas, 1992) and display their own petrological and geochemical characteristic. Anorthosites from these two complexes and other layered intrusions display fractionation trends in which Fe and Na were enriched with the evolution of the residual magma. The highest concentrations were found at the highest stratigraphic levels. Generally, an enrichment in major and trace elements which have affinities for plagioclase is proposed, whereas other elements are generally depleted. Judging from geochemical results of this study in combination with petrological investigations, the meta-anorthosites and their related mafic rocks of the Mavuradonha Layered Complex reflect common iron and sodium enrichment trends caused by fractionation.

Considering the three different units of the Mavuradonha Layered Complex as a whole chemical and mineralogical data indicate the following main fractionation scheme: First, clinopyroxene was fractionated, thereby reducing the CaO and also the $\text{CaO}/\text{Al}_2\text{O}_3$ ratio. During the second stage clinopyroxene fractionation was accompanied by plagioclase fractionation which led to a further reduction of MgO and the $\text{CaO}/\text{Al}_2\text{O}_3$ ratio. The fractionation of these two phases reduced the Al_2O_3 concentration ($\text{Al}_2\text{O}_3/\text{MgO}$ variation diagram) towards lower MgO prior to its marked increase in the meta-anorthosites. The increase of Na_2O whereas Al_2O_3 decreases indicates that fractionating plagioclase and clinopyroxene are progressively more sodic.

5.4.2 Nyamhanda Inlier and Chimwaya Hill Inlier

Major and trace element analyses of metagabbros and ferro-metagabbros of the Nyamhanda and Chimwaya Hill Inliers display systematic variations which are different to those observed for the Mavuradonha Layered Complex. Overall, the inlier samples are

enriched in incompatible elements such as TiO_2 , K_2O , MnO and P_2O_5 , LIL and HFS elements with respect to the Mavuradonha Layered Complex samples. These elements correlate well with Fe_2O_3 concentrations which can be taken as an indicator for an evolved magma enriched in these oxides (TiO_2 , K_2O , MnO and P_2O_5). Although a kind of layering was observed during fieldwork, major and trace element geochemistry of the inlier samples provided no evidence for a layered occurrence of these rocks. Even though different mineral assemblages occur in the inlier samples, the rock types cannot be distinguished by their chemistry as it is possible for the Mavuradonha Layered Complex samples.

High MgO , Cr and Ni concentrations characterise olivine- and clinopyroxene-bearing samples in the inlier metagabbros, whereas other metagabbroic rocks of both inliers display a good correlation between Fe_2O_3 and TiO_2 . Strong iron enrichment with decreasing MgO , which is also evident in the AFM diagram (Fig. 5-2), and the correlation between Fe_2O_3 and TiO_2 indicate crystallisation of Fe-Ti oxides. The significantly decreasing CaO with decreasing MgO is indicative of fractionation of clinopyroxene. Fairly constant Al_2O_3 (not displayed in diagram) in combination with increasing Na_2O suggests that crystallising plagioclase became richer in sodium. The major and trace element variation of the inlier metagabbros and ferro-metagabbros is typical for continental tholeiites (Le Maitre, 1976; Tarney, 1992) showing a marked increase of incompatible LIL and HFS elements with decreasing MgO . A proportional variation of Ni and Cr with MgO suggests their control by fractionation of olivine and pyroxene.

Considering the results of the geochemistry, the inlier rocks show similar major and trace element characteristics (variation diagrams, Fig. 5-6) as the Mashonaland Sills (Stubbs et al., 1999) and the metagabbros of the Lufilian Arc (Tembo, 1994). Both are interpreted as continental tholeiites, the latter are interpreted to have evolved in an extensional setting. The investigated inlier metagabbroic rocks are also similar in their major and trace element characteristics to other Pan-African metagabbros from the Masoso, the Rushinga and Chimanda Metamorphic Suites (Barton et al. 1993). Additionally, trace element ratios for the inlier metagabbroic rocks are also similar to ratios in the metagabbros of the Lufilian Arc, the Masoso, the Rushinga and Chimanda Metamorphic Suites: Ti/Zr ratios range between 43 and 409 in the inliers which are similar to the ratios of the metagabbros of the Masoso, the Rushinga and Chimanda Metamorphic Suites (68 – 480) and the Lufilian Arc (35 – 198). Zr/Y ratios (1.2 – 8) are more homogeneous and similar to the above rocks ranging from 0.6 to 4.3 for the Masoso, the Rushinga and Chimanda Metamorphic Suite, and 5 to 11.8 for the metagabbros of the Lufilian Arc.

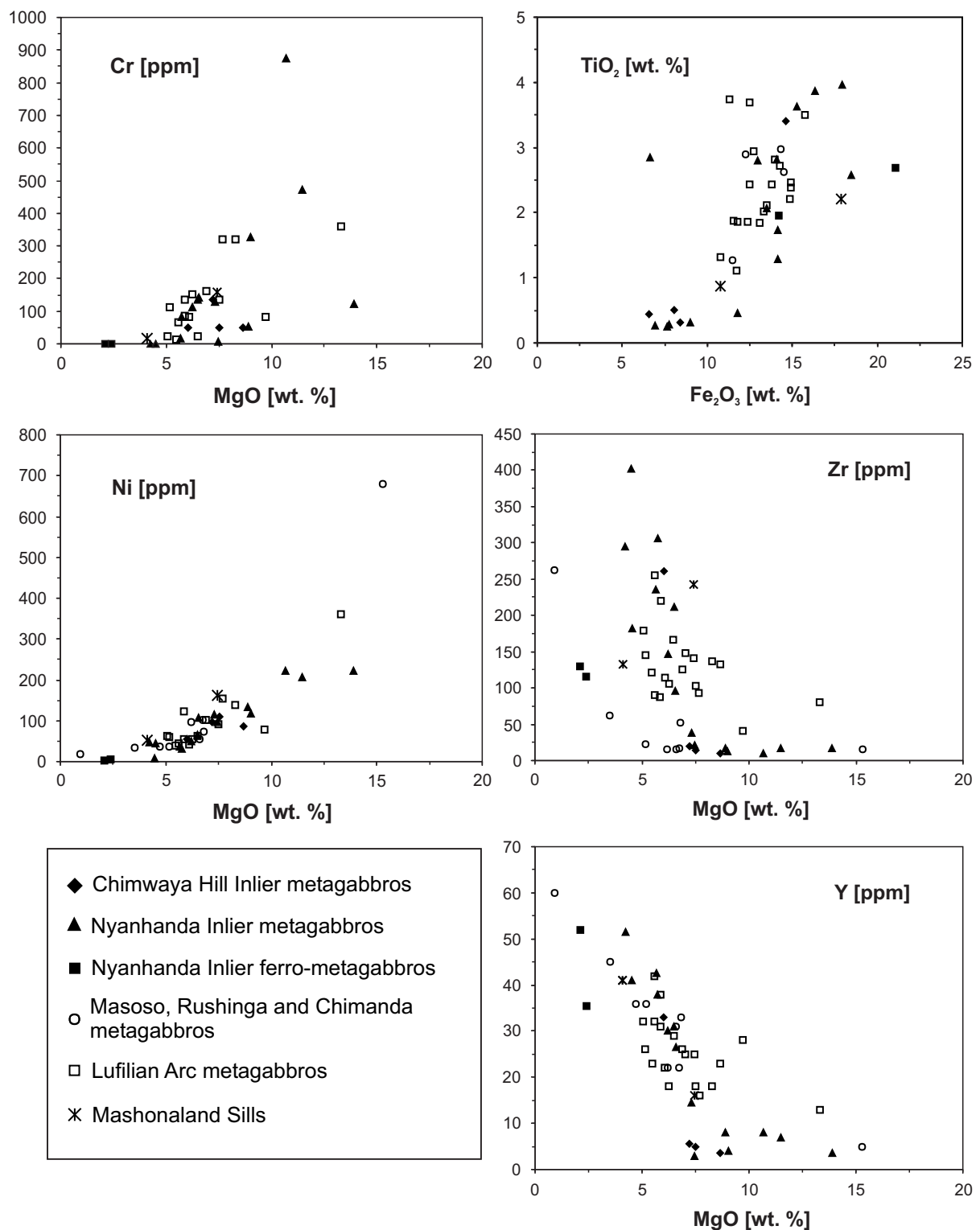


Figure 5-6. Selected major and trace element variations versus MgO and Fe₂O₃ for metagabbroic rocks of the Nyanhanda Inlier and the Chimwaya Hill Inlier in comparison with metagabbros of the Copperfeld and the Domes region, Lufilian Arc (Tembo, 1994), metagabbros of the Masoso Suite, Rushinga and Chimanda Metamorphic Suites, Zimbabwe (Barton et al., 1991) and the Mashonaland Sills, Zimbabwe (Stubbs et al., 1999).

Taking into account the geochemical data presented in this chapter, the major and trace element variations of metagabbroic rocks of the Nyamhanda and Chimwaya Hill Inliers are different to those of the metagabbroic rocks of the Mavuradonha Layered Complex. In contrast to the large variety of rock types of the Mavuradonha Layered Complex, the inliers are composed only of metagabbros and ferro-metagabbros and, therefore, it is impossible to distinguish different rock types as in the Mavuradonha Layered Complex. According to their geochemistry, no genetic relation between the inlier metagabbros and the Mavuradonha Layered Complex metagabbroic rocks is evident. In the case of the Nyamhanda Inlier metagabbros, this supports the hypothesis of an independent evolutionary history for these rocks which were probably extracted from a different source region in the mantle. For the Chimwaya Hill Inlier the geochemistry provides no evidence for a genetic relation with the Mavuradonha Layered Complex metagabbroic rocks. Based on the geochemical data, a similar evolution as for the Nyamhanda Inlier rocks can be assumed for the metagabbros of the Chimwaya Hill Inlier.

6 Geochronology and zircon-cathodoluminescence

U-Pb and Pb-Pb zircon geochronology was carried out in order to obtain precise age information on the Mavuradonha Layered Complex and the Ocellar Gneiss. Zircon analyses were carried out in combination with cathodoluminescence (CL) studies. Because of the high closure temperature (T_c) for the U-Pb isotopic system of $> 900\text{ }^{\circ}\text{C}$ (Mezger & Krogstad, 1997), the U-Pb system in zircons is not generally affected by high-grade metamorphism (Kröner et al., 1987). Therefore, zircons preserve their primary age information under most crustal conditions. Discordant data points may be related to a mixture of core and rim components with distinct isotopic composition of overgrown zircons. Discordant data points also reflect Pb loss during metamorphic events or may be the result of recent lead loss. Metamictisation caused by U decay and the resulting lattice damage and diffusion is the most common mechanism leading to Pb loss (Mezger & Krogstad, 1997).

The thermal history of the Mavuradonha Layered Complex was reconstructed by U-Pb isotope studies on rutile, a mineral with a lower T_c for the U-Pb isotopic system than zircon (Mezger, 1990). The T_c for the U-Pb system in rutile is considered to range from 380 to 420 $^{\circ}\text{C}$ (Mezger et al., 1989). Rutile is the only accessory mineral which occurs in sufficient amounts in rocks of the Mavuradonha Layered Complex. Therefore, rutile may be the only mineral, using the U-Pb isotopic system, which can be used to date cooling following high-grade metamorphism in the northeastern part of the Zambezi belt. The combination of mineral ages and PT estimates are useful indicators for the geological evolution of the Mavuradonha Layered Complex.

Additionally, because the Sm-Nd whole-rock system is believed to be unaffected by high-grade metamorphism (see below), this system is used for age determination. Two garnet-bearing granulite-facies metagabbros and two meta-anorthosites of the upper meta-anorthosite suite of the Mavuradonha Layered Complex were selected for age determinations in order to characterise either the age of high-grade metamorphism or cooling immediately after the metamorphic peak using garnet – whole-rock pairs.

6.1 Cathodoluminescence imaging

Dating of high-grade metamorphic rocks using zircons requires additional methods to identify different types of zircon domains, such as metamorphic overgrowth around a zircon of magmatic origin. If zircons of high-grade metamorphic rocks are used for geochronology, characterisation by typology (Pupin, 1980; Vavra, 1990) alone is

unsatisfactory. Therefore, cathodoluminescence imaging is a powerful tool to obtain information on internal structures and on the origin of zircons (Vavra, 1994)

In CL images, alternating zones of bright and dark luminescence generally characterise zircons. CL emission within these zones depends on trace element as well as REE, Th and U concentrations. Dark zones or bands reflect high trace element contents. Areas depleted in these elements show bright luminescence (Hanchar & Rudnick, 1995; Rubatto & Gebauer, 1999). Using CL imaging, different types of zircon growth and internal structures can be distinguished. Sector zoning and/or oscillatory zoning are considered to be the result of magmatic crystallisation, characterised by relatively high Th/U ratios. Contrary to these two zonation types metamorphic zircons commonly show only sector zoning in CL images (Vavra, 1990).

Metamorphic overgrowth areas are difficult to recognise under an optical microscope, especially where overgrowth follows the morphology of the zircon, or the grain is more or less rounded. Therefore, CL imaging is a useful tool to examine such overgrowth structures. These overgrowth areas are often irregularly shaped and show only weak zoning, or no zoning at all. Metamorphic overgrowth domains have low, homogenous U-contents and are highly luminescent in CL images. Metamorphic domains are frequently, but not always, characterised by low Th/U ratios (Rubatto & Gebauer, 1999). CL imaging also allows identification of additional internal structures such as recrystallisation, resorption of older grains, inherited cores within oscillatory zoned grains and mineral inclusions.

Discordant data points are often the result of metamorphic overgrowth and characterise zircons which experienced lead loss. Therefore, CL imaging prior to age investigation is a convenient way to avoid such discordant data points. The most suitable methods for U-Pb dating of high-grade metamorphic rocks are SHRIMP analysis and CLC-TIMS zircon geochronology (Poller et al. 1997). Both methods provide the ability to select specific zircons with 'normal' internal structures. For further information and details of CL as a tool in mineral analysis the reader is referred to Hanchar and Miller (1993), Vavra et al. (1996), Pagel et al. (1999 and reference therein); Kempe et al. (1999) and Poller (1999).

Zircons of every sample analysed for U-Pb and Pb-Pb isotopic composition in this study were investigated by CL. All isotopic data obtained are discussed in combination with CL observations in the following part.

6.2 Zircon U-Pb and Pb-Pb geochronology

Two metagabbro samples of the Mavuradonha Layered Complex which contained sufficient zircons, were chosen for zircon geochronology. Zircons from three different samples of the Ocellar Granitoid Gneiss, exposed south of the Mavuradonha Layered Complex, were also dated. In addition, two pegmatites which are discordant to the magmatic layering within the Mavuradonha Layered Complex were collected for zircon dating. Pegmatites are exposed in the Mavuradonha Layered Complex as well as in the Ocellar Gneiss and reflect one of the last major tectonothermal events in the Zambezi Allochthonous Terrain as discussed in the first two chapters.

Zircons from the metagabbroic, granitoid and pegmatitic samples were dated by both vapour transfer and single grain evaporation techniques. Zircons from metagabbro sample ZZB 123 were additionally analysed by A. Kröner using a SHRIMP ion microprobe in Perth, Australia. Sample locations are illustrated in Figure 2-2 and in Appendix I. The U-Pb data are reported in Appendix V in Tables V-1, V-2 and V-3, SHRIMP data are presented in Table V-4 and Pb-Pb data are listed in Table V-5.

6.2.1 Mavuradonha Layered Complex

6.2.1.1 Metagabbro sample ZZB 123

Zircons from this metagabbroic sample were dated by three methods: SHRIMP ion-microprobe, vapour transfer (VT), and single grain evaporation. The sample ZZB 123 contains four morphologically distinct populations of zircons.

The first population is pale brown, clear and long prismatic. These zircons are nearly euhedral, some, however, have rounded terminations. In CL images the zircons show an oscillatory zonation with a highly luminescent core. After a period of corrosion the zircons were overgrown by a small younger metamorphic rim (Fig. 6-1 a, b). Both zircons seem to have recrystallised interiors, whereas grain (b) also shows recrystallisation in its outer parts. Some zircons contain inclusions of apatite and quartz.

Zircon of this population yielded two $^{207}\text{Pb}/^{206}\text{Pb}$ evaporation ages within 100 Ma: A mean age of 894 ± 2 Ma (Fig. 6-2), integrated from 372 ratios of two single zircons and one small zircon fraction, and an age of 805 ± 2 Ma (296 ratios on 3 grains). Both ages are well constrained by repeated measurements yielding similar $^{207}\text{Pb}/^{206}\text{Pb}$ ratios. Most zircons were evaporated as single grains to avoid mixing of different populations.

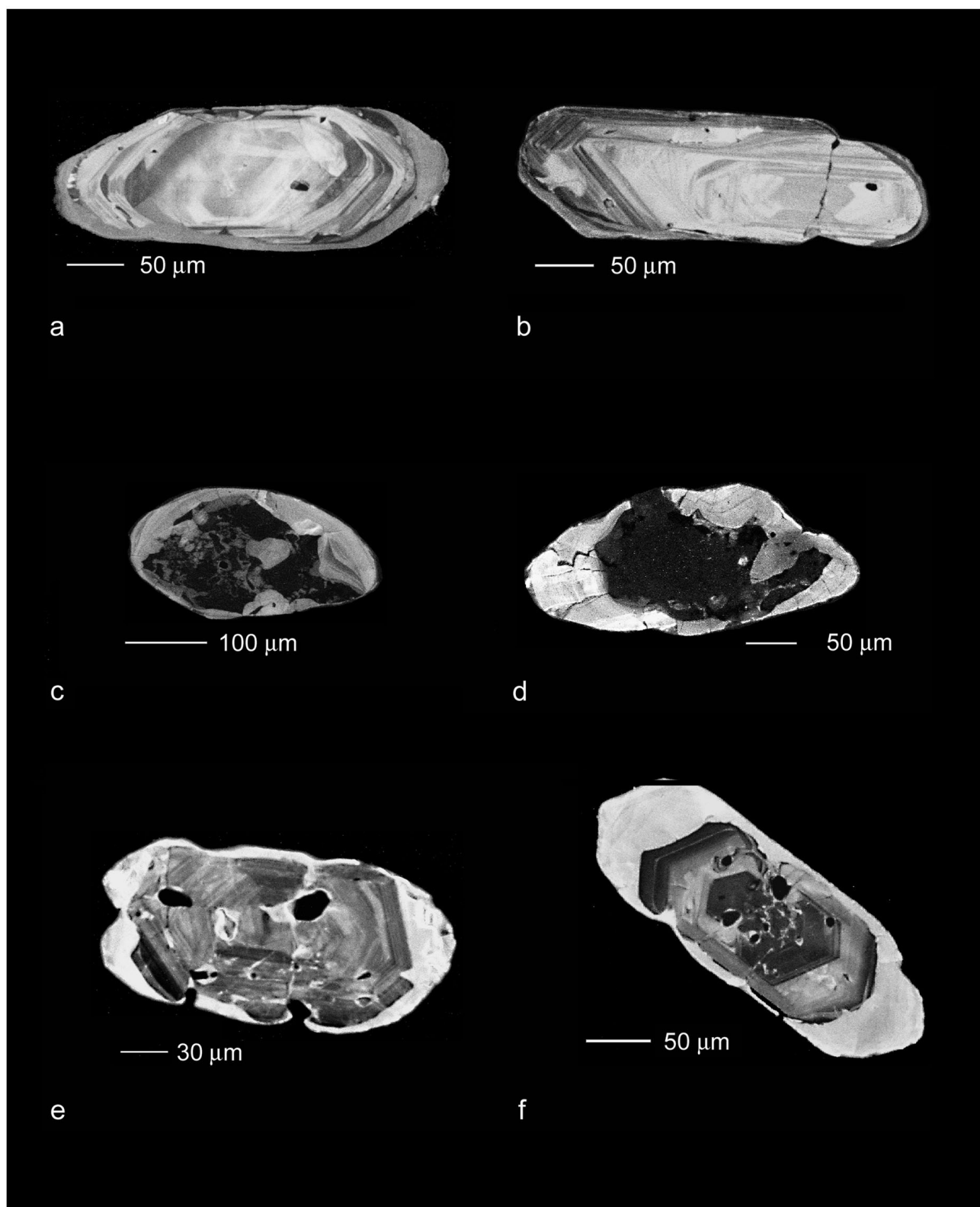


Figure 6-1. CL-images of zircons from MLC samples ZZB 123 and ZZB 166

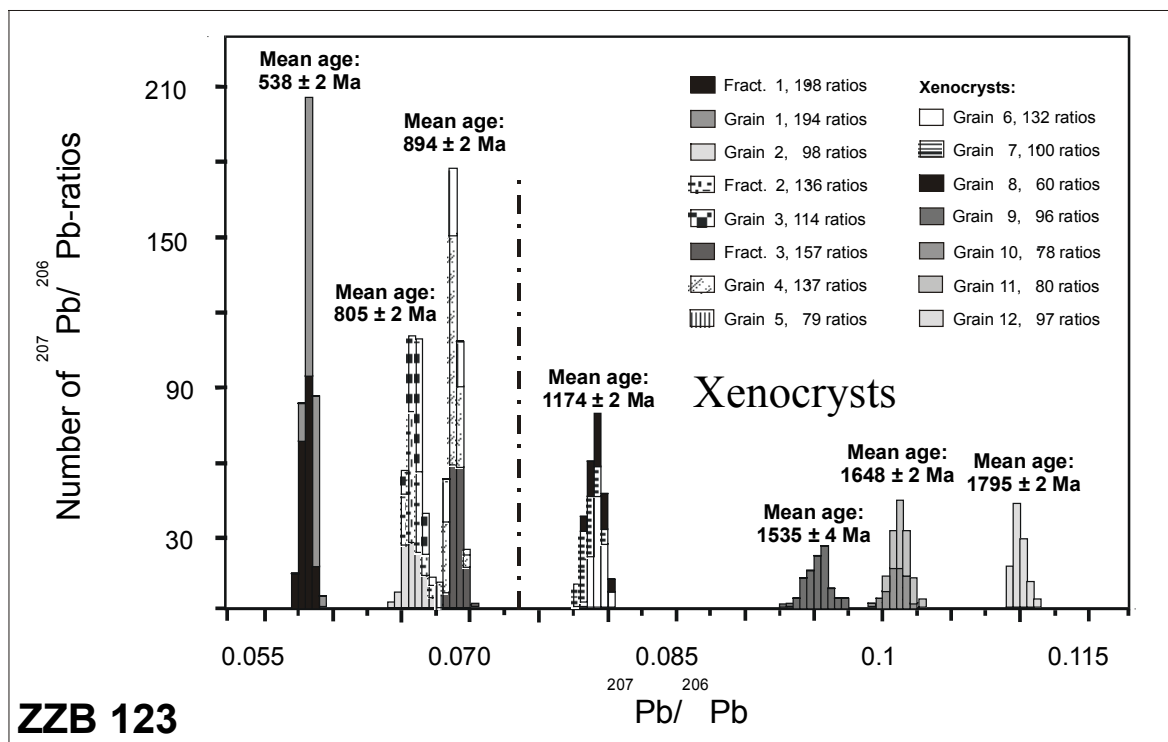


Figure 6-2. Histogram showing distribution of evaporation ages derived from zircons of metagabbro sample ZZB 123.

Five zircons of the first population were additionally analysed by the vapour transfer method. The data points are arranged along a discordia (MSWD: 1.9) that was forced through the origin. The discordia defines an upper intercept age of 872 ± 21 Ma with the concordia. The zircons are 10 to 30 % discordant (Fig. 6-3), indicating that variable recent lead loss has occurred.

The second population consists of dark brown, cloudy, prismatic zircons and zircon aggregates. Most of these are rounded to anhedral. In CL images the zircons of this population show either diffuse internal structures or none at all. The interior (Fig. 6-1 c) consists mainly of low luminescent recrystallised fragments and shows signs of strong resorption. The zircon exhibits younger overgrowth which follows the former corroded grain surface. Most of the zircons contain inclusions of hornblende, apatite and K-feldspar.

Single zircon evaporation yielded $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 1535 ± 4 Ma, 1648 ± 2 Ma and 1795 ± 2 Ma (Fig. 6-2) for this population.

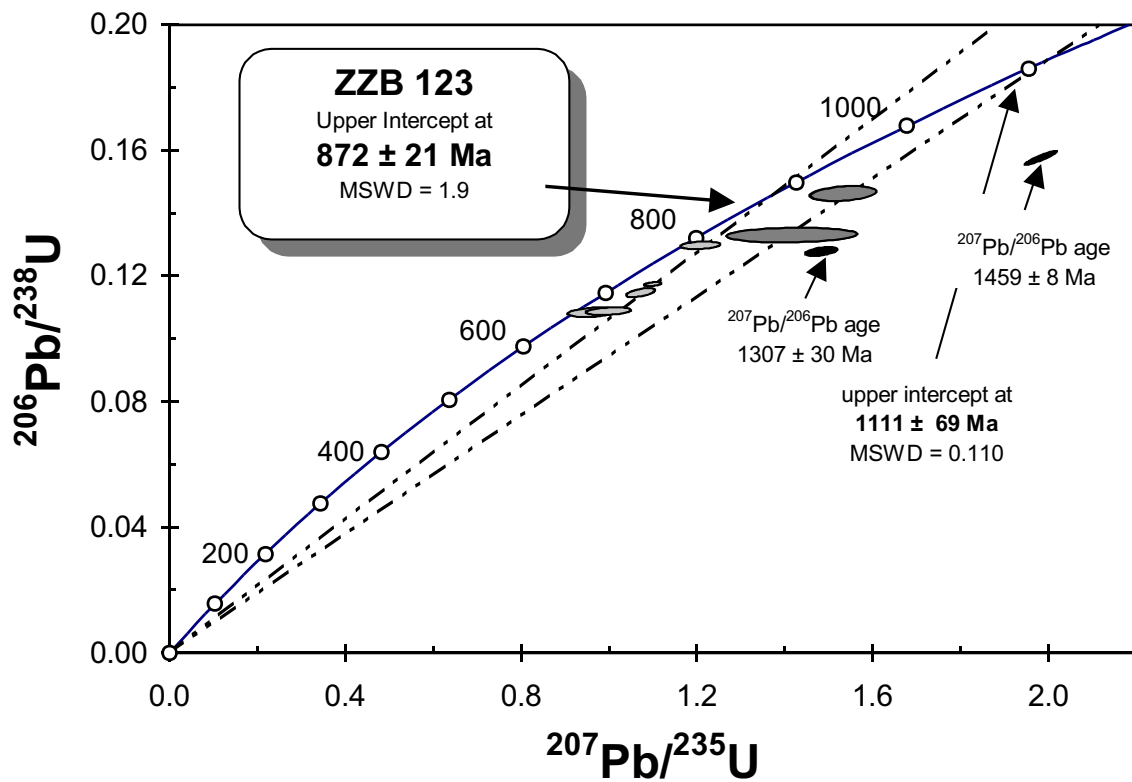


Figure 6-3. Concordia diagram showing analytical data for zircons from layered metagabbro sample ZZB 123. Dark grey and black symbols represent the second and third morphological populations.

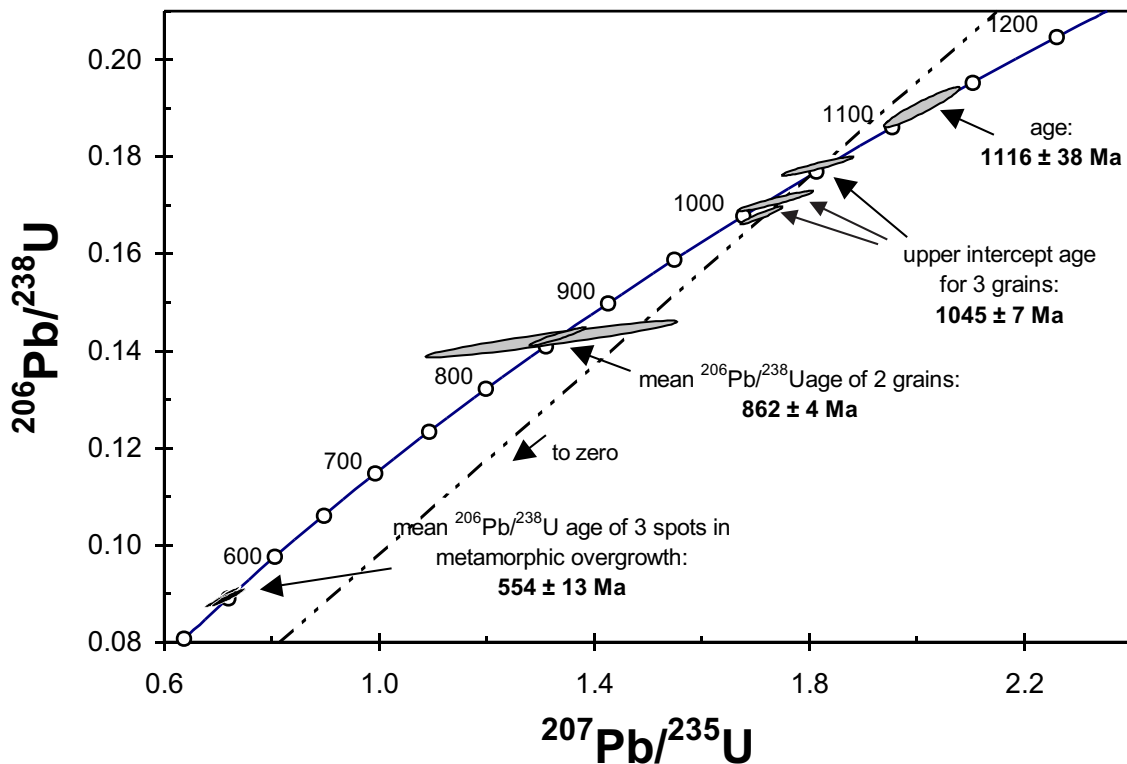


Figure 6-4. Concordia diagram showing distribution of SHRIMP data derived for zircons from layered metagabbro sample ZZB 123. Errors are 1 σ .

Two zircons of the second population analysed by VT yielded $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 1307 ± 59 Ma and 1459 ± 8 Ma as shown in the concordia diagram (Fig. 6-3). These zircons show a major influence of late Pan-African metamorphism because they plot near the lower concordia intercept of 449 ± 71 Ma (not shown). The lower intercept reflects the combined effects of metamorphic overgrowth and lead loss.

Zircons of the third population are pinkish-brown, slightly clouded, long prismatic and have subrounded terminations. The CL image (6-1 d) shows a corroded grain. The low luminescent core exhibits diffuse and heterogeneous internal structures. A highly luminescent, nearly homogenous overgrowth resulted in a rounded zircon grain. Hornblende, rutile, quartz and plagioclase are abundant inclusions.

Evaporation of single zircons of the third population yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1174 ± 2 Ma (Fig. 6-2). An upper intercept age of 1111 ± 71 Ma (Fig. 6-3; MSWD: 0.1) is defined by two zircons analysed by VT. The discordia defined by these two data points was forced through the origin.

Population four consists of round, football-shaped zircons which are clear and light brown. Also typical for this population is the formation of anhedral zircon aggregates. These aggregates were probably the result of agglomeration and amalgamation of pre-existing zircons during a phase in which the growth of zircon was possible such as a metamorphic event.

Evaporation of two grains of this population yielded a mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 538 ± 2 Ma (Fig. 6-2, TableV-1, Appendix V).

Eight grains of this sample were analysed on the SHRIMP II ion-microprobe at Curtin University, Australia. The resulting ages are shown in the concordia diagram of Figure 6-4, and Figure 6-5 illustrates CL images of the investigated zircons.

Two grains exhibit a highly luminescent core and only a very weak (Fig. 6-5 a, b) oscillatory zoning in CL image. In addition, the prismatic grain in Figure 6-5 a shows an outer part which has lower luminescence than the core. After a period of resorption, both grains were overgrown by a small rim, probably of metamorphic origin. These grains yielded a $^{206}\text{Pb}/^{238}\text{U}$ age of 862 ± 4 Ma, which agrees within error with the age of 872 ± 21 Ma, determined by the vapour transfer method for the first morphological population. Zircons of metagabbros with similar weak internal structures were reported for the metagabbros of the Val Barlas-ch, Switzerland (Poller, 1997). Because of these weak internal structures in cathodoluminescence imaging the zircons of the Val Barlas-ch metagabbros were interpreted to reflect the crystallisation age of those metagabbros.

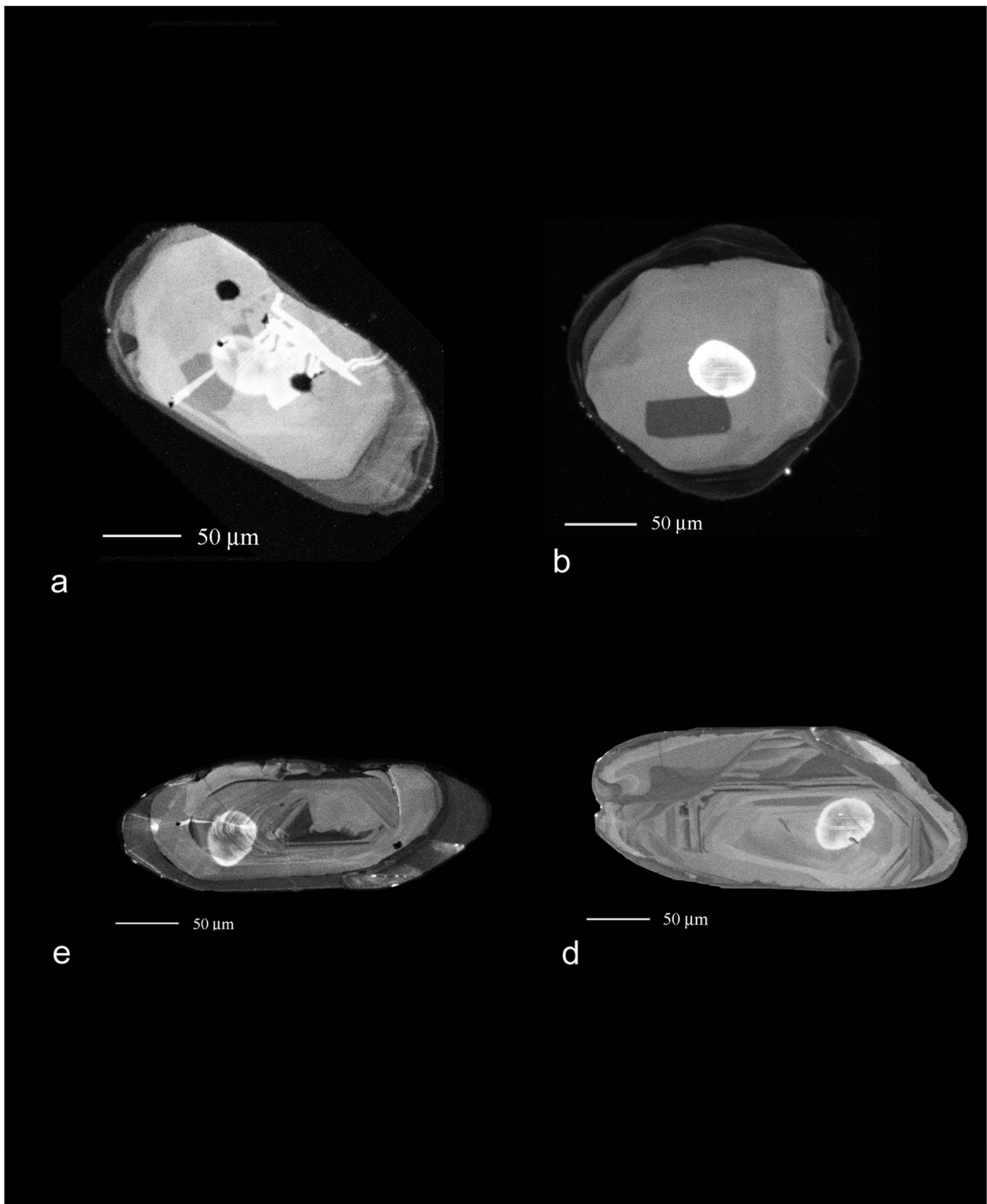


Figure 6-5. CL-images of zircons from MLC sample ZZB 123 measured on the SHRIMP II ion-microprobe.

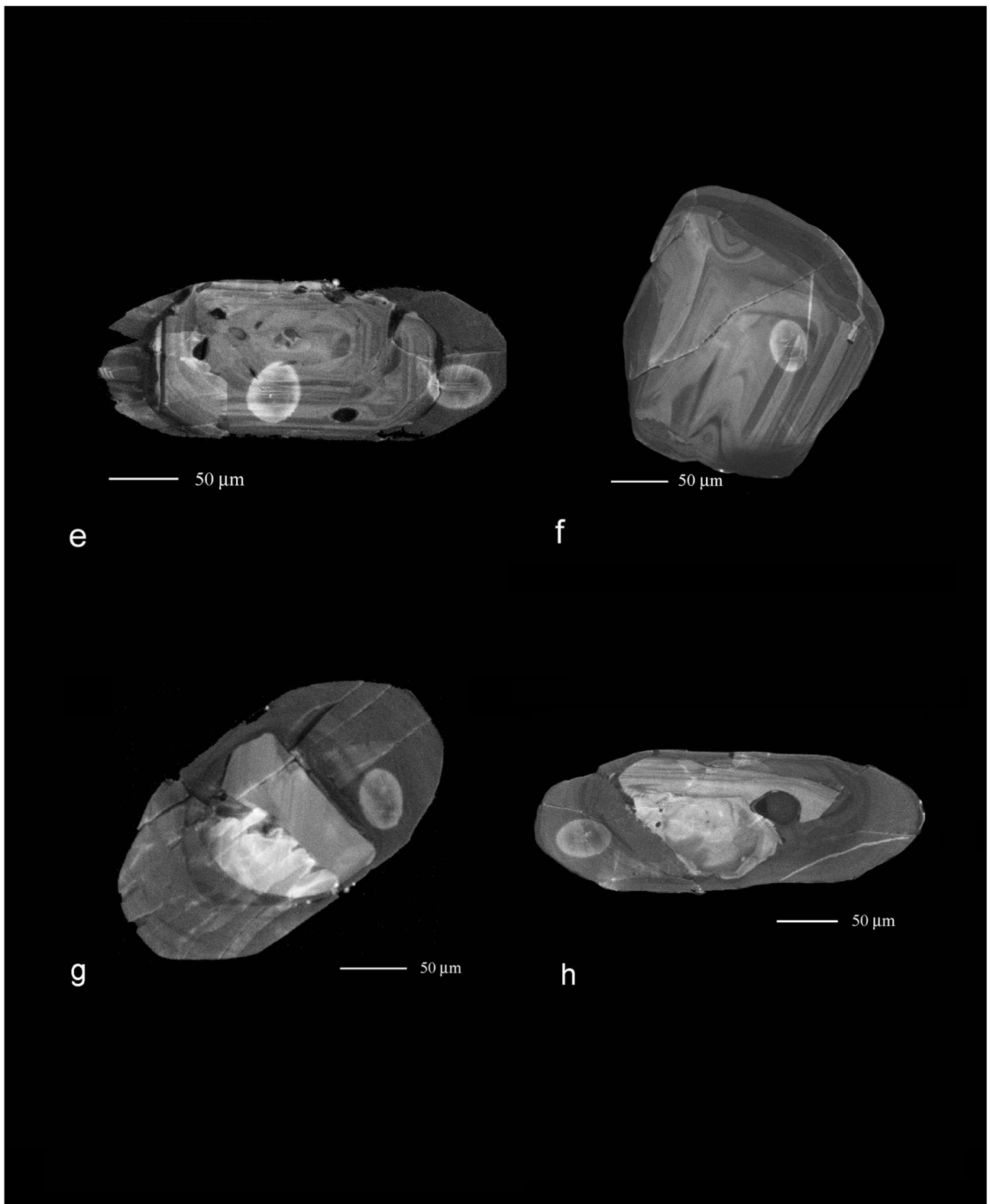


Figure 6-5. continued

Another prismatic zircon displays homogeneous oscillatory zoning and a partly recrystallised core in CL image (Fig. 6-5 c). The grain in figure 6-5 d shows a partly oscillatory zoned, diffuse interior. After a period of resorption both grains were overgrown by an inner rim. A third zircon (Fig. 6-5 e) contains an inherited core which is overgrown by an inner, oscillatory zoned rim. After marginal resorption, the these three grains were overgrown by an outer rim which is lower luminescent than the inner rim and the core.

Data points of these three grains are somewhat discordant but define a chord with an upper intercept age of 1045 ± 7 Ma (Fig. 6-4).

A sixth grain (Fig. 6-5 f) shows diffuse structures in its interior and appears zoned in the outer part on the right side. A spot located on this zoned side produced a $^{206}\text{Pb}/^{238}\text{U}$ age of 1136 ± 38 Ma (Fig. 6-4).

Grain g exhibits a weakly zoned, diffusely structured inherited core, which was partly resorbed. Grain h also contains an inherited core without internal structures. Both grains were overgrown by an inner rim that is lower luminescent than the core, and by an outer rim which is slightly higher luminescent than the inner rim. Three spots were positioned on this outer rim which is interpreted to be a metamorphic overgrowth (Fig. 6-5 e, g and h). These spots on three zircons provided concordant SHRIMP results with a mean $^{206}\text{Pb}/^{238}\text{U}$ age of 554 ± 13 Ma (Fig. 6-4).

6.2.1.2 Ferro-metagabbro sample ZZB 166

The zircons of this sample are pale brown to pinkish-brown, clear and long prismatic, with slightly to well rounded terminations. In CL images (Fig. 6-1 e, f) the zircons show oscillatory magmatic zonation. In nearly all cases the older grain surface is corroded. The zircons are overgrown by a younger, locally dominant highly luminescent rim. Nearly all zircons contain inclusions of apatite, titanite, plagioclase, K-feldspar, hornblende and quartz. Only long prismatic zircons with slightly rounded terminations were used for U-Pb and Pb-Pb analyses.

Seven zircons were analysed from this sample. Three zircons with a mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 851 ± 4 Ma are weakly discordant. These and the four other analysed zircons yielded an upper intercept age of 871 ± 19 Ma in the $^{207}\text{Pb}/^{235}\text{U}$ vs. $^{206}\text{Pb}/^{238}\text{U}$ diagram (Fig. 6-6; MSWD: 0.54). The zircons experienced variable recent lead loss to define a discordia line. Application of the Pb-Pb single zircon evaporation technique yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 849 ± 2 Ma (Fig. 6-6, inset).

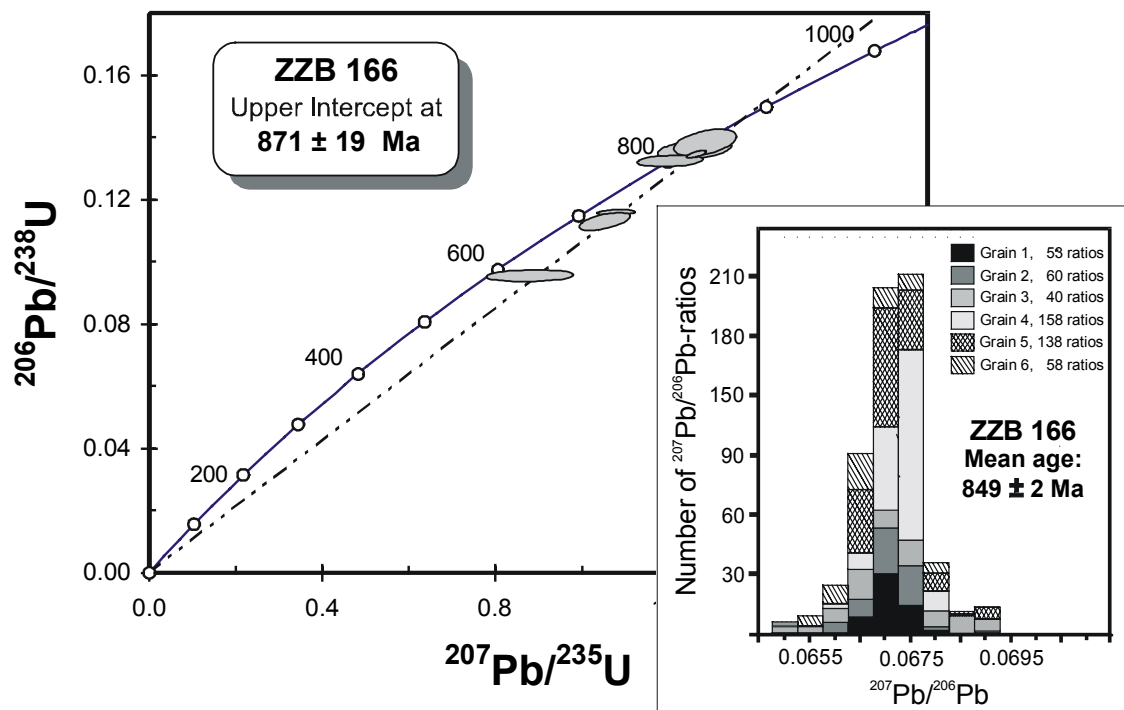


Figure 6-6. Concordia diagram showing distribution of vapour transfer analysed zircons of ferro-metagabbro sample ZZB 166. Inset: Histogram showing the results of zircon evaporation as integrated from 507 ratios.

6.2.1.3 Summary of age interpretation for zircons from the Mavuradonha Layered Complex

Zircons of a metagabbro and a ferro-metagabbro sample were used to determine the age of crystallisation of the Mavuradonha Layered Complex as well as the age of a metamorphic overprint. Since four zircon populations were found in the metagabbro sample (ZZB 123), the results of the age determination permit to draw some conclusion about the evolution of the Mavuradonha Layered Complex.

The first population of sample ZZB 123 and the zircons of sample ZZB 166 are interpreted as a magmatic population based on the results of CL imaging. All zircons of this population are nearly euhedral in shape and exhibit oscillatory zoning. Zircons from both samples yielded the same upper intercept age ($872 \pm 21 \text{ Ma}$ and $871 \pm 19 \text{ Ma}$) within error as determined by two different methods (SHRIMP and vapour transfer method). Zircons analysed by SHRIMP are interpreted to reflect the crystallisation age because of concordance of the data points and their weak internal structures in cathodoluminescence imaging. Therefore, the concordant SHRIMP age of $862 \pm 4 \text{ Ma}$ is interpreted as reflecting

the time of emplacement of the Mavuradonha Layered Complex magma and crystallisation of the layered rocks. This age confirms early Neoproterozoic emplacement of the Mavuradonha Layered Complex.

Interpretation of the discordance of the magmatic zircon populations is a reflection of recent lead loss and late Pan-African metamorphism. The evaporation technique was not successfully to provide primary emplacement ages for both samples. The Pb-Pb age of 849 ± 2 Ma for zircons of sample ZZB 166, together with most of the evaporation data for zircons of sample ZZB 123, are interpreted to reflect lead loss that affected the entire grain. The evaporation technique provided significant variation in age data, where zircons comprising ages greater than 900 Ma are probably inherited because of increasing Pb-Pb ratios during evaporation and their shape.

Age determinations on zircon overgrowths around an oscillatory zoned zircon (Fig. 6-5 e) and two zircon fragments (Fig. 6-5 g and h) using SHRIMP yielded a concordant age of 554 ± 13 Ma. This is interpreted as the age of the high-grade metamorphic event during which this overgrowth formed. Such overgrowths are visible on nearly all of the investigated zircons. The $^{207}\text{Pb}/^{206}\text{Pb}$ age of 538 ± 2 Ma which was determined by evaporation on the fourth population of sample ZZB 123 can be ascribed to the same metamorphic event. The slightly lower age may be the result of lead loss. This metamorphic age indicates that the complex was overprinted under high-grade metamorphic conditions in late Pan-African times, at least 300 Ma after the emplacement of the complex.

The four oldest SHRIMP-analyses are interpreted to reflect xenocrysts. In addition, the internal structures (Fig. 6-5 c, d, e, f) observed during CL imaging show that some zircons were overgrown twice, one grain contains an inherited core component, and some of the zircons also show diffuse internal structures. The zircons of the second and third morphological populations are also interpreted as xenocrysts. The zircon grains are well rounded and show metamorphic overgrowth by a luminescent rim in CL images. Internal structures are diffuse, and an oscillatory zoning is lacking. Some of these zircons recorded decreasing $^{207}\text{Pb}/^{206}\text{Pb}$ ratios during evaporation. This is due to evaporation of outer zones of the zircons being younger in age, and of a different Pb-Pb isotopic composition than the core. These xenocrysts provide evidence for a crustal setting of the Mavuradonha Layered Complex and document contamination of the mafic magma with lower to mid-crustal rocks. Therefore, the xenocrysts of sample ZZB 123 may have been picked up during intrusion of the mafic melts at the base of lower continental crust.

6.2.2 Pegmatites within the Mavuradonha Layered Complex

Pegmatites are found in the entire Mavuradonha Mountain range and in the Ocellar Gneiss. Two pegmatites cross-cutting the igneous layering of the Mavuradonha Layered Complex were investigated. The petrographic description is given in Chapter 4.

6.2.2.1 ZZB 60: pegmatite, northern Mavuradonha Mountains

The zircons of this pegmatite exhibit homogeneous magmatic oscillatory zoning in CL images. Some crystals show clear evidence for very tiny younger overgrowth (Fig 6-7, grain a).

Figure 6-7 b shows a partly recrystallised interior. Other zircons of this sample show evidence of metamictisation in their interior. Metamictisation and recrystallisation are mechanisms causing Pb-loss and, therefore, discordant ages. This sample contains zircons that show clear evidence for inherited core components. These older cores are overgrown by oscillatory zoned rims of different size.

Six zircons were dated from this pegmatite. Three of these yielded concordant data points with ages of 524 ± 6 Ma, 497 ± 7 Ma and 471 ± 10 Ma, respectively (Table V-2, appendix A-V). Three zircons define a chord with an upper intercept age of 540 ± 19 Ma (Fig 6-8 a) because they are weakly to significantly discordant, but well aligned (MSWD: 0.1). The lower intercept of this chord was forced through the origin. The concordant zircons are interpreted to be shifted along a chord which is indistinguishable from the concordia.

The differences in age are probably the result of lead loss due to metamictisation and recrystallisation which were seen in some of the zircons during the CL imaging study (Fig. 6-7 b).

Evaporation of three identical grains yielded a mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 555 ± 2 Ma (Fig 6-8 a inset) which is consistent, within error, with the upper intercept age of 540 ± 19 Ma.

6.2.2.2 ZZB 128: pegmatite, south of the Mavuradonha Mts.

The zircons of this pegmatite are elongated, clear and colourless. The terminations vary from weakly to well rounded. Typical for most of these zircons is a low luminescence (Fig. 6-7 c) and the lack of any significant internal structure (Fig. 6-7 c, d). Most grains show only weak oscillatory zoning.

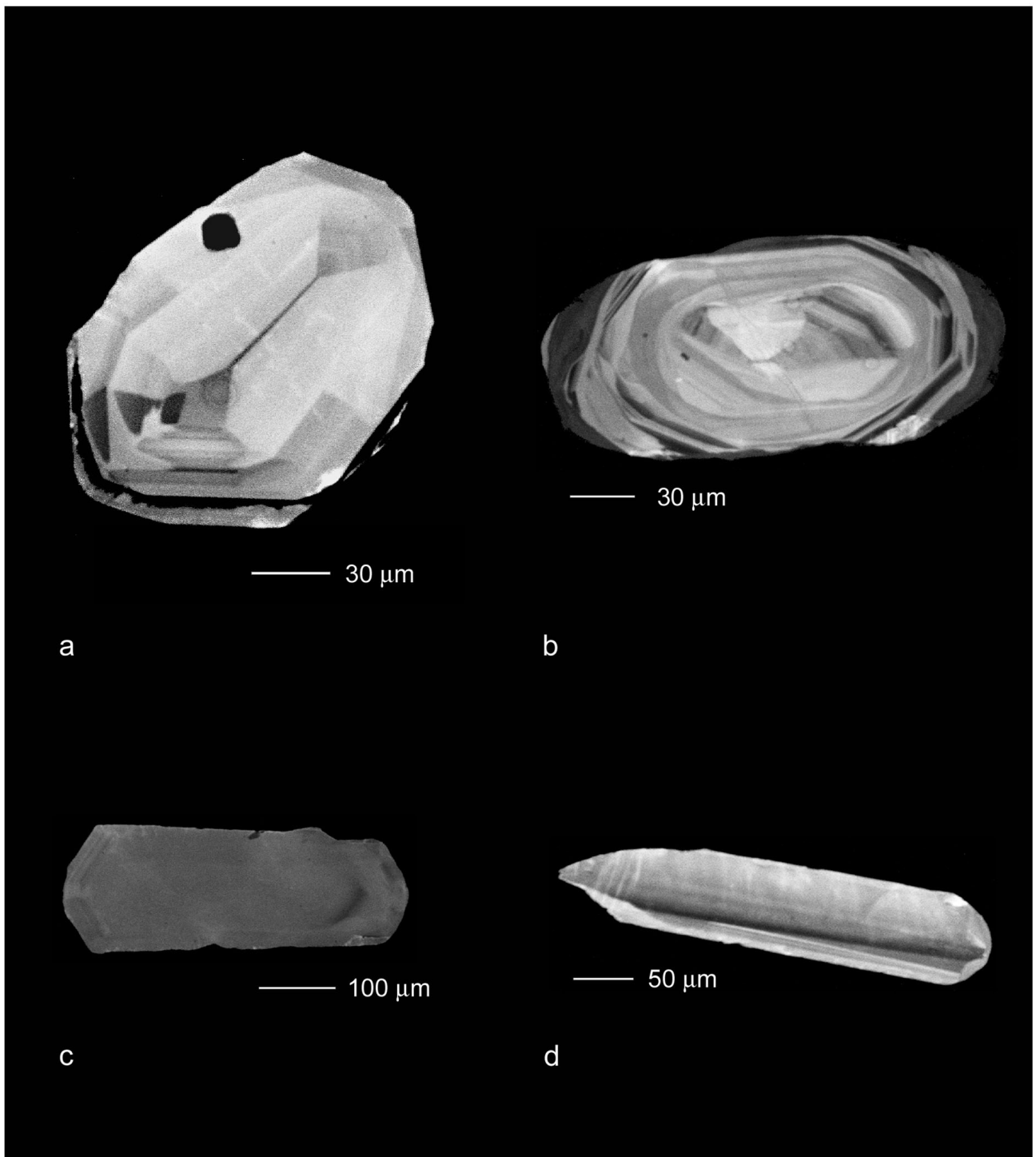


Figure 6-7. Zircons from pegmatite samples ZZB 60 and ZZB 128

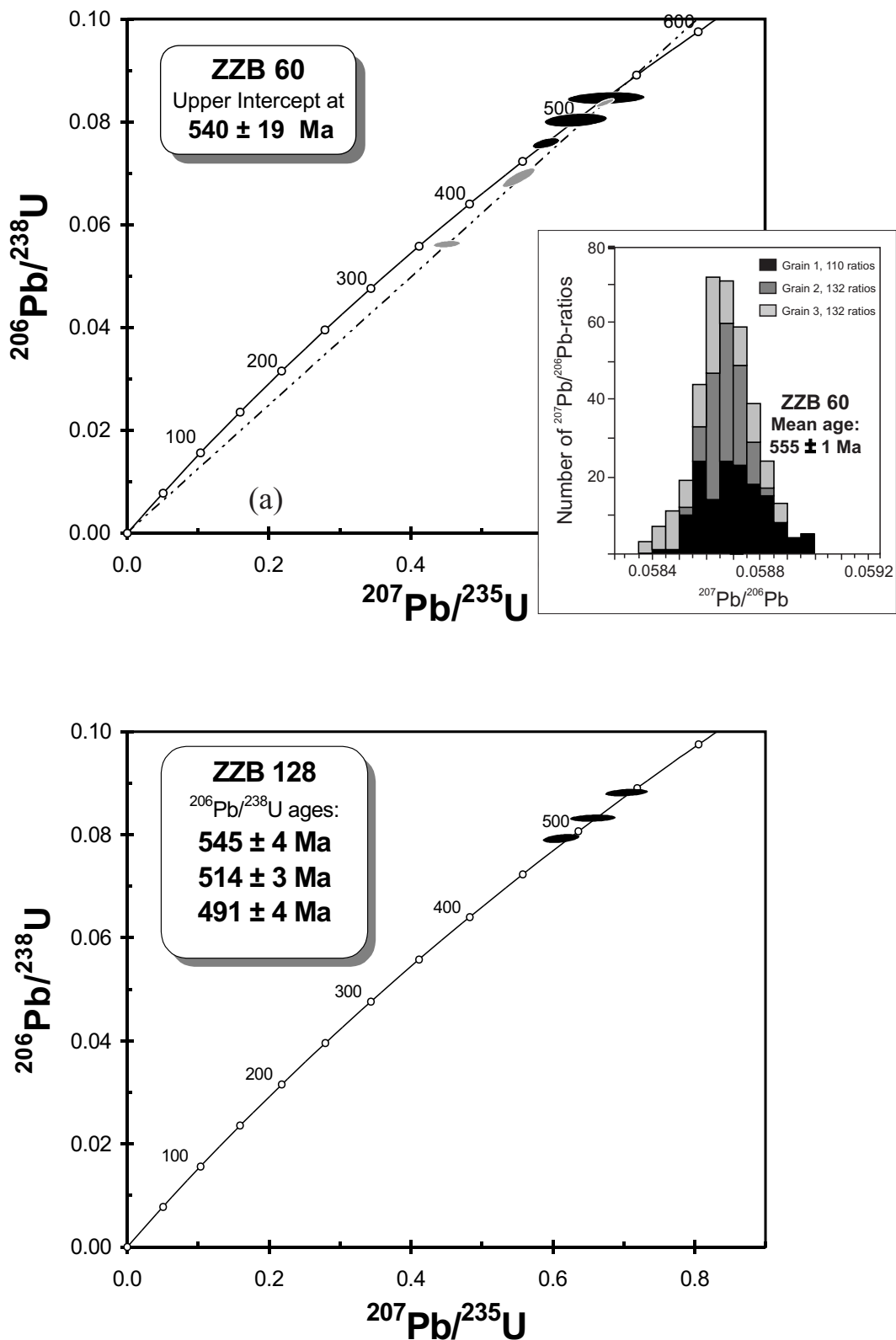


Figure 6-8 a and b. Concordia diagrams and histogram showing the analytical results for zircons from two pegmatites ZZB 60 (a) and ZZB 128 (b). The inset in figure (a) shows the distribution of the integrated 374 radiogenic Pb isotope ratios. Black ellipses mark concordant data points, whereas grey ellipses represent three data points that define the chord.

