

Petrology and associated woody vegetation of the Koedoesfontein Complex in the Vredefort dome

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DECLARATION

I declare that the work presented in this *Magister Scientiae* dissertation is my own work, that it has not been submitted for any degree or examination at any other university, and that all the sources I have used or quoted have been acknowledged by complete reference.

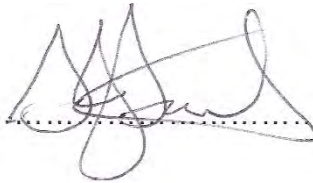
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*Leaning on my elbow, I turn to my own thoughts.
The pages of my life, I have cast, on the great millwheel of fortune.
In my hunt for the golden crown, I have sought clues, in labours long
and wide.
I am resolute in my dangerous pursuit.
Dangerous you laugh?
Yes, I say, for Wisdom is demanding.
She does not release her captives readily, but demands inconsiderate
labor in searching her ciphers.
So as the millwheel turns, I entrust myself between the columns where
the University of Antiquity awaits.
There I am sure the dance of the labyrinth will lead me to the emerald
city.
And Ariadne's silken thread, woven from chains of light, will support
my journey to the midnight sun.
The turn of the millwheel has brought me luck, I cast again...
...God goes with the brave.*

Stuart Nettleton, (1998)

ABSTRACT

The Vredefort dome impact crater hosts several Neoarchaeon (2800 Ma) to Mesoproterozoic (1000 Ma) peralkaline to ultramafic intrusive bodies that are exposed amongst its core and collar. The Koedoesfontein Complex (KC), in particular, forms part of an alkaline-dioritic suite of post-Transvaal, syn-Bushveld (~2050 to 2060 Ma) intrusions including the Schurwedraai-, Lindequesdrift-, Roodekraal- and Rietfontein- Complexes, as well as the Winddam wehrlite.

Due to the current level of erosion, the geology of the KC is relatively less well understood and has been historically understudied, probably because of its relatively small size, poor outcrop visibility and dense vegetation cover. These characteristics make the Complex difficult to interpret, however, since several plant species are well established amongst the diverse geology, it posed a unique opportunity for a multidisciplinary approach to re-evaluate the petrology of the underlying intrusions and to simultaneously study their effects on the hosted woody vegetation.

With regards to the geology, a number of small dyke structured outcrops of the KC were identified and described during previous studies. These consist of ultramafic wehrlite and dioritic lamprophyre (termed spessartite), as well as a few dispersed outcrops of sodic (alkali) granite. More recently, investigation of the KC revealed a broader textural and mineralogical spectrum within the spessartite body, as well as the presence of small unmapped sills of mafic gabbro, hornblendite, olivine $\pm$ clinopyroxenite and quartz alkali syenite. These intrusive rock types are hypothesised to be related to the known syn-Bushveld intrusions of the KC, and to collectively form a layered series which is derived from magmatic differentiation.

The KC was mapped, and a complete set of rock samples were taken across the entire intrusive sequence. The rock samples were prepared for the production of thin sections, as well as X-ray diffraction and X-ray fluorescence analysis to determine the mineral composition and geochemistry.

Structural evidence, petrographical observations and geochemical trends of the known and unmapped outcrops, in their entirety, indicate that the intrusions collectively form a coherent stratigraphical, mineralogical and geochemical sequence that supports the notion for differentiation. It is therefore concluded that the KC represents a layered differentiated intrusion quite similar and probably closely related to the nearby Rietfontein Complex.

With regards to the botany, the broad difference in mineral composition of the intrusions, as well as the abundant quartzite and shale country rocks, are believed to cause abrupt transitions in soil chemistry that ultimately affect the floristic, physiognomic and biochemical characteristics of the hosted woody vegetation. Ultramafic intrusions, in particular, are known to produce soils that are

rich in magnesium, chrome, cobalt and nickel, and may cause adverse affects on species that are not adapted to these environments. In spite of various efforts to study the ultramafic-adapted vegetation (also known as “serpentine flora”) of southern Africa, little is known about the effect these particular intrusions have on the soil and vegetation components of the Vredefort dome.

The KC was established to collectively form a mineralogical gradient ranging from purely mafic- to felsic-dominant silicates. Three primary components (geology, soil and woody vegetation) were sampled along this gradient and analysed to determine the effect of these intrusions on the residual soil chemistry, and the floristics, physiognomy and biochemistry of the supported woody vegetation.

Weathering of the ultramafic to felsic alkaline rock types greatly affected the soil's chemical properties. Major chemical variation in the soil, including calcium and sodium availability and the total chromium and nickel content, is correlated with the difference in abundance of calcium-, sodium-, iron- and magnesium-rich silicate minerals.

Significant variations in woody species composition (particularly across members of *Senegalia* and *Vachellia*) have been established between soils originating from ultramafic, mafic, dioritic and felsic rock types, and can be utilised as indicators for different geological substrates. The ability of *V. karroo* to outcompete *S. caffra* and *V. robusta* on ultramafic intrusions was confirmed by higher species performance index values for this substrate. It is concluded, therefore, that *V. karroo* is an indicator for ultramafic rocks types at the KC.

The dominance of *V. karroo* on ultramafic rock types was hypothesised to be due to a higher tolerance of magnesium, chrome, cobalt and nickel by means of element exclusion. Results are in support of this notion, and it is concluded that *V. karroo* is able to more effectively suppress its magnesium, chrome, cobalt and nickel content compared to other species such as *Grewia. flava*, *S. caffra* and *V. robusta*.

Key terms

Vredefort dome, syn-Bushveld intrusions, Koedoesfontein Complex, ultramafic–mafic intrusions, serpentine flora, *Vachellia*, *Senegallia*.

UITTREKSEL

Die Vredefortkoepel impakkrater bevat verskeie Neoargeïese (2800 Ma) tot Mesoproterosoïese (1000 Ma) peralkaliese tot ultramafiese intrusiewe liggamete wat in die kern en kraag van die koepel dagsoom. Die Koedoesfontein Kompleks (KC) in die besonder, vorm deel van 'n alkaliese-dioritiese suite met indringings van post-Transvaal, sin-Bosveld ouderdom (~ 2050 tot 2060 Ma) en sluit in die Schurwedraai-, Lindequesdrift-, Roodekraal- en Rietfontein Komplekse, asook die Winddam wehrliet.

As gevolg van die huidige mate van erosie, is die geologie van die KC minder bekend en is dit histories onderbestudeer, moontlik omdat dit relatief klein is, swak dagsoom en met digte plantegroei bedek is. Hierdie eienskappe maak die Kompleks egter moeilik om te interpreteer, maar aangesien verskeie plantspesies goed gevestig is onder die diverse geologie, het dit 'n unieke geleentheid gebied vir 'n multidisiplinêre aanslag om die petrologie van die onderliggende indringings te herevalueer en om gelyktydig hul effek op die teenwoordige houtagtige plantegroei te bestudeer.

Met betrekking tot die geologie, is verskeie klein gangagtige dagsome van die KC deur vorige studies geïdentifiseer en beskryf. Hierdie bestaan uit ultramafiese wehrliet en dioritiese lamprofier (ook genoem spessartiet), sowel as 'n paar verspreide dagsome van natrium-ryke (alkaliese) graniet. Meer onlangse ondersoek van die KC het 'n breër tekstuur- en mineralogiese spektrum binne die spessartiet liggaam bekend gemaak, sowel as die teenwoordigheid van klein ongekarteerde plate van mafiese gabbro, horingblendiet, olivien ± klinopirokseniet en kwarts alkali sieniet. Daar word voorgestel dat hierdie indringings verband hou met die bekende sin-Bosveld indringings van die KC, en dat hierdie rotstipes gesamentlik 'n gelaagde reeks vorm wat afkomstig is van magmatiese differensiasie.

Die KC is gekarteer, en 'n volledige stel rotsmonsters is oor die hele gesteente reeks van die Kompleks geneem. Die rotsmonsters was voorberei vir die maak van slypplaatjies, asook X-straaldiffraksie en X-straal-fluoressensie ontledings om die mineraal samestelling en geochemie te bepaal.

Strukturele bewyse, petrografiese waarnemings en geochemiese tendense van die bekende en onbekende dagsome in die geheel, dui daarop dat die indringings 'n samehangende stratigrafiese, mineralogiese en geochemiese volgorde volg wat die hipotese vir differensiasie ondersteun. Daar word derhalwe tot die gevolgtrekking gekom dat die KC 'n gelaagde gedifferensieerde indringing verteenwoordig wat soortgelyk en waarskynlik nou verwant is aan die nabygeleë Rietfontein Kompleks.

Met betrekking tot die plantegroei word daar beweer dat die breë verskil in mineralogiese samestelling van die indringings, asook die oorvloedige kwartsiet- en skalie newegesteentes, skerp oorgange in grondchemie veroorsaak wat uiteindelik die floristiese, fisiese en biochemiese eienskappe van die gevestigte houtagtige plantegroei beïnvloed. Dit is veral bekend dat ultramafiese indringings grondsoorte produseer wat ryk is aan magnesium, chroom, kobalt en nikkel, en kan nadelige gevolge vir spesies inhou wat nie aangepas is to hierdie omgewings nie. Ten spyte van verskeie pogings om die ultramafiese aangepaste plantegroei (ook bekend as "serpentine flora") van Suider-Afrika te bestudeer, is min bekend oor die effek wat hierdie besondere indringings het op die grond- en plantegroei komponente van die Vredefort-koepel.

Die KC is bevestig om gesamentlik 'n mineralogiese gradiënt te vorm wat wissel van dominante suiwer mafies tot felsiese silikate. Drie primêre komponente (geologie, grond en houtagtige plantegroei) is langs hierdie gradiënt gemonster en ontleed om die effek van hierdie indringings op die residuele grondchemie, en die floristiek, fisiese en biochemie van die teenwoordige houtagtige plantegroei te bepaal.

Die verwerking van die ultramafiese tot felsiese alkaliese rotstipes het grootliks die chemiese eienskappe van die grond beïnvloed. Merkwaaardige chemiese variasie in die grond, insluitende kalsium en natrium beskikbaarheid en die totale chroom- en nikkelinhoud, word gekorreleer met die verskil in die verspreiding van kalsium-, natrium-, yster- en magnesiumryke silikaatminerale.

Betekenisvolle variasies in houtagtige spesiesamestelling (veral tussen lede van *Senegalia* en *Vachellia*) is uitgekien tussen gronde wat ontstaan uit ultramafiese, mafiese, dioritiese en felsiese gesteentes, en kan gebruik word as aanwysers vir verskillende rots tipes. *V. karroo* se vermoë om *S. caffra* en *V. robusta* uit te kompeteer op ultramafiese indringings, is bevestig deur hoër spesie prestasie indekswaardes op hierdie substraat. Daarom word die gevolgtrekking gemaak dat *V. karroo* 'n aanwyser is vir ultramafiese gesteentes by die KC.

Die dominansie van *V. karroo* op ultramafiese gesteentes is vermoedelik te wyte aan 'n hoër verdraagsaamheid van magnesium, chroom, kobalt en nikkel deur middel van element-uitsluiting. Resultate is ter ondersteuning van hierdie idee, en daar word tot die gevolgtrekking gekom dat *V. karroo* in staat is om sy magnesium-, chroom-, kobalt- en nikkelinhoud meer effektief te onderdruk in vergelyking met ander spesies soos *Grewia flava*, *S. caffra* en *V. robusta*.

Sleutel terme

Vredefort koepel, sin-Bosveld indringings, Koedoesfontein Kompleks, ultramafies-mafiese indringings, serpentyne flora, *Vachellia*, *Senegalia*.

ACRONYMS USED

BAG: Baviaankraans alkali granite
Bushveld: Bushveld igneous Complex
CCA: Canonical Correspondence Analyses
CEC: Cation exchange capacity
HFS: High field strength
HITIS: Hi-Titanium Igneous Suite
IV: Species performance Index Value
KC: Koedoesfontein Complex
LDC: Lindequesdrift Complex
LOI: Loss on ignition
MDPS: Mafic – dioritic – peralkaline series
PCA: Principal Component Analyses
PPL: Plane polarized light
RC: Roodekraal Complex
RTC: Rietfontein Complex
SAG: Schurwedraai alkali granite
SC: Schurwedraai Complex = **BAG** + **SAG**
SC: Stem circumference
SOC: Soil organic carbon
SOM: Soil organic material
Ventersdorp: Ventersdorp Supergroup
WW: Winddam wehrlite
XPL Crossed polarized light

MINERAL ABBRIVIATIONS USED

Ab: Albite
Amp: Amphibole
An: Anorthite
Ap: Apatite
Aug: Augite
Bt: Biotite
Cpx: Clinopyroxene
Fhb: Ferro-hornblende / Fe-Hornblende
Hbl: Hornblende
Mag: Magnetite
Mc: Microcline
Mhb: Magnesio-hornblende / Mg-Hornblende
Nph: Nepheline
OI: Olivine
Opx: Orthopyroxene
Or: Orthoclase / K-feldspar
Pl: Plagioclase
Px: Pyroxene
Qz: Quartz
Spn: Sphene
Sr: Sericite
Srp: Serpentine
Zr: Zircon

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CHAPTER 1

GENERAL INTRODUCTION

1.1 BACKGROUND

The Vredefort dome is located approximately 120 km south-west of Johannesburg (Figure 1), centered within the geologically renowned Witwatersrand Basin. The area consists of a structurally uplifted terrain which is commonly accepted to be the result of a large meteorite impact, dated around $\sim 2023 \pm 4$ Ma (Kamo *et al.*, 1996). As a result of the upliftment and overturning of the supracrustal rocks in the northern and north-western sectors of the dome, several of South Africa's major stratigraphic units are exposed within this area. These include rocks which originated from as old as ~ 3500 Ma to more recent ~ 180 Ma old fluvial depositions. Ages for the following major units were obtained from Gibson and Reimold (2008), Reimold and Gibson (2005), Reimold (2006) and Rob *et al.*, (2006).

The dome comprises a 40 to 50 km wide central core consisting of Archaean granitoid gneisses (c. 3200-3100 Ma) which in addition, contain subsidiary greenstone inliers as well as several ultramafic, mafic, pelithic and ironstone xenoliths, which is thought to be around ~ 3400 -3500 Ma old (Hart *et al.*, 1981; Armstrong *et al.*, 2006). With the exception of unusually high metamorphic grades, it is suggested that the core rocks resemble typical Archean granite-greenstone sequences observed in adjoining areas of the Kaapvaal craton (Gibson and Reimold, 2008). The granitic core rocks have been subdivided into an inner core of granulite facies gneisses termed the Inlandsee Lecogranofels (ILG), and an outer core of amphibolite-grade migmatitic gneisses termed the Outer Granite Gneiss (OGG) (Stephens, 1990).

The granitoid core is rimmed by a 20 to 25 km wide collar of up- to overturned layers including volcanic and sedimentary sequences of the Dominion Group (c. 3074 Ma), and strata of the Witwatersrand (c. 2950 to 2710 Ma), Ventersdorp (c. 2714 Ma) and Transvaal (c. 2650 to 2150 Ma) Supergroups. At present day, only remnants of the dome's once vast structure remain visible. Regional geophysical modelling (Henkel & Reimold, 1998) in addition to the distribution of impact related deformation structures, would suggest an original diameter of 180 to 300 km (Brink, *et al.*, 1997; Hart *et al.*, 2000; Wieland *et al.*, 2005; Reimold, 2006). The overturned layers become progressively younger from the core outwards and are partially covered by horizontal Palaeozoic sedimentary formations and dolerite sills of the Karoo Supergroup, forming the youngest group of rocks (c. 300 to 180 Ma) in the south-eastern part of the structure.

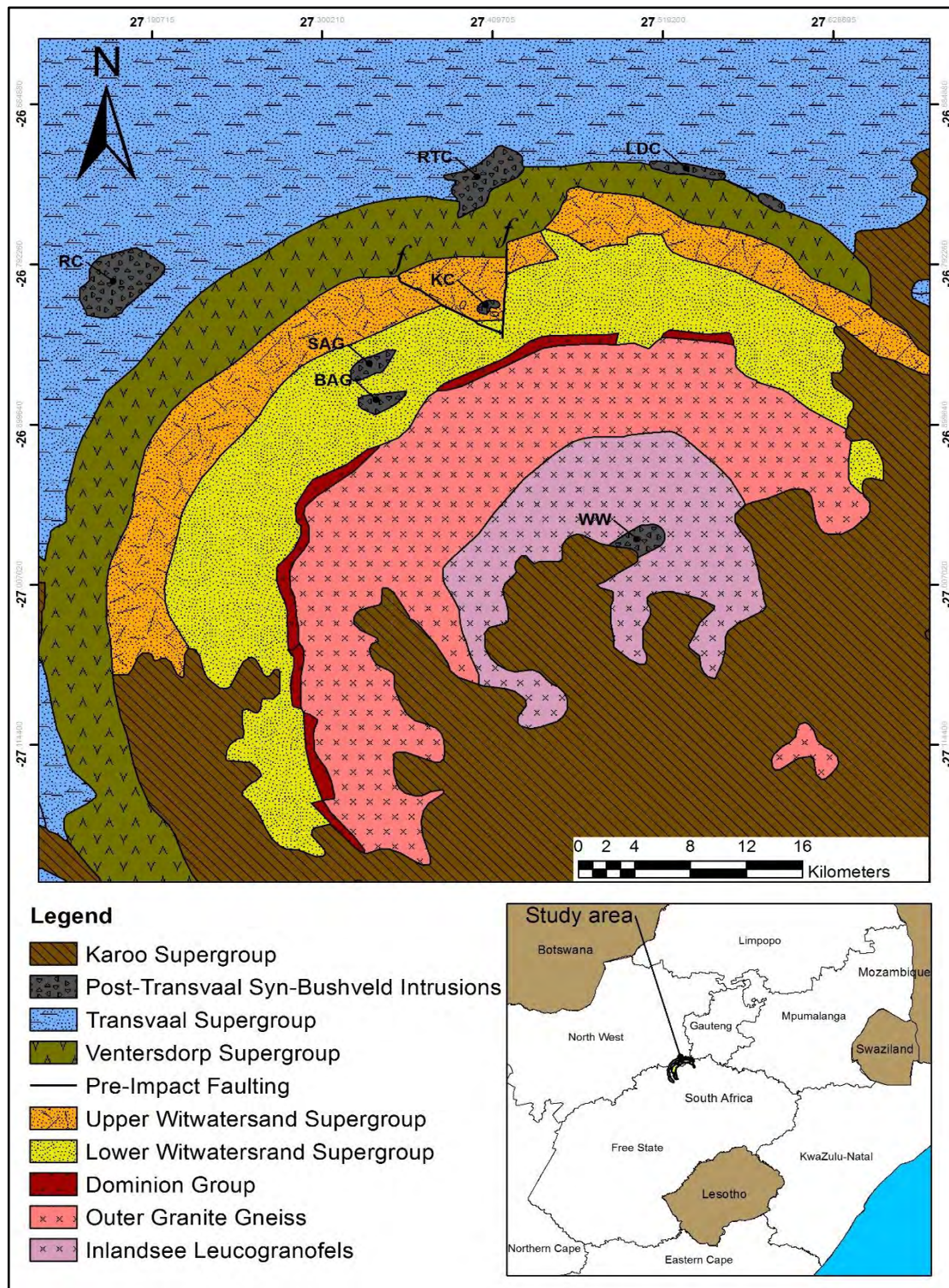


Figure 1: Simplified geological map of the Vredefort dome, depicting the locality of the uplifted structure, and distribution of post-Transvaal syn-Bushveld intrusions. BAG: Baviaan Kranz alkali granite, SAG: Schurwedraai alkali granite, KC: Koedoesfontein Complex, LDC: Lindequesdrift Complex, RC: Roodekraal Complex, RTC: Rietfontein Complex, WW: Winddam wehrlite.

Although the panoramic views of the Vredefort dome are largely dominated by uplifted siliceous supracrustal rocks of the Witwatersrand Supergroup, a variety of plutonic and hypabyssal igneous rock types are distributed amongst the core and collar rocks of the dome (Coetzee, 2001). These include several Neoarchean to Mesoproterozoic dioritic, mafic and ultramafic intrusions comprising sills and dykes, as well as layered differentiated intrusions (Bisschoff, 1969, 1973, 1999; Reimold *et al.*, 2000; Reimold & Gibson, 2005; Graham *et al.*, 2005; Anhaeusser, 2006; de Waal *et al.*, 2006, 2008; de Waal, 2008; Gibson & Reimold, 2008). The intrusions are broadly grouped based on their relationship to the 2023 Ma old impact deformation features of the dome (planar micro-deformation, pseudotachylite and shatter cones), and are referred to as either pre-or post-impact intrusions. Anhaeusser, (2004, 2006) summarises these intrusions as follows:

- (i) “Primitive” mafic to ultramafic intrusions including the Steynskraal Metamorphic Suite and several adjacent dykes within the inner granitoid core termed the Inlandsee Leucogranofels (ILG) (Stepito, 1990; Hart *et al.*, 1990; Bisschoff, 1999).
- (ii) An older and younger set of metamorphosed (also shock-affected) epidiorite intrusions of possible Ventersdorp to lower Witwatersrand age (Bisschoff, 1969, 1973, 1999; Stepito, 1990).
- (iii) A tholeiitic suite of sills, dykes and layered intrusions of possible Bushveld age (Bisschoff, 1969, 1972(a); Coetzee, 2001; Coetzee *et al.*, 2006).
- (iv) A younger suite of post-Transvaal syn-Bushveld age intrusions varying from ultramafic to dioritic sills and dykes, layered intrusions as well as alkali granite complexes (Bisschoff, 1969, 1972(b), 1973, 1999; Clark, 1972; Elsenbroek, 1991; Stepito, 1990; Graham *et al.*, 2005; de Waal *et al.*, 2006, 2008; de Waal, 2008).
- (v) Mafic granophyre dykes representing melt rock produced by the impact event. (Bisschoff, 1996; Therriault *et al.*, 1997; Reimold & Gibson, 2006; Leiger & Riller, 2012).
- (vi) Post-Waterberg igneous intrusions, including the Anna’s Rust gabbroic sheet, Vredefort Mafic Complex and the Wonderfontein nepheline-syenite dyke that crosscuts the Rietfontein Complex (Anhaeusser, 2006; Bate *et al.*, 1995; Bisschoff, 1973, 1999; Reimold *et al.*, 2000).

The Koedoesfontein Complex (KC) in particular, forms part of the younger suite of post-Transvaal syn-Bushveld (2050-2060 Ma) intrusions comprising a variety of rock types. This includes ultramafic to dioritic sills and dykes, as well as peralkaline alkali granite plutons, which

were emplaced in the vicinity of major strike-faults in the dome (Bisschoff, 1969, 1973, 1999; Elsenbroek, 1991; Graham *et al.*, 2005).

Along with the KC the following are included, the Baviana Kranz (BAG) and Schurwedraai alkali granite (SAG), the Lindequesdrift (LDC), Roodekraal (RC) and Rietfontein (RTC) Complexes, which are distributed within the collar of the dome, as well as the Winddam wehrlite (WW) located within the granitoid core (Stephens, 1990) (Figure 1).

1.2 PROBLEM STATEMENT

The primary focus regarding the KC is divided into two main research facets, encompassing both the geological and botanical aspects of the Complex. The study will, therefore, discourse from both a geological and botanical viewpoint.

1.2.1 Geological facet

The majority of post-Transvaal syn-Bushveld intrusions of the Vredefort dome are well known and documented by various authors (Bisschoff, 1969, 1999; Stephens, 1990; Anhaeusser, 2004; 2006; de Waal *et al.*, 2006, 2008). The KC, however, remains understudied amongst these well-known intrusive complexes because of its relatively small size, poor outcrop visibility (mostly attributed to the current erosion level), and dense vegetation cover. Despite having difficulties obtaining visible outcrops, Bisschoff (1969, 1972(b), 1973, 1999) identified several small dyke structured outcrops consisting of wehrlite and dioritic lamprophyre (termed spessartite), as well as numerous dispersed outcrops of sodic-alkali granite, which is believed to represent a larger underlying body that is probably associated with the SAG and BAG intrusions (Bisschoff, 1996). Substantial petrological descriptions of these rock types are presented, however, published geochemical data is generally limited or centred around the alkali granite intrusion and outcrops of wehrlite and dioritic lamprophyres occurring on the north-western slope of the KC (Bisschoff, 1969, 1972(b); de Waal *et al.*, 2008).

Until recently, during an initial geobotanical investigation of the KC (Boneschans, 2012), small unmapped sills of mafic gabbro as well as hornblende, olivine clinopyroxenite (previously identified as wehrlite; Bisschoff, 1969) and quartz alkali syenite were discovered occurring on the northern and north-eastern slope of the Complex respectively. Initial qualitative observations made from these diverse unmapped outcrops indicated a strong structural and mineralogical association with the documented intrusions, and seem to collectively form a rock series which is indicative of magmatic differentiation. It is therefore considered that the KC represents a layered differentiated intrusion quite similar, and probably closely related to the RTC. This

hypothesis still requires further testing and forms the basis for this petrological and geochemical investigation of the KC.

1.2.2 Botanical facet

Because of distinct geochemical characteristics associated with ultramafic rock bodies, they are generally known to host a wide range of uniquely adapted vegetation commonly referred to as “serpentine flora” (Balkwill *et al.*, 1989; Mengel & Kirkby, 2001; Rodenkirchen & Roberts 1993). Studies regarding plant species adaption to ultramafic ecosystems is a broadly researched discipline. In southern Africa, the rich floras of ultramafic areas have received much attention, especially the Great Dyke of Zimbabwe (Balkwill & Campbell-Young, 2001), Barberton Greenstone Belt (Smith *et al.*, 2001), Sekhukhuneland (Siebert *et al.*, 2002), Witwatersrand (Reddy *et al.*, 2001, 2009) and Swaziland (McCallum, 2006). Nonetheless, vegetation characteristics associated with specific ultramafic, mafic and dioritic rock types within the Vredefort dome remains undocumented (Balkwill, 2005). The KC, therefore, was selected as a case study to investigate the possible effect of these rock types, on the floristic and physiognomic features of its supported vegetation.

Initial results (Boneschans, 2012) revealed that the woody vegetation of the KC (*Senegalia* and *Vachellia* species above all) show distinctive changes in density and species composition between ultramafic to felsic rock types. The majority of the species (including *S. caffra* and *V. robusta*) is mainly restricted to dioritic and felsic intrusions, nonetheless, species such as *V. karroo* are unaffected and tend to show a higher species abundance on the ultramafic outcrops. The exact explanation for these abrupt species changes across the varying rock types of the KC is believed to be unclear and requires further study. It is believed to be due to a tolerance mechanism of *V. karroo* to better exclude specific trace elements generally associated with ultramafic rocks (Cr, Co, Ni), as well as to maintain a high leaf Ca:Mg ratio.

1.3 RESEARCH SUBSTANTIATION

This dissertation aims to address the lack of knowledge regarding the aforementioned aspects of the woody vegetation and petrology of the ultramafic to felsic rock associations of the KC. Specifically by providing an updated geological map and lithological record of the KC and studying the relationship between *Senegalia* and *Vachellia* species, and their underlying soil and geology.