Vapour transfer analyses of three zircons resulted in concordant data points (Fig. 6-8 b), which yielded ages of 545 ± 4 Ma, 514 ± 3 Ma and 491 ± 4 Ma, respectively. The latter zircon is slightly reversely discordant. The two youngest grains are interpreted to define a discordia which is indistinguishable from the concordia. This discordance indicates a disturbed U/Pb system, in which Pb-loss has occurred.

6.2.2.3 Interpretation

The U-Pb upper intercept age of 540 ± 19 Ma for sample ZZB 60 is considered to reflect the crystallisation age of this pegmatite. The $^{207}\text{Pb}/^{206}\text{Pb}$ evaporation age of 555 ± 1 Ma is slightly older than the U-Pb ages and probably indicates the influence of inherited cores, a texture which was observed during CL imaging in several zircons. The three concordant zircons of this sample display minor lead loss and, therefore, it seems that these data points were shifted along the concordia.

The concordant age of 545 ± 4 Ma of sample ZZB 128 is also interpreted as the time of crystallisation of this pegmatite. Pb-loss is a reasonable explanation for the discordant zircons.

Both pegmatite samples yielded similar late Pan-African crystallisation ages around 545 Ma and reflect a widespread tectonothermal event in the Zambezi (Vinyu et al. 1999; Hanson et al. 1998). This tectono-thermal event seems to have been contemporaneous with the amphibolite-facies retrogression which was dated by the garnet - whole-rock technique.

6.2.3 Ocellar Gneiss

6.2.3.1 ZIM 30 (Fig. 6-9 a & b; Fig. 6-10 a)

Most zircons from this sample are long-prismatic with slightly rounded terminations. Some grains are well rounded. The zircons are predominantly clear, but some are yellow-brown and cloudy, some grains contain visible inclusions. In CL images the prismatic zircons display homogeneous oscillatory zoning (Fig 6-9 a) which is typical for magmatic crystallisation. After marginal resorption one zircon was overgrown by a highly luminescent rim. A second zircon (Fig. 6-9 b) from this sample is a “three-phase” grain and contains an inherited core which is overgrown by an oscillatory zoned first rim. After a period of resorption the zircon was overgrown by a second highly luminescent rim.

Vapour transfer results for air-abraded single zircon grains of this sample are well aligned in the $^{206}\text{Pb}/^{238}\text{U}$ vs. $^{207}\text{Pb}/^{235}\text{U}$ concordia diagram (Fig. 6-10 a) and define a discordia line (MWSD: 0.7) with an upper intercept age of 1026 ± 28 Ma. The lower

intercept of 107 ± 63 Ma seems to have no geological significance. The discordance is interpreted to have resulted from recent lead loss. Alternatively the lower intercept could be interpreted as due to recent lead loss combined with the metamorphic influence by minor, younger zircon overgrowth (perhaps at ~ 550 Ma) around old zircon grains with an age of ~ 1030 Ma.

The upper intercept age is consistent with the $^{207}\text{Pb}/^{206}\text{Pb}$ single zircon evaporation age of 1011 ± 1 Ma (Fig. 6-10 a, inset).

6.2.3.2 ZIM 56 (Fig. 6-9 c & d; Fig. 6-10 b)

The zircons of this sample are morphologically similar to those in sample ZIM 30, they are long-prismatic, colourless and cloudy. Another population has a brown colour with colourless pyramids. The third population is dark brown and has a long-prismatic and pyramidal shape. Many grains exhibit low luminescence. In CL images the zircons are oscillatory zoned (Fig. 6-9 c) and are interpreted to have grown during a single magmatic phase. The core of one grain (Figure 6-9 c) is partly recrystallised, and the primary zoning is lost. Another zircon (Fig 6-9 d) has a more complex internal structure. The grain contains a large recrystallised core, and an oscillatory zoned outer part. The upper pyramid is in its interior recrystallised. Both zircons were partly overgrown by a thin highly luminescent rim after a period of resorption.

Sample ZIM 56 yielded one concordant data point, with a $^{206}\text{Pb}/^{238}\text{U}$ age of 1022 ± 7 Ma. Three further grains define a discordia which was forced through the concordant data point of 1022 ± 7 Ma (Fig. 6-10 b; MSWD: 2.3) taken as the upper intercept. The lower intercept of 190 ± 10 Ma is considered to be without geological significance and is probably the result of variable lead loss. An alternative interpretation would be a combined effect of lead loss and minor younger zircon growth around the older zircons.

The concordant age of 1022 ± 7 Ma is consistent with the $^{207}\text{Pb}/^{206}\text{Pb}$ evaporation age of 1018 ± 1 Ma (Fig. 6-10 b, inset).

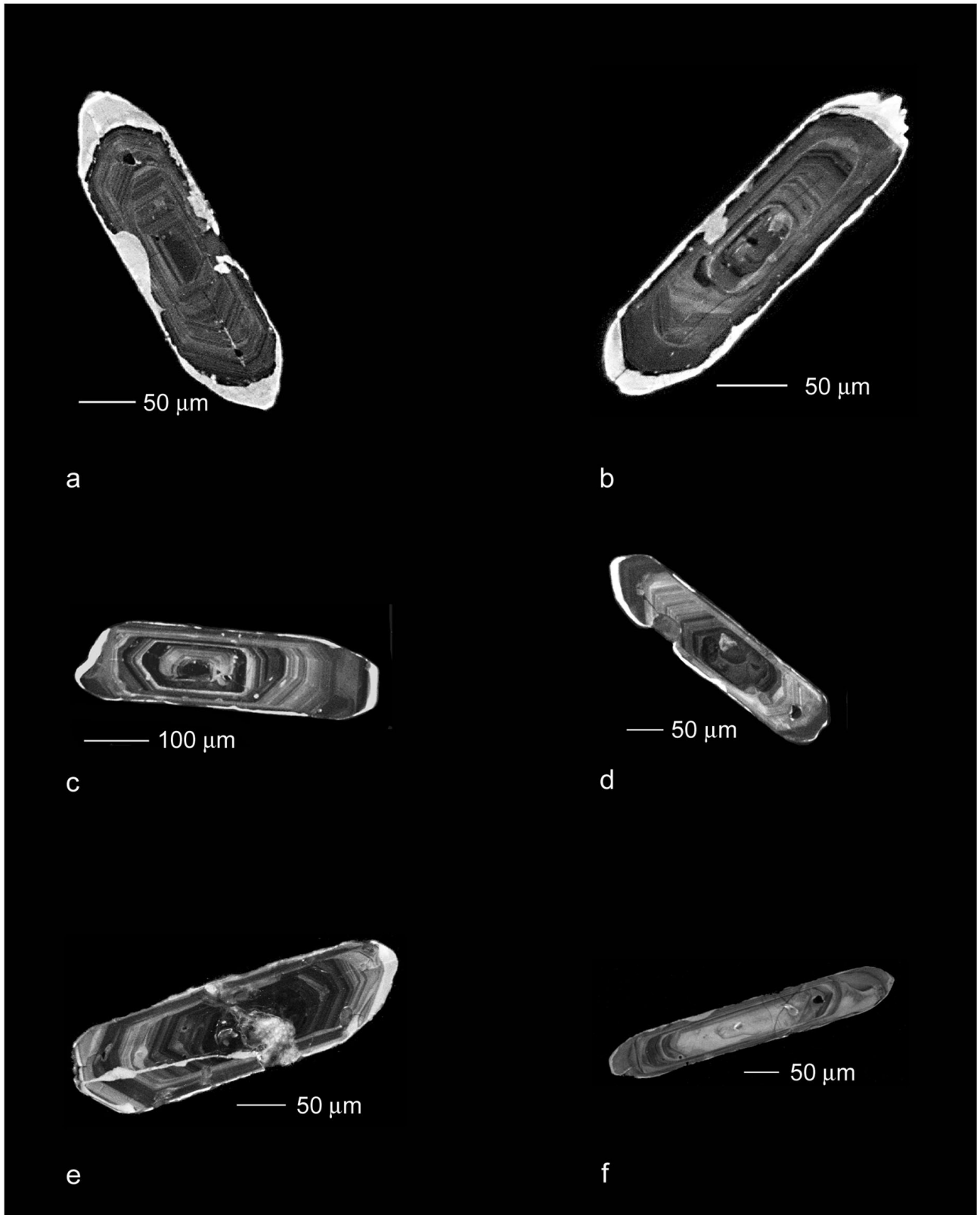


Figure 6-9. CL-images of typical zircons from Ocellar Gneiss samples ZIM 30, ZIM 55 and ZIM 56

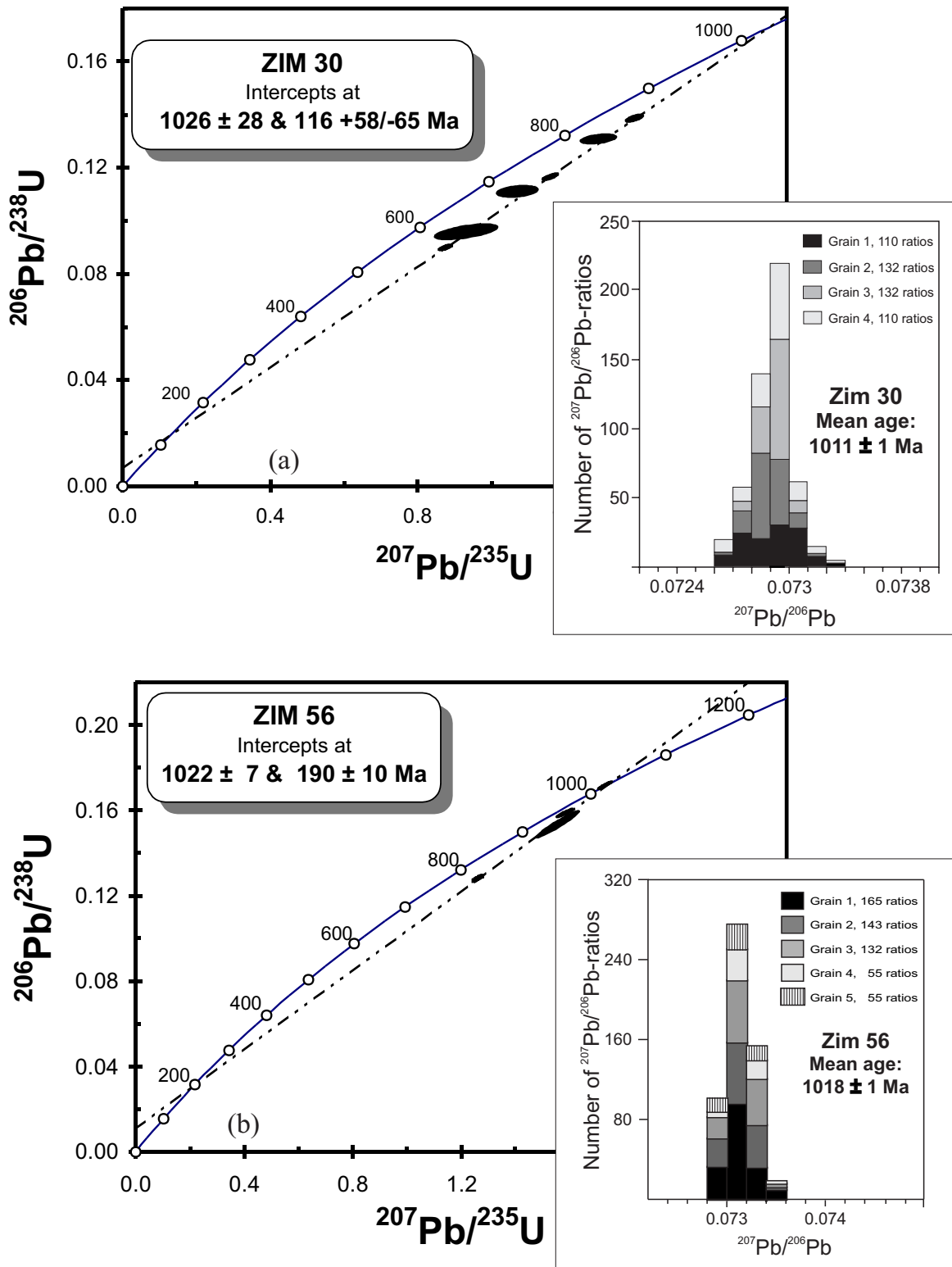


Figure 6-10 a and b. Concordia diagrams and histograms showing vapour transfer and evaporation data for Ocellar Gneiss samples ZIM 30 and ZIM 56. Note the good agreement of U-Pb and Pb-Pb ages in both samples. The insets show the spectrum for zircons of both samples, integrated from 484 ratios (a) and 550 ratios (b).

6.2.3.3 ZIM 55 (Fig. 6-9 e & f; Fig. 6-10 c)

The zircons of this sample have a pyramidal and long-prismatic shape. They are slightly to well rounded, cloudy and colourless, or have a brown colour (typical zircons from this sample: Fig. 6-9 e, f). They also contain inclusions of K-feldspar, quartz and apatite. In CL image one zircon (Fig. 6-9 e) has a small, rounded, inherited component. The central part is low luminescent, and cracks are healed. The inherited core is overgrown by an oscillatory zoned part which is partly corroded and recrystallised. The second grain (Fig 6-9 f) is elongated with a nearly homogeneous, high luminescent core, whereas the low luminescent outer part is oscillatory zoned. The pyramids are homogeneous and recrystallised. Both zircons examined under CL are overgrown, after a period of resorption, by a thin highly luminescent rim. Most of the other zircons analysed by CL images show completely recrystallised and diffuse interiors, as well as inherited cores.

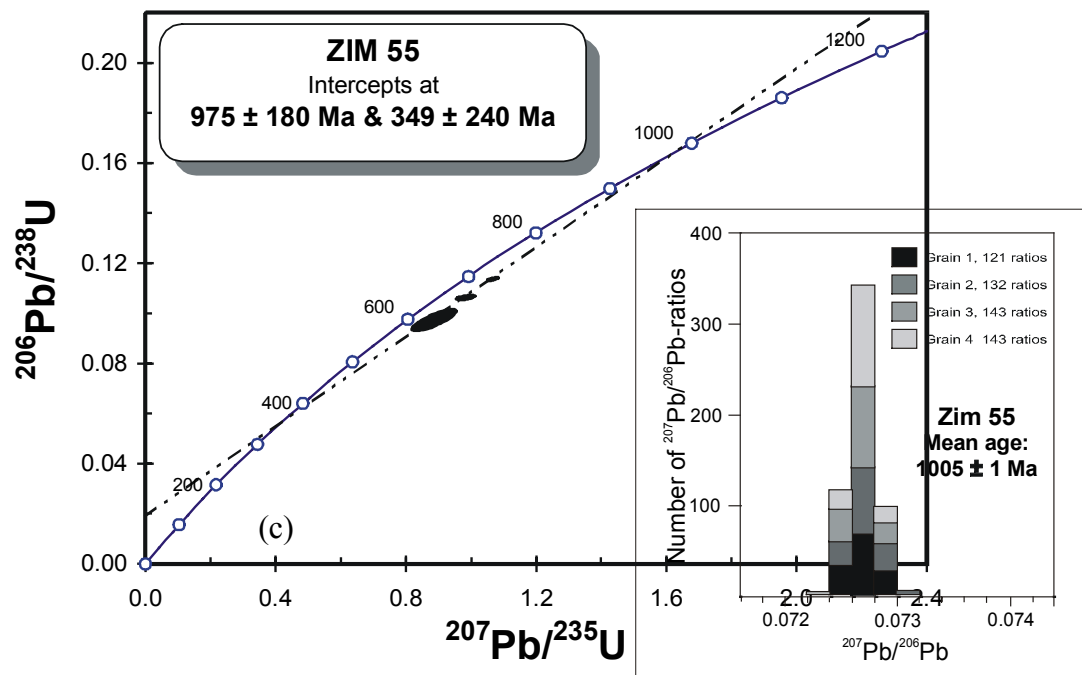


Figure 6-10 c. Concordia diagram and histogram for zircons from Ocellar Gneiss sample ZIM 55. Note large error in U-Pb age for the vapour transfer analyses, whereas the Pb-Pb age is consistent with ages obtained for the other two Ocellar Gneiss samples. Inset shows distribution of radiogenic Pb ratios for four zircons, integrated from 539 ratios.

The sample was dated by vapour transfer analyses of three cloudy, colourless and elongated zircons. They define a discordia with an upper intercept age of 975 ± 180 Ma (Fig. 6-10 c; MSWD: 0.02). The zircons were about 30 - 50 % discordant indicating significant Pb loss. This upper intercept age of 975 ± 180 Ma was not considered in the following discussion because of the unreliably large errors. The lower intercept of $349 \pm$

240 Ma is considered to be without geological significance.

In contrast, the $^{207}\text{Pb}/^{206}\text{Pb}$ evaporation age yielded a mean age of 1005 ± 1 Ma (Fig. 6-10 c, inset) which is in the range of the two other samples dated (ZIM 30 and ZIM 56).

6.2.3.4 Summary and interpretation of the Ocellar Gneiss

Two samples of the Ocellar Gneiss (ZIM 30 and ZIM 56) exhibit nearly identical U-Pb ages, which are also consistent with the single zircon Pb-Pb evaporation ages. The third sample (ZIM 55) yielded a Pb-Pb age which is in the range of the two other samples. The U-Pb data of sample ZZB 55 of 975 ± 180 Ma is, within error, consistent with U-Pb ages of ZIM 30 and ZIM 56.

Considering the U-Pb and Pb-Pb ages, especially the concordant age of sample ZIM 56, and taking into account the results of the CL-images, the concordant age of 1022 ± 7 Ma is interpreted as the most likely time of emplacement of the granitoids from which the Ocellar Gneiss evolved.

6.3 Rutile U-Pb geochronology

Rutile fractions were separated from a granulite-facies meta-anorthosite and considered to be metamorphic in origin because of its occurrence and shape in thin section. The results for rutile fractions of this meta-anorthosite are listed in Appendix V, Table V-6 and shown in Fig. 6-11.

Three fractions from the same sample were analysed for U-Pb isotopic composition. The fractions exhibit nearly similar U concentrations (5.535 to 5.668 ppm), whereas Pb concentrations and the $^{206}\text{Pb}/^{204}\text{Pb}$ ratios differ in the range of 0.567 to 2.371 ppm and 33 to 203, respectively.

Two rutile fractions yielded identical ages which are concordant and provide $^{206}\text{Pb}/^{238}\text{U}$ ages of 502 ± 6 Ma and 501 ± 6 Ma. The third fraction (Rt ZZB 150-4) yielded a slightly reversely discordant $^{206}\text{Pb}/^{238}\text{U}$ age of 526 ± 7 Ma, which is probably the result of the low $^{206}\text{Pb}/^{204}\text{Pb}$ ratio of 33.

The ages for the rutile fractions suggest cooling of the Mavuradonha Layered Complex to ~ 400 °C in an interval of 30 to 50 Ma considering the T_c of 420 ± 30 °C of rutile for the U-Pb system (Mezger et al., 1989).

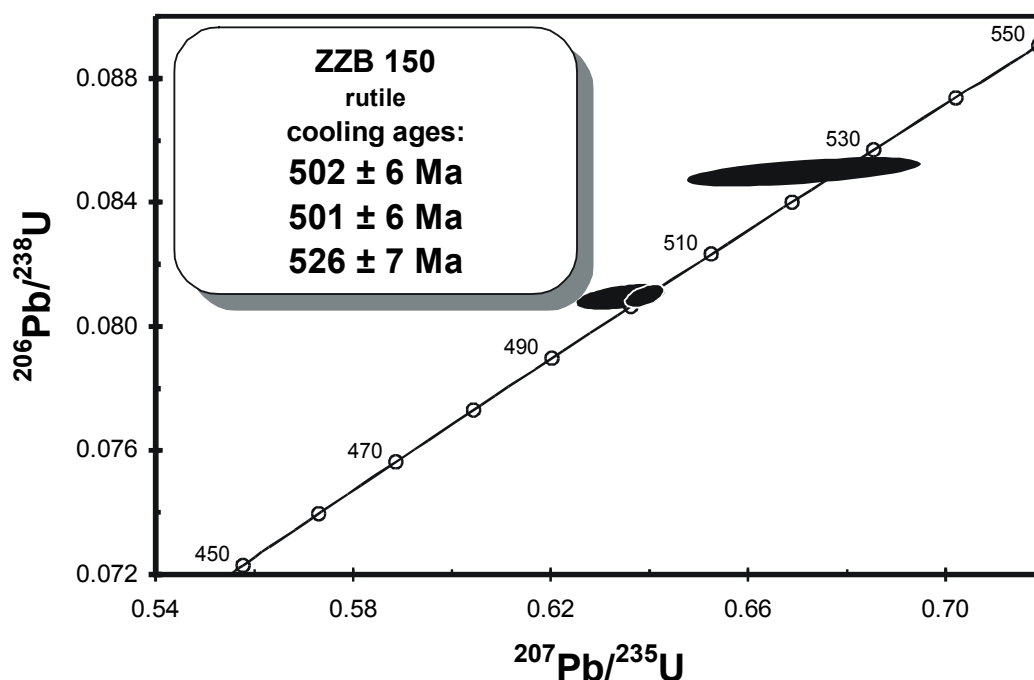


Figure 6-11. Concordia diagram showing the U-Pb isotopic composition of the three rutile fractions from meta-anorthosite sample ZZB 150 from the upper meta-anorthosite suite.

6.4 Sm-Nd garnet - whole-rock geochronology

Garnets are major-rock forming minerals in meta-anorthosites and ferro-metagabbros of the upper meta-anorthosite suite. These garnets record, in addition to the *PT* conditions for high-grade metamorphism (see Chapter 4), essential information on the age of metamorphism. Due to the fractionation of REE and the enrichment of heavy REE in garnet, garnet generally has high Sm-Nd ratios. Therefore, garnets can be used to obtain geochronological information using the Sm-Nd system if they are combined with other mineral phases or whole-rock data containing low Sm-Nd ratios. The use of the Sm-Nd dating method also has the advantage of measuring the isotopic composition of rock-forming minerals which are in equilibrium with the whole-rock. However, a limiting factor for the application of the Sm-Nd dating technique is the closure temperature. There is considerable controversy about the closure temperature (T_c) for Sm-Nd in garnet which depends on the grain size and the cooling rate of rocks (Cohen et al., 1988; Mezger et al. 1992; Burton et al., 1995; Hensen & Zhou, 1995; Jung & Mezger, 2001). Considering a T_c for slowly cooling terrains of 600 ± 30 °C of Mezger et al. (1992;) and the T_c of 640 – 820 °C of Burton et al. (1995) only cooling ages for HT terrains can be determined and,

therefore, the Mavuradonha Layered Complex garnets may record cooling ages. Therefore, the ages provided by these metamorphic garnets can be used to determine the age of cooling on the *PT* path after peak metamorphism. Results are given in Appendix V, Table V-7.

6.4.1 Garnet - whole-rock age determinations

Four samples were chosen which show a range in $^{147}\text{Sm}/^{144}\text{Nd}$ ratios from 0.771 to 0.901, as well as a spread of 0.514811 ± 69 to 0.515292 ± 5 in the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio. The similarity in the isotopic composition of the garnets derived from meta-anorthosites and those from the metagabbros is striking. The garnet - whole-rock pairs yielded identical ages of 532 ± 18 Ma for ZZB 147, 545 ± 6 Ma for ZZB 148 (containing an inclusion-rich fraction, 547 ± 7 Ma for an inclusion-free garnet fraction), 545 ± 6 Ma for ZZB 149 and 549 ± 6 Ma for ZZB 146 (see Appendix V). These ages demonstrate that leaching was successful in removing inclusions. Any remaining and undetected inclusions have not altered the Sm-Nd ratios of the garnets significantly. In addition to the inclusion-free separate, an inclusion-rich split was analysed from one sample (ZZB 148). The inclusion-rich fraction was processed to check whether inclusions in garnets have any significant effect on the $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios. The large number of inclusions in the separate prior to leaching suggested a disturbance of the garnet isotopic composition. However, the measured ratios ($^{147}\text{Sm}/^{144}\text{Nd}$: 0.7249; $^{143}\text{Nd}/^{144}\text{Nd}$: 0.514686) indicate the opposite. The data point is shifted in the direction of the whole-rock and plots directly on the tie-line of this sample. This suggests that other minerals with a low Sm-Nd ratio like apatite were dissolved during the three leaching steps or were not present. Furthermore, it would appear that remaining inclusions have apparently been reset during garnet growth and thus yielded the same age. Alternatively, the minor disturbance indicates that the remaining inclusions contain concentrations of Sm and Nd or an isotope signature which did not lead to a significant change of the isotopic systematics of the garnets. Considering these arguments, the indistinguishable ages from the leached fraction as well as from the inclusion-rich fraction ascertain that the ages do not reflect the effect of mixing.

The previously described four garnet whole-rock pairs, if combined, yield an age of 546 ± 9 Ma with an initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.512090 ± 33 (Fig. 6-12), which corresponds to an $\varepsilon_{\text{Nd}(t)}$ of 3.0 ± 0.6 . All these tie-line ages are identical, within error, and considering the Sm-Nd age to be a cooling age, the age of the garnet - whole-rock pairs indicates rapid cooling of the Mavuradonha Layered Complex after the high-grade metamorphic event. This event was dated at 554 ± 13 Ma by SHRIMP on metamorphic rims around older cores.

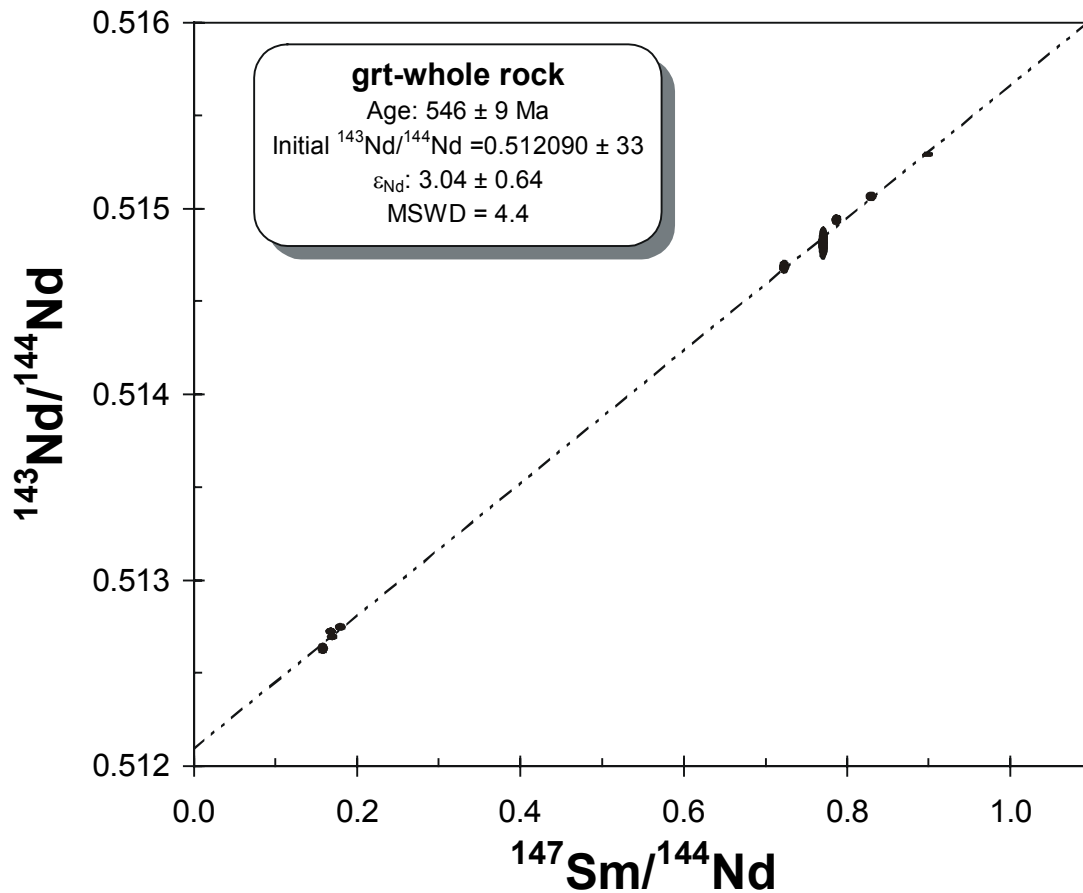


Figure 6-12. Isochron diagram for garnet whole-rock pairs from samples of the upper meta-anorthosite suite of the Mavuradonha Layered Complex.

6.5 Summary

Zircons of inferred magmatic origin were analysed isotopically to determine the age of crystallisation of the Mavuradonha Layered Complex. A concordant SHRIMP age of 862 ± 4 Ma is taken as the age of emplacement of this layered mafic complex. Isotope dilution ages determined on two metagabbroic samples yielded a U-Pb crystallisation age of $\sim 870 \pm 20$ Ma which is identical to the SHRIMP age. These results point to early Neoproterozoic emplacement of the Mavuradonha Layered Complex at around 860 Ma.

The interpretation of the concordant SHRIMP age of 862 ± 4 Ma as the crystallisation age for the Mavuradonha Layered Complex is in contradiction to the proposed minimum age of ca. 1800 Ma (Hargrove et al., 2003) based on U-Pb isotope dilution data for several single zircon analysis. None of the reported ages by Hargrove et al. (2003) is the result of inheritance of zircons and, therefore, their interpretation is contrary to the interpretation of the zircon populations as presented in this study.

Metamorphism dated on overgrowth around older zircons yielded an age of 554 ± 13 Ma for the high-grade metamorphic event during the late Pan African orogeny. A four point garnet whole-rock regression line yielded an age of 546 ± 9 Ma. Considering the T_c of garnet (700 ± 30 °C which is estimated for fast cooling terrains; Mezger et al., 1992) this age probably reflects cooling of the Mavuradonha Layered Complex within a short time after the peak of metamorphism. Subsequent cooling of the complex to ~ 400 °C occurred at around 526 to 501 Ma and is dated by U-Pb ages on rutile (considering the T_c of 400 ± 30 °C for the U-Pb system of rutile).

Judging from the results of thermobarometry and geochronology, the Mavuradonha Layered Complex was metamorphosed during the late Pan-African orogeny at around 554 Ma under HP granulite-facies conditions. Afterwards it cooled within a short time interval to amphibolite-facies conditions and reached lower amphibolite- to greenschist-facies conditions at around 501 Ma. Our result for the age of high-grade metamorphism at around 550 Ma is consistent with the lower intercept U-Pb zircon age for the Mavuradonha Layered Complex in combination with U-Pb titanite age data reported by Hargrove et al. (2003).

Another interesting feature are the zircon xenocrysts in Sample ZZB 123. These xenocrysts of various ages indicate a crustal setting for the emplacement of the Mavuradonha Layered Complex.

An age of 1022 ± 7 Ma is estimated for the Ocellar Gneiss which is interpreted as the time of emplacement of the protolith granitoids. This age is younger than the age of 1.15 Ga which was estimated by Hargrove et al. (1998) but is in line with the reported age of 1051.7 ± 1.3 Ma (Hargrove, pers. comm., 2000; Hargrove et al., 2003) for the Ocellar Gneiss north of the Mavuradonha Mountains.

Syn- or slightly post-metamorphic pegmatites yielded identical late Pan-African crystallisation ages around 545 Ma which are identical to other pegmatite ages in the NE Zambezi belt (Hargrove et al. 1998; Vinyu et al 1999).

7 Sm-Nd, Rb-Sr and Pb-Pb isotopic systematics

Isotopic systematics are a powerful tool to ascertain mantle processes and evolution of oceanic and continental crust (White, 1997). In addition to precise age dating they are useful tracers to identify geological processes, i.e. regional vs. contact metamorphism or igneous evolution. In addition, combining data of different isotopic systems is also of advantage for the determination of the geological setting of mafic and felsic complexes.

The Sm-Nd isotopic system is considered to remain unchanged during metamorphism and alteration because of the similar chemical behaviour and the general immobility of both elements (Faure, 1986). Therefore, this isotopic system is a powerful tool for investigations of initial isotopic compositions in high-grade metamorphic rocks. Additionally, the system can be used for age determinations of mafic rocks, using whole-rock or mineral isochrons, because of sufficiently variable Sm-Nd ratios. However, a large variation in the $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of rocks or minerals is required to yield precise age data.

Another important issue is whether mafic complexes of inferred mantle origin experienced crustal contamination during emplacement. Initial isotope ratios of differentiated rocks may be modified either by mixing between single layers of different origin or by assimilation of crustal material during magma ascent through the continental crust. Contamination or mixing between layers of different origin may result in significant changes in the $^{147}\text{Sm}/^{144}\text{Nd}$ ratio and, hence, the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio. A good example of this is provided by the Pongola Supergroup and the Usushwana Intrusive Suite, Swaziland (Hegner et al. 1984). These authors have shown that the Sm-Nd whole-rock isochron for the Usushwana intrusive suite represents a mixing line of the mafic magma and continental crust with an inferred initial ϵ_{Nd} value that was not representative of the source. Due to contamination with old crustal material the resulting Sm-Nd age was older than the crystallisation age that was determined using U-Pb zircon geochronology. However, processes leading to crustal contamination in the lower crust are sometimes not easy to detect because only small amounts of old and therefore isotopically evolved continental crust may contaminate relatively large amounts of mantle-derived magma. This does not lead to a significant change in the original isotope signature of the mafic magma (Rollinson, 1993).

The behaviour of the Rb-Sr system is controlled by the relative mobility of Rb under most crustal conditions. Therefore, this isotopic system is a sensitive tracer for late-stage metamorphic overprints as well as alteration and crustal contamination of mafic rocks. In contrast to the Sm-Nd system, the Rb-Sr isotope system is not a useful tool for the characterisation of high-grade metamorphic rocks since this system is likely to be disturbed during high-grade metamorphism and/or infiltration of concomitant fluid phases.

Therefore, Rb-Sr ages and initial ratios have to be interpreted with care. In this case, the Rb-Sr isotopic system only provides meaningful information if another system is involved in the interpretation.

In feldspars common Pb is enriched relatively to U. Therefore, the Pb composition of feldspars undergoes no appreciable changes with time and reflects the initial Pb systematics after the last resetting of the system. Chemical leaching of feldspars before analysis ensures that the lead isotope composition of the feldspar is not biased towards an isotope composition of different origin, such as coatings on the mineral surface. Due to the strong fractionation of U from Pb, the isotopic composition of Pb generally indicates a mantle-derived source (low μ -value) or a continental source (high μ -value). Additionally, in favourable circumstances, mixing between the two sources is revealed by the Pb isotopic composition (Stacey & Kramers, 1975). Therefore, the Pb isotopic composition is suitable to provide information about the petrogenesis of mafic rocks associated with crustal lithologies such as in the Mavuradonha Layered Complex.

The following chapter presents the isotopic characteristics of the Mavuradonha Layered Complex, the Nyamhanda Inlier and the Chimwaya Hill Inlier in order to characterise the metagabbroic complexes and to place constraints on their possible tectonic setting. A combination of the Rb-Sr and Sm-Nd isotope data is used to determine the characteristics of the source region from which the magmas of the Mavuradonha Layered Complex, the Nyamhanda Inlier and the Chimwaya Hill Inlier were extracted. Furthermore, an interpretation of the different isotopic systems is used to recognise contamination processes during ascent and emplacement of the magmas.

Seventeen samples from three cross-sections across the Mavuradonha Layered Complex were chosen for the determination of Rb-Sr, Sm-Nd and Pb-Pb isotope ratios. These cross-sections contain different rock types ranging from metapyroxenite to meta-anorthosite and are described in detail in the previous chapters. Ten metagabbro samples of the Nyamhanda Inlier and the Chimwaya Hill Inlier were selected for isotope analysis. The analytical results are listed in Appendix VI, Table VI-1 for the Sm-Nd and Rb-Sr systems and in Table VI-2 for the common Pb isotopic composition. Initial Sr and Nd isotopic ratios were recalculated for a crystallisation age of 860 Ma as determined by U-Pb dating of zircons (Chapter 6).

7.1 Sm-Nd isotopic systematics

7.1.1 Mavuradonha Layered Complex

Most of the analysed samples of the Mavuradonha Layered Complex have low Sm and Nd concentrations ranging from 0.2 to 2 ppm, except for the garnet-bearing samples which have generally higher Sm and Nd concentrations between 0.8 and 14.8 ppm, indicating that these represent more evolved rocks in the complex. Since the rocks of all three sections are cumulates, these differences in the isotopic systematics are strongly dependent on the mineralogy and chemistry of the single cross-section. Plagioclase, clinopyroxene and with ongoing differentiation also Fe-Ti oxides and apatite, played an important role during the accumulation process as major cumulus as well as intercumulus phases. The upper and lower meta-anorthosite suites consist of ferro-metagabbros and meta-anorthosites with varying amounts of mafic, felsic and ore minerals with $^{147}\text{Sm}/^{144}\text{Nd}$ ratios (Fig. 7-1) ranging from 0.138 to 0.206. These fairly constant $^{147}\text{Sm}/^{144}\text{Nd}$ ratios are explained as the result of accumulation of plagioclase, amphibole and clinopyroxene in these suites.

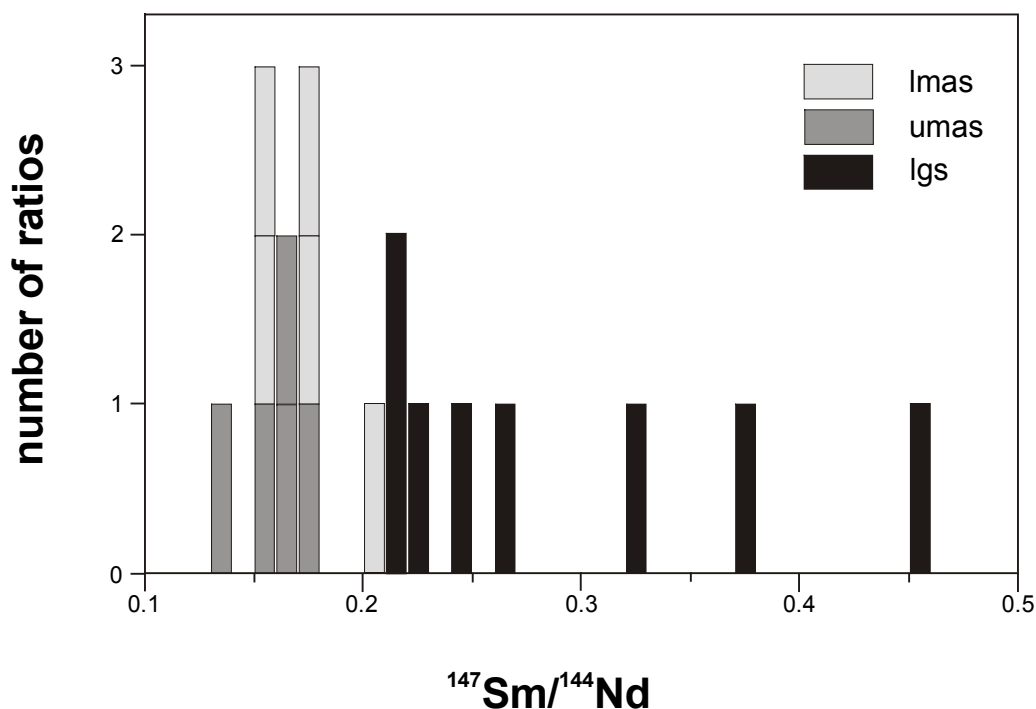


Figure 7-1. $^{147}\text{Sm}/^{144}\text{Nd}$ ratios for all analysed samples of the Mavuradonha Layered Complex. Note the wide range of the samples from the layered gabbro series (lgs), whereas the samples of the upper meta-anorthosite suite (umas) and the lower meta-anorthosite suite (lmas) are consistent.

In contrast, the $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of the metagabbroic rocks of the cross-section in the layered gabbro series vary over a wide range from 0.215 to 0.455. Due to varying differentiation stages, the relatively high $^{147}\text{Sm}/^{144}\text{Nd}$ ratios probably reflect accumulation of early, REE-depleted clinopyroxene and minor amounts of plagioclase as intercumulus phases in metapyroxenites. Metagabbros with similar amounts of plagioclase and clinopyroxene have lower $^{147}\text{Sm}/^{144}\text{Nd}$ ratios. A leuco-metagabbro has the lowest $^{147}\text{Sm}/^{144}\text{Nd}$ ratio in this cross-section which is indicative of plagioclase enrichment in this sample.

The measured $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of ferro-gabbros and meta-anorthosites from the upper and lower meta-anorthosite suite range from 0.512498 ± 11 to 0.512910 ± 12 (Table VI-1; Appendix VI) and are generally lower than in the layered gabbro series. In this series the samples exhibit a wider range in measured $^{143}\text{Nd}/^{144}\text{Nd}$ ratios between 0.512970 ± 8 and 0.513884 ± 13 . The calculated initial $\epsilon_{\text{Nd}}(860 \text{ Ma})$ values for the metagabbroic rocks of the layered gabbro series vary from + 4.4 to + 6.6, whereas the initial $\epsilon_{\text{Nd}}(860 \text{ Ma})$ in the ferro-metagabbros and meta-anorthosites of both meta-anorthosite suites are lower ranging from + 3.3 to + 5.1. However, the initial ϵ_{Nd} values of all investigated samples are slightly lower than the depleted mantle value at 860 Ma ($\epsilon_{\text{Nd}} = + 7.6$, calculation based on the data given in Liew & Hofmann, 1988). Therefore, the initial $\epsilon_{\text{Nd}}(860 \text{ Ma})$ can be interpreted as a source feature of the Mavuradonha Layered Complex, which is most likely a moderately depleted mantle. Since most samples display lower ϵ_{Nd} values than the depleted mantle, contamination processes during magma ascent or during the early stages of differentiation within the Mavuradonha Layered Complex are likely.

In Figure 7-2 (a) the $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of samples from all three cross-sections are correlated over a wide range of isotopic ratios. Taken the different rock types together, the relatively large range in the $^{147}\text{Sm}/^{144}\text{Nd}$ ratios is suitable for age determination. On the $^{147}\text{Sm}/^{144}\text{Nd}$ vs. $^{143}\text{Nd}/^{144}\text{Nd}$ diagram the samples can be fitted to a regression line (MWSD = 11) that defines an age of $899 \pm 46 \text{ Ma}$. This is similar, within error, to the previously described zircon age of $862 \pm 4 \text{ Ma}$. From this it can be seen that the Sm-Nd system was not significantly disturbed during the high-grade metamorphic overprint.

The linear array of the data points could also be interpreted as a two-component mixing line. In this case, a straight line should result in a diagram $^{143}\text{Nd}/^{144}\text{Nd}_{(i)}$ vs. $1/\text{Nd}$ (Fig. 7-3). However, there is no linear relationship between the $^{143}\text{Nd}/^{144}\text{Nd}$ initial ratio and the Nd concentration. Therefore, a two component mixing model is not applicable to the Sm-Nd isotopic systematics. This implies that mixing between layers or mixing of different layers is not very likely. In addition, the large spatial distance between the three cross-sections is a further argument against mixing between single layers.

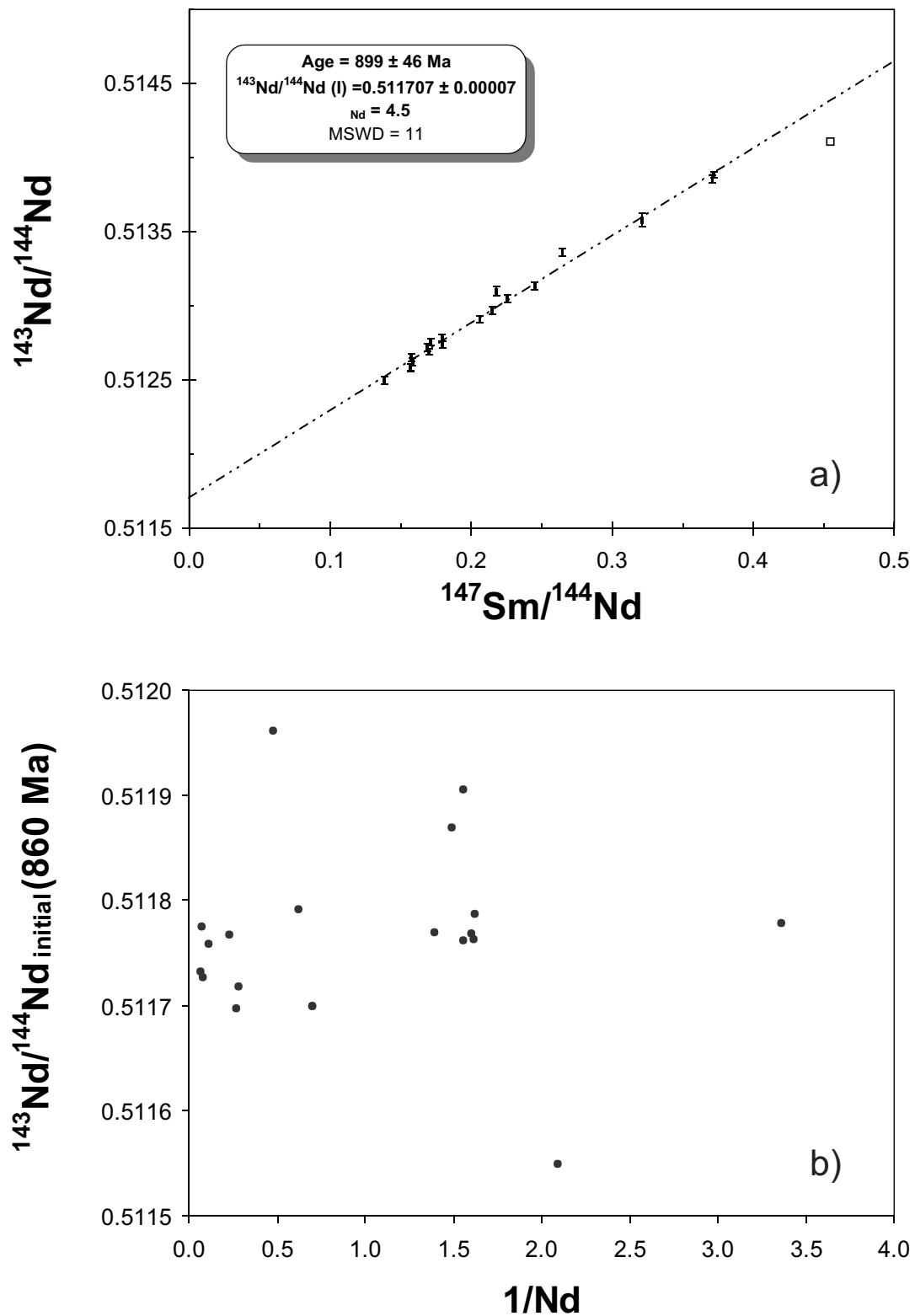


Figure 7-2. a) Whole-rock correlation diagram for the Mavuradonha Layered Complex. The square marks a sample which was excluded from discussion because of its anomalous isotopic data.

b) $^{143}\text{Nd}/^{144}\text{Nd}_{(860\text{ Ma})}$ vs. $1/\text{Nd}$ diagram for samples of the Mavuradonha Layered Complex.

As an alternative interpretation the large scatter in the arrangement of the data points probably points to heterogeneity of the isotopic systematics in different parts of the intrusion at the time of emplacement. Initial isotopic variations have been shown to vary with stratigraphic heights in many other layered intrusions (e.g. dePaolo, 1985; Eales et al., 1986; Gray & Goode, 1989; Oberthür et al., 2002).

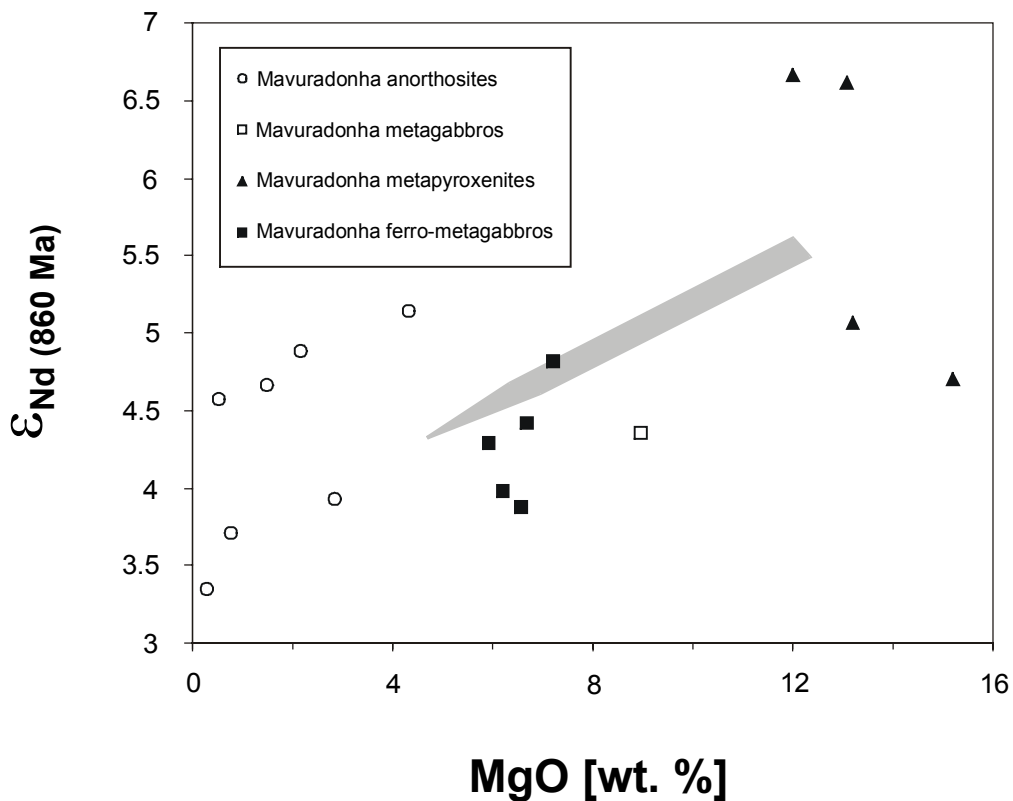


Figure 7-3. Progressively decreasing ϵ_{Nd} values in the diagram $\epsilon_{\text{Nd}} (860 \text{ Ma})$ vs. MgO for samples of the Mavuradonha Layered Complex. The grey arrow indicates the increase in contamination with decreasing MgO concentrations.

In order to substantiate the interpretations of the results $\epsilon_{\text{Nd}} (860 \text{ Ma})$ are plotted vs. MgO [wt.%] which provides further information about possible contamination of the analysed rocks with crustal material. In figure 7-3 the $\epsilon_{\text{Nd}} (860 \text{ Ma})$ values progressively decrease with decreasing MgO [wt.%] which indicates an increase in contamination. This diagram also implies that the ferro-metagabbros and meta-anorthosites underwent higher degrees of contamination than the other rocks. Additionally, crustal contamination is also supported by abundant xenoliths in the layered gabbro series as well as the previously described inherited zircons in metagabbro sample ZZB 123. Therefore, the large scatter (Fig. 7-2a) in the arrangement of the data points probably reflects contamination of the Mavuradonha Layered Complex rocks by lower-crustal rocks which led to changes in the isotopic systematics. Therefore, it is more likely that the errorchron reflects contamination

of the mafic magma by crustal rocks rather than an isotope heterogeneity in different parts of the complex.

7.1.2 Nyamhanda Inlier and Chimwaya Hill Inlier

The Nyamhanda and the Chimwaya Hill Inlier samples show a large range in Sm and Nd concentrations as well as in their isotopic ratios. Two metagabbroic complexes have been mapped (Fig. 2-2) and can also be distinguished in the Nyamhanda Inlier on the basis of their Sm-Nd isotopic composition (Fig. 7-4). The first group consists of a small metagabbro in the NE and a large complex in the W of the inlier. The initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios vary between 0.511601 ± 12 and 0.51164 ± 7 .

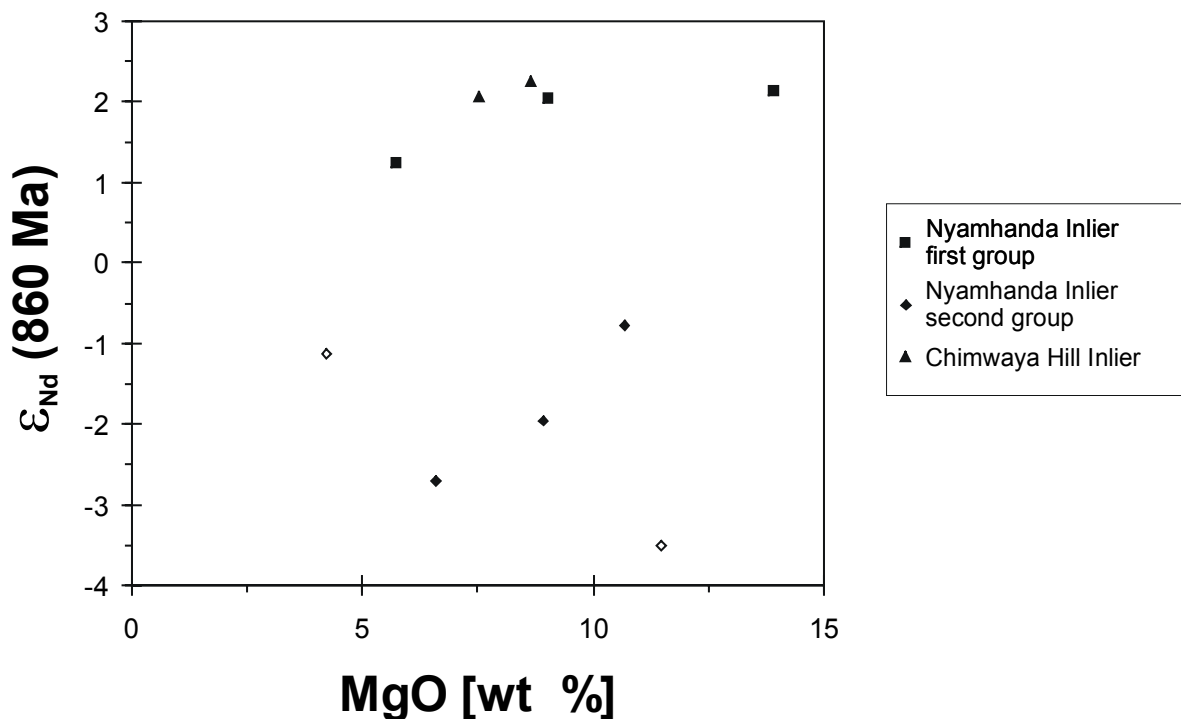


Figure 7-4. Fairly constant ϵ_{Nd} values in the diagram $\epsilon_{\text{Nd}} (860 \text{ Ma})$ vs. MgO for samples of first and second group of the Nyamhanda Inlier and the Chimwaya Hill Inlier. An open diamond marks samples of the second Nyamhanda Inlier group that were not considered for calculation of the isochron.

The $\epsilon_{\text{Nd}(t)}$ for these two metagabbros are + 1.2 and + 2.1. In a diagram of MgO vs. $\epsilon_{\text{Nd}(t)}$ (Fig. 7-4) the metagabbros provide evidence for increasing contamination by decreasing ϵ_{Nd} at decreasing MgO concentrations. Therefore, the metagabbros of this group are assumed to be derived from a moderately depleted upper mantle without or only minor crustal contamination.

The second group consists of the central metagabbroic complex and a small metagabbro in the SE. Four metagabbros and a ferro-metagabbro have initial $^{143}\text{Nd}/^{144}\text{Nd}$ values between 0.511361 ± 13 and 0.511479 ± 16 . The $\varepsilon_{\text{Nd}(t)}$ values vary between -0.7 and -3.5. These values are somewhat lower than the CHUR value, and probably reflect crustal contamination. In the diagram MgO vs. $\varepsilon_{\text{Nd}(t)}$ (Fig. 7-4) they provide no evidence of increasing contamination at decreasing MgO concentrations as a whole. Considering relations that were observed during field work, three samples (DKZ 16, DKZ 18 and DKZ 25) show a correlation of decreasing $\varepsilon_{\text{Nd}(t)}$ with decreasing MgO concentrations. According to the field observations the other two metagabbros are interpreted as single gabbro bodies.

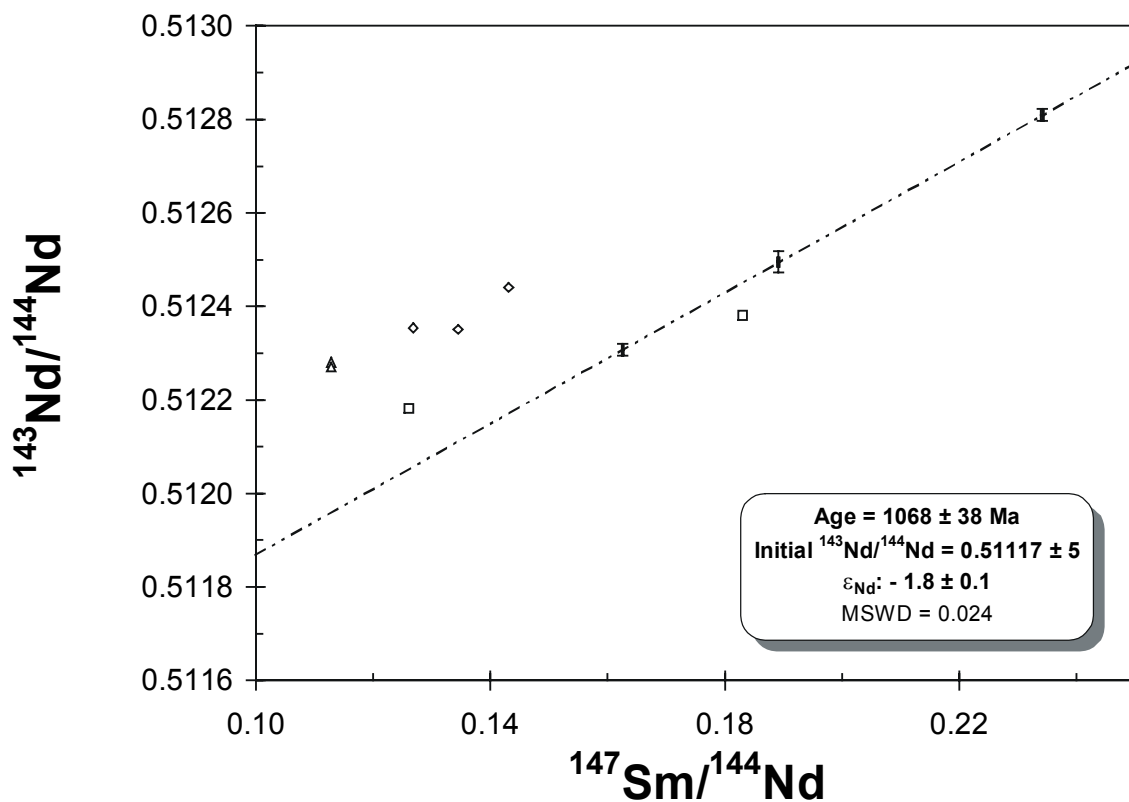


Figure 7-5. Diagram of $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{147}\text{Sm}/^{144}\text{Nd}$ for whole-rock samples from the Nyamhanda Inlier and the Chimwaya Hill Inlier; Nyamhanda Inlier samples of the first group are marked by diamonds, the second group by squares and the Chimwaya Hill Inlier samples by triangles. Errors on the $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ ratios are smaller than the symbol size.

In Figure 7-5 the two groups from the Nyamhanda Inlier are plotted separately, and they define two different linear trends. The samples of the first group cannot be used for age determination, because only a few analyses are available that show little isotopic variation. Three samples of the second group show a larger range in $^{147}\text{Sm}/^{144}\text{Nd}$ isotope

composition than the first group which allows the calculation of a best-fit line. The resulting age of 1068 ± 38 Ma is only of minor significance because of the very small MSWD of 0.024 which reflects an overassessment of uncertainties during measurement (Fig. 7-5). If this age reflects the emplacement age of these metagabbros it has to be interpreted carefully because of the previously described contamination with crustal material and further age investigations using accessory minerals like zircons have to be carried out in order to substantiate this age. In a diagram $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $1/\text{Nd}$ (not shown in diagram) no linear relationship is visible, which precludes any mixing relationship.

Both analysed metagabbro samples of the Chimwaya Hill Inlier have identical Nd isotopic composition (Table VI-1). These rocks have initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of 0.511651 ± 7 and 0.511641 ± 7 and $\varepsilon_{\text{Nd}(860 \text{ Ma})}$ values of 1.9 and 2.1. They compare well with the first group of the Nyamhanda Inlier (Fig. 7-5), although they exhibit the same mineralogical textures as the metagabbros of the Mavuradonha Layered Complex.

7.2 Rb-Sr isotopic system

7.2.1 Mavuradonha Layered Complex

Low Rb and variable Sr concentrations characterise the samples of the Mavuradonha Layered Complex. Since plagioclase played an important role during accumulation of this complex, the meta-anorthosites contain the highest Sr concentrations but the lowest $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ ratios. In contrast, a metapyroxenite, the most mafic rock type analysed during this study, contains the highest $^{87}\text{Rb}/^{86}\text{Sr}$ ratio of 0.46 and the lowest Sr concentration. The analytical results are reported in Table VI-1.

The $^{87}\text{Rb}/^{86}\text{Sr}$ ratios for the Mavuradonha Layered Complex samples range from 0.007 to 0.46, the measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vary from 0.702533 ± 10 to 0.709145 ± 12 . Ferro-metagabbros and meta-anorthosites of the upper meta-anorthosite suite show the most consistent and lowest $^{87}\text{Rb}/^{86}\text{Sr}$ (0.007 to 0.041) and measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.702533 ± 10 to 0.704201 ± 9). In contrast, some ferro-metagabbros, meta-anorthosites of the lower meta-anorthosite suite and some of the metagabbroic rocks of the layered gabbro series display significant disturbance of the Rb/Sr system. This probably reflects the inferred contamination by crustal material although some of the samples seem to show also an effect of infiltration of crustally-derived fluids.

In Figure 7-6 the samples show a considerable scatter, but linear relationships between single samples are also visible. Three groups of samples can be used for calculating regression lines (Fig. 7-6), but the calculated ages vary significantly between 1014 ± 95 Ma and 5675 ± 1700 Ma. The resulting ages, when best-fit lines are calculated,

seem to have no geological significance taking into account the large errors of the ages, the MSWD values and especially if the emplacement age of 862 Ma is considered. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $1/\text{Sr}$ (not shown) illustrates no direct relationship between the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio and the Sr concentration. This indicates no binary mixing between the samples and fluids of an unknown composition. Therefore, the scatter is probably the result of the inferred disturbance of the system due to crustal contamination, and in several cases, also to alteration. Since the ferro-metagabbros and meta-anorthosites of the upper meta-anorthosite suite do not appear significantly altered in thin section the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios may be used to some extent in addition to other systems to determine the source region.

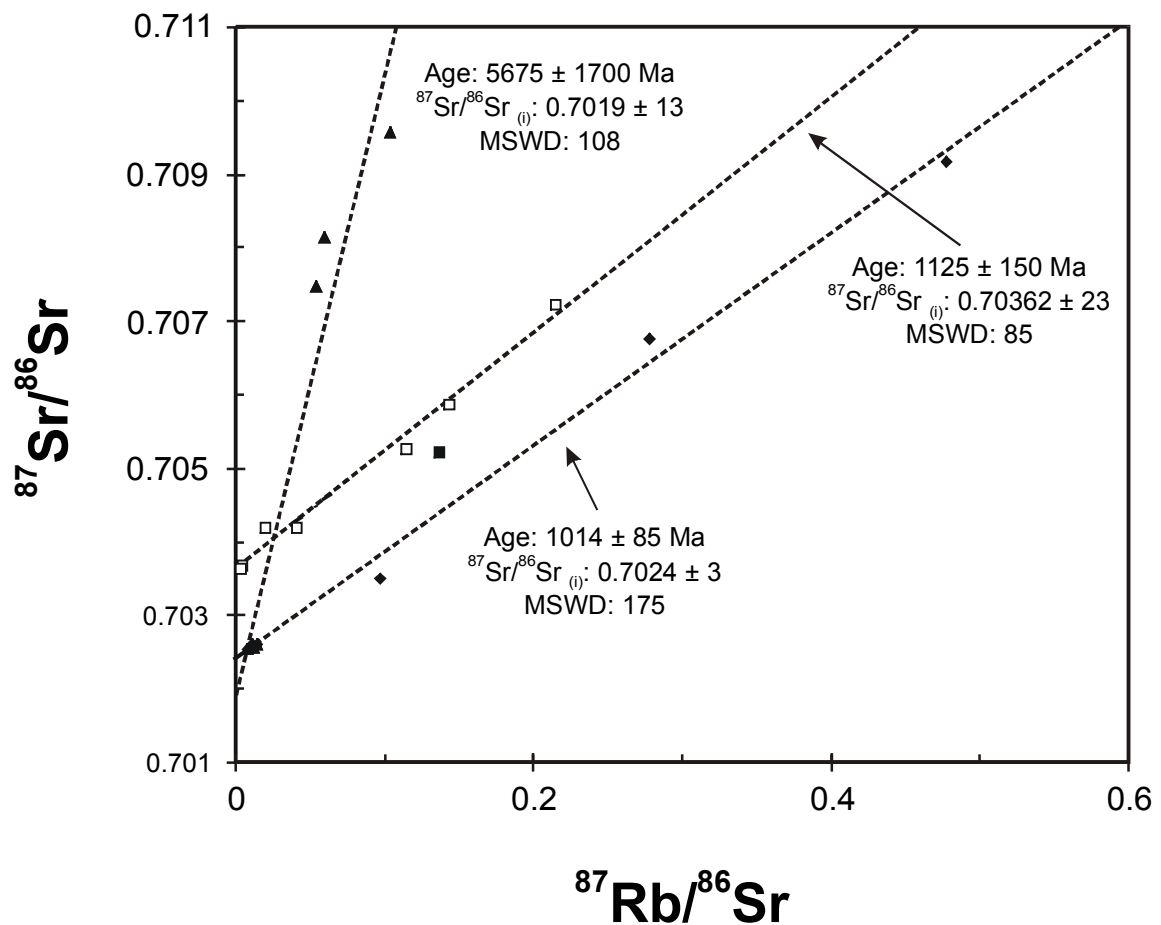


Figure 7-6. $^{87}\text{Rb}/^{86}\text{Sr}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ correlation diagram for samples of the Mavuradonha Layered Complex. Note the wide range of isotopic ratios in the samples. One sample (filled square) was not considered during calculation of the best-fit line for this sample group.

7.2.2 Nyamhanda Inlier and Chimwaya Hill Inlier

The Rb-Sr isotopic ratios and element concentrations from the Nyamhanda Inlier show a large range. The samples have low $^{87}\text{Rb}/^{86}\text{Sr}$ ratios between 0.01 and 0.231 and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios range from 0.704801 ± 5 to 0.712496 ± 11 . The isotopic composition of the Nyamhanda Inlier metagabbroic rocks is reported in Tab. VI-1.

Two types of metagabbroic rocks (Fig. 2-2) can be distinguished in the Nyamhanda Inlier on the basis of their Rb-Sr isotopic composition: The first group comprises the metagabbros in the NE and NW with low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.704801 ± 5 to 0.705122 ± 13 (Fig. 7-7), whereas the second group, consisting of the metagabbros in the centre and the SE of the inlier, has higher $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.707652 ± 13 to 0.712496 ± 11 .

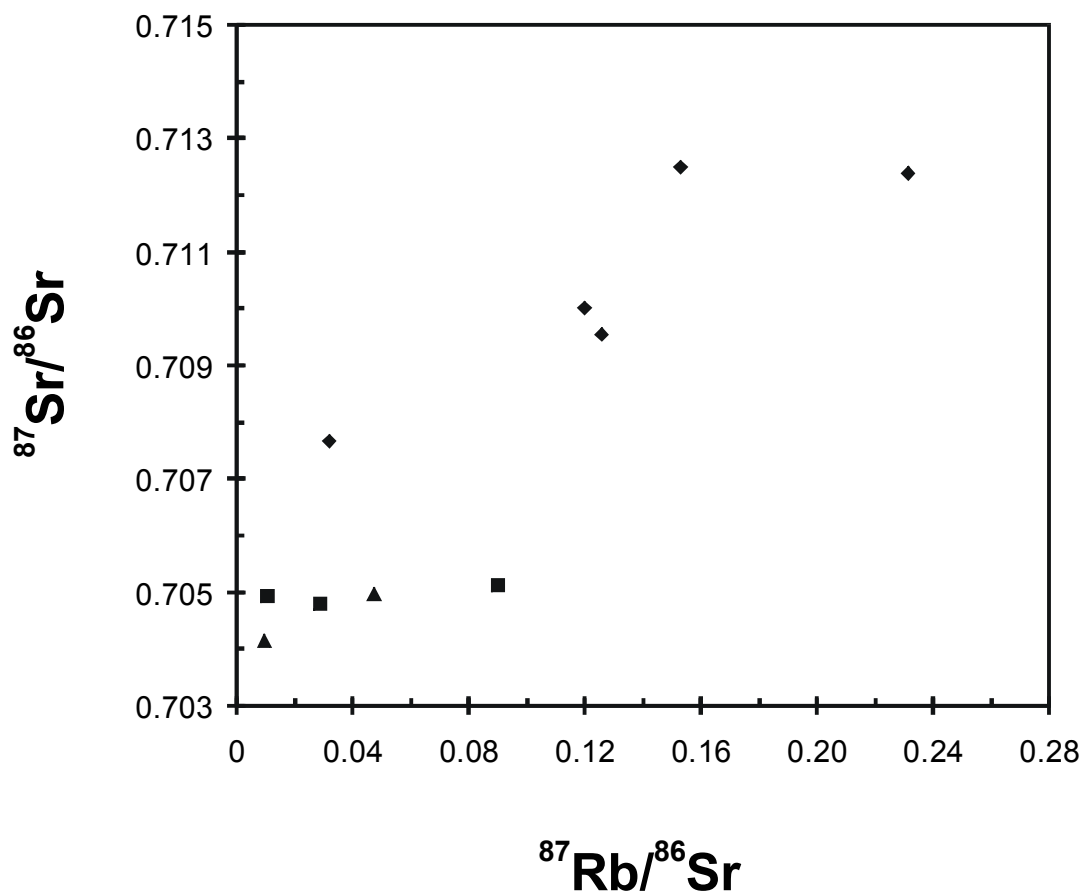


Figure 7-7. $^{87}\text{Rb}/^{86}\text{Sr}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ correlation diagram for samples of the Nyamhanda Inlier and the Chimwaya Hill Inlier. Nyamhanda Inlier samples of the first group are marked by squares, the second group by diamonds and the Chimwaya Hill Inlier samples by triangles. Note that errors on the $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are smaller than the symbol size.

Considering contamination with crustally-derived material as shown for the Sm-Nd isotope system in combination with diagrams of $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $\text{FeO}_2/\text{Fe}_2\text{O}_3$ and $^{87}\text{Sr}/^{86}\text{Sr}$ vs. MgO

(both not shown in diagram) it is likely that this disturbance rather reflects contamination by crustal material than major effects of alteration and metamorphism on the Rb-Sr isotope system. The diagrams of $\text{FeO}_2/\text{Fe}_2\text{O}_3$ and MgO vs. $^{87}\text{Sr}/^{86}\text{Sr}$ show some relations for single samples with increasing $^{87}\text{Sr}/^{86}\text{Sr}$ ratios at decreasing MgO content.

Because of the observed uralitisation of pyroxene and corona-textures around pyroxene and opaques which are the result of fluid infiltration and retrogression, an influence of infiltration of fluid phases and/or a metamorphic overprint cannot be ruled out. However, this influence was only of a minor significance, because these mineralogical textures are observed in all samples investigated during this study. Judging from the results of the Rb-Sr isotope investigations, contamination occurred to a minor extent in metagabbros of the first group if the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are compared with the ones of the second group in which higher radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ ratios indicate higher degrees of contamination.

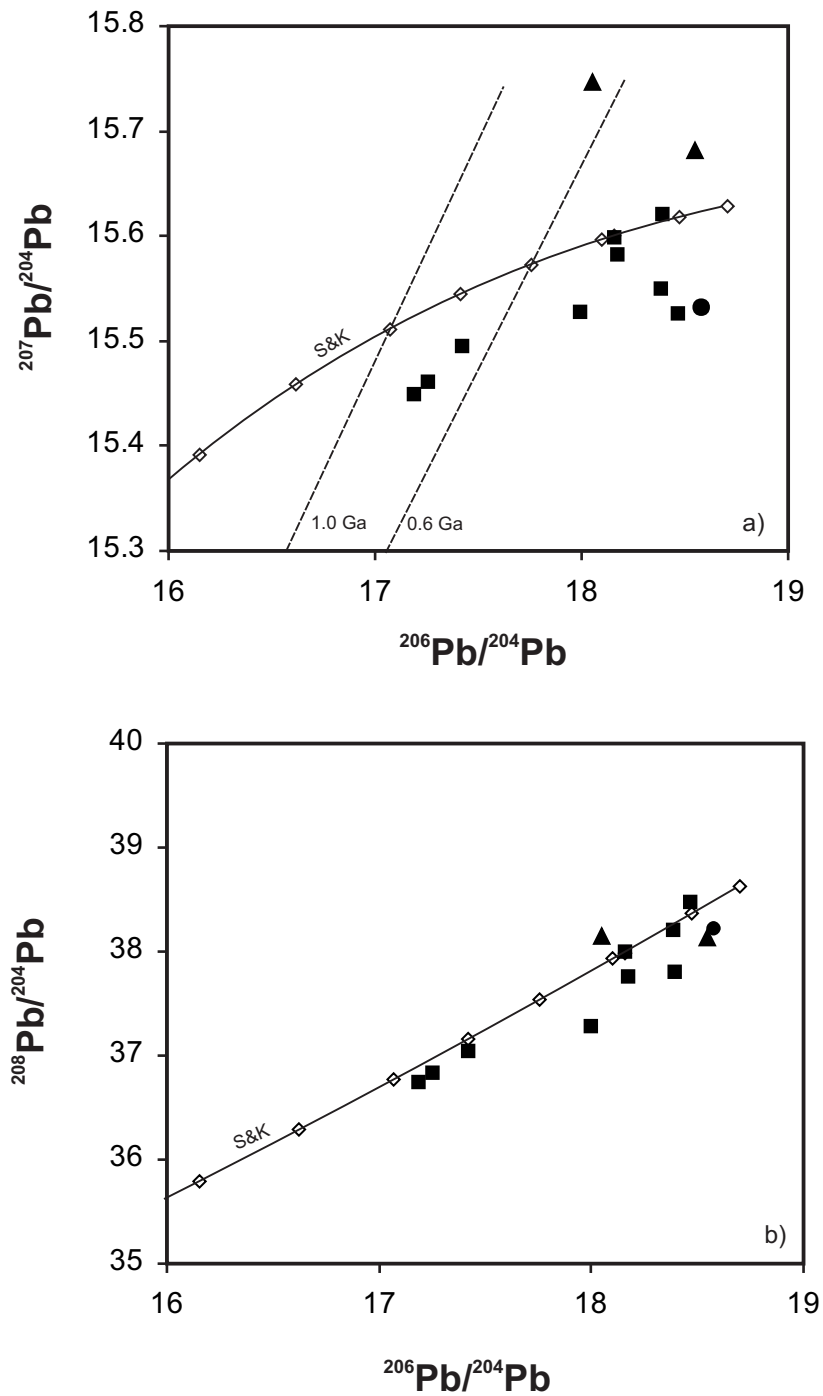
The Rb-Sr isotopic systematics (Table VI-2) of the Chimwaya Hill Inlier samples show similarities to the first group of the Nyamhanda Inlier with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.704973 ± 11 and 0.704138 ± 10 . Based on the isotopic data presented here, crustal contamination can also be assumed for the Chimwaya Hill Inlier metagabbros. The minor shift to higher ratios than the Bulk Earth, which was observed for one of the samples, is probably the result of high-grade metamorphism or infiltration of fluids. This effect is shown by garnet coronas and the nearly complete replacement of plagioclase by scapolite.

7.3 Pb isotopic systematics of feldspars

7.3.1 Mavuradonha Layered Complex

Plagioclase was separated from nine samples in order to determine their lead isotopic compositions. Two samples (ZZB 166 and ZZB 123) were also chosen for common Pb corrections of zircons, a third sample for Pb corrections of rutile, both accessories were dated during this study. The isotopic composition of leached plagioclases is reported in Table VI-2.

In a $^{207}\text{Pb}/^{206}\text{Pb}$ diagram, the common Pb isotopic composition suggests an evolution with a μ -value similar or slightly lower than the average continental crust as proposed by Stacey & Kramers (1975; S&K), because most of the analysed samples plot on or below the Pb evolution curve. These low μ -values are an indication for their reservoirs, which may be the average crust or the mantle.



- Mavuradonha Layered Complex
- ▲ Nyamhanda Inlier, 2nd group
- Chimwaya Hill Inlier

Figure 7-8. Lead isotopic composition of leached plagioclases from the Mavuradonha Layered Complex, the Nyamhanda Inlier and the Chimwaya Hill Inlier. The curve represents the lead evolution of the two stage model of Stacey & Kramers (1975; S&K) with $\alpha=9.74$. "Palaeogeochrons" are shown for 1.0 and 0.6 Ga in the $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagram.

The Pb data allow to distinguish between two groups of the layered rocks from the Mavuradonha Layered Complex (Fig. 7-8). The first group of three samples is composed of meta-anorthosites of the upper meta-anorthosite suite next to the Zambezi Escarpment. These samples exhibit the least evolved common Pb compositions on the $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams. They plot below the S&K evolution curve and to the right of the 1 Ga geochron, suggesting that plagioclase in these samples preserved its isotopic composition and was not reset during granulite-facies metamorphism. The second group consists of samples, which plot to the right of the 600 Ma geochron and slightly above or significantly below the S&K curve in Figure 7-8 a. This group exhibits a large spread, and it is possible that these samples record the last recrystallisation and resetting event of plagioclase during the late Pan-African high-grade metamorphic overprint around 550 Ma. In Fig. 7-8 b the samples show the same feature, i.e. the samples plot to the left and right of the U/Pb evolution curve.

Applying the model of Doe & Zartman (1979) and Zartman & Doe (1981) the Pb isotope composition indicates derivation of the magma from a mantle source. This is because the data points plot in $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams between the Pb-evolution curve for the mantle and the upper crust (Fig. 7-9 a and b) which can be taken as a clear indication for contamination by upper crustal material. Nevertheless, this underlines the existence of a depleted mantle source beneath North-eastern Zimbabwe as derived from Nd-isotopes. Additionally, contamination of the extracted parental magma either during ascent into crustal levels or during fractionation as proposed by Nd isotope evolution is outlined. Some of the samples of the Mavuradonha Layered Complex provide additional information for resetting during an orogeny because they plot on, or slightly above, the Pb-evolution curve for orogens (Fig. 7-9 a and b).

7.3.2 Nyamhanda Inlier and Chimwaya Hill Inlier

Two samples of the previously mentioned second group of the Nyamhanda Inlier were investigated for Pb-Pb isotopic composition. The common Pb composition of these samples shows an evolution of the metagabbros with a higher μ -value than 9.74, because both samples plot above the S&K Pb evolution curve (Fig 7-8 a). Both samples exhibit a similar Th/Pb evolution but show disturbance of the lead isotopic system because they plot at similar $^{208}\text{Pb}/^{204}\text{Pb}$ but different $^{206}\text{Pb}/^{204}\text{Pb}$ on different sides of the Pb evolution curve (Fig. 7-8 b).

The Pb isotopic composition of these two samples provides indications for contamination by upper crustal material (Doe & Zartman, 1979; Zartman & Doe, 1981),

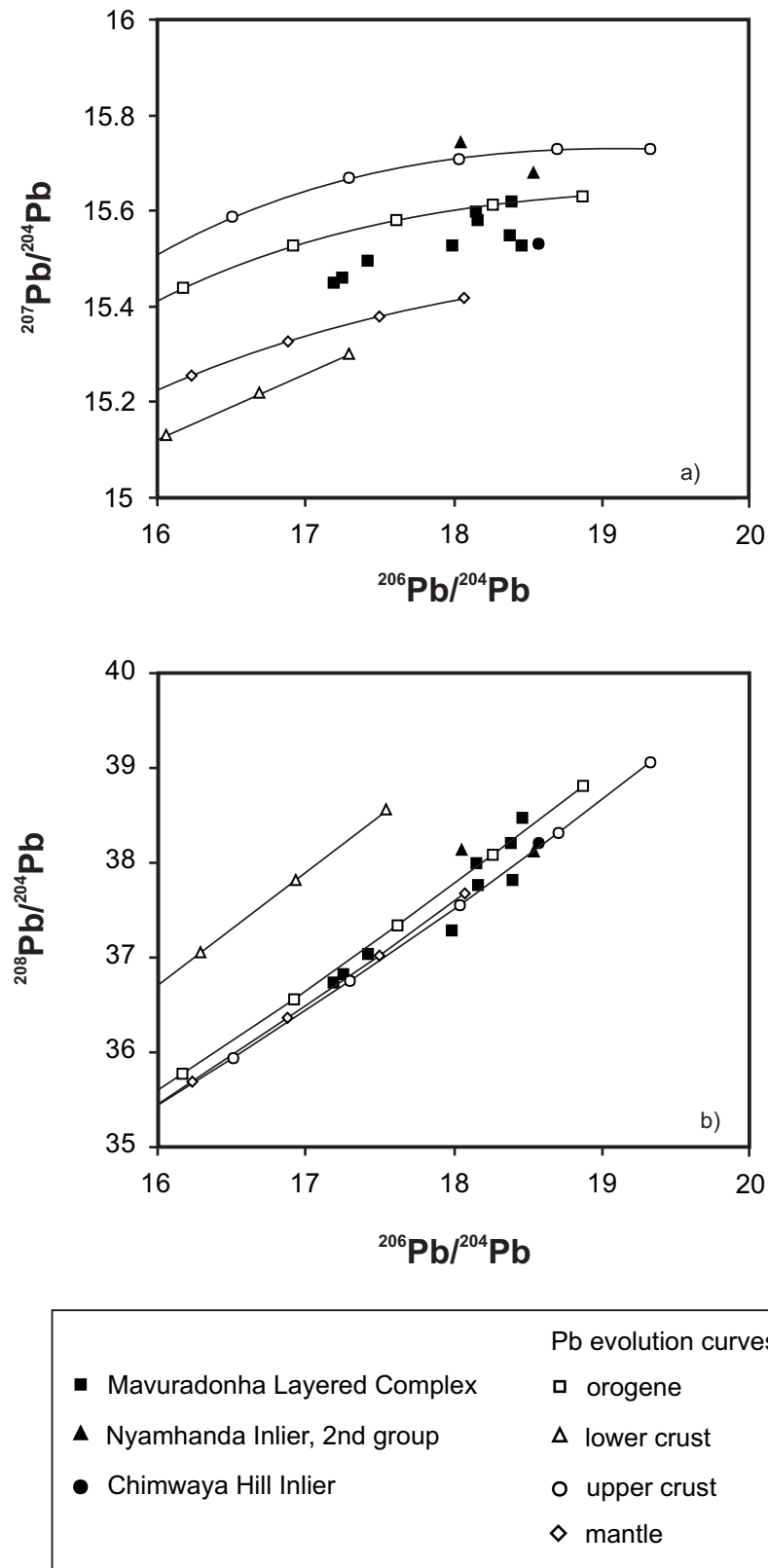


Figure 7-9. Lead isotopic evolution of leached plagioclases from the Mavuradonha Layered Complex, the Nyamhanda Inlier and the Chimwaya Hill Inlier with given evolution curves for lead as proposed by Zartman & Doe (1981).

because in $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams the data points plot on or above the Pb-evolution curve of the upper crust (Fig. 7-9 a and b).

The Pb evolution of the Chimwaya Hill Inlier samples is similar to those of the Mavuradonha Layered Complex. They evolved from a source with low μ -value and plot below the S&K evolution curve (Fig. 7-8).

Again, when the model of Doe & Zartman (1979) and Zartman & Doe (1981) is applied, the Pb isotopic composition of the Chimwaya Hill Inlier sample indicated a similar evolution as observed for the Mavuradonha Layered Complex samples. However, derivation of the magma from a mantle source with a clear indication for contamination by upper crustal material, due to the plot in $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams between the Pb-evolution curve for the mantle and the upper crust (Fig. 7-9 a and b) can be proposed for the Chimwaya Hill Inlier.

7.4 Summary and geotectonic implications

7.4.1 Summary

The isotopic compositions portray a polyphase evolution of the studied metagabbros in the Mt. Darwin area. The metagabbroic rocks of the Mavuradonha Layered Complex show a wide range in Sm-Nd isotopic ratios that is suitable for age determinations in order to obtain additional information about the emplacement age. An errorchron corresponds to an age of 899 ± 46 Ma (MSWD: 11), which is consistent with an emplacement age of 862 ± 4 Ma determined using concordant SHRIMP zircon data points. On the other hand, the Sm-Nd isotopic system reflects clear evidence for crustal contamination, which is also supported by inherited zircons in the metagabbro samples. Rb-Sr isotope investigations substantiate the assumption of contamination by crustal material; additional results were the effects of fluid phases and metamorphism that are observed in single samples. Taking into account the progressive decrease of initial ϵ_{Nd} values with increasing differentiation as well as the primitive character of some metapyroxenites, only relatively minor contamination with crustal material can be suggested for the layered gabbro series.

Considering the isotopic composition of the Mavuradonha Layered Complex rocks, the meta-anorthosites and ferro-metagabbros of the meta-anorthosite suites show the most consistent, low $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ and high $^{143}\text{Nd}/^{144}\text{Nd}_{(i)}$ ratios. Furthermore, resetting during the high-grade metamorphic overprint or any noteworthy effect of fluid phases on these rocks is unlikely. The resulting low $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ ratios are still relatively primitive in respect to the $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic composition as reported by Barton et al. (1991) for other Pan-African

gabbroic rocks (0.70688 ± 38 to 0.70663 ± 1059) or Archaean to Kibaran continental crust in NE Zimbabwe (0.70873 ± 187 to 0.71479 ± 424). The Pb isotopic composition of the Mavuradonha Layered Complex samples is an indication for the source region, which was a moderately depleted mantle. Additional to the unaltered Pb isotopic composition of the samples from the upper meta-anorthosite suite, some samples indicate an influence of metamorphism, because the Pb isotopic composition seems to be reset during an event at around 550 Ma. This age was determined in a previous chapter to be the age of the high-grade metamorphism. Nevertheless, the Pb isotopic compositions underline the extraction from a depleted mantle source, as proposed for the Nd isotope composition, and indicate variable amounts of contamination by crustal material and the effects of orogenic activity during the final evolution of the Zambezi belt.

The samples of the Nyamhanda Inlier and Chimwaya Hill Inlier have a different isotopic composition from those of the Mavuradonha Layered Complex. Two isotopically distinct groups of the Nyamhanda Inlier samples show a significant isotopic heterogeneity in this inlier. The results of isotope investigations also indicate no relationship for the metagabbros of the Nyamhanda Inlier and revise previous interpretations of the inlier to represent one complex (Leitner & Phaup, 1974, Bache et al., 1984). Whereas the northern and the western part of the Nyamhanda Inlier seem to be derived from a moderately depleted mantle source with a minor disturbance of the isotopic systematics due to crustal contamination, the metagabbros in the centre and SE display a major disturbance of their isotopic composition in all three isotopic systems. The best indication for contamination processes with other crustal rocks is provided by the Sm-Nd isotope system, because the Nyamhanda Inlier metagabbros of the second group yielded negative ϵ_{Nd} values. This contamination is underlined by the Pb isotopic composition of two samples of this group as shown in Figs. 7-8 and 7-9.

The isotopic data for the Chimwaya Hill Inlier indicate a genetic relation to the first group of metagabbros from the Nyamhanda Inlier. This is due to similar Sm-Nd and Rb-Sr isotope compositions of the metagabbros. Based on the isotopic systematics, contamination by crustal material of a magma that was extracted from a moderately depleted mantle can also be assumed for the Chimwaya Hill Inlier metagabbros. Pb isotopic composition corroborates the results that were derived from Nd/Sr isotopic investigations.

7.4.2 Geotectonic implications

The $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ and $^{143}\text{Nd}/^{144}\text{Nd}_{(i)}$ ratios for the metagabbroic rocks of the layered gabbro series and the meta-anorthosite suites of the Mavuradonha Layered Complex could be interpreted as representative for a moderately depleted mantle without any significant crustal contamination. Meta-anorthosites, metagabbros and metapyroxenites of the

Mavuradonha Layered show a wide variation in the Sr/Nd correlation diagram (Fig. 7-10), which cannot be explained by heterogeneity of the mantle source. Crustal contamination of Mavuradonha Layered Complex metagabbroic rocks is displayed by the isotopic systematics in terms of scattering ϵ_{Nd} ratios in the range of 3.3 to 6.7, but also in their $^{87}\text{Sr}/^{86}\text{Sr}$ ratios.

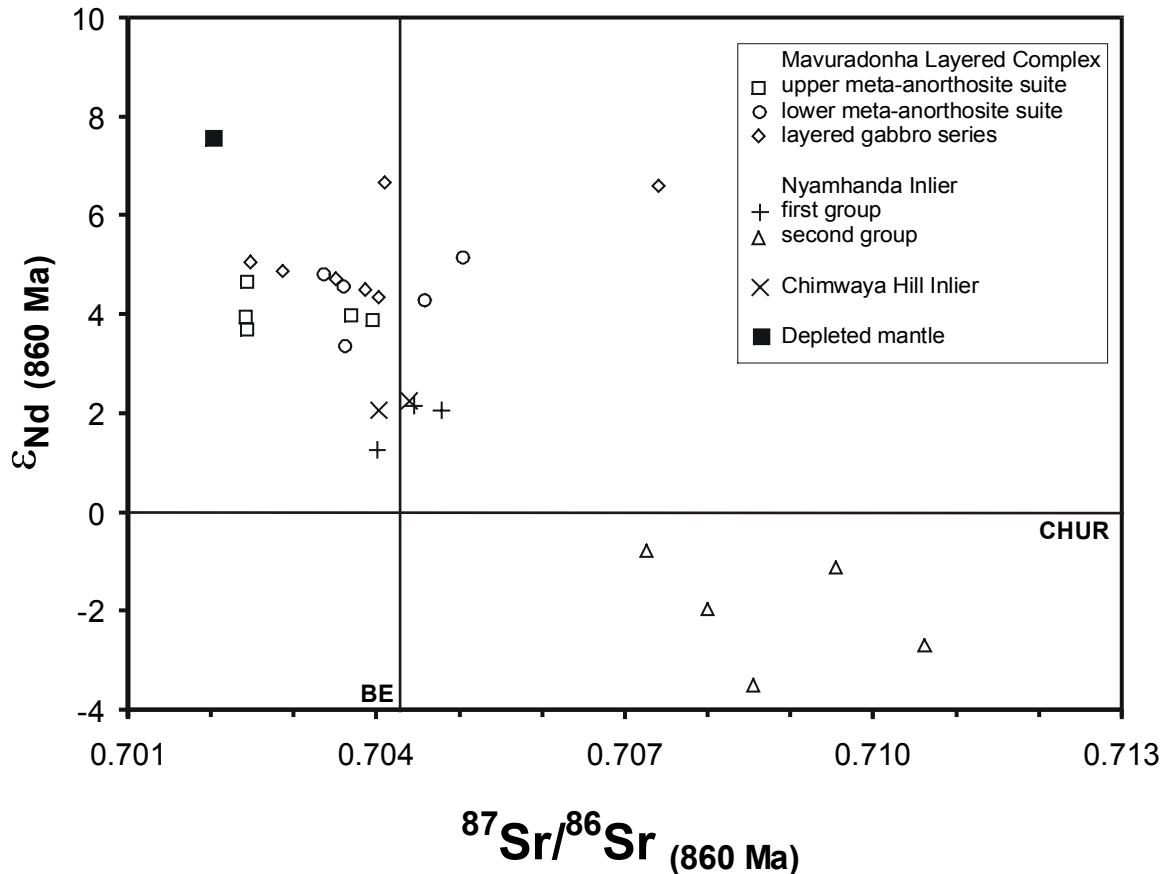


Figure 7-10. ϵ_{Nd} vs. $^{87}\text{Sr}/^{86}\text{Sr}$ diagram (recalculated for 860 Ma) for whole-rock samples of the Mavuradonha Layered Complex, the Nyamhanda Inlier and the Chimwaya Hill Inlier showing variable mixing with older continental crust. Note homogeneity of meta-anorthosite suite samples and similarity of the first group of the Nyamhanda Inlier and samples of the Chimwaya Hill Inlier. A wide compositional range is shown for the second group of the Nyamhanda Inlier and the layered gabbro series of the Mavuradonha Layered Complex. BE: Bulk Earth; CHUR: Chondritic uniform reservoir.

In general, rocks with two distinct features can be observed in Fig. 7-10 for the Mavuradonha Layered Complex:

- (1) The arrangement of the data points of the layered gabbro series to ferro-metagabbros and meta-anorthosites of the upper meta-anorthosite suite point to an increase of crustal contamination of a depleted mantle-derived magma with

crustal material with different isotopic compositions. Due to the decreasing ϵ_{Nd} values with increasing differentiation within the complex (metapyroxenites of the layered gabbro series have the most primitive ϵ_{Nd} values, ferro-metagabbros and meta-anorthosites of the upper meta-anorthosite suite the lowest ϵ_{Nd} values in the complex) an increase in contamination with stratigraphic height can be assumed in the Mavuradonha Layered Complex. Most of the Mavuradonha Layered Complex rocks have fairly constant $^{87}\text{Sr}/^{86}\text{Sr}$ ratios which are still relatively primitive with respect to the Bulk Earth.

- (2) The relative translation of some data point along the $^{87}\text{Sr}/^{86}\text{Sr}$ axis is probably the result of alteration caused by fluid rock interactions. Since fluids are significantly enriched in Sr but contain negligible amounts of Nd (Jacobsen & Wasserburg, 1979), the isotope chemistry of the rock is modified towards higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios.

The parental magma of the inlier metagabbros may be interpreted to be extracted from a moderately depleted mantle source because of the still relatively primitive isotopic composition of the metagabbros. However, higher degrees of crustal contamination than observed for the Mavuradonha Layered Complex metagabbroic rocks are displayed in Fig. 7-10 for the inlier samples. The scatter of the samples of the second group of the Nyamhanda Inlier along both $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ axes reflects higher degrees of contamination with crustal material than observed for the first group of the Nyamhanda Inlier. Contrary to this, the first group of the Nyamhanda Inlier and Chimwaya Hill Inlier samples display similar amounts of contamination. Again, fluid-rock interaction that caused a shift of the $^{87}\text{Sr}/^{86}\text{Sr}$ towards more radiogenic values must be considered. This interaction is also documented by petrographic observations.

Taking these considerations into account, emplacement of the Mavuradonha Layered Complex and the inlier metagabbros into a crustal setting can be proposed. Crustal contamination played a significant role during crystallisation of the magmas and, for single samples, fluid-rock interaction during retrogression following the high-grade metamorphic event has also been observed. Due to the lack of isotopic data for possible contaminants and the lack of contaminants with an age of 1.8 Ga in close relation to the metagabbros in the Mt. Darwin area, calculation of mixing lines by simple two component mixing or AFC calculations (DePaolo, 1985) were not carried out.

8 Petrogenetic and geotectonic implications

8.1 Mavuradonha Layered Complex

Based on the results and interpretations presented in this study the Mavuradonha Layered Complex can be considered as a large layered intrusion that was emplaced in a crustal setting around 860 Ma ago. Judging from petrography and geochemistry, the Mavuradonha Layered Complex rocks can be interpreted as comagmatic cumulates that contain variable amounts of clinopyroxene and plagioclase as major cumulus minerals. With ongoing differentiation apatite and Fe-Ti oxides were accumulated in ferro-gabbros and anorthosites as additional cumulus phases. Accumulation of these mineral phases is inferred, based on evidence from thin sections, and could be verified by the major and trace element variations (Fig. 5-3). Olivine-bearing ultramafic rocks are not exposed in the Mavuradonha Layered Complex, and the lack of late silica-rich differentiated rocks indicates that the Mavuradonha Layered Complex magma chamber behaved as an open system from which these differentiates may have escaped. Xenoliths occur to a minor extent and are restricted to the metagabbros of the layered gabbro series.

Although numerous fine-grained layers are found in the Mavuradonha Layered Complex, especially at the base of macrorhythmic units in the layered gabbro series, these layers cannot be related to intraplutonic quench zones as described for the Kap Edvard Holm Layered Gabbro Complex, East Greenland (Tegner et al., 1993). These fine-grained layers are rather the result of grain-size reduction due to metamorphism than quenching of the fresh magma. This indicates a similar evolution for the Mavuradonha Layered Complex as recorded for the Skaergaard Intrusion and Bushveld Complex (McBirney, 1996; Cawthorn & Walraven, 1998), where intraplutonic quench zones in the sense of Tegner et al. (1993) are also lacking.

In the following discussion the evolutionary history of the Mavuradonha Layered Complex is subdivided into two phases. These are the early Neoproterozoic emplacement, combined with the magmatic evolution, and the late Neoproterozoic / early Phanerozoic metamorphic evolution.

8.1.1 Early Neoproterozoic magmatic evolution

8.1.1.1 Crystallisation age and implications for the parental magma

The inferred magmatic evolution of the Mavuradonha Layered Complex comprises several phases: In the first phases a mafic magma was extracted from a moderately depleted mantle which differentiated into a layered complex at the base of, or within, the

lower continental crust. The extracted magma that gave rise to the layered gabbro series and the meta-anorthosite suites has a tholeiitic and subalkaline character. This tholeiitic character is displayed by progressive iron-enrichment with ongoing differentiation in the AFM-diagram (Fig. 5-2). On the basis of internal zircon structures that were observed by cathodoluminescence imaging, the U/Pb zircon ages measured by SHRIMP and the vapour transfer method constrain the crystallisation age of the complex. The concordant SHRIMP age of 862 ± 4 Ma is interpreted as the time of crystallisation of the complex; this age is also confirmed by two independent ages of ~ 870 Ma (determined by the vapour transfer method on two samples). However, the single zircon evaporation data (849 ± 2 Ma) are somewhat younger than the crystallisation age but verify indirectly the crystallisation age. A seventeen point Sm-Nd whole-rock errorchron (899 ± 46 Ma), based on mafic and anorthositic samples also reflects an event of crust formation that falls within the range of the zircon age. Despite the large error, the errorchron age is similar to the zircon crystallisation age of the complex.

This crystallisation age for the Mavuradonha Layered Complex corroborates the age of <900 Ma suggested by Hanson et al. (1998), but is in contradiction to a proposed minimum age of 1.8 Ga (Hargrove et al., 1998; 2003). These authors interpreted an upper intercept U-Pb age that was determined using the vapour transfer method on single zircons, as the time of crystallisation. As shown by U-Pb and Pb-Pb analyses using SHRIMP, vapour transfer and evaporation techniques combined with cathodoluminescence studies on zircons of metagabbro sample ZZB 123 the history is quite complex. Therefore, it is difficult to determine the crystallisation age solely by vapour transfer analysis without careful interpretation of internal zircon structures.

The concordant SHRIMP age of 862 ± 4 Ma reflects the first event of mafic plutonism in this part of the Zambezi belt which can be related to extension (Dirks et al., 1998). The crystallisation age of the Mavuradonha Layered Complex is similar to the age of 846 ± 68 Ma estimated by Barr et al. (1977) for the Lusaka Granite, but somewhat older than the age of 820 ± 7 Ma determined by Hanson et al. (1988a) for the Ngoma Gneiss in the western part of the Zambezi belt in Zambia. Considering the model of Dirks et al. (1998) these ages reflect the same episode of rifting in the Zambezi belt but probably characterise different stages, although the rift-related granitoids were previously interpreted to be synchronous to closure of a basin.

8.1.1.2 Geochemical evolution during crystallisation of the complex

The second magmatic phase was dominated by fractionation and contamination of the parental magma of the Mavuradonha Layered Complex. Consideration of chemical and mineralogical data indicates a general differentiation to strong iron enrichment for the

Mavuradonha Layered Complex: First, fractionation of clinopyroxene reduced the CaO and CaO/Al₂O₃ ratio. This reduction led to the formation of Mg-rich cumulates now present as metapyroxenites. The residual magma was enriched in iron, alumina and incompatible elements. The onset of plagioclase fractionation accompanying the accumulation of clinopyroxene led to a decrease of the CaO/Al₂O₃ ratio and a further reduction of MgO. The fractionation of these two phases also reduced the Al₂O₃ concentration (Al₂O₃/MgO variation diagram) towards lower Al₂O₃ prior to its marked increase in the meta-anorthosites. Additional to the fractionation of plagioclase and clinopyroxene, orthopyroxene was fractionated as observed in several metagabbros.

The observed trends point to crystallisation of early Mg-rich cumulates in the layered gabbro series, such as pyroxenite and gabbro/norite, that contain the most primitive geochemical and isotopic signature in the complex, comprising the highest Mg#, Cr and Ni contents and the most primitive ϵ_{Nd} values. In addition, the most mafic and most primitive rock type found during field work occurs in the layered gabbro series, i.e. the serpentinite. With higher degrees of differentiation the parental magma merged into anorthosites and ferro-gabbros with an enrichment of the Fe₂O₃_{tot} content in whole-rocks and an onset of fractionation of apatite and Fe-Ti oxides (high phosphorus and titanium concentrations). The increase in Na₂O while Al₂O₃ decreases in the gabbros indicates that fractionating plagioclase and clinopyroxene become progressively more sodic. This trend is verified by the fractionation trends exposed in clinopyroxene (Al_{tot}, Na), and decreasing An-content in plagioclase from high values in the layered gabbro series to relatively low values in the upper meta-anorthosite suite.

Taking into account the serpentinite as the most mafic rock type in the Mavuradonha Layered Complex and the results of geochemistry and petrography, a main differentiation scheme for the Mavuradonha Layered Complex of orthopyroxene – clinopyroxene – plagioclase+clinopyroxene+orthopyroxene – plagioclase+clinopyroxene+apatite+Fe-Ti oxides with a possible crystallisation sequence of harzburgite, now represented by serpentinite, pyroxenites, gabbro/norite, anorthosite+ferro-gabbros could be suggested. The evolution of the Mavuradonha Layered Complex is similar to that proposed for classical layered intrusions such as the Bushveld Complex, the Skaergaard Intrusion or the Kiglapait Intrusion. Differentiation trends from Mg-rich compositions to strong Fe-rich compositions as well as mineral fractionation trends with ongoing differentiation of decreasing An-content in plagioclase and MgO concentration in pyroxenes are reported from the Bushveld Complex (Sharpe, 1985; Eales et al., 1986; Eales & Cawthorn, 1996; Mitchell et al., 1998). The Skaergaard Intrusion is interpreted to be a tholeiitic intrusion that shows differentiation trends of decreasing MgO, Cr and Ni concentrations in whole-rock chemistry, and a strong iron enrichment is verified

in mineral fractionation trends (McBirney & Naslund, 1990; McBirney, 1995; McBirney, 1996). Current studies confirm this tholeiitic trend by showing increasing iron concentration in plagioclase (Tegner, 1997; McBirney, 2000). The Kiglapait Intrusion represents the layered intrusion with the highest Fe-enrichment and, generally, a decrease of An-content in plagioclase and a depletion in MgO concentration in pyroxene with ongoing differentiation was observed (Morse 1981; 1990). Progressive crystallisation of a tholeiitic magma is verified for the Hasvik Layered Intrusion by the depletion in MgO concentration, decrease of An-content in plagioclase as well as apatite as further cumulus phase with ongoing differentiation (Tegner et al. 1999).

8.1.1.2.1 Aspects of layering in the Mavuradonha Layered Complex

Judging from the results of petrography, geology and geochemistry, some conclusions about the observed compositional layering can be drawn: In addition to the general fractionation trends in the complex, compositional layering is best developed in the layered gabbro series, which can also be interpreted as the least evolved part of the Mavuradonha Layered Complex. The layered gabbro series contains several macrorhythmic units in which cyclic differentiation processes in the magma chamber are recorded. Fractionation within the cyclic units of the layered gabbro series mainly produced the crystallisation sequence pyroxenite – gabbro/norite – leuco-gabbro which is often described for sequences with modal/gradual macro – to inch-scale (metre to centimetre) layering of layered intrusions (e.g. McBirney & Noyes, 1979; Mitchell et al., 1998). In the layered gabbro series, sharp basal contacts occur between coarse-grained leuco-metagabbro layers of the preceding unit and the succeeding fine-grained metapyroxenite layer. The observed prominent modal and gravitational layering can be explained by sedimentation processes due to density contrasts in combination with cooling and convection within the magma chamber (Sparks et al. 1993; Naslund & McBirney, 1996). This process resulted in depletion of the residual magma in ferro-magnesian minerals due to sedimentation of these minerals. As pointed out by Sparks et al. (1993) for layering in the Rhum intrusion, the smaller size of the metapyroxenite layers in the Mavuradonha Layered Complex can probably be explained by a lower critical concentration of pyroxene, whereas the formation of leucogabbros and anorthosites demonstrates that the critical concentration of plagioclase was reached.

After the formation of the layered gabbro series the magma became enriched in phases which are not necessary for the formation of ferro-magnesian minerals. The evolved residual magma was enriched in Al_2O_3 , Na_2O , TiO_2 , P_2O_5 , REE, LIL and HFS elements and has a high Fe concentration which led to the accumulation of plagioclase to form anorthosites, whereas the enriched residual magma formed ferro-gabbroic layers.

Meta-anorthosites of both meta-anorthosite suites seem to represent a more transitional type between massive anorthosites (described in detail in Ashwal, 1993) and anorthosites that occur in layered intrusions (Weiblen & Morey, 1980; Haskin & Salpas, 1992; Ashwal, 1993). In the Mavuradonha Layered Complex the transitional character is expressed in the whole-rock geochemistry, the chemical composition of plagioclase and observed textures. Although these meta-anorthosites are related to a layered series of metagabbros/metanorites, they also show typical characteristics of massif-type anorthosites. The plagioclase composition, especially in the upper meta-anorthosite suite, ranges from An₃₉ to An₄₈, similar to plagioclases of massif-type anorthosites that have an An-content of An₄₀₋₆₀ (Ashwal, 1993). The association with metagabbros rich in Fe-Ti oxide is also indicative of massive type anorthosites (Mitchell et al., 1995; Bhattacharya et al., 1998; Markl & Frost, 1999). The lack of ultramafic rocks is a further indication for this transitional occurrence (Ashwal, 1993). Characteristic of anorthosites in layered intrusions is the occurrence of modal layering, cryptic variation in clinopyroxene and plagioclase mineral compositions, and the 'mottled' and 'spotted' appearance (Ashwal, 1993) as observed in the Mavuradonha Layered Complex (Plate I-4 and I-5). Similar transitional-type anorthosites are reported from the Tete-Complex in Mozambique (Evans et al., 1999).

Ferro-metagabbros of the upper and lower meta-anorthosite suites always occur in association with meta-anorthosites. Therefore, the close relationship of ferro-metagabbros and meta-anorthosites as observed in the field points to mechanical accumulation of plagioclase that crystallised from a mafic magma (Haskin & Salpas, 1992). This accumulation also led to an evolved residual magma that was enriched in incompatible elements. Ferro-metagabbros of the Mavuradonha Layered Complex are enriched in Al₂O₃, Na₂O, TiO₂, P₂O₅, and have a high Fe₂O₃ concentrations as seen in geochemistry and mineral chemistry. This chemical characteristic and the correlation between TiO₂ and the LIL and HFS elements indicate that these metagabbroic rocks and the associated meta-anorthosites reflect late stage differentiation products, because the absolute concentrations increase due to the incompatibility of the former elements with higher degrees of fractionation. A further indication for the joint evolution from a single parental magma is the similarity in Sr and Nd isotope characteristic of the meta-anorthosites and ferro-metagabbros.