Furthermore, by linking the petrological study with the soil and vegetation components of the KC, different correlations can be drawn with regards to useful rock-soil-plant relationships. These associations may reveal important environmental and ecological characteristics

regarding species adaptation to specific toxic trace elements within an ecosystem such as the Vredefort dome, and as a result, provide specific floristic associations which may prove useful for identifying indicator species and phytoremediation applications.

1.4 RESEARCH AIMS

The primary research aims for this study are to:

- (i) Provide an updated detailed geological map and lithological record of the KC, and to determine and discuss the petrological character (structure, mineralogy, and geochemistry) of each intrusive rock type comprising the Complex.
- (ii) Determine and discuss the extent with which the variable geology affects the chemical characteristics of the soil profile, and identify chemical variables that correlate with changes in dominant rock types.
- (iii) Produce checklists of woody species which occur on the KC, and indicator species which may be commonly associated with ultramafic, dioritic and felsic rock formations elsewhere within the Vredefort dome.
- (iv) Investigate the prominence of indicator species on the ultramafic dykes compared to the other rock types present at the KC, and determine the group's ability to tolerate high Mg and toxic trace element concentrations.

1.5 RESEARCH HYPOTHESES

The following basic hypotheses were postulated regarding the geology and woody vegetation of the KC:

- (i) If each dyke and intrusive rock type of the KC follows a coherent structural, mineralogical, and geochemical sequence, which indicates that they are likely related in terms of magmatic origin, then the Complex can be regarded collectively as an overturned, layered intrusive structure.
- (ii) If the variation in the geological bedrock of the KC affects soil chemical properties which limit the spatial distribution of specific associated woody species, then changes in woody vegetation patterns will be associated with specific rock types at the KC.
- (iii) If *V. karroo* is more prevalent on the ultramafic intrusions, and outperforms co-occurring woody vegetation, then the species is an indicator of ultramafic rock types at the KC.

- (iv) If *V. karroo* is more effectively adapted to exclude toxic amounts of major- and trace elements associated with ultramafic rock types, then this species will have a higher Ca:Mg ratio and lower Cr, Co, Ni content than the co-occurring woody vegetation.

1.6 STUDY AREA

The KC is located on the Koedoesfontein 12 farm in North-West, South Africa, roughly 12 kilometres north-northwest from Parys, situated at the following coordinates: 26°49'38.48" S 27°23'46.17" E. The study area is demarcated by the R53 road between Parys and Potchefstroom on the western side, the farm boundary to the north-east, and the Enselspruit (dry river bed) to the south (Figure 2).



Figure 2: Orthophotograph of the Koedoesfontein Complex, illustrating the locality and demarcation of the study area, as well as the density of woody vegetation cover (predominantly *Senegalia* and *Vachellia*). Sparse woody vegetation is observed on the northern slope of the Complex along the dioritic-ultramafic intrusions, as well as the upper parts of the exposed quartzite ridges (Google Earth, 2014).

1.6.1 Geological overview

The Koedoesfontein Complex consists of metamorphosed quartzite and hornfelses (shales) of the upper Neoarchaeon Witwatersrand Supergroup (>2710 Ma), with an altered andesitic epidiorite intrusion considered to be of Ventersdorp age (2714 Ma) (Bisschoff, 1999). Scattered outcrops of a number of small overturned sills of wehrlite, gabbro, hornblendite, spessartite, as

well as sodic quartz alkali syenite and alkali granite, can be observed as part of the Complex. A general summary of the mineralogical composition of each rock type is provided in Table 1. The younger intrusive series intruded through the Main Quartzite Formation of the Johannesburg Subgroup (Bisschoff, 1969) and forms part of the upper Witwatersrand Supergroup that outcrop in the collar of the dome. The alkali granite and closely related quartz–alkali syenite is considered comagmatic with the SAG and BAG intrusions, which together with the ultramafic-intermediate intrusions, are accepted as part of the Bushveld magmatic event (2050–2060 Ma, Graham *et al.*, 2005; Anhaeusser, 2006, de Waal *et al.*, 2006, 2008). The presence of shatter cone structures and small pseudotachylite veins within the intrusive rocks comprising the Complex (Bisschoff, 1999), as well as the deformation of the Witwatersrand rocks, confirms the pre-impact occurrence of these intrusions.

Table 1: Summary of the general mineralogy of the igneous and metamorphic rock types constituting the Koedoesfontein Complex.

Rock type	Mineralogy	
	Major	Minor
Alkali granite	Quartz, microcline-feldspar, Albite-plagioclase	Na-rich pyroxene and/or amphibole, muscovite
Quartz alkali syenite	Albite-plagioclase, microcline-feldspar,	Quartz, Na-rich pyroxene and/or amphibole, muscovite
Spessartite	Hornblende, plagioclase (Ca < Na)	Augite, apatite, chlorite, muscovite, quartz
Hornblendite	Fe-Hornblende	Augite, olivine, magnetite, talc
Opx-Hbl gabbro	Augite, hornblende, plagioclase (Ca > Na)	Orthopyroxene, magnetite, quartz, talc
Wehrlite/ Ol ± Clinopyroxenite	Olivine, augite	Hornblende, magnetite, serpentine, talc
Epidiorite	Actinolite, plagioclase (Ca<Na)	Augite, epidote, chlorite, magnetite, quartz
Hornfels	Cordierite, cummingtonite, anthophyllite	Quartz, Ca-rich plagioclase
Quartzite	Quartz	Hematite, muscovite, biotite,

1.6.2 Soil overview

The majority of the preserved and metamorphosed igneous intrusions that comprise the KC are poorly exposed, owing to their thin outcrop thicknesses and relatively high weathering rates. The outcrops are predominantly covered by their residual soils mixed with colluvial and alluvial quartzite overburden (Figure 3). Apart from areas situated on the crest and scarp of the ridge, soil cover occurring along the mid-slope down towards to the valley is fairly well developed,

comprising largely red-brown gravelly topsoil, situated on various weathered saprolitic substrates and/or bedrock.

Because of the substantial range in rock lithology, an extensive mineral gradient exists within the soil's parental material, ranging from purely siliceous to sodic-alkaline, to ferromagnesian dominated silicate minerals (Table 1). The presence and abundance of these minerals which is determined by the prevailing rock type(s), is expected to produce similar gradients within the soil chemistry and, ultimately, the supported vegetation.



Figure 3: Typical red-brown residual and colluvial soil cover associated with the ultramafic-dioritic intrusions of the Koedoesfontein Complex. Note the patchy distribution of the hosted grassy vegetation, which is frequently encountered on weathered wehrlite, Ol $\pm$ clinopyroxenite, and hornblendite outcrops.

The soil cover of the study area is broadly classified utilizing the Taxonomic Classification System for South Africa (MacVicar *et al.*, 1991). The shallow (roughly 30 cm deep) rocky soils on the ridges and mid-slopes of the Complex vary between a Mispah and Glenrosa soil forms. Along the foot-slopes and valley, bordering the dry river bed, reddish alluvial soils consisting of a deeply weathered Hutton soil form are encountered.

On the southern aspect of the quartzite ridge, (along the mid-slope), thin highly organic topsoil is produced as a result of higher vegetation densities, as well as a lower and more humid micro-

climate. These quartzite-rich colluvial soils are shallow and rocky (likewise classified as a Mispah soil form). However, the topsoil is strikingly darker as a result from accumulated soil organic material. This characteristic is also observed in the soils along the alkali granite, quartz alkali syenite and altered wehrlite intrusions on the north-eastern slope (Figure 4). Similarly, this area corresponds with higher densities of woody vegetation, which seems to be related to more favourable soil conditions.



Figure 4: Typical dark-brown organic topsoil found on the north-eastern slope of the Koedoesfontein Complex, occurring along the alkali granite, quartz alkali syenite and altered wehrlite intrusions.

1.6.3 Vegetation type and climate

Due to the geo-diversity of the Vredefort dome, along with its related soil, topography, and microclimate, as well as the proximity of the Vaal River, distinct conditions are produced under which different plant communities have established themselves (Balkwill, 2005). The vegetation map compiled by Mucina and Rutherford (2006), shows correspondence between the spatial distribution of major vegetation communities in the dome, and the outcrop patterns of the dome's major geological units. For example, the Andesite Mountain Bushveld vegetation unit (SVcb11), occurs along the andesitic outcrops of the Ventersdorp Supergroup, and the Gold Reef Mountain Bushveld vegetation unit (SVcb9), is mainly situated on the quartzite and shales of the

Witwatersrand Supergroup. It is therefore apparent that the geology is a key factor controlling species distribution and physiognomy in the dome.

A similar observation was previously noted by Mogg (1961) who indicated floristic variations amongst andesite, norite and quartzite outcrops within the Witwatersrand area. Bezuidenhout and Bredenkamp (1988) also supportively concluded from vegetation studies on the Baviaanskraan and Schurwedraai granitic intrusions, south-west of the KC, that there is a definite correlation between the distribution of plant communities in the dome and specific smaller scale geological formations. Several indicator species including *Combretum molle*, *Nuxia congesta*, *Senegalia caffra* and *Searsia magalismontana*, were identified for quartzite outcrops and quartzitic substrates, whereas *Diospyros lycioides* subsp. *lycioides*, *Sporobolus fimbriatus*, and *Ziziphus mucronata* were identified as indicator species for alkali granite outcrops. *Vachellia karroo* and *Z. mucronata* were also documented as indicators for the lower lying alluvial soil habitats.

The KC, is situated within the Gold Reef Mountain Bushveld vegetation unit (SVcb9) (Mucina & Rutherford, 2006). The area is generally characterised by rocky hills and ridges of quartzite, containing higher densities of woody vegetation on the south-facing slopes (Figure 5), and is associated with distinct floristic differences, such as *S. caffra* which displays higher abundances on the southern aspects. The area receives summer rainfall (600-750 mm per annum), with very dry winters and frequent frost (especially on the south facing slopes), but less common on the ridges and hills (Mucina & Rutherford, 2006).



Figure 5: The southern aspect of the study area, depicting the high density of woody vegetation on the southern slope (left) of the quartzite ridge, and typical *V. karroo* (stands in front) which illustrates a high abundance on the alluvial soils.

The study area is specifically dominated by three microphyllous woody species, namely *S. caffra*, *V. karroo* and *V. robusta*. These species possess reduced foliar surfaces and deep-penetrating taproot systems and are generally resistant to drought and high concentrations of trace elements (Nyakudya *et al.*, 2011). *S. caffra* is generally restricted to the quartzite-rich environments, whereas *V. karroo* and *V. robusta* contribute the greatest portion of woody vegetation on and around the intrusive rock types, as well as the lower lying alluvial soils.

Other relevant woody taxa of the study area include (Mucina & Rutherford, 2006):

Tree species: *Celtis africana*, *Dombeya rotundifolia*, *Pappea capensis*, *Searsia leptodictya* and *Vangueria parvifolia*.

Tall woody shrubs: *Asparagus laricinus*, *Ehretia rigida*, *Euclea crispa* subsp. *crispa*, *Grewia flava*, *Grewia occidentalis*, *Gymnosporia buxifolia*, *Mystroxydon aethiopicum*, *Pavetta zeyheri* and *Searsia pyroides*,

Small woody shrubs: *Asparagus suaveolens* and *Searsia rigida*.

1.7 FORMAT OF DISSERTATION

This dissertation is compiled according to the standard guidelines set forth by the North-West University (NWU) and comprises seven chapters. All cited literature is recorded in a reference list at the end of the manuscript. Sampling coordinates, raw analysis results, maps and supplementary material are provided as Annexures. The preliminary data from chapter 5 and 6 have been published (Boneschans *et al.* 2015).

Chapter 2: Relevant background literature regarding each research facet is reviewed and discussed. This includes a general description and discussion of post-Transvaal syn-Bushveld intrusions, physical, chemical and ecological features of ultramafic soils and serpentine flora, as well as previous findings related to trace element studies on *Senegalia* and *Vachellia* species.

Chapter 3: The method and materials for investigation are validated and discussed. The discussion reviews sampling techniques used to obtain rock, soil and plant samples from the KC, sample preparation procedures, analyses and analytical equipment utilised, as well as statistical analysis of botanical and environmental data.

Chapter 4: The results of the petrological investigation of the KC are presented and discussed. An updated geological map is included, along with photomicrographs and corresponding descriptions of the structure, mineralogy and geochemistry of each intrusive and metamorphic rock type.

Chapter 5: The results of the soil chemical analyses of the KC are presented. A discussion of the soil chemical data including the pH, organic carbon, cation exchange capacity, macro-, micro- and phytotoxic trace elements are provided.

Chapter 6: The results of the botanical investigation of the KC are presented and discussed. Species and plot ordination results are provided, along with the total trace element analysis results.

Chapter 7: Conclusions and recommendations of the study are discussed and the study hypotheses are evaluated. The closing discussion is presented by means of an integrated summary of the chapters that relate the geological (Chapter 4), soil chemical (Chapter 5) and botanical (Chapter 6) features of the KC. Research recommendations and closing notes are also included.

CHAPTER 2

LITERATURE REVIEW

2.1 INTRODUCTION

In order to clearly recognise the geological setting of the KC, and how the geology of the Complex relates with the larger groups of post-Transvaal syn-Bushveld igneous intrusions within the dome, the following discussion will provide a geological overview of this specific group of intrusions, and assess specific connections / similarities between the KC and the surrounding intrusive complexes. (Specific descriptions of the KC, however, will be covered in chapter 4, and will only be referred to in this chapter). Secondly, to define the physical, chemical and ecological characteristics that are associated with ultramafic-mafic rock bodies, literature is reviewed on ultramafic-mafic related soils and their associated vegetation, which is referred to as “serpentine- or ultramafic flora”. In addition, a review on relevant *Senegalia* and *Vachellia* spp. (previously *Acacia* spp.) is presented in order to draw attention to their general taxonomy as well as previous nutrient and trace element studies conducted on these species in similar environments.

2.2 POST-TRANSVAAL SYN-BUSHVELD AGE INTRUSIONS OF THE VREDEFORT DOME

The KC forms part of a mafic-dioritic-peralkaline group of post-Transvaal (c. <2150 Ma) syn-Bushveld (c. 2050-2060 Ma) intrusive complexes, which outcrops within the core and collar rocks of the Vredefort dome (Figure 1). This includes the BAG and SAG intrusions, which are now collectively termed the Schurwedraai Complex (SC), the Roodekraal- (RC), Rietfontein- (RTC), Lindequesdrift Complexes (LDC), as well as the Winddam Wehrlite (WW) (Stepito, 1990; Bisschoff, 1999).

Apart from the WW (Stepito, 1990), which was later included, Bisschoff (1969) originally grouped the former intrusions based on structural relationships as well as geochemical and mineralogical features. The term “rock-suite” was used to broadly group these rocks based on associations, however, no genetic relationships were initially implied. As a result, a distinction was made between the younger mafic-dioritic-peralkaline intrusions (termed the “alkaline suite”) from older magnesium- and silica-rich tholeiitic intrusions (termed the “tholeiitic suite”), which were established to coincide with earlier stages of the Bushveld-magmatic event (Bisschoff, 1972(a); Anhaeusser, 2006; Coetzee *et al.*, 2006). The mafic tholeiitic intrusions have a strong noritic character which is mineralogically similar to parts of Critical- and Main Zone of the Rustenburg Layered Suite, comprising primarily norite with orthopyroxenite, dolerite-norite, and dolerite

intrusions. By sub-dividing the tholeiitic group based on mineralogical composition, Coetzee (2001) identified three related intrusions, particularly the Wittekopjes norite, the poikilitic Parsons Rust dolerite-norite and the Reebokkop dolerite.

Succeeding the tholeiitic suite, the alkaline suite is believed to correspond with later stages of the Bushveld-magmatic event. This relative age relationship was first confirmed by the alkali granite intrusions of the SC and RTC, which were observed crosscutting the rocks of the tholeiitic suite (Bisschoff, 1969). Hargraves (1970, 1987) illustrated that the LC, RTC and SC bears a contemporaneous age and palaeomagnetic vector relation to the Bushveld Igneous Complex (Bisschoff, 1973; Elsenbroek, 1991).

The alkaline suite has a dioritic character similar to the Upper Zone of the Rustenburg Layered Suite, comprising dioritic and mafic rock types which contain albitic plagioclase ($An_{<40}$), whereas the rocks of the tholeiitic suite predominantly contain labradoritic plagioclase (An_{50-70}) (Coetzee, 2001). Orthopyroxene, which is common in the rocks of the tholeiitic suite, is also virtually absent in the rocks of the alkaline suite (which predominantly contain augite clinopyroxenes) (Bisschoff, 1969; Coetzee, 2001).

More recently, de Waal *et al.* (2006) and de Waal *et al.* (2008) postulated that the majority of the intrusive complexes from Bisschoff's (1969) proposed "alkaline suite", indicate a genetic relationship with adjoining intrusions to the north and north-east of the Vredefort structure. These include the Marble Hall sills and related intrusives, the Basal Gabbro Unit of the Uitkomst Complex, the Schoongezicht Complex, the Heidelberg intrusions, and the high-Ti basalt and FeTiP basaltic lava of the Rooiberg Group. These intrusions were grouped along with the intrusions of the "alkaline suite" based on similar age constraints, geochemical trends and corresponding petrographical observations. de Waal *et al.* (2006) and de Waal *et al.* (2008) proposed the existence of a high-Ti igneous suite of mafic to felsic rocks (HITIS), that emplaced in coeval with earlier stages of the Bushveld Magmatic event around ~ 2055 Ma. The KC, in particular, is also preliminarily included in the HITIS based on petrographic observations made by Bisschoff (1973, 1999), which share similarities with the RTC and LDC. It was stated by de Waal *et al.* (2008), however, that the published information on the KC is too limited to make any conclusive interpretations.

Trace element analyses (de Waal *et al.*, 2008) indicated that the proposed HITIS is derived from an alkali to traditional basaltic parental magma, and comprises predominantly of mafic-dioritic rocks containing clinopyroxene, magnetite-ilmenite, amphibole, fosteritic olivine, orthopyroxene and albite plagioclase ($An_{<50}$), as well as felsic rocks of syenodiorite and alkali granite with K-feldspar, albite, sodic-pyriboles, and biotite.

The main focus of this study is on the intrusive complexes situated within the Vredefort structure and, therefore, excludes the surrounding intrusions grouped within the HITIS (de Waal *et al.*, 2006; de Waal *et al.*, 2008). The collective term “mafic-dioritic-peralkaline series” (MDPS) will be used to refer to intrusions contained within the Vredefort dome impact structure.

2.2.1 Distribution, field relations and petrography of the mafic-dioritic-peralkaline series

The MDPS intrusions (particularly the RC, RTC, KC, and SC) are situated near fault zones within the dome and occur predominantly along major strike and dip faults. These faults displaced tholeiitic rocks, but not the intrusions of the younger MDPS (Bisschoff, 1999). Therefore, the emplacement is thought to have been controlled to some extent by pre-impact faulting (Bisschoff, 1969; Elsenbroek, 1991; Graham *et al.*, 2005). This observation is confirmed by the sharp contacts between the alkali granite intrusions of the SC and the Witwatersrand country-rock, which is ascribed to forceful injection of magma, along with the plane of these faults, as well as secondary faults and fractures (Bisschoff, 1969, 1973; Elsenbroek, 1991).

Elsenbroek (1991) indicated from the distribution and orientation of localized shattercone structures, that the major strike-slip fault occurring at the SC, which duplicates the Jeppestown Subgroup, must be of pre-impact (c. 2023 ± 4 Ma; Kamo *et al.*, 1996) tectonic-origin. It is, therefore, reasonable to assume the alkali granite plutons must have been emplaced during an earlier stage of localized folding and deformation within subgroups of the Witwatersrand Supergroup prior to the impact event.

This notion is in agreement with the findings of Mayer and Albat (1990) which investigated the geometry of the axial orientations of small-scale folds in banded ironstone formations of the lower Hospital Hill Subgroup. Their findings indicated that the small-scale folds, and in some localities small thrust faults, existed prior to the impact event. This was confirmed by the presumably transgressive nature of an emplaced epidiorite sill, as well as the presence of shatter cones and planar shock fractures, which cut across the small-scale folding. It was thus concluded by Mayer and Albat (1990) that the Witwatersrand Supergroup was likely subjected to a compressive event directed approximately north-east south-west, prior to the outpouring of the Ventersdorp lavas.

Corresponding to the findings of Mayer and Albat (1990), a similar conclusion can be drawn with regards to two major faults, occurring on either side of the KC in the northern collar of the dome (Figure 1). As indicated by Tinker *et al.* (2002) and later supported by Wieland *et al.* (2005), these particular faults, displacing upper Central Rand Group rocks, are believed to be of pre-Transvaal age, having a slip, orientation and timing consistent with Ventersdorp-age rifting (c.

2714 Ma). The larger fault occurs approximately 1.5 km east of the KC and is north north-east to south south-west trending, whereas the second fault lies 2 km to the west, and trends north-west to south-east. Subsequent emplacement of the KC along the plane of these faults as well as associated secondary fault and fracture systems would thus seem to fit its structural setting, as well as conform to the age constraints for emplacement (c. 2050-2060 Ma).

2.2.1.1 The Rietfontein Complex (RTC)

The RTC is located roughly ~20 km north north-west of Parys on the farms Rietfontein 446 and 447 (Bisschoff, 1973, 1999; Anhaeusser, 2006; de Waal *et al.*, 2008). Covering about ~18 km², the RTC consists of a southern portion of layered, differentiated, mafic-ultramafic intrusions, and a northern portion of alkali granite, which grades into a scarcely distributed syenitic phase. The emplaced bodies are elongated along the strike of the overturned country rock, intruding into the upper lavas of the Ventersdorp Supergroup, as well as the Black Reef Formation and dolomite of the Chuniespoort Group. The ultramafic to mafic portion of the RTC consists mainly of wehrlite and olivine diorite, with subordinate augite-bearing- and non-augite-bearing troctolites. These intrusions are essentially seen as two distinct series of wehrlite and olivine diorite, and troctolite and olivine diorite which differentiated as separate intrusive bodies (de Waal *et al.*, 2008). Bisschoff (1972b, 1973) also reported finding loose blocks of clinopyroxenite on the northern contact of the eastern wehrlite outcrops, as well as samples of porphyritic spessartite which seem to be related to the olivine diorite.

Three types of wehrlite were initially recognised, distributed amongst the RTC and KC (Bisschoff, 1969, 1973). The first type, containing interstitial olivine, occurs as an overturned sill on the KC, whereas the second and third type (concluded as a result of magma pulsing) are found in the southern mafic portion of the RTC, grading into olivine diorite. According to Bisschoff (1969, 1999) the wehrlite intrusions of the KC, are possibly related to the mafic portion of the RTC and may represent remnants of a larger differentiated intrusion. If this hypothesis is correct, it has been suggested that the parental magma which could have produced these intrusions, should have had a similar composition to the magma which produced the olivine diorite and wehrlite series of the RTC. In close association with the ultramafic-mafic intrusions, de Waal *et al.* (2008) states that the petrographic similarity between the alkali granite of the RTC and the Baviaanskraal pluton (BAG) indicates a magmatic co-lineage with the felsic intrusions of the LDC.

Petrography

The wehrlite of the RTC is distributed along southern and middle-northern parts of the mafic portion, comprising augite (~65 modal %), olivine (~20 modal %) and plagioclase (An₂₀ - <5

modal %), with late-stage brown amphibole-hornblende (~4%) and magnetite (~1%) as accessory. The southern contact of wehrlite (the second type) is medium-grained allotriomorphic granular, while the third type of wehrlite is a hypidiomorphic granular and contains between 5 – 10 modal % plagioclase (An_{20}) which poikilitically enclose the olivine and augite, and in some cases orthopyroxene.

The olivine diorite is medium-grained hypidiomorphic granular and comprises olivine (~40 modal %), clinopyroxene (~30 modal %), plagioclase (~20 modal %), amphibole (~8 modal %) and magnetite (~1 modal %). The magnetite is either enclosed by olivine and clinopyroxene, or interstitial.

In contrast, the troctolite consists of small olivine crystals (~30-70 modal %) which are poikilitically enclosed in clinopyroxene (~2-10 modal %) and feldspar (~25 modal %), and contain accessory magnetite.

Bisschoff (1972) describes a sample of spessartite found at the RTC as composed of subhedral to euhedral brown amphibole phenocrysts (<10 mm) set in a medium-grained ground-mass of plagioclase (An_5) and augite. The augite is partly altered to green hornblende, and in some cases are poikilitically enclosed by the plagioclase.

The alkali granite and subordinate alkali-syenitic rocks are medium-grained hypidiomorphic to allotriomorphic and mainly composed of alkali feldspar (microcline-antiperthite), quartz and sodic-amphibole (magnesio-arfvedsonite). Sodic-pyroxene, specifically aegirine is also a constituent of the syenitic rocks, with magnetite, sphene, and biotite as an accessory.

2.2.1.2 The Koedoesfontein Complex (KC)

As the KC forms the main focus of the study an overview of the distribution and geology will be presented in Section 4.1 & 4.2. The petrographic descriptions follows in section 4.3.

2.2.1.3 The Winddam Wehrlite (WW)

The WW intrusion outcrops within the core of the dome, approximately 10 km south south-east of Parys on the farms Windam 124 and Oakdale 349 (Stepito 1990, Bisschoff, 1999; Anhaeusser, 2006). The intrusive body measures 1.7 km long and 0.7 km across and comprises predominantly wehrlite which is considered to be similar and probably related to the RTC and KC. In accordance with Bisschoff (1996, 1973), Stepito (1990) (as cited by Bisschoff, 1999 & Anhaeusser, 2006) proposed that the WW was “coeval” with the wehrlite intrusions of the RTC and KC. No pseudotachylite veins, however, have been reported within the WW samples (Stepito 1990). Bisschoff (1999) stated that this is mainly attributed to lack of significant outcrop exposure

since veins of pseudotachylite produced by the impact event are commonly encountered within this area.

Petrography

The wehrlite varies from fine-grained at the contacts to coarse-grained towards the centre of the body, and comprises augite, olivine and skeletal altered amphibole which is surrounded by poikilitic inclusions of plagioclase (Bisschoff, 1999).

2.2.1.4 The Roodekraal Complex (RC)

The RC occurs ~10 km south-east of Potchefstroom on the farms Roodekraal 37, Plessiskraal 43 and Prinsloosrust 93 (Bisschoff, 1972 b, 1999; Clark, 1972; de Waal *et al.*, 2006, 2008; Anhaeusser, 2006; de Waal, 2008). The intrusion unconformably overlies the Pretoria Group sediments, occupying an estimated area of 12 to 15 km². The RC has a general concentric distribution, and comprises porphyritic and amygdaloidal sodium-rich andesite into which a core of varying porphyritic, medium- to fine-grained diorite and dykes of albatite are intrusive. A fine-grained diorite was dated at 2053.9 ± 9.2 Ma (SHRIMP zircon ²⁰⁷Pb/²⁰⁶Pb; de Waal *et al.*, 2006, 2008).

Petrography

The andesite is described as amygdaloidal to massive, porphyritic to non-porphyritic with albite-oligoclase (An₀₋₃₀) phenocrysts that are generally tabular (<5 mm). The groundmass is composed of plagioclase, amphibole (particularly ferrorichterite and ferrowinchite; de Waal *et al.*, 2006), augite, magnetite-ilmenite, biotite, and apatite. Common alteration products include chlorite, actinolite, sericite, calcite, sphene, epidote-clinozoisite and small amounts of quartz.

The diorite is composed of zoned plagioclase (An₀₋₃₀) (~60-70 Modal %), augite, magnetite, biotite, apatite and seldom minor traces of olivine and K-feldspar. Late-stage alteration of the augite to amphibole was also observed. Alteration minerals include chlorite, sericite, calcite, sphene, epidote-clinozoisite, magnetite and small amounts of quartz. The coarse to medium grained diorite is porphyritic with tubular (<15 mm long) plagioclase phenocrysts, and generally greater amounts of augite. The fine-grained diorites are equigranular and are mainly composed of plagioclase (non-tubular) and amphibole.

Clark (1972) as referenced by de Waal *et al.*, 2006, reported samples of layered olivine diorite occurring in a 30 m borehole section at the RC. The layered sequence comprises of a magnetite-rich (<40 Modal %) basal layer, followed by a clinopyroxene-rich (<80 modal %) lower and middle layer, and a top layer containing plagioclase (<45 modal %), orthopyroxene (~40 modal

%) and olivine (<10 modal %). Late-stage amphibole was also noted replacing the clinopyroxene.

The youngest dykes of albatite are porphyritic and comprise of zoned plagioclase (An₀₋₃₀) phenocrysts with cores of oligoclase grading into rims of albite. The phenocrysts are set into a groundmass of plagioclase, containing accessory minerals such as quartz, chlorite and amphibole particularly barkevikite (Bisschoff, 1972b).

2.2.1.5 The Lindequesdrift Complex (LDC)

The LDC is situated about 20 km north-east from Parys on the farms Oorbietjesfontein 20, Woodlands 192, Boschdraai 33 and Vaaldraai 277 (Bisschoff, 1972 b, 1999; de Waal *et al.*, 2006, 2008; Anhaeusser, 2006). The LDC comprises of an 11 km long elongated body with an estimated thickness of 150 to 200 m, which was emplaced into the dolomite of the Chuniespoort Group of the Transvaal Supergroup. It principally consists of equigranular to porphyritic lamprophyre (spessartite), with cross-cutting dykes and sills of diorite, syenodiorite and albite-syenite (normatively leucodiorite - Ab>Hb). The spessartite contains feldspathic pegmatitic schlieren and is similar to the spessartite of KC and RC. Veins of pseudotachylite are present signifying pre-impact occurrence. A SHRIMP zircon ²⁰⁷Pb/²⁰⁶Pb date for the spessartite revealed an age of 2054.8 ± 5.7 Ma (de Waal *et al.*, 2006, 2008).

Petrography

The spessartite consists of euhedral to anhedral amphibole phenocrysts ranging from 1-20 mm in diameter, set within a fine- to medium-grained groundmass of augite, amphibole, sodic-plagioclase (An₀₋₄₅), biotite, magnetite-ilmenite, and apatite. The phenocrysts vary from yellow-brown edenite and hastingsite that grades into green rims of magnesiohastingsite and magnesiohornblende (de Waal *et al.*, 2006). Two main varieties of spessartite are distinguished. The more primitive type is porphyritic, containing larger phenocrysts set within a darker medium-grained ground-mass with higher amounts of augite. While the more evolved type is equigranular and contains less augite, more plagioclase and larger acicular amphibole grains within the ground-mass.

The diorite and subordinate differentiates of syenodiorite and albite-syenite comprise primarily sodic-plagioclase (An₀₋₄₅) and brown-green hornblende, with accessory minerals of augite, magnetite, apatite, micro-perthite, sphene, and quartz. The syenodiorite is hypidiomorphic granular with alkali feldspar as a cumulus phase, whereas the albite-syenite is allotriomorphic-granular comprising mainly albite and hornblende (20-30 Modal %).

2.2.1.6 The Schurwedraai Complex (SC)

The SC consists of two related peralkaline plutonic bodies, occurring along the Vaal River ~10 km west north-west from Parys (Bisschoff, 1973, 1999; Elsenbroek, 1991; Graham *et al.*, 2005). The SC comprises primarily of sodic-alkali granite with associated rocks and xenoliths of albite- and nepheline syenite (termed mariupolite). Bisschoff (1999) conveys that the alkali granite pluton which intrudes through the ultramafic to mafic rocks of the RTC may be correlated with the SC, as well as the small scattered outcrops of alkali granite on the KC, which illustrate a strong association with spessartite and wehrnite intrusions. The alkali granite was dated at 2052 ± 14 Ma. (SHRIMP zircon U-Pb; Graham *et al.*, 2005).

Petrography

The alkali granite is hypidiomorphic granular with interlocking grains of albite (~43 modal %), quartz (~25 modal %), microcline-feldspar (~22 modal %), aegirine and arfvedsonite (~9 modal %), and biotite (~1 modal %). Magnetite, ilmenite, apatite, and zircon are minor constituents. Near the contact zones the alkali granite grades into an albitic-alkali syenite containing more albite (~59 modal %) and aegirine/arfvedsonite (~20 modal %), and less microcline (~19 modal %) and quartz (~2 modal %).

The nepheline syenite has a more coarsely-grained allotriomorphic texture compared to the alkali granite and contains subhedral to euhedral crystals of albite (~55 modal %), nepheline (~25 modal %), microcline (~15 modal %), aegirine (~4 modal %) and minor amounts of hornblende. Accessory minerals include magnetite, ilmenite, and sphene.

Based on feldspar phase relations, the alkali granite of the RTC is a hypersolvus granite containing mainly orthoclase antiperthite. The intrusions of the the SC, however, are subsolvus granites, containing both albite and microcline. Their feldspars contain no exsolution textures however, perthite and antiperthite were found in some samples of the SC (Bisschoff, 1999). The difference in feldspars of the alkali granite from the RTC and the SAG and BAG complexes are mainly attributed to differences in H₂O pressure during crystallisation. This pressure difference most likely resulted from variances in emplacement depth before over folding (Impact deformation). The alkali granite of the KC illustrates characteristics which are intermediate between the RTC alkali granite, and the BAG and SAG intrusions. According to Bisschoff (1973), a sample of alkali granite found from a water borehole on the KC contains albite crystals with no exsolved microcline, which are enclosed or partly enclosed by microcline perthite. Pure albite crystals with no exsolution textures can only be expected in subsolvus rocks or hypersolvus rocks with very high albite concentrations, while the perthite can only occur in hypersolvus rocks (Bisschoff 1973). As an explanation for this occurrence, it was suggested that conditions must

have changed during emplacement and crystallisation. This was probably caused by assimilation of potassium rich country rock, which increased the K_2O , Al_2O_3 and SiO_2 concentration of the magma during emplacement.

2.2.2 Conclusion

The general consensus among the reviewed authors is that the MDPS intrusions occurring in the rim and core of the Vredefort dome are closely related, and the series of intrusions comprising the KC illustrate a strong structural and mineralogical association with the rest of the complexes within this group. From past studies, however, it seems that a large portion of the KC's lithology and structure has been overlooked, since only limited descriptions for this Complex are available. These factors restricts efforts to determine how closely the KC is related to the surrounding intrusions, as the Complex still remains vaguely understood.

From a structural perspective, it would seem the emplacement of the majority of the MDPS intrusions are controlled by pre-impact fault-systems, originating within the Witwatersrand Supergroup, possibly as a result from Ventersdorp-age rifting as well as later post-Transvaal tectonic-reactivation. This observation also explains why the complexes generally occur in the upper Witwatersrand Supergroup, with the exception of WW, which occurs in the Archean core rocks.

Several key mineralogical characteristics could be identified which can link the intrusions of KC to several of the related MDPS intrusions. In an attempt to summarise the corresponding rock types of each complex, a proposed lithostratigraphic column is presented in Table 2. When both the documented and undocumented intrusions of the KC are considered, in relation to the adjoining complexes, it appears that this Complex can provide valuable information with regards to unraveling the relationship between the MDPS intrusions as well as the related intrusions included within the HITIS Suite.

Table 2: Proposed lithostratigraphic summary for the reviewed mafic-dioritic-peralkaline (MDPS) intrusions of the Vredefort dome.

	Windam Wehrlite (WW)	Koedoesfontein Complex (KC)	Rietfontein Complex (RTC)	Roodekraal Complex (RC)	Lindequesdrift Complex (LDC)	Schurwedraai Complex (SC)	Mineral composition*
Peralkaline		Alkali granite (Mc-perthite, Ab) ^{1,4,7}	Alkali granite (Ab-antiperthite) ^{1,4,7,10,11}			Alkali granite (Mc ± perthite, Ab ± antiperthite) ^{1,4,6,7,8,11}	Qz, Mc, Ab, (± Na, Fe)-Amp, Na-Px, Mag, Bt, Ap, Spn, Zr
		Qz Alkali-syenite (Ab)*	Qz Alkali syenite (Ab) ^{4,10}	Albite syenite ^{1,2,7,10}	Albite syenite ^{2,7,9,10,11}	Alkali syenite (Ab) ^{1,4,6,7,8}	Ab (An ₀₋₁₀) > Mc, ± Qz, Na-Amp, Na-Px, Mag
						Albite-Nepheline-syenite ^{1,4,6,7,8}	Ab (An ₀₋₁₀), Mc or Nph, ± Qz, Amp, Na-Px, Spn, Mag
Dioritic					Syenodiorite ^{2,7,9,10,11}		Pl (An<50) > Aug, Amp, Or ± Mag, Bt, Ap,
				Sodic-Andesite (Volcanic) ^{2,3,7,9,10,11}	Diorite ^{2,7,9,10,11}		Pl (An<50) > Aug, Amp ± Or, Mag, Bt, Ap,
		Leuco-spessartite ^{1,4,7}	Spessartite ^{1,4}	Diorite (± Or) ^{2,3,9,10,11}	Spessartite ^{1,2,7,9,10,11}		Pl (An<50), Amp, and/or Aug ± Or, Mag, Bt, Ap
		Hornblendite cumulate (± Pl)* / Mela-spessartite*			Hornblendite / "Dark Spessartite" (± Pl) ¹		Amp > Aug ± Pl (An<50),
			Ol Diorite, (± Amp, Opx) ^{1,4,7,10,11}	Ol Diorite / Websterite cumulate (± Amp, Opx) ^{3,9,11}			Aug > Ol, Pl (An<50) ± Amp, Opx, Mag, Ap, Bt, Spn
Mafic		Opx Hbl Gabbro*					Aug, Pl (An>50) ± Amp, Opx, Mag, Ap
			Troctolite (± Aug) ^{1,4,7,10,11}				Ol, Pl (An>50) > Aug ± Mag, Amp, Bt
Ultramafic		Uralitite*					Amp ± Aug, Ol, Mag, Pl
		Ol ± Clinopyroxenite* (± Amp, Pl) /	Ol ± Clinopyroxenite (± Amp, Pl) ⁴	Ol Diorite (?) - Borehole (Aug <80 modal%) ^{3,9,11}			Aug >> Ol ± Amp, Mag, Pl
	Wehrlite (± Pl, Opx) ^{5,7}	Wehrlite (± Pl) ^{1,4,7}	Wehrlite (± Pl An ₂₀ , Opx) ^{1,4,7,10,11}	Ol Diorite (?) - Borehole (Mag <40 modal%) ^{3,9,11}			Aug > Ol ± Amp, Mag, Pl
		Mela-Dolerite*					Mag, Aug ± Amp, Pl

¹. Bisschoff (1969); ². Bisschoff (1972 b); ³. Clark (1972); ⁴. Bisschoff (1973); ⁵. Stepto (1990); ⁶. Elsenbroek (1991); ⁷. Bisschoff (1999); ⁸. Graham *et al.* (2005); ⁹. de Waal *et al.* (2006); ¹⁰. Anhaeusser (2006); ¹¹. de Waal *et al.* (2008); *Undocumented intrusions included by this study (2017).

2.3 ULTRAMAFIC SOILS AND SERPENTINE FLORA

2.3.1 Introduction

Ultramafic or “serpentine” soils are derived from weathered ultramafic rocks. The term ultramafic refers to a mineral content of more than 90% ferromagnesian silicate, or “mafic” minerals (Winter, 2010). Therefore, the majority of minerals present in this rock type contain a high amount of iron (Fe) and magnesium (Mg) relative to silicon (Si) and include olivine, pyroxene, hornblende, biotite, phlogopite, talc, and serpentine. The term serpentine in the strict sense refers in general to a specific group of hydrated mafic minerals with the general formula $Mg_6Si_4O_{10}(OH)$ and includes antigorite, lizardite, and chrysotile. These minerals are a common metamorphic product of olivine, which can be produced by hydrothermal alteration of ultramafic and mafic rocks (Klein & Dutrow, 2008; Best, 2003). The metamorphic derivative of ultramafic rocks composed primarily (>90%) of olivine, such as dunite, is accordingly termed serpentinite when the parent rock is completely altered to serpentine minerals. The terms ‘serpentine’ and ‘serpentinite’ are, therefore, commonly indicative of metamorphically altered environments, and are also frequently applied by biologists to roughly describe igneous ultramafic rocks and their derived soil products in general, even if they do not contain minerals of the serpentine group (McCallum 2006). Thus, it has been recognized that ‘serpentine soil’ is, in fact, a misnomer (Brady *et al.* 2005), and should rather be referred to as ultramafic soil, since the term relates in more general context to soils produced largely from weathered bodies of ferromagnesian silicates.

2.3.2 Ultramafic soil characteristics

2.3.2.1 Ca:Mg ratio

Ultramafic soils are characterized by their low to very low calcium (Ca):Mg ratios (Brady *et al.*, 2005; O’Dell *et al.*, 2006), which is generally <1 (Mengel & Kirkby, 2001). This is accredited to the high Mg content in ultramafic rocks, which is often dominant over Ca because of (1) the presence of olivine or serpentine (a major source of Mg), and (2) lower concentrations of plagioclase feldspars (Ca-sodium (Na)-rich aluminosilicate minerals), compared to mafic and intermediate igneous rocks (containing relatively higher amounts of Ca and Si). This reduces the availability of Ca in the weathered product (McGrath & Smith, 1990), and in some cases results in soil Ca deficiency. Conversely, a large increase in the soil Mg concentration can also result in Mg toxicity (Brady *et al.*, 2005). Generally, soils with Ca:Mg ratios between 4 and 1 will typically illustrate a decrease in Ca availability, and become Ca deficient at ratios lower than 1 (Hazelton & Murphy, 2007).

2.3.2.2 Trace elements (Cr, Co & Ni)

Soils originating from ultramafic rocks often have elevated levels of phytotoxic trace elements, particularly Chromium (Cr), Cobalt (Co) and Nickel (Ni) (McGrath & Smith, 1990; Robinson *et al.*, 1997; Brady *et al.*, 2005; Mengel & Kirkby, 2001; O'Dell *et al.*, 2006; Kabata-Pendias, 2011). These siderophilic trace metals have a strong affinity for mafic minerals since they are commonly substituted for Fe, Mg or Manganese (Mn) in olivine, serpentine and pyroxene mineral structures (Alloway, 1990; Best, 2003; Tashakor *et al.*, 2011), and can become potentially toxic in the weathering product (in some cases, concentrated along with hydrated Fe and Mn oxides - McGrath & Smith, 1990).

Apart from silicate minerals, spinels, such as magnetite and chromite are also regarded as a primary source of Cr in ultramafic soils (Oze *et al.*, 2004). However, due to the discrete occurrences of chromite bodies, extremely high concentrations of Cr are commonly encountered along confined areas of ultramafic outcrops. Cheng *et al.* (2011) illustrated that although high localized concentrations of Cr are associated with bodies of chromite in some areas, in terms of total mass, mafic silicates are generally more prolific than Cr-spinel minerals.

A world average of soil background levels for Cr, Co and Ni is indicated as 59.5 mg/kg, 11.3 mg/kg and 29 mg/kg respectively (Kabata-Pendias, 2011). Concentration levels of 50 to 100 mg/kg for Cr and Ni and 50 mg/kg for Co are generally considered as phytotoxic for most species with regards to agricultural soils (Kabata-Pendias & Pendias, 1992). Ultramafic rocks and derived surface soils, however, can contain up to 40 times more Cr and Ni, and about 10 times more Co (Mengel & Kirkby, 2001). In this regard, it becomes clear why these elements form one the major factors influencing plant-soil interactions on ultramafic bodies.

Ni is particularly regarded as the main limiting trace element in ultramafic soils, having a higher potential for phytotoxicity compared to high concentrations of Cr and Co (McGrath & Smith, 1990). This is because Ni is more soluble and mobile in the form of Ni^{2+} , whereas Cr in soil tends to occur as Cr^{3+} forming insoluble precipitates ($\text{pH} > 5.5$). The mobility and distribution of Ni are primarily controlled by the soil pH, however, factors such as the clay content and concentration of Fe and Mn oxides are of secondary importance. It is indicated by McGrath and Smith (1990) that mobility of Ni increases with decreasing pH and cation exchange capacity (CEC). Therefore, plants growing in acidic soils with low concentrations of clay and organic material will be likely to have a higher availability of Ni.

Similarly to Cr, Co derived from ultramafic rocks is largely immobile in alkaline soils, and occur as insoluble oxides, hydroxides, and carbonates, which are often associated with Mn oxides. In acid conditions, however, Co is more susceptible to dissolution and likely to be removed by leaching.

Therefore, concentrations of total Co are usually higher in alkaline soils, but less available compared to acid soils.

There is a strong association between concentrations of Co and Ni in soil and Mn and Fe oxides (Alloway, 1990), which is related to their ability to isomorphically substitute one another in the crystal lattice of their oxide minerals (McGrath & Smith, 1990). This results in co-precipitation of these elements in soils, which can inversely affect the individual availability of each element and its related toxicity.

2.3.2.3 Macronutrient deficiency

Ultramafic soils are deficient in essential macronutrients specifically nitrogen (N), phosphorus (P) and potassium (K) (O'Dell *et al.*, 2006; Rajakaruna & Boyd, 2009). The lack of N is primarily attributed to low concentrations of soil organic material (SOM) in the weathered product, as well as high concentrations of Ni (100-1000 mg/kg) being known to inhibit nitrification and N-mineralisation (Kabata-Pendias & Pendias, 1992). The lack of K and P is mainly ascribed to the absence of specific inorganic minerals in ultramafic parent materials. K-feldspar, leucite, nepheline, muscovite and some secondary clay minerals are major sources of K, whereas P is largely derived from apatite phosphates (Gregory & Nortcliff, 2013). These minerals are commonly associated with felsic and dioritic igneous rock types and are absent in ultramafic rocks. Therefore, the weathered product is generally lacking in these essential macro nutrients.

Apart from a general deficiency due to the absence of sufficient organic and inorganic nutrient sources, the specific availability of these elements is correspondingly low, as a result of the soil's CEC, which is negatively affected owing to the low concentrations of SOM and clay minerals.

2.3.2.4 Structure and water retention capacity

In addition to chemical attributes, physical conditions also prove uncongenial to most vegetation. Ultramafic outcrops are highly susceptible to weathering and are often steep and rock-strewn. This results in shallow litocutanic soils with a minimal silt and clay content (Brady *et al.*, 2005), as most of the smaller, clay-sized particles are usually quickly transported and localised within lower regions of the landscape. In these conditions, ultramafic soils will generally illustrate weak soil structure, as well as lower water retention capacity. This is largely due to the absence of clay minerals (Proctor, 2003), but also in part due to the coarse-textured nature of the soil, which characteristically tends to retain less SOM (Spargo *et al.*, 2013). Hosted serpentine vegetation is therefore commonly subjected to arid soil conditions, which requires the associated species to be well adapted to the lack of soil water.

2.3.3 Ultramafic vegetation traits

The influence of ultramafic soils, on the physiology and phytogeographical distribution of vegetation, is a universally well-known phenomenon (Kruckeberg, 1984; Robinson *et al.*, 1997; Reddy *et al.*, 2001; Smith *et al.*, 2001; Proctor, 2003; Prasad, 2004; Brady *et al.*, 2005; O'Dell *et al.*, 2006; Rajakaruna & Boyd, 2009; Kabata-Pendias, 2011; Anacker, 2014). Referred to as systems with “serpentine syndrome” these biological and geochemically unique habitats are quite inhospitable to vegetation, due to a complex interaction of chemical, physical and biological factors. Ultramafic soils vary considerably depending on the parent mineralogy, However, three mutual traits relating to its supported vegetation have been documented (Brady *et al.*, 2005; Reddy *et al.*, 2009). This includes poor vegetation productivity and decline in species diversity, high rates of endemism, and distinct floristic differences between ultramafic zones and adjacent non-ultramafic rock types. Some ultramafic systems are also known to have remarkably high biodiversity and are classified as biodiversity hotspots (Harrison and Inouye, 2002).

Typical ultramafic plant species often display stress tolerance features, such as xenomorphic foliage, a decrease in height and density, and often more developed root systems (O'Dell *et al.*, 2006). Several edaphic factors of ultramafic soils (which are ultimately related to the specific geochemistry of ultramafic rock types) have been identified to affect speciation, and limit plant productivity (Kruckeberg, 1984; Mengel & Kirkby, 2001; O'Dell *et al.*, 2006). These include the following:

2.3.3.1 Ca deficiency / Mg toxicity

It is proposed that the lack of sufficient Ca concentrations is one of the principal plant stressors on ultramafic soils, along with increased Mg which further suppresses Ca availability. Several studies summarised by Brady *et al.* (2005) indicated that native serpentine species are more tolerant to low Ca concentrations, and are able to absorb satisfactory amounts of Ca without absorbing excessive Mg. Non serpentine species, introduced to the same environments, are more commonly affected by low Ca availability and show a decrease in growth and biomass yields. It was also noted by Brady *et al.* (2005) that the non-serpentine species tend to show improvement with the addition of Ca to the soil, but did not react noticeably to fertilizer (N, P & K) amendments. Corresponding observations made by O'Dell *et al.* (2006) also indicate that serpentine species maintain higher foliar Ca:Mg ratios than non-serpentine species. Ca deficiency symptoms in plants are characterised by the reduced growth of meristematic tissues, and is initially observed in the youngest leaves and apical meristems. The affected tissues become deformed and show signs of chlorosis and necrosis (Mengel & Kirkby, 2001).

In addition to Ca deficiency, a large increase in available soil Mg can also result in Mg toxicity. Proctor (2003) conveys that several serpentine species have shown limited Mg absorption, while others illustrate tolerance for high tissue concentrations of Mg. Symptoms of Mg toxicity in plants nevertheless, are unknown, and not generally discussed in the literature. It has been established, however, that imbalances between K, Ca and Mg can result in reduced growth when soil Mg concentrations are exceptionally high (Katalin, 2011).

2.3.3.2 Phytotoxic trace elements

Although toxicity and accumulation values for specific trace elements vary broadly between different plant species (depending on a variety of factors), typical native serpentine species have developed different mechanisms to survive these conditions. Some species are able to tolerate high Mg and heavy trace-metal concentrations, and can, in some cases, hyperaccumulate these elements (particularly Ni) (Kabata-Pendias & Pendias, 1992; McGrath *et al.*, 2001; Wenzel *et al.*, 2003; Prasad, 2004; Brady *et al.*, 2005; Ghaderian *et al.*, 2007; Kabata-Pendias, 2011), while other species endure such conditions by excluding these elements (Orcutt & Nilsen 2000; Mengel & Kirkby, 2001; Reddy *et al.*, 2001; Williams Araújo do Nascimento & Xing, 2006).

While the majority of serpentine-tolerant species do not necessarily fall in this category, Kabata-Pendias (2011) indicates there are an estimated 400 species of known hyperaccumulators. Some of which can contain foliar concentrations of Co and Ni above 1000 mg/kg. Several Ni hyperaccumulator species which are predominantly of the Boraginaceae, Cruciferae, Caryophyllaceae, Myrtaceae and Leguminosae families, have also been identified to accumulate Co (Kabata-Pendias & Pendias, 1992). Some grass species were reported to contain up to 96 mg/kg (dry weight) Co.

Typical toxic plant tissue levels for the following elements are indicated as ~10-100 mg/kg for Ni, ~5-30 mg/kg for Cr, and ~15-50 mg/kg for Co (Orcutt & Nilsen, 2000; Kabata-Pendias, 2011). Toxicity symptoms for these elements are primarily observed as wilting tops with brownish red leaves, root injury (particularly for Cr toxicity), as well as chlorosis of the young leaves and new growth tips (commonly associated with Ni, Co & Cr toxicity, Kabata-Pendias, 2011).

2.3.3.3 Nutrient deficiency

General deficiency of primary macronutrients such as N, P, and K can cause a variety of symptoms in plants and is discussed to some extent by Mengel and Kirkby (2001).

Nitrogen deficiency can cause poor growth leading to thin stems, small leaves (which fall prematurely) and restricted root branching. Furthermore, disturbance in chloroplast development and related collapse of chloroplasts causes chlorosis which is generally evenly distributed.

Necrosis normally occurs in cases of severe deficiency, which is different from K deficiency that appears earlier.

Phosphorus deficiency is characterized by stunted growth, lower fruit yields, and a rigid erect appearance. Young leaves appear blueish-green, while older leaves are often dark-green. An early symptom of Phosphorus deficiency is constrained leaf surface areas and hindered leaf expansion, as a result from decreased water delivery. Root and stem growth, however, is less inhibited.

Potassium deficiency is not normally indicated by early symptoms and usually first becomes visible by a reduction in growth rate and necrosis, and later chlorosis of the older leaves. Plants also usually show a decrease in turgor and resilience to drought conditions.

2.3.4 Conclusion

Ultramafic intrusions and their related soils have been intercontinentally established to produce unique ecosystems, in which plant species have adopted a variety of tolerance and avoidance mechanisms to survive. Chemical and physical characteristics of ultramafic soil environments require hosted species to tolerate lack of soil water and specific macronutrients (N, P, and K), high soil Ca:Mg ratios and high levels of phytotoxic trace elements, particularly Cr, Co, and Ni. Studies indicate that typical serpentine flora is able to maintain higher concentrations of Ca (despite depression by Mg in soils), are less dependent on fertilizer amendments, and are known to exclude or hyperaccumulate relevant toxic trace elements.

If the primary characteristics of “the serpentine effect” are evaluated, it appears that some of the relevant mafic-ultramafic intrusions of the Vredefort dome may encompass the geochemical features required to produce localised occurrences of ultramafic soil traits. In this regard, it can most likely be hypothesised that the hosted vegetation of the dome will show similarly established relationships with several of these intrusions. Although, to date, there is a general lack of supporting literature regarding this smaller-scale geo-ecological factor.

2.4 SENEGALIA AND VACHELLIA SPP.

2.4.1 Introduction

With regard to *Senegalia* and *Vachellia* literature references, it should be noted that these specific genera were recently phylogenetically segregated from *Acacia* Miller s.l., formalizing a new separate classification specifically for African species (Kyalangalilwa *et al.*, 2013). Several morphological characteristics are now distinguished between the two genera, of which the presence of stipular spines on *Vachellia* sp. and prickles on *Senegalia* sp. is particularly specified.

In general, species of *Senegalia* and *Vachellia* occurring on the KC are well adapted to the dome's environment. This is because of several attributes which make them robust pioneer species in the Southern African region (Hayward, 2004; Nyakundya *et al.*, 2011; Spicer *et al.*, 2005). They are able to tolerate various climate extremes such as hot arid environments, variable annual rainfall (from 150 to 1200 mm per annum) (Hamad and Kamal, 2006), cold winters ($>-12^{\circ}\text{C}$) and periodic flooding (Hayward, 2004), and can occur across a variety of soil types including highly alkaline and saline soils (Nyakundya *et al.*, 2011).