Interpretations concerning the evolution of ferro-gabbros that are related to anorthosites include models that explain the magmas to be products of deep crustal melting of basic rocks (Duchesne, 1990) and fractionated high-Al gabbroic magmas that were the parental magmas to ferro-gabbros, ferro-diorites and anorthosites (Mitchell et al. 1995; 1996). The evolution of ferro-gabbros and anorthosites of the Laramie Anorthosite Complex from a single parental magma was proposed because of the close relation of

anorthosites and ferro-gabbros, the similarity in isotope systematic and the chemical similarity to ferro-gabbros of the Harpe Lake intrusion (Mitchell et al., 1995).

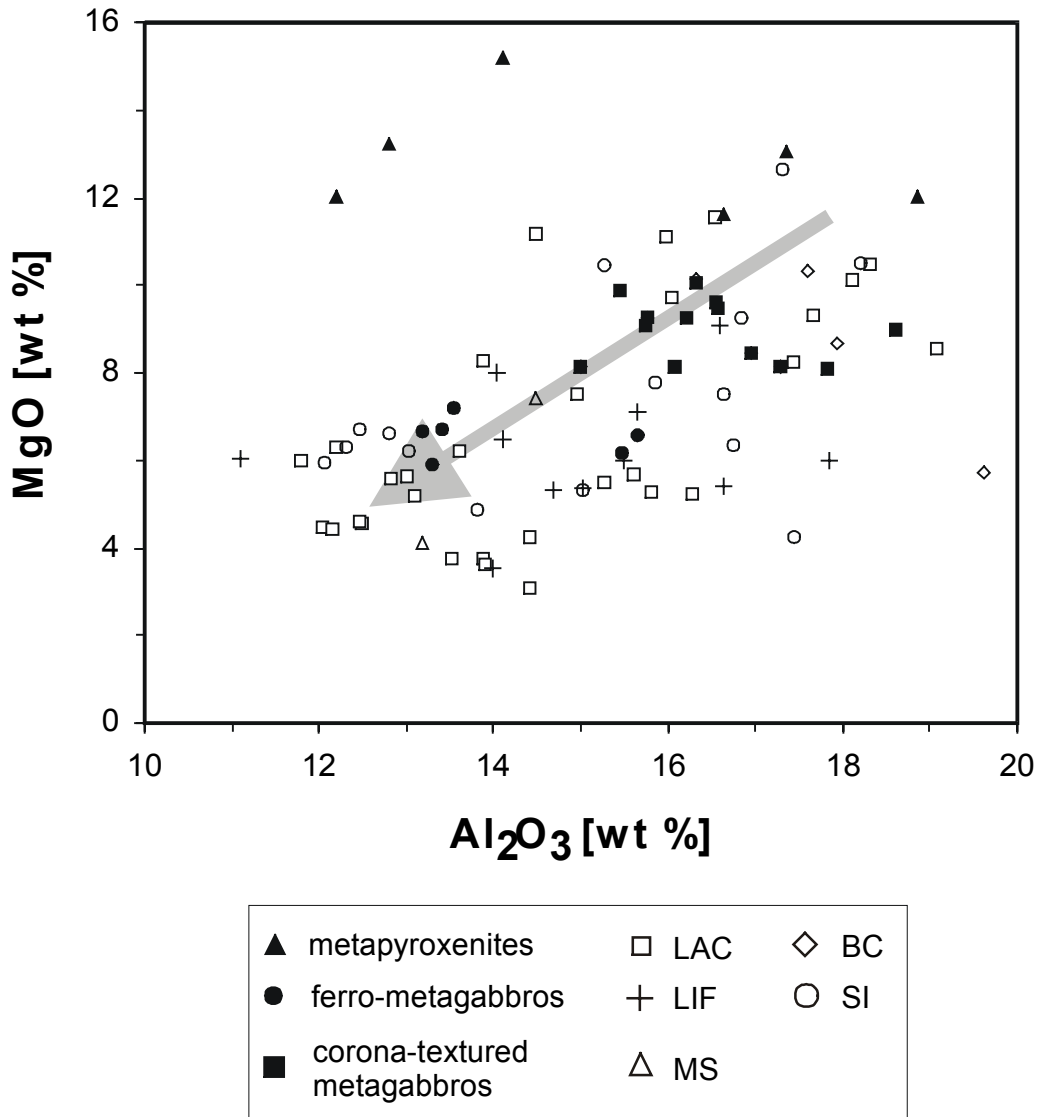


Figure 8-1. Diagram MgO vs. Al₂O₃ (after Markl & Frost, 1999) showing evolution of the Mavuradonha Layered Complex metagabbroic rocks (filled symbols) due to fractionation. For comparison the ferro-gabbros of the Laramie Anorthosite Complex (LAC; Mitchell et al., 1995; 1996), the Lofoten Island Ferrogabbros (LIF; Markl & Frost, 1999), gabbros of the Bushveld Complex (BC; Mitchell et al., 1998), the Skaergaard Intrusion (SI; McBirney, 1998), and the Mashonaland Sills continental tholeiites (MS; Stubbs et al., 1999) are plotted. The grey arrow indicates fractionation trend of the Laramie Anorthosite Complex (Mitchell et al., 1995).

In a diagram Al₂O₃ vs. MgO (Fig. 8-1) the evolution of the mafic metagabbroic rocks of the Mavuradonha Layered Complex is compared with the evolution of the high Al-gabbros of the Laramie Anorthosite Complex and other complexes (McBirney, 1998; Markl & Frost, 1999; Mitchell et al., 1998). Fractionation of the parental magma of the

Mavuradonha Layered Complex defines the same trend as was observed for the high Al-gabbros of the Laramie Anorthosite Complex: after crystallisation of Mg-rich, Al-poor pyroxenites, Al-rich gabbros and afterwards ferro-gabbros were formed. Judging from the arguments discussed above and because of the similarity of ferro-metagabbros of the Mavuradonha Layered Complex to those in other complexes (Fig. 8-1), it is more likely that the parental magma to the ferro-metagabbros and the meta-anorthosites of the Mavuradonha Layered Complex represented a fractionated, Al_2O_3 -rich but SiO_2 -poor, residual magma rather than a different magma pulse which was originated due to anatexis of lower crustal mafic rocks as proposed by Duchesne (1990). Considerations of geochemical variations and comparisons with other ferro-gabbros of layered intrusions (e.g. McBirney, 1998; Fig. 8-1) support the argument that the ferro-metagabbros originated from an evolved, tholeiitic magma which was enriched in incompatible elements.

8.1.1.3 Isotopic evolution and contamination of the complex

The isotopic signature of primitive mafic magmas is generally determined by the source composition. Isotopic investigations therefore, provide essential information about the source region. This is important due to the fact that the cumulate nature may obscure the major and trace element signature. Layered mafic intrusions are well suited for investigations of crustal contamination of mafic magmas, because they provide information on processes within the evolving magma chamber due to their inferred long residence times within the lower continental crust (e.g. Sørensen & Wilson, 1995). Contamination of massive type anorthosites (Ashwal, 1993; Ashwal & Wooden, 1983; Ashwal et al., 1998; Dempster et al., 1999) as well as layered intrusions (Gray & Goode, 1981; Gray et al., 1981; Lambert et al., 1988; Stewart & DePaolo, 1990; Sørensen & Wilson, 1995; Nielsen et al., 1996; Evans et al., 1999; Maier et al., 2000) is a common feature and has affected most of the investigated complexes to different proportions. Figure 8-2 presents the Nd isotope composition of the Mavuradonha Layered Complex in combination with the Nd isotope compositions and the effects of crustal contamination of other layered intrusions and Proterozoic anorthosite complexes. Common to all these complexes is the extraction of the primary magma from a depleted mantle source prior to fractionation and contamination. Displacement from the primary magma composition lowers the ϵ_{Nd} value and increases the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio due to assimilation of granitoid or mafic country rocks of different ages, as proposed for these complexes.

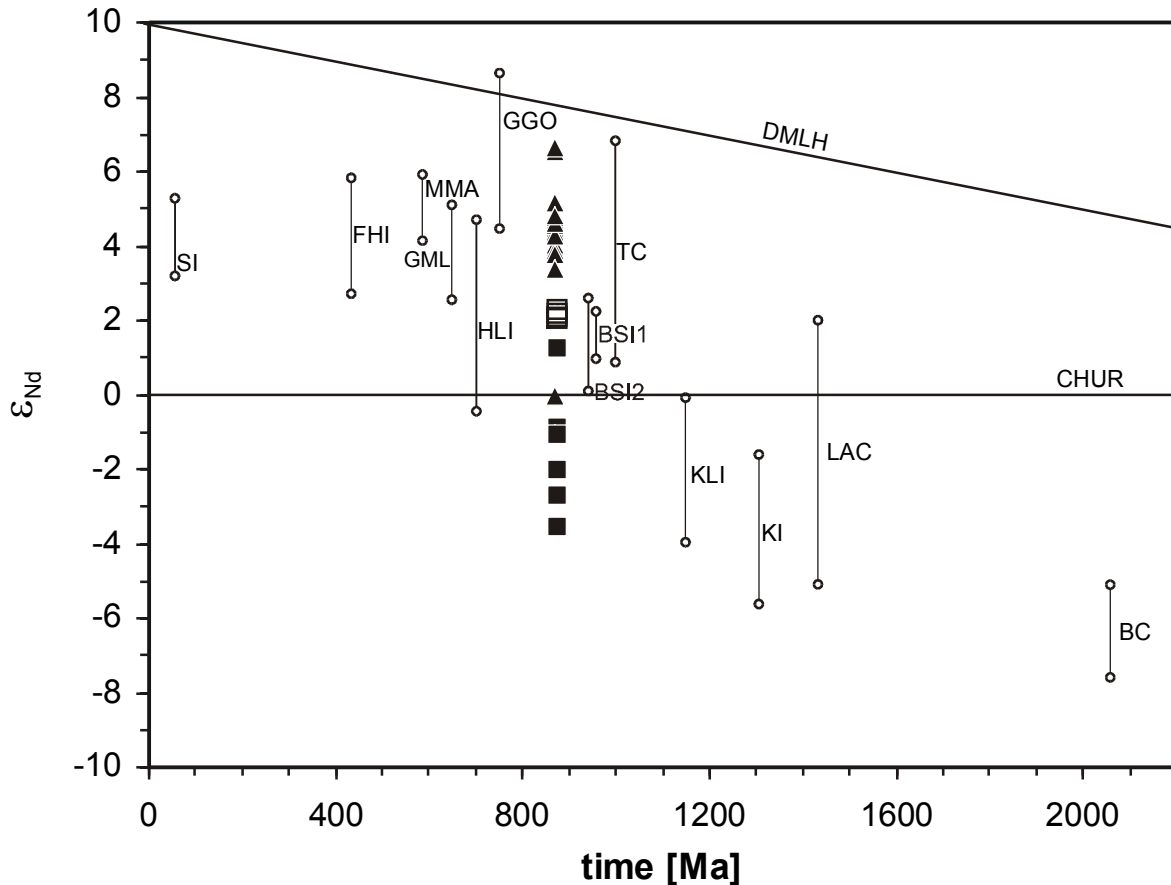


Figure 8-2. ϵ_{Nd} vs. time diagram (after DePaolo, 1981) showing ϵ_{Nd} values for the Mavuradonha Layered Complex and the inlier metagabbroic rocks. For comparison ϵ_{Nd} data for other layered intrusions, Phanerozoic massif-type anorthosites and an ophiolite are plotted. All given data except for some gabbros of the ophiolite plot below the depleted mantle evolution curve of Liew & Hofmann (1988; DMLH), indicating different proportions of crustal contamination. Triangles: Mavuradonha Layered Complex; filled squares: Nyamhandanda Inlier second group; open squares: Nyamhandanda Inlier first group and Chimwaya Hill Inlier metagabbros; BC: Bushveld Complex (Maier et al., 2000); BSI1: Bjerkreim-Sokndal Layered Intrusion (Barling et al., 2000); BSI2: Bjerkreim-Sokndal Layered Intrusion (Nielsen et al., 1996); FHI: Fongen Hyllingen Layered Intrusion (Sørensen & Wilson, 1995); GGO: Gabal Gerf Ophiolite (Zimmer et al., 1995); GML: Glen Mountains Layered Intrusion (Lambert et al., 1988); HLI: Hasvik Layered Intrusion (Tegner et al., 1999); KI: Kiglapait Intrusion (DePaolo, 1981); KLI: Kalka Layered Intrusion (Gray et al., 1981); MMA: Madagascar massif-type Anorthosite (Ashwal et al., 1998); LAC: Laramie Anorthosite Complex (Mitchell et al., 1995); SI: Skaergaard Layered Intrusion (Stewart & DePaolo, 1990); TC: Tete Complex (Evans et al., 1999).

However, the trend from negative ϵ_{Nd} values in Archaean intrusions to positive ϵ_{Nd} values for recent intrusions shown in this diagram is a reflection of the crystallisation age in conjunction with the age of the country rocks. Positive ϵ_{Nd} values point to assimilation of Proterozoic basement, e.g. the Fongen-Hyllingen Layered Intrusion (Sørensen & Wilson, 1995), whereas negative ϵ_{Nd} values, e.g. the Bushveld Complex (Maier et al., 2000) are the result of assimilation of evolved crustal material of Archaean age (Longhi et al., 1999; Fig. 8-2). In the Mavuradonha Layered Complex the effects of crustal contamination are

displayed by a shift of the ϵ_{Nd} values for the metagabbroic rocks to lower values than the depleted mantle evolution curve. The Nd isotopic composition in combination with the estimated age of the complex fits well to the general trend of “the younger the age the younger the contaminant” that was proposed by Longhi et al. (1999). The diagram demonstrates that the Mavuradonha Layered Complex shows the same Nd isotope evolution/composition as has been described for other layered intrusions of similar age.

Additionally, analogies of the Mavuradonha Layered Complex with several other major layered intrusions suggest that crustal contamination was a significant aspect in the petrogenesis of this complex. These analogies are:

- (1) The Nd and Pb isotopic signatures of the Mavuradonha Layered Complex point to a depleted mantle source.
- (2) The effects of crustal contamination may have lowered the ϵ_{Nd} values of the metapyroxenites, metagabbros and meta-anorthosites. However, the isotopic composition of the samples from the Mavuradonha Layered Complex is still relatively primitive in comparison to the inlier rocks and other complexes (Fig. 8-2).
- (3) As pointed out earlier, metapyroxenites of the Mavuradonha Layered Complex have the most primitive ϵ_{Nd} composition and low Nd concentration in the complex, ferro-metagabbros, in contrast, yielded low ϵ_{Nd} values but high Nd concentrations. The decrease of ϵ_{Nd} with stratigraphic height and the increase of Nd concentrations with ongoing differentiation as observed in the Hasvik Layered Intrusion (Tegner et al., 1999), the Bushveld Complex (Maier et al., 2000) or the Kiglapait Intrusion (DePaolo, 1985) is, therefore, also characteristic for the evolution of the Mavuradonha Layered Complex.
- (4) The evolution of the isotopic systems indicates assimilation of wall rocks in upper parts of the intrusion by a progressive decrease of the ϵ_{Nd} values from the layered gabbro series to the upper meta-anorthosite suite. Assimilation and contamination occurred mainly in the upper parts of the Bjerkreim-Sokndal Layered Intrusion (Nielsen et al., 1996), the Skaergaard Intrusion (Stewart & DePaolo, 1990), and the Fongen-Hyllingen Layered Intrusion (Sørensen & Wilson, 1995), whereas crystallisation took place at the base of these complexes. This resulted in a decrease in ϵ_{Nd} and increase in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the direction of the roof of the magma chambers.
- (5) The variation in the initial $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, especially in rocks of the meta-anorthosite suites, indicates that the system was not closed and that

there was at least some assimilation of country rocks or melts of granitoid wall rocks. This probably occurred in a similar manner as proposed for the Hasvik Layered Intrusion (Sørensen & Wilson, 1995) or the Kiglapait Intrusion (DePaolo, 1985) and is also supported by relatively homogeneous Pb, Nd and Sr isotopic compositions in the meta-anorthosite suites. A buoyant, partial molten layer of wall rocks at the roof of the basaltic magma for some layered intrusions has been suggested (DePaolo, 1985; Stewart & DePaolo, 1990; Sørensen & Wilson, 1995). As pointed out by these authors, assimilation of this molten material took place in the immediate contact zone to the mafic magma. Therefore, the observed relatively homogeneous composition in the meta-anorthosites suite from the Mavuradonha Layered Complex may result from mixing of the molten wall rock layer and the magma. As proposed by Stewart & DePaolo (1990) the relatively low rate of contamination may be the result of inefficient mixing of layers of contrasting density.

- (6) Abundant xenoliths of wall rocks in the layered gabbro series are further evidence for crustal contamination (Stewart & DePaolo, 1990). This is also corroborated by inherited zircons that were only found in the layered gabbro series. However, xenoliths are not found in the upper parts of the Mavuradonha Layered Complex. Therefore, it is likely that the xenoliths were dislodged from the roof or the base of the chamber during the emplacement of the parental magma to the layered gabbro series and, similar to the Hasvik Layered Intrusion, remained within the layered gabbros series. For the Hasvik Layered Intrusion the presence of xenoliths within the magma chamber is proposed because of an increase of xenoliths at the top of the Layered Series and distinct isotopic compositions in the intrusion (Tegner et al. 1999). Large numbers of wall rock xenoliths are also reported from the lower parts of the Bjerkreim-Sokndal Layered Intrusion (Nielsen et al., 1996).

Considering the ages of inherited zircons, possible contaminants with ages of 1.8 Ga are lacking in the Mt. Darwin area, which precludes deducing the properties of contaminants from the surrounding rocks. However, a systematic isotopic study of the country rocks has not been carried out. Therefore, it is also premature to consider detailed mixing or assimilation and fractional crystallisation models (AFC; DePaolo, 1981; 1985) for the metagabbroic, meta-anorthositic and surrounding country rocks. On the other hand, knowledge of the isotopic composition of possible contaminants is necessary for quantitative calculations of AFC processes (DePaolo, 1981).

Based on the isotopic systematics, ϵ_{Nd} varies from +3.0 to +5.2 in the Skaergaard

Intrusion which was contaminated with 2 - 4% of the surrounding Precambrian gneisses (Stewart & DePaolo, 1990). A larger range for ϵ_{Nd} from 1.6 to 5.8 was observed for the Fongen-Hyllingen Layered Intrusion and a maximum of 10 % of assimilation of crustal material was proposed (Sørensen & Wilson, 1995). In contrast, the evolution of the Kalka Layered Complex or the Hasvik Layered Intrusion were different. Assimilation rates of 25 % were calculated for the Kalka Layered Intrusion which caused a variation in ϵ_{Nd} of 0 to – 3.9 (Gray et al., 1981). Tegner et al. (1999) interpreted the Hasvik Layered Intrusion as one of the most contaminated intrusions because of assimilation rates of ~21% with ϵ_{Nd} values ranging from +4.76 to –3.26. Comparing the isotopic composition of the Mavuradonha Layered Complex with other complexes, the assumed contamination rates and the scatter in ϵ_{Nd} values (+3.3 to +6.7) are similar to the rates of the Skaergaard Intrusion (Stewart & DePaolo, 1990) and the Fongen-Hyllingen Layered Intrusion (Sørensen & Wilson, 1995). Therefore, contamination of less than 10 % by country rocks can be estimated for the Mavuradonha Layered Complex.

Dirks et al. (1998) proposed initiation of rifting for the Zambezi belt in NE Zimbabwe at around 900 Ma, and the results of this study are in line with this proposal. The relatively primitive character of the $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic composition of the layered gabbro series, indications for contamination of a mantle-derived melt in the magma chamber with lower crustal rocks as well as the crystallisation age of 862 ± 4 Ma all points to upwelling of the asthenosphere caused by initiation of rifting. Similar processes with similar isotopic compositions (Fig. 8-2) were described for the Glen Mountains Layered Complex which was interpreted to indicate initial rifting of the Oklahoma aulacogen (Lambert et al., 1988) based on the isotope composition in conjunction with the crystallisation age.

However, the Pb and Sr isotopic systems show major changes that cannot solely be explained by crustal contamination. Although both systems show a depleted mantle signature for single samples, the influence of metamorphism and the infiltration of fluid phases are evident in the evolution of both isotopic systems in addition to the effects of crustal contamination. These effects led to equilibration of the Pb isotopic system at around 550 Ma and a shift to higher radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. These changes were also corroborated by investigations in thin-section such as the formation of chlorine-rich scapolites and the formation of corona-textures, which are related to fluid infiltration and metamorphism.

8.1.2 Implications for the tectono-metamorphic evolution of the Mavuradonha Layered Complex

Pressure/temperature estimates in combination with the age determinations of major rock-forming and accessory minerals provide constraints on the tectonic evolution of this part of the Zambezi belt. It is proposed that the granulite-facies metamorphic overprint of the Mavuradonha Layered Complex occurred during the Pan-African orogeny. An age of 554 ± 13 Ma was determined, based on metamorphic zircon overgrowth that can be assigned to granulite-facies metamorphism. This event also produced large amounts of garnet in the upper meta-anorthosite suite next to the Zambezi Escarpment. These garnets are taken to reflect the *PT* conditions for the Mavuradonha area; the observed granulite-facies *PT* conditions reached a peak metamorphic pressure of ~ 13 kbar at a temperature of ~ 840 °C. The lack of prograde major element zoning in garnet, especially the lack of zoning in the grossular component, indicates temperature conditions in excess of 750 °C. It has been proposed that above this temperature this characteristic prograde feature becomes erased (Chakraborty & Ganguly, 1990). Age determinations for the high-grade event, using these garnets of the upper meta-anorthosite suite, yielded a five point Sm-Nd garnet - whole-rock isochron age of 545 ± 9 Ma. This age indicates rapid cooling of the complex after granulite-facies conditions. The *PT* data for the high-grade metamorphic event reported by Hargrove et al. (2003) are consistent with the data reported in this study and suggest that the complex was overprinted under high-temperature, high pressure metamorphic conditions. Based on age data that were calculated for a similar metamorphic event in a neighbouring terrain, Hargrove et al. (2003) assumed an age of 870–850 Ma for the granulite-facies overprint. However, the age data for the granulite-facies event presented in this study indicate that the high-grade metamorphic event occurred at least 300 Ma later. This age for the high-grade metamorphic event is the first Pan-African age for granulite-facies metamorphism reported for the Zambezi Allochthonous Terrain of NE Zimbabwe.

After high-temperature metamorphism cooling with slight decompression occurred in the complex. Amphibolite-facies conditions attained a pressure of 11 kbar at 680 °C, followed by further cooling of the complex at 526 to 501 Ma, an age interval that was determined using the U/Pb isotopic system applied to rutile. The cooling event under amphibolite-facies conditions can be regarded as a static retrogression, forming corona textures of hornblende and quartz or pargasitic amphibole around clinopyroxene. It was accompanied by the infiltration of highly saline fluids inferred from the occurrence of chlorine-rich amphiboles and scapolites. The decompression textures, e.g. plagioclase coronas around garnet in garnet-bearing amphibolites, are further evidence for cooling and uplift. This is consistent with the interpretation of the U-Pb titanite and zircon data of

550-530 Ma for amphibolite-facies metamorphism as estimated by Hargrove et al. (2003). The age of amphibolite-facies metamorphism is also in line with the age of metamorphism proposed by Vinyu et al. (1999) for the Rushinga area, a neighbouring terrane west of the Mt. Darwin area. They proposed a similar age (535 Ma) for amphibolite-facies metamorphism in leucomigmatites, based on U/Pb isotope systematics in zircons and sphene. This age for the cooling event is only slightly younger than the age for the granulite-facies event, supporting petrographic investigations and cathodoluminescence imaging that provide no indications for a long period of cooling. Only small zones of retrograde exchange at garnet rims and flat major element patterns were observed, and no evidence for a second metamorphic event was found.

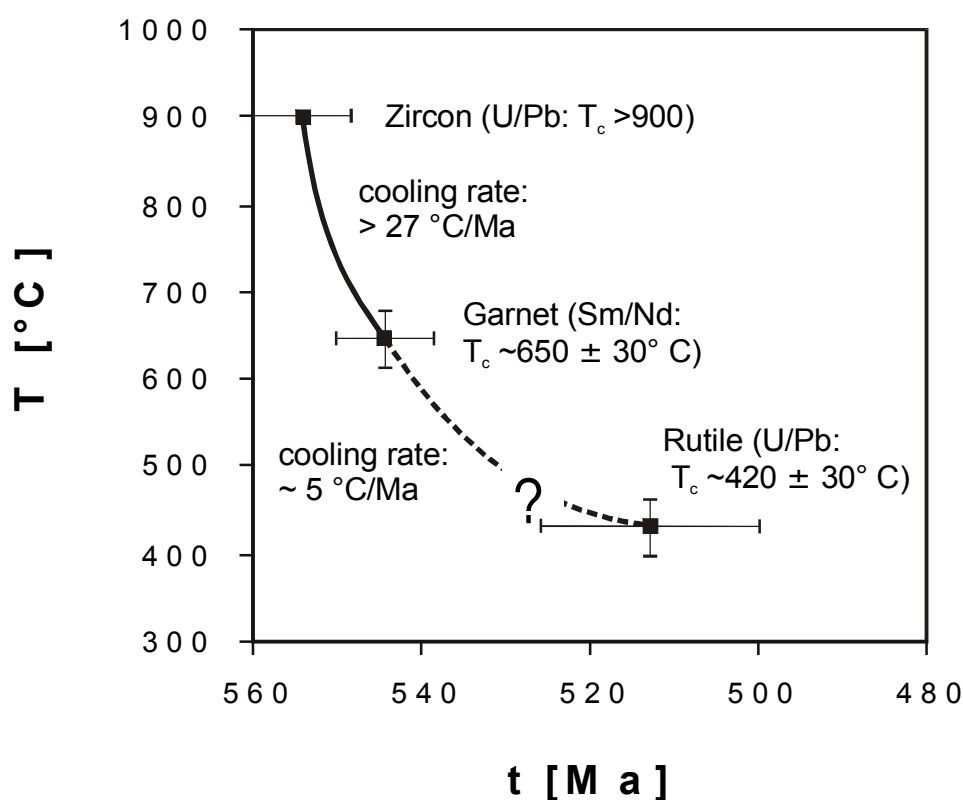


Figure 8-3. Tt-path for the metamorphic evolution of the Mavuradonha Layered Complex. Note short time interval and rapid cooling after peak metamorphism and slower cooling to greenschist-facies conditions in the following time interval of ~40 Ma.

The sequence of ages presented leads to the interpretation that granulite-facies metamorphism occurred at 554 Ma and was followed immediately by rapid cooling from HP/HT conditions to lower amphibolite-facies conditions. The similarity of the U-Pb zircon ages and the Sm-Nd garnet whole-rock isochron age indicate that initial cooling rates have been larger than 20 °C/Ma (Fig. 8-3). At about 30 to 50 Ma after peak metamorphism, cooling reached greenschist-facies conditions with cooling rates of

~5 °C/ Ma as indicated by U-Pb rutile ages (Fig. 8-3). At the beginning relatively high cooling rates prevailed in the Mavuradonha Layered Complex, whereas with prolonged cooling the rates were only a little higher than the 1-4 °C/Ma reported from other mountain belts, such as the Grenville orogen (Mezger et al. 1993), the Eastern Ghats belt (Mezger & Cosca, 1999) or the Pikwitonei granulite domain (Metzger et al., 1989). Cooling rates of about 10 °C/Ma were obtained by Mezger et al. (1993) in the Grenville orogen, and immediately after the high-grade event these authors also reported cooling rates of 16 °C/Ma. Therefore, the relatively high cooling rates immediately after the peak metamorphic event point to rapid uplift during the first cooling steps.

The high-grade metamorphic overprint at around 554 Ma reflects an event of crustal thickening (Dirks et al. 1998), which may be related to the final amalgamation of Gondwana (Stern, 1994) and is consistent with the interpretation of Hargrove et al. (2003). This Pan-African granulite-facies event is the same major Pan-African tectono-thermal event as reported from Madagascar and Malawi (Kröner et al., 1996; 1997; 1999; 2001), the Lufilian Arc (John et al., 2001; John, 2001) and that led to a high-grade overprint in the Damara belt (Jung & Mezger, 2001). Collisional tectonics accompanied by crustal thickening is one of the processes in orogenic belts that is responsible for granulite-facies metamorphism (Ellis, 1987; Bohlen, 1987). The Mavuradonha Layered Complex is characterised by nearly isobaric cooling which has also been observed for other granulite-facies terrains (Bohlen & Mezger, 1989; Harley, 1989). After this period of cooling and slight decompression, the Mavuradonha Layered Complex was rapidly uplifted and exhumed. Rapid exhumation, as stated earlier, occurred later during in close relation to the Pan-African orogeny that also caused overthrusting of the Mavuradonha Layered Complex to its present position. Since no relics of the prograde evolution were observed during this study it is not possible to decide whether $P_{\max}$ was reached prior to $T_{\max}$ (clockwise *PT* path) or $T_{\max}$ was reached prior to $P_{\max}$ (anti-clockwise *PT* path).

8.2 Nyamhanda Inlier and Chimwaya Hill Inlier

The magmas that gave rise to the formation of the inlier metagabbroic rocks had a tholeiitic, subalkaline character (Fig. 5-2). Considering the results of the geochemistry, the inlier rocks show similar major and trace element characteristics (variation diagrams, Fig. 5-6) as continental tholeiites in southern Africa such as the Mashonaland Sills (Stubbs et al., 1999), metagabbros of the Lufilian Arc (Tembo, 1994) and to other Pan-African metagabbros from the Masoso, the Rushinga and Chimanda Metamorphic Suites (Barton et al. 1993). In contrast to the depleted geochemical signature of the Mavuradonha Layered Complex rocks, the metagabbros of both inliers are enriched in FeO, which is typical for

tholeiitic magmas (Wilson, 1989). Additionally, the metagabbroic rocks are more enriched in LIL and HFS elements than the Mavuradonha Layered Complex cumulate rocks.

Comparing the isotopic composition of the Nyamhanda and Chimwaya Hill Inlier metagabbros with other complexes, crustal contamination also played an important role during crystallisation of the metagabbroic rocks especially for the second group of the Nyamhanda Inlier metagabbros. The observed initial isotopic compositions rule out direct derivation from a depleted mantle source. However, the composition of the inlier metagabbros suggest mixing between a tholeiitic magma derived from a depleted mantle source with crustal rocks that are characterised by low ϵ_{Nd} values and high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Petrographic evidence and field observations suggest that several magmas having different source characteristics coexisted in the Nyamhanda Inlier, an interpretation which is consistent with the isotopic data. Alternatively, the isotopic data suggest that the amount of assimilation was different in the different intrusions. This difference is best displayed in the lower, negative initial ϵ_{Nd} value and more radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the second Nyamhanda Inlier group compared to the first Nyamhanda Inlier group or the Chimwaya Hill Inlier metagabbros. In Figure 8-2 the similarity in ϵ_{Nd} composition of the second group of the Nyamhanda Inlier to pre-Pan-African complexes is obvious and suggests similar degrees of contamination.

The isotopic signature of the Sm/Nd and Rb/Sr isotope systems of the inlier rocks is also different from to the metagabbros of the Mavuradonha Layered Complex. The initial Sr and Nd isotopic compositions of the inlier metagabbros suggest that either the parental magma had a different source region or the amount of contamination was much higher than was suggested for the Mavuradonha Layered Complex. This is also evident in the common Pb composition, because the two samples from the second group of the Nyamhanda Inlier seem to be derived from a source with higher μ -value than the gabbros of the Mavuradonha Layered Complex and Chimwaya Hill Inlier. Since no AFC calculations were undertaken contamination rates (Fig. 8-2) as high as proposed for the Hasvik Layered Intrusion (~21 %, Tegner et al., 1999), the Kiglapait Intrusion (DePaolo, 1985) and the Kalka Layered Intrusion (~25 %, Gray et al., 1981) can be assumed for the second Nyamhanda Inlier group. Contamination of the Chimwaya Hill Inlier and first group Nyamhanda Inlier metagabbros occurred to lesser extend than for the second Nyamhanda Inlier group. The small shift to higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ values than the Bulk Earth could either be interpreted as crustal contamination and/or as the result of the high-grade metamorphic overprint and/or infiltration of fluid phases.

The whole-rock geochemistry, petrography and the different isotopic compositions of the inlier metagabbros are strong evidence for a different evolution of these rocks. Interpretations of the Nyamhanda Inlier data point to several different metagabbroic bodies

in this inlier. These interpretations are in contrast to previous views for the evolution of this inlier (Leitner & Phaup, 1974; Bache et al., 1983, 1984). These authors interpreted the metagabbros as part of a single complex, which was complexly infolded within the underlying gneisses. Additionally, based on the results of this study, the inlier metagabbros cannot be related to the Mavuradonha Layered Complex metagabbroic rocks.

9 Interpretation in the regional context

Most conclusions from this study have already been summarized in the previous chapters. These lead to the revision of the geological setting for the Mavuradonha Mountains. This chapter discusses the implications of this study for the Zambezi Allochthonous Terrain and for the evolution of the Zambezi belt in NE Zimbabwe.

9.1 Implications for the evolution of the Zambezi Allochthonous Terrain

The combination of petrological, geochemical, geochronological and isotopic investigations has unravelled the origin of the metagabbro/meta-anorthosite association of the Mavuradonha Layered Complex. This association and the contemporaneous evolution of meta-anorthosites and metagabbros was previously recognised by Barton et al. (1991), Hargrove et al. (1998), Makanza (1993) and Muzuwa (1993). However, an evolutionary model in which the Mavuradonha Layered Complex is interpreted as a layered intrusion can now be proposed for the entire meta-anorthosite/metagabbro/ amphibolite association that is exposed in the Mavuradonha Mountains. The conclusions presented here contradict the geological setting as a large metagabbroic sill as proposed by Bache et al. (1983). The main conclusions from this study as follows:

- The parental magma of the Mavuradonha Layered Complex recognised in early Neoproterozoic times at 862 ± 4 Ma.
- The metagabbro/meta-anorthosite/amphibolite association recognised from a tholeiitic, subalkaline magma. The Mavuradonha Layered Complex samples have a typical continental tholeiite geochemistry (LeMaitre, 1976; Cox, 1980).
- The initial Nd isotopic data suggest that the parental magma for rocks of the Mavuradonha Layered Complex was derived from a moderately depleted mantle source.
- The isotopic composition of the highly differentiated rocks, the abundant xenoliths in the layered gabbro series, as well as inherited zircons, are convincing evidence for crustal contamination and for a crustal setting. Crustal contamination is indicated by variable initial ϵ_{Nd} values. Similar signatures of crustal contamination in southern Africa are found in the Tete Complex of NW Mozambique (Evans et al., 1999), the Malagasy anorthosites (Ashwal et al., 1998) and the Messina

Layered Intrusion of the Limpopo belt (Barton, 1996). In Tanzania, different isotopic signatures are reported for Pan-African anorthosites (Möller, 1995; Maboko, 1995) implying a late Archaean contaminant (Möller, 1995; Möller et al., 1998).

- Macro-rhythmic units and small-scale layering has been identified in the Mavuradonha Layered Complex. Three major sequences have been distinguished based on their geochemical characteristics and mineral chemistry:
 - The **layered gabbro series** comprise several macro-rhythmic units each showing small-scale layering. Fractionation in the layered gabbro series produced sequences of pyroxenites, gabbro/norites, leucogabbros and anorthosites.
 - The **lower meta-anorthosite suite** contains anorthosites with internal layering of mainly gravitational origin. Additionally, large volumes of ferro-gabbroic units occur in this suite.
 - The **upper meta-anorthosite suite** consists of large volumes of anorthosite-ferro-gabbro suites. Each suite is separated from the one below by sharp contacts.
- The meta-anorthosites of the Mavuradonha Layered Complex belong to a transitional type of anorthosite. The meta-anorthosites and the related ferro-metagabbros originated from an evolved residual magma enriched in incompatible elements.
- The complex was metamorphosed in latest Neoproterozoic times at 554 Ma under high-pressure granulite-facies conditions with peak *PT* conditions of 840 ± 30 °C at 13 ± 2 kbar.
- Cooling under amphibolite-facies to greenschist-facies conditions occurred within 50 Ma until the closure temperature of rutile (420 °C) was reached. This cooling was accompanied by infiltration of highly enriched saline fluids as shown by the abundance of late, chlorine-rich scapolites and amphiboles.

- Numerous pegmatites were emplaced during metamorphism, and a crystallisation age of 545 Ma was determined on zircons from pegmatites. This age reflects the same event as reported by Hargrove et al. (1998) for a pegmatite sampled in the Ocellar Gneiss and by Vinyu et al. (1999) for a pegmatite in the nearby Rushinga area.
- The latest event affecting the Mavuradonha Layered Complex was the formation of extensional faults which are related to the formation of the Zambezi Valley. This event was accompanied by brittle deformation.

Several other metagabbroic bodies are exposed in the Mt. Darwin area. Two of them are exposed in inliers and are also the subject of this study. The parental magma of these two metagabbroic bodies have a subalkaline, continental tholeiitic composition and are thus different from the metagabbros of the Mavuradonha Layered Complex. They are enriched in incompatible elements and REE. Both metagabbros show neither geochemical nor petrological evidence for igneous layering, although in some cases the field occurrence implies layering. The isotopic evidence for interaction with Archaean or Kibaran-age crust has important implications for the evolution of the inlier metagabbros. It can be assumed that variable assimilation is required to produce the wide range of the observed isotopic compositions. The older gneisses surrounding the inlier metagabbros are possible contaminants for the inlier gabbros. The initial Sr and Nd isotopic compositions require either a parental magma composition distinct from that which produced the Mavuradonha Layered Complex gabbros, or it required more contamination by Archaean or Proterozoic crust. Additionally, these isotopic characteristics also imply a crustal setting for the inlier gabbros.

A schematic three-stage model (Fig. 9-1) for the evolution of the Zambezi Allochthonous Terrain in the Mt. Darwin area is as follows:

First stage ~1020 Ma: Formation of a granitoid basement

Age determinations on zircons from the granitoid Ocellar Gneiss which structurally underlies the Mavuradonha Layered Complex indicate that this is the oldest rock unit in the study region. The age of 1022 ± 7 Ma is consistent with the age of 1052 ± 1 Ma determined for another sample of the Ocellar Gneiss SE of the

Mavuradonha Layered Complex (Hargrove et al., 2003) and points to granitoid magmatism during this Kibaran event. This age is similar to ages reported for A-type granitoids in northern Mozambique (Sacchi et al., 1984; Pinna et al., 1993; Kröner et al., 1997; 2001).

Second stage ~870 Ma: Magmatic emplacement of the Mavuradonha Layered Complex

Geochronological and Nd isotope data show that the Mavuradonha Layered Complex was emplaced at 862 ± 4 Ma and originated from a depleted mantle-derived source. The parental magma with a subalkaline and tholeiitic character intruded into, or at the base of, the lower continental crust. Emplacement occurred in an extensional regime, probably above a zone of upwelling of the asthenosphere. During emplacement, a magmatic layering was formed and contamination with crustal material of unknown sources occurred. Other gabbroic bodies, now exposed in several inliers, were probably emplaced at the same time and also in the same crustal setting. The parental magmas had a similar tholeiitic and subalkaline chemistry but were probably derived from different mantle sources than the Mavuradonha Layered Complex.

Third stage ~550 Ma: Metamorphic evolution and tectonic emplacement:

The third stage of the model, more than 300 Ma after crystallisation of the layered complex, comprises the metamorphic evolution, emplacement of pegmatites and thrusting of the complex and other metagabbros onto the granitoid Ocellar gneiss. This stage probably occurred during an episode of compression during the final phase of the Zambezi orogeny. Peak metamorphic conditions are estimated at 840 ± 30 °C and 13 ± 2 kbar, followed by a period of near isobaric cooling. Retrograde conditions under amphibolite-facies conditions (680 ± 20 °C at 11 ± 2 kbar) were followed by further decompression and cooling as observed in greenschist-facies mineral assemblages. The age of granulite-facies metamorphism is defined by a zircon age of 554 ± 13 Ma, followed by cooling as shown by a Sm-Nd garnet - whole-rock age of 545 ± 9 Ma and rutile cooling ages in the range 526 to 501 Ma. The amphibolite-facies metamorphic event was accompanied by extensive pegmatite formation at around 550 Ma, which is documented by various ages in the Zambezi Allochthonous Terrain.

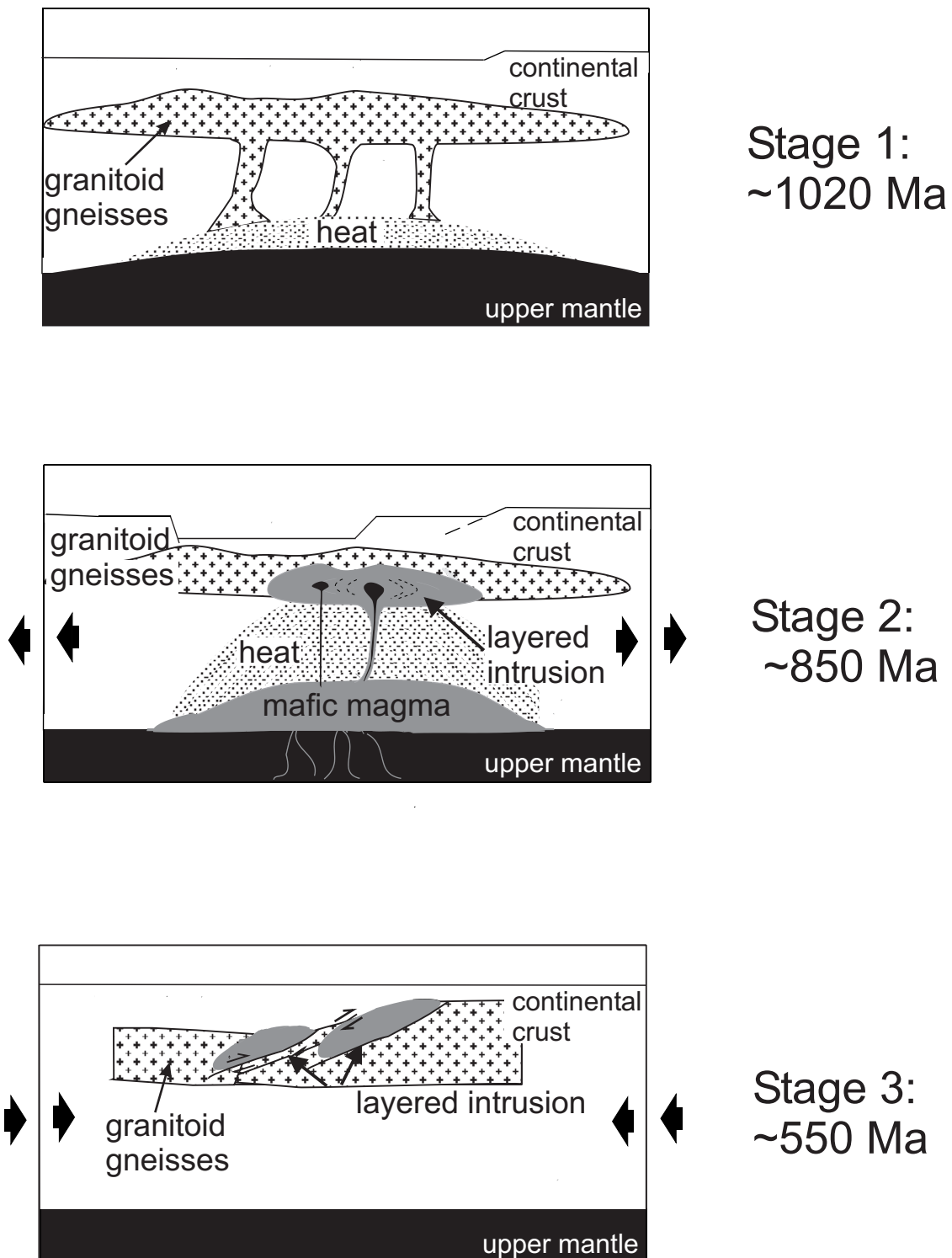


Fig. 9-1. Schematic evolutionary model for the Zambezi Allochthonous Terrain in the Mt. Darwin area, northern Zimbabwe. The model is not to scale. For explanations see text.

9.2 Implications for the Pan-African orogenic belt system

The results obtained in this study provide insights for the evolution of the Pan-African mobile belt system in southern Africa. Dirks et al. (1998) proposed a phase of lithospheric extension for the NE part of the Zambezi belt, beginning at around 900 Ma and followed by a compressional event more than 300 Ma later. Ages of 900 Ma for the extensional event and 550 Ma for the compression as proposed by these authors have been constrained in this study. The distinction between emplacement of the Mavuradonha Layered Complex at 862 ± 4 Ma and the more than 300 Ma younger granulite-facies metamorphic event is important for an understanding of the evolution of the central position of the Zambezi belt.

The crystallisation age of 862 ± 4 Ma reflects a phase of initial rifting in which magma was extracted from the mantle and emplaced in the lower or middle crust, probably during an event of magmatic underplating. After the period of rifting the Mavuradonha Layered Complex was metamorphosed some 300 Ma later under HP granulite-facies conditions during a compressional event at ~ 550 Ma. The complex was thrust onto the Ocellar gneiss prior to, or during, the high-grade thermal event. In addition, the lack of intrusive contacts between the metagabbros and their host rocks, the lack of Archaean gneisses in the Zambezi Allochthonous Terrain, and the present position of the Mavuradonha Layered Complex on top of the Ocellar Gneiss, all favour tectonic emplacement of the metagabbros. These rocks experienced an uncertain amount of lateral displacement. Cooling after peak metamorphism was accompanied by the formation of chlorine-rich scapolites. This is a widespread phenomenon in southern Africa and has also been recognised in the Lufilian Arc (Cosi et al., 1992; Tembo, 1994), in Malawi, and the Tete Complex of Mozambique (Andreoli, 1984; Evans et al., 1999). The Mavuradonha Layered Complex was uplifted to higher structural levels during compression.

The above model for the Zambezi Allochthonous Terrain supports a model of early Pan-African rifting which has also been proposed for the evolution of the Lufilian Arc and the Zambezi belt in Zambia (Porada, 1989; Porada & Berhorst, 2000). However, in the case of the Zambezi belt in northeastern Zimbabwe, no gabbroic rocks with MORB signature are preserved, typical sequences of oceanic crust are lacking and HP-assemblages were not discovered. Due to the different settings of metagabbros in the Lufilian Arc and in the Zambezi belt of Zambia (Tembo, 1994; John, 2001), a genetic link and similar evolution of the Zambezi belt and the Lufilian Arc are not supported.

The age of ~ 870 Ma for rift-related magmatism in the Mavuradonha Layered Complex is similar to ages for rifting in the East African Orogen (Kröner et al., 1991; 1992) and the Lufilian Arc. Rift-related metagabbros in the Lufilian Arc were dated at 935 ± 60 Ma (Carr et al., 1986; cited in Tembo, 1994), whereas in the Damara belt the first

recognised rifting occurred at around 760 Ma (Miller, 1983). An evolution as a failed rift (Porada, 1989) may be proposed for the Zambezi belt, if this age of ~870 Ma of initial rifting is considered. A further argument for such a rift is the difference in tectonic setting between the Zambezi belt and the Lufilian Arc, because there are reported MORB characteristics in parts of the Lufilian Arc (Oliver et al. 1998; John et al., 2000; Johnson & Oliver, 2000; John, 2001), but not in the Zambezi belt. A second possibility is the formation of several small basins along strike of the present-day transcontinental network of mobile belts (Porada & Berhorst, 2000). In this case, the basins in the Lufilian Arc were probably larger and had evolved to an ocean as proposed by John (2001), which may be an explanation for the MORB signature in this part of the Lufilian Arc.

The results of this study constrain two events for the Zambezi Allochthonous Terrain in the Mt. Darwin area that were also proposed for the Zambezi belt of NE-Zimbabwe by Dirks et al. (1998). The first event led to the formation of a failed rift, or to an intracratonic basin, with the extraction of primitive depleted-mantle derived magmas during an early stage of rifting caused by upwelling of the asthenosphere. The Mavuradonha Layered Complex and the inlier gabbros represent gabbroic complexes that were emplaced during this episode of intense plutonic activity. Thereafter, during compression as part of the late Pan-African orogeny, the rift/basin was closed. Closure of the basin was accompanied by high-grade metamorphism, intense deformation (Barton et al., 1991), and the emplacement of numerous pegmatites. This late Pan-African event led to the formation of the present-day transcontinental network of mobile belts in southern Africa.

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Appendix

A-I	Geology of the study areas.....	i
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A-I Geology of the study areas

SW-NE cross-section

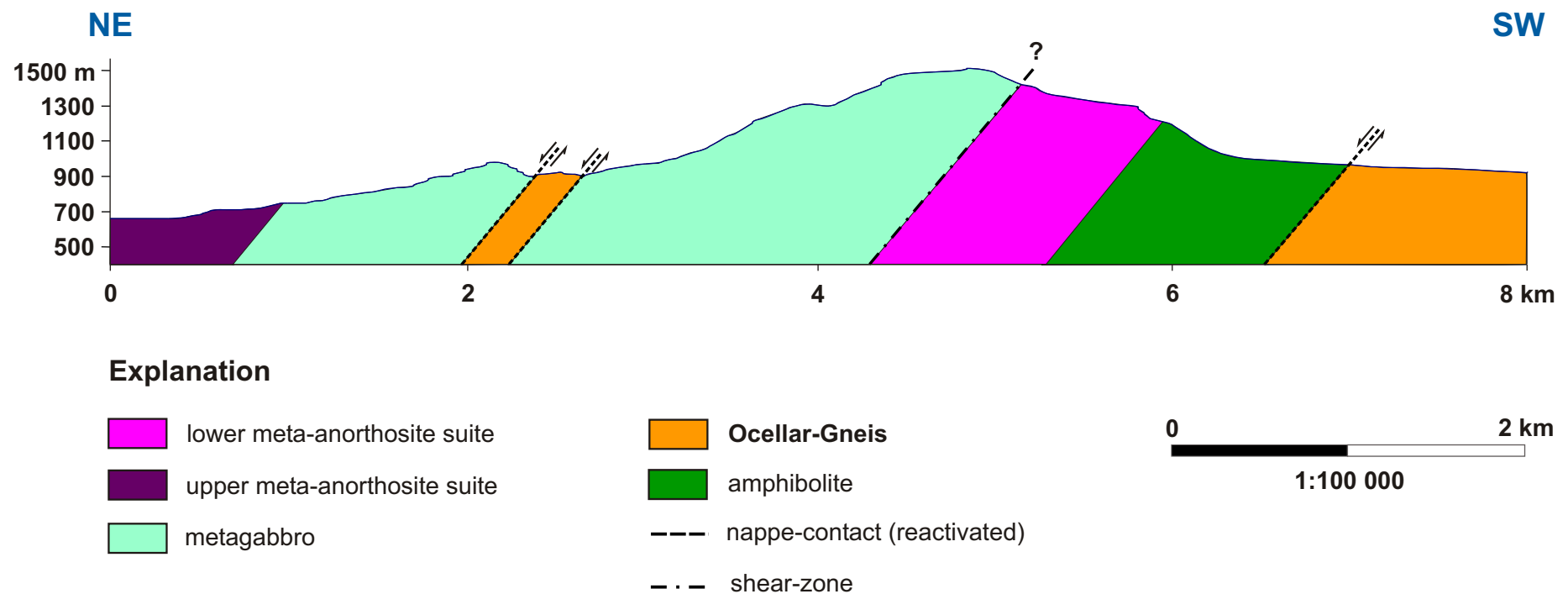
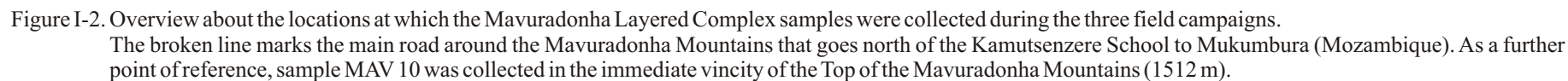


Figure I-1: Cross-section across the Mavuradonha Layered Complex as marked in Figure 2-1.



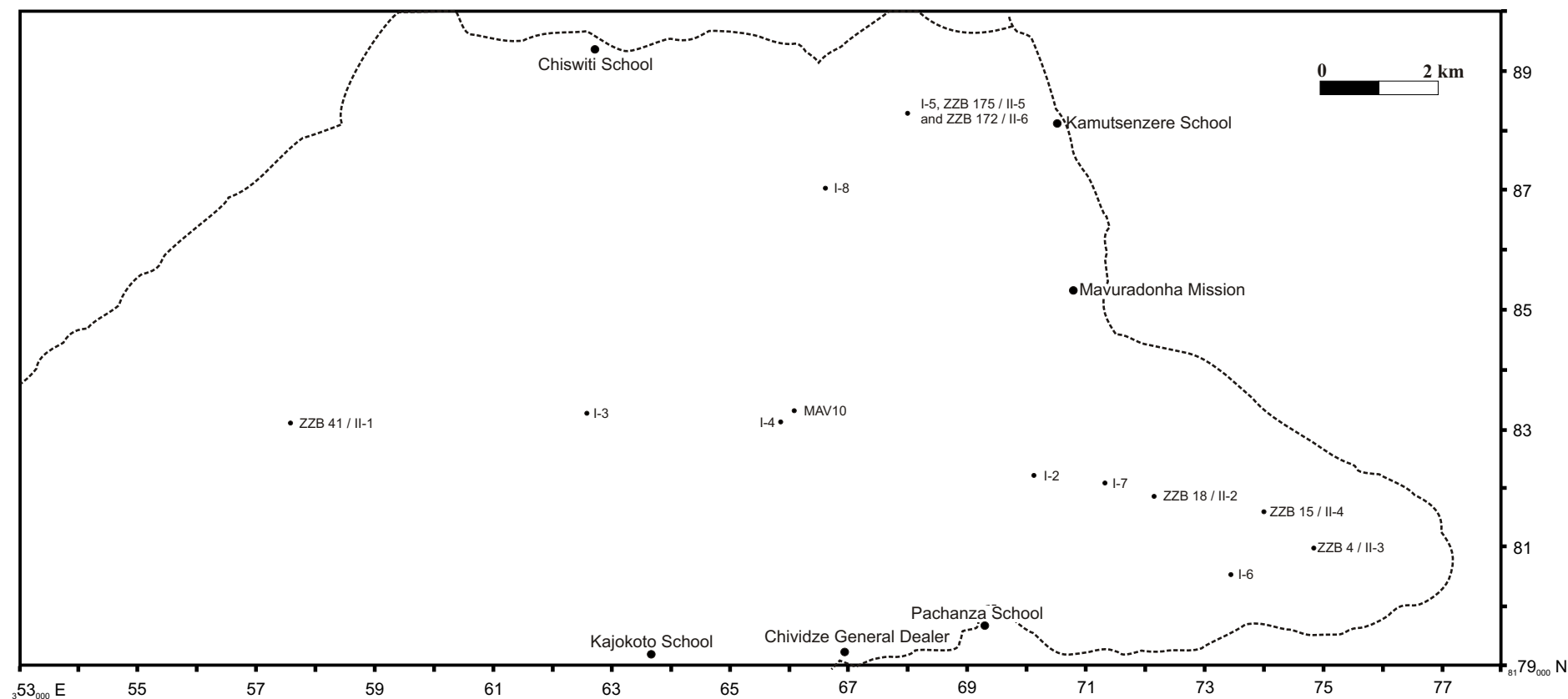


Figure I-3. Locations of samples that are shown in thin section images (plate II) and locations of photographs that are shown in plate I. The broken line marks the main road around the Mavuradonha Mountains that goes north of the Kamutsenzere School to Mukumbura (Mozambique). As a further point of reference, sample MAV 10 was collected in the immediate vicinity of the Top of the Mavuradonha Mountains (1512 m).