Specific differences in species occurrences are observed amongst varying rock substrates of the Vredefort dome. As a fundamental aspect of the geobotanical study, it is hypothesised that amongst the woody vegetation structure of the KC, of which *S. caffra*, *V. robusta*, and *V. karroo* is the most prevalent (Boneschans, 2012), *V. karroo* tends to outcompete the other species on the mafic-ultramafic intrusions, due to better tolerance mechanisms for ultramafic soil characteristics. In support of this view, several characteristics could be identified from previous studies which could account for the tenacity that *V. karroo* displays in these particular environments.

2.4.2 Rock, soil and nutrient association

According to Nair (1993), as referenced by Nyakundya *et al.* (2011), species of *Vachellia* have extensive root systems that can access deep seated water and mineral sources, which would otherwise not be available to shallowly rooted species. The nutrients are in turn recycled in the upper soil horizons via leaf fall and litter accumulation (Nyakundya *et al.*, 2011). This, in turn, improves soil properties such as porosity, moisture retention, structure, and erosion resistance.

Vachellia species are members of the legume plant family and are able to fix nitrogen from the air with the aid of symbiotic rhizobia that produce nodules on the roots (Hayward, 2004). This, as a result, stimulates soil nitrogen enrichment (Spicer *et al.*, 2005).

Mshumi (2006) confirmed the biochemical prospecting characteristics of *V. karroo* in the Blue Dot Mine area, situated within the Amalia Greenstone Belt in the Northwest Province, South Africa. Results indicated that individuals of *V. karroo* were able to concentrate higher anomalous values of gold and several indicator REEs, compared to *Boscia albitrunca*. In addition, Mshumi (2006) noted from analysed leaf samples that the Ca- (~2.3% avg. - dry wt.) and Mg content (~0.4% avg. - dry wt.) of *V. karroo* accounted for almost ~41% of the total macronutrient variance between individuals and is lithologically controlled, indicating similar, increased distribution patterns in individuals situated on mafic-ultramafic- tholeiitic greenstones. As a result, a strong association was seen between the Ca content of individuals particularly situated in mineralized zones of the study area.

In regards to more recent studies, Nyakundya *et al.* (2011) illustrated that *S. polyacantha*, *V. gerrardii*, and *V. karroo*, show a strong early survival performance on ultramafic-greenstone related Ni-tailings in Zimbabwe. Results indicated that *V. karroo* outperformed both *V. gerrardii* and *S. polyacantha* with regards to growth in height and root collar diameter. Corresponding to similar findings of Ibrahim *et al.* (2001), it was determined that *V. karroo* possesses a robustness due to an inherent characteristic to make opportunistic growth.

2.4.3 Conclusion

The observations made by Mshumi (2006) may strongly suggest that *V. karroo* tends to maintain or increase Ca uptake when situated on mafic-ultramafic rocks. However, from the lack of point specific soil data, this could not yet be concluded with much certainty. The findings of Nyakundya *et al.* (2011) may also support the proposed ability of *V. karroo* to maintain a higher Ca content, seeing that this nutrient is particularly associated with growth performance and biomass production, but due to the lack of plant tissue analysis, a direct correlation could also not be drawn from published results, hence analysis of heavy metal nutrient uptake was recommended. It is in this respect that the floristic, as well as the biochemical evaluation of the geobotanical features of KC, may reveal the specific nature of this proposed association.

CHAPTER 3

METHODS AND MATERIALS

3.1 INTRODUCTION

This study focuses on two major aspects regarding the KC. Specific methods were applied to re-map and describe the geological features (structure, mineralogy, and geochemistry) in detail, and secondly to map and describe the interaction between the chemical characteristics of the soil profile and the floristic and physiognomic responses of the KC's supported woody vegetation (particularly of *Senegalia* and *Vachellia* species). Therefore, related sampling and statistical analyses were performed by combining three primary environmental features: geology, soil, and vegetation.

3.2 ALLOCATION OF HOMOGENEOUS GEOBOTANICAL UNITS

In order to re-map the geology of the KC and simultaneously stratify the study area for soil and vegetation sampling, a geological survey was performed, utilising the geological map (scale 1:50000) of the Council for Geoscience (Bisschoff, 1999), and the detailed map (scale 1:10000) of Bisschoff (1969). Strike and dip measurements were taken where possible, along with GPS coordinates of the major outcrop positions of each intrusive and metamorphic rock type, and digitally mapped using a digital orthophoto of the Koedoesfontein Complex (Appendixes I1 & I2)

Soil and vegetation sampling were conducted utilizing a geobotanical approach (Heshmatti & Squires, 1995). To identify different sampling units within the study area according to this approach, and to avoid sampling in ecotones (both ecological and pedo-chemical), the focus was placed on combining the geomorphology of the landscape with the floristic and physiognomic characteristics of the supported vegetation. Therefore, data from the geological survey, along with woody vegetation composition and density (obtained from quantitative field observations), was used to stratify the area into six homogeneous geobotanical units (Figure 6). Each unit was selected according to the absence and presence of specific groups of intrusive (ranging from ultramafic to felsic) and metamorphic rock types, as well as key woody species present. Figure 7 visually illustrates this division as seen from the south-eastern aspect of the KC.

By rather dividing each unit subjectively according to rock type and their related key woody species, as opposed to random stratification, a corresponding geochemical and ecological gradient is more clearly established for sampling. As a result, each proposed unit represents an area of a specific homogeneous rock, soil, and woody species. By sampling within the allocated homogeneous units, changes in the mineralogical and chemical composition of the bedrock

geology can be correlated with changes in soil chemistry and vegetation characteristics. Ultimately by correlating these variations, the effect of each rock type on specific soil and vegetation features can be determined.

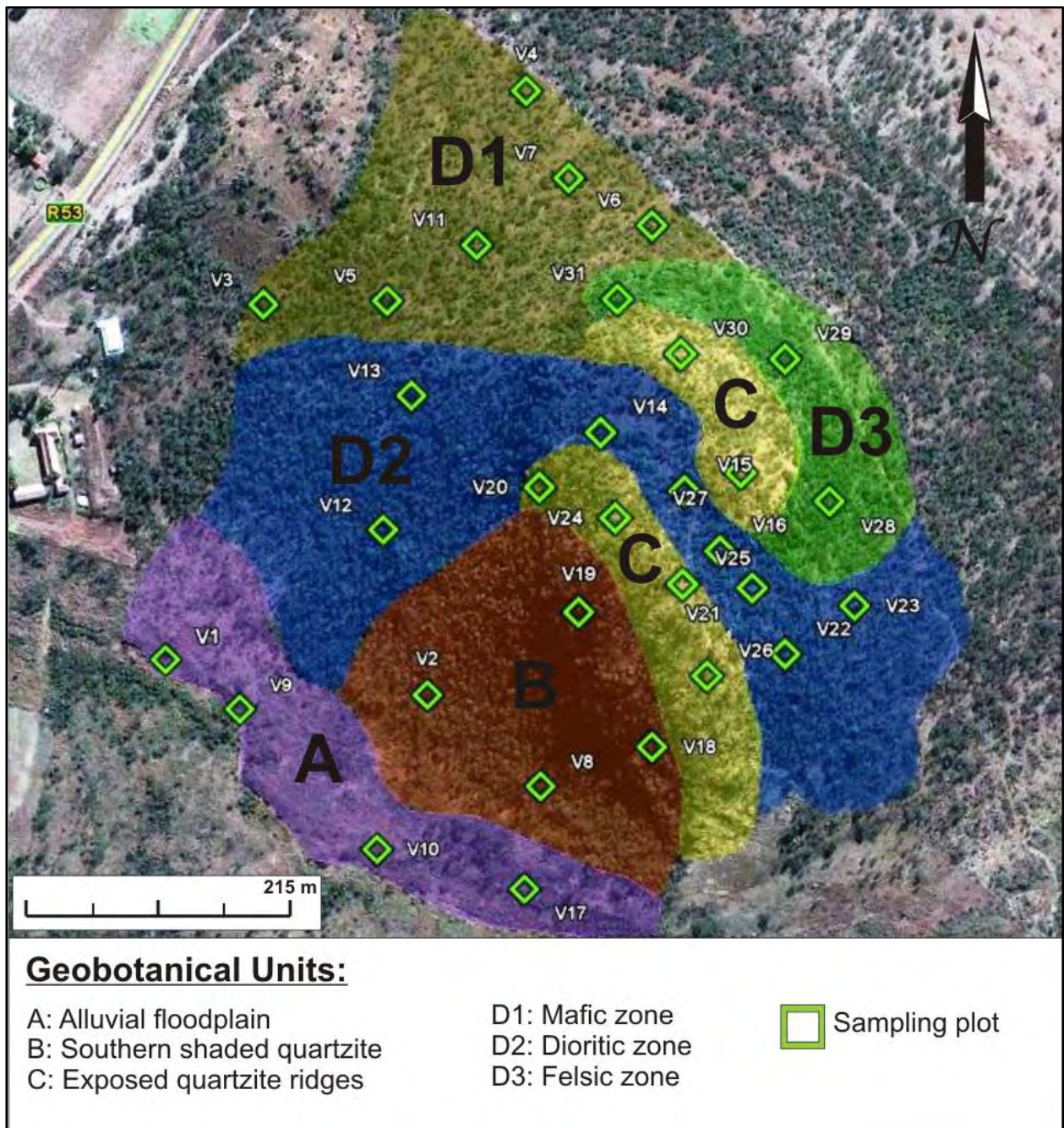


Figure 6: Aerial photograph of the Koedoesfontein Complex, illustrating the division of the study area into homogeneous geobotanical units, and allocation of sampling quadrants for soil and botanical surveys. The units are named accordingly after the predominant bedrock geology or overburden.

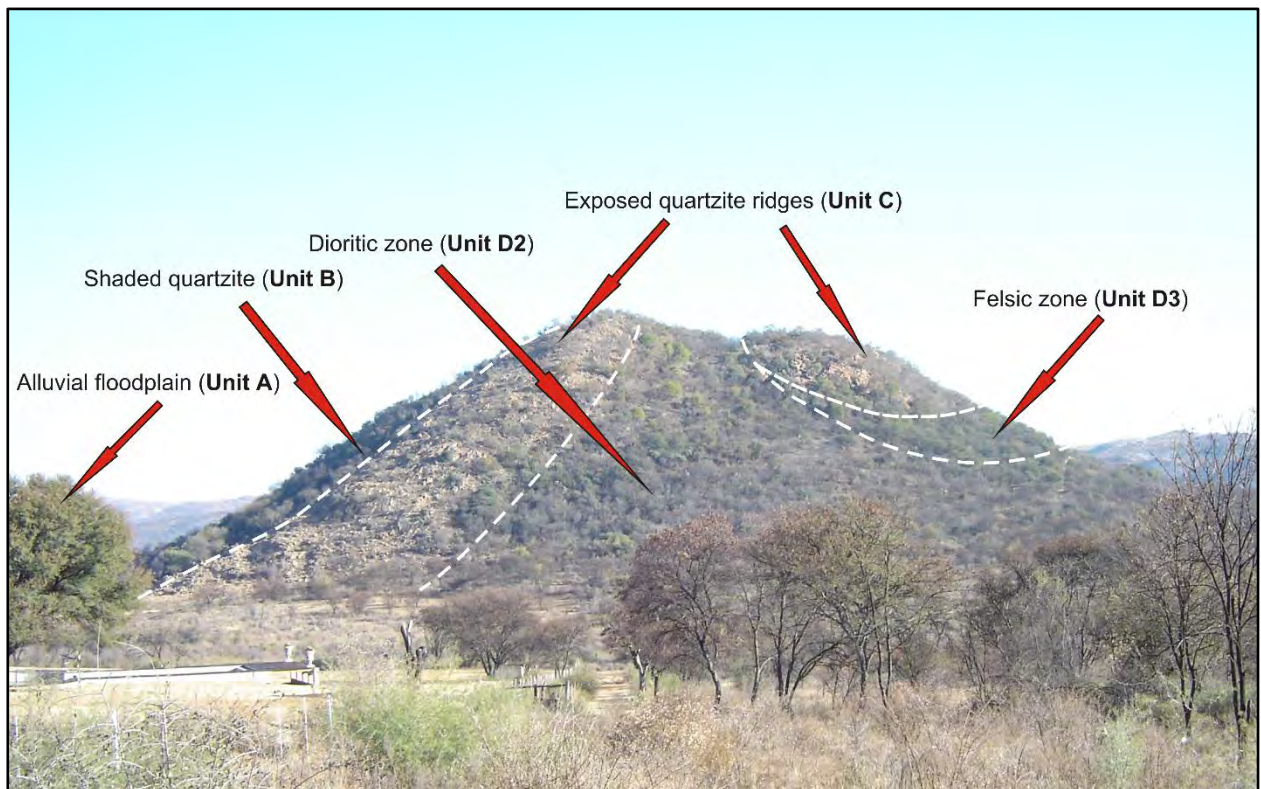


Figure 7: Photograph of the south-eastern part of the study area, facing the strike of the overturned epidiorite sill. Visible changes in woody vegetation density and composition can be observed between the indicated geobotanical units.

3.2.1 Stratification and characterisation

The KC was stratified into the following geobotanical units (Figure 6 and Figure 7):

3.2.1.1 The alluvial floodplain (Unit A):

This area is largely covered with alluvial soils which consist of a deeper weathered Hutton soil form, occurring primarily upon quartzite bedrock. Woody vegetation is mainly sparse and dominated by, *Pyracantha angustifolia*, *P. crenulata*, *V. karroo*, and *Ziziphus. mucronata*.

3.2.1.2 The southern shaded quartzite (Unit B):

This area is similarly dominated by quartzite but contains small scattered outcrops of dark, porphyroblastic hornfels (metamorphosed shale). The related soils are shallower and consist of a Mispah soil form. The soil is also more organic and hosts very dense woody vegetation, comprising mainly *Celtis africana*, *Diospyros lycioides*, *Dombeya rotundifolia*, *Grewia occidentalis* and *S. caffra*.

3.2.1.3 The exposed quartzite ridges (Unit C):

These areas have a very shallow Glenrosa soil from, comprising fine quartzite gravel and minor amounts of organic material. Woody vegetation density is very low and is dominated by *Combretum molle*, *Nuxia congesta*, *S. caffra* and *Vangueria parvifolia*.

3.2.1.4 The mafic zone (Unit D1):

Termed collectively as “The mafic zone”, this area was selected as the most probable zone for ultramafic/serpentine soil characteristics to occur, as it encompasses the main zone where ultramafic, mafic and intermediate intrusions are encountered. Particular rock varieties include wehrlite which grades into an $\pm$ olivine clinopyroxenite, hornblendite, gabbro and spessartite. The soil is shallow and consists of a Mispah soil form, with low woody density and even sparseness of grassy vegetation, which seems patchily distributed along the outcrops of wehrlite and Ol clinopyroxenite (Figure 3). The woody vegetation is almost entirely dominated by *Ehretia rigida*, *V. karroo*, *V. robusta* and *Z. mucronata*.

3.2.1.5 The dioritic zone (Unit D2):

This area follows the outcrop pattern of the epidiorite intrusion, which is primarily mafic to dioritic in composition. The soil is similarly shallow and rocky, but nonetheless hosts higher densities of woody vegetation compared to unit D1, particularly *D. rotundifolia*, *S. caffra*, *V. robusta* and *Z. mucronata*.

3.2.1.6 The Felsic zone (Unit D3):

This small area contains very thin outcrops of highly weathered uralitite, however, is dominated by larger bodies of felsic igneous rocks. These bodies consist particularly a Na-rich quartz alkali syenite, which seems to grade into a Na-K-rich alkali granite. The soil cover is slightly thicker and contains high amounts of organic material. Woody vegetation density is particularly high and follows the outcrop pattern of the overturned felsic sills. The area is dominated by *E. rigida*, *G. flava*, *Pappea capensis* and *V. robusta*.

3.3 SAMPLE ACQUISITION

3.3.1 Geological sampling

After compiling a preliminary geological map of the study area, samples from the various intrusive and metamorphic rock types were collected for petrographic investigation, as well as mineralogical and chemical analyses (Appendix A2). Care was taken as far as possible to acquire

fresh unweathered samples, which was not always possible. Quartzites and hornfelses were excluded from sampling since this study focuses mainly on describing the intrusive and related metamorphic rock types. Furthermore, any major variation in soil element concentration would mainly be attributed to the intrusive rock types, as they are the primary sources of mafic minerals as well as feldspars, and have restricted localities within the abundant quartzite country rock. A total of 25 samples from eight different rock types was collected for textural description, as well as chemical and mineralogical analyses. Sample numbers and corresponding coordinates are provided in Appendix A2.

3.3.2 Soil sampling

Soil sampling consisted of two sample groups termed “KSA” and “KSB”. The first group of samples (KSA) was taken simultaneously with the vegetation survey in allocated sampling plots of 20 × 20 m (Figure 8), in order to gain an overall view of the soil chemical attributes of each geobotanical unit. One soil composite sample for every plot was compiled by taking a sample, at the four corner points of each sampling plot, with a 1 m barrel auger. These four samples were then mixed thoroughly. Sampling depth was roughly 30 cm due to the shallow Mispah and Glenrosa soil forms. A total of 31 KSA samples was obtained (see Appendix B1).

The second group of samples (KSB) was taken simultaneously with the sampling of plant material for trace element analysis, to determine the concentration and availability of specific trace metals in the soil. A total of four composite soil samples from the KC were compiled by mixing ten samples per composite. Sampling areas consisted of a 20 m radius, surrounding major outcrop positions of the different ultramafic to felsic intrusions at which plant material was collected (Appendix B2). A total of 35 composite samples (31 KSA samples and four KSB samples) were taken for preparation at the Soil Analytical Laboratory of the North-West University (NWU), Potchefstroom.

3.3.3 Vegetation sampling

3.3.3.1 Species abundance and physiognomic data

To gather data on woody species abundances for the different geobotanical units, a stratified vegetation survey was performed with a combination of the quadrant and transect method, as described by Kent and Coker (1992). Three to eight sampling plots of 20 m x 20 m were allocated to each geobotanical unit depending on its size (Figure 6). Each sampling plot (Figure 8) contained five transects, spaced 1 m from the sampling border, and 4.5 m from each other.

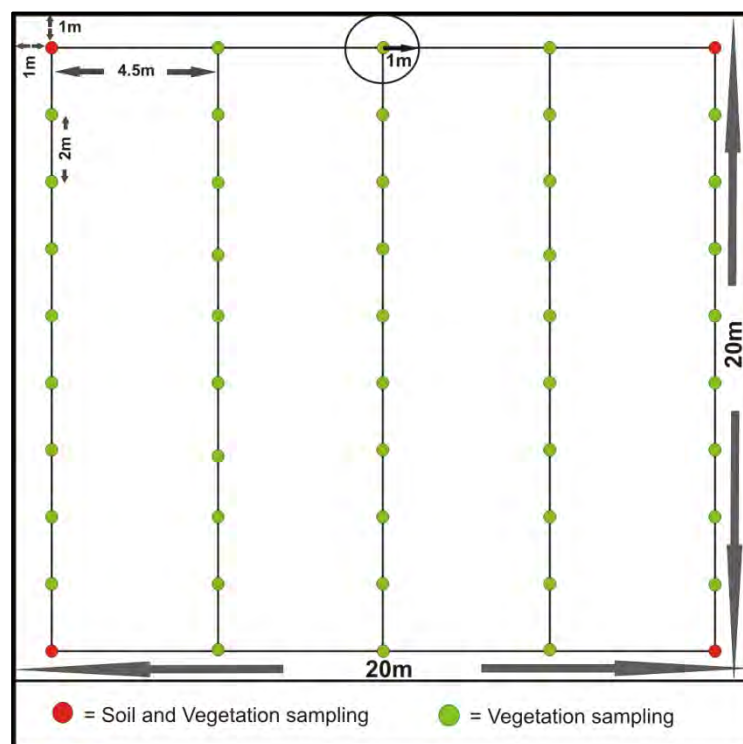


Figure 8: Sampling plot design to obtain species abundance and physiognomic data, as well as the first group of 31 KSA soil samples.

Each transect contained ten sampling points spaced 2 m apart, resulting in 50 sampling points per plot. At each sampling point, the nearest adult tree and woody shrub species were identified and recorded within a 1 m radius. To determine what effect variations in soil chemistry have on growth characteristics of particular tree species, stem circumferences were measured for all the encountered trees. Measurements were taken 1 m above the base of each individual and recorded in centimeters. The locality of each plot was determined with a GPS. A total of 31 plots were sampled, and all the relevant data recorded (Appendices C1 and C2).

3.3.3.2 Plant material for chemical analysis

Leaf and stem composite samples were taken from four selected woody plant species on the various intrusive rock types to determine their ability to exclude or hyperaccumulate phytotoxic trace elements, as well as to determine what effect the geology has on the concentration of specific major and trace elements in the plant tissue. Dominant woody species were selected which illustrated the best correlation with the distribution of the geobotanical units containing intrusive rock types. These species include *G. flava*, *S. caffra*, *V. karroo* and *V. robusta*.

Composite samples were made by harvesting leaves and stems of about 20 individuals within a 20 m radius around four different groups of ultramafic to felsic intrusions of the KC. A total of 24

samples (12 leaf samples and 12 stem samples) were obtained (see Appendix C3) and taken for preparation at the Soil Analytical Laboratory of the Potchefstroom Campus (NWU).

3.4 SAMPLE PREPARATION

3.4.1 Rock samples

The rock samples were cleaned to remove any weathered material and air-dried before being cut into appropriate sizes for the production of thin sections and crushing respectively. Thin sections were made from 23 samples for petrographic investigation.

To determine the chemical and mineralogical composition of each rock type, 20 samples were prepared for X-ray fluorescence spectrometry (XRF) and 19 samples for X-ray diffraction (XRD) analysis (Appendix A2). Preparation for these analyses required the respective samples to be crushed into fine gravel size using a jaw crusher and transferred to a tungsten carbide ring mill, which was used to grind the samples into a fine powder ($\leq 75 \mu\text{m}$).

3.4.2 Soil samples

The 36 composite soil samples were air dried to remove soil moisture. Thereafter samples were pulverized and put through a 2 mm sieve to break down aggregates and remove any gravel or organic material on the surface. The first group of 31 (KSA) samples was reduced for analyses. This was done by compositing samples using a riffler. New composite samples were made from samples that showed similar characteristics, and were located within the same homogenous unit (Figure 4). A total of 17 new soil composite samples termed “KKSA” were prepared for XRF and wet chemical analyses (see Appendix B1), as well as four KSB samples for trace metal content analysis (see Appendix B2). Soil samples for the XRF analysis required further crushing and were ground into a fine powder ($\leq 75 \mu\text{m}$) using a tungsten carbide ring mill.

3.4.3 Plant tissue samples

Leaf and stem samples were washed with deionized water (dH_2O) to remove any soil and major inorganic material. Thereafter the samples were washed with a 1 molar hydrogen chloride (HCl) solution and soaked in deionized water for 20 seconds to remove any excess HCl. This process was repeated three times with clean deionized water. After cleaning, the samples were packed in individual brown paper bags and transferred to a drying oven. The samples were dried at 50°C for two days. The dried samples were then transferred to a tungsten carbide ring mill, which was used to powder the samples. A total 26 powdered samples were prepared for soluble and total trace metal analysis (Appendix C3).

3.5 LABORATORY ANALYSIS

3.5.1 XRD analysis

XRD analysis of the rock samples was performed at XRD Analytical & Consulting cc, Lynnwood Glen, Pretoria 0081. The analysis was performed according to the method described by Jenkins and Snyder (1996), using a PANalytical X'Pert Pro instrument. The samples were prepared using a back loading technique and were loaded onto a spinner stage inside the XRD. The samples were then irradiated using a Co X-Ray tube, and the different mineral structures were detected using a X'Celerator detector. A unique diffractogram for each sample was generated which shows the different mineral phases (peak positions) and their concentrations (peak heights) present in each sample (Appendix D1). The minerals were then identified using X'Pert Highscore plus version 4 software, and a database from the International Centre for Diffraction Data (ICDD) (PDF 4+) program. The Rietveld quantification technique (Kniess *et al.*, 2012) was applied to determine the weight percentage of each identified mineral in the sample (Appendix D2).

3.5.2 XRF analysis

XRF analysis of the soil, as well as the rock samples, were performed at the XRD and XRF Laboratory, NWU, Potchefstroom. The analysis was performed according to the method described by Markowicz and Grieken (1992). For the analysis, a PANalytical Axios XRF instrument was used. The samples were poured into aluminium sample cups and pressed using a 25 ton Lenton powder press. The samples were then irradiated using a Rh X-ray tube, and the elemental concentrations of each sample were determined (see Appendix E1 & E2) using a Super Q database, which was set up with a predetermined calibrated range of standards. The Wide Range Oxides (WROXI) application was used to determine major element oxide concentrations and the Pro-Trace application for trace elements. The volatile content of the rock samples was determined by loss on ignition (LOI).

3.5.3 Soil chemical analyses

The first group of 17 KSA samples was analysed for pH, cation exchange capacity (CEC), exchangeable cations, available phosphorus (P-Bray1), macro- and micro-nutrients and soil organic carbon (SOC) (see Appendix F1). The second group of five KSB samples obtained for trace element analysis were analysed for total and soluble trace metal content (see Appendix F2). All analyses were performed at Eco Analytica Laboratory, NWU, Potchefstroom, according to the method and procedures prescribed by the Council of Soil Testing and Plant Analysis (1997), the International Organization for Standardization (ISO19730, 2008) and the United States Environmental Protection Agency (EPA 3050B, 1996).

3.5.3.1 pH (H₂O)

The pH values of the samples were determined using a 1:2.5 extract solution. This method determines the pH of the soil in a 1:2.5 soil/water ratio suspension based on mass. By definition, pH is the negative logarithm of H⁺ ions. These ions occur within the soil solution as well as cation exchange sites on soil colloids. In the presence of dissolved cations with a higher affinity for adsorption, H⁺ ions are replaced in the exchange sites and transferred to the soil solution. The pH of the soil solution will, therefore, be related to the amount of H⁺ ions released from the exchange sites (Coleman & Thomas, 1967; Mclean, 1982).

Twenty grams of soil (<2 mm particle size) was weighed into a 100 ml plastic beaker to which 50 ml of deionized water (dH₂O) was added. The suspension was stirred for five seconds using a glass rod and left for at least four hours. Before the pH was determined the suspension was stirred again and left for ten minutes. The pH of the overlying fluid was then determined with a pH meter, by allowing the pH electrode to stabilize for three minutes in solution. The pH was then recorded.

3.5.3.2 Cation exchange capacity

Soil colloidal particles such as clay usually hold a negative charge, which is neutralized by the adsorption of cations. The ability of a soil to absorb cations is referred to as the Cation Exchange Capacity (CEC). The CEC of the samples was determined using the sodium acetate method. This method uses the concentration of adsorbed sodium ions to determine the exchange capacity of a soil (Council of Soil Testing and Plant Analysis, 1997).

Five grams of soil (<2mm particle size) was weighed in a Schott-bottle to which 50 ml of sodium acetate (C₂H₃NaO₂) was added and shaken for 15 minutes, at 180 rpm. A leaching tube was prepared with Filter Floc which was gently compacted with a glass rod inside the tube to flocculate residual solids. The suspension was then quantitatively transferred to the leach tube by spraying out the soil from the Schott-bottle with sodium acetate. The excess sodium was removed by leaching 50 ml ethanol/dH₂O (1:1) through the tube. This process was repeated five times with 50 ml concentrated ethanol, and the excess liquid removed with compressed air. To remove the adsorbed sodium from the exchange sites of the sample, the leach tube was washed three times with 30 ml ammonium acetate (C₂H₃O₂NH₄), which was collected in a 100 ml volumetric flask at the bottom of the tube. The excess fluid was removed with compressed air. The 100 ml flask was filled to volume with ammonium acetate and thoroughly mixed.

A tenfold dilution was made by transferring five ml from the 100 ml flask to a 50 ml volumetric flask, and filling it to volume with ammonium acetate. The sodium concentration of the diluted

sample was then determined with an Atomic Absorption Spectrometer (AAS), and these concentrations were used to calculate the CEC.

3.5.3.3 Exchangeable cations

Exchangeable cation concentrations from the samples were determined by the ammonium acetate method. This method utilizes the ammonium ion to displace different cations from the exchange sites of soil colloids. The concentration of extracted ions (Ca, Mg, Na, and K) was then determined with AAS in cmol(+)/kg.

The rate of extraction is a function of temperature, therefore, the extraction process was carried out at 20°C. Cations were extracted at a 1:10 g/ml ratio of soil: extraction solution.

Five grams of soil (<2mm particle size) was weighed in a Schott-bottle to which 50 ml ammonium acetate (20°C) was added and shaken for 15 min at 180 rpm. Two to three drops (1%) of superfloc flocculant was added and the bottle gently swirled. The overlying liquid was filtered into a clean Schott-bottle. Ten ml of the filtrate was transferred with a pipette to a 50 ml volumetric flask, to which ten ml lanthanum (La) (500 ppm) was added. The flask was then filled to volume with ammonium acetate, and the concentration of cations determined with an AAS.

3.5.3.4 Plant available phosphorus (P-Bray-1)

Plant available phosphorus of the samples was determined with the P-Bray 1 extraction which determines the soluble portion of phosphates in soil by using P-Bray 1 solution as extractant (Bray & Kurtz, 1945).

P-Bray 1 solution was prepared by dissolving 2.22 g ammonium fluoride (NH₄F) in deionized water which was transferred to a two-liter volumetric flask. Concentrated hydrochloric acid (five ml) was then added and the flask filled to volume with deionized water.

Ten grams of soil (< 2mm particle size) was weighed in a Schott-bottle, to which 75 ml of P-Bray 1 (20°C) solution was added and immediately shaken for 40 seconds. Two drops of Superfloc® flocculant was added to the solution, which was then gently swirled. Immediately after the suspension settled, the overlying liquid was filtered into a clean Schott bottle, and the concentration of phosphate was determined using an auto-analyser (Industrial Method 4-68W, 1996).

3.5.3.5 Macro-and micro-nutrients

The macro and micronutrient content of the samples was determined with a 1:2 extract method. The samples for this analysis did not undergo any pre-chemical treatment (Sonneveld & Van Den Ende, 1971).

Two to three drops (1%) of superfloc® flocculant was added to 200 ml of deionized water (dH₂O) in a plastic shaking bottle. 100 ml of soil was transferred systematically to the shaking bottle and shaken for 30 min. The clear overlying liquid was then decanted to low-speed centrifuge tubes, and centrifuged for 10 min at 2000 rpm. The resulted overlying liquid was then decanted into an Erlenmeyer flask. Two high-speed centrifuge tubes per sample were filled from the flask and centrifuged for 12 min at 16500 rpm. Liquid from the high-speed centrifuge tubes was then filtered into a bottle marked A. The same funnel and filter paper was used to filter the remaining liquid from the Erlenmeyer flask to a bottle marked B. The bicarbonate (HCO₃) concentration was determined from the liquid in bottle A by means of titration. Cations (Ca, Mg, K, Na & Fe) were determined from the liquid in bottle A with AAS. Anions (Cl, NO₃, NO₂, F, SO₄ & PO₄) were determined from the liquid in bottle A with ion chromatography. The ammonium (NH₄) concentration was determined from the liquid in bottle B using ion selective electrodes.

3.5.3.6 Soil organic carbon

The organic carbon content of the composite samples was determined by loss on ignition (LOI). This method determines the amount of organic carbon in soils gravimetrically, by burning off the organic matter content of the sample under controlled conditions in a furnace.

Dried soil (<2 mm particle size) was weighed out to 15 g in a high-temperature crucible. The crucible was then placed in a furnace and burned/baked for six hours at 650°C. After cooling the crucible was weighed again and the gravimetrical difference used to calculate the percentage soil organic carbon content of the sample.

3.5.3.7 Soluble trace metal content

The soluble/plant available trace metal concentration of the four soil samples was determined by means of the ammonia nitrate (NH₄NO₃) solution method. This analysis was done according to procedures set forth by the DIN 19730 method (ISO19730, 2008).

Samples of 20 g each were placed in a 100 to 150 ml shaking bottle, to which exactly 50 ml of the ammonium nitrate solution was added. The mixture was shaken for two hours at 20 rpm (25°C). Solid particles were allowed to settle for 15 min before the supernatant solution was filtered through a 0.45 µm filter. The first five ml was disposed of, and the solution that remained

was collected in a 50 ml bottle for analysis. The samples were analysed by inductively coupled plasma–mass spectrometry (ICP-MS) for soluble trace metals such as V, Cr, Co, Ni, Cu, Zn, As, Se, Mo, Pd, Ag, Cd, Ba, Pt, Au, Hg, Tl and U (Appendix F2).

3.5.3.8 Total trace metal content

The total trace metal concentration of the four soil samples was determined by means of acid digestion. This analysis was done according to the procedures set forth by the EPA 3050B method (EPA 3050B, 1996).

Samples of 25 g each were placed in a 150 ml beaker, to which 15 ml nitric acid (HNO_3) was added. The samples were immediately covered with watch glasses and placed on a sand stove inside a fume hood, which was set at a medium temperature of $\pm 95^\circ\text{C}$. The mixture was left to reflux for an hour during which fumes were generated. When the fumes diminished the samples were fully digested and the watch glasses were removed. The acid was evaporated by heating each sample until the volume was reduced to $\pm$ five ml. Each sample was then cooled in order to add three ml 30% hydrogen peroxide (H_2O_2). After cooling, 10 ml of 3N HCl was added and then covered again with a watch glass. The mixture was placed in the sand bath in the fume hood and was refluxed for about an hour. After the hour reflux, the sample was cooled down to 25°C . The mixture was filtered through a Whatman 40 filter paper into a 50 ml volumetric flask. The filter paper was washed with deionized water. The volumetric flask was filled to volume with deionized water. Samples were tested with ICP-MS according to the EPA 3050B method (EPA 3050B, 1996) for total trace metals such as V, Cr, Co, Ni, Cu, Zn, As, Se, Mo, Pd, Ag, Cd, Ba, Pt, Au, Hg, Tl, and U (Appendix F2).

3.5.4 Plant tissue analysis

The 24 powdered leaf and stem samples which were obtained from selected woody species were analysed for their total trace metal content (Appendix G1). The same method and procedure (EPA 3050B, 1996) were used as previously discussed for the soil analysis. The analysis was performed at Eco Analytica Laboratory, NWU, Potchefstroom.

3.6 STATISTICAL ANALYSES

Frequency values for each plant species in every plot were calculated and analysed utilising multivariate analyses. Primary ordination techniques, as described by Kent and Coker (1992), were applied with PRIMER (Clarke and Gorley 2006) and CANOCO (Ter Braak and Smilauer 2002) software to analyse the vegetation data.

3.6.1 Principal Component Analyses and Non-metric testing

Principal Component Analyses (PCA) was performed on non-transformed species count data (Appendix C1), to assess the floristic variation of the different plant assemblages in each geobotanical unit. This multivariate analysis represents a complex matrix of sampling plots and their associated species within a visual dimensional space, illustrating clustering of sampling plots based on similar floristic composition.

To reduce visual scattering and improve data normality, the species count data were additionally square root transformed and analysed using non-metric multidimensional scaling (NMDS). This provided an improved graphical representation of dissimilarity between the floristic assemblages of each geobotanical unit. In conjunction with the NMDS ordination, analysis of similarities (ANOSIM), a non-metric test based on dissimilarity measures, was performed to test for significant differences in species composition among geobotanical units in PAST (Hammer et al. 2001). In ANOSIM, the Bray–Curtis similarity index was implemented. This index takes into account the relative abundance of species and reveals changes in species community composition that are due to changes in the relative abundances of species.

3.6.2 Canonical Correspondence Analyses

Canonical Correspondence Analyses (CCA) was done to determine species correlations with specific soil chemical variables related to changes in rock geochemistry, which could influence the floristics of these assemblages. These analyses indicate the strongest linear correlations between specific woody species, and soil chemical variables depicted as diagonal vectors. Corresponding statistical summaries are provided in Appendix H1.

3.6.3 Species performance indexing

The stem circumferences of each tree species were computed to a performance index value for each sampling plot (Appendix H2) to determine the response of specific species to variations in the soil chemical element concentration. The Index values (*IV*) were calculated by multiplying the median stem circumference (*SC*) of each species with its frequency (*F*) which is divided by 50 points per plot.

$$IV = \frac{SC \times F}{50}$$

The index value, therefore, incorporates both the species abundance and productivity and was used as an indicator value for species performance.

3.7 MAPPING

3.7.1 Geology

Geological mapping of the KC was initially performed by plotting the locality of major outcrop positions on an orthophotograph. Thereafter by using the available strike and dip measurements along with established plant-rock type associations, a proposed geological map of the KC was produced, which is presented and discussed in Chapter 4 (Figure 9). Due to the shallow occurrence and weathered nature of the outcrops, strike and dip measurements for some intrusive rock types could not be obtained. A larger A3-sized geological map is also provided in Appendix I1.

3.7.2 Geobotany

In order to visually summarise the findings of this study, a geobotanical map of the KC was additionally created to provide an overview of the main geological, soil chemical and floristic associations of each geobotanical unit (Appendix I2). The map was produced by compiling three layers, including the base geology, the mean soil pH and Cr, Ni and Co concentration of each unit, as well as correlated dominant woody species.

CHAPTER 4

PETROLOGY AND GEOCHEMISTRY OF THE KOEDOESFONTEIN COMPLEX

4.1 INTRODUCTION

Investigation of the published literature and site geology of the KC revealed a general shortage of structural, petrological and geochemical knowledge with regards to both documented and undocumented intrusions. In particular, a large discontinuity in the current lithological record was noted, which may have left previous attempts to describe and interpret the complete association and petrogenesis of the KC partially unsolved. As a result, it was unclear how the syn-Bushveld intrusions are related to one another, and if the KC should be regarded as a singular body or a series of multiple intrusions.

The geological features of the KC were reassessed and the full extent and relationship of the intrusive rock series was documented. The justification, ultimately, is to construct an updated geological record that may provide an improved understanding of the field relation and petrology of the KC, so as to better understand the relationship between this Complex and the other members of the MDPS and HITIS (de Waal *et al.*, 2006, 2008) intrusions.

Specific objectives for the geological facet of this study were to map and describe the localised structure of the Complex (section 4.2) and to determine and discuss the mineralogical (section 4.3), petrographical (section 4.4) and geochemical (section 4.5) characteristics of the intrusive rock types.

Due to specific structural occurrences and mineralogical associations observed during field work, it is hypothesised that the group of intrusive rocks, apart from the older epidiorite intrusion, are genetically related, and collectively form a series of intrusive bodies, which is possibly the result of magmatic differentiation. This view is in accordance with Bischoff (1969,1999) who proposed that the KC is or most likely was, connected to the larger RTC, which is currently accepted as a layered differentiated intrusion.

4.2 STRUCTURE AND FIELD RELATION

The proposed structure and lithological distribution of the KC are illustrated in Figure 9. The general scarcity of large definable outcrops amongst the surface geology proved to be a restrictive factor in defining the rock boundaries for the intrusive series when conventional field observations methods are applied.

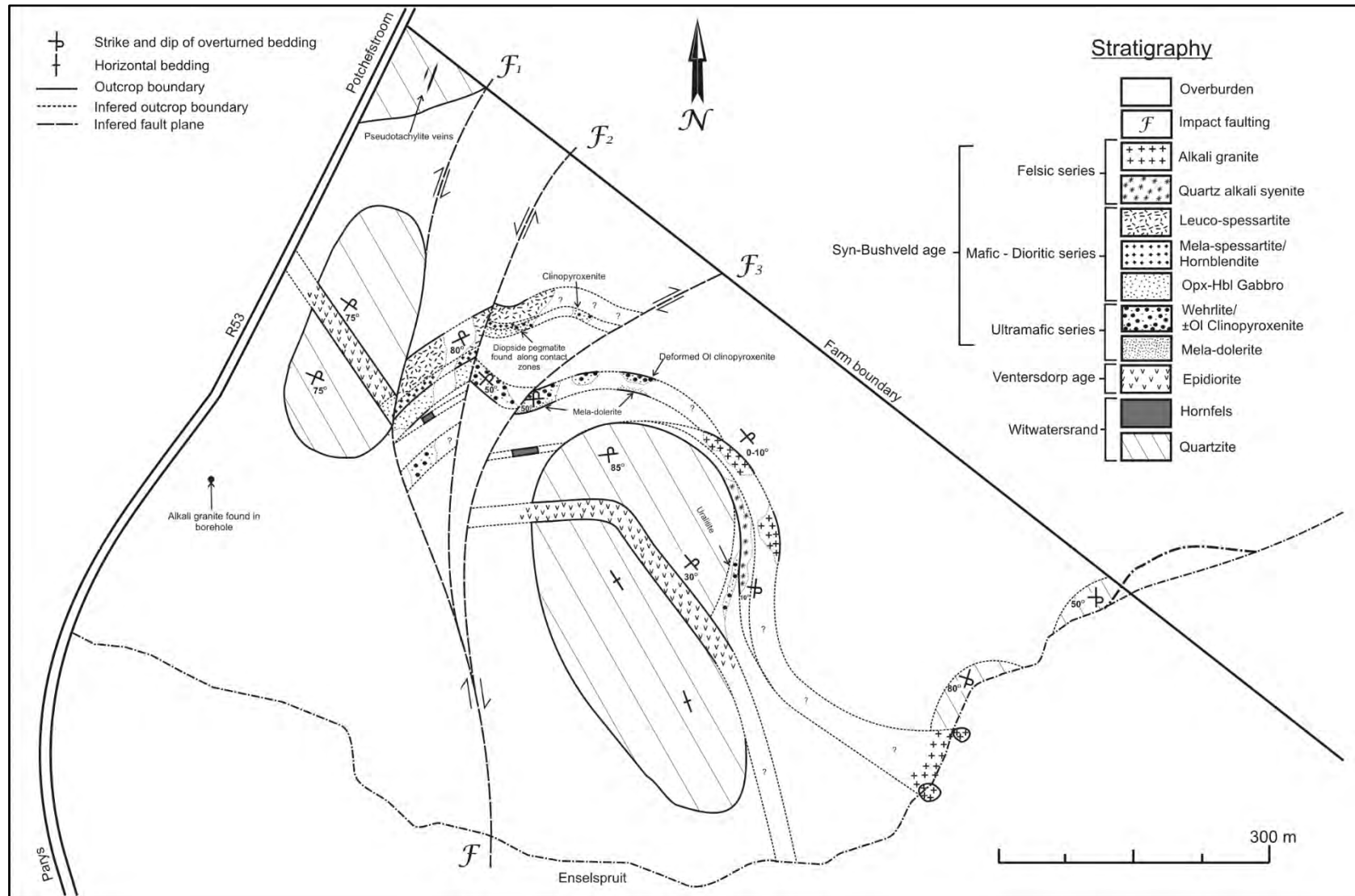


Figure 9: Detailed geological map of the Koedoesfontein Complex, displaying the proposed structure and lithostratigraphy of each intrusive and metamorphic rock type. Minor synthetic fault systems (F_1 , F_2 , and F_3) were inferred based on structural relationships (section 4.2).

Some boundaries and outcrop thicknesses were, therefore, inferred based on the distribution of loose bedrock fragments, as well as a number of soil and plant species indicators, that were identified to have a strong correlation with the geology.

The Complex comprises a relatively small area of about 1 km² and is composed of several rock types that are stratigraphically divided into three main age groups as summarised in Table 1. The intrusive rocks occur as thin dyke-like outcrops, which strike concordant along the bedding of the older deformed Witwatersrand metasedimentary sequences. Because these rocks have been subsequently overturned in this area, this observation, amongst several, suggests that the intrusions were originally emplaced as sills before the impact event. Another confirmation of this deduction is when the Witwatersrand rocks are rotated to their original horizontal position, the syn-Bushveld intrusions of the KC illustrate corresponding modal and phase layering sequences indicative of differentiation, with the ultramafic intrusions occurring stratigraphically lower than the succeeding mafic, dioritic and felsic counterparts. More information regarding this characteristic will be discussed throughout the succeeding sections.

Small scaled compressional folding and signs of related thrust faulting, consistent with impact deformation, are prominent throughout the Complex (Figure 9). The south-eastern portion of the structure forms a near vertical plunging anticline of which the fold axis trends north-eastward. The orientation of this fold is consistent with a north-westward compressional force applied parallel across the bedding. This is particularly interesting, as this observation indicates that the strike of the Witwatersrand rocks, in this specific area, was far more rotated towards the center of the dome, during impact deformation, than the neighboring outcrops surrounding the Complex. This is possibly the result of part of the lower portion of Central Rand Group rocks which have been displaced earlier within the sequences of the West Rand Group along the two major faults of Venterdorp age (Wieland *et al.*, 2005). According to Wieland *et al.* (2005), there seems to be little post-tectonic reactivation of these fault systems during the deposition of the Transvaal Supergroup. Therefore, it can be reasonably assumed given the larger throw of the eastern fault, and thickening of the Venterdorp lavas to the west of this feature (directly above the KC), that the bedding of the lower Central Rand Group was already discordant and somewhat rotated prior to emplacement of the younger intrusions, and subsequent overturning during impact. This observation supports Bisschoff's (1973) view that this particular fault zone, including possible secondary fault systems, could have played a significant role in the emplacement of these specific intrusions.

In the north-western portion of the study area, the western limb of the anticlinal fold connects to what resembles a severely faulted sinclinal fold, situated between the two ridges of quartzite (bordered by F₁ north-west and F₃ south-west). In this area, the bulk of the syn-Bushveld

intrusions is encountered straddling across remnants of what must have been a much larger igneous structure. On the north-westward facing slope of the eastern quartzite ridge (F_3), outcrops generally strike east to west in accordance with the western limb of the anticline. On the lower south-eastern slope of the western quartzite ridge (F_1), however, outcrops of the syn-Bushveld intrusions were found to strike northeast-southwest as opposed to northwest-southeast such as the epidiorite intrusion partly exposed further to the northwest (west of F_1). This structural feature, along with mineralogical and petrographical observations, indicate that different portions of the sills must have been displaced relative to one another, and therefore a proposed system of synthetic northwest trending thrust faults (F_1 , F_2 & F_3 ; Figure 9) is inferred on the fault zone indicated by Bisschoff (1999).

The south-eastern portion of the Complex, east of F_3 , is thus thought to represent the main foot wall of the structure, which has been displaced underneath the hanging wall west of F_1 . This resulted in an upward compressional slip of the hanging walls west of F_2 and F_3 and effectively shortening of the intrusive bodies at their current erosion level. To account for some rotational movement of the syn-Bushveld intrusions towards the epidiorite dyke situated on the north-western portion (western ridge), and away from the dyke on the south-eastern portion (eastern ridge), faults F_1 and F_3 has a minor dextral strike-slip component. To account for the more northerly occurrence and folded nature of the ultramafic dykes on the westward facing slope, fault F_2 was inferred with a sinistral strike-slip component.

General thickening of the sills towards the western portion of the faulted syncline, suggests that they might have been lopoliths, and based on the folded nature had some structural relation to the folding event. In this case, it can be argued that some initial folding, as a result of the former displacement event, must have occurred prior to impact. This feature, in turn, might have exerted some structural control over emplacement of the younger intrusions.

4.2.1 Witwatersrand metasedimentary rocks

Metamorphosed quartzite and hornfels (shales) of the upper Witwatersrand Supergroup is the oldest bedrock lithologies of the study area. The largest portion of exposed strata, which dips to the southwest, forms part of the Main Quartzite Formation (Bischoff, 1969, 1999) that gives rise to the prominent ridge. South of the concordant epidiorite intrusion, the bedding is tilted vertical, while to the north of the intrusion the bedding has been overturned, attaining an average dip angle of 30° .

Fragments of dark hornfels containing anthophyllite were commonly encountered along the westward facing slope and near the southern base of an outcrop of mafic rocks. This is believed to represent an altered shale bed, however, no solid outcrops was encountered.

On a more localised scale, the quartzite bedding displays common impact related deformation. Shear zones (Figure 10), shatter cone structures (Figure 11), as well as small veinlets of pseudotachylite (Figure 12), were regularly encountered throughout most of the exposed strata.

The two smaller outcrops of quartzite that occur to the northwest of fault zone F_1 differs from the southeast portion. The beds seem generally void of metapelites, less brittle, more mineralogically mature, and contain better-preserved veinlets of pseudotachylite. If the proposed fault zones have some amount of oblique slip as suggested, these strata might represent different sections of the Main Quartzite Formation which have been displaced vertically, however, the largest slip along F_1 is thought to be across the dip of the fault plane. The bedding was measured to strike in a southeast direction with an average dip angle of 75° .



Figure 10: Shear zones displaying smooth surface striations (red arrows) along the oxidized bedding planes of an outcrop of quartzite, exposed in north-western portion the Koedoesfontein Complex. Note how the white quartzite in the centre of the lower bed (white marker lines) transitions sharply towards a light brown colour at the contact margins. This colour transition, (due to oxidation), suggests a prolonged exposure to water, which was able infiltrate along the fracture planes that weakened the contact margins the individual beds.



Figure 11: Small shatter cone structures (red arrows) typically found along the bedding planes of the south-eastern quartzite ridge on the Koedoesfontein Complex.



Figure 12: Veinlets of glassy pseudotachylite (red arrows), found in the quartzite bedrock of the north-western portion of the Koedoesfontein Complex.

4.2.2 Ventersdorp-age rocks

The altered intrusion of epidiorite is situated stratigraphically lower than the succeeding ultramafic intrusions and is accepted to be the oldest igneous lithology of the Complex. The intrusion outcrops as a differentially weathered indent along the ridges of quartzite occurring on either side of the sheared zone where it is cut discordantly by displaced portions of younger mafic and dioritic rocks.

The largest part of the overturned sill is exposed laterally in the south-eastern portion of the Complex, dipping around 30° southwest. The sill trends parallel with the bedding of the Witwatersrand metasedimentary sequences which become more sharply inclined towards the north-western portion, attaining an average dip angle of around 85° southwest. Samples of this rock type, along with hornfels were primarily found on the highest parts of the outcrop occurring on the escarpment of the quartzite ridge (Figure 13).

In the north-western portion of the Complex, no definable outcrops of epidiorite apart from loose fragments could be observed along the weathered indent. Remnants of the eroded sill in the north-western portion, also seem to cut more discordant across the quartzite bedding.



Figure 13: Quartzite xenolith (red arrow) found in an exposed sample of epidiorite at the south-eastern escarpment of the Koedoesfontein Complex. Note the epidiorite's greenish colour due to the presence of fibrous hornblende (uralite).

4.2.3 Syn-Bushveld ultramafic rocks

Syn-Bushveld ultramafic outcrops are exposed primarily at four localities within the Complex. On the north-western slope of the eastern quartzite ridge, scattered samples (Figure 14) of an overturned sill (appearing to be a dyke) of wehrlite are found outcropping as separate dykes striking in alternating directions about 100 m from one another. These seem to form a single overturned ultramafic body dipping roughly at 50° south-southwest. Outcrops on the northern lower end of the slope strikes northwest-southeast, while outcrops to the south, along the mid-slope, strike in an east-westward direction. If these bodies are regarded as a single structured sill, then it forms a small moderately folded syncline which joins with the anticlinal fold to the southeast.

The northern end of the southern wehrlite dyke appears to be cut sharply by a fault striking approximately northeast-southwest. This feature was interpreted as part of fault zone F₃. Based on differences in modal olivine content, it is assumed that the southern dyke (which is in actual fact an olivine clinopyroxenite) has been partly displaced to the south.

In the southern contact of the wehrlite / olivine clinopyroxenite dyke, very fine-grained samples varying from almost pure magnetite to spinel-rich mafic rocks were found (sample RB1A; Appendix A2). Although no solid outcrops could be verified these seem to be closely associated with the wehrlite body and probably forms part of a chilled contact layer.



Figure 14: A typical sample of moderate- to coarse-grained wehrlite / olivine clinopyroxenite, found scattered along the ultramafic outcrops on the north-western slope of the eastern quartzite ridge. These rocks can be distinguished by their speckled, grey-silverish to almost sub-metallic luster produced by the dark augite crystals.

In the south-eastern sector of the Complex, on the upper eastern slope of the quartzite ridge, another thin portion of the ultramafic sill is encountered which at a glance seems texturally similar to the dykes of wehrlite, however, the augite is almost entirely replaced by fined-grained uralite (Figure 41). The sill has been completely overturned and occurs less than two meters above a sill of felsic rocks, which dips at 10° to the west.

Severe alteration due to weathering could also be observed in some of the hand samples, of which the crust had oxidised red (Figure 15) or completely converted to talc and serpentine. Impact deformation features were also prominent in the form of small V-shaped striations that may represent partial shatter cone structures or shatter fans (Figure 15). Unfortunately, large enough outcrops could not be found to clearly verify this observation. Very minute pieces of fine-grained magnetite-rich rocks were also found along this area, but no indication of a structural layer could be observed.

Following the general strike pattern of the uralitised section of the sill down towards the northern slope, just west of an intrusion of alkali granite, another small ultramafic outcrop occurs of what appears to be a impact deformed olivine clinopyroxenite sharply inclined to the southwest.