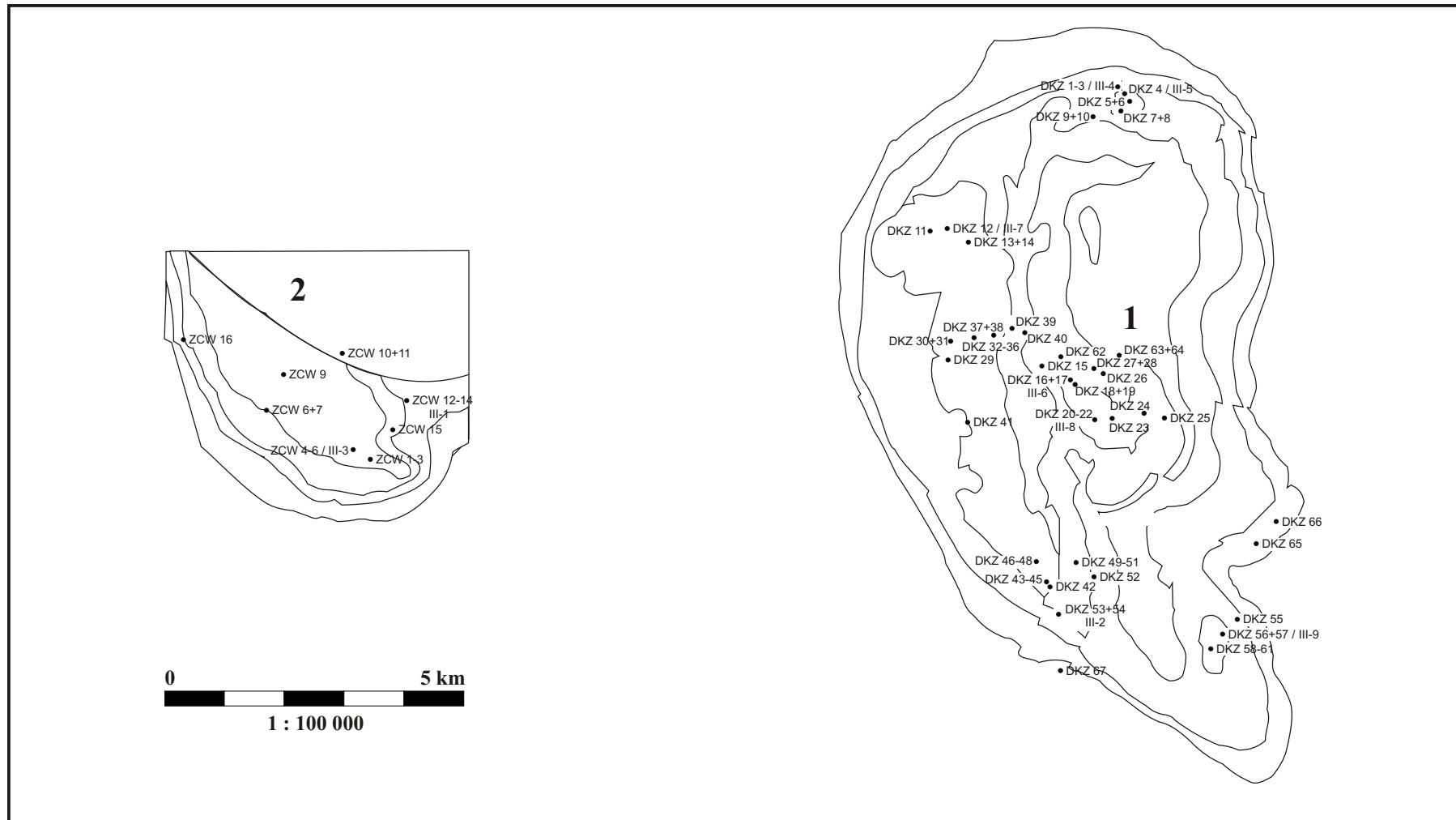


Figure I-4. Overview about the locations for samples of the Nyamhanda Inlier (1) and Chimwaya Hill Inlier (2). The contours of the different rock units are based on the geological map of Bache et al. (1983).

A-II Analytical methods

Table II-1. Accuracy of major elements analyzed by repeated measurements of an internal standard (mean of 10 times)

<i>element</i>	<i>x (wt. %)</i>	<i>s (wt. %)</i>	<i>s (% rel)</i>
SiO ₂	44.2	0.06	0.13
TiO ₂	2.17	0.003	0.16
Al ₂ O ₃	12.33	0.031	0.25
Fe ₂ O ₃	12.27	0.014	0.11
MnO	0.19	0.002	0.88
MgO	12.23	0.029	0.23
CaO	10.06	0.016	0.16
Na ₂ O	2.7	0.016	0.61
K ₂ O	1.08	0.005	0.45
P ₂ O ₅	0.62	0.003	0.48
Cr ₂ O ₃	0.063	0	0.54
NiO	0.043	0	0.56

x: concentration of each element; s: standard deviation

Table II-2. Accuracy of trace elements analyzed by repeated measurements of an internal standard (mean of 10 times)

<i>element</i>	<i>x (ppm)</i>	<i>s (ppm)</i>	<i>s (% rel)</i>	<i>dl (ppm)</i>
V	220	1.8	0.8	1.1
Cr	435	5.1	1.2	2.1
Co	57	2.3	3.9	1.9
Ni	345	2.9	0.8	1.1
Cu	60	2.2	3.7	0.4
Zn	105	1.5	1.3	1.8
Ga	18	0.7	3.9	1.4
Rb	44	1	2.4	1.3
Sr	755	2.5	0.3	1.4
Y	26	0.6	2.4	1.6
Zr	223	1.5	0.7	1.1
Nb	77	0.3	0.4	1
Cd	29	0.8	2.8	3.8
Sn	57	0.4	0.8	6.9
Sb	17	0.4	2.4	9.3
Ba	585	3.6	0.6	6.6

<i>element</i>	<i>x (ppm)</i>	<i>s (ppm)</i>	<i>s (% rel)</i>	<i>dl (ppm)</i>
La	278	1.9	0.7	3.6
Ce	406	3.7	0.9	9.1
Pr	39	1.9	4.9	2.5
Nd	109	2.1	2	5.1
Sm	8	1.8	22	6.9
Hf	24	1.2	5	2.5
Ta	23	1	4.3	3.2
Pb	30	0.6	2	2
Th	13	0.4	3.1	1
U	5	0.3	6	1.6

x: concentration of each element; s: standard deviation;
dl: detection limit

A-III Petrology and metamorphic evolution

Table III-1. Thin section description for samples from the Mavuradonha Layered Complex.

sample	rock type	Pl	Cpx	Opx	Grt	Am	Kfs	Bt	Ttn	Ap	Rt	Qtz	Zrn	Opaques	Am _c	Am _g	Am _{gb}	Am _b	Grt	Scp	Ep	Ttn	Qtz	Chl	Bt	Ms	Cal	Srp	Spl	
	primary mineral assemblage														secondary mineral assemblage															
ZZB 1	fgmg	X	X								X			X		X				X	X	X	X							
ZZB 2	cgmng	X	X								X			X		X			X	X	X		X							
ZZB 3	fgmg	X	X								X			X		X			X	X	X		X		X					
ZZB 4	cgmng	X	X	X							X			X	X	X				X	X		X		X					
ZZB 5	cgmng	X	X											X		X	X		X	X	X		X							
ZZB 6	cgmng	X	X								X			X		X	X		X	X	X	X	X							
ZZB 7	gtap	X			X	X				X	X	X		X						X	X	X								
ZZB 8	ap	X				X				X	X	X		X						X	X	X								
ZZB 9	ap	X				X				X	X	X		X						X	X									
ZZB 10	cgmng	X	X								X			X						X	X	X	X							
ZZB 12	cgmng	X	X	X							X			X	X	X			X	X	X		X							
ZZB 14	fgmg	X	X											X		X				X	X		X							
ZZB 15	cgmng	X	X								X			X		X				X	X	X	X			X				
ZZB 18	fgmg	X	X								X			X		X				X	X	X	X							
ZZB 27	fgmg	X	X								X	X		X		X					X	X	X							

ap = amphibolite; cgmng = coarse grained metagabbro; fgmg = fine grained metagabbro; fmg = ferro-metagabbro; ggns: granitoid gneiss; gtap = garnet-amphibolite; lmg = leuco-metagabbro; ma = meta-anorthosite; mg = metagabbro; mp = metapyroxenite; pgmatite = pegmatite; spt = serpentinite; am c = colourless amphibole; am g = green amphibole; am gb = greenish-blue amphibole; am b = brown amphibole.

Table III-1. Continued.

sample	rock type	Pl	Cpx	Opx	Grt	Am	Kfs	Bt	Ttn	Ap	Rt	Qtz	Zrn	Opaques	Am c	Am g	Am gb	Am b	Grt	Scp	Ep	Ttn	Qtz	Chl	Bt	Ms	Cal	Srp	Spl	
	primary mineral assemblage														secondary mineral assemblage															
ZZB 28	gtap	X			X	X				X	X	X	X	X						X	X	X								
ZZB 41	gtap	X			X	X				X	X	X		X						X	X	X								
ZZB 46	ma	X	X		X				X			X				X	X			X	X									
ZZB 55	spt													X	X													X		
ZZB 58	cgmg	X	X								X			X		X			X	X	X		X							
ZZB 60	pgt																													
ZZB 62	fmg	X	X		X				X			X				X	X			X	X									
ZZB 63	fmg	X	X		X					X	X	X		X			X					X		X	X					
ZZB 65	gtap	X			X	X				X	X	X		X						X	X	X								
ZZB 66	fgmg	X	X											X		X				X	X		X	X						
ZZB 67	cgmg	X									X			X		X				X	X	X	X							
ZZB 69	fgmg	X	X											X		X			X	X	X	X	X							
ZZB 70	fgmg	X	X											X		X			X	X	X	X		X						
ZZB 77	gtap	X			X	X				X		X		X						X	X	X	X							
ZZB 79	cgmg	X	X											X		X				X	X	X	X							

ap = amphibolite; cgmg = coarse grained metagabbro; fgmg = fine grained metagabbro; fmg = ferro-metagabbro; ggns: granitoid gneiss; gtap = garnet-amphibolite; lmg = leuco-metagabbro; ma = meta-anorthosite; mg = metagabbro; mp = metapyroxenite; pgt = pegmatite; spt = serpentinite. am c = colourless amphibole; am g = green amphibole; amph gb = greenish-blue amphibole; am b = brown amphibole.

Table III-1. Continued.

sample	rock type	Pl	Cpx	Opx	Grt	Am	Kfs	Bt	Ttn	Ap	Rt	Qtz	Zrn	Opaques	Am _c	Am _g	Am _{gb}	Am _b	Grt	Scp	Ep	Ttn	Qtz	Chl	Bt	Ms	Cal	Srp	Spl	
	primary mineral assemblage														secondary mineral assemblage															
ZZB 80	fmg	X	X											X		X				X	X	X								
ZZB 81	lmg	X												X		X					X	X	X							
ZZB 83	gtap	X			X	X				X	X	X		X						X	X	X								
ZZB 85	fmg	X												X		X	X			X	X	X								
ZZB 96	fmg	X	X											X		X	X			X	X	X								
ZZB 97	fmg	X	X		X					X	X	X		X		X	X			X	X	X								
ZZB 100	mg	X	X								X			X				X			X								X	
ZZB 101	mp		X											X		X														
ZZB 102	mg	X	X								X			X				X	X		X		X							
ZZB 105	mp		X								X			X		X				X		X					X			
ZZB 110	mg	X	X															X	X	X	X						X			
ZZB 115	fmg	X	X		X					X	X			X		X				X	X	X		X	X					
ZZB 122	fmg	X	X		X					X	X				X	X		X		X	X		X							
ZZB 123	cgm	X	X								X		X	X		X					X	X	X		X					
ZZB 124	lmg	X	X		X					X	X					X				X										

ap = amphibolite; cgm = coarse grained metagabbro; fmg = fine grained metagabbro; fmg = ferro-metagabbro; ggns: granitoid gneiss; gtap = garnet-amphibolite; lmg = leuco-metagabbro; ma = meta-anorthosite; mg = metagabbro; mp = metapyroxenite; pgt = pegmatite; spt = serpentinite; am c = colourless amphibole; am g = green amphibole; am gb = greenish-blue amphibole; am b = brown amphibole.

Table III-1. Continued.

sample	rock type	Pl	Cpx	Opx	Grt	Am	Kfs	Bt	Ttn	Ap	Rt	Qtz	Zrn	Opaques	Am _c	Am _g	Am _{gb}	Am _b	Grt	Scp	Ep	Ttn	Qtz	Chl	Bt	Ms	Cal	Srp	Spl	
	primary mineral assemblage														secondary mineral assemblage															
ZZB 125	ap	X				X				X	X	X		X						X	X	X								
ZZB 128	pgt																													
ZZB 135	fgmg	X								X	X			X		X					X	X	X			X				
ZZB 138	fgmg	X	X							X	X		X	X		X					X	X	X							
ZZB 143	fmg	X	X		X					X	X			X		X				X	X		X		X					
ZZB 145	fmg	X	X		X					X	X	X				X				X	X		X							
ZZB 147	fmg	X	X		X					X	X	X		X		X				X	X		X							
ZZB 149	fmg	X	X		X					X	X	X		X		X				X	X	X	X		X					
ZZB 152	ma	X	X		X					X	X					X	X			X	X		X		X	X				
ZZB 153	fmg	X	X		X						X	X		X		X				X	X	X	X		X					
ZZB 154	ma	X			X					X	X			X		X		X		X	X		X							
ZZB 160	ma	X				X			X	X							X				X					X				
ZZB 161	gtap	X				X				X	X	X	X	X						X	X	X				X				
ZZB 162	ma	X				X			X	X							X				X			X	X	X				
ZZB 166	gtap	X			X	X				X	X	X	X	X					X	X	X	X								
ZZB 167	fmg	X	X		X					X	X	X		X		X	X			X	X		X							

ap = amphibolite; cgmg = coarse grained metagabbro; fgmg = fine grained metagabbro; fmg = ferro-metagabbro; ggns: granitoid gneiss; gtap = garnet-amphibolite; lmg = leuco-metagabbro; ma = meta-anorthosite; mg = metagabbro; mp = metapyroxenite; pgmt = pegmatite; spt = serpentinite; am c = colourless amphibole; am g = green amphibole; amph gb = greenish-blue amphibole; am b = brown amphibole.

Table III-1. Continued.

sample	rock type	Pl	Cpx	Opx	Grt	Am	Kfs	Bt	Ttn	Ap	Rt	Qtz	Zrn	Opaques	Am _c	Am _g	Am _{gb}	Am _b	Grt	Scp	Ep	Ttn	Qtz	Chl	Bt	Ms	Cal	Srp	Spl	
	primary mineral assemblage														secondary mineral assemblage															
ZZB 172	ma	X	X		X					X						X				X	X		X			X				
ZZB 173	fmg	X	X		X					X	X	X		X		X	X			X	X	X	X		X					
ZZB 175	fmg	X	X		X					X	X	X		X		X				X	X		X							
ZZB 176	ma	X	X		X					X						X				X	X		X		X					
ZZB 179	fmg	X	X		X					X	X	X		X		X	X			X	X		X							
ZZB 183	fmg	X	X		X					X	X	X		X		X				X	X	X	X		X					
ZZB 193	ap	X			X					X	X	X		X		X				X	X	X	X		X					
ZZB 195	mg	X	X		X					X	X	X				X				X	X		X							
ZZB 196	mp	X	X													X							X				X			
ZZB 197	lmg	X	X		X					X	X			X	X	X				X	X	X	X		X					
ZZB 199	ggns	X				X	X	X	X	X		X	X				X			X	X									
ZZB 200	ggns	X				X	X	X	X	X		X	X				X			X	X									
ZZB 201	ap	X				X				X	X	X		X						X	X	X								
ZZB 202	fgmg	X	X								X			X		X				X	X	X	X							
ZZB 203	fmg	X	X		X					X	X	X		X		X				X	X		X		X					
ZZB 215	cgmng	X	X	X							X				X	X				X	X		X							
ZZB 216	cgmng	X	X								X			X	X	X			X	X	X		X							
ZZB 219	cgmng	X	X	X	X						X			X	X	X				X	X		X							

ap = amphibolite; cgmng = coarse grained metagabbro; fgmg = fine grained metagabbro; fmg = ferro-metagabbro; ggns: granitoid gneiss; gtap = garnet-amphibolite; lmg = leuco-metagabbro; ma = meta-anorthosite; mg = metagabbro; mp = metapyroxenite; pgt = pegmatite; spt = serpentinite; am c = colourless amphibole; am g = green amphibole; amph gb = greenish-blue amphibole; am b = brown amphibole.

Table III-1. Continued.

sample	rock type	Pl	Cpx	Opx	Grt	Am	Kfs	Bt	Ttn	Ap	Rt	Qtz	Zrn	Opaques	Am _c	Am _g	Am _{gb}	Am _b	Grt	Scp	Ep	Ttn	Qtz	Chl	Bt	Ms	Cal	Srp	Spl
		primary mineral assemblage														secondary mineral assemblage													
ZZB 220	cgmg	X	X	X							X			X	X	X				X	X		X						
MAV 1	mp	X				X								X															
ZIM 30	ggns	X				X	X	X		X		X	X	X							X	X			X				
ZIM 55	ggns	X			X	X	X	X	X	X		X	X	X							X	X			X				
ZIM 56	ggns	X				X	X	X		X		X	X	X							X	X			X				

ap = amphibolite; cgmg = coarse grained metagabbro; fgmg = fine grained metagabbro; fmg = ferro-metagabbro; ggns: granitoid gneiss; gtap = garnet-amphibolite; lmg = leuco-metagabbro; ma = meta-anorthosite; mg = metagabbro; mp = metapyroxenite; pgt = pegmatite; spt = serpentinite; am c = colourless amphibole; am g = green amphibole; amph gb = greenish-blue amphibole; am b = brown amphibole.

Table III-2. Thin section description for samples from the Chimwaya Hill Inlier.

<i>sample</i>	<i>rock type</i>	<i>Cpx</i>	<i>Pl</i>	<i>Ol</i>	<i>Grt</i>	<i>Am</i>	<i>Kfs</i>	<i>Bt</i>	<i>Ttn</i>	<i>Ap</i>	<i>Rt</i>	<i>Qtz</i>	<i>Zrn</i>	<i>Opaques</i>	<i>Am c</i>	<i>Am gb</i>	<i>Grt</i>	<i>Scp</i>	<i>Ep</i>	<i>Ttn</i>	<i>Qtz</i>	<i>Ms</i>	<i>Chl</i>
		primary mineral assemblage														secondary mineral assemblage							
ZCW 2	cgmg	X	X	X							X			X	X			X		X	X		X
ZCW 3	fgmg	X	X								X			X	X			X	X	X	X		
ZCW 6	cgmg	X	X	X							X			X	X			X		X	X		X

ap = amphibolite; cgmg = coarse grained metagabbro; fgmg = fine grained metagabbro; ggns: granitoid gneiss; gtap = garnet-amphibolite; am g = green amphibole; amph gb = greenish-blue amphibole.

Table III-2. Continued.

<i>sample</i>	<i>rock type</i>	<i>Cpx</i>	<i>Pl</i>	<i>Ol</i>	<i>Grt</i>	<i>Am</i>	<i>Kfs</i>	<i>Bt</i>	<i>Ttn</i>	<i>Ap</i>	<i>Rt</i>	<i>Qtz</i>	<i>Zrn</i>	<i>Opaques</i>	<i>Am c</i>	<i>Am gb</i>	<i>Grt</i>	<i>Scp</i>	<i>Ep</i>	<i>Ttn</i>	<i>Qtz</i>	<i>Ms</i>	<i>Chl</i>
	primary mineral assemblage														secondary mineral assemblage								
ZCW 8	ap		X			X				X	X	X		X				X	X	X			X
ZCW 10	ggns		X			X	X	X	X	X		X	X			X		X	X				
ZCW 11	ggns		X			X	X	X	X	X		X	X			X		X	X				
ZCW 12	ggns		X			X	X	X	X	X		X	X			X		X	X				
ZCW 14	gtap		X		X	X				X	X	X		X				X	X	X			

ap = amphibolite; cgmng = coarse grained metagabbro; fgmg = fine grained metagabbro; ggns: granitoid gneiss; gtap = garnet-amphibolite; am g = green amphibole; amph gb = greenish-blue amphibole.

Table III-3. Thin section description for samples from the Nyamhanda Inlier.

sample	rock type	Cpx	Pl	Opx	Grt	Am	Kfs	Bt	Ttn	Ap	Rt	Qtz	Zrn	Opaques	Am c	Am g	Am b	Grt	Scp	Ep	Ttn	Qtz	Bt	Ms	Cal	Spl	Chl	
	primary mineral assemblage														secondary mineral assemblage													
DKZ 1	mg	X	X								X			X	X	X			X	X		X						
DKZ 2	ap		X			X			X	X			X	X		X	X		X	X					X			
DKZ 4	mg	X	X	X										X		X		X		X		X	X			X		
DKZ 5	mg	X	X								X			X		X		X		X								
DKZ 6	mg	X	X								X			X		X		X		X	X							

ap = amphibolite; ggns: granitoid gneiss; gtap = garnet-amphibolite; lmg = leuco-metagabbro; mg = metagabbro; am c = colourless amphibole; am g = green amphibole; am b = brown amphibole.

Table III-3. Continued.

sample	rock type	Cpx	Pl	Opx	Grt	Am	Kfs	Bt	Ttn	Ap	Rt	Qtz	Zrn	Opaques	Am c	Am g	Am b	Grt	Scp	Ep	Ttn	Qtz	Bt	Ms	Cal	Spl	Chl	
	primary mineral assemblage														secondary mineral assemblage													
DKZ 11	gtap		X		X	X				X	X	X		X					X	X	X	X	X					
DKZ 12	mg	X	X								X			X	X	X	X			X			X			X		
DKZ 13	mg	X	X	X										X		X				X		X						
DKZ 16	gtap	X	X		X	X				X				X					X	X	X	X						
DKZ 18	lmg		X									X		X						X		X		X				
DKZ 20	gtap		X		X	X			X	X	X	X		X			X		X	X								
DKZ 22	mg	X	X	X										X	X			X		X		X	X					
DKZ 25	ap		X			X			X	X	X	X		X			X		X	X								
DKZ 41	ggns		X		X	X	X		X	X		X	X	X					X	X								
DKZ 42	ggns		X			X	X		X	X		X	X						X	X								
DKZ 53	gtap		X		X	X				X	X	X	X	X					X	X	X							
DKZ 54	gtap		X		X	X				X	X	X	X	X					X	X	X							
DKZ 55	ggns		X			X	X	X	X			X		X						X				X				
DKZ 57	mg	X	X							X	X			X	X	X	X	X		X			X			X		
DKZ 60	mg	X	X	X							X			X		X		X		X		X						
DKZ 62	gtap	X	X		X	X				X				X					X	X	X	X						

ap = amphibolite; ggns: granitoid gneiss; gtap = garnet-amphibolite; lmg = leuco-metagabbro; mg = metagabbro; am c = colourless amphibole; am g = green amphibole; am b = brown amphibole.

Table III-4. Representative microprobe analyzes of clinopyroxenes from samples from the Mavuradonha Layered Complex.

sample	ZZB 2	ZZB 2	ZZB 2	ZZB 2	ZZB 2	ZZB 2	sample	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	sample	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2
analysis no.	173	125	136	152	164	185	analysis no.	3	64	122	42	60	107	analysis no.	189	195	173	149	152	176
SiO ₂	53.22	54.11	53.57	53.37	53.11	53.26	SiO ₂	53.82	54.01	53.61	53.98	53.94	54.23	SiO ₂	53.57	53.80	53.92	52.79	52.70	53.81
TiO ₂	0.73	0.16	0.17	0.16	0.14	0.21	TiO ₂	0.13	0.14	0.24	0.10	0.09	0.03	TiO ₂	0.10	0.12	0.11	0.23	0.25	0.05
Al ₂ O ₃	2.92	2.23	3.29	2.90	2.95	2.20	Al ₂ O ₃	2.58	1.72	2.46	2.25	1.85	1.62	Al ₂ O ₃	2.22	2.47	2.16	3.28	3.97	1.60
FeO	3.61	3.96	3.15	3.23	2.85	3.14	FeO	4.37	4.73	4.49	4.28	4.99	4.19	FeO	4.90	4.94	4.81	5.34	5.41	5.41
MnO	0.02	0.13	0.11	0.05	0.05	0.05	MnO	0.20	0.22	0.18	0.19	0.24	0.22	MnO	0.07	0.09	0.13	0.11	0.13	0.12
MgO	16.74	15.90	15.56	15.82	16.43	16.41	MgO	15.48	15.78	15.36	15.83	15.74	16.07	MgO	14.73	14.61	14.72	14.20	13.92	15.22
CaO	21.45	23.33	23.69	23.84	23.98	24.30	CaO	22.88	23.54	23.30	23.44	23.05	23.75	CaO	23.91	23.50	23.79	23.14	23.26	22.97
Na ₂ O	0.70	0.78	0.83	0.70	0.61	0.55	Na ₂ O	0.80	0.71	0.83	0.83	0.68	0.66	Na ₂ O	0.73	0.87	0.71	0.87	0.98	0.82
Cr ₂ O ₃	0	0	0	0	0	0	Cr ₂ O ₃	0	0	0	0	0	0	Cr ₂ O ₃	0	0	0	0	0	0
Total	99.42	100.60	100.38	100.07	100.13	100.13	Total	100.28	101.89	100.56	100.91	100.60	100.78	Total	100.25	100.42	100.37	99.96	100.61	100.01
Si	1.94	1.96	1.94	1.94	1.92	1.93	Si	1.96	1.96	1.95	1.95	1.96	1.96	Si	1.96	1.96	1.97	1.94	1.92	1.97
Al(IV)	0.06	0.04	0.06	0.06	0.08	0.07	Al(IV)	0.04	0.04	0.05	0.05	0.04	0.04	Al(IV)	0.04	0.04	0.03	0.06	0.08	0.03
Al(VI)	0.07	0.05	0.08	0.06	0.05	0.03	Al(VI)	0.07	0.03	0.06	0.05	0.04	0.03	Al(VI)	0.05	0.07	0.06	0.08	0.09	0.04
Fe ²⁺	0.09	0.08	0.06	0.05	0.02	0.02	Fe ²⁺	0.11	0.08	0.09	0.07	0.11	0.08	Fe ²⁺	0.11	0.12	0.13	0.12	0.11	0.12
Fe ³⁺	0.02	0.04	0.03	0.04	0.07	0.07	Fe ³⁺	0.03	0.06	0.04	0.06	0.05	0.05	Fe ³⁺	0.04	0.03	0.02	0.04	0.05	0.05
Mg	0.99	0.86	0.84	0.85	0.89	0.88	Mg	0.84	0.85	0.83	0.85	0.85	0.87	Mg	0.80	0.80	0.80	0.78	0.76	0.83
Ti	0.02	0	0	0	0	0.01	Ti	0	0	0	0	0	0	Ti	0	0	0	0.01	0.01	0
Cr	0	0	0	0	0	0	Cr	0	0	0	0	0	0	Cr	0	0	0	0	0	0
Na	0.05	0.05	0.06	0.05	0.04	0.04	Na	0.06	0.05	0.06	0.06	0.05	0.05	Na	0.05	0.06	0.05	0.06	0.07	0.06
Ca	0.84	0.90	0.92	0.93	0.93	0.94	Ca	0.89	0.91	0.91	0.91	0.90	0.92	Ca	0.94	0.92	0.93	0.91	0.91	0.90
Jd%	4.03	3.32	4.19	2.99	1.86	1.05	Jd%	4.20	1.68	3.42	2.54	2.29	1.79	Jd%	3.13	4.48	4.11	4.22	4.63	2.68
Di%	86.19	86.10	87.66	89.26	93.58	93.74	Di%	83.57	86.48	84.57	87.04	84.56	87.66	Di%	83.07	81.13	81.49	80.77	80.64	82.35
Hd%	8.72	8.28	6.30	5.65	2.00	2.27	Hd%	10.63	8.39	9.46	6	6.98	22.87	Hd%	11.61	12.60	13.40	12.83	12.13	11.71
Ae%	1.06	2.29	1.84	2.10	2.57	2.93	Ae%	0.75	1.67	2.56	6.98	10.52	7.60	Ae%	2.19	1.79	1.00	2.18	2.61	3.25

Structural formulas are calculated on the basis of 4 cations. Endmembers are calculated after Morimoto et al. (1988).

Table III-4. Continued.

sample	ZZB 12	ZZB 12	ZZB 12	ZZB 12	ZZB 12	ZZB 12	sample	ZZB 15	ZZB 15	ZZB 15	ZZB 15	ZZB 15	ZZB 15	sample	ZZB 18	ZZB 18	ZZB 18	ZZB 18	ZZB 18	ZZB 18
analysis-no.	96	29	84	15	2	119	analysis-no.	2	42	30	40	62	11	analysis-no.	66	1	23	37	54	45
SiO ₂	51.47	60.46	50.83	53.44	53.75	53.02	SiO ₂	53.55	53.62	53.65	53.75	53.46	53.77	SiO ₂	53.82	53.39	53.76	53.74	53.50	52.22
TiO ₂	0.33	0.14	3.81	0.24	0.18	1.97	TiO ₂	0.12	0.14	0.19	0.15	0.15	0.18	TiO ₂	0.05	0.03	0.06	0.15	0.10	0.14
Al ₂ O ₃	4.89	2.78	3.38	3.54	2.97	2.95	Al ₂ O ₃	2.71	2.55	2.47	2.37	2.92	1.94	Al ₂ O ₃	1.42	1.46	1.53	1.87	2.34	2.73
FeO	5.99	2.73	5.22	3.99	3.32	3.12	FeO	4.63	5.01	4.68	4.65	4.55	4.62	FeO	4.04	3.89	4.25	4.17	4.29	4.32
MnO	0.08	0.06	0.16	0.02	0.08	0.10	MnO	0.17	0.15	0.16	0.25	0.21	0.16	MnO	0.10	0.13	0.12	0.10	0.14	0.08
MgO	15.77	13.50	14.11	15.47	15.89	15.57	MgO	14.99	15.02	15.09	15.01	14.92	15.06	MgO	15.97	15.71	15.70	15.44	15.26	14.91
CaO	19.91	20.44	21.47	22.51	23.56	24.11	CaO	22.85	22.55	23.13	22.81	23.23	23.56	CaO	24.22	24.35	24.04	24.31	23.97	23.70
Na ₂ O	0.81	0.66	0.91	1.04	0.74	0.58	Na ₂ O	0.88	0.88	0.88	0.85	0.88	0.80	Na ₂ O	0.32	0.57	0.55	0.62	0.66	0.76
Cr ₂ O ₃	0	0	0	0	0	0	Cr ₂ O ₃	0	0	0	0	0	0	Cr ₂ O ₃	0.21	0.10	0.16	0.29	0.22	0.22
Total	99.32	100.77	99.91	100.28	100.51	101.43	Total	99.89	99.92	100.25	99.86	100.32	100.10	Total	100.15	99.63	100.19	100.69	100.47	99.08
Si	1.89	2.21	1.88	1.94	1.94	1.91	Si	1.96	1.96	1.96	1.97	1.95	1.97	Si	1.97	1.96	1.96	1.95	1.95	1.93
Al(IV)	0.11	0	0.12	0.06	0.06	0.09	Al(IV)	0.04	0.04	0.04	0.03	0.05	0.03	Al(IV)	0.03	0.04	0.04	0.05	0.05	0.07
Al(VI)	0.10	0.12	0.03	0.09	0.07	0.04	Al(VI)	0.08	0.07	0.06	0.07	0.07	0.05	Al(VI)	0.03	0.02	0.03	0.03	0.05	0.04
Fe ²⁺	0.12	0.08	0.12	0.08	0.07	0.06	Fe ²⁺	0.12	0.13	0.10	0.12	0.10	0.10	Fe ²⁺	0.09	0.05	0.08	0.07	0.08	0.05
Fe ³⁺	0.06	0	0.05	0.04	0.03	0.03	Fe ³⁺	0.02	0.02	0.04	0.02	0.04	0.04	Fe ³⁺	0.03	0.06	0.05	0.05	0.05	0.08
Mg	0.85	0.74	0.78	0.83	0.85	0.83	Mg	0.82	0.83	0.82	0.82	0.81	0.82	Mg	0.87	0.86	0.86	0.84	0.82	0.82
Ti	0.01	0	0.11	0.01	0.01	0.05	Ti	0	0	0.01	0	0	0	Ti	0	0	0	0	0	0
Cr	0	0	0	0	0	0	Cr	0	0	0	0	0	0	Cr	0.01	0	0	0.01	0.01	0.01
Na	0.06	0.05	0.07	0.07	0.05	0.04	Na	0.06	0.06	0.06	0.06	0.06	0.06	Na	0.02	0.04	0.04	0.04	0.05	0.05
Ca	0.78	0.80	0.85	0.87	0.91	0.93	Ca	0.90	0.88	0.90	0.89	0.91	0.92	Ca	0.95	0.96	0.94	0.95	0.93	0.94
Jd%	3.79	5.49	2.75	5.13	3.66	2.24	Jd%	4.79	4.78	3.90	4.94	4.18	3.27	Jd%	1.11	0.94	1.43	1.66	2.36	2.00
Di%	82.21	84.88	81.04	84.44	87.70	89.43	Di%	81.93	80.88	83.13	81.43	83.30	83.65	Di%	88.14	90.13	87.71	87.97	86.67	88.70
Hd%	11.65	9.63	12.01	7.99	6.97	6.32	Hd%	11.72	12.73	10.52	12.41	10.32	10.57	Hd%	9.53	5.74	8.29	7.56	8.54	5.62
Ae%	2.35	0	4.21	2.44	1.68	2.01	Ae%	1.57	1.61	2.45	1.23	2.21	2.51	Ae%	1.22	3.19	2.57	2.81	2.43	3.69

Structural formulas are calculated on the basis of 4 cations. Endmembers are calculated after Morimoto et al. (1988).

Table III-4. Continued.

sample	ZZB 41	ZZB 41	ZZB 41	ZZB 41	ZZB 41	ZZB 41	sample	ZZB 58-3	ZZB 58-3	ZZB 58-3	ZZB 58-3	ZZB 58-3	ZZB 58-3	sample	ZZB 101	ZZB 101	ZZB 101	ZZB 101	ZZB 101	ZZB 101
analysis-no.	116	81	70	86	71	95	analysis-no.	127	120	142	115	118	139	analysis-no.	13	18	27	69	74	82
SiO ₂	51.26	51.68	52.68	52.13	51.65	51.68	SiO ₂	54.08	53.84	53.78	53.60	52.44	53.24	SiO ₂	47.90	48.00	47.44	47.08	47.73	48.13
TiO ₂	0.26	0.21	0.16	0.23	0.16	0.29	TiO ₂	0.19	0.25	0.10	0.31	0.50	0.30	TiO ₂	1.21	1.15	1.22	1.17	1.25	1.38
Al ₂ O ₃	4.81	3.86	2.55	4.18	4.21	4.38	Al ₂ O ₃	2.43	2.89	1.83	3.08	3.19	3.18	Al ₂ O ₃	8.27	8.49	8.35	8.46	8.21	7.81
FeO	8.82	8.70	6.86	8.60	8.96	8.56	FeO	4.02	4.24	4.58	4.67	5.05	5.28	FeO	7.97	7.96	8.18	7.76	7.56	6.84
MnO	0.01	0.09	0.06	0.06	0.11	0.04	MnO	0.09	0.08	0.15	0.21	0.17	0.16	MnO	0.23	0.21	0.19	0.21	0.24	0.18
MgO	11.28	11.88	13.65	11.86	11.89	11.81	MgO	15.19	15.09	15.01	14.84	15.15	14.49	MgO	11.20	11.20	10.96	11.15	11.17	11.43
CaO	21.44	22.12	23.30	22.04	21.44	22.23	CaO	24.37	23.74	24.64	23.70	22.43	23.71	CaO	22.48	22.10	22.15	22.14	22.13	23.13
Na ₂ O	1.81	1.40	0.96	1.52	1.61	1.47	Na ₂ O	0.64	0.76	0.45	0.71	0.60	0.75	Na ₂ O	1.13	1.23	1.23	1.27	1.27	1.09
Cr ₂ O ₃	0.28	0.09	0	0.06	0.01	0.03	Cr ₂ O ₃	0.27	0.06	0.05	0.03	0.28	0.03	Cr ₂ O ₃	0.01	0.03	0.07	0.01	0.04	0.01
Total	99.99	100.04	100.23	100.69	100.04	100.49	Total	101.28	100.96	100.59	101.15	99.86	101.15	Total	100.42	100.36	99.81	99.26	99.60	100.03
Si	1.90	1.92	1.94	1.92	1.91	1.90	Si	1.95	1.95	1.96	1.94	1.92	1.93	Si	1.77	1.77	1.76	1.75	1.77	1.78
Al(IV)	0.10	0.08	0.06	0.08	0.09	0.10	Al(IV)	0.05	0.05	0.04	0.06	0.08	0.07	Al(IV)	0.23	0.23	0.24	0.25	0.23	0.22
Al(VI)	0.11	0.08	0.05	0.10	0.09	0.09	Al(VI)	0.06	0.07	0.04	0.07	0.06	0.07	Al(VI)	0.13	0.14	0.13	0.12	0.13	0.12
Fe ²⁺	0.16	0.17	0.13	0.18	0.17	0.17	Fe ²⁺	0.09	0.10	0.11	0.11	0.11	0.11	Fe ²⁺	0.09	0.10	0.09	0.06	0.08	0.07
Fe ³⁺	0.12	0.10	0.08	0.09	0.11	0.10	Fe ³⁺	0.03	0.03	0.03	0.03	0.04	0.05	Fe ³⁺	0.15	0.15	0.17	0.18	0.15	0.14
Mg	0.62	0.66	0.75	0.65	0.65	0.65	Mg	0.82	0.81	0.82	0.80	0.83	0.78	Mg	0.61	0.61	0.60	0.62	0.62	0.63
Ti	0.01	0.01	0	0.01	0	0.01	Ti	0.01	0.01	0	0.01	0.01	0.01	Ti	0.03	0.03	0.03	0.03	0.03	0.04
Cr	0.01	0	0	0	0	0	Cr	0.01	0	0	0	0.01	0.00	Cr	0	0	0	0	0	0
Na	0.13	0.10	0.07	0.11	0.12	0.11	Na	0.05	0.05	0.03	0.05	0.04	0.05	Na	0.08	0.09	0.09	0.09	0.09	0.08
Ca	0.85	0.88	0.92	0.87	0.85	0.88	Ca	0.94	0.92	0.96	0.92	0.88	0.92	Ca	0.89	0.87	0.88	0.88	0.88	0.92
Jd%	6.59	4.91	2.67	6.03	5.68	5.43	Jd%	3.06	4.05	1.97	3.73	2.68	3.28	Jd%	4.18	4.84	4.38	4.27	4.79	4.04
Di%	68.97	70.74	78.98	69.61	69.72	70.83	Di%	85.71	83.92	84.97	83.04	84.03	82.45	Di%	78.92	77.61	78.56	81.54	79.16	82.06
Hd%	17.27	18.72	13.97	19.06	18.17	18.13	Hd%	9.66	10.58	11.76	11.83	11.51	12.06	Hd%	11.86	12.44	11.31	7.97	10.47	9.09
Ae%	7.17	5.63	4.38	5.30	6.43	5.61	Ae%	1.57	1.45	1.30	1.40	1.78	2.22	Ae%	5.04	5.11	5.75	6.22	5.57	4.80

Structural formulas are calculated on the basis of 4 cations. Endmembers are calculated after Morimoto et al. (1988).

Table III-4. Continued.

sample	ZZB 105	ZZB 105	ZZB 105	ZZB 105	ZZB 105	ZZB 105	sample	ZZB 145	ZZB 145	ZZB 145	ZZB 145	ZZB 145	ZZB 145	sample	ZZB 147	ZZB 147	ZZB 147	ZZB 147	ZZB 147	ZZB 147
analysis- no.	43	42	45	33	38	46	analysis- no.	15	35	42	20	9	2	analysis- no.	94	130	145	100	118	106
SiO ₂	50.99	49.60	50.75	50.27	50.48	49.90	SiO ₂	51.18	50.93	50.16	51.36	50.03	49.60	SiO ₂	51.41	51.17	51.14	50.95	50.89	50.72
TiO ₂	0.48	0.38	0.72	0.78	0.97	0.69	TiO ₂	0.64	0.35	0.65	0.41	0.67	0.68	TiO ₂	0.40	0.55	0.47	0.36	0.48	0.58
Al ₂ O ₃	4.33	4.71	5.12	5.18	5.45	6.23	Al ₂ O ₃	5.22	4.92	6.33	4.74	6.44	6.71	Al ₂ O ₃	4.52	4.73	4.89	4.73	4.80	4.97
FeO	10.69	10.71	10.08	9.53	9.15	9.40	FeO	7.76	8.15	8.17	8.48	8.82	9.35	FeO	8.31	8.58	8.69	8.89	9.01	9.39
MnO	0.17	0.15	0.15	0.14	0.13	0.16	MnO	0	0.08	0.06	0.08	0.04	0.03	MnO	0.06	0.09	0.06	0.09	0.08	0.10
MgO	10.20	9.83	10.91	11.22	11.04	10.53	MgO	12.40	12.03	11.64	12.26	11.64	11.24	MgO	12.21	11.52	11.78	11.84	11.71	11.64
CaO	21.63	21.27	21.43	21.75	21.79	22.07	CaO	21.85	21.56	20.74	21.49	20.73	20.90	CaO	21.02	21.28	21.05	21.04	20.95	20.92
Na ₂ O	1.54	1.71	1.57	1.47	1.40	1.56	Na ₂ O	1.08	1.23	1.36	1.23	1.29	1.32	Na ₂ O	1.30	1.47	1.36	1.29	1.46	1.25
Cr ₂ O ₃	0.06	1.12	0.13	0	0.09	0.01	Cr ₂ O ₃	0.07	0.01	0.07	0.03	0.17	0.03	Cr ₂ O ₃	0.05	0.03	0.07	0.10	0.12	0.02
Total	100.21	99.98	99.28	99.18	99.83	100.06	Total	100.21	99.28	99.18	100.09	99.84	99.85	Total	99.28	99.41	99.51	99.30	99.49	99.59
Si	1.91	1.87	1.88	1.86	1.87	1.85	Si	1.89	1.90	1.87	1.90	1.86	1.84	Si	1.92	1.91	1.91	1.90	1.90	1.89
Al(IV)	0.09	0.13	0.12	0.14	0.13	0.15	Al(IV)	0.11	0.10	0.13	0.10	0.14	0.16	Al(IV)	0.08	0.09	0.09	0.10	0.10	0.11
Al(VI)	0.10	0.08	0.10	0.09	0.11	0.12	Al(VI)	0.12	0.11	0.15	0.11	0.14	0.14	Al(VI)	0.11	0.12	0.12	0.11	0.11	0.11
Fe ²⁺	0.24	0.17	0.19	0.16	0.19	0.16	Fe ²⁺	0.19	0.19	0.19	0.19	0.20	0.20	Fe ²⁺	0.21	0.20	0.21	0.21	0.19	0.22
Fe ³⁺	0.09	0.17	0.12	0.13	0.10	0.13	Fe ³⁺	0.05	0.07	0.06	0.07	0.08	0.09	Fe ³⁺	0.05	0.07	0.06	0.07	0.09	0.07
Mg	0.57	0.55	0.60	0.62	0.61	0.58	Mg	0.68	0.67	0.65	0.68	0.64	0.62	Mg	0.67	0.64	0.65	0.66	0.67	0.65
Ti	0.01	0.01	0.02	0.02	0.03	0.02	Ti	0.02	0.01	0.02	0.01	0.02	0.02	Ti	0.01	0.02	0.01	0.01	0.01	0.02
Cr	0	0.03	0	0	0	0	Cr	0	0	0	0	0.01	0	Cr	0	0	0	0	0	0
Na	0.11	0.12	0.11	0.11	0.10	0.11	Na	0.08	0.09	0.10	0.09	0.09	0.10	Na	0.09	0.11	0.10	0.09	0.11	0.09
Ca	0.87	0.86	0.85	0.86	0.86	0.87	Ca	0.86	0.86	0.83	0.85	0.82	0.83	Ca	0.84	0.85	0.84	0.84	0.84	0.84
Jd%	6.03	4.23	5.46	4.60	5.72	5.80	Jd%	5.61	5.95	7.50	5.61	6.45	6.15	Jd%	6.77	7.13	6.88	6.08	6.14	5.94
Di%	61.95	66.28	66.59	70.09	68.20	68.71	Di%	72.14	70.73	68.79	70.61	68.95	68.13	Di%	69.07	67.56	67.82	68.49	68.52	67.15
Hd%	26.30	20.05	21.36	18.55	21.01	19.13	Hd%	19.67	19.89	20.67	20.11	21.01	21.55	Hd%	21.10	21.28	21.85	21.63	20.32	23.27
Ae%	5.72	9.44	6.59	6.76	5.07	6.36	Ae%	2.57	3.43	3.04	3.67	3.59	4.16	Ae%	3.06	4.03	3.45	3.80	5.02	3.64

Structural formulas are calculated on the basis of 4 cations. Endmembers are calculated after Morimoto et al. (1988).

Table III-4. Continued.

<i>sample</i>	<i>ZZB</i> <i>152</i>	<i>ZZB</i> <i>152</i>	<i>ZZB</i> <i>152</i>	<i>ZZB</i> <i>152</i>	<i>ZZB</i> <i>152</i>	<i>ZZB</i> <i>152</i>	<i>sample</i>	<i>ZZB</i> <i>153</i>	<i>ZZB</i> <i>153</i>	<i>ZZB</i> <i>153</i>	<i>ZZB</i> <i>153</i>	<i>ZZB</i> <i>153</i>	<i>ZZB</i> <i>153</i>	<i>sample</i>	<i>ZZB</i> <i>167</i>	<i>ZZB</i> <i>167</i>	<i>ZZB</i> <i>167</i>	<i>ZZB</i> <i>167</i>	<i>ZZB</i> <i>167</i>	<i>ZZB</i> <i>167</i>
<i>analysis-</i> <i>no.</i>	<i>154</i>	<i>132</i>	<i>169</i>	<i>123</i>	<i>141</i>	<i>126</i>	<i>analysis-</i> <i>no.</i>	<i>25</i>	<i>27</i>	<i>31</i>	<i>34</i>	<i>36</i>	<i>41</i>	<i>analysis-</i> <i>no.</i>	<i>61</i>	<i>87</i>	<i>108</i>	<i>93</i>	<i>70</i>	<i>100</i>
SiO ₂	51.10	50.79	51.01	51.13	51.65	50.44	SiO ₂	52.38	51.53	51.71	51.71	51.50	52.29	SiO ₂	51.57	51.58	51.29	51.02	51.35	50.70
TiO ₂	0.35	0.46	0.43	0.63	0.40	0.52	TiO ₂	0.28	0.44	0.35	0.44	0.33	0.28	TiO ₂	0.38	0.26	0.43	0.43	0.34	0.50
Al ₂ O ₃	5.30	5.62	5.16	5.91	4.81	6.73	Al ₂ O ₃	3.30	4.47	3.78	4.05	3.87	3.66	Al ₂ O ₃	3.87	3.93	4.49	4.56	3.83	4.99
FeO	5.56	5.62	5.73	6.03	6.22	6.40	FeO	7.48	8.55	8.08	8.37	8.63	7.49	FeO	7.46	7.69	8.45	8.55	9.03	9.19
MnO	0.02	0.05	0.03	0.06	0.07	0.09	MnO	0.09	0.05	0.05	0.06	0.07	0	MnO	0.07	0	0.10	0.08	0.05	0.07
MgO	13.68	13.74	13.83	13.27	13.65	13.52	MgO	13.11	12.14	13.03	12.80	12.50	13.36	MgO	12.85	12.75	12.24	11.98	12.45	12.49
CaO	22.65	22.14	22.40	22.01	22.68	21.39	CaO	22.89	22.05	22.28	22.16	21.98	22.84	CaO	21.93	21.45	20.65	20.95	21.14	19.85
Na ₂ O	0.86	0.96	0.89	1.11	0.87	0.99	Na ₂ O	0.93	1.38	1.11	1.18	1.35	1.33	Na ₂ O	1.46	1.47	1.73	1.68	1.40	1.55
Cr ₂ O ₃	0.02	0.04	0	0.03	0.01	0.10	Cr ₂ O ₃	0.27	0.06	0.05	0.03	0.28	0.03	Cr ₂ O ₃	0.09	0.03	0.10	0.14	0.11	0.07
Total	99.57	99.42	99.51	100.18	100.35	100.19	Total	100.73	100.67	100.43	100.80	100.51	101.27	Total	99.68	99.17	99.49	99.39	99.71	99.42
Si	1.88	1.87	1.88	1.87	1.89	1.85	Si	1.92	1.90	1.90	1.90	1.90	1.90	Si	1.91	1.92	1.90	1.90	1.91	1.88
Al(IV)	0.12	0.13	0.12	0.13	0.11	0.15	Al(IV)	0.08	0.10	0.10	0.10	0.10	0.10	Al(IV)	0.09	0.08	0.10	0.10	0.09	0.12
Al(VI)	0.11	0.12	0.10	0.13	0.10	0.14	Al(VI)	0.07	0.09	0.07	0.07	0.06	0.06	Al(VI)	0.07	0.09	0.10	0.10	0.07	0.10
Fe ²⁺	0.11	0.10	0.11	0.12	0.13	0.12	Fe ²⁺	0.16	0.16	0.15	0.16	0.14	0.10	Fe ²⁺	0.12	0.14	0.15	0.15	0.17	0.17
Fe ³⁺	0.06	0.07	0.07	0.06	0.06	0.07	Fe ³⁺	0.07	0.10	0.10	0.10	0.13	0.13	Fe ³⁺	0.11	0.10	0.11	0.12	0.11	0.11
Mg	0.75	0.76	0.76	0.73	0.75	0.74	Mg	0.72	0.66	0.71	0.70	0.68	0.73	Mg	0.71	0.71	0.68	0.66	0.69	0.70
Ti	0.01	0.01	0.01	0.02	0.01	0.01	Ti	0.01	0.01	0.01	0.01	0.01	0.01	Ti	0.01	0.01	0.01	0.01	0.01	0.01
Cr	0	0	0	0	0	0	Cr	0.01	0	0	0	0.01	0	Cr	0	0	0	0	0	0
Na	0.06	0.07	0.06	0.08	0.06	0.07	Na	0.07	0.10	0.08	0.08	0.10	0.09	Na	0.10	0.11	0.12	0.12	0.10	0.11
Ca	0.89	0.87	0.88	0.86	0.89	0.84	Ca	0.90	0.87	0.88	0.87	0.87	0.89	Ca	0.87	0.85	0.82	0.83	0.84	0.79
Jd%	4.26	4.56	4.02	5.72	4.08	4.99	Jd%	3.45	4.84	3.34	3.71	3.48	3.05	Jd%	4.31	5.30	6.12	5.81	4.27	5.68
Di%	81.31	81.50	81.77	78.16	79.49	79.17	Di%	75.91	72.17	75.90	74.57	74.55	79.25	Di%	76.49	73.92	71.15	71.14	71.68	70.41
Hd%	12.19	11.18	11.46	13.43	14.01	13.18	Hd%	17.19	17.43	15.75	16.58	15.23	10.90	Hd%	12.50	15.03	15.73	16.05	17.72	17.70
Ae%	2.24	2.76	2.74	2.69	2.42	2.67	Ae%	3.46	5.56	5.02	5.15	6.74	6.80	Ae%	6.69	5.76	7.01	7.00	6.32	6.20

Structural formulas are calculated on the basis of 4 cations. Endmembers are calculated after Morimoto et al. (1988).

Table III-4. Continued.

sample	ZZB 172	ZZB 172	ZZB 172	ZZB 172	ZZB 172	ZZB 172	sample	ZZB 173	ZZB 173	ZZB 173	ZZB 173	ZZB 173	ZZB 173	sample	ZZB 175	ZZB 175	ZZB 175	ZZB 175	ZZB 175	ZZB 175
analysis- no.	122	160	173	205	175	172	analysis- no.	216	254	266	218	239	245	analysis- no.	26	45	35	51	8	17
SiO ₂	51.59	50.20	49.72	49.42	48.48	47.57	SiO ₂	53.02	50.60	51.00	50.50	50.55	50.57	SiO ₂	51.92	50.86	50.86	50.97	51.42	51.23
TiO ₂	0.35	0.46	0.66	0.79	0.77	1.08	TiO ₂	0.15	0.55	0.44	0.62	0.58	0.68	TiO ₂	0.31	0.38	0.38	0.36	0.41	0.45
Al ₂ O ₃	3.72	5.46	6.17	6.29	7.28	8.94	Al ₂ O ₃	3.02	4.95	4.99	5.51	5.65	6.11	Al ₂ O ₃	3.53	4.58	4.57	4.38	4.57	4.95
FeO	7.77	9.00	9.10	8.75	9.11	9.24	FeO	6.75	8.61	8.55	8.18	8.43	8.54	FeO	8.73	9.20	9.45	9.48	9.50	10.35
MnO	0.12	0.14	0.12	0.13	0.14	0.07	MnO	0.12	0.06	0.09	0.08	0.10	0.08	MnO	0.09	0.11	0.05	0.10	0.01	0.08
MgO	12.58	11.44	11.00	11.24	10.72	9.82	MgO	14.14	11.89	12.00	12.03	11.90	11.37	MgO	12.35	11.52	11.30	11.59	11.67	11.46
CaO	23.72	21.98	22.24	22.13	21.77	21.71	CaO	22.05	21.94	21.47	21.37	21.16	20.50	CaO	21.91	21.41	20.77	20.95	21.32	20.14
Na ₂ O	0.88	1.31	1.26	1.17	1.17	1.36	Na ₂ O	1.24	1.40	1.56	1.50	1.77	1.71	Na ₂ O	1.44	1.65	1.63	1.60	1.70	1.65
Cr ₂ O ₃	0.03	0.02	0.01	0.03	0.04	0.03	Cr ₂ O ₃	0.08	0.09	0.16	0.04	0.01	0.06	Cr ₂ O ₃	0.03	0.04	0.04	0.03	0	0
Total	100.75	100.01	100.29	99.95	99.49	99.83	Total	100.56	100.09	100.28	99.85	100.15	99.64	Total	100.32	99.74	99.05	99.46	100.61	100.32
Si	1.90	1.86	1.84	1.84	1.81	1.77	Si	1.94	1.87	1.88	1.87	1.86	1.88	Si	1.92	1.89	1.91	1.90	1.89	1.90
Al(IV)	0.10	0.14	0.16	0.16	0.19	0.23	Al(IV)	0.06	0.13	0.12	0.13	0.14	0.12	Al(IV)	0.08	0.11	0.09	0.10	0.11	0.10
Al(VI)	0.06	0.10	0.11	0.11	0.13	0.17	Al(VI)	0.07	0.09	0.10	0.11	0.11	0.14	Al(VI)	0.07	0.09	0.11	0.09	0.09	0.11
Fe ²⁺	0.14	0.16	0.17	0.16	0.17	0.16	Fe ²⁺	0.12	0.14	0.14	0.14	0.12	0.18	Fe ²⁺	0.16	0.16	0.20	0.18	0.17	0.23
Fe ³⁺	0.10	0.12	0.11	0.11	0.12	0.13	Fe ³⁺	0.08	0.12	0.12	0.11	0.14	0.08	Fe ³⁺	0.11	0.13	0.09	0.11	0.12	0.09
Mg	0.69	0.64	0.61	0.62	0.60	0.54	Mg	0.77	0.66	0.66	0.66	0.66	0.63	Mg	0.68	0.64	0.63	0.64	0.64	0.63
Ti	0.01	0.01	0.02	0.02	0.02	0.03	Ti	0	0.02	0.01	0.02	0.02	0.02	Ti	0.01	0.01	0.01	0.01	0.01	0.01
Cr	0	0	0	0	0	0	Cr	0	0	0	0	0	0	Cr	0	0	0	0	0	0
Na	0.06	0.09	0.09	0.08	0.08	0.10	Na	0.09	0.10	0.11	0.11	0.13	0.12	Na	0.10	0.12	0.12	0.12	0.12	0.12
Ca	0.93	0.87	0.88	0.88	0.87	0.87	Ca	0.86	0.87	0.85	0.85	0.83	0.81	Ca	0.87	0.85	0.83	0.84	0.84	0.80
Jd%	2.54	4.73	4.88	4.61	5.00	6.29	Jd%	3.98	4.45	5.28	5.60	5.83	8.31	Jd%	4.18	5.20	6.61	5.57	5.48	6.81
Di%	77.31	71.50	70.72	72.26	70.76	68.68	Di%	78.40	73.41	72.52	73.18	73.23	67.42	Di%	72.22	70.11	66.39	68.26	69.00	64.49
Hd%	16.05	18.35	19.45	18.55	19.84	20.22	Hd%	12.53	15.84	15.57	15.28	13.17	19.43	Hd%	17.00	17.28	21.17	19.54	18.18	23.00
Ae%	4.10	5.43	4.95	4.58	4.40	4.81	Ae%	5.08	6.31	6.63	5.94	7.77	4.84	Ae%	6.60	7.41	5.84	6.63	7.35	5.70

Structural formulas are calculated on the basis of 4 cations. Endmembers are calculated after Morimoto et al. (1988).