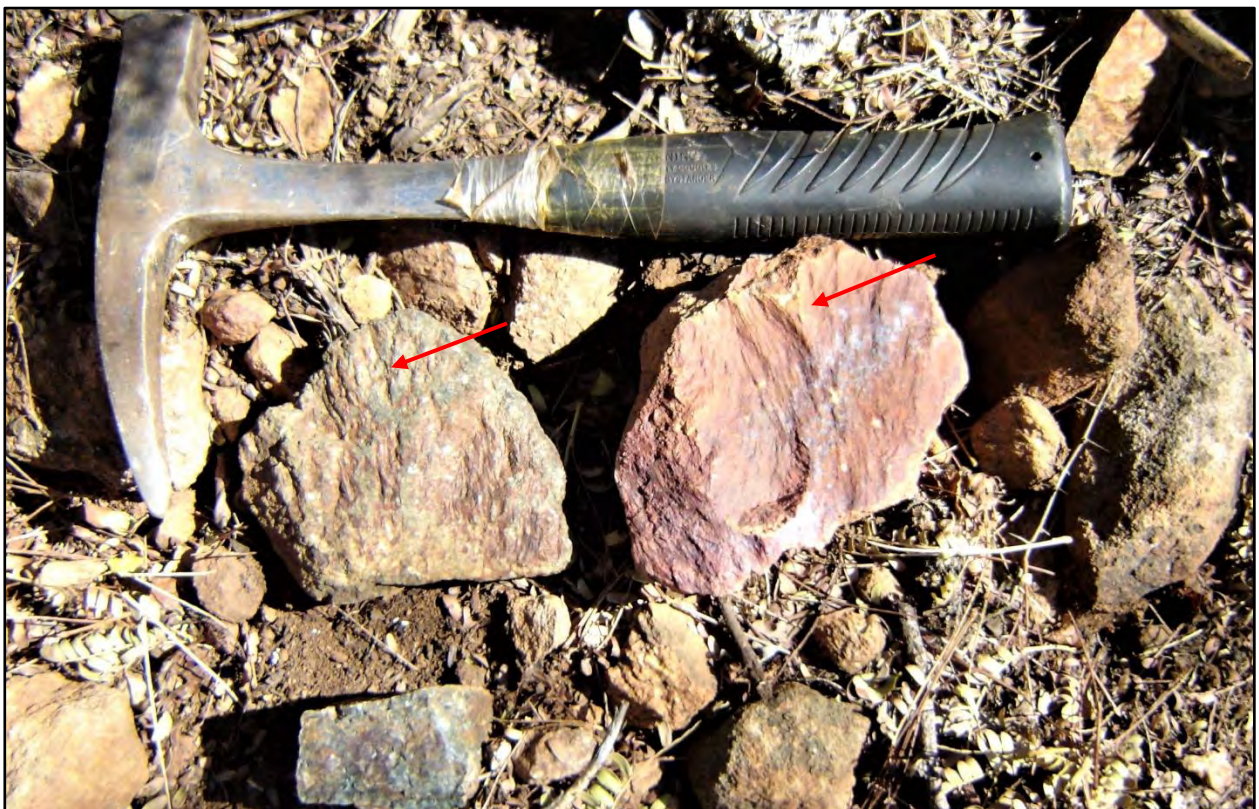


Figure 15: Samples of uralitised clinopyroxenite obtained from a small sill structured outcrop on the eastern slope of the eastern quartzite ridge. Some red surface discoloration due to weathering can be observed, as well as small deformation structures consisting of V-shaped striations (red arrows).

Compared to the wehrlite and olivine clinopyroxenite outcrops of the lower north-western end, larger amounts of alteration and deformation could be observed in samples of the northern slope which have small prominent veinlets of pseudotachylite. In this area, the sill was observed to also contain a partly exposed layer of highly fractured and weathered magnetite rich rocks.

About 80 m north of the north-western body of wehrlite, fine-grained uralitised and non-uralitised samples of clinopyroxenite containing accessory intergranular plagioclase was encountered (Figure 16). These rocks also grades towards the northerly displaced layer of gabbroic rocks occurring to the east of fault zone F₂. No magnetite-rich rocks were encountered along this layer, however, a sample of pegmatitic schlieren (RB18) was found that comprises almost entirely of green diopside augite.



Figure 16: A sample of fine-grained non-uralitised clinopyroxenite, found grading westward towards a partly displaced sill of gabbroic rocks, in the north-western portion of the Complex. Note the orange-brown crust due to surface oxidation (white arrow), where the augite is converted to talc and serpentine. Surface striations along the fracture planes (red arrows) are also clearly visible: these are indicative of shock deformation.

4.2.4 Syn-Bushveld mafic to dioritic rocks

Outcrops of mafic-dioritic intrusions are located mainly in the north-western portion of the Complex, scattered along the sheared zone. On the western and north-western contacts of the northern wehrlite outcrop (west of F₂), sparsely distributed samples of medium-grained gabbro are erratically encountered (Figure 17 and Figure 18). These outcrops seem to form a layer that grades northward to an overturned and partially rotated sill of what can be generally classified as lamprophyric rock, which strikes almost perpendicular to the dyke of wehrlite.

The lamprophyric sill is tilted almost vertical, dipping steeply at 80° to the southeast. Moving perpendicular across the strike of the outcrop, from the south to the northern end, a large amount of textural and mineralogical variability (modal and phase layering) could be observed amongst the surface rocks.

This feature proved rather enigmatic towards classifying the lamprophyric body as a single entity since four different rock types could ultimately be distinguished within the apparent layers.

At the southern base of the outcrop, the rock is generally more melanocratic, varying from a cumulate phase of very coarse-grained equigranular hornblendite (Figure 19 and Figure 20) to a porphyritic spessartite with phenocrysts of dark brown hornblende (Figure 19 and Figure 21). In the north, the majority of the spessartite appears more leucocratic and poikilitic as the oikocrysts of brown to green hornblende are mantled by a layer of plagioclase grains, which also become more modally pronounced within the matrix (Figure 21). In these rocks schlieren of coarser-grained equigranular varieties also occur (Figure 22).

It is therefore proposed that the sill consists of two principle layers: a thinner melanocratic basal layer of hornblendite that modally grades to mela-spessartite containing lesser amounts plagioclase, (hornblende > plagioclase), followed by a main leucocratic layer of mostly poikilitic spessartite, and pegmatitic schlieren of plagioclase and hornblende (hornblende < plagioclase). The pegmatitic schlieren is mainly found along the upper portion of the sill.

A similar sequence of rocks as described above is found to the east of fault zone F₂. In this area more scattered pieces of gabbro, texturally similar to those in the northwest, were found, which seems also to grade towards a northerly body of dark to light colored porphyritic spessartite (Figure 20). Due to the fairly weak outcrop exposure, however, this relationship could not be clearly identified.



Figure 17: An example of a gabbro specimen containing a thin vinelet of labradoritic plagioclase, found on the eastern slope of the western quartzite ridge.



Figure 18: Typical medium-grained gabbroic specimens, from west of fault zone F_2 bordering the northern wehrlite outcrop.



Figure 19: Samples of the melanocratic, equigranular (left) and porphyritic (right) rock types found along the southern portion of the main spessartite body.



Figure 20: A contact sample of quartzite and melanocratic spessartite, found at the base of a northerly outcrop, occurring east of fault zone F₂.



Figure 21: Leucocratic spessartite displaying a poikilitic texture, found along the northern portion of the main spessartite body.



Figure 22: A sample displaying the contact between quartzite and plagioclase schlieren, found along the northern portion of the main spessartite body. The schlieren contains aggregated monomineralic glomerocrysts of brown-green hornblende that measure 1 to 1.5 cm in diameter (red arrow).

4.2.5 Syn-Bushveld felsic rocks

Highly weathered intrusions of felsic rocks are partly exposed at the south-eastern portion of the Complex. On the eastern slope of the eastern quartzite ridge, thin displaced sills of quartz alkali syenite, with sharp contacts, are found emplaced between the bedding planes of quartzite (Figure 23), occurring just below, and j to the east of the sill of uraltised clinopyroxenite crosscutting this altered clinopyroxenite sill. The quartz alkali syenite dips along the overturned meta-sedimentary strata at roughly 10° to the west, and are observed to grade downward (that means upward in the stratigraphy) towards an intrusion of alkali granite (Figure 24).

On the contact surface of the quartzite bedding (F_a in; Figure 23), long striations indicative of fault movement are observed. This strengthens the notion that the emplacement of the sills was most probably controlled by pre-impact faulting as formerly suggested. The sill and bedding are cut vertically by younger faults which display minor strike-slip movement (F_b in; Figure 23). If these faults are accepted as of impact-related origin, then this would place the time of emplacement before the impact event. Another indication of this relationship is the presence of fine-grained pseudotachylite, which was found in samples of the alkali syenite (Figure 25).

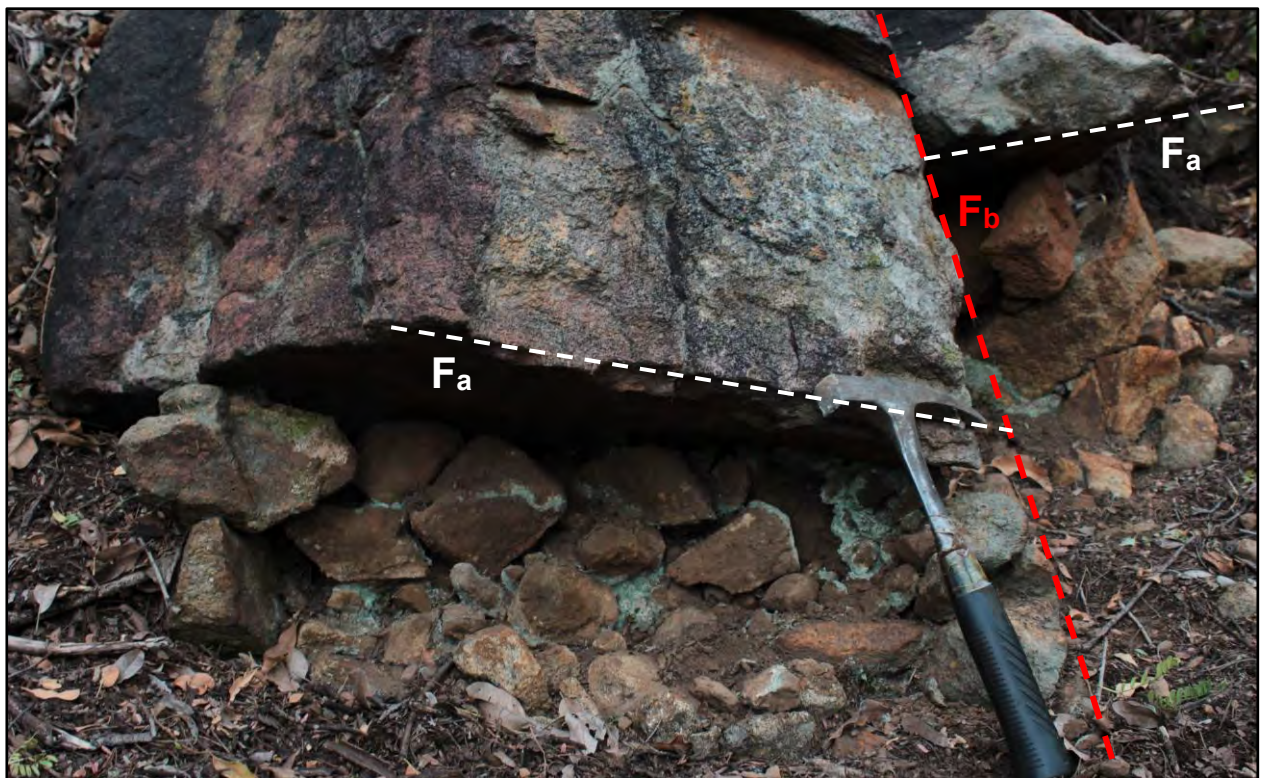


Figure 23: Exposed lower contact of a tilted sill of alkali syenite, found outcropping on the south-eastern portion of the Koedoesfontein Complex. The bedding plane surface of the quartzite bedrock (F_a) is smooth and contains parallel surface striations. This feature is vertically displaced by a younger lateral fault (F_b).



Figure 24: An alkali syenitic rock sample containing a vein of microcline pegmatite, found in the lower transitional contact of the sill of alkali syenite where it grades towards the exposed layer of weathered alkali granite.



Figure 25: An alkali syenite rock sample containing fine-grained fractured pseudotachylite, found along the upper base contact of the alkali syenite sill.

Spanning the north-eastern slope of the eastern ridge, about 20 m below the sill of quartz alkali syenite, small stock-like outcrops of alkali granite occur which have been almost completely converted to saprolite (Figure 26). It seems, to some extent, that these outcrops form an elongated body along the strike of the quartzite bedding, similar to the alkali granite of the RTC (Bisschoff, 1973), however, the exact outcrop pattern was difficult to interpret due to a high amount of colluvial cover.

The outcrops strike in a south-eastern trend, from the slope of the eastern ridge down towards the carved out stream bed where another highly weathered outcrop is encountered. Due to the lack of exposure and degree of weathering, no measurements could be obtained for these outcrops. Fresh rock exposures were also extremely limited, and only one sample (RB15; Appendix A2) could be obtained from the upper outcrops where they grade towards the sill of alkali syenite.

In the western portion of the Complex, no felsic outcrops were found, although, Bisschoff (1973) does make note of alkali granite being encountered in the chips obtained from a water borehole to the south-west (Figure 9). This was interpreted as another possible pluton which may underlie this area.



Figure 26: An outcrop of weathered alkali granite occurring on the north-eastern slope of the eastern quartzite ridge, about 20 m below the base contact of the alkali syenite displayed in Figure 23. In the center and lower right corner of the photo thin veinlets of preserved microcline pegmatite can be observed.

4.3 MINERALOGY AND NOMENCLATURE

The mineralogical composition of representative samples from each intrusive rock type, as determined in weight percentages from XRD analysis, is displayed in Appendix D2. Accessory minerals that occur below 1 to 2 wt. % was confirmed petrographically (section 4.4) and are not detected by the XRD analysis. This is because their diffraction peaks were too weak to be identified, as they are masked by overlapping peaks of more abundant minerals or background noise (Appendix D1). This factor relates to a small degree of error when the mineralogical composition is recalculated to mode percentages based on volume (Appendix D3), however, due to the fact that they are accessory, this error was permissibly small and did not influence their nomenclatural classification based on normalised values.

When the optically determined anorthite content of the plagioclase was compared with the XRD results (Appendix D1), it was noticed that some members of the plagioclase series, initially classified based on peak identification (Boneschans *et al.* 2015), were also under-valued by the software. Instead, the anorthite content of the plagioclase was estimated using the Michael-Lévy method (Nesse, 2004), to assist with proper classification.

Samples were classified according to the classification schemes and recommendations set forth by the International Union of Geological Sciences (IUGS) Subcommission on the Systematics of Igneous Rocks (Le Maitre, 2002) and Subcommission on the Systematics of Metamorphic Rocks (SCMR) (Fettes & Desmons, 2007). Since all the sills are originally igneous in origin and show some degree of alteration due to autometasomatism and associated hydrothermal metamorphism, samples apart from the discernible aphanitic epidiorites, are classified as metamorphic only when the secondary mineral phases are extensively predominant over primary ones. This distinction is primarily applicable to the clinopyroxenite sill and its samples, which have been completely uralitised to fibrous Mg-hornblende (uralite). Although the cumulus phase of hornblende (RB6B) and lower layer of porphyritic rocks (RB6A), do also border this distinction, the igneous classification scheme will still receive preference above the scheme for metamorphic rocks. This decision was made in order to retain nomenclatural unity within the entire series of syn-Bushveld rocks, as well as in respect to the surrounding intrusive complexes of the Vredefort dome (MDPS intrusions).

The classification of samples in the following discussion will be dealt with according to their established relative age and field relationships. Samples from the epidiorite dyke are classed first, followed by the samples of younger intrusions of possible syn-Bushveld age. The latter samples are compiled into an ultramafic, mafic-dioritic and felsic series, and will be named based on the primary mineral's normalised weight percentages (Tables 3, 4, 5 & 6). Petrographical

observations (section 4.4) that follow will have to underwrite the classification otherwise the naming will have to be reconsidered. The same applies for geochemical trends.

4.3.1 Ventersdorp-age Epidiorite

Epidiorite is currently regarded as a field term for mafic intrusive rocks that have been altered by low-grade hydration metamorphism, which caused recrystallization and alteration of the original plagioclase and pyroxene (Bisschoff, 1999; IUGS, Le Maitre, 2002). The term is now specifically restricted to samples in which either relict minerals or relict textures, or both, are still preserved, as epidiorites are accepted to be the transitional link between mafic igneous rocks and true orthoamphibolites in which the original minerals and textures have been completely recrystallized (Epidiorite, 2011).

Epidiorites are in fact dioritic in terms of their secondary phase mineralogy, although based on their silica content they are still distinctly more basic than typical diorites (Encyclopedia Britannica, Epidiorite, 2011). This difference relates to mineral alterations that include the limited or complete conversion of augite to actinolite or fibrous Mg-hornblende (uralitisation - Best, 2003; Buseck *et al.*, 1980; Nesse, 2004), the crystallisation of sphene, and the saussuritisation of plagioclase members to albite and epidote or ziosite. As the grade of metamorphism increases the albite becomes more anorthite rich as epidote simultaneously diminishes (Best, 2003; Bisschoff, 1969; Epidiorite, 2011).

Sample RB16 and RB17 (Appendix D3), which was taken from the exposed sill tentatively classified as epidiorite, display the typical characteristics as specified to confirm this deduction. The samples are generally dioritic composition, although quite variable, comprising primarily fibrous Mg-hornblende (uralite) (~38 to 43vol. %), oligoclase (~19 to 41 vol. %), quartz (~19 to 39 vol. %) and minor amounts of preserved diopsidic augite (< 2 vol. %). Other accessory minerals include epidote, sericite and magnetite.

On the basis of their altered dioritic mineralogy and remnant portions of augite, fine-grained recrystallized texture (section 4.4.1), and corresponding low silica content (~40 to 51 wt. %; Appendix E1) samples RB16 and RB17 are verified as epidiorite.

4.3.2 Syn-Bushveld igneous and metamorphic rocks

Mineral compositions of the intrusive rock types that are probably of Bushveld age are consistent with the original thin section descriptions of Bisschoff (1969). In addition, detailed sampling of the outcrop areas revealed variations in texture and mineralogy, which, along with the Complex's structure, indicate that the individual sills are modally layered. The corresponding rock types are grouped into series primarily based on their mineralogical trends. Some textural, structural and

whole-rock chemical characteristics were also taken into account to avoid contradicting relative age and field relationships. For instance, the fine-grained mafic rock that occurs as layer at the base of the main wehrlite body, is taken as the most primitive in terms of geochemistry (section 4.5), and will be treated as part of the ultramafic series in spite of its mafic character, *i.e.* its appreciable plagioclase content. Secondly, the equigranular cumulus phase of hornblende, which is classified as ultramafic, will be grouped along with the mafic-dioritic series as it occurs along the base layer of the lamprophyric sill and is considered to be a younger formation than the rocks of the ultramafic series.

4.3.2.1 Ultramafic series

Sample RB1A (Table 3 & Figure 27) is the most primitive of the ultramafic series representing a chilled contact zone, which is mafic in composition. The rock comprises primarily of brown hornblende (~50 vol. %), diopsidic augite (~26 vol. %) and labradorite (~24 vol. %), and is associated with other spinel-rich varieties, however, magnetite in this particular sample is accessory along with orthopyroxene, serpentine and sericite.

Based on sample RB1's primary mineral content it can be classified as a pyroxene hornblende gabbro (Figure 27c & d). Since the sample is very fine-grained (section 4.4.2), the name pyroxene hornblende dolerite would thus be more suitable (Le Maitre, 2002). For ease of reference, the sample will be termed a mela-dolerite, as termed appropriate for its high relative percentage of mafic minerals (Figure 27d).

The main outcrops of the ultramafic series (samples RB7, RB8, RB9 & RB10) are composed primarily of augite (~45 to 66 vol. %), forsterite (~9 to 49 vol. %) and secondary brown hornblende (≤ 17 vol. %). Minor alteration products include talc ($\leq \sim 7$ vol. %), lizardite (≤ 6 vol. %) and magnetite, while sericitised plagioclase and apatite are accessory in some samples (see Figure 40, sample RB31).

Samples from the north-western portion of the ultramafic sill (RB7 & RB8) display a greater modal composition of forsterite (~49 to 47 vol. %) in relation to augite (~48 to 45 vol. %) and contain no hornblende. Samples of the southern portion (RB9 & RB10) show a sharp decrease in modal forsterite (~9 to 40 vol. %), while augite (~51 to 66 vol. %) and hornblende (~7 to 17 vol. %) increases more gradually.

According to the classification scheme for ultramafic rocks (Figure 27a & b), samples RB7, RB8 and RB9 are classified as wehrlites, whereas sample RB11 is classified as olivine clinopyroxenite.

Sample RB14, which was taken from the upper sill near the felsic intrusions, has almost completely been altered and is composed mainly of fibrous Mg-hornblende (~84 vol. %) and minor

amounts muscovite (~8 vol. %) and clinochlore (~5 vol. %). Diopsidic augite (~3 vol. %) and magnetite, which can be observed optically (Figure 41), is accessory.

Table 3: Normalised modal compositions (vol. % - see Appendix D3) of analysed primitive-mafic (RB1A) and ultramafic rock samples.

Sample	Pyroxene	Hornblende	Plagioclase
RB1a	26	50	24 _(An>50) [#]
	Olivine	Pyroxene	Hornblende
RB7	50	50	0
RB8	52	48	0
RB9	41	52	7
RB10	10	72	19
RB14	0	3	97

[#](An) = analytically determined with XRD

Because sample RB14 is regarded as metamorphic, it is classified based on the scheme and recommendations of the SCMR (Fettes & Desmons, 2007). According to the scheme, the rock can be generally regarded as a hornblende granofels, although in this particular case the more specific term uralite can be applied, as to denote its specific amphibole uralite, which is the fibrous variety of Mg-hornblende specifically derived from the deuteritic alteration of augite (Best, 2003, Nesse, 2004).

4.3.2.2 Mafic - dioritic series

Outcrops of the mafic - dioritic series (samples RB3, RB4, RB5, RB6A & RB6B) are composed primarily of amphibole (~18 to 91 vol. %), ± plagioclase (~2 to 56 vol. %), ± quartz (~0.1 to 21 vol. %) and ± diopsidic augite (~11 to 18 vol. %). The amphibole grades from hornblende to Mg-hornblende, which is generally the dominant variety. Apatite, magnetite and orthopyroxene are observed optically as minor or accessory minerals. Clinocllore (~2 to 8 vol. %), together with saussurite containing epidote and sericite (observed optically) are alteration products.

The mafic - dioritic rocks are classified based on their prominent feldspar and mafic mineral phases (Table 4), and the distinction between the terms mafic and dioritic is based mainly on the anorthite content of the feldspar. Equigranular varieties (RB3 & RB6B) are named in accordance with the schemes for phaneritic rocks (Figure 27c & d), while the porphyritic varieties (RB4, RB5, and RB6B) will be named according to the nomenclatorial scheme for lamprophyres (Table 5). In order to illustrate the relationship amongst the samples in terms of their mineralogical composition, the lamprophyric samples are also included in the schemes for phaneritic rocks (Figure 27c, d & Figure 28).

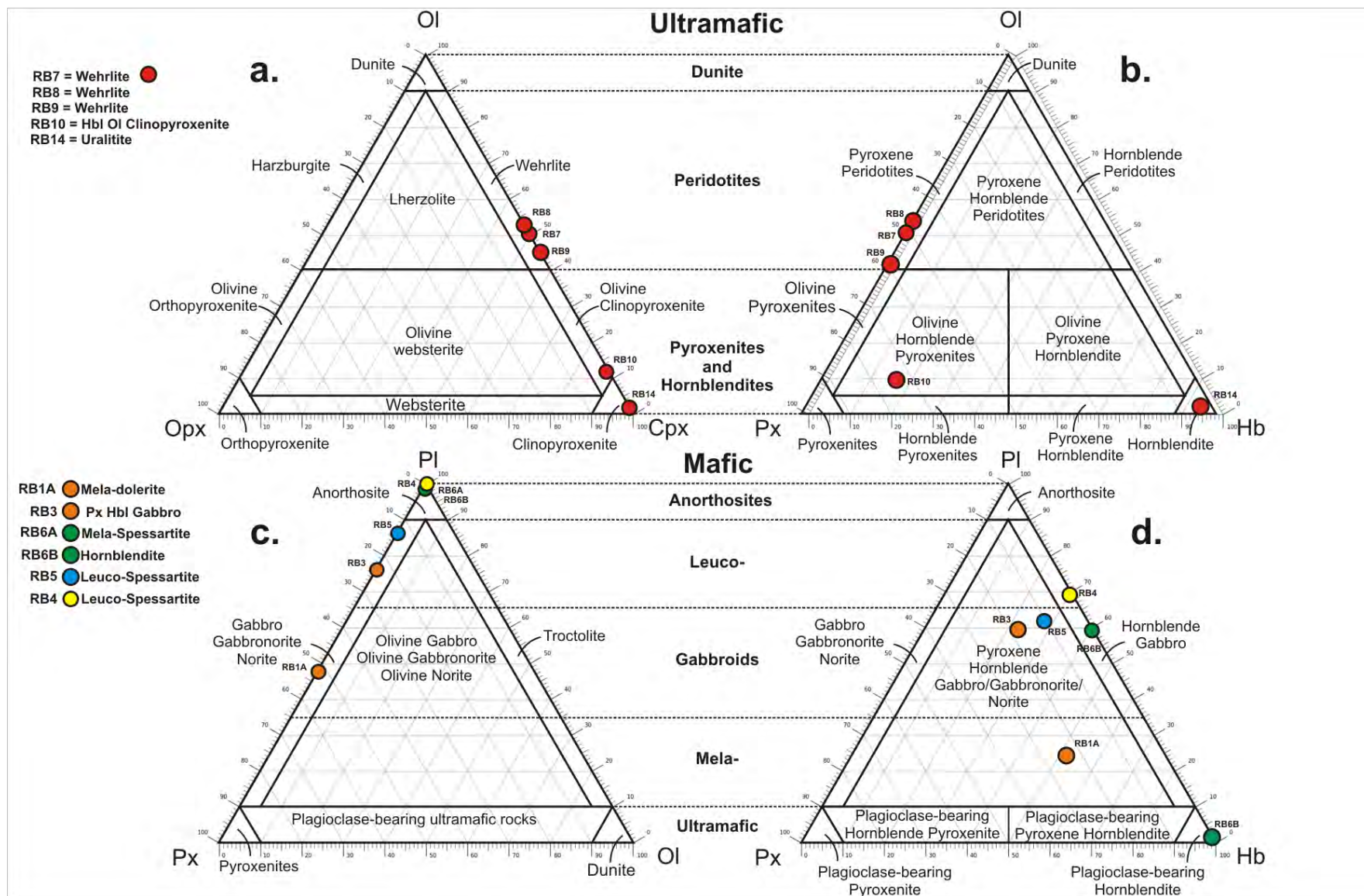


Figure 27: Modal classification of ultramafic (a & b) and mafic (c & d) rocks based on the proportions of olivine (Ol), orthopyroxene (Opx), clinopyroxene (Cpx), pyroxene (Px), hornblende (Hbl) and plagioclase (Pl). Adapted from Le Maitre (2002; Figures 2.6 & 2.9), after Streckeisen (1973; Figures 2a & 2b, 1976; Figure

Table 4: Normalised modal compositions (vol. % - see Appendix D3) of analysed mafic and dioritic rock samples.

Sample	Pyroxene	Hornblende	Plagioclase
RB3	18	23	59 _{(An>50)*}
RB6a	0	41	59 _{(An<50)*}
RB6b	0	99	1 _{(An<50)[#]}
RB5	12	26	62 _{(An<50)*}
RB4	0	26	74 _{(An<50)[#]}

* (An) = optically determined. # (An) = analytically determined (XRD)

Table 5: Classification and nomenclature of lamprophyres based on their mineralogy. Adapted from Le Maitre (2002; p.19), after Streckeisen (1978; p.11).

Light coloured minerals		Predominant dark coloured mafic minerals		
Feldspar	Foid	Biotite > hornblende, ± diopsidic augite, (± olivine)	Hornblende, ± diopsidic augite, (± olivine)	Brown amphibole, Ti-augite, olivine, biotite
Or > Pl	-	minette	vogsite	
Pl > Or	-	kersanite	spessartite	
Or > Pl	Feld > foid			sannaite
Pl > Or	Feld > foid			camptonite
-	Glass or foid			monchiquite

Or =alkali feldspar; Pl=Plagioclase; feld = feldspar; foid = feldspathoid.

Sample RB3 is mafic and comprises mainly of andesine plagioclase (~55 vol. %), Mg-hornblende (~21 vol. %) and diopsidic augite (~18 vol. %). Small amounts of clinocllore and muscovite are present (~2 to 8 vol. %), and the occurrence of kaolinite (< 2 vol. %) is attributed to muscovite weathering (≤ 2 vol. %). Although the hornblende is secondary, it is not predominant over the total primary minerals and will thus still be regarded as igneous. The sample is classified as a pyroxene hornblende gabbro (Figure 27d), and in this case the term hornblende will be included since it is the dominant mafic mineral. Some accessory orthopyroxene can also be observed optically (Figure 43), and therefore the term orthopyroxene hornblende gabbro will be used to recognise the occurrence of this pyroxene.

Sample RB6B is ultramafic and composed primarily of granular brown to green hornblende (~91 vol. %). Clinocllore occurs as a minor alteration mineral (~8 vol. %), whereas oligoclase (~1 vol. %) and quartz (< 1 vol. %) are accessory. Optical observation (section 4.4.6) reveals that the hornblende is secondary after clinopyroxene, but since the hornblende grains form a cumulate texture (Figure 44), as a result from crystal settling during deuteric alteration, the rock will still be treated as igneous. The sample is accordingly classified as a hornblendite (Figure 27d).

Sample RB6A is dioritic, containing mainly oligoclase (~57 vol. %) and brown to green hornblende (~40 vol. %). Clinocllore (~2 vol. %), magnetite and apatite are accessory. Mineralogically, the sample is equivalent to a hornblende diorite (Figure 27d), although, due to its texture, it will be classified as a spessartite lamprophyre (Table 5). To denote its darker appearance, as a result of more hornblende, the term mela-spessartite will be used.

Samples RB4 and RB5 are classified as lighter, more poikilitic varieties of spessartite lamprophyre. They contain almost exactly the same amount of oligoclase (~56 vol. %), similar to sample RB6A, although about half as much Mg-hornblende (~20 to 24 vol. %). Quartz becomes a minor to major constituent (~7 to 22 vol. %) along with remnant diopsitic augite in sample RB5 (~11 vol. %). Clinocllore (~2 vol. %), magnetite and apatite are also accessory.

Based on their feldspar and mafic mineral content (Table 4) samples RB5 and RB4 are similarly equivalent to a diorite (Figure 27d), although if the quartz - alkali feldspar - plagioclase (QAP) scheme is applied (Figure 28) they vary from quartz diorite (RB5) to tonalite (RB4). These lighter coloured samples will, in fact, be termed leuco-spessartite to distinguish them from sample RB6a.

4.3.2.3 Felsic series

Rocks of the felsic series comprise primarily of albite (~58 to 60 vol. %), quartz (~10 to 30 vol. %) and microcline feldspar (~8 to 10 vol. %). Minor to accessory minerals include $\pm$ Fe-hornblende (~7 vol. %), $\pm$ muscovite (~7 vol. %) and secondary kaolinite (~7 to 2 vol. %), clinocllore (~2 vol. %) and sericite.

Sample RB13, which was taken near the base contact of the felsic sill, contains a lower amount of quartz (~10 vol. %) and microcline (~8 vol. %), while Fe-hornblende (~7 vol. %) and muscovite (~7 vol. %) are minor constituents. Sample RB15, which was taken lower down (above in stratigraphy), where the highly weathered outcrops of felsic rocks are encountered (Figure 26), contains a greater amount of quartz (~30 vol. %) and microcline (~10 vol. %). Fe-hornblende and muscovite, however, were scarcely observable, and are not detected by XRD analysis.

Samples were classified based on their felsic mineral content (Table 6), applying the QAP scheme as illustrated in Figure 28. In contrast with the recommendations of the IUGS (Le Maitre, 2002), the albite endmember ($An_{0.5}$) of the plagioclase series was be normalised together with microcline as part of the the alkali feldspar component (A). This decision was made in order to synchronise with the classification methods applied by previous authors including Bisschoff (1969, 1972, 1999) and Elsenbroek (1991), as referenced by de Waal *et al.* (2008) and Graham *et al.* (2005). Sample RB13 and RB15 are therefore respectively classified as quartz alkali syenite and alkali granite.

Table 6: Normalised modal compositions (vol. % - see Appendix D3) of analysed dioritic and felsic rock samples.

Sample	Quartz	K + Na-feldspar	Plagioclase feldspar
RB5	10	0 (An<5)	90 (An>5)
RB4	28	0 (An<5)	72 (An>5)
RB13	12	88 (An<5)	0 (An>5)
RB15	31	69 (An<5)	0 (An>5)

(An<5) = Microcline + Albite; (An>5) oligoclase to andesine

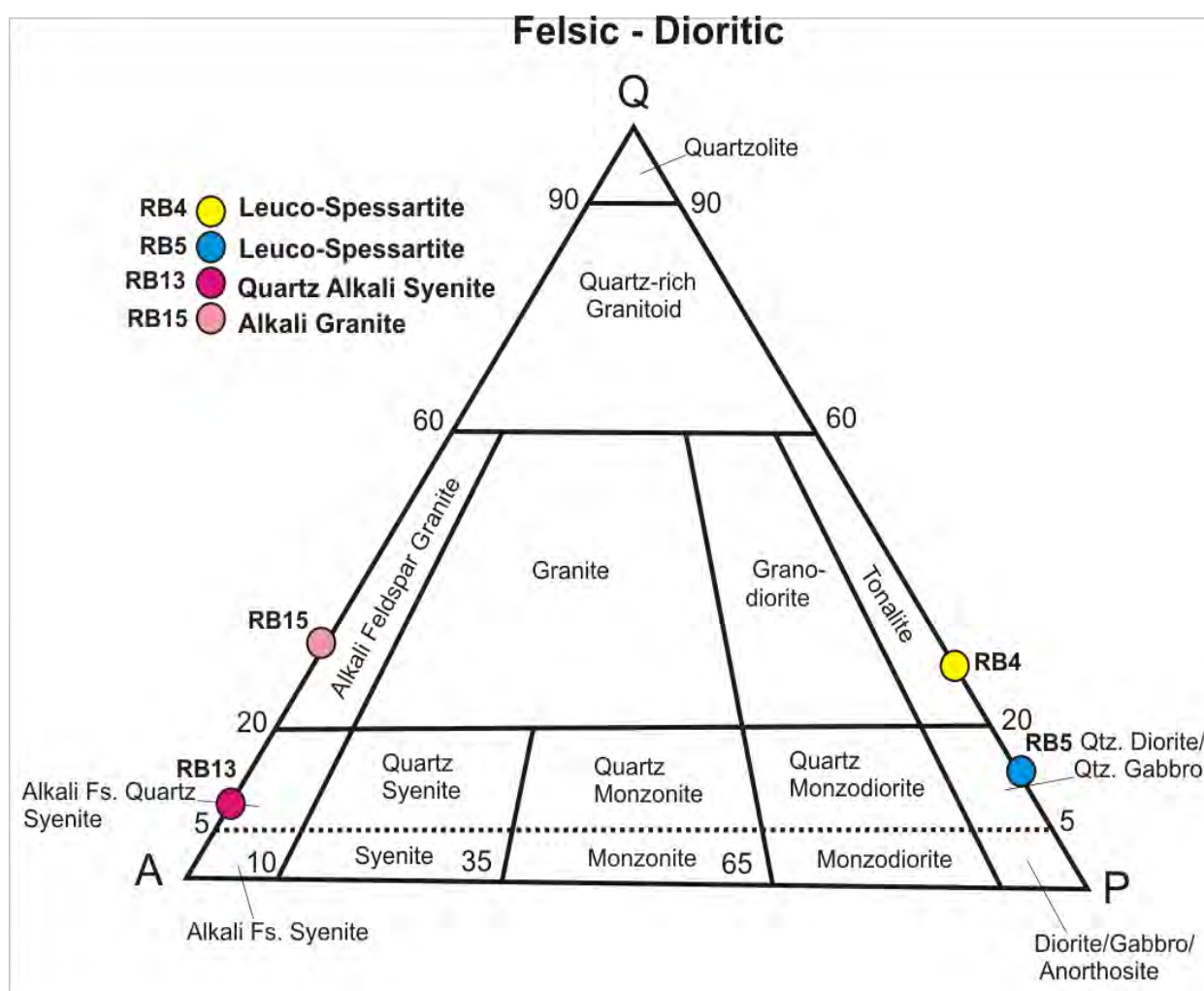


Figure 28: QAP modal classification of plutonic rocks. Adapted from Le Maitre (2002; Figure 2.4), after Streckeisen (1976; Figure 1a). Q = quartz, A = alkali feldspar, P = plagioclase. Alkali Fs = alkali feldspar; Qtz = quartz.

4.4 PETROGRAPHY

The petrographical descriptions of selected thin sections as catalogued in Appendix A2 will now be discussed according to relative age and name specified in section 4.3. Additional thin sections of samples that were not analysed with XRD, will also be referred to, on the basis of field and petrographic classification. This particularly includes an impact deformed sample of olivine clinopyroxenite (RB21) and a sample of non-uralitised clinopyroxenite (RB31). In accordance with the recommendations of the IUGS (SCMR- Fettes & Desmons, 2007) the collective term olivine $\pm$ clinopyroxenite is used when referring to both these modal varieties. The mineral abbreviations used in the supplementary photomicrographs are explained on page vii.

4.4.1 Epidiorite

The epidiorite (Figure 29) remains quite homogeneous throughout the sill and is mostly aphanitic hypidiomorphic granular with a microcrystalline texture composed of subhedral to euhedral grains of Mg-hornblende, and subhedral to anhedral grains of plagioclase and interstitial quartz.

The Mg-hornblende is secondary after augite and has a slender prismatic to acicular habit forming almost spherulitic clusters, which range between 50 to 300 μm in size (Figure 29). The clusters commonly penetrate, or divide the feldspar grains but rarely enclose them entirely. The hornblende grains are strongly pleochroic and vary from a pale yellow to green to blue-green colour. Due to the small size of the crystals, no cleavage planes can be detected, although, simple twinning is occasionally displayed, and occurs parallel to [001].

The plagioclase crystals are relatively the same size as the grains of hornblende but tend to be more skeletal and irregular. Some feldspar grains still retain their original lathlike shape, since the rock is only partially altered, and veinlets of subhedral plagioclase are also present, which is protruded by the newly formed Mg-hornblende (Figure 29). The plagioclase is colourless and intersertal, containing small fussy inclusions of epidote as a result of saussuritisation (Figure 30), which is the alteration of the original plagioclase to albite and epidote (Winter, 2010). The subhedral grains commonly display Carlsbad twins, while polysynthetic twinning is occasionally observed.

The quartz grains are interstitial and sometimes difficult to distinguish from the plagioclase by their interference figures, due to small grain size, and the distortion of grain boundaries as a result from saussuritisation (Figure 30). The quartz grains do, however, commonly display undulose extinction as a result of deformation, while this feature is less prominent within the plagioclase crystals.

Accessory magnetite occurs as small irregular to equant grains, typically in the center of Mg-hornblende grains (Figure 31). The magnetite grains seems to have crystallised initially and were enclosed by pyroxene before uralitisation occurred. At the contact boundaries of the magnetite, the hornblende typically appears dark brown, which is due to the adsorption and increase in iron content (Nesse, 2004).

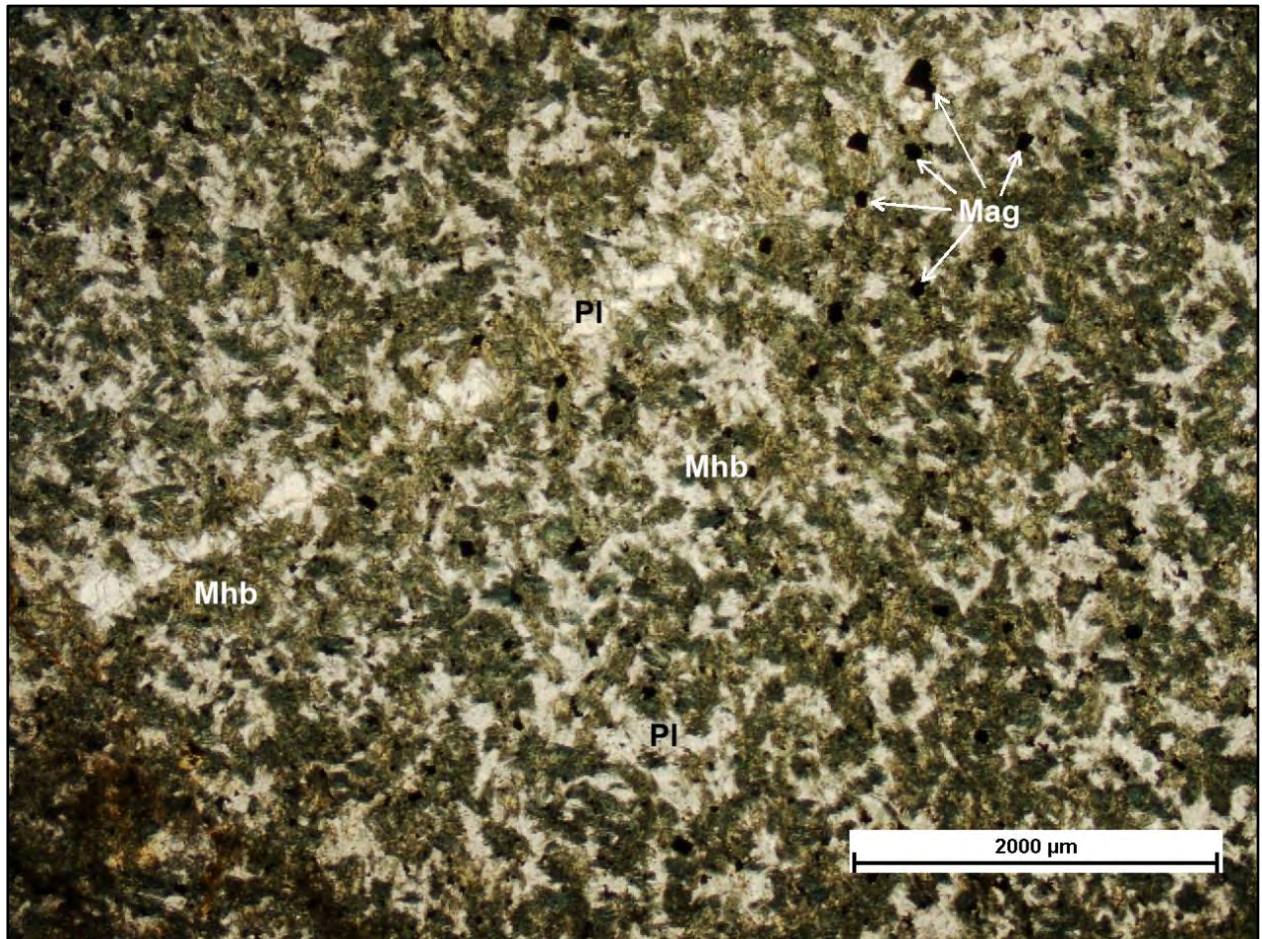


Figure 29: Photomicrograph of epidiorite (sample RB16) viewed under plane polarized light (PPL), illustrating the rock's microcrystalline structure that is composed of clusters of slender pale yellow to blue-green crystals of Mg-hornblende (uralite) and skeletal grains of colourless anhedral plagioclase. Note how the veinlet of plagioclase is protruded by the Mg-hornblende, which has completely recrystallised after pyroxene. Small magnetite grains are also commonly seen throughout the rock.

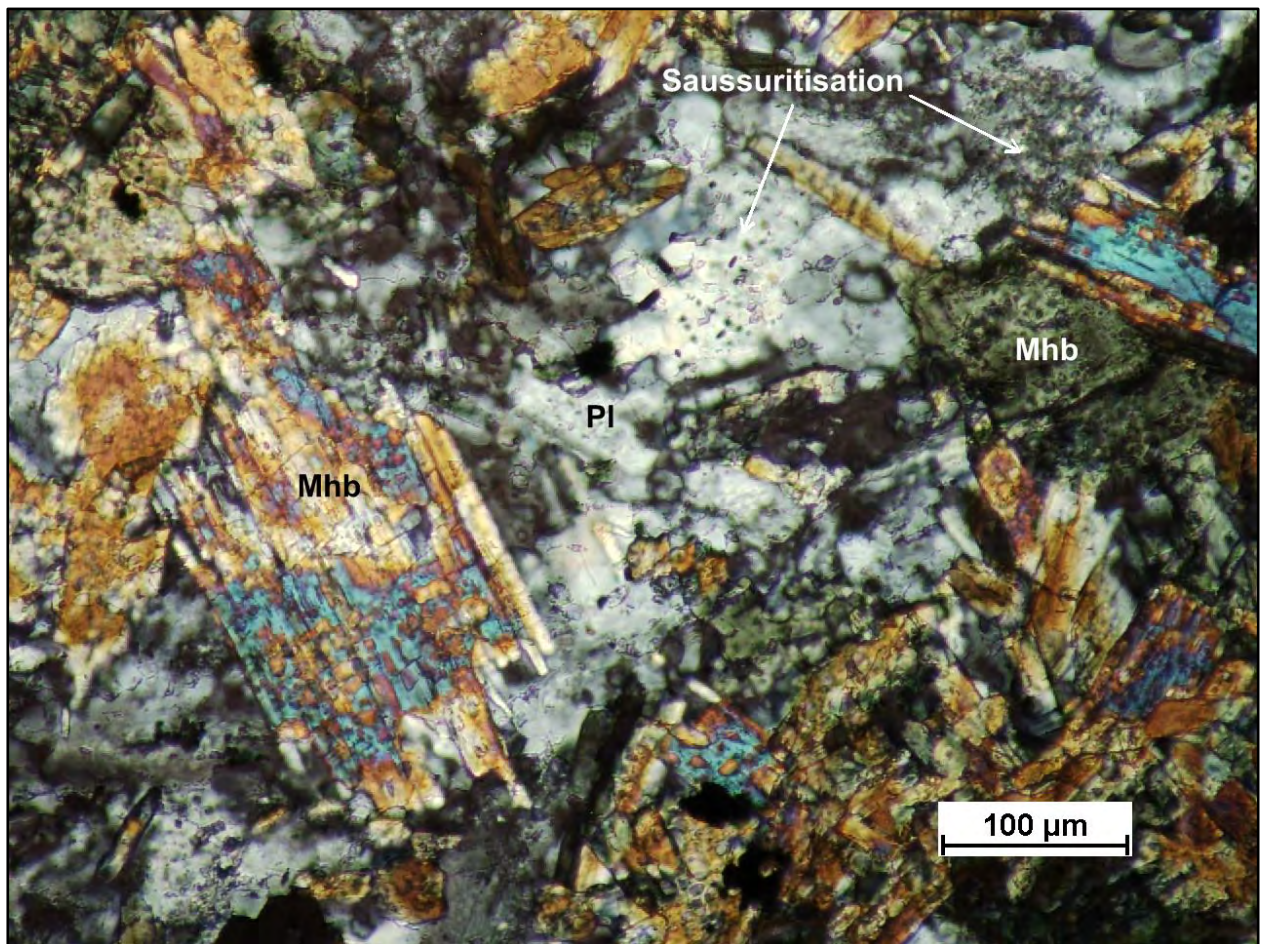


Figure 30: Photomicrograph of epidiorite (sample RB17) viewed under crossed polarized light (XPL), displaying the slender prismatic habit of the irregular green Mg-hornblende that penetrated the irregular grains of saussuritised plagioclase.

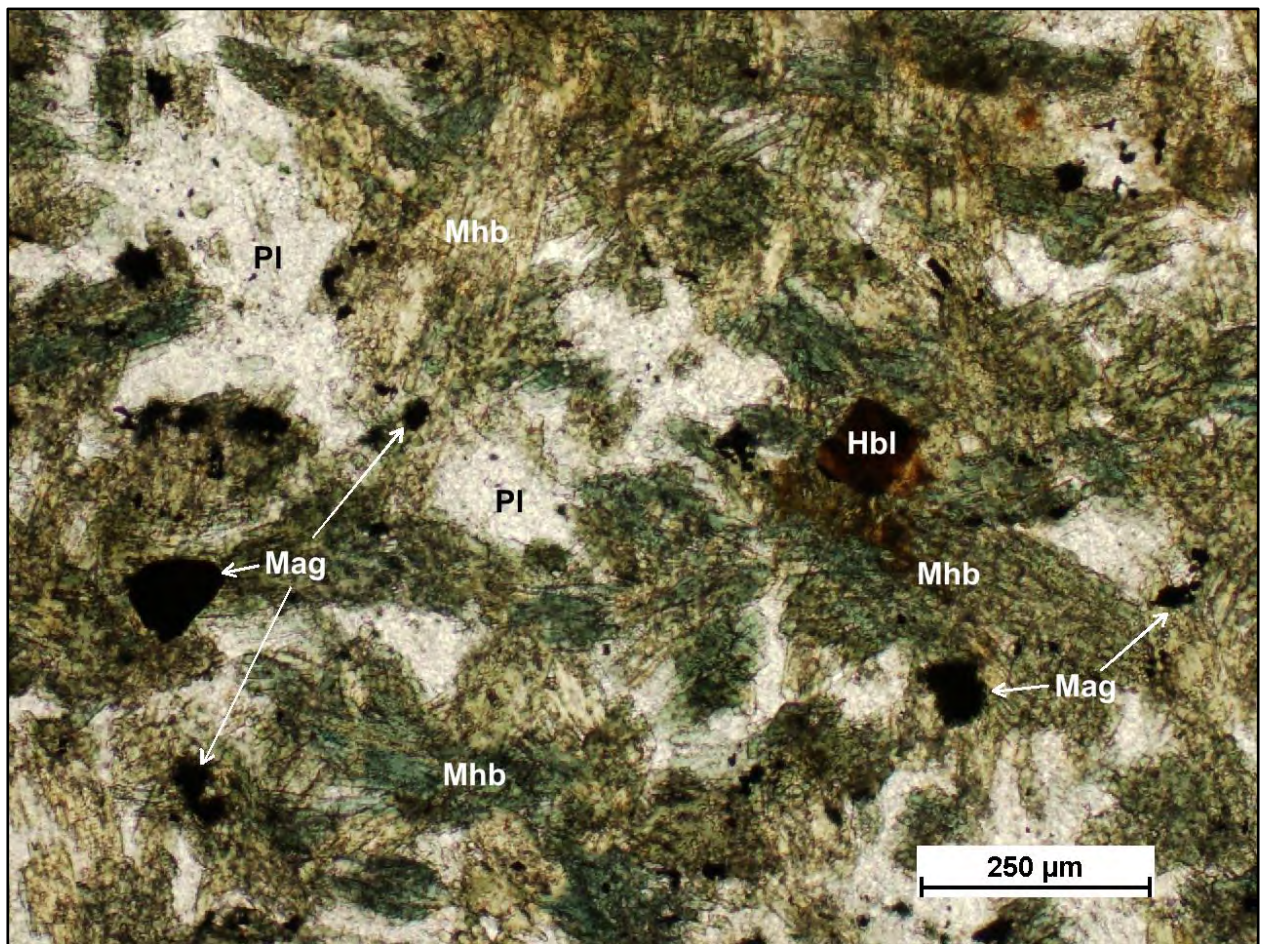


Figure 31: Photomicrograph of epidiorite (sample RB16) under PPL, illustrating the occurrence of subhedral, equant magnetite grains that are enclosed by yellow-green subhedral Mg-hornblende. Note how some of the magnetite grains have been completely replaced by brown hornblende.

4.4.2 Mela-dolerite

The contact sample of mela-dolerite is aphanitic inequigranular (Figure 32) and comprises of dispersed subophitic glomerocrysts of subhedral to anhedral augite and plagioclase (200 to 1000 μm), which are surrounded by finer masses of anhedral grains (50 to 500 μm) composed of augite, plagioclase, and hornblende.

The hornblende is visually most abundant and forms irregular, yellow-brown to dark brown pleochroic grains, which is frequently observed to replace augite. Visually, the degree of uralitisation is variable, as smaller augite grains have thicker coronal rings of hornblende, while larger grains are only slightly altered at their boundaries and fracture zones (Figure 32).

The augite is colourless to light grey and commonly displays herringbone textures (Figure 33), which in this sample consists of simple twinning on the (100) plane and thin exsolution lamellae

of enstatite or pigeonite on the (001) plane. The thin exsolution lamellae can be easily mistaken for lamellar twins, however, these are slightly inclined in reference to the crystal lattice of the host grain, which suggests that they are more likely lamellae. Orthopyroxene lamellae can be distinguished from pigeonite lamellae based on their extinction angles. Orthopyroxene lamellae are extinct parallel to the polarizers, while pigeonite lamellae extinguish at an oblique angle (Gill, 2013).

The plagioclase is colourless and forms irregular to rounded grains that are partially to completely enclosed by augite and hornblende (Figure 33). The plagioclase grains show signs of saussuritisation, while Carlsbad and albite twinning are regularly observed.

Accessory magnetite is fairly abundant in this rock and occur as intergranular, anhedral grains, strongly clustered along the contact margins of hornblende. In some cases the magnetite is also poikilitically enclosed by augite and hornblende (Figure 33).