Table III-4. Continued.

<i>sample</i>	<i>ZZB</i> 176	<i>ZZB</i> 176	<i>ZZB</i> 176	<i>ZZB</i> 176	<i>ZZB</i> 176	<i>ZZB</i> 176	<i>sample</i>	<i>ZZB</i> 195	<i>ZZB</i> 195	<i>ZZB</i> 195	<i>ZZB</i> 195	<i>ZZB</i> 195	<i>ZZB</i> 195	<i>sample</i>	<i>ZZB</i> 196	<i>ZZB</i> 196	<i>ZZB</i> 196	<i>ZZB</i> 196	<i>ZZB</i> 196	<i>ZZB</i> 196
<i>analysis-no.</i>	64	79	46	85	89	90	<i>analysis-no.</i>	18	35	48	11	25	172	<i>analysis-no.</i>	1	69	7	87	90	93
SiO ₂	51.05	49.99	49.68	49.86	49.77	50.53	SiO ₂	51.66	51.01	51.88	51.58	51.86	51.44	SiO ₂	53.81	53.14	53.45	52.68	53.17	53.19
TiO ₂	0.59	0.85	0.92	0.86	0.97	0.73	TiO ₂	0.48	0.67	0.53	0.52	0.56	0.60	TiO ₂	0.10	0.02	0.13	0.17	0.09	0.13
Al ₂ O ₃	6.63	7.24	8.20	7.81	7.80	7.72	Al ₂ O ₃	5.83	6.99	5.83	5.92	6.08	6.12	Al ₂ O ₃	1.63	1.51	1.74	2.90	1.96	1.61
FeO	4.54	4.87	5.08	5.09	5.25	5.52	FeO	4.04	4.05	4.10	4.20	4.20	4.22	FeO	10.24	10.03	10.22	9.98	9.98	10.27
MnO	0.02	0.03	0.04	0.02	0.06	0.02	MnO	0.02	0.07	0.10	0.09	0.05	0.02	MnO	0.12	0.07	0.07	0.11	0.08	0.14
MgO	13.53	13.25	12.62	12.72	12.41	12.66	MgO	14.21	13.79	14.34	14.46	14.03	14.05	MgO	11.94	11.87	11.89	11.25	11.71	11.85
CaO	21.84	22.35	22.11	21.98	22.01	21.63	CaO	23.02	22.90	23.44	23.18	23.03	23.18	CaO	21.24	21.67	20.97	19.92	20.14	20.91
Na ₂ O	1.20	0.87	1.03	1.06	1.04	1.16	Na ₂ O	0.87	0.99	0.85	0.92	0.89	0.92	Na ₂ O	2.33	2.10	2.28	2.91	2.36	2.26
Cr ₂ O ₃	0	0.10	0.04	0.05	0.06	0.06	Cr ₂ O ₃	0	0	0.02	0.09	0.04	0.11	Cr ₂ O ₃	0.05	0	0.13	0.04	0	0.06
Total	99.42	99.59	99.77	99.45	99.39	100.03	Total	100.14	100.47	101.09	100.96	100.77	100.67	Total	101.45	100.41	100.87	99.97	99.53	100.42
Si	1.87	1.84	1.82	1.84	1.84	1.85	Si	1.88	1.85	1.87	1.86	1.88	1.87	Si	1.96	1.96	1.96	1.94	1.98	1.96
Al(IV)	0.13	0.16	0.18	0.16	0.16	0.15	Al(IV)	0.12	0.15	0.13	0.14	0.12	0.13	Al(IV)	0.04	0.04	0.04	0.06	0.02	0.04
Al(VI)	0.16	0.15	0.18	0.18	0.18	0.19	Al(VI)	0.13	0.15	0.12	0.11	0.14	0.13	Al(VI)	0.03	0.03	0.04	0.07	0.06	0.03
Fe ²⁺	0.10	0.10	0.11	0.12	0.13	0.14	Fe ²⁺	0.09	0.07	0.07	0.05	0.10	0.07	Fe ²⁺	0.15	0.15	0.16	0.12	0.18	0.15
Fe ³⁺	0.04	0.05	0.04	0.04	0.03	0.03	Fe ³⁺	0.04	0.05	0.05	0.07	0.03	0.06	Fe ³⁺	0.16	0.16	0.16	0.19	0.13	0.16
Mg	0.73	0.72	0.69	0.70	0.68	0.69	Mg	0.77	0.75	0.77	0.78	0.76	0.76	Mg	0.65	0.65	0.65	0.62	0.65	0.65
Ti	0.02	0.02	0.03	0.02	0.03	0.02	Ti	0.01	0.02	0.01	0.01	0.02	0.02	Ti	0	0	0	0	0	0
Cr	0	0	0	0	0	0	Cr	0	0	0	0	0	0	Cr	0	0	0	0	0	0
Na	0.09	0.06	0.07	0.08	0.07	0.08	Na	0.06	0.07	0.06	0.06	0.06	0.06	Na	0.16	0.15	0.16	0.21	0.17	0.16
Ca	0.86	0.88	0.87	0.87	0.87	0.85	Ca	0.90	0.89	0.91	0.90	0.89	0.90	Ca	0.83	0.86	0.83	0.79	0.80	0.83
Jd%	7.42	5.15	6.50	6.74	6.91	7.84	Jd%	5.18	5.67	4.46	4.23	5.52	4.84	Jd%	2.94	2.23	3.25	5.70	5.60	2.79
Di%	79.82	81.74	79.07	78.54	77.11	75.41	Di%	83.89	84.16	85.62	87.09	82.59	84.98	Di%	67.63	68.94	67.24	66.07	64.70	67.45
Hd%	11.05	11.50	12.85	13.21	14.77	15.66	Hd%	9.55	8.32	8.01	5.98	10.76	8.04	Hd%	15.53	15.72	16.17	12.46	18.06	16.02
Ae%	1.72	1.61	1.58	1.51	1.21	1.08	Ae%	1.38	1.85	1.91	2.71	1.13	2.14	Ae%	13.90	13.11	13.34	15.77	11.64	13.74

Structural formulas are calculated on the basis of 4 cations. Endmembers are calculated after Morimoto et al. (1988).

Table III-4. Continued.

<i>sample</i>	<i>ZZB</i> <i>203</i>	<i>ZZB</i> <i>203</i>	<i>ZZB</i> <i>203</i>	<i>ZZB</i> <i>203</i>	<i>ZZB</i> <i>203</i>	<i>ZZB</i> <i>203</i>	<i>sample</i>	<i>ZZB</i> <i>220</i>	<i>ZZB</i> <i>220</i>	<i>ZZB</i> <i>220</i>	<i>ZZB</i> <i>220</i>	<i>ZZB</i> <i>220</i>	<i>ZZB</i> <i>220</i>
<i>analysis-</i> <i>no.</i>	<i>2</i>	<i>10</i>	<i>27</i>	<i>33</i>	<i>37</i>	<i>28</i>	<i>analysis-</i> <i>no.</i>	<i>182</i>	<i>167</i>	<i>172</i>	<i>193</i>	<i>158</i>	<i>160</i>
SiO ₂	51.89	50.65	51.96	52.11	52.26	51.94	SiO ₂	54.09	53.01	52.78	52.81	51.57	52.59
TiO ₂	0.42	0.39	0.27	0.32	0.33	0.27	TiO ₂	0.04	0.15	0.11	0.19	1.11	0.09
Al ₂ O ₃	4.33	4.42	3.80	3.82	3.95	3.76	Al ₂ O ₃	1.75	1.92	2.17	2.19	2.65	2.49
FeO	8.99	10.32	8.79	8.99	9.33	9.46	FeO	4.18	7.15	7.16	6.57	7.92	7.00
MnO	0.04	0.08	0.09	0.03	0.14	0.06	MnO	0.07	0.19	0.20	0.23	0.20	0.25
MgO	11.98	12.50	12.11	12.05	11.64	12.54	MgO	15.62	14.36	14.59	14.85	13.73	13.97
CaO	21.29	20.86	22.13	21.70	21.79	21.13	CaO	24.56	22.56	22.16	23.03	22.23	22.43
Na ₂ O	1.74	1.32	1.49	1.52	1.49	1.42	Na ₂ O	0.57	0.74	0.76	0.67	0.87	0.88
Cr ₂ O ₃	0.05	0.03	0.05	0.04	0.01	0.04	Cr ₂ O ₃	0.16	0.09	0.11	0.06	0.09	0.06
Total	100.73	100.92	100.69	100.58	100.94	100.63	Total	101.04	100.18	100.04	100.60	100.36	99.77
Si	1.91	1.86	1.91	1.92	1.92	1.91	Si	1.96	1.95	1.94	1.93	1.90	1.94
Al(IV)	0.09	0.14	0.09	0.08	0.08	0.09	Al(IV)	0.04	0.05	0.06	0.07	0.10	0.06
Al(VI)	0.09	0.05	0.08	0.09	0.09	0.08	Al(VI)	0.03	0.04	0.04	0.03	0.02	0.05
Fe ²⁺	0.16	0.15	0.16	0.18	0.21	0.18	Fe ²⁺	0.08	0.16	0.15	0.12	0.14	0.15
Fe ³⁺	0.11	0.17	0.11	0.09	0.08	0.11	Fe ³⁺	0.05	0.06	0.07	0.09	0.11	0.07
Mg	0.66	0.69	0.67	0.66	0.64	0.68	Mg	0.85	0.79	0.81	0.81	0.75	0.77
Ti	0.01	0.01	0.01	0.01	0.01	0.01	Ti	0	0	0	0.01	0.03	0
Cr	0	0	0	0	0	0	Cr	0	0	0	0	0	0
Na	0.12	0.09	0.11	0.11	0.11	0.10	Na	0.04	0.05	0.05	0.05	0.06	0.06
Ca	0.84	0.82	0.87	0.86	0.86	0.83	Ca	0.95	0.89	0.87	0.90	0.88	0.89
Jd%	5.83	2.42	4.49	5.38	6.06	4.40	Jd%	1.59	2.01	1.96	1.14	0.99	2.85
Di%	69.76	73.87	71.76	69.52	67.03	70.46	Di%	87.93	78.65	79.46	83.18	79.18	78.22
Hd%	17.22	15.91	17.09	19.16	21.89	18.92	Hd%	7.98	15.94	14.93	11.89	14.25	15.28
Ae%	7.19	7.81	6.66	5.95	5.02	6.22	Ae%	2.50	3.40	3.65	3.79	5.58	3.65

Structural formulas are calculated on the basis of 4 cations. Endmembers are calculated after Morimoto et al. (1988).

Table III-5. Representative microprobe analyses of garnets from samples from the Mavuradonha Layered Complex

sample	ZZB 2	ZZB 2	ZZB 2	ZZB 2	ZZB 2	ZZB 2	sample	ZZB 7	ZZB 7	ZZB 7	ZZB 7	ZZB 7	ZZB 7	sample	ZZB 12	ZZB 12	ZZB 12	ZZB 12	ZZB 12	ZZB 12
analysis-no.	140	166	128	168	188	183	analysis-no.	16	57	47	27	52	1	analysis-no.	3	34	20	122	95	31
SiO ₂	39.99	39.84	40.46	40.21	39.78	39.99	SiO ₂	38.46	38.57	38.28	38.45	38.34	38.54	SiO ₂	39.96	40.07	40.15	40.26	40.39	40.38
TiO ₂	0.06	0	0	0.06	0	0.04	TiO ₂	0.08	0.07	0.13	0.11	0.07	0.08	TiO ₂	0.043	0.033	0	0	0.012	0
FeO	19.37	18.38	18.11	19.38	18.42	17.68	FeO	23.53	24.25	25.19	23.98	24.64	25.11	FeO	18.69	19.33	18.23	19.14	19.05	19.51
Al ₂ O ₃	22.79	22.82	22.73	22.69	22.84	22.99	Al ₂ O ₃	21.11	21.23	20.96	21.26	21.15	21.27	Al ₂ O ₃	22.41	22.11	22.42	22.44	22.27	22.2
MgO	11.41	12.11	11.26	11.32	12.51	13.06	MgO	2.66	2.97	2.38	2.78	2.04	2.42	MgO	10.75	10.13	9.85	10.42	10.81	10.65
CaO	7.22	6.84	8.16	6.98	7.09	6.60	CaO	13.50	12.43	11.59	12.22	12.54	11.58	CaO	8.27	8.59	9.93	8.29	8.17	7.91
MnO	0.73	0.68	0.74	0.85	0.71	0.70	MnO	1.72	1.98	2.44	2.51	2.62	2.67	MnO	0.687	0.745	0.56	0.645	0.616	0.643
Total	101.63	100.74	101.50	101.52	101.35	101.11	Total	101.23	101.58	101.03	101.34	101.43	101.71	Total	100.832	101.019	101.177	101.195	101.338	101.328
Si	2.95	2.95	2.98	2.97	2.92	2.93	Si	2.98	2.98	3.00	2.98	2.99	2.99	Si	2.97	2.99	2.98	2.99	2.99	3.00
Ti	0	0	0	0	0	0	Ti	0	0	0.01	0.01	0	0	Ti	0	0	0	0	0	0
Al	1.98	1.99	1.97	1.97	1.98	1.99	Al	1.93	1.93	1.93	1.94	1.94	1.95	Al	1.96	1.94	1.96	1.96	1.94	1.94
Fe ³⁺	0.13	0.13	0.07	0.08	0.18	0.15	Fe ³⁺	0.12	0.10	0.06	0.08	0.07	0.06	Fe ³⁺	0.09	0.08	0.08	0.05	0.07	0.07
Fe ²⁺	1.07	1.01	1.04	1.11	0.95	0.93	Fe ²⁺	1.41	1.46	1.59	1.48	1.53	1.57	Fe ²⁺	1.08	1.13	1.06	1.14	1.11	1.14
Mn	0.05	0.04	0.05	0.05	0.04	0.04	Mn	0.11	0.13	0.16	0.17	0.17	0.18	Mn	0.04	0.05	0.04	0.04	0.04	0.04
Mg	1.25	1.34	1.24	1.25	1.37	1.43	Mg	0.31	0.34	0.28	0.32	0.24	0.28	Mg	1.19	1.13	1.09	1.15	1.19	1.18
Ca	0.57	0.54	0.64	0.55	0.56	0.52	Ca	1.12	1.03	0.97	1.02	1.05	0.96	Ca	0.66	0.69	0.79	0.66	0.65	0.63
Al IV	0.05	0.05	0.02	0.03	0.08	0.07	Al IV	0.02	0.02	0	0.02	0.01	0.01	Al IV	0.03	0.01	0.02	0.01	0.01	0
Al VI	1.93	1.94	1.95	1.94	1.90	1.92	Al VI	1.91	1.92	1.93	1.93	1.93	1.94	Al VI	1.94	1.93	1.95	1.96	1.94	1.94
Almandine	36.31	34.46	35.12	37.56	32.50	31.85	Almandine	47.71	49.36	53.02	49.62	51.24	52.54	Almandine	36.23	37.74	35.53	38.00	37.03	38.18
Andradite	6.08	6.01	3.54	4.02	8.41	7.21	Andradite	5.80	5.12	2.82	3.76	3.71	2.98	Andradite	4.21	3.86	3.74	2.60	3.66	3.47
Grossular	13.34	12.49	18.14	14.60	10.68	10.55	Grossular	32.15	29.60	29.36	30.30	31.29	29.19	Grossular	17.98	19.11	22.85	19.46	18.04	17.58
Pyrope	42.72	45.59	41.64	42.02	46.88	48.91	Pyrope	10.43	11.55	9.25	10.78	7.93	9.37	Pyrope	40.13	37.71	36.70	38.59	39.97	39.43
Spessartine	1.55	1.45	1.55	1.80	1.52	1.48	Spessartine	3.83	4.37	5.38	5.53	5.79	5.87	Spessartine	1.46	1.58	1.19	1.36	1.29	1.35
Uvarovite	0	0	0	0	0	0	Uvarovite	0.08	0	0.18	0	0.04	0.06	Uvarovite	0	0	0	0	0	0

Structural formulas are calculated on the basis of 8 cations. Endmembers are calculated after Droop (1987).

Table III-5. Continued.

<i>sample</i>	<i>ZZB 28</i>	<i>ZZB 28</i>	<i>ZZB 28</i>	<i>ZZB 28</i>	<i>ZZB 28</i>	<i>ZZB 28</i>	<i>sample</i>	<i>ZZB 41</i>	<i>ZZB 41</i>	<i>ZZB 41</i>	<i>ZZB 41</i>	<i>ZZB 41</i>	<i>ZZB 41</i>	<i>sample</i>	<i>ZZB 145</i>	<i>ZZB 145</i>	<i>ZZB 145</i>	<i>ZZB 145</i>	<i>ZZB 145</i>	<i>ZZB 145</i>
<i>analysis-no.</i>	79	82	85	92	97	98	<i>analysis-no.</i>	75	69	78	79	84	88	<i>analysis-no.</i>	1	7	25	28	34	4
SiO ₂	37.86	37.57	37.52	37.64	37.42	37.97	SiO ₂	38.74	38.51	39.05	39.06	38.81	38.57	SiO ₂	39.01	38.96	38.94	39.12	38.76	38.97
TiO ₂	0.12	0.08	0.09	0.12	0.08	0.03	TiO ₂	0.231	0.037	0.079	0.087	0.367	0.292	TiO ₂	0.08	0.11	0.14	0.10	0.11	0.11
FeO	26.81	27.25	28.34	27.05	26.53	25.74	FeO	23.16	24.14	23.56	23.52	23.6	23.96	FeO	23.02	22.42	22.38	23.13	22.30	22.73
Al ₂ O ₃	20.86	20.60	20.66	20.66	20.61	21.76	Al ₂ O ₃	21.45	21.24	21.48	21.45	21.23	20.86	Al ₂ O ₃	21.90	21.89	21.75	21.84	21.65	21.87
MgO	2.44	2.25	2.65	2.21	2.42	4.48	MgO	3.83	3.59	3.81	3.83	3.45	3.22	MgO	8.03	8.33	8.41	8.18	8.32	8.38
CaO	10.95	10.93	9.62	10.81	10.82	9.18	CaO	13.3	13.06	13.1	13.27	13.65	13.41	CaO	8.09	8.03	8.04	8.24	8.07	7.95
MnO	1.39	1.06	1.29	1.40	1.30	0.77	MnO	0.678	0.687	0.687	0.691	0.703	0.728	MnO	0.57	0.55	0.53	0.50	0.48	0.53
Total	100.52	99.90	100.32	99.91	99.27	100.03	Total	101.48	101.38	101.83	101.94	101.86	101.30	Total	100.77	100.35	100.24	101.16	99.77	100.59
Si	2.98	2.98	2.97	2.99	2.98	2.97	Si	2.97	2.97	2.99	2.99	2.98	2.98	Si	2.96	2.96	2.97	2.96	2.97	2.96
Ti	0.01	0	0.01	0.01	0	0	Ti	0.01	0	0	0.01	0.02	0.02	Ti	0.00	0.01	0.01	0.01	0.01	0.01
Al	1.94	1.93	1.93	1.93	1.94	2.00	Al	1.94	1.93	1.94	1.93	1.92	1.90	Al	1.96	1.96	1.95	1.95	1.95	1.96
Fe ³⁺	0.08	0.09	0.12	0.08	0.08	0.05	Fe ³⁺	0.09	0.13	0.07	0.09	0.09	0.09	Fe ³⁺	0.10	0.09	0.10	0.12	0.10	0.11
Fe ²⁺	1.69	1.72	1.76	1.72	1.69	1.63	Fe ²⁺	1.40	1.42	1.44	1.42	1.42	1.46	Fe ²⁺	1.36	1.34	1.33	1.34	1.33	1.34
Mn	0.09	0.07	0.09	0.09	0.09	0.05	Mn	0.04	0.04	0.04	0.04	0.05	0.05	Mn	0.04	0.04	0.03	0.03	0.03	0.03
Mg	0.29	0.27	0.31	0.26	0.29	0.52	Mg	0.44	0.41	0.43	0.44	0.39	0.37	Mg	0.91	0.95	0.95	0.92	0.95	0.95
Ca	0.92	0.93	0.82	0.92	0.92	0.77	Ca	1.09	1.08	1.07	1.09	1.12	1.11	Ca	0.66	0.65	0.66	0.67	0.66	0.65
Al IV	0.02	0.02	0.03	0.01	0.02	0.03	Al IV	0.03	0.03	0.01	0.01	0.02	0.02	Al IV	0.04	0.04	0.03	0.04	0.03	0.04
Al VI	1.92	1.91	1.90	1.92	1.92	1.97	Al VI	1.91	1.90	1.93	1.92	1.90	1.88	Al VI	1.92	1.93	1.92	1.91	1.92	1.92
Almandine	56.40	57.59	59.17	57.43	56.51	54.81	Almandine	46.99	48.11	48.10	47.44	47.70	48.77	Almandine	45.95	44.97	44.67	45.28	44.76	45.06
Andradite	3.95	4.37	5.62	3.76	3.96	2.66	Andradite	4.41	6.39	3.38	4.35	4.43	4.62	Andradite	4.78	4.40	4.72	5.82	4.69	5.18
Grossular	26.68	26.24	21.35	26.87	26.68	22.92	Grossular	32.25	29.80	32.31	32.05	33.10	31.92	Grossular	17.19	17.42	17.18	16.58	17.34	16.47
Pyrope	9.59	8.91	10.51	8.73	9.63	17.58	Pyrope	14.74	13.94	14.52	14.63	13.21	12.43	Pyrope	30.64	31.81	32.11	31.11	31.93	31.99
Spessartine	3.10	2.37	2.91	3.14	2.94	1.72	Spessartine	1.48	1.52	1.49	1.50	1.53	1.60	Spessartine	1.23	1.18	1.15	1.09	1.05	1.14
Uvarovite	0.28	0.51	0.45	0.07	0.29	0.31	Uvarovite	0.13	0.24	0.20	0.02	0.03	0.66	Uvarovite	0.21	0.21	0.16	0.13	0.23	0.16

Structural formulas are calculated on the basis of 8 cations. Endmembers are calculated after Droop (1987).

Table III-5. Continued.

<i>sample</i>	<i>ZZB</i> 147	<i>ZZB</i> 147	<i>ZZB</i> 147	<i>ZZB</i> 147	<i>ZZB</i> 147	<i>ZZB</i> 147	<i>sample</i>	<i>ZZB</i> 166	<i>ZZB</i> 166	<i>ZZB</i> 166	<i>ZZB</i> 166	<i>ZZB</i> 166	<i>ZZB</i> 166	<i>sample</i>	<i>ZZB</i> 167	<i>ZZB</i> 167	<i>ZZB</i> 167	<i>ZZB</i> 167	<i>ZZB</i> 167	<i>ZZB</i> 167
<i>analysis-no.</i>	93	120	123	126	129	132	<i>analysis-no.</i>	122	119	156	160	147	130	<i>analysis-no.</i>	62	71	80	83	89	92
SiO ₂	38.62	38.58	38.52	38.78	38.63	38.63	SiO ₂	38.05	38.20	38.46	38.44	38.66	38.49	SiO ₂	39.00	38.83	38.84	38.92	38.85	38.99
TiO ₂	0.10	0.12	0.12	0.12	0.12	0.11	TiO ₂	0.18	0.10	0.12	0.07	0.07	0.17	TiO ₂	0.14	0.09	0.12	0.13	0.14	0.13
FeO	24.02	23.92	24.39	24.30	24.52	24.41	FeO	26.02	25.79	24.74	25.61	25.41	24.99	FeO	24.42	24.75	24.88	24.45	24.68	24.68
Al ₂ O ₃	21.77	21.86	21.88	21.69	21.68	21.72	Al ₂ O ₃	21.24	21.04	21.19	21.27	21.26	21.24	Al ₂ O ₃	22.17	22.10	22.14	22.09	21.84	22.00
MgO	7.06	7.25	6.86	7.13	7.05	7.01	MgO	2.75	2.68	2.83	3.03	2.86	2.58	MgO	8.36	7.90	8.21	8.09	8.00	8.14
CaO	7.94	7.88	7.83	7.94	7.91	7.97	CaO	12.02	11.76	12.59	10.73	11.41	11.95	CaO	6.35	6.41	6.02	6.09	6.36	6.43
MnO	0.60	0.59	0.59	0.58	0.61	0.61	MnO	0.61	0.92	1.11	2.10	2.14	2.23	MnO	0.64	0.59	0.63	0.58	0.61	0.58
Total	100.17	100.23	100.22	100.58	100.54	100.55	Total	100.91	100.48	101.13	101.34	101.84	101.74	Total	101.15	100.74	100.86	100.40	100.52	100.99
Si	2.97	2.96	2.97	2.97	2.96	2.96	Si	2.97	3.00	2.99	2.99	2.99	2.98	Si	2.96	2.96	2.96	2.98	2.97	2.97
Ti	0.01	0.01	0.01	0.01	0.01	0.01	Ti	0.01	0.01	0.01	0.00	0.00	0.01	Ti	0.01	0.01	0.01	0.01	0.01	0.01
Al	1.97	1.98	1.98	1.96	1.96	1.96	Al	1.95	1.94	1.94	1.95	1.94	1.94	Al	1.98	1.99	1.99	1.99	1.97	1.97
Fe ³⁺	0.07	0.08	0.07	0.08	0.10	0.09	Fe ³⁺	0.10	0.05	0.08	0.07	0.07	0.07	Fe ³⁺	0.09	0.07	0.08	0.04	0.07	0.08
Fe ²⁺	1.47	1.45	1.50	1.47	1.47	1.47	Fe ²⁺	1.60	1.64	1.53	1.60	1.57	1.55	Fe ²⁺	1.46	1.51	1.50	1.53	1.51	1.49
Mn	0.04	0.04	0.04	0.04	0.04	0.04	Mn	0.04	0.06	0.07	0.14	0.14	0.15	Mn	0.04	0.04	0.04	0.04	0.04	0.04
Mg	0.81	0.83	0.79	0.81	0.81	0.80	Mg	0.32	0.31	0.33	0.35	0.33	0.30	Mg	0.94	0.90	0.93	0.92	0.91	0.92
Ca	0.65	0.65	0.65	0.65	0.65	0.65	Ca	1.00	0.99	1.05	0.89	0.95	0.99	Ca	0.52	0.52	0.49	0.50	0.52	0.52
Al IV	0.03	0.04	0.03	0.03	0.04	0.04	Al IV	0.03	0	0.01	0.01	0.01	0.02	Al IV	0.04	0.04	0.04	0.02	0.03	0.03
Al VI	1.94	1.94	1.95	1.93	1.92	1.93	Al VI	1.92	1.94	1.93	1.94	1.93	1.92	Al VI	1.94	1.95	1.94	1.97	1.94	1.94
Almandine	49.52	48.94	50.50	49.51	49.65	49.62	Almandine	53.99	54.61	51.34	53.60	52.62	51.84	Almandine	49.35	50.77	50.62	51.11	50.56	50.07
Andradite	3.47	4.01	3.36	4.04	4.76	4.48	Andradite	4.69	2.62	3.90	3.31	3.55	3.63	Andradite	4.12	3.50	4.03	1.89	3.52	3.95
Grossular	18.34	17.74	18.28	17.72	17.04	17.29	Grossular	29.18	30.30	31.21	26.53	28.10	29.46	Grossular	13.06	13.96	12.49	14.67	13.89	13.58
Pyrope	27.20	27.95	26.49	27.35	27.14	27.00	Pyrope	10.78	10.44	11.01	11.78	11.04	9.99	Pyrope	31.87	30.28	31.44	30.91	30.61	31.05
Spessartine	1.30	1.29	1.29	1.26	1.33	1.33	Spessartine	1.36	2.03	2.46	4.64	4.69	4.91	Spessartine	1.38	1.29	1.37	1.25	1.33	1.27
Uvarovite	0.17	0.08	0.09	0.12	0.08	0.29	Uvarovite	0	0	0.09	0.13	0	0.18	Uvarovite	0.22	0.19	0.05	0.16	0.08	0.10

Structural formulas are calculated on the basis of 8 cations. Endmembers are calculated after Droop (1987).

Table III-5. Continued.

<i>sample</i>	<i>ZZB</i> 175	<i>ZZB</i> 175	<i>ZZB</i> 175	<i>ZZB</i> 175	<i>ZZB</i> 175	<i>ZZB</i> 175	<i>sample</i>	<i>ZZB</i> 176	<i>ZZB</i> 176	<i>ZZB</i> 176	<i>ZZB</i> 176	<i>ZZB</i> 176	<i>ZZB</i> 176
<i>analysis-no.</i>	13	16	19	31	34	37	<i>analysis-no.</i>	45	72	78	81	87	48
SiO ₂	38.80	38.56	38.80	38.64	38.80	38.65	SiO ₂	39.79	39.48	39.61	39.84	39.77	39.74
TiO ₂	0.12	0.13	0.14	0.10	0.12	0.10	TiO ₂	0.11	0.11	0.08	0.11	0.12	0.12
FeO	25.25	25.04	25.20	25.62	25.44	25.73	FeO	16.39	16.65	16.68	16.53	16.36	16.52
Al ₂ O ₃	21.62	21.75	21.78	21.69	21.71	21.70	Al ₂ O ₃	22.58	22.65	22.67	22.74	22.52	22.63
MgO	6.90	7.00	6.91	6.77	6.90	6.66	MgO	12.35	12.90	12.28	12.22	12.11	12.30
CaO	7.38	7.36	7.19	7.33	7.31	7.48	CaO	8.32	7.04	8.24	8.42	8.32	8.24
MnO	0.65	0.67	0.63	0.62	0.62	0.63	MnO	0.30	0.29	0.30	0.25	0.29	0.29
Total	100.74	100.56	100.73	100.83	100.93	101.00	Total	99.87	99.14	99.89	100.14	99.59	99.91
Si	2.98	2.96	2.98	2.97	2.97	2.96	Si	2.95	2.95	2.94	2.95	2.96	2.95
Ti	0.01	0.01	0.01	0.01	0.01	0.01	Ti	0.01	0.01	0	0.01	0.01	0.01
Al	1.96	1.97	1.97	1.96	1.96	1.96	Al	1.97	1.99	1.98	1.98	1.98	1.98
Fe ³⁺	0.07	0.09	0.05	0.09	0.08	0.10	Fe ³⁺	0.10	0.10	0.12	0.10	0.08	0.10
Fe ²⁺	1.55	1.52	1.56	1.55	1.55	1.55	Fe ²⁺	0.91	0.94	0.91	0.92	0.94	0.92
Mn	0.04	0.04	0.04	0.04	0.04	0.04	Mn	0.02	0.02	0.02	0.02	0.02	0.02
Mg	0.79	0.80	0.79	0.77	0.79	0.76	Mg	1.37	1.44	1.36	1.35	1.35	1.36
Ca	0.61	0.61	0.59	0.60	0.60	0.61	Ca	0.66	0.56	0.66	0.67	0.66	0.66
Al IV	0.02	0.04	0.02	0.03	0.03	0.04	Al IV	0.05	0.05	0.06	0.05	0.04	0.05
Al VI	1.93	1.93	1.95	1.93	1.93	1.92	Al VI	1.93	1.94	1.92	1.94	1.94	1.93
Almandine	51.82	51.13	52.36	52.30	52.06	52.29	Almandine	30.84	31.70	30.95	31.24	31.72	31.18
Andradite	3.63	4.37	2.66	4.38	3.86	4.70	Andradite	5.02	4.91	5.88	4.79	3.74	4.96
Grossular	16.63	15.86	16.91	15.74	16.20	15.83	Grossular	17.23	14.10	16.31	17.71	18.34	17.01
Pyrope	26.45	27.00	26.48	26.07	26.45	25.64	Pyrope	46.18	48.61	46.15	45.63	45.29	46.04
Spessartine	1.40	1.47	1.37	1.35	1.35	1.38	Spessartine	0.63	0.62	0.64	0.53	0.62	0.62
Uvarovite	0.07	0.17	0.23	0.16	0.08	0.16	Uvarovite	0.10	0.05	0.07	0.09	0.28	0.19

Structural formulas are calculated on the basis of 8 cations. Endmembers are calculated after Droop (1987).

Table III-6. Representative microprobe analyses of plagioclases from samples from the Mavuradonha Layered Complex

sample	Mav1	Mav1	Mav1	Mav1	Mav1	Mav1	sample	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	sample	ZZB 5	ZZB 5	ZZB 5	ZZB 5	ZZB 5	ZZB 5
analysis-no.	12	1	8	6	15	19	analysis-no.	68	90	95	106	48	2	analysis-no.	88	99	116	92	93	108
SiO ₂	44.39	43.40	43.48	43.93	43.96	43.65	SiO ₂	49.08	52.09	51.44	51.67	48.63	49.84	SiO ₂	58.97	56.9	59.2	58.15	58.35	56.17
Na ₂ O	0.53	0.34	0.40	0.32	0.38	0.41	Na ₂ O	2.83	4.42	4.16	3.99	2.83	3.40	Na ₂ O	7.43	6.48	7.62	7.18	7.29	6.31
K ₂ O	0	0	0	0	0.01	0.02	K ₂ O	0.03	0.01	0.02	0.06	0.05	0.03	K ₂ O	0.03	0.03	0.06	0.04	0.03	0.02
TiO ₂	0	0	0.06	0	0	0.02	TiO ₂	0	0	0.07	0	0	0.03	TiO ₂	0.01	0	0	0.01	0	0.01
FeO	0.10	0.09	0.13	0.06	0.03	0.21	FeO	0.02	0.04	0.04	0.01	0.04	0.05	FeO	0.03	0.07	0.04	0.21	0.26	0.06
Al ₂ O ₃	35.17	35.31	35.47	35.75	35.42	35.59	Al ₂ O ₃	31.78	29.56	29.76	29.99	31.56	30.41	Al ₂ O ₃	25.39	26.76	25.42	25.89	25.46	27.23
MgO	0.01	0.02	0	0.01	0.02	0	MgO	0	0.01	0.01	0.01	0	0	MgO	0.01	0.01	0.01	0	0	0
CaO	19.39	19.62	19.67	19.74	19.77	20.06	CaO	15.73	13.35	13.49	13.65	15.57	14.73	CaO	7.57	9.31	7.56	8.39	7.91	9.71
MnO	0.01	0	0.02	0	0.01	0	MnO	0	0	0.04	0	0	0	MnO	0.01	0.05	0.02	0	0	0.05
Total	99.66	98.83	99.23	99.82	99.60	99.98	Total	99.46	99.48	99.02	99.38	98.68	98.5	Total	99.45	99.61	99.93	99.86	99.3	99.57
Si	8.23	8.13	8.12	8.14	8.17	8.10	Si	9.03	9.52	9.45	9.45	9.02	9.24	Si	10.59	10.29	10.59	10.44	10.52	10.15
Ti	0	0	0.01	0	0	0	Al	6.88	6.36	6.44	6.46	6.89	6.64	Al	5.37	5.68	5.35	5.47	5.41	5.79
Al	7.69	7.80	7.80	7.81	7.76	7.78	Ti	0	0	0.01	0	0	0	Ti	0	0	0	0	0	0
Fe ²⁺	0.02	0.01	0.02	0.01	0.00	0.03	Fe ²⁺	0	0.01	0.01	0	0.01	0.01	Fe ²⁺	0	0.01	0.01	0.03	0.04	0.01
Mn	0	0	0	0	0	0	Mn	0	0	0.01	0	0	0	Mn	0	0.01	0	0	0	0.01
Mg	0	0.01	0	0	0.01	0	Mg	0	0	0	0	0	0	Mg	0	0	0	0	0	0
Ca	3.85	3.94	3.93	3.92	3.94	3.99	Ca	3.1	2.61	2.66	2.68	3.09	2.93	Ca	1.46	1.8	1.45	1.61	1.53	1.88
Na	0.19	0.12	0.15	0.12	0.14	0.15	Na	1.01	1.57	1.48	1.42	1.02	1.22	Na	2.59	2.27	2.64	2.5	2.55	2.21
K	0	0	0	0	0	0	K	0.01	0	0	0.01	0.01	0.01	K	0.01	0.01	0.01	0.01	0.01	0.01
Or	0	0	0	0	0.06	0.10	Ab	24.50	37.40	35.80	34.50	24.70	29.40	Ab	63.90	55.70	64.40	60.60	62.40	54.00
Ab	4.70	3.07	3.57	2.87	3.39	3.57	An	75.30	62.50	64.10	65.20	75.00	70.40	An	36.00	44.20	35.30	39.20	37.40	45.90
An	95.30	96.93	96.43	97.13	96.55	96.33	Or	0.10	0	0.10	0.30	0.30	0.20	Or	0.20	0.20	0.30	0.20	0.20	0.10

Structural formulas are calculated on the basis of 32 oxygens. Endmembers are calculated after Deer et al. (1986).

Table III-6. Continued.

sample	ZZB 7	ZZB 7	ZZB 7	ZZB 7	ZZB 7	ZZB 7	sample	ZZB 15	ZZB 15	ZZB 15	ZZB 15	ZZB 15	ZZB 15	Sample	ZZB 41	ZZB 41	ZZB 41	ZZB 41	ZZB 41	ZZB 41
analysis-no.	2	23	39	48	54	66	analysis-no.	16	19	25	34	49	52	analysis-No.	104	106	108	110	112	114
SiO ₂	60.26	62.04	59.33	61.34	59.17	61.70	SiO ₂	51.70	51.89	51.88	52.54	51.74	51.51	SiO ₂	60.74	60.66	60.37	60.96	61.44	61.11
Na ₂ O	8.36	8.64	7.64	8.84	7.67	8.83	Na ₂ O	4.04	4.32	4.33	4.49	4.29	3.91	Na ₂ O	8.83	8.56	8.21	8.69	8.86	8.38
K ₂ O	0.06	0.09	0.11	0.08	0.05	0.10	K ₂ O	0.02	0.04	0.03	0.03	0.02	0.05	K ₂ O	0.09	0.04	0.07	0.07	0.03	0.05
TiO ₂	0	0.03	0	0	0	0.02	TiO ₂	0	0.05	0.01	0.02	0.01	0.48	TiO ₂	0.03	0	0.04	0.01	0.02	0
FeO	0.29	0.18	0.11	0.14	0.36	0.12	FeO	0.04	0.11	0.04	0.04	0.02	0.06	FeO	0.10	0.06	0.13	0.17	0.04	0.09
Al ₂ O ₃	24.31	23.62	24.85	23.80	25.43	23.41	Al ₂ O ₃	29.94	29.87	29.66	30.00	29.91	30.39	Al ₂ O ₃	24.17	24.02	24.27	23.90	23.39	24.06
MgO	0	0.01	0.01	0.01	0	0	MgO	0	0.04	0.01	0	0	0.01	MgO	0	0	0.02	0	0	0
CaO	6.64	5.70	7.29	5.75	7.30	5.28	CaO	13.40	13.30	12.77	13.04	13.42	13.67	CaO	6.03	6.04	6.33	6.01	5.54	6.10
MnO	0	0	0	0.01	0	0.01	MnO	0.02	0.03	0.03	0.02	0	0	MnO	0	0	0	0.01	0	0.03
Total	99.98	100.34	99.38	100.04	100.04	99.48	Total	99.16	99.64	98.76	100.18	99.42	100.08	Total	100.16	99.42	99.44	99.86	99.36	99.89
Si	10.76	10.98	10.66	10.91	10.56	11.01	Si	9.47	9.47	9.53	9.52	9.46	9.37	Si	10.83	10.86	10.81	10.87	10.99	10.89
Ti	0	0	0	0	0	0	Al	6.46	6.42	6.42	6.40	6.44	6.51	Ti	0	0	0	0	0	0
Al	5.11	4.93	5.26	4.99	5.35	4.92	Ti	0	0.01	0	0	0	0.07	Al	5.08	5.07	5.12	5.02	4.93	5.05
Fe ²⁺	0.04	0.03	0.02	0.02	0.05	0.02	Fe ²⁺	0.01	0.02	0.01	0.01	0	0.01	Fe ²⁺	0.01	0.01	0.02	0.03	0.01	0.01
Mn	0	0	0	0	0	0	Mn	0.02	0.04	0.01	0	0	0	Mn	0	0	0	0	0	0
Mg	0	0	0	0	0	0	Mg	0	0.01	0	0	0	0	Mg	0	0	0	0	0	0
Ca	1.27	1.08	1.40	1.10	1.40	1.01	Ca	2.63	2.60	2.51	2.53	2.63	2.66	Ca	1.15	1.16	1.21	1.15	1.06	1.16
Na	2.89	2.97	2.66	3.05	2.65	3.05	Na	1.44	1.53	1.54	1.58	1.52	1.38	Na	3.05	2.97	2.85	3.00	3.07	2.89
K	0.01	0.02	0.02	0.02	0.01	0.02	K	0	0.01	0.01	0.01	0.01	0.01	K	0.02	0.01	0.02	0.02	0.01	0.01
Or	0.33	0.50	0.61	0.45	0.27	0.53	Or	0.10	0.20	0.20	0.10	0.10	0.30	Or	0.47	0.22	0.40	0.36	0.17	0.25
Ab	69.27	72.92	65.08	73.23	65.35	74.76	Ab	35.30	36.90	38.00	38.30	36.60	34.00	Ab	72.26	71.79	69.84	72.09	74.20	71.13
An	30.40	26.58	34.32	26.32	34.37	24.70	An	64.60	62.80	61.90	61.50	63.30	65.70	An	27.27	27.99	29.76	27.55	25.64	28.61

Structural formulas are calculated on the basis of 32 oxygens. Endmembers are calculated after Deer et al. (1986).

Table III-6. Continued.

sample	ZZB 122	ZZB 122	ZZB 122	ZZB 122	ZZB 122	sample	ZZB 145	ZZB 145	ZZB 145	ZZB 145	ZZB 145	ZZB 145	sample	ZZB 147	ZZB 147	ZZB 147	ZZB 147	ZZB 147	ZZB 147
analysis- no.	92	95	94	91	60	analysis- no.	3	24	18	36	8	27	analysis- no.	137	146	95	119	125	143
SiO ₂	49.56	50.81	53.30	49.81	44.43	SiO ₂	59.36	59.87	59.58	59.81	59.67	59.23	SiO ₂	61.04	60.99	61.24	61.44	61.60	61.35
Na ₂ O	2.99	3.56	4.57	3.19	0.74	Na ₂ O	7.41	7.29	7.38	7.55	7.41	7.46	Na ₂ O	8.12	7.96	7.93	8.35	8.37	8.32
K ₂ O	0.04	0.04	0.06	0.06	0.03	K ₂ O	0.17	0.13	0.16	0.17	0.15	0.15	K ₂ O	0.35	0.29	0.38	0.32	0.35	0.32
TiO ₂	0.01	0	0.05	0.01	0	TiO ₂	0	0	0	0	0	0	TiO ₂	0	0	0	0	0.03	0
FeO	0	0.05	0.05	0.07	0.26	FeO	0.04	0.05	0.08	0.05	0.07	0.08	FeO	0	0	0.02	0.16	0	0.08
Al ₂ O ₃	32.16	31.20	29.48	32.04	34.90	Al ₂ O ₃	25.27	25.39	25.31	24.46	25.21	25.37	Al ₂ O ₃	22.97	23.18	24.22	24.21	24.13	23.18
MgO	0	0.01	0.01	0	0.10	MgO	0.01	0	0	0.03	0	0	MgO	0	0	0.01	0	0	0
CaO	15.48	14.31	12.61	15.18	19.15	CaO	7.16	7.06	7.28	7.17	7.30	7.11	CaO	5.53	5.84	5.97	5.78	5.64	5.72
MnO	0	0.02	0.02	0.04	0.01	MnO	0	0.02	0.06	0	0.02	0	MnO	0	0	0	0	0.03	0
Total	100.24	99.99	100.18	100.40	99.68	Total	99.42	99.81	99.85	99.28	99.83	99.40	Total	98.01	98.26	99.76	100.34	100.18	98.97
Si	9.04	9.25	9.64	9.07	8.24	Si	10.65	10.68	10.64	10.74	10.66	10.63	Si	11.05	11.02	10.91	10.89	10.93	11.02
Ti	0	0	0.01	0	0	Ti	0	0	0	0	0	0	Ti	0	0	0	0	0	0
Al	6.91	6.70	6.28	6.87	7.63	Al	5.34	5.34	5.33	5.18	5.31	5.36	Al	4.90	4.94	5.08	5.06	5.05	4.91
Fe ²⁺	0	0.01	0.01	0.01	0.04	Fe ²⁺	0.01	0.01	0.01	0.01	0.01	0.01	Fe ²⁺	0	0	0	0.02	0	0.01
Mn	0	0	0	0.01	0	Mn	0	0	0.01	0	0	0	Mn	0	0	0	0	0	0
Mg	0	0	0	0	0.03	Mg	0	0	0	0.01	0	0	Mg	0	0	0	0	0	0
Ca	3.02	2.79	2.44	2.96	3.81	Ca	1.38	1.35	1.39	1.38	1.40	1.37	Ca	1.07	1.13	1.14	1.10	1.07	1.10
Na	1.06	1.26	1.60	1.13	0.27	Na	2.58	2.52	2.56	2.63	2.57	2.59	Na	2.85	2.79	2.74	2.87	2.88	2.90
K	0.01	0.01	0.01	0.01	0.01	K	0.04	0.03	0.04	0.04	0.03	0.03	K	0.08	0.07	0.09	0.07	0.08	0.07
Or	0.25	0.23	0.35	0.35	0.17	Or	0.95	0.78	0.93	0.95	0.84	0.84	Or	2.02	1.68	2.15	1.79	1.96	1.80
Ab	25.84	30.97	39.47	27.45	6.52	Ab	64.57	64.63	64.12	64.96	64.20	64.95	Ab	71.19	69.96	69.10	71.04	71.44	71.16
An	73.91	68.79	60.18	72.19	93.31	An	34.48	34.59	34.95	34.09	34.95	34.21	An	26.79	28.36	28.75	27.17	26.60	27.04

Structural formulas are calculated on the basis of 32 oxygens. Endmembers are calculated after Deer et al. (1986).

Table III-6. Continued.

sample	ZZB 152	ZZB 152	ZZB 152	ZZB 152	ZZB 152	ZZB 152	sample	ZZB 160	ZZB 160	ZZB 160	ZZB 160	ZZB 160	ZZB 160	sample	ZZB 166	ZZB 166	ZZB 166	ZZB 166	ZZB 166	ZZB 166
analysis- no.	133	130	170	149	155	121	analysis- no.	68	70	71	52	58	6	analysis- no.	120	159	166	169	172	174
SiO ₂	57.83	56.79	57.75	55.18	57.33	57.00	SiO ₂	51.47	51.77	50.65	48.68	48.88	49.38	SiO ₂	61.19	60.56	62.55	61.42	61.48	60.90
Na ₂ O	6.83	6.59	6.83	6.03	6.78	6.38	Na ₂ O	3.96	4.03	3.7	2.85	3.06	3.04	Na ₂ O	8.70	8.45	9.09	8.77	8.74	8.62
K ₂ O	0.26	0.34	0.30	0.23	0.25	0.25	K ₂ O	0.04	0.04	0.05	0.01	0	0.02	K ₂ O	0.08	0.07	0.05	0.07	0.11	0.13
TiO ₂	0	0	0	0	0.02	0	TiO ₂	0.021	0	0	0.05	0	0	TiO ₂	0	0	0	0.02	0	0
FeO	0	0.01	0.03	0.07	0.11	0.11	FeO	0.07	0.03	0.03	0	0	0	FeO	0.22	0.25	0.07	0.09	0.01	0.06
Al ₂ O ₃	26.17	26.52	26.39	26.92	26.36	26.55	Al ₂ O ₃	30.54	30.18	30.73	31.97	31.45	31.44	Al ₂ O ₃	23.66	23.88	23.28	23.75	23.66	23.44
MgO	0	0.01	0.01	0	0.03	0	MgO	0.02	0	0	0	0.01	0.01	MgO	0	0.02	0	0.01	0	0.04
CaO	8.36	9.02	8.44	9.69	8.58	8.88	CaO	13.7	13.44	14.38	15.9	15.38	15.26	CaO	5.59	5.77	4.93	5.60	5.50	5.45
MnO	0	0.01	0	0	0.04	0.01	MnO	0	0	0.02	0	0.04	0	MnO	0.05	0.05	0.01	0.00	0	0.01
Total	99.50	99.29	99.75	98.13	99.52	99.19	Total	99.82	99.49	99.56	99.46	98.82	99.15	Total	99.54	99.19	100.11	99.76	99.55	98.71
Si	10.41	10.28	10.38	10.12	10.34	10.31	Si	9.38	9.45	9.28	8.97	9.05	9.1	Si	10.93	10.87	11.09	10.94	10.98	10.97
Ti	0	0	0	0	0	0	Al	6.55	6.49	6.63	6.93	6.86	6.82	Ti	0	0	0	0	0	0
Al	5.55	5.66	5.59	5.82	5.60	5.66	Ti	0	0	0	0.01	0	0	Al	4.98	5.05	4.86	4.99	4.98	4.97
Fe ²⁺	0	0	0	0.01	0.02	0.02	Fe ²⁺	0.01	0	0.01	0	0	0	Fe ²⁺	0.03	0.04	0.01	0.01	0	0.01
Mn	0	0	0	0	0.01	0	Mn	0	0	0	0	0.01	0	Mn	0.01	0.01	0	0	0	0
Mg	0	0	0	0	0.01	0	Mg	0	0	0	0	0	0	Mg	0	0.01	0	0	0	0.01
Ca	1.61	1.75	1.62	1.90	1.66	1.72	Ca	2.68	2.63	2.82	3.14	3.05	3.01	Ca	1.07	1.11	0.94	1.07	1.05	1.05
Na	2.38	2.31	2.38	2.14	2.37	2.24	Na	1.4	1.43	1.31	1.02	1.1	1.09	Na	3.01	2.94	3.12	3.03	3.03	3.01
K	0.06	0.08	0.07	0.05	0.06	0.06	K	0.01	0.01	0.01	0	0	0.01	K	0.02	0.02	0.01	0.01	0.02	0.03
Or	1.48	1.91	1.68	1.29	1.43	1.43	Or	0.20	0.20	0.30	0	0	0.10	Or	0.44	0.41	0.30	0.36	0.59	0.72
Ab	58.77	55.85	58.42	52.28	58.01	55.72	Ab	34.30	35.10	31.70	24.50	26.50	26.50	Ab	73.47	72.31	76.71	73.65	73.76	73.57
An	39.75	42.24	39.89	46.43	40.56	42.86	An	65.50	64.70	68.00	75.50	73.50	73.40	An	26.09	27.29	22.99	25.99	25.65	25.70

Structural formulas are calculated on the basis of 32 oxygens. Endmembers are calculated after Deer et al. (1986).

Table III-6. Continued.

Sample	ZZB 167	ZZB 167	ZZB 167	ZZB 167	ZZB 167	ZZB 167	sample	ZZB 175	ZZB 175	ZZB 175	ZZB 175	ZZB 175	ZZB 175	sample	ZZB 220	ZZB 220	ZZB 220	ZZB 220	ZZB 220	ZZB 220
analysis- No.	82	97	64	76	88	73	analysis- no.	9	52	36	18	54	21	analysis- no.	147	148	156	176	181	177
SiO ₂	62.13	61.02	61.62	61.62	62.21	62.12	SiO ₂	61.73	60.77	61.40	61.37	61.30	61.43	SiO ₂	52.61	52.54	52.1	52.62	52.08	52.57
Na ₂ O	8.66	8.54	8.83	8.71	8.96	8.77	Na ₂ O	8.60	8.73	8.69	8.80	8.61	8.69	Na ₂ O	0.01	0	0	0	0	0
K ₂ O	0.35	0.38	0.32	0.44	0.39	0.41	K ₂ O	0.17	0.14	0.14	0.12	0.13	0.12	K ₂ O	29.28	29.34	29.71	29.37	29.11	29.59
TiO ₂	0	0	0.01	0	0	0	TiO ₂	0	0	0	0	0	0	TiO ₂	0.11	0.10	0.08	0.04	0.07	0.01
FeO	0.02	0.04	0.06	0.07	0.11	0.14	FeO	0.06	0.08	0.09	0.11	0.11	0.12	FeO	0	0.02	0.02	0.02	0.05	0.03
Al ₂ O ₃	23.51	23.68	23.11	23.60	23.19	23.26	Al ₂ O ₃	23.94	23.69	23.77	23.72	24.03	24.18	Al ₂ O ₃	0.02	0.01	0.03	0	0.03	0
MgO	0	0	0	0	0	0	MgO	0	0.01	0	0	0.01	0	MgO	12.44	12.77	13.07	12.32	12.20	12.61
CaO	5.04	5.42	5.00	5.24	5.01	5.21	CaO	5.49	5.54	5.57	5.63	5.72	5.62	CaO	4.82	4.48	4.31	4.77	4.80	4.58
MnO	0	0	0	0	0	0	MnO	0.02	0.02	0.01	0	0	0.03	MnO	0.01	0	0.03	0.02	0.02	0.03
Total	99.77	99.08	98.95	99.71	99.87	99.94	Total	100.03	99.04	99.71	99.76	99.91	100.21	Total	99.3	99.26	99.34	99.16	98.35	99.51
Si	11.05	10.95	11.06	10.99	11.07	11.05	Si	10.96	10.92	10.95	10.94	10.91	10.90	Si	9.61	9.60	9.52	9.62	9.61	9.58
Ti	0	0	0	0	0	0	Ti	0	0	0	0	0	0	Al	6.30	6.31	6.40	6.32	6.32	6.35
Al	4.93	5.01	4.89	4.96	4.86	4.87	Al	5.01	5.02	4.99	4.98	5.04	5.06	Ti	0	0	0	0	0	0
Fe ²⁺	0	0.01	0.01	0.01	0.02	0.02	Fe ²⁺	0.01	0.01	0.01	0.02	0.02	0.02	Fe ²⁺	0.02	0.02	0.01	0.01	0.01	0.02
Mn	0	0	0	0	0	0	Mn	0	0	0	0	0	0	Mn	0	0	0	0	0.01	0.01
Mg	0	0	0	0	0	0	Mg	0	0	0	0	0	0	Mg	0	0	0.01	0	0.01	0
Ca	0.96	1.04	0.96	1.00	0.95	0.99	Ca	1.04	1.07	1.06	1.08	1.09	1.07	Ca	2.44	2.50	2.56	2.41	2.41	2.46
Na	2.99	2.97	3.07	3.01	3.09	3.02	Na	2.96	3.04	3.00	3.04	2.97	2.99	Na	1.71	1.59	1.53	1.70	1.72	1.62
K	0.08	0.09	0.07	0.10	0.09	0.09	K	0.04	0.03	0.03	0.03	0.03	0.03	K	0	0	0.01	0.01	0	0.01
Or	1.94	2.10	1.79	2.43	2.12	2.27	Or	0.93	0.76	0.77	0.65	0.74	0.65	Or	0.10	0	0.20	0.10	0.10	0.20
Ab	74.19	72.48	74.80	73.23	74.78	73.57	Ab	73.24	73.47	73.27	73.40	72.60	73.19	Ab	41.20	38.80	37.30	41.20	41.60	39.60
An	23.86	25.42	23.41	24.34	23.10	24.15	An	25.83	25.76	25.95	25.95	26.65	26.16	An	58.70	61.20	62.50	58.70	58.30	60.20

Structural formulas are calculated on the basis of 32 oxygens. Endmembers are calculated after Deer et al. (1986).

Table III-7. Representative microprobe analyses of orthopyroxenes from samples from the Mavuradonha Layered Complex.

sample	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	sample	ZZB 220	ZZB 220	ZZB 220	ZZB 220	ZZB 220	ZZB 220
analysis no.	26	58	107	99	120	123	analysis no.	151	166	135	194	171	155
SiO ₂	53.47	52.52	54.48	53.85	54.91	54.70	SiO ₂	51.99	52.49	52.27	52.29	52.93	52.96
TiO ₂	0.01	0.97	0.01	0.13	0.02	0.03	TiO ₂	0.02	0.12	0.06	0.07	0	0.01
Al ₂ O ₃	1.85	2.32	1.38	1.75	1.23	1.49	Al ₂ O ₃	1.96	1.49	1.51	1.56	1.65	1.70
FeO	16.95	18.03	18.20	17.32	17.45	16.93	FeO	22.30	22.89	21.38	22.31	22.23	22.31
MnO	0.65	0.56	0.50	0.57	0.70	0.71	MnO	0.56	0.52	0.58	0.53	0.61	0.63
MgO	25.48	25.46	27.02	27.08	27.46	27.72	MgO	22.73	23.03	22.91	23.39	23.23	23.42
CaO	1.87	1.26	0.19	0.60	0.17	0.33	CaO	0.28	0.35	0.39	0.33	0.33	0.29
Na ₂ O	0.12	0.20	0.01	0.05	0	0.04	Na ₂ O	0.03	0	0.01	0.03	0.01	0
Cr ₂ O ₃	0	0	0	0	0	0	Cr ₂ O ₃	0.03	0.08	0.04	0.04	0.06	0.07
Total	100.41	101.37	101.81	101.35	101.94	101.95	Total	99.90	101.08	99.16	100.54	101.06	101.38
Si	1.93	1.89	1.94	1.92	1.95	1.94	Si	1.96	1.93	1.95	1.93	1.94	1.94
Al(IV)	0.07	0.11	0.06	0.08	0.05	0.06	Al(IV)	0.07	0.07	0.05	0.07	0.06	0.06
Al(VI)	0.01	0	0	0	0	0	Al(VI)	0.02	0	0.02	0	0.01	0.01
Fe ²⁺	0.45	0.43	0.48	0.43	0.47	0.44	Fe ²⁺	0.64	0.64	0.64	0.61	0.64	0.63
Fe ³⁺	0.06	0.11	0.06	0.08	0.05	0.07	Fe ³⁺	0.05	0.07	0.03	0.08	0.05	0.05
Mg	1.37	1.37	1.43	1.44	1.45	1.44	Mg	1.26	1.26	1.28	1.28	1.27	1.28
Ti	0	0.03	0	0	0	0	Ti	0	0.01	0	0	0	0
Cr	0	0	0	0	0	0	Cr	0	0	0	0	0	0
Na	0.01	0.01	0	0	0	0	Na	0	0	0	0	0	0
Ca	0.07	0.05	0.01	0.02	0.01	0.01	Ca	0.01	0.01	0.02	0.01	0.01	0.01
En%	74.79	75.00	74.80	76.56	75.61	76.88	En%	66.23	66.50	66.58	67.74	66.60	67.04
Fs%	25.21	25.00	25.20	23.44	24.39	23.22	Fs%	33.77	33.50	33.42	32.26	33.40	32.96

Structural formulas are calculated on the basis of 4 cations. Endmembers are calculated after Morimoto et al. (1988).

Table III-8. Representative microprobe analyses of amphiboles from samples from the Mavuradonha Layered Complex

sample	Mav1	Mav1	Mav1	Mav1	Mav1	Mav1	sample	ZZB 7	ZZB 7	ZZB 7	ZZB 7	ZZB 7	ZZB 7	sample	ZZB 8	ZZB 8	ZZB 8	ZZB 8	ZZB 8	ZZB 8
analysis-no.	22	20	44	17	18	21	analysis-no.	9	32	45	49	53	61	analysis-no.	190	191	192	200	202	203
SiO ₂	41.63	41.41	45.71	47.88	56.30	53.48	SiO ₂	42.35	42.60	42.19	42.12	42.07	41.46	SiO ₂	48.27	46.19	51.36	45.62	47.12	51.64
Na ₂ O	2.06	2.23	1.56	1.39	0.23	0.51	Na ₂ O	1.81	1.71	1.68	1.85	1.75	1.71	Na ₂ O	0.96	1.20	0.57	1.30	1.09	0.61
K ₂ O	0.71	0.78	0.41	0.24	0.03	0.08	K ₂ O	0.13	0.12	0.13	0.15	0.15	0.13	K ₂ O	0.27	0.44	0.18	0.41	0.35	0.20
TiO ₂	0.21	0.29	0.19	0.22	0	0.09	TiO ₂	0.64	0.76	0.71	0.70	0.67	0.73	TiO ₂	0.38	0.45	0.20	0.47	0.41	0.27
FeO	10.26	9.86	7.46	6.69	4.35	5.07	FeO	17.24	17.38	18.10	17.56	18.42	18.64	FeO	10.89	11.65	9.72	12.09	11.76	9.57
Al ₂ O ₃	17.38	17.03	13.30	11.50	2.22	4.89	Al ₂ O ₃	14.31	13.86	14.03	14.51	14.26	13.98	Al ₂ O ₃	9.06	11.17	5.92	11.56	10.51	6.01
MgO	12.49	12.55	15.55	16.87	21.47	20.26	MgO	8.61	9.03	8.44	8.88	8.14	8.19	MgO	14.74	13.24	16.81	12.96	13.96	16.61
CaO	12.32	12.45	12.90	12.51	12.57	12.65	CaO	11.70	11.49	11.50	11.75	11.55	11.44	CaO	12.51	12.48	12.63	12.53	12.38	12.61
Cr ₂ O ₃	0.17	0.26	0.17	0.18	0.17	0.24	Cr ₂ O ₃	0.08	0.07	0.05	0	0.07	0.05	Cr ₂ O ₃	0	0	0	0	0	0
MnO	0.16	0.13	0.12	0.14	0.15	0.16	MnO	0.23	0.21	0.34	0.18	0.35	0.31	MnO	0.29	0.26	0.24	0.35	0.22	0.24
Total	97.40	96.99	97.37	97.62	97.49	97.44	Total	97.10	97.23	97.16	97.70	97.42	96.63	Total	97.36	97.09	97.63	97.29	97.81	97.75
Si	6.03	6.05	6.52	6.72	7.72	7.39	Si	6.31	6.31	6.28	6.23	6.27	6.23	Si	6.93	6.72	7.27	6.64	6.76	7.31
Al IV	1.97	1.95	1.48	1.28	0.28	0.61	Al IV	1.69	1.69	1.72	1.77	1.73	1.77	Al IV	1.07	1.28	0.73	1.36	1.24	0.69
Ti	0.02	0.03	0.02	0.02	0	0.01	Ti	0.07	0.08	0.08	0.08	0.08	0.08	Ti	0.04	0.05	0.02	0.05	0.04	0.03
Al VI	1.00	0.98	0.75	0.62	0.08	0.19	Al VI	0.83	0.73	0.75	0.77	0.77	0.70	Al VI	0.46	0.63	0.25	0.62	0.53	0.31
Cr	0.02	0.03	0.02	0.02	0.02	0.03	Cr	0.01	0.01	0.01	0	0.01	0.01	Cr	0	0	0	0	0	0
Fe 3+	0.39	0.23	0.24	0.42	0.45	0.51	Fe 3+	0.43	0.62	0.63	0.56	0.59	0.71	Fe 3+	0.36	0.24	0.42	0.28	0.45	0.30
Fe 2+	0.85	0.98	0.64	0.36	0.05	0.08	Fe 2+	1.72	1.53	1.62	1.62	1.70	1.63	Fe 2+	0.94	1.17	0.73	1.19	0.96	0.83
Mn	0.02	0.02	0.01	0.02	0.02	0.02	Mn	0.03	0.03	0.04	0.02	0.04	0.04	Mn	0.03	0.03	0.03	0.04	0.03	0.03
Mg	2.70	2.73	3.31	3.53	4.39	4.17	Mg	1.91	1.99	1.87	1.96	1.81	1.83	Mg	3.15	2.87	3.55	2.81	2.98	3.50
Ca	1.91	1.95	1.97	1.88	1.85	1.87	Ca	1.87	1.82	1.83	1.86	1.84	1.84	Ca	1.92	1.94	1.91	1.95	1.90	1.91
Na	0.58	0.63	0.43	0.38	0.06	0.14	Na	0.52	0.49	0.49	0.53	0.51	0.50	Na	0.27	0.34	0.16	0.37	0.30	0.17
K	0.13	0.15	0.07	0.04	0	0.01	K	0.02	0.02	0.02	0.03	0.03	0.02	K	0.05	0.08	0.03	0.08	0.06	0.04

Structural formulas are calculated on the basis of 13 cations except Ca, Na and K. Endmembers are calculated after Leake et al. (1997).

Table III-8. Continued.

Sample	ZZB 9	ZZB 9	ZZB 9	ZZB 9	ZZB 9	ZZB 9	sample	ZZB 28	ZZB 28	ZZB 28	ZZB 28	ZZB 28	ZZB 28	sample	ZZB 41	ZZB 41	ZZB 41	ZZB 41	ZZB 41	ZZB 41
analysis-no.	130	140	132	164	143	127	analysis-no.	101	102	104	80	58	76	analysis-no.	83	89	103	105	109	113
SiO ₂	44.30	45.19	44.17	45.86	48.41	47.76	SiO ₂	42.60	42.24	41.37	40.46	40.19	39.22	SiO ₂	40.91	40.91	41.97	42.11	42.10	41.93
Na ₂ O	1.60	1.32	1.60	1.28	0.83	0.96	Na ₂ O	2.03	2.05	2.03	1.99	2.01	2.08	Na ₂ O	2.66	2.66	2.48	2.54	2.56	2.55
K ₂ O	0.52	0.42	0.43	0.37	0.34	0.43	K ₂ O	0.89	0.91	1.01	1.12	1.13	1.12	K ₂ O	1.62	1.62	0.84	0.83	0.83	0.84
TiO ₂	0.42	0.46	0.43	0.40	0.41	0.41	TiO ₂	0.73	0.65	1.10	1.05	0.88	0.89	TiO ₂	0.71	0.86	1.00	1.06	1.20	1.10
FeO	12.76	12.09	12.17	11.66	10.73	11.11	FeO	17.64	17.64	18.16	19.58	20.26	20.99	FeO	17.39	17.58	16.87	16.88	16.07	16.53
Al ₂ O ₃	13.43	12.23	12.98	11.86	9.32	10.12	Al ₂ O ₃	13.45	13.57	14.14	14.34	15.09	15.49	Al ₂ O ₃	14.14	14.09	13.72	13.65	13.47	13.54
MgO	12.23	12.97	12.26	13.39	14.79	14.19	MgO	9.61	9.57	8.51	7.16	6.68	6.22	MgO	9.12	9.01	9.85	10.10	10.30	10.03
CaO	12.06	12.17	11.95	12.12	12.24	12.50	CaO	11.45	11.39	11.18	11.29	11.36	11.28	CaO	11.74	11.53	11.67	11.67	11.53	11.44
Cr ₂ O ₃	0	0	0	0	0	0	Cr ₂ O ₃	0.11	0	0.12	0.28	0.27	0.08	Cr ₂ O ₃	0.02	0.08	0.01	0.06	0.06	0.06
MnO	0.29	0.26	0.33	0.28	0.32	0.28	MnO	0.13	0.17	0.15	0.16	0.10	0.06	MnO	0.05	0.02	0.05	0.07	0.09	0.08
Total	97.60	97.10	96.32	97.22	97.39	97.76	Total	98.63	98.19	97.76	97.42	97.97	97.43	Total	98.37	98.35	98.46	98.95	98.20	98.11
Si	6.43	6.56	6.49	6.62	6.92	6.85	Si	6.27	6.25	6.19	6.15	6.09	5.99	Si	6.16	6.16	6.22	6.20	6.23	6.22
Al IV	1.57	1.44	1.51	1.38	1.08	1.15	Al IV	1.73	1.75	1.81	1.85	1.91	2.01	Al IV	1.84	1.84	1.78	1.80	1.77	1.78
Ti	0.05	0.05	0.05	0.04	0.04	0.04	Ti	0.08	0.07	0.12	0.12	0.10	0.10	Ti	0.08	0.10	0.11	0.12	0.13	0.12
Al VI	0.73	0.65	0.73	0.63	0.49	0.56	Al VI	0.61	0.61	0.68	0.72	0.78	0.78	Al VI	0.67	0.66	0.61	0.57	0.59	0.59
Cr	0	0	0	0	0	0	Cr	0.01	0	0.01	0.03	0.03	0.01	Cr	0	0.01	0	0.01	0.01	0.01
Fe 3+	0.46	0.47	0.39	0.49	0.47	0.32	Fe 3+	0.60	0.62	0.53	0.42	0.44	0.49	Fe 3+	0.13	0.19	0.37	0.43	0.36	0.41
Fe 2+	1.09	1.00	1.11	0.92	0.81	1.01	Fe 2+	1.57	1.56	1.74	2.07	2.13	2.19	Fe 2+	2.06	2.03	1.72	1.65	1.63	1.64
Mn	0.04	0.03	0.04	0.03	0.04	0.03	Mn	0.02	0.02	0.02	0.02	0.01	0.01	Mn	0.01	0	0.01	0.01	0.01	0.01
Mg	2.65	2.81	2.68	2.88	3.15	3.03	Mg	2.11	2.11	1.90	1.62	1.51	1.42	Mg	2.05	2.02	2.18	2.22	2.27	2.22
Ca	1.87	1.89	1.88	1.87	1.87	1.92	Ca	1.81	1.81	1.79	1.84	1.84	1.85	Ca	1.89	1.86	1.85	1.84	1.83	1.82
Na	0.45	0.37	0.46	0.36	0.23	0.27	Na	0.58	0.59	0.59	0.59	0.59	0.62	Na	0.78	0.78	0.71	0.73	0.73	0.73
K	0.10	0.08	0.08	0.07	0.06	0.08	K	0.17	0.17	0.19	0.22	0.22	0.22	K	0.31	0.31	0.16	0.15	0.16	0.16

Structural formulas are calculated on the basis of 13 cations except Ca, Na and K.

Table III-8. Continued.

Sample	ZZB 58-3	ZZB 58-3	ZZB 58-3	ZZB 58-3	ZZB 58-3	ZZB 58-3	sample	ZZB 122	ZZB 122	ZZB 122	ZZB 122	ZZB 122	ZZB 122	sample	ZZB 123	ZZB 123	ZZB 123	ZZB 123	ZZB 123	ZZB 123
analysis- no.	146	134	138	131	140	145	analysis- no.	59	97	74	103	69	99	analysis- no.	10	15	2	11	3	13
SiO ₂	46.98	43.82	41.66	42.30	43.44	43.39	SiO ₂	44.00	42.80	41.00	42.89	43.42	41.39	SiO ₂	55.12	55.19	54.27	52.09	52.00	50.37
Na ₂ O	1.20	1.81	2.12	2.07	1.62	1.77	Na ₂ O	1.76	2.36	2.48	2.24	1.77	2.47	Na ₂ O	0.64	0.73	0.61	0.96	0.90	1.18
K ₂ O	0.40	0.38	0.38	0.34	0.52	0.54	K ₂ O	0.50	0.31	0.86	0.63	0.61	0.48	K ₂ O	0.05	0.11	0.08	0.22	0.18	0.27
TiO ₂	0.43	0.45	0.25	0.31	0.79	0.65	TiO ₂	0.12	0.08	0.07	0.51	0.21	0.10	TiO ₂	0.08	0.10	0.09	0.25	0.18	0.39
FeO	10.94	11.51	12.73	12.92	13.39	14.02	FeO	7.53	7.97	7.98	8.08	8.73	9.67	FeO	9.57	10.06	10.21	11.18	11.74	12.01
Al ₂ O ₃	10.31	15.27	17.42	16.61	13.13	13.78	Al ₂ O ₃	15.92	18.67	19.81	17.27	16.68	19.25	Al ₂ O ₃	1.76	1.78	2.40	4.34	4.10	5.42
MgO	13.80	12.17	10.92	11.07	11.49	11.12	MgO	14.42	13.79	13.59	14.49	13.90	12.30	MgO	18.16	18.08	17.55	15.99	15.99	15.44
CaO	12.43	12.42	11.95	11.97	12.45	12.27	CaO	12.48	11.66	12.07	12.17	12.47	11.62	CaO	11.56	11.69	12.02	12.13	11.81	12.12
Cr ₂ O ₃	0.20	0.12	0.10	0.26	0.19	0.14	Cr ₂ O ₃	0.02	0.13	0.04	0.12	0	0.02	Cr ₂ O ₃	0.08	0.03	0.08	0.05	0.16	0.16
MnO	0.17	0.21	0.23	0.26	0.14	0.22	MnO	0.08	0.06	0.05	0.06	0.09	0.05	MnO	0.14	0.10	0.09	0.11	0.08	0.09
Total	96.86	98.17	97.76	98.12	97.16	97.89	Total	96.83	97.83	97.94	98.47	97.88	97.34	Total	97.16	97.87	97.42	97.33	97.13	97.45
Si	6.83	6.32	6.05	6.12	6.41	6.36	Si	6.30	6.03	5.83	6.05	6.18	5.94	Si	7.75	7.73	7.67	7.46	7.44	7.25
Al IV	1.17	1.68	1.95	1.88	1.59	1.64	Al IV	1.70	1.97	2.17	1.95	1.82	2.06	Al IV	0.25	0.27	0.33	0.54	0.56	0.75
Ti	0.05	0.05	0.03	0.03	0.09	0.07	Ti	0.01	0.01	0.01	0.05	0.02	0.01	Ti	0.01	0.01	0.01	0.03	0.02	0.04
Al VI	0.59	0.91	1.03	0.95	0.69	0.74	Al VI	0.99	1.13	1.15	0.92	0.98	1.19	Al VI	0.04	0.02	0.07	0.20	0.14	0.17
Cr	0.02	0.01	0.01	0.03	0.02	0.02	Cr	0	0.01	0	0.01	0	0	Cr	0.01	0	0.01	0.01	0.02	0.02
Fe 3+	0.21	0.27	0.49	0.50	0.22	0.30	Fe 3+	0.27	0.60	0.49	0.51	0.39	0.50	Fe 3+	0.54	0.50	0.42	0.25	0.48	0.38
Fe 2+	1.12	1.12	1.05	1.06	1.43	1.41	Fe 2+	0.64	0.34	0.46	0.44	0.64	0.65	Fe 2+	0.59	0.67	0.78	1.09	0.93	1.06
Mn	0.02	0.03	0.03	0.03	0.02	0.03	Mn	0.01	0.01	0.01	0.01	0.01	0.01	Mn	0.02	0.01	0.01	0.01	0.01	0.01
Mg	2.99	2.61	2.36	2.39	2.53	2.43	Mg	3.08	2.90	2.88	3.05	2.95	2.63	Mg	3.80	3.77	3.70	3.42	3.41	3.31
Ca	1.93	1.92	1.86	1.86	1.97	1.93	Ca	1.92	1.76	1.84	1.84	1.90	1.79	Ca	1.74	1.75	1.82	1.86	1.81	1.87
Na	0.34	0.51	0.60	0.58	0.46	0.50	Na	0.49	0.64	0.68	0.61	0.49	0.69	Na	0.17	0.20	0.17	0.27	0.25	0.33
K	0.07	0.07	0.07	0.06	0.10	0.10	K	0.09	0.06	0.16	0.11	0.11	0.09	K	0.01	0.02	0.01	0.04	0.03	0.05

Structural formulas are calculated on the basis of 13 cations except Ca, Na and K.

Table III-8. Continued.

<i>sample</i>	<i>ZZB</i> <i>166</i>	<i>ZZB</i> <i>166</i>	<i>ZZB</i> <i>166</i>	<i>ZZB</i> <i>166</i>	<i>ZZB</i> <i>166</i>	<i>ZZB</i> <i>166</i>	<i>sample</i>	<i>ZZB</i> <i>173</i>	<i>ZZB</i> <i>173</i>	<i>ZZB</i> <i>173</i>	<i>ZZB</i> <i>173</i>	<i>ZZB</i> <i>173</i>	<i>ZZB</i> <i>173</i>
<i>analysis-</i> <i>no.</i>	<i>127</i>	<i>145</i>	<i>155</i>	<i>165</i>	<i>167</i>	<i>168</i>	<i>analysis-</i> <i>no.</i>	<i>212</i>	<i>241</i>	<i>244</i>	<i>247</i>	<i>263</i>	<i>268</i>
SiO ₂	41.93	42.71	41.28	42.33	42.48	42.32	SiO ₂	42.16	41.08	42.28	41.16	42.19	41.84
Na ₂ O	2.04	1.99	2.13	2.10	2.21	2.04	Na ₂ O	2.29	2.19	2.22	2.19	2.11	2.03
K ₂ O	0.60	0.53	0.62	0.55	0.58	0.59	K ₂ O	0.83	0.93	0.71	1.02	0.73	1.04
TiO ₂	0.82	0.72	0.68	0.77	1.00	0.90	TiO ₂	1.79	1.21	1.21	1.46	0.90	1.76
FeO	17.88	15.40	19.19	17.90	17.62	17.34	FeO	12.86	16.43	15.87	15.34	15.46	14.84
Al ₂ O ₃	13.30	14.31	14.98	13.50	13.38	13.42	Al ₂ O ₃	13.65	12.50	11.42	12.84	12.39	12.81
MgO	8.70	10.17	7.50	9.11	9.35	9.33	MgO	12.34	10.22	10.85	10.83	11.12	10.80
CaO	11.12	11.54	11.21	11.30	11.21	11.38	CaO	12.04	11.52	11.45	11.68	11.63	12.02
Cr ₂ O ₃	0.12	0.01	0.01	0.03	0.07	0	Cr ₂ O ₃	0.07	0.03	0.08	0.01	0.04	0.08
MnO	0.20	0.10	0.24	0.18	0.15	0.14	MnO	0.05	0.17	0.10	0.06	0.07	0.00
Total	96.71	97.49	97.85	97.78	98.04	97.46	Total	98.08	96.27	96.20	96.59	96.64	97.21
Si	6.31	6.29	6.18	6.29	6.29	6.30	Si	6.17	6.22	6.37	6.19	6.30	6.26
Al IV	1.69	1.71	1.82	1.71	1.71	1.70	Al IV	1.83	1.78	1.63	1.81	1.70	1.74
Ti	0.09	0.08	0.08	0.09	0.11	0.10	Ti	0.20	0.14	0.14	0.17	0.10	0.20
Al VI	0.67	0.78	0.82	0.65	0.62	0.65	Al VI	0.52	0.45	0.40	0.47	0.48	0.52
Cr	0.01	0	0	0	0.01	0	Cr	0.01	0	0.01	0	0	0.01
Fe 3+	0.55	0.46	0.52	0.59	0.58	0.52	Fe 3+	0.33	0.49	0.47	0.41	0.55	0.17
Fe 2+	1.70	1.43	1.88	1.63	1.60	1.64	Fe 2+	1.24	1.59	1.53	1.52	1.38	1.69
Mn	0.03	0.01	0.03	0.02	0.02	0.02	Mn	0.01	0.02	0.01	0.01	0.01	0
Mg	1.95	2.23	1.67	2.02	2.06	2.07	Mg	2.69	2.31	2.44	2.43	2.48	2.41
Ca	1.79	1.82	1.80	1.80	1.78	1.81	Ca	1.89	1.87	1.85	1.88	1.86	1.93
Na	0.60	0.57	0.62	0.60	0.63	0.59	Na	0.65	0.64	0.65	0.64	0.61	0.59
K	0.11	0.10	0.12	0.10	0.11	0.11	K	0.15	0.18	0.14	0.19	0.14	0.20

Structural formulas are calculated on the basis of 13 cations except Ca, Na and K.

Table III-9. Representative microprobe analyses of chlorine-rich amphiboles from samples from the Mavuradonha Layered Complex

Sample	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	Sample	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2	Sample	ZZB 5	ZZB 5	ZZB 5	ZZB 5	ZZB 5	ZZB 5
analysis-no.	246	93	81	30	25	246	analysis-no.	132	156	157	163	148	136	analysis-no.	77	100	122	125	127	129
SiO ₂	39.14	37.01	39.85	38.75	39.53	39.14	SiO ₂	42.26	42.41	42.48	41.77	42.12	42.07	SiO ₂	42.21	41.80	41.40	42.72	42.15	41.44
Na ₂ O	2.50	2.83	2.87	2.82	2.80	2.50	Na ₂ O	2.30	2.30	2.30	2.31	2.34	2.47	Na ₂ O	2.34	2.34	2.35	2.40	2.40	2.42
K ₂ O	0.96	0.70	0.58	0.57	0.56	0.96	K ₂ O	0.29	0.29	0.30	0.34	0.28	0.28	K ₂ O	0.24	0.32	0.32	0.33	0.34	0.29
TiO ₂	0.80	0.67	0.52	0.48	0.96	0.80	TiO ₂	0.85	0.81	0.79	0.89	0.92	0.74	TiO ₂	0.38	0.81	0.29	0.57	0.63	0.58
FeO	17.53	17.71	15.91	15.73	15.19	17.53	FeO	13.52	13.52	13.41	13.97	13.87	11.39	FeO	12.76	13.22	15.00	13.28	13.50	13.92
Al ₂ O ₃	13.24	16.85	14.84	16.60	15.31	13.24	Al ₂ O ₃	14.46	14.54	14.16	14.42	14.45	15.95	Al ₂ O ₃	15.72	14.82	15.61	14.20	14.94	14.99
MgO	8.93	7.45	9.91	9.39	10.07	8.93	MgO	11.37	11.37	11.44	10.65	10.88	11.89	MgO	11.60	11.40	10.30	11.51	11.17	10.66
CaO	11.79	11.13	11.38	11.50	11.48	11.79	CaO	11.68	11.93	12.13	12.02	11.82	12.04	CaO	11.80	11.93	11.47	11.89	11.73	11.53
MnO	0.19	0.24	0.18	0.20	0.18	0.19	MnO	0.13	0.13	0.09	0.17	0.19	0.08	MnO	0.13	0.13	0.18	0.09	0.17	0.15
Cl	4.85	5.20	4.15	4.31	3.92	4.85	Cl	1.66	1.72	1.68	1.90	1.85	2.17	Cl	1.66	1.73	2.02	1.80	1.92	1.84
Total	99.97	99.79	100.23	100.40	100.05	99.97	Total	98.59	99.07	98.82	98.50	98.83	99.10	Total	98.85	98.58	98.98	98.81	98.97	97.81
-O equ. Cl	1.09	1.17	0.94	0.97	0.88	1.09	-O equ. Cl	0.37	0.39	0.38	0.43	0.42	0.49	-O equ. Cl	0.37	0.39	0.46	0.41	0.43	0.42
Total	98.88	98.62	99.29	99.43	99.17	98.88	CTotal	98.22	98.68	98.44	98.07	98.41	98.61	CTotal	98.48	98.19	98.52	98.40	98.54	97.39

Table III-9. Continued

Sample	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	Sample	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2	Sample	ZZB 5	ZZB 5	ZZB 5	ZZB 5	ZZB 5	ZZB 5
<i>analysis-no.</i>	246	93	81	30	25	246	<i>analysis-no.</i>	132	156	157	163	148	136	<i>analysis-no.</i>	77	100	122	125	127	129
Si	6.10	5.76	6.03	5.86	5.97	6.10	Si	6.22	6.23	6.27	6.23	6.24	6.16	Si	6.16	6.17	6.11	6.30	6.21	6.18
Al IV	1.90	2.24	1.97	2.14	2.04	1.90	Al IV	1.78	1.77	1.73	1.77	1.76	1.84	Al IV	1.84	1.83	1.89	1.70	1.80	1.82
Ti	0.09	0.08	0.06	0.06	0.11	0.09	Ti	0.09	0.09	0.09	0.10	0.10	0.08	Ti	0.04	0.09	0.03	0.06	0.07	0.07
Al VI	0.53	0.85	0.67	0.81	0.69	0.53	Al VI	0.73	0.74	0.73	0.77	0.76	0.92	Al VI	0.86	0.75	0.83	0.76	0.80	0.81
Fe 3+	0.31	0.52	0.54	0.56	0.49	0.31	Fe 3+	0.47	0.38	0.28	0.23	0.33	0.22	Fe 3+	0.50	0.39	0.63	0.31	0.41	0.44
Fe 2+	1.97	1.79	1.47	1.43	1.42	1.97	Fe 2+	1.20	1.28	1.38	1.52	1.39	1.17	Fe 2+	1.06	1.24	1.22	1.32	1.25	1.29
Mn	0.03	0.03	0.02	0.03	0.02	0.03	Mn	0.02	0.02	0.01	0.02	0.02	0.01	Mn	0.02	0.02	0.02	0.01	0.02	0.02
Mg	2.07	1.73	2.23	2.12	2.27	2.07	Mg	2.50	2.49	2.52	2.37	2.40	2.60	Mg	2.52	2.51	2.27	2.53	2.45	2.37
Ca	1.97	1.86	1.84	1.86	1.86	1.97	Ca	1.84	1.88	1.92	1.92	1.88	1.89	Ca	1.85	1.89	1.81	1.88	1.85	1.84
Na	0.76	0.86	0.84	0.83	0.82	0.76	Na	0.66	0.66	0.66	0.67	0.67	0.70	Na	0.66	0.67	0.67	0.69	0.69	0.70
K	0.19	0.14	0.11	0.11	0.11	0.19	K	0.05	0.05	0.06	0.06	0.05	0.05	K	0.05	0.06	0.06	0.06	0.06	0.05
Cl	1.28	1.37	1.06	1.11	1.00	1.28	Cl	0.41	0.43	0.42	0.48	0.46	0.54	Cl	0.41	0.43	0.51	0.45	0.48	0.47

Structural formulas are calculated on the basis of 13 cations except Ca, Na and K.

Table III-9. Continued

Sample	ZZB 15	ZZB 15	ZZB 15	ZZB 15	ZZB 15	ZZB 15	Sample	ZZB 18	ZZB 18	ZZB 18	ZZB 18	ZZB 18	ZZB 18	Sample	ZZB 172	ZZB 172	ZZB 172	ZZB 172	ZZB 172	ZZB 172
analysis-no.	47	8	44	22	13	33	analysis-no.	2	3	64	11	8	20	analysis-no.	114	123	128	152	153	158
SiO ₂	41.29	41.34	42.08	41.15	40.39	40.13	SiO ₂	47.37	47.53	47.79	47.38	47.04	47.07	SiO ₂	38.48	38.72	39.22	38.01	38.42	38.81
Na ₂ O	2.33	2.39	2.47	2.61	2.67	2.67	Na ₂ O	1.18	1.20	1.20	1.25	1.30	1.35	Na ₂ O	0.96	0.93	1.06	1.13	1.09	1.03
K ₂ O	0.36	0.35	0.29	0.37	0.37	0.38	K ₂ O	0.26	0.26	0.27	0.31	0.30	0.28	K ₂ O	2.66	2.82	2.51	2.56	2.68	2.85
TiO ₂	0.99	0.93	0.89	0.77	0.81	0.72	TiO ₂	0.44	0.45	0.51	0.46	0.47	0.52	TiO ₂	2.06	1.89	1.53	1.00	1.78	1.97
FeO	14.86	14.71	14.77	15.61	15.77	15.82	FeO	9.15	9.20	8.29	8.76	9.40	9.31	FeO	18.05	17.91	18.31	16.58	16.83	17.61
Al ₂ O ₃	14.33	13.94	13.74	14.41	15.19	15.44	Al ₂ O ₃	10.34	10.06	9.89	10.44	10.80	10.51	Al ₂ O ₃	14.06	14.19	14.09	17.05	14.66	14.66
MgO	9.65	10.43	10.61	9.80	9.50	9.34	MgO	15.48	15.64	15.72	15.54	15.10	15.17	MgO	8.05	8.36	7.99	8.32	8.42	8.29
CaO	11.51	11.82	11.75	11.68	11.69	11.86	CaO	12.69	12.77	12.79	12.71	12.72	12.63	CaO	12.02	12.10	11.67	12.00	11.51	11.91
MnO	0.17	0.23	0.24	0.22	0.18	0.17	MnO	0.14	0.11	0.08	0.11	0.08	0.09	MnO	0.06	0.17	0.11	0.13	0.09	0.10
Cl	2.51	2.39	2.37	2.76	2.90	2.96	Cl	0.31	0.29	0.28	0.27	0.31	0.29	Cl	1.65	1.63	1.48	1.25	1.39	1.60
Total	98.03	98.65	99.25	99.43	99.50	99.52	Total	97.35	97.51	96.86	97.23	97.61	97.26	Total	96.46	97.18	96.52	96.82	95.50	97.27
-O equ. Cl	0.57	0.54	0.53	0.62	0.65	0.67	-O equ. Cl	0.07	0.06	0.06	0.06	0.07	0.07	-O equ. Cl	0.37	0.37	0.33	0.28	0.31	0.36
Total	97.46	98.11	98.72	98.81	98.85	98.85	Total	97.28	97.45	96.80	97.17	97.54	97.19	Total	96.09	96.81	96.19	96.54	95.19	96.91

Table III-9. Continued

Sample	ZZB 15	ZZB 15	ZZB 15	ZZB 15	ZZB 15	ZZB 15	Sample	ZZB 18	ZZB 18	ZZB 18	ZZB 18	ZZB 18	ZZB 18	Sample	ZZB 172	ZZB 172	ZZB 172	ZZB 172	ZZB 172	ZZB 172
analysis-no.	47	8	44	22	13	33	analysis-no.	2	3	64	11	8	20	analysis-no.	114	123	128	152	153	158
Si	6.25	6.21	6.27	6.18	6.08	6.06	Si	6.79	6.80	6.88	6.80	6.75	6.78	Si	5.97	5.96	6.05	5.80	5.96	5.95
Al IV	1.75	1.79	1.73	1.82	1.92	1.95	Al IV	1.21	1.20	1.12	1.20	1.25	1.23	Al IV	2.03	2.04	1.95	2.20	2.04	2.05
Ti	0.81	0.68	0.68	0.73	0.77	0.80	Ti	0.53	0.50	0.56	0.56	0.58	0.56	Ti	0.24	0.22	0.18	0.12	0.21	0.23
Al VI	0.11	0.11	0.10	0.09	0.09	0.08	Al VI	0.05	0.05	0.06	0.05	0.05	0.06	Al VI	0.54	0.53	0.61	0.86	0.64	0.60
Fe 3+	0.23	0.33	0.34	0.33	0.35	0.29	Fe 3+	0.31	0.30	0.11	0.23	0.24	0.23	Fe 3+	0.19	0.25	0.32	0.36	0.29	0.21
Fe 2+	1.65	1.52	1.50	1.64	1.64	1.70	Fe 2+	0.78	0.80	0.89	0.82	0.89	0.89	Fe 2+	2.15	2.06	2.04	1.75	1.90	2.05
Mn	0.02	0.03	0.03	0.03	0.02	0.02	Mn	0.02	0.01	0.01	0.01	0.01	0.01	Mn	0.01	0.02	0.01	0.02	0.01	0.01
Mg	2.18	2.34	2.36	2.20	2.13	2.10	Mg	3.31	3.34	3.38	3.32	3.23	3.26	Mg	1.86	1.92	1.84	1.89	1.95	1.90
Ca	1.87	1.90	1.88	1.88	1.89	1.92	Ca	1.95	1.96	1.97	1.95	1.96	1.95	Ca	2.00	1.99	1.93	1.96	1.91	1.96
Na	0.68	0.70	0.71	0.76	0.78	0.78	Na	0.33	0.33	0.34	0.35	0.36	0.38	Na	0.29	0.28	0.32	0.33	0.33	0.31
K	0.07	0.07	0.06	0.07	0.07	0.07	K	0.05	0.05	0.05	0.06	0.06	0.05	K	0.53	0.55	0.49	0.50	0.53	0.56
Cl	0.64	0.61	0.60	0.70	0.74	0.76	Cl	0.08	0.07	0.07	0.07	0.07	0.07	Cl	0.43	0.43	0.39	0.32	0.37	0.42

Structural formulas are calculated on the basis of 13 cations except Ca, Na and K.

Table III-10. Representative microprobe analyses of scapolites from samples from the Mavuradonha Layered Complex.

sample	ZZB 2	ZZB 2	ZZB 2	ZZB 2	ZZB 2	ZZB 2	sample	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	ZZB 4	sample	ZZB 5	ZZB 5	ZZB 5	ZZB 5	ZZB 5	ZZB 5
analysis-no.	131	150	169	175	182	183	analysis-no.	10	29	105	94	98	91	analysis-no.	79	86	89	102	115	126
SiO ₂	51.24	49.52	50.99	49.58	48.82	50.39	SiO ₂	52.86	52.52	52.28	53.55	52.76	52.29	SiO ₂	52.40	52.52	52.61	51.28	52.56	53.59
TiO ₂	0	0	0	0.04	0	0	TiO ₂	0	0	0.02	0	0	0.03	TiO ₂	0.00	0.00	0.00	0.00	0.04	0.00
Al ₂ O ₃	24.26	24.28	24.18	24.65	24.64	24.07	Al ₂ O ₃	23.83	24.30	23.94	23.11	23.36	24.05	Al ₂ O ₃	23.75	23.80	23.53	24.72	24.03	23.24
FeO	0.04	0.05	0.20	0	0.06	0.07	FeO	0.09	0.10	0.02	0.35	0.05	0.08	FeO	0.01	0.04	0.30	0.06	0	0.07
MnO	0.01	0.04	0.05	0	0.05	0	MnO	0.05	0.06	0	0.02	0.02	0.02	MnO	0	0.04	0.04	0	0.08	0.04
MgO	0.01	0	0.04	0.02	0.03	0.01	MgO	0	0.02	0.01	0.09	0.01	0.02	MgO	0.02	0	0.02	0	0	0.01
CaO	12.92	13.76	12.97	14.10	14.50	13.00	CaO	10.53	10.94	10.80	8.77	10.09	10.81	CaO	10.85	11.11	10.43	12.07	10.41	8.77
Na ₂ O	6.58	5.98	6.63	6.12	5.73	6.60	Na ₂ O	8.45	8.24	8.02	9.46	8.76	8.08	Na ₂ O	8.24	7.84	8.06	7.06	8.21	9.15
K ₂ O	0.31	0.28	0.28	0.29	0.31	0.32	K ₂ O	0.29	0.26	0.27	0.23	0.24	0.18	K ₂ O	0.12	0.17	0.15	0.19	0.19	0.22
Cl	2.22	2.05	2.25	1.99	1.83	2.22	Cl	3.29	3.07	3.05	3.67	3.22	3.15	Cl	2.56	2.63	2.72	2.37	2.96	3.41
Total	97.58	95.96	97.58	96.79	95.97	96.68	Total	99.38	99.51	98.41	99.24	98.51	98.71	Total	97.95	98.14	97.86	97.75	98.47	98.50
Si	7.70	7.61	7.70	7.57	7.53	7.68	Si	7.84	7.77	7.80	7.96	7.89	7.78	Si	7.82	7.82	7.86	7.66	7.80	7.94
Al	4.30	4.39	4.30	4.43	4.47	4.32	Al	4.16	4.23	4.20	4.04	4.11	4.22	Al	4.18	4.18	4.14	4.35	4.20	4.06
Ti	0	0	0	0	0	0	Ti	0	0	0	0	0	0	Ti	0	0	0	0	0	0
Fe ²⁺	0	0.01	0.03	0	0.01	0.01	Fe ²⁺	0.01	0.01	0	0.04	0.01	0.01	Fe ²⁺	0	0	0.04	0.01	0	0.01
Mn	0	0.01	0.01	0	0.01	0	Mn	0.01	0.01	0	0	0	0	Mn	0	0.01	0.01	0	0.01	0
Mg	0	0	0.01	0.01	0.01	0	Mg	0	0.01	0	0.02	0	0	Mg	0	0	0	0	0	0
Ca	2.08	2.27	2.10	2.31	2.40	2.12	Ca	1.67	1.73	1.73	1.40	1.62	1.72	Ca	1.74	1.77	1.67	1.93	1.66	1.39
Na	1.92	1.78	1.94	1.81	1.71	1.95	Na	2.43	2.36	2.32	2.73	2.54	2.33	Na	2.39	2.27	2.34	2.04	2.36	2.63
K	0.06	0.05	0.05	0.06	0.06	0.06	K	0.05	0.05	0.05	0.04	0.05	0.03	K	0.02	0.03	0.03	0.04	0.04	0.04
Cl	0.57	0.53	0.58	0.52	0.48	0.57	Cl	0.83	0.77	0.77	0.93	0.82	0.80	Cl	0.65	0.66	0.69	0.60	0.75	0.86
EqAn	43.17	46.43	43.33	47.73	49.13	44.00	EqAn	38.73	41.07	40.13	34.80	37.10	40.53	EqAn	39.20	39.20	38.00	44.87	40.00	35.23
XMe	51.28	55.36	51.73	55.29	57.67	51.46	XMe	40.49	42.17	42.14	34.50	38.57	42.29	XMe	41.90	43.63	42.02	48.25	40.99	34.42

Structural formulas are calculated on the basis of 16 cations. EqAn: equivalent anorthit (calculated after Ellis, 1978), XMe: meionit content of a scapolite (calculated after Shaw, 1960).

Table III-10. Continued.

sample	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2	ZZB 6-2	sample	ZZB 8	ZZB 8	ZZB 8	ZZB 8	ZZB 8	sample	ZZB 9	ZZB 9	ZZB 9	ZZB 9
analysis-no.	162	169	171	193	143	analysis-no.	193	195	197	219	225	analysis-no.	139	157	161	167
SiO ₂	51.22	51.53	52.08	52.86	53.67	SiO ₂	49.87	49.59	49.68	49.76	49.34	SiO ₂	50.17	50.85	50.88	52.28
TiO ₂	0.02	0	0	0	0.02	TiO ₂	0	0.04	0.01	0.03	0.02	TiO ₂	0.02	0	0	0
Al ₂ O ₃	24.35	24.15	24.23	22.74	23.78	Al ₂ O ₃	24.31	24.09	24.36	24.54	24.27	Al ₂ O ₃	24.59	24.45	24.55	24.02
FeO	0.18	0.23	0.18	0.10	0.03	FeO	0.02	0.07	0.08	0.16	0.09	FeO	0.02	0.04	0.06	0.05
MnO	0	0.02	0	0	0.02	MnO	0.01	0.05	0.02	0.03	0	MnO	0	0.03	0.02	0
MgO	0	0.02	0.02	0.02	0.02	MgO	0	0.01	0.01	0.01	0.03	MgO	0	0	0	0
CaO	11.74	11.89	11.78	8.87	9.90	CaO	13.45	13.86	14.23	13.82	14.08	CaO	12.73	12.23	12.73	10.93
Na ₂ O	7.36	7.21	7.29	8.64	8.83	Na ₂ O	6.11	6.01	5.81	6.23	5.90	Na ₂ O	6.39	6.89	6.68	7.51
K ₂ O	0.17	0.15	0.19	0.17	0.14	K ₂ O	0.58	0.55	0.57	0.59	0.49	K ₂ O	0.50	0.61	0.55	0.64
Cl	2.48	2.33	2.41	3.42	3.18	Cl	1.79	1.76	1.66	1.74	1.60	Cl	2.01	2.17	2.09	2.6
Total	97.51	97.52	98.19	96.82	99.57	Total	96.15	96.02	96.42	96.91	95.82	Total	96.44	97.26	97.56	98.03
Si	7.69	7.73	7.75	7.96	7.89	Si	7.62	7.63	7.61	7.59	7.60	Si	7.61	7.66	7.65	7.79
Al	4.31	4.27	4.25	4.03	4.11	Al	4.38	4.37	4.39	4.41	4.40	Al	4.39	4.34	4.35	4.21
Ti	0	0	0	0	0	Ti	0	0.01	0	0	0	Ti	0	0	0	0
Fe ²⁺	0.02	0.03	0.02	0.01	0	Fe ²⁺	0	0.01	0.01	0.02	0.01	Fe ²⁺	0	0	0.01	0.01
Mn	0	0	0	0	0	Mn	0	0.01	0	0	0	Mn	0	0	0	0
Mg	0	0	0.01	0.01	0	Mg	0	0	0	0	0.01	Mg	0	0	0	0
Ca	1.89	1.91	1.88	1.43	1.56	Ca	2.20	2.29	2.33	2.26	2.32	Ca	2.07	1.97	2.05	1.74
Na	2.14	2.10	2.10	2.52	2.52	Na	1.81	1.79	1.73	1.84	1.76	Na	1.88	2.01	1.95	2.17
K	0.03	0.03	0.04	0.03	0.03	K	0.11	0.11	0.11	0.12	0.10	K	0.10	0.12	0.11	0.12
Cl	0.63	0.59	0.61	0.87	0.79	Cl	0.46	0.46	0.43	0.45	0.42	Cl	0.52	0.55	0.53	0.66
EqAn	43.57	42.23	41.57	34.47	37.13	EqAn	45.90	45.53	46.43	46.97	46.73	EqAn	46.40	44.63	44.90	40.47
XMe	46.77	47.71	47.11	36.18	38.02	XMe	53.37	54.80	56.09	53.77	55.75	XMe	51.20	48.21	50.09	43.31

Structural formulas are calculated on the basis of 16 cations. EqAn: equivalent anorthit (calculated after Ellis, 1978), XMe: meionit content of a scapolite (calculated after Shaw, 1960).

Table III-10. Continued.

sample	ZZB 12	ZZB 12	ZZB 12	ZZB 12	ZZB 12	ZZB 12	sample	ZZB 15	ZZB 15	ZZB 15	ZZB 15	ZZB 15	ZZB 15	sample	ZZB 18	ZZB 18	ZZB 18	ZZB 18
analysis-no.	4	16	24	25	26	11	analysis-no.	17	51	53	57	29	14	analysis-no.	18	32	33	51
SiO ₂	52.65	52.72	52.39	52.21	52.72	52.05	SiO ₂	49.45	50.65	50.91	52.77	51.38	52.05	SiO ₂	51.62	51.18	50.67	50.23
TiO ₂	0.02	0	0.01	0.01	0	0	TiO ₂	0.00	0.03	0.00	0.01	0.02	0.00	TiO ₂	0	0	0	0
Al ₂ O ₃	24.00	23.55	23.52	23.19	23.24	24.21	Al ₂ O ₃	25.26	24.72	24.57	24.07	24.54	24.13	Al ₂ O ₃	24.29	24.04	23.07	24.65
FeO	0.04	0	0.49	0.09	0.08	0.07	FeO	0.01	0.03	0	0.08	0.01	0.02	FeO	0.03	0.04	0.42	0.01
MnO	0	0.01	0	0	0.04	0.01	MnO	0	0.05	0.01	0.04	0.01	0	MnO	0	0.02	0.01	0
MgO	0.02	0.02	0.02	0.01	0.01	0.02	MgO	0.02	0.01	0	0	0.01	0.03	MgO	0.03	0.00	0.02	0.04
CaO	10.81	10.44	10.81	11.01	10.55	11.31	CaO	14.09	12.76	12.58	10.87	12.55	11.05	CaO	11.96	11.86	11.46	12.85
Na ₂ O	7.92	7.23	6.86	7.92	8.27	7.54	Na ₂ O	5.90	6.60	6.72	8.06	6.79	7.75	Na ₂ O	7.30	7.10	7.38	6.41
K ₂ O	0.22	0.27	0.20	0.20	0.23	0.18	K ₂ O	0.19	0.26	0.20	0.24	0.22	0.22	K ₂ O	0.53	0.50	0.54	0.42
Cl	2.95	2.91	3.07	3.01	3.18	2.86	Cl	1.85	2.22	2.30	2.82	2.26	2.74	Cl	2.00	1.84	2.01	1.57
Total	98.63	97.15	97.37	97.65	98.32	98.24	Total	96.77	97.33	97.29	98.95	97.79	97.99	Total	97.75	96.58	95.58	96.18
Si	7.81	7.86	7.85	7.88	7.90	7.75	Si	7.49	7.62	7.65	7.81	7.68	7.76	Si	7.72	7.73	7.81	7.61
Al	4.19	4.14	4.15	4.12	4.10	4.25	Al	4.51	4.38	4.35	4.19	4.32	4.24	Al	4.28	4.27	4.19	4.40
Ti	0	0	0	0	0	0	Ti	0	0	0	0	0	0	Ti	0	0	0	0
Fe ₂	0.01	0	0.06	0.01	0.01	0.01	Fe ₂ ⁺	0	0	0	0.01	0	0	Fe ₂ ⁺	0	0.01	0.05	0
Mn	0	0	0	0	0.01	0	Mn	0	0.01	0	0.01	0	0	Mn	0	0	0	0
Mg	0	0	0.01	0	0	0	Mg	0.01	0	0	0	0	0.01	Mg	0.01	0	0	0.01
Ca	1.72	1.67	1.74	1.78	1.69	1.81	Ca	2.29	2.06	2.03	1.72	2.01	1.77	Ca	1.92	1.92	1.89	2.08
Na	2.28	2.09	1.99	2.32	2.40	2.18	Na	1.73	1.93	1.96	2.31	1.97	2.24	Na	2.12	2.08	2.21	1.88
K	0.04	0.05	0.04	0.04	0.04	0.03	K	0.04	0.05	0.04	0.05	0.04	0.04	K	0.10	0.10	0.11	0.08
Cl	0.74	0.74	0.78	0.77	0.81	0.72	Cl	0.48	0.57	0.59	0.71	0.57	0.69	Cl	0.51	0.47	0.53	0.40
EqAn	39.73	37.87	38.33	37.37	36.67	41.57	EqAn	50.23	46.00	44.97	39.80	44.00	41.27	EqAn	42.63	42.47	39.63	46.50
XMe	42.71	43.90	47.07	43.27	41.17	45.13	XMe	56.43	51.10	50.36	42.44	50.00	43.74	XMe	46.44	46.95	45.71	51.59

Structural formulas are calculated on the basis of 16 cations. EqAn: equivalent anorthit (calculated after Ellis, 1978), XMe: meionit content of a scapolite (calculated after Shaw, 1960).

Table III-11. Results of thermobarometric calculations for peak granulite-facies metamorphism.

<i>sample</i>	<i>E/E&G</i>	<i>P&H/E&G</i>	<i>N&P/E&G</i>	<i>E/K</i>	<i>P&H/K</i>	<i>N&P/K</i>	<i>TWQ</i>
	<i>P [kbar] and T [° C]</i>						
ZZB 145-13	14.7/848	11.9/840	10.8/836	14.4/817	11.6/807	10.5/804	16.2/855
ZZB 147-9	15.5/858	12.6/849	11.2/845	15.1/824	12.3/814	10.9/814	16.5/877
ZZB 152-14	13.2/831	11.0/833	10.1/841	12.8/800	10.7/793	9.8/789	15.6/855
ZZB 167-12	14.8/840	11.9/830	10.5/824	14.2/780	11.4/772	10.0/768	13.3/869
ZZB 175-11	15.3/845	12.2/841	10.6/835	14.8/807	11.7/798	10.2/793	15.9/874

E: Eckart et al. (1991); N&P: Newton & Perkins (1982); P&H: Powell & Holland (1988); K: Krogh (1988); E&G: Ellis & Green (1979). Sample numbers are representative for the estimated conditions of the sample.

Table III-12. Results of thermobarometric calculations for retrograde amphibolite-facies metamorphism.

<i>sample</i>	<i>G&P</i>	<i>K&S mg 1</i>	<i>K&S fe 1</i>	<i>K&S mg 2</i>	<i>K&S fe 2</i>	<i>TWQ</i>
	<i>T [° C]</i>	<i>P [kbar]</i>				
ZZB 7-45	688	8.3	9.0	9.3	10.0	9.5/694
ZZB 7-52	669	8.3	9.2	9.3	10.2	9.2/970
ZZB 28-79	709	10.6	12.0	12.3	12.6	12.0/710
ZZB 166-156	707	9.9	10.5	11.1	11.7	11.4/710

G&P: Graham & Powell (1984); K&S: Kohn & Spear (1989); mg 1, mg 2 and fe2 refer to the models for barometric calculations which were published by Kohn & Spear (1989). Sample numbers are representative for the estimated conditions of the sample and also refer to the microprobe analyses.

A-IV Geochemistry

Table IV-1. Whole rock geochemistry for Mavuradonha Layered Complex samples analyzed by XRF.

<i>sample</i>	<i>MAV 1</i>	<i>ZZB 5</i>	<i>ZZB 8</i>	<i>ZZB 9</i>	<i>ZZB 10</i>	<i>ZZB 11</i>	<i>ZZB 12</i>	<i>ZZB 14</i>	<i>ZZB 18</i>	<i>ZZB 54</i>	<i>ZZB 65</i>	<i>ZZB 66</i>
SiO ₂	46.77	51.14	51.91	52.21	51.37	51.26	51.30	50.35	50.39	47.02	49.30	48.75
Al ₂ O ₃	14.10	16.08	16.94	17.28	15.75	16.33	15.75	12.20	16.54	17.36	13.41	16.63
Fe ₂ O ₃ calc.	1.88	1.99	1.73	1.72	1.36	1.23	1.17	1.89	1.16	1.44	4.35	0.93
FeO meas.	6.26	6.07	5.48	5.82	6.79	4.96	4.29	4.82	3.67	5.05	10.58	5.15
MgO	15.18	8.12	8.44	8.11	9.27	10.04	9.07	12.03	9.62	13.06	6.68	11.60
CaO	13.16	13.06	12.41	11.79	12.98	13.64	15.56	16.48	16.22	13.98	11.52	15.27
Na ₂ O	1.97	2.80	2.31	2.32	1.70	1.93	2.12	1.52	1.89	1.59	1.82	1.28
K ₂ O	0.37	0.15	0.29	0.30	0.26	0.19	0.11	0.29	0.17	0.15	0.22	0.09
TiO ₂	0.16	0.42	0.30	0.28	0.35	0.26	0.51	0.30	0.26	0.20	1.79	0.20
P ₂ O ₅	nd	nd	0.01	0.01	0.01	nd	nd	nd	0.01	0.01	0.12	nd
MnO	0.14	0.16	0.17	0.16	0.15	0.14	0.12	0.13	0.07	0.13	0.19	0.09
LOI	0.5	0.41	0.81	0.85	0.86	0.94	0.27	1.06	0.73	0.75	0.55	0.96
Fe ₂ O ₃ meas.	8.6	8.55	7.65	8	8.67	6.59	5.84	7.05	5.13	6.87	15.79	6.52
Mg#	0.71	0.57	0.61	0.58	0.58	0.67	0.68	0.71	0.72	0.72	0.39	0.69
Sc	37	49	47	44	52	47	50	61	52	30	51	43
V	117	229	180	172	206	178	196	221	200	109	460	147
Cr	1409	229	71	171	598	402	481	1323	1146	690	77	1153
Co	47	35	36	37	38	37	29	38	31	44	54	45
Ni	271	98	76	90	115	109	127	174	154	280	69	235
Cu	81	44	88	126	154	30	43	56	78	17	51	64
Zn	46	52	63	69	77	43	46	43	17	38	74	33
Ga	10	16	16	16	14	13	15	10	13	11	20	13
Rb	7	4	6	6	7	9	6	6	3	4	8	4
Sr	41	148	219	281	244	88	147	89	178	136	137	174
Y	9	12	10	8	11	7	16	11	9	7	34	7
Zr	7	12	15	9	15	10	18	12	11	10	79	8
Nb	2	1	2	2	2	2	1	2	2	1	8	2
Ba	42	55	77	76	126	45	67	32	23	40	39	29

Table IV-1. Continued.

<i>sample</i>	<i>MAV 1</i>	<i>ZZB 5</i>	<i>ZZB 8</i>	<i>ZZB 9</i>	<i>ZZB 10</i>	<i>ZZB 11</i>	<i>ZZB 12</i>	<i>ZZB 14</i>	<i>ZZB 18</i>	<i>ZZB 54</i>	<i>ZZB 65</i>	<i>ZZB 66</i>
La	1	2	1	I.L.	1	1	1	2	I.L.	1	3	1
Ce	2	8	6	12	11	I.L.	4	1	I.L.	1	21	1
Pr	1	1	2	1	I.L.	I.L.	1	I.L.	1	1	2	I.L.
Nd	1	4	3	3	3	1	2	I.L.	1	2	8	1
Sm	1	1	I.L.	1	1	1	2	3	1	1	4	I.L.
CIPW												
Q	0	0	0.7	3.79	1.2	0	0	0	0	3.5	12.6	0
Or	0.7	0.4	0.7	0.7	0.6	0.5	0.3	0.7	0.5	1.8	0.5	0.2
Ab	2.6	11.2	8.2	8.0	5.7	6.9	8.3	5.8	7.7	10.5	5.8	4.7
An	17.1	27.7	27.7	27.5	25.7	28.0	28.9	21.9	32.7	20.8	19.7	32.3
Ne	2.7	0	0	0	0	0	0	0	0	0	0	0
Lc	0	0	0	0	0	0	0	0	0	0	0	0
Di	21.9	30.3	20.8	17.2	22.0	25.5	38.3	47.8	39.8	9.3	20.1	29.6
Hv	0	15.5	38.7	39.9	42.3	31.4	14.3	4.9	3.3	38.3	31.6	12.3
Ol	52.8	10.5	0	0	0	5.4	6.5	15.2	13.5	0	0	19.0
Mt	1.96	3.1	2.4	2.3	1.8	1.7	1.8	2.8	1.8	6.2	5.4	1.3
Hm	0	0	0	0	0	0	0	0	0	0	0	0
Il	0.3	1.3	0.8	0.7	0.9	0.7	1.5	0.9	0.8	7.7	4.4	0.6
Ap	0	0	0	0	0	0	0	0	0	1.8	0.2	0

I.L. = lower limit; nd = not detected

Table IV-1. Continued.

sample	ZZB 67	ZZB 68	ZZB 69	ZZB 70	ZZB 100	ZZB 101	ZZB 111	ZZB 146	ZZB 147	ZZB 148	ZZB 149	ZZB 150
SiO ₂	48.25	49.38	47.07	49.64	49.50	47.67	47.17	53.16	49.93	53.71	48.74	52.42
Al ₂ O ₃	18.61	29.14	18.85	12.81	17.83	8.32	20.03	24.73	15.47	23.91	15.65	28.32
Fe ₂ O ₃ calc.	1.82	0.62	5.54	2.50	1.33	0.00	5.73	0.87	3.60	1.41	2.85	0.28
FeO meas.	4.53	1.03	2.09	5.06	3.44	7.73	1.84	2.70	8.54	3.18	9.50	1.07
MgO	8.98	2.17	12.01	13.21	8.07	11.25	11.57	2.84	6.19	1.50	6.57	0.75
CaO	14.39	14.91	12.49	14.96	16.90	21.70	10.71	10.44	11.39	9.36	11.89	12.40
Na ₂ O	2.74	2.55	1.50	1.24	1.98	1.08	2.44	4.40	2.60	5.12	2.56	4.24
K ₂ O	0.32	0.09	0.14	0.12	0.10	0.04	0.15	0.41	0.53	0.62	0.44	0.28
TiO ₂	0.29	0.08	0.12	0.21	0.76	1.17	0.22	0.36	1.35	1.11	1.38	0.19
P ₂ O ₅	nd	nd	nd	nd	0.02	nd	0.02	0.02	0.18	0.02	0.19	0.02
MnO	0.08	0.03	0.18	0.25	0.08	0.17	0.11	0.06	0.21	0.06	0.22	0.03
LOI	1.03	0.46	1.21	0.97	0.1	0.58	0.08	0.07	0.1	0.02	0.1	0.17
Fe ₂ O ₃ meas.	6.61	1.73	7.71	7.95	5.08	8.59	7.69	3.86	12.86	4.9	13.22	1.46
Mg#	0.66	0.68	0.85	0.72	0.70	1.00	0.86	0.51	0.42	0.32	0.41	0.41
Sc	37	14	27	58	47	51	15	10	44	13	43	8
V	154	34	77	192	150	169	37	51	310	123	309	25
Cr	498	81	418	1067	649	227	244	23	137	18	130	10
Co	37	7	57	42	21	21	55	19	44	18	43	3
Ni	183	31	367	181	91	59	199	50	62	23	65	13
Cu	56	16	132	48	40	-	10	4	133	15	68	1
Zn	29	16	31	45	21	-	42	32	98	41	94	18
Ga	15	19	14	10	14	12	14	21	19	24	18	18
Rb	9	4	6	4	2	2	3	3	5	4	3	3
Sr	216	149	94	54	521	76	737	641	219	699	253	659
Y	10	3	5	11	11	54	4	5	34	6	35	3
Zr	11	5	8	9	24	115	3	10	109	12	92	8
Nb	2	1	1	1	2	8	2	2	6	2	5	2
Ba	48	53	47	42	50	l.l.	97	115	152	212	137	107

Table IV-1. Continued.

<i>sample</i>	<i>ZZB 67</i>	<i>ZZB 68</i>	<i>ZZB 69</i>	<i>ZZB 70</i>	<i>ZZB 100</i>	<i>ZZB 101</i>	<i>ZZB 111</i>	<i>ZZB 146</i>	<i>ZZB 147</i>	<i>ZZB 148</i>	<i>ZZB 149</i>	<i>ZZB 150</i>
La	1	I.I.	I.I.	I.I.	1	12	2	4	9	1	7	2
Ce	4	I.I.	5	I.I.	I.I.	56	1	1	25	I.I.	17	I.I.
Pr	1	1	I.I.	2	3	5	1	1	4	1	4	I.I.
Nd	3	2	5	1	4	36	1	2	14	4	13	4
Sm	1	1	2	2	4	8	2	3	6	3	5	2
CIPW												
Q	0	0	0	0	0	0	0	0	2.2	0	0	0
Or	0.7	0.3	0.3	0.3	0.3	0	0.4	1.3	1.3	2.2	1.1	1.0
Ab	5.7	11.7	5.2	4.2	6.8	0	9.2	21.3	9.6	27.1	10.1	21.5
An	28.9	69.8	34.1	22.2	33.6	14.3	36.4	50.5	24.1	47.9	26.3	66.4
Ne	3.8	0	0	0	0.7	3.9	0	0	0	0	0	0.4
Lc	0	0	0	0	0	0.1	0	0	0	0	0	0
Di	25.8	6.1	13.5	34.4	37.5	62.9	8.1	5.3	21.6	6.7	24.3	4.3
Hv	0	9.5	28.3	23.9	0	0	15.0	12.8	31.9	4.7	19.4	0
Ol	31.9	1.2	11.0	11.1	16.9	15.6	22.0	5.7	0	3.9	9.9	5.0
Mt	2.4	1.1	5.9	3.3	2.0	0	5.3	1.7	5.2	2.9	4.4	0.6
Hm	0	0	1.5	0	0	0	3.1	0	0	0	0	0
Il	0.8	0.3	0.3	0.6	2.3	3.3	0.7	1.4	3.9	4.6	4.2	0.8
Ap	0	0	0	0	0	0	0.0	0	0.3	0.1	0.3	0.1

I.I. = lower limit; nd = not detected

Table IV-1. Continued.

<i>sample</i>	<i>ZZB 160</i>	<i>ZZB 161</i>	<i>ZZB 162</i>	<i>ZZB 164</i>	<i>ZZB 166</i>	<i>ZZB 215</i>	<i>ZZB 216</i>	<i>ZZB 220</i>	<i>ZZB 224</i>	<i>ZZB 225</i>	<i>ZZB 231</i>	<i>ZZB 270</i>
SiO ₂	48.33	50.58	48.38	49.74	47.70	51.37	51.22	51.80	48.70	53.58	48.58	51.18
Al ₂ O ₃	32.03	13.30	31.65	25.13	13.54	15.45	16.58	16.22	28.09	14.26	13.18	15.00
Fe ₂ O ₃ calc.	0.47	4.27	0.44	0.98	3.72	1.40	3.48	1.13	0.88	5.68	0.00	3.31
FeO meas.	0.17	10.50	0.31	2.75	12.17	6.08	4.08	6.05	2.10	1.69	13.09	7.20
MgO	0.28	5.92	0.50	4.31	7.20	9.87	9.46	9.26	2.08	4.85	6.64	8.14
CaO	16.11	11.11	16.06	14.34	11.01	13.52	12.93	12.65	14.41	9.25	10.61	12.00
Na ₂ O	2.49	1.38	2.51	2.40	2.08	1.76	1.73	2.37	3.38	5.91	3.51	2.10
K ₂ O	0.04	0.53	0.05	0.12	0.48	0.04	0.13	0.05	0.17	0.24	0.59	0.39
TiO ₂	0.06	2.10	0.08	0.17	1.72	0.35	0.24	0.31	0.14	4.49	2.06	0.46
P ₂ O ₅	nd	0.15	nd	nd	0.15	nd	nd	0.02	0.01	0.02	0.16	0.02
MnO	0.01	0.15	0.01	0.05	0.21	0.15	0.15	0.15	0.03	0.05	0.13	0.20
LOI	0.38	0.9	0.46	0.5	0.39	0.22	0.76	0.03	0.1	0.13	0.91	0.76
Fe ₂ O ₃ meas.	0.66	15.51	0.78	3.98	16.85	8.04	7.8	7.8	3.15	7.51	14.55	11.11
Mg#	0.62	0.36	0.62	0.61	0.37	0.62	0.70	0.60	0.50	0.74	1.00	0.53
Sc	10	51	10	27	50	48	46	42	16	34	48	58
V	9.5	492	17	88	433	210	168	178	54	158	411	268
Cr	2	100	3	177	101	414	594	144	25	18	178	165
Co	l.l.	42	3	19	55	42	29	40	11	19	44	48
Ni	2.5	42	7	52	73	111	140	94	29	30	71	85
Cu	0	42	1	7	89	52	45	16	2	3	24	188
Zn	5.5	35	7	14	99	53	34	52	15	21	62	76
Ga	17.5	21	19	17	22	15	13	15	17	19	20	16
Rb	1.5	7	2	3	13	3	4	4	4	4	7	14
Sr	278.5	60	282	283	104	74	137	97	164	204	146	100
Y	1.5	46	3	4	36	11	8	11	5	28	48	21
Zr	4	100	5	11	109	13	9	14	20	101	113	35
Nb	1.5	10	2	2	10	2	2	2	2	19	4	2
Ba	18.5	16	18	21	51	21	37	51	24	l.l.	25	69

Table IV-1. Continued.

<i>sample</i>	<i>ZZB 160</i>	<i>ZZB 161</i>	<i>ZZB 162</i>	<i>ZZB 164</i>	<i>ZZB 166</i>	<i>ZZB 215</i>	<i>ZZB 216</i>	<i>ZZB 220</i>	<i>ZZB 224</i>	<i>ZZB 225</i>	<i>ZZB 231</i>	<i>ZZB 270</i>
La	I.I.	3	1	1	8	I.I.	3	1	2	2	14	3
Ce	I.I.	23	I.I.	1	26	3	3	3	1	5	28	9
Pr	0.5	3	I.I.	1	3	1	1	1	2	6	2	2
Nd	1	10	2	2	11	2	4	1	4	7	10	5
Sm	3	3	3	3	4	1	2	2	3	4	3	1
CIPW												
Q	0	24.4	0	0	0	0.3	9.1	0	0	-	-	-
Or	0.13	1.0	0.2	0.3	1.1	0.1	0.3	0.1	0.5	-	-	-
Ab	11.73	4.0	11.5	10.2	7.4	5.9	5.6	8.4	11.1	-	-	-
An	84.26	18.3	82.4	54.4	20.8	25.5	26.7	26.6	61.5	-	-	-
Ne	0.59	0	0.8	0	0	0	0	0	4.2	-	-	-
Lc	0	0	0	0	0	0	0	0	0	-	-	-
Di	2.15	16.3	4.0	12.8	21.7	24.5	19.4	23.0	10.3	-	-	-
Hv	0	26.4	0	17.1	32.6	41.1	34.0	35.1	0	-	-	-
Ol	0	0	0	3.1	6.3	0	0	4.3	10.3	-	-	-
Mt	0.5	4.8	0.8	1.6	5.1	1.8	4.4	1.6	1.5	-	-	-
Hm	0.41	0	0	0	0	0	0	0	0	-	-	-
Il	0.23	4.7	0.3	0.6	4.7	0.9	0.6	0.9	0.5	-	-	-
Ap	0	0.2	0	0	0.2	0	0	0	0	-	-	-

I.I. = lower limit; nd = not detected

Table IV-2. Whole rock geochemistry for Nyamhanda Inlier (DKZ) and Chimwaya Hill Inlier (ZCW) samples analyzed by XRF.

<i>sample</i>	<i>DKZ 1</i>	<i>DKZ 2</i>	<i>DKZ 4</i>	<i>DKZ 5</i>	<i>DKZ 6</i>	<i>DKZ 11</i>	<i>DKZ 12</i>	<i>DKZ 13</i>	<i>DKZ 16</i>	<i>DKZ 18</i>	<i>DKZ 20</i>
SiO ₂	48.33	51.46	46.65	43.72	45.89	47.59	43.63	49.29	51.19	51.63	54.26
Al ₂ O ₃	13.79	17.22	14.63	17.67	24.81	13.00	18.97	14.11	14.15	16.65	13.19
Fe ₂ O ₃ calc.	3.34	1.90	2.53	8.03	0.97	4.68	3.40	2.79	4.12	1.69	3.41
FeO meas.	10.03	4.39	11.86	3.60	5.47	10.75	9.81	9.33	9.15	5.59	9.32
MgO	6.50	4.49	5.66	13.90	7.45	5.73	9.02	6.22	6.57	8.90	4.53
CaO	10.98	12.17	9.78	10.64	12.52	10.13	10.00	10.71	10.43	13.10	11.44
Na ₂ O	2.84	3.80	3.23	1.55	2.18	2.80	3.01	3.21	2.18	1.76	0.75
K ₂ O	0.72	0.29	1.16	0.17	0.25	0.61	0.24	0.92	0.58	0.20	0.52
TiO ₂	2.82	2.85	3.63	0.48	0.29	3.87	1.75	2.81	1.31	0.30	2.08
P ₂ O ₅	0.37	1.34	0.64	0.03	0.05	0.58	0.02	0.38	0.11	0.01	0.23
MnO	0.27	0.10	0.24	0.21	0.13	0.24	0.15	0.23	0.21	0.16	0.27
LOI	0.5	0.71	0.08	0.99	0.38	0.04	0.16	-0.23	-0.18	0.85	0.23
Fe ₂ O ₃ meas.	14.04	6.63	15.27	11.76	6.89	16.33	14.14	12.96	14.12	7.76	13.5
Mg#	0.39	0.51	0.32	0.79	0.58	0.35	0.48	0.40	0.42	0.61	0.33
Sc	37	50	29	11	11	39	17	33	38	39	40
V	328	41	337	72	35	409	336	346	367	145	446
Cr	136	I.I.	18	122	6	82	328	114	144	53	1
Co	44	14	45	83	43	39	70	43	55	45	39
Ni	67	9	37	224	96	31	117	50	107	133	45
Cu	48	I.I.	35	25	15	17	72	35	197	8	39
Zn	177	40	153	90	65	193	96	118	110	51	67
Ga	21	27	27	16	20	24	22	24	20	17	22
Rb	19	4	23	4	5	13	5	20	25	10	13
Sr	401	759	584	247	875	375	693	492	139	181	269
Y	31	74	43	4	3	38	4	30	27	8	41
Zr	212	402	236	18	22	306	12	147	96	18	182
Nb	30	66	43	4	5	36	4	32	6	3	13
Ba	314	115	671	132	192	326	128	446	154	77	116

Table IV-2. Continued.

<i>sample</i>	<i>DKZ 1</i>	<i>DKZ 2</i>	<i>DKZ 4</i>	<i>DKZ 5</i>	<i>DKZ 6</i>	<i>DKZ 11</i>	<i>DKZ 12</i>	<i>DKZ 13</i>	<i>DKZ 16</i>	<i>DKZ 18</i>	<i>DKZ 20</i>
La	23	80	43	2	5	30	3	25	10	1	23
Ce	58	191	100	13	12	76	14	59	35	4	60
Pr	12	26	14	I.I.	1	12	1	11	6	0	9
Nd	31	117	58	4	5	36	4	31	16	2	24
Sm	5	26	11	1	4	9	I.I.	9	3	3	6
CIPW											
Q	0	13.5	0	0	0	5.5	0	0	14.9	5.0	43.9
Or	1.9	0.8	2.7	0.4	0.5	1.5	0.5	2.5	1.2	0.4	0.8
Ab	11.2	15.1	10.9	5.5	5.7	10.1	5.5	13.4	7.0	5.8	1.8
An	20.1	25.9	17.5	31.9	40.8	17.0	24.8	20.0	19.4	27.3	16.9
Ne	0	0	0.6	0	1.2	0	3.4	0	0	0	0
Lc	0	0	0	0	0	0	0	0	0	0	0
Di	25.7	19.9	17.7	9.1	3.1	20.3	7.8	27.2	16.9	20.6	13.0
Hv	19.2	10.9	0	16.0	0	27.5	0	12.0	32.2	37.8	16.3
Ol	7.5	0	36.2	24.9	46.6	0.0	50.2	10.4	0	0.0	0
Mt	5.1	2.9	3.5	9.6	1.2	6.5	3.9	4.5	5.1	2.2	3.2
Hm	0	0	0	1.3	0	0	0	0	0	0	0
Il	8.6	8.8	10.0	1.3	0.7	10.8	4.0	9.1	3.2	0.8	3.9
Ap	0.6	2.3	1.0	0.1	0.1	0.9	0	0.7	0.2	0	0.3

I.I. = lower limit; nd = not detected

Table IV-2. Continued.

<i>sample</i>	<i>DKZ 22</i>	<i>DKZ 25</i>	<i>DKZ 53</i>	<i>DKZ 54</i>	<i>DKZ 57</i>	<i>DKZ 60</i>	<i>sample</i>	<i>ZCW 2</i>	<i>ZCW 3</i>	<i>ZCW 6</i>	<i>ZCW 7</i>
SiO ₂	51.26	48.93	49.99	46.38	48.25	43.76	SiO ₂	46.30	45.80	46.42	47.28
Al ₂ O ₃	15.45	15.72	17.87	14.78	13.15	15.16	Al ₂ O ₃	22.93	23.07	22.92	15.11
Fe ₂ O ₃ calc.	0.81	1.84	4.03	4.37	9.28	4.71	Fe ₂ O ₃ calc.	1.69	1.63	1.81	3.36
FeO meas.	6.15	6.57	9.37	15.61	7.87	12.57	FeO meas.	5.94	4.53	5.98	10.42
MgO	11.47	10.68	2.40	2.10	4.23	7.31	MgO	7.52	7.21	8.66	6.02
CaO	13.11	14.07	8.07	8.86	8.08	11.09	CaO	11.66	14.27	11.55	9.10
Na ₂ O	1.20	1.56	4.56	2.86	2.85	2.41	Na ₂ O	2.87	2.63	2.02	3.15
K ₂ O	0.11	0.15	0.77	0.75	1.62	0.18	K ₂ O	0.39	0.26	0.15	1.46
TiO ₂	0.27	0.34	1.96	2.69	3.97	2.59	TiO ₂	0.52	0.46	0.33	3.42
P ₂ O ₅	0.01	nd	0.67	1.10	0.49	0.03	P ₂ O ₅	0.03	0.03	0.02	0.47
MnO	0.15	0.14	0.30	0.50	0.21	0.18	MnO	0.15	0.10	0.13	0.20
LOI	0.06	0.62	-0.47	-0.70	-0.92	-0.47	LOI	0.28	0.72	0.18	0.27
Fe ₂ O ₃ meas.	7.61	8.96	14.2	21.06	17.89	18.44	Fe ₂ O ₃ meas.	8.06	6.55	8.38	14.6
Mg#	0.65	0.62	0.20	0.12	0.35	0.37	Mg#	0.56	0.61	0.59	0.37
Sc	38	41	22	40	28	34	Sc	11	27	11	27
V	132	152	11	16	385	310	V	45	87	41	332
Cr	471	875	n.d.	l.l.	n.d.	129	Cr	50	136	51	50
Co	49	54	15	15	48	65	Co	41	36	48	51
Ni	206	223	5	4	46	116	Ni	109	98	85	55
Cu	37	30	n.d.	l.l.	69	n.d.	Cu	30	1	8	27
Zn	47	64	149	223	166	136	Zn	77	51	64	131
Ga	16	17	30	28	29	24	Ga	20	18	20	25
Rb	7	4	17	15	60	5	Rb	24	4	3	40
Sr	127	237	985	667	297	332	Sr	1271	601	737	521
Y	7	8	36	52	52	15	Y	5	6	4	33
Zr	17	10	115	130	295	38	Zr	15	20	10	261
Nb	2	2	49	60	41	5	Nb	6	4	4	37
Ba	69	26	1083	673	483	133	Ba	341	23	121	367

Table IV-2. Continued.

<i>sample</i>	<i>DKZ 22</i>	<i>DKZ 25</i>	<i>DKZ 53</i>	<i>DKZ 54</i>	<i>DKZ 57</i>	<i>DKZ 60</i>	<i>sample</i>	<i>ZCW 2</i>	<i>ZCW 3</i>	<i>ZCW 6</i>	<i>ZCW 7</i>
La	3	I.I.	37	41	54	3	La	9	4	2	35
Ce	3	4	89	114	125	26	Ce	14	7	5	90
Pr	nd	I.I.	11	18	16	3	Pr	1	I.I.	I.I.	12
Nd	3	2	53	71	57	7	Nd	10	4	1	46
Sm	3	1	9	17	9	3	Sm	1	I.I.	2	9
CIPW							CIPW				
Q	0.7	0	0	-	25.5	0	Q	0	0	0	0
Or	0.2	0.4	2.4	-	3.7	0.4	Or	0.8	0.6	0.4	4.2
Ab	3.7	5.7	21.0	-	9.9	7.0	Ab	6.5	4.0	7.9	13.7
An	25.0	28.7	26.9	-	14.2	21.4	An	35.4	38.7	46.4	22.2
Ne	0	0	0	-	0	0.7	Ne	2.9	5.0	0	0
Lc	0	0	0	-	0	0	Lc	0	0	0	0
Di	19.4	27.6	9.8	-	14.3	17.6	Di	7	15	3	19
Hy	49.4	19.3	18.4	-	8.6	0.0	Hy	0	0	10.9	2.8
Ol	0.0	14.8	6.0	-	0	40.7	Ol	44.4	32.8	27.1	20.4
Mt	1.0	2.6	7.2	-	12.5	5.8	Mt	2.1	2.2	2.7	5.7
Hm	0	0	0	-	0	0	Hm	0	0	0	0
Il	0.7	1.0	7.0	-	10.7	6.4	Il	1.3	1.2	1.0	11.5
Ap	0	0	1.4	-	0.8	0	Ap	0	0.1	0	0.9

I.I. = lower limit; nd = not detected

A-V Geochronology and cathodoluminescence

Table V-1. Results of vapour transfer U-Pb isotopic analyses of zircons of metagabbroic rocks from the Mavuradonha Layered Complex.

sample no.		U/Pb^*	$^{206}Pb/^{204}Pb^m$	$^{207}Pb/^{206}Pb^m$	$^{206}Pb/^{238}U^x$	$^{207}Pb/^{235}U^x$	$^{207}Pb/^{206}Pb^x$	$^{206}Pb/^{238}U$ age [Ma]	$^{207}Pb/^{235}U$ age [Ma]	$^{207}Pb/^{206}Pb$ age [Ma]	Correl coeff.
ZZB 123-6	c	7.41	155 ± 2	0.15686 ± 29	0.1080 ± 12	0.949 ± 44	0.6371 ± 29	661 ± 7	677 ± 23	732 ± 100	0.20
ZZB 123-7	c	7.34	189 ± 3	0.14069 ± 40	0.1075 ± 10	1.010 ± 50	0.0682 ± 2	657 ± 6	709 ± 25	875 ± 63	0.17
ZZB 123-8	c	4.90	1212 ± 16	0.10308 ± 16	0.1578 ± 17	1.995 ± 29	0.0917 ± 4	945 ± 10	1114 ± 10	1459 ± 8	0.93
ZZB 123-10	c	6.50	315 ± 5	0.12891 ± 22	0.1278 ± 11	1.491 ± 31	0.0847 ± 13	776 ± 6	927 ± 13	1307 ± 30	0.47
ZZB 123-11	c	6.41	376 ± 11	0.10683 ± 30	0.1293 ± 10	1.231 ± 35	0.0691 ± 17	784 ± 6	815 ± 16	900 ± 51	0.25
ZZB 123-12	c	7.45	860 ± 21	0.08366 ± 41	0.1137 ± 11	1.051 ± 23	0.0671 ± 10	694 ± 6	729 ± 15	840 ± 31	0.29
ZZB 123-13	c	4.45	338 ± 9	0.11779 ± 47	0.1462 ± 17	1.533 ± 54	0.0761 ± 28	880 ± 10	944 ± 22	1096 ± 74	0.27
ZZB 123-15	c	7.15	407 ± 4	0.10300 ± 14	0.1168 ± 6	1.096 ± 15	0.0678 ± 8	712 ± 4	751 ± 7	869 ± 24	0.40
ZZB 166-2		7.58	306 ± 6	0.11294 ± 40	0.1151 ± 8	1.074 ± 36	0.0677 ± 16	702 ± 5	741 ± 18	858 ± 52	0.22
ZZB 166-3		8.41	105 ± 3	0.20279 ± 40	0.0948 ± 17	0.874 ± 81	0.0669 ± 57	584 ± 10	638 ± 45	833 ± 190	0.38
ZZB 166-16		6.34	322 ± 11	0.11178 ± 78	0.1357 ± 27	1.260 ± 71	0.0674 ± 29	820 ± 15	828 ± 32	849 ± 94	0.22
ZZB 166-19		6.56	545 ± 25	0.08876 ± 102	0.1319 ± 15	1.202 ± 63	0.0661 ± 24	799 ± 9	801 ± 29	809 ± 80	0.62
ZZB 166-20		7.70	1009 ± 68	0.07989 ± 64	0.1124 ± 22	1.048 ± 49	0.0676 ± 17	687 ± 13	728 ± 25	856 ± 54	0.16
ZZB 166-25	c	6.25	312 ± 6	0.11324 ± 33	0.1378 ± 35	1.282 ± 60	0.0675 ± 30	832 ± 20	838 ± 27	853 ± 97	0.46
ZZB 166-29	c	9.13	1057 ± 21	0.07768 ± 29	0.0975 ± 6	0.864 ± 13	0.0642 ± 7	600 ± 3	632 ± 7	749 ± 22	0.48
ZZB 166-30	c	6.48	1385 ± 43	0.07848 ± 13	0.1342 ± 11	1.263 ± 19	0.0683 ± 5	812 ± 6	830 ± 9	877 ± 16	0.71

*: total U/radiogenic lead ratio; m: corrected for fractionation; x: corrected for common lead, blank, spike and fractionation. Lead isotope compositions for correction of sample ZZB 166: 6/4: 18.395; 7/4: 15.621; 8/4: 37.811 and sample ZZB 123: 6/4: 18.469; 7/4: 15.527; 8/4: 38.474 as analyzed on cogenetic plagioclase; c: U and Pb were separated by column chemistry; all errors are 2 σ .

Table V-2. Results of vapour transfer U-Pb isotopic analyses of zircons of pegmatites from the Mavuradonha Layered Complex.

sample no.	U/Pb^*	$^{206}Pb/^{204}Pb^m$	$^{207}Pb/^{206}Pb^m$	$^{206}Pb/^{238}U^x$	$^{207}Pb/^{235}U^x$	$^{207}Pb/^{206}Pb^x$	$^{206}Pb/^{238}U$ age [Ma]	$^{207}Pb/^{235}U$ age [Ma]	$^{207}Pb/^{206}Pb$ age [Ma]	Correl coeff.
ZZB 60-3	11.20	699 ± 22	0.07886 ± 64	0.0563 ± 6	0.451 ± 15	0.0581 ± 17	353 ± 4	378 ± 11	535 ± 64	0.48
ZZB 60-4	9.12	553 ± 5	0.08431 ± 44	0.0690 ± 15	0.553 ± 19	0.0581 ± 9	430 ± 9	447 ± 12	533 ± 34	0.82
ZZB 60-6	10.62	2023 ± 37	0.06514 ± 21	0.0837 ± 7	0.673 ± 10	0.0584 ± 4	518 ± 4	523 ± 6	543 ± 14	0.79
ZZB 60-7	7.99	939 ± 125	0.07229 ± 22	0.0802 ± 11	0.633 ± 36	0.0572 ± 25	497 ± 7	498 ± 23	499 ± 98	0.26
ZZB 60-9	9.80	564 ± 7	0.08167 ± 32	0.0758 ± 8	0.591 ± 16	0.0565 ± 9	471 ± 5	471 ± 10	473 ± 34	0.54
ZZB 60-10	10.71	565 ± 22	0.08225 ± 178	0.0846 ± 10	0.675 ± 45	0.0579 ± 31	524 ± 6	524 ± 27	524 ± 121	0.17
ZZB 128-1	10.65	256 ± 3	0.11171 ± 34	0.0881 ± 6	0.706 ± 25	0.0581 ± 12	545 ± 4	542 ± 15	532 ± 44	0.30
ZZB 128-2	11.18	808 ± 30	0.07432 ± 99	0.0831 ± 7	0.657 ± 26	0.0574 ± 17	514 ± 3	513 ± 16	506 ± 66	0.17
ZZB 128-5	11.90	319 ± 3	0.09966 ± 23	0.0791 ± 7	0.613 ± 21	0.0561 ± 10	491 ± 4	485 ± 13	458 ± 41	0.40

*: total U/radiogenic lead ratio; m: corrected for fractionation; x: corrected for common lead, blank, spike and fractionation. Pegmatites were corrected for common lead for 550 Ma after Stacey & Kramers (1975).

Table V-3. Results of vapour transfer U-Pb isotopic analyses of zircons from Ocellar Gneisses.

sample no.	U/Pb*	$^{206}\text{Pb}/^{204}\text{Pb}^m$	$^{207}\text{Pb}/^{206}\text{Pb}^m$	$^{206}\text{Pb}/^{238}\text{U}^x$	$^{207}\text{Pb}/^{235}\text{U}^x$	$^{207}\text{Pb}/^{206}\text{Pb}^x$	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{235}\text{U}$	$^{207}\text{Pb}/^{206}\text{Pb}$	Correl coeff.
							age [Ma]	age [Ma]	age [Ma]	
ZIM 30-a	5.84	643 ± 40	0.09226 ± 80	0.1499 ± 35	1.460 ± 95	0.0707 ± 29	900 ± 20	914 ± 40	947 ± 86	0.47
ZIM 30-1	8.90	1766 ± 310	0.07813 ± 75	0.1001 ± 52	0.975 ± 92	0.0706 ± 32	615 ± 30	691 ± 48	947 ± 95	0.18
ZIM 30-2	6.43	407 ± 35	0.10181 ± 86	0.1385 ± 15	1.272 ± 92	0.0666 ± 44	836 ± 9	833 ± 42	827 ± 146	0.14
ZIM 30-3	7.25	1412 ± 193	0.07925 ± 104	0.1228 ± 50	1.172 ± 92	0.0692 ± 32	747 ± 29	788 ± 44	906 ± 99	0.64
ZIM 30-5	7.71	1630 ± 286	0.07908 ± 47	0.1138 ± 14	1.105 ± 46	0.0704 ± 22	695 ± 8	756 ± 22	941 ± 65	0.33
ZIM 30-6	6.50	1440 ± 143	0.08154 ± 42	0.1334 ± 14	1.319 ± 41	0.0717 ± 15	807 ± 8	854 ± 18	977 ± 43	0.43
ZIM 30-7	6.19	511 ± 4	0.10064 ± 23	0.1411 ± 12	1.415 ± 21	0.0728 ± 6	851 ± 7	895 ± 9	1007 ± 78	0.68
ZIM 30-9	9.63	662 ± 4	0.09242 ± 30	0.0932 ± 9	0.911 ± 19	0.0709 ± 7	574 ± 5	657 ± 8	954 ± 20	0.71
ZIM 55-1	14.91	809 ± 99	0.08299 ± 33	0.0548 ± 4	0.509 ± 26	0.0673 ± 27	344 ± 3	418 ± 18	848 ± 84	0.56
ZIM 55-2	7.62	3208 ± 389	0.07089 ± 45	0.1066 ± 10	0.978 ± 24	0.0666 ± 11	653 ± 6	693 ± 13	824 ± 34	0.46
ZIM 55-3	23.64	867 ± 10	0.08351 ± 47	0.0346 ± 3	0.320 ± 6	0.0671 ± 9	219 ± 2	281 ± 5	840 ± 29	0.51
ZIM 55-4	8.49	2040 ± 169	0.07303 ± 94	0.0975 ± 38	0.888 ± 57	0.0661 ± 21	600 ± 22	645 ± 31	808 ± 69	0.76
ZIM 55-12	7.32	4222 ± 197	0.07088 ± 32	0.1137 ± 8	1.060 ± 15	0.0676 ± 5	695 ± 5	734 ± 8	855 ± 16	0.66
ZIM 56-7	5.01	4167 ± 36	0.07643 ± 17	0.1718 ± 15	1.736 ± 21	0.0733 ± 2	1022 ± 8	1022 ± 8	1022 ± 6	0.95
ZIM 56-8	c 5.56	911 ± 18	0.08932 ± 32	0.1535 ± 51	1.560 ± 66	0.0737 ± 7	921 ± 29	955 ± 26	1034 ± 20	0.96
ZIM 56-9	c 7.02	2112 ± 15	0.07812 ± 9	0.1280 ± 11	1.260 ± 13	0.0714 ± 2	776 ± 6	828 ± 6	968 ± 5	0.96
ZIM 56-12	5.19	3488 ± 41	0.07625 ± 23	0.1665 ± 7	1.648 ± 6	0.0718 ± 8	993 ± 4	989 ± 2	980 ± 24	0.95

*: total U/isotopic lead ratio; m: corrected for fractionation; x: corrected for common lead after Stacey & Kramers (1975), blank, spike and fractionation; c: U and Pb were separated by column chemistry; all errors are 2 σ .

Table V-4. Results of SHRIMP U-Th-Pb isotopic analyses for zircons from layered metagabbro sample ZZB 123.

<i>grain spot.</i>	<i>U (ppm)</i>	<i>Th (ppm)</i>	$^{206}\text{Pb}/^{204}\text{Pb}$ (<i>measured</i>)	$^{208}\text{Pb}/^{206}\text{Pb}$	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{235}\text{U}$	$^{207}\text{Pb}/^{206}\text{Pb}$	$^{206}\text{Pb}/^{238}\text{U}$ <i>age [Ma]</i>	$^{207}\text{Pb}/^{235}\text{U}$ <i>age [Ma]</i>	$^{207}\text{Pb}/^{206}\text{Pb}$ <i>age [Ma]</i>
ZZB 123-1	139	133	6703	0.2389 ± 48	0.1779 ± 16	1.819 ± 53	0.0742 ± 20	1055 ± 9	1052 ± 19	1046 + 53/- 55
ZZB 123-2	52.6	117	9291	0.6690 ± 80	0.1707 ± 17	1.741 ± 57	0.0739 ± 22	1016 ± 9	1024 ± 21	1040 + 59/- 62
ZZB 123-3	127	67.3	23124	0.1605 ± 17	0.1679 ± 15	1.731 ± 22	0.0748 ± 6	1000 ± 8	1020 ± 8	1062 ± 16
ZZB 123-4	26.1	14.3	5760	0.1743 ± 41	0.1435 ± 15	1.390 ± 32	0.0703 ± 14	864 ± 9	885 ± 14	936 + 39/- 40
ZZB 123-5	151	113	4495	0.2334 ± 29	0.1906 ± 34	2.102 ± 46	0.0800 ± 9	1121 ± 18	1120 ± 19	1116 + 38/- 39
ZZB 123-6	414	183	15306	0.1401 ± 34	0.0901 ± 15	0.724 ± 22	0.0583 ± 14	556 ± 9	553 ± 13	541 + 50/- 52
ZZB 123-7	249	164	2500	0.2008 ± 57	0.0895 ± 15	0.725 ± 32	0.0857 ± 23	553 ± 9	553 ± 19	556 + 82/- 86
ZZB 123-8	296	193	4811	0.2012 ± 44	0.0897 ± 15	0.730 ± 25	0.0590 ± 17	554 ± 9	556 ± 15	568 + 60/- 62
ZZB 123-9	32.4	19.1	648	0.175 ± 22	0.1426 ± 30	1.326 ± 188	0.0675 ± 93	859 ± 17	857+79/-86	852 + 263/-316

Ratios refer to radiogenic lead. All errors are 1 σ .

Table V-5. Pb-Pb isotopic data from evaporation of zircons from the Mavuradonha Layered Complex, from a pegmatite and the Ocellar Gneiss.

<i>sample no.</i>	<i>zircon colour and morphology</i>	<i>grain no.</i>	<i>mass scans¹</i>	<i>mean ²⁰⁷Pb/²⁰⁶Pb ratio² and 2-σ error</i>	<i>²⁰⁷Pb/²⁰⁶Pb age [Ma] and 2-σ error</i>
pegmatite					
ZZB 60	light brown,	1	110	0.058727 ± 21	557.0 ± 0.8
	idiomorphic, longprismatic	2	132	0.058674 ± 11	555.0 ± 0.4
	slightly rounded terminations	3	132	0.058635 ± 25	553.6 ± 0.9
mean of 4				0.058679 ± 19	*555 ± 1
MLC					
ZZB 123	magmatic and metamorphic zircons				
	bright light brown, clear,	fract. 1	198	0.058047 ± 43	532 ± 2
	round	1	194	0.058393 ± 40	545 ± 1.5
mean of 2			392	0.05822 ± 4	538 ± 2
	pale brown, clear	2	98	0.06581 ± 11	800 ± 4
	idiomorphic, longprismatic	fract. 2	136	0.065873 ± 70	802 ± 2
	euhebral	3	114	0.066167 ± 74	811 ± 2
mean of 3			348	0.06595 ± 17	805 ± 5
	pale brown, clear	fract. 3	157	0.086794 ± 38	893 ± 2
	idiomorphic, longprismatic	4	137	0.069062 ± 66	900 ± 2
	euhebral	5	79	0.068585 ± 70	886 ± 2
mean of 3			373	0.06881 ± 21	893 ± 6

¹Number of ²⁰⁷Pb/²⁰⁶Pb ratios evaluated for age assessment. ²Observed mean ratio corrected for non-radiogenic Pb where necessary. Errors based on uncertainties in counting statistics. *Errors of combined mean ages (bold print) are based on the reproducibility of internal standard at 0.000027 (2σm).

Table V-5. Continued.

<i>sample no.</i>	<i>zircon colour and morphology</i>	<i>grain no.</i>	<i>mass scans¹</i>	<i>mean ²⁰⁷Pb/²⁰⁶Pb ratio² and 2-σ error</i>	<i>²⁰⁷Pb/²⁰⁶Pb age [Ma] and 2-σ error</i>
ZZB 123	xenocrysts				
		6	132	0.079315 ± 82	1182 ± 5
	pinkish-brown, slightly clouded,	7	100	0.078627 ± 77	1162 ± 2
	longprismatic, subrounded terminations	8	60	0.079253 ± 45	1178 ± 1
mean of 3			292	0.079065 ± 340	1174 ± 9
		9	96	0.095337 ± 138	1535 ± 4
	dark brown, cloudy	10	78	0.101126 ± 108	1645 ± 2
	prismatic, well rounded or	11	80	0.101031 ± 145	1643 ± 3
mean of 3	anhedral		158	0.101079 ± 60	1644 ± 1
		12	97	0.109963 ± 71	1799 ± 2
ZZB 166		1	53	0.067114 ± 81	838 ± 4
	pale to pinkish-brown, clear,	2	60	0.067126 ± 107	842 ± 4
	longprismatic,	3	40	0.067434 ± 306	851 ± 10
	slightly to well rounded	4	158	0.0677379 ± 45	850 ± 2
		5	138	0.07213 ± 65	844 ± 3
		6	58	0.066706 ± 160	823 ± 6)
mean of 6			449	0.067325 ± 236	849 ± 2

¹Number of ²⁰⁷Pb/²⁰⁶Pb ratios evaluated for age assessment. ²Observed mean ratio corrected for non-radiogenic Pb where necessary. Errors based on uncertainties in counting statistics.

Table V-5. Continued.

<i>sample no.</i>	<i>zircon colour and morphology</i>	<i>grain no.</i>	<i>mass scans¹</i>	<i>mean ²⁰⁷Pb/²⁰⁶Pb ratio² and 2-σ error</i>	<i>²⁰⁷Pb/²⁰⁶Pb age [Ma] and 2-σ error</i>
Ocellar Gneiss					
Zim 30	brown,	1	110	0.072903 ± 27	1011.3 ± 0.8
	long-prismatic	2	132	0.072889 ± 13	1010.9 ± 0.4
	idiomorphic with rounded	3	132	0.072927 ± 11	1011.9 ± 0.3
	terminations	4	110	0.072908 ± 22	1011.4 ± 0.6
mean of 4			484	0.072907 ± 19	*1011.4 ± 1
Zim 55	brown	1	121	0.072692±24	1005.4 ± 0.7
	long-prismatic	2	132	0.072704±23	1005.7 ± 0.7
	rounded terminations	3	143	0.072679±19	1005.0 ± 0.5
		4	143	0.072686±15	1005.2 ± 0.4
mean of 4			539	0.072690±10	*1005.3 ± 0.7
Zim 56	yellow-brown	1	165	0.073156 ± 23	1018.3 ± 0.6
	long-prismatic	2	143	0.073143 ± 26	1017.9 ± 0.7
	idiomorphic with	3	132	0.073165 ± 24	1018.5 ± 0.7
	slightly rounded	4	55	0.073161 ± 35	1018.4 ± 1.0
	terminations	5	55	0.073150 ± 41	1018.1 ± 1.1
mean of 5			550	0.073155 ± 12	*1018.3 ± 1

¹Number of ²⁰⁷Pb/²⁰⁶Pb ratios evaluated for age assessment. ²Observed mean ratio corrected for non-radiogenic Pb where necessary. Errors based on uncertainties in counting statistics. *Errors of combined mean ages (bold print) are based on the reproducibility of internal standard at 0.000027 (2σm).

Table V-6. Results of U-Pb isotopic analyses of rutiles from the Mavuradonha Layered Complex.

sample no.		U/Pb^*	$^{206}Pb/^{204}Pb^m$	$^{207}Pb/^{206}Pb^m$	$^{206}Pb/^{238}U^x$	$^{207}Pb/^{235}U^x$	$^{207}Pb/^{206}Pb^x$	$^{206}Pb/^{238}U$ age [Ma]	$^{207}Pb/^{235}U$ age [Ma]	$^{207}Pb/^{206}Pb$ age [Ma]	Correl coeff.
Rt ZZB 150-1	c	8.4500	132.1 ± 0.1	0.1656 ± 114	0.0809 ± 3	0.6387 ± 33	0.0573 ± 2	502 ± 6	502 ± 8	501 ± 9	0.52
Rt ZZB 150-3	c	10.0045	203.3 ± 0.3	0.1267 ± 375	0.0809 ± 3	0.6339 ± 37	0.0568 ± 3	501 ± 6	499 ± 9	486 ± 11	0.50
Rt ZZB 150-4	c	2.3348	32.8 ± 0.1	0.4975 ± 121	0.0850 ± 4	0.6710 ± 192	0.0573 ± 16	526 ± 7	521 ± 28	502 ± 60	0.59

*: total U/radiogenic lead ratio; m: corrected for fractionation; x: corrected for common lead, blank, spike and fractionation. Lead isotope composition for correction of sample ZZB 150: $\underline{6/4}$: 17.419; $\underline{7/4}$: 15.495; $\underline{8/4}$: 37.04 as analyzed on cogenetic plagioclase; c: U and Pb were separated by column chemistry; all errors are 2 σ .

Table V-7. Results of wholerock analytical data of Sm and Nd isotopic analyses for ferro-metagabbros and meta-anorthosites from the Mavuradonha Layered Complex.

sample	rock type	MSWD	$^{143}Nd/^{144}Nd_{(l)}$	$E_{Nd(l)}$	age (Ma)
	garnet – whole rock	4.4	0.51210 ± 03	3.0	546 ± 09
ZZB 146	meta-anorthosite	—	0.51205 ± 03	2.3	549 ± 06
ZZB 147	ferro-metagabbro	—	0.51209 ± 03	3.0	538 ± 18
ZZB 148	meta-anorthosite	—	0.51212 ± 02	3.5	547 ± 07
ZZB 148	garnet + inclusion rich fraction	1.9	0.51212 ± 02	3.5	546 ± 06
ZZB 149	ferro-metagabbro	—	0.51210 ± 02	3.2	545 ± 06
ZZB 150	meta-anorthosite	—	0.51200 ± 02	1.6	560 ± 04

Results of Sm-Nd isotopic analyses for garnets are given in Table VI-1.

A-VI Sm-Nd, Rb-Sr and Pb-Pb isotopic systematics

Table VI-1. Sm-Nd isotopic composition for whole rocks and garnets of the Mavuradonha Layered Complex, and whole rocks of the Nyamhanda and Chimwaya Inliers.

sample	rock type/ mineral	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}_{(i)}$	$E_{\text{Nd}} (i)$	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$
Mavuradonha Layered Complex												
ZZB 54	metapyroxenite	0.545	1.244	0.2647	0.513360 ± 09	0.511867	6.6	3.0	147	0.060	0.708140 ± 16	0.707410
ZZB 65	garnet-amphibolite	3.136	8.789	0.2154	0.512970 ± 08	0.511755	4.4	5.5	138	0.115	0.705264 ± 10	0.703854
ZZB 66	metapyroxenite	0.359	0.477	0.4547	0.514110 ± 10	0.511545	0.3	2.0	145	0.040	0.707508 ± 17	0.707022
ZZB 67	metagabbro	0.854	2.106	0.2425	0.513134 ± 21	0.511751	4.4	7.9	221	0.104	0.705309 ± 14	0.704033
ZZB 68	meta-anorthosite	0.111	0.297	0.2256	0.513051 ± 23	0.511778	4.9	2.6	144	0.053	0.703509 ± 14	0.702863
ZZB 69	metagabbro	0.242	0.672	0.2179	0.513099 ± 16	0.511870	6.7	5.0	101	0.143	0.705859 ± 15	0.704097
ZZB 70	metapyroxenite	0.379	0.617	0.3717	0.513884 ± 13	0.511787	5.1	0.2	53	0.011	0.702605 ± 13	0.702469
MAV 1	metapyroxenite	0.382	0.719	0.3210	0.513580 ± 47	0.511769	4.7	6.9	43	0.462	0.709179 ± 12	0.703503
ZZB 146	meta-anorthosite	1.433	5.467	0.1585	0.512623 ± 16	0.511730	3.9	2.7	646	0.012	0.702569 ± 10	0.702421
ZZB 146	garnet	1.472	0.989	0.9007	0.515292 ± 05	0.512076	—					
ZZB 147	ferro-metagabbro	4.164	14.81	0.1699	0.512691 ± 11	0.511732	4.0	3.0	210	0.041	0.704201 ± 09	0.703696
ZZB 147	garnet	2.558	2.006	0.7714	0.514811 ± 69	0.512057	—					
ZZB 148	meta-anorthosite	1.211	4.340	0.1687	0.512719 ± 10	0.511767	4.7	3.3	691	0.014	0.702602 ± 08	0.702432
ZZB 148	garnet	1.947	1.494	0.7884	0.514938 ± 17	0.512123	—					
ZZB 148	garnet + inclusions	1.826	1.524	0.7249	0.514686 ± 25	0.512098	—					
ZZB 149	ferro-metagabbro	3.939	13.25	0.1797	0.512740 ± 12	0.511727	3.9	1.6	241	0.020	0.704187 ± 18	0.703945
ZZB 149	garnet	2.698	1.966	0.8301	0.515061 ± 14	0.512097	—					
ZZB 150	meta-anorthosite	0.820	3.585	0.1382	0.512498 ± 11	0.511718	3.7	1.7	673	0.007	0.702533 ± 10	0.702442
ZZB 160	meta-anorthosite	0.370	1.424	0.1570	0.512585 ± 15	0.511700	3.3	0.4	277	0.004	0.703676 ± 11	0.703623
ZZB 161	ferro-metagabbro	3.824	11.22	0.2061	0.512910 ± 12	0.511748	4.3	5.3	71	0.215	0.707221 ± 09	0.704577
ZZB 162	meta-anorthosite	0.168	0.643	0.1575	0.512651 ± 20	0.511763	4.6	0.4	277	0.004	0.703640 ± 10	0.703594

(i). Initial $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are recalculated to an age of 860 Ma. Errors of the measured isotope ratios are given as 2σ .

Table VI-1. Continued.

sample	rock type/ mineral	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}_{(i)}$	$E_{\text{Nd}} (i)$	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$
Mavuradonha Layered Complex												
ZZB 164	meta-anorthosite	0.453	1.604	0.1709	0.512756 ± 20	0.511792	5.1	1.4	294	0.014	0.705213 ± 14	0.705043
ZZB 166	ferro-metagabbro	4.135	13.95	0.1793	0.512786 ± 12	0.511775	4.8	9.6	100	0.278	0.706770 ± 06	0.703360
Nyamhanda Inlier												
DKZ 5	metagabbro	0.671	3.197	0.1269	0.512354 ± 12	0.511639	2.1	2.4	246	0.029	0.704801 ± 05	0.704450
DKZ 11	metagabbro	8.730	39.24	0.1345	0.512351 ± 12	0.511592	1.2	11	364	0.090	0.705122 ± 13	0.704014
DKZ 12	metagabbro	0.645	2.725	0.1431	0.512441 ± 07	0.511634	2.1	2.6	697	0.011	0.704927 ± 14	0.704793
DKZ 16	garnet-amphibolite	3.546	13.19	0.1625	0.512307 ± 13	0.511390	- 2.7	7.1	135	0.153	0.712496 ± 11	0.710615
DKZ 18	metagabbro	0.788	2.517	0.1891	0.512495 ± 23	0.511429	- 2.0	9.0	207	0.126	0.709550 ± 11	0.708002
DKZ 22	metagabbro	0.731	2.413	0.1830	0.512381 ± 13	0.511349	- 3.5	5.4	130	0.120	0.710017 ± 14	0.708546
DKZ 25	amphibolite	0.775	2.001	0.2341	0.512809 ± 07	0.511498	- 0.8	2.8	253	0.032	0.707652 ± 13	0.707260
DKZ 57	metagabbro	10.96	52.52	0.1261	0.512182 ± 16	0.511471	- 1.1	27	333	0.231	0.712385 ± 17	0.709542
Chimwaya Inlier												
ZCW 2	metagabbro	1.377	7.376	0.1129	0.512280 ± 07	0.511644	2.2	21	1263	0.048	0.704973 ± 11	0.704390
ZCW 6	metagabbro	0.650	3.484	0.1129	0.512271 ± 07	0.511634	2.1	2.4	708	0.010	0.704138 ± 10	0.704019

(i). Initial $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are recalculated to an age of 860 Ma. Errors of the measured isotope ratios are given as 2σ .

Table VI-2. Pb-Pb isotopic composition for selected plagioclases of the Mavuradonha Layered Complex, the Nyamhanda and Chimwaya Inliers.

<i>sample</i>	Pb^{206}/Pb^{204}	Pb^{207}/Pb^{204}	Pb^{208}/Pb^{204}
Mavuradonha Layered Complex			
ZZB 67	18.155 ± 2	15.599 ± 3	38.004 ± 6
ZZB 123	18.469 ± 2	15.527 ± 3	38.474 ± 8
ZZB 146	17.188 ± 2	15.449 ± 3	36.744 ± 6
ZZB 147	17.994 ± 3	15.528 ± 4	37.289 ± 8
ZZB 148	17.252 ± 6	15.462 ± 7	36.834 ± 16
ZZB 149	18.172 ± 7	15.583 ± 5	37.764 ± 11
ZZB 150	17.419 ± 3	15.495 ± 3	37.040 ± 7
ZZB 162	18.384 ± 2	15.550 ± 4	38.213 ± 5
ZZB 166	18.395 ± 7	15.621 ± 6	37.811 ± 13
Chimwaya Inlier			
ZCW 2	18.577 ± 2	15.532 ± 4	38.217 ± 6
Nyamhanda Inlier			
DKZ 22	15.681 ± 9	15.746 ± 6	38.144 ± 15
DKZ 57	15.746 ± 3	15.681 ± 3	38.129 ± 8

Errors of the measured isotope ratios are given as 2 σ . Ratios are corrected for mass fractionation with a factor of 0.15 % per amu.

Curriculum Vitae

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