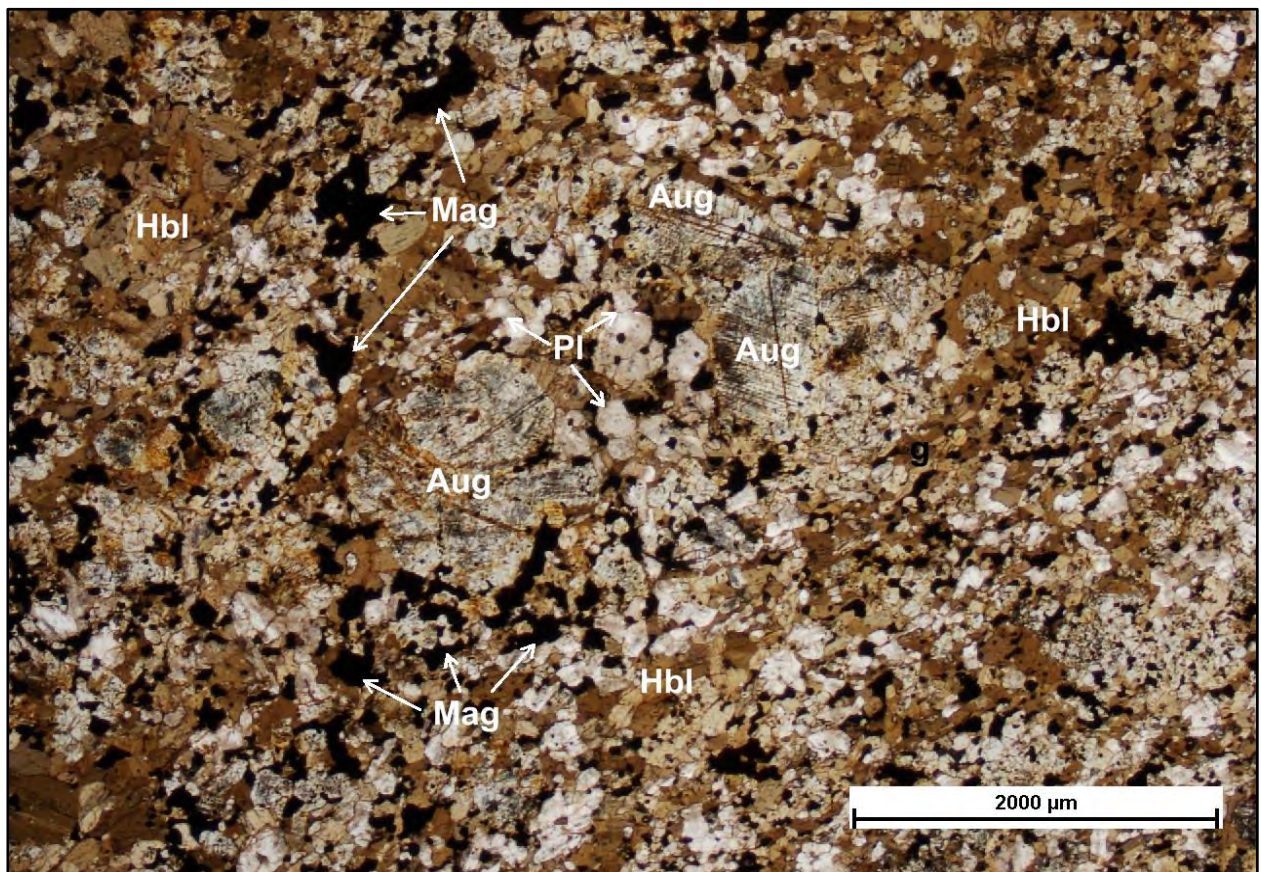


Figure 32: Photomicrograph of mela-dolerite (sample RB1A) under PPL, displaying the rock's seriate texture that comprises of coarser-grained glomerocrysts of light-grey augite and colourless plagioclase, which is surrounded by finer grains of anhedral yellow-brown hornblende, plagioclase, augite, and magnetite. Note how the hornblende forms brown coronas around the smaller augite grains, and how the larger grains of augite are replaced by hornblende along their fracture planes.

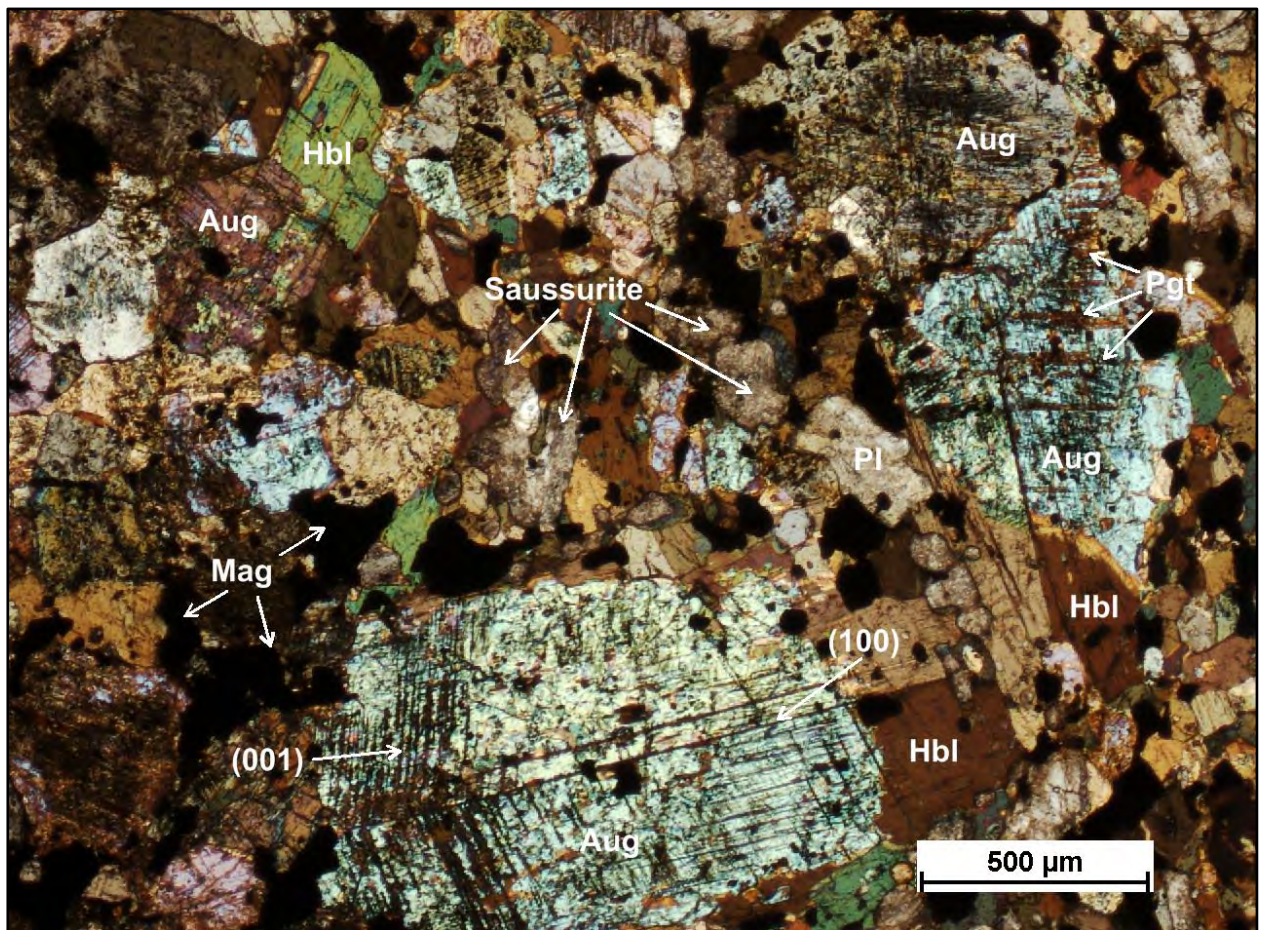


Figure 33: Photomicrograph of mela-dolerite (sample RB1A) under XPL, showing a coarser-grained glomerocryst of subhedral to anhedral augite and plagioclase with secondary saussurite, and anhedral hornblende after clinopyroxene. The larger augite grains display herringbone textures composed of simple twinning (100) and pigeonite/ enstatite lamellae (001). Note how the twinning is inherited by the hornblende, which partially replaces the augite grains pseudomorphically at the bottom right hand side.

4.4.3 Wehrlite and Olivine $\pm$ clinopyroxenite

The samples of wehrlite (Figure 34 & Figure 37), olivine clinopyroxenite (Figure 38 & Figure 39) and clinopyroxenite (Figure 40) are medium-grained hypidiomorphic- to allotriomorphic-granular, and comprise primarily of subhedral to anhedral grains of partially corroded, oval to irregularly shaped augite (500 to 3000 μm), and highly fractured anhedral olivine, which varies from abundant interconnected crystals (500 to 2000 μm), to interstitial and isolated accessory grains.

The ultramafic samples contain prominent veinlets that are indicative of alterations of the olivine and augite grains, namely serpentinisation and uralitisation. Related saussuritisation of accessory interstitial plagioclase is also common, which occurs mostly in the upper clinopyroxenite.

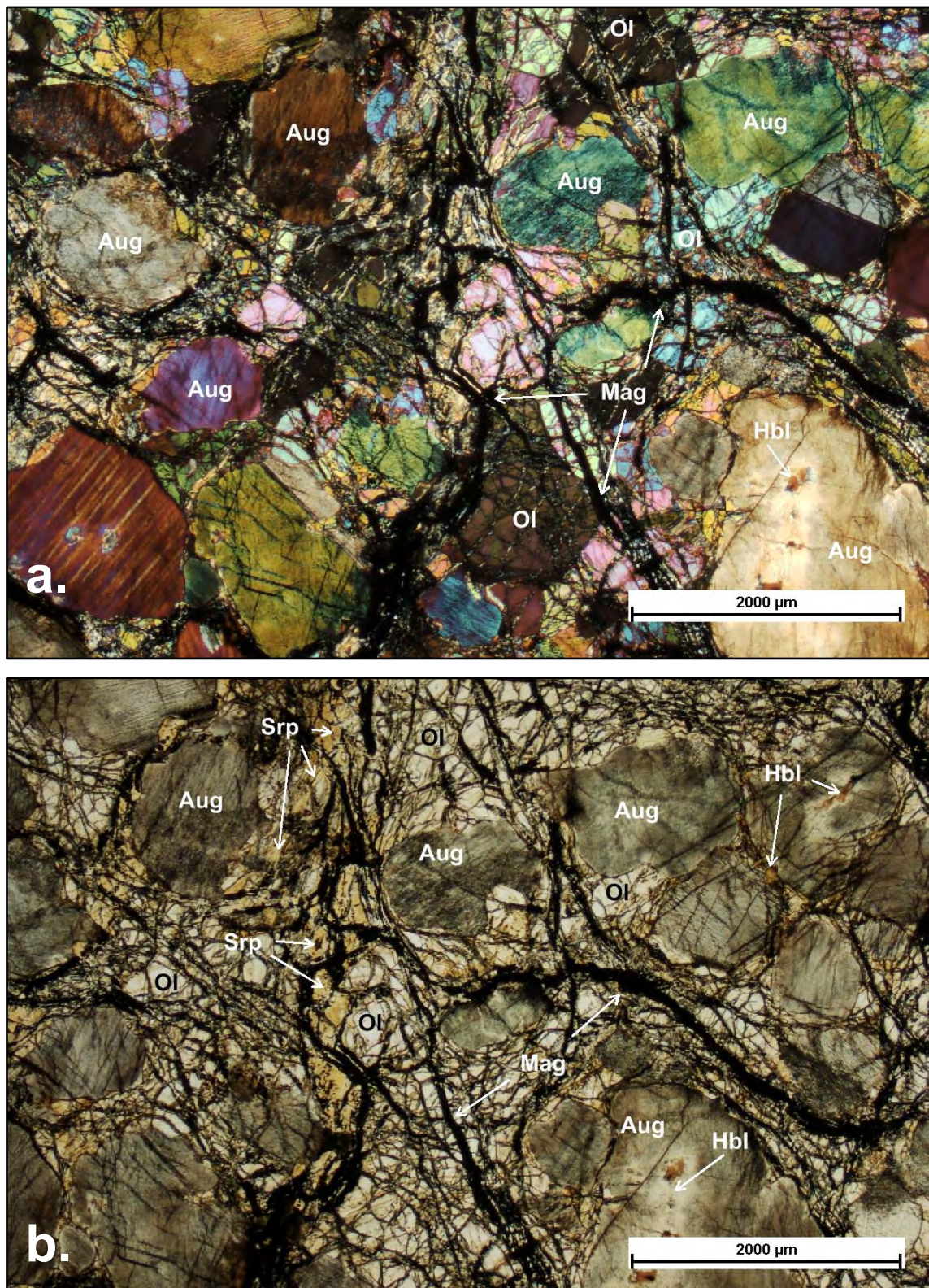


Figure 34: Photomicrographs of wehrlite (sample RB7) under (a) PPL, displaying the hypidiomorphic-granular texture of the rock that comprises of light to dark grey subhedral to anhedral augite, and fractured colourless anhedral olivine, and (b) XPL, illustrating the difference in birefringence of the third order in olivine and the second order in augite. Note in both photomicrographs the presence of dark symplectic veinlets that are composed of opaque magnetite and yellow lizardite that displays a low first order birefringence.

The olivine is forsteritic, typically colourless, and displays a mesh texture (Figure 35), which is the result of serpentinisation (Best, 2003). The grains are partly altered to yellow serpentine, specifically lizardite, and accompanied by the release of very fine grained opaque magnetite crystals that form dark symplectic veinlets. These veinlets are mostly concentrated within olivine, although some run through augite as well. As with typical serpentinisation patterns, lizardite occurs along fractures in the olivine where it must have been infiltrated by late-stage magmatic fluids which caused the alteration.

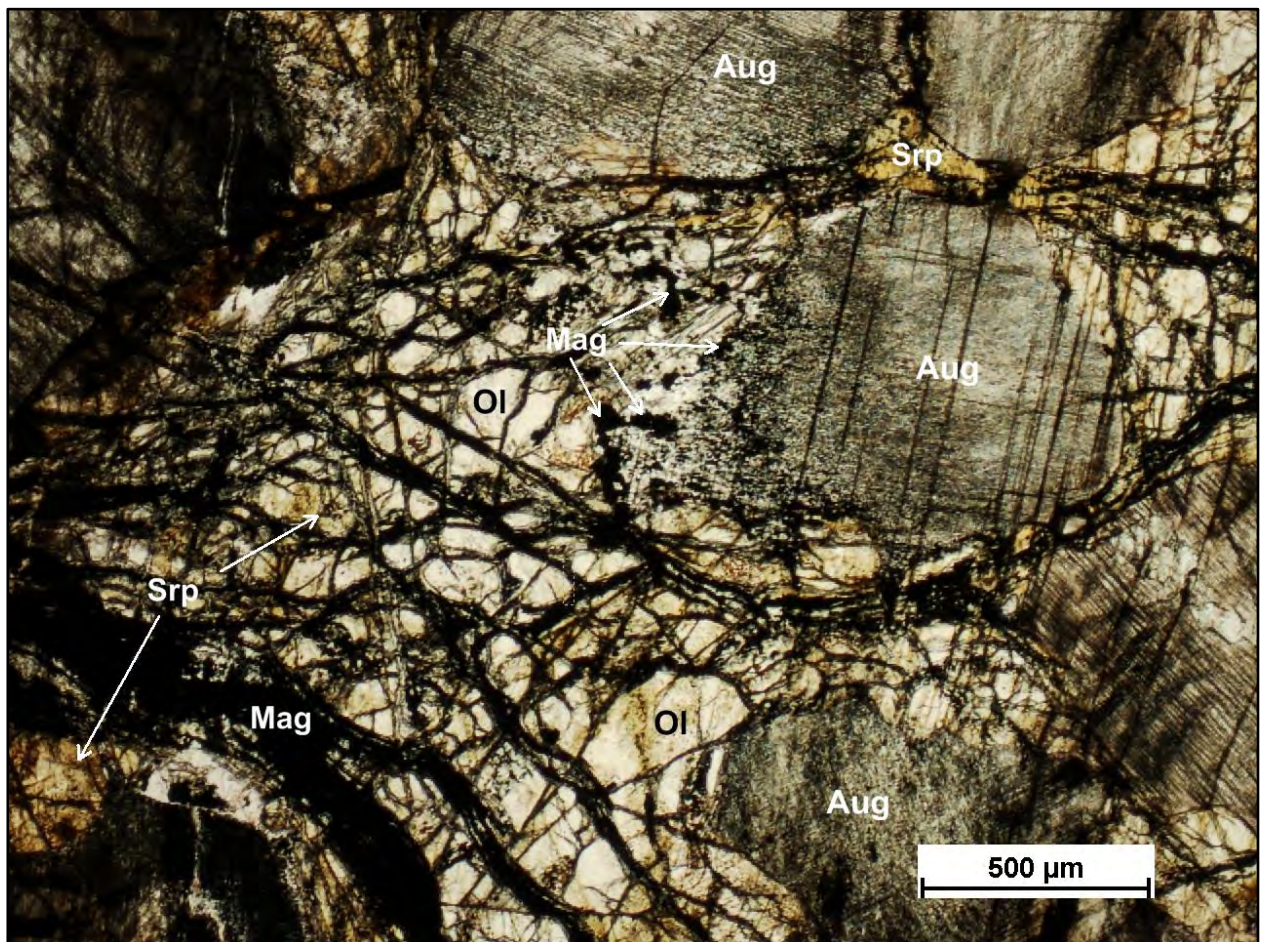


Figure 35: Photomicrograph of wehrlite (sample RB7) under PPL, illustrating the typical mesh texture of fractured olivine grains that resulted from serpentinisation, as can be seen by the protruding yellow and opaque veinlets of lizardite and magnetite respectively, which are the products of this alteration. Note how the grey colour of the partially assimilated augite grain is darkened from alteration. This is ascribed to crystallization of secondary needle-like intergrowths.

The augite grains generally show a lesser extent of physical deformation and is partly serpentinised pseudomorphically, producing the particular texture known as bastite (Best, 2003). In portions where dark symplectic veinlets penetrate augite or touch a boundary, patchy irregular grains of light to dark brown hornblende occurs as a secondary reaction product (Figure 36).

The particular grey to dark grey colour of augite in the wehrlites can be ascribed to orientated secondary intergrowths that compose of very fine needlelike inclusions of opaque minerals (probably Fe-, Cr- and Ti-oxides such as ilmenite or Cr-spinel). These seem to have formed as a by-product during the alteration process that resulted in exsolution of these opaque crystals along the lattice of the host grain. The intergrowths are particularly concentrated where the rim of the grain has reacted to form serpentine (Figure 35) or where veinlets cut into the grain and have been altered to hornblende (Figure 36).

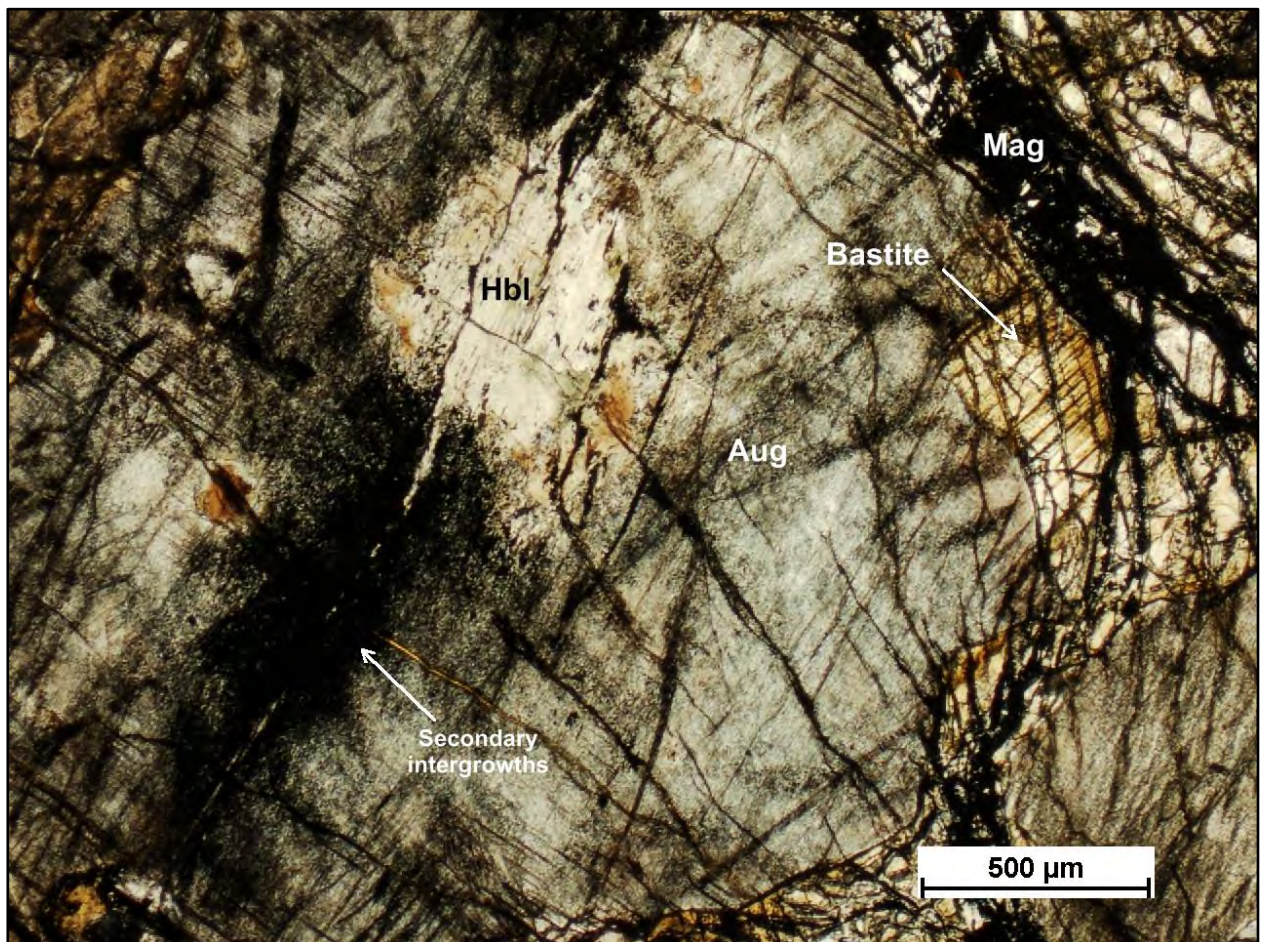


Figure 36: Photomicrograph of wehrlite sample RB7 (PPL), displaying an altered subhedral grain of light to dark-grey augite. On the right-hand border of the grain, the typical bastite texture can be observed where the grain is replaced pseudomorphically by yellow lizardite. In the left-hand corner, a thin veinlet of light yellow-brown hornblende protrudes the grain and expands towards the centre where it forms an anhedral crystal. Note how the augite gets progressively darker towards the vein as a result of secondary needle-like intergrowths of oxide minerals, possibly ilmenite or Cr-spinel.

The lower portion of wehrlite (RB7 & RB8) that outcrops at the north-western sill (west of F₃, Figure 9), displays a cumulate texture (Figure 34), where both the olivine and augite are more or

less equally distributed, and some grains poikilitically enclose one another indicating that they have crystallized simultaneously, as Bisschoff (1973) suggests.

The olivine grains tend to form interconnected frameworks that partly or fully surround augite (Bisschoff's (1969) "sponge texture"), and since the olivine grains taper or "pinch" around the boundaries of the augite, it indicates that olivine had ceased crystallising before augite, which allowed olivine to be partly shaped by the expanding grains of pyroxene. This texture is particularly more noticeable in the upper wehrlite and olivine clinopyroxenite outcrops of the southern portion of the sill (east of F_3), which contains less olivine (RB9 & RB10). In these rocks, the texture can be described as ortho- to mesocumulate because the olivine grains becomes interstitial to isolated, as the grain boundaries of augite become more connected, some even forming triple point boundaries (Figure 34) that are indicative of Ostwald ripening (Winter, 2010).

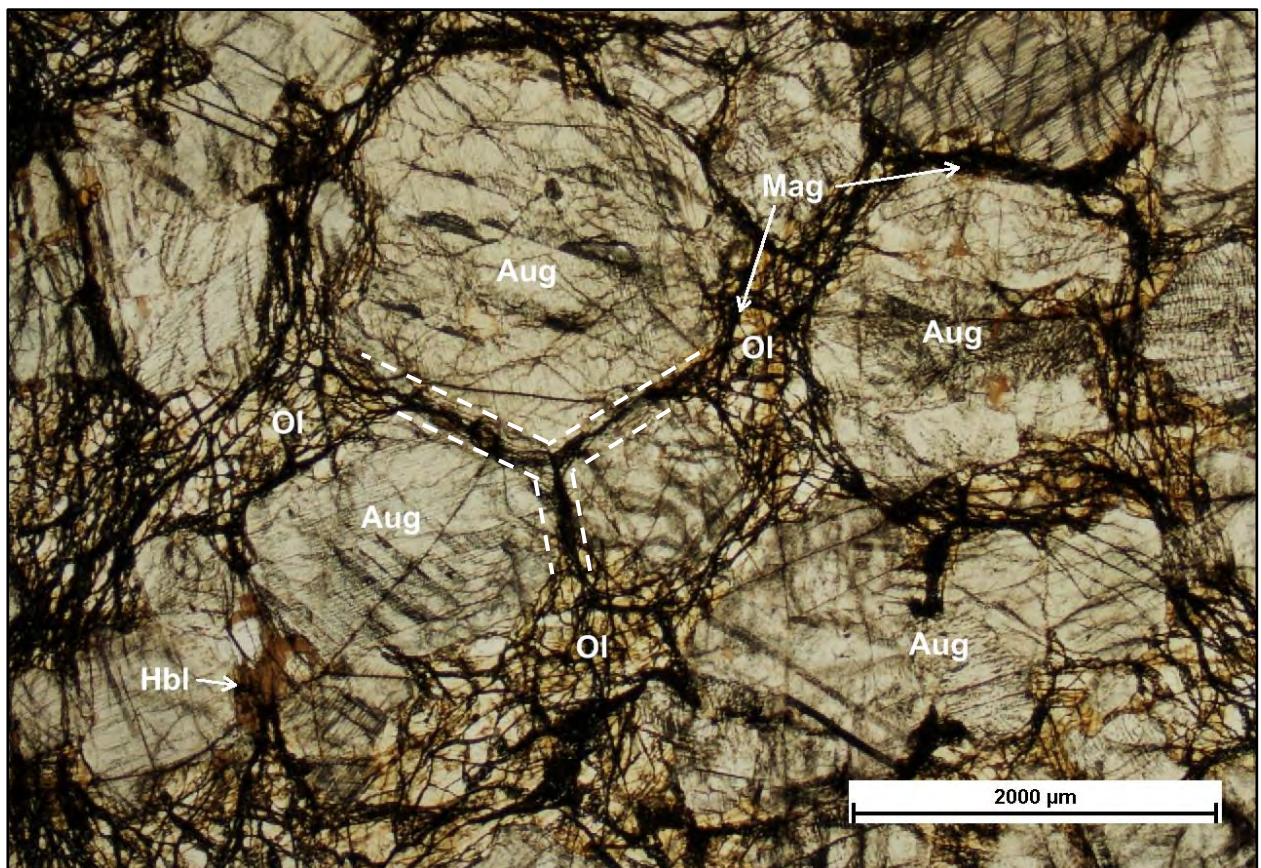


Figure 37: Photomicrograph of wehrlite (sample RB9) under PPL, displaying the interstitial character of partially serpentinised olivine, which forms an interconnected framework between subhedral to anhedral grains of augite. Secondary brown hornblende can be seen where opaque magnetite veinlets penetrate augite grains, while augite itself appears more colourless to light grey. Note how the boundaries of augite grains display 120° intercepts that are indicative of Ostwald ripening.

The olivine clinopyroxenite (Figure 38) is allotriomorphic granular, as augite appears more sutured along its grain boundaries. Olivine in this sample (RB10) is scattered and isolated, forming irregular anhedral grains that are partly serpentised. The alteration of olivine progresses towards brown-red iddingsite, which consists of a mixture of hydrated silicates and Fe and Mg oxides, while augite is primarily altered along thicker veinlets of secondary colourless to light brown hornblende, which runs across the sample.

The sample from the impact deformed olivine clinopyroxenite (RB21), which occurs north-west of the main exposed outcrop, displays a somewhat different texture (Figure 39). Olivine is similarly patchy and isolated, while the anhedral augite is particularly more rounded, and also partly converted to brown hornblende. Together, these minerals form fractured or brecciated portions that are separated by prominent dark veinlets of pseudotachylite that contain small broken pieces of olivine and augite.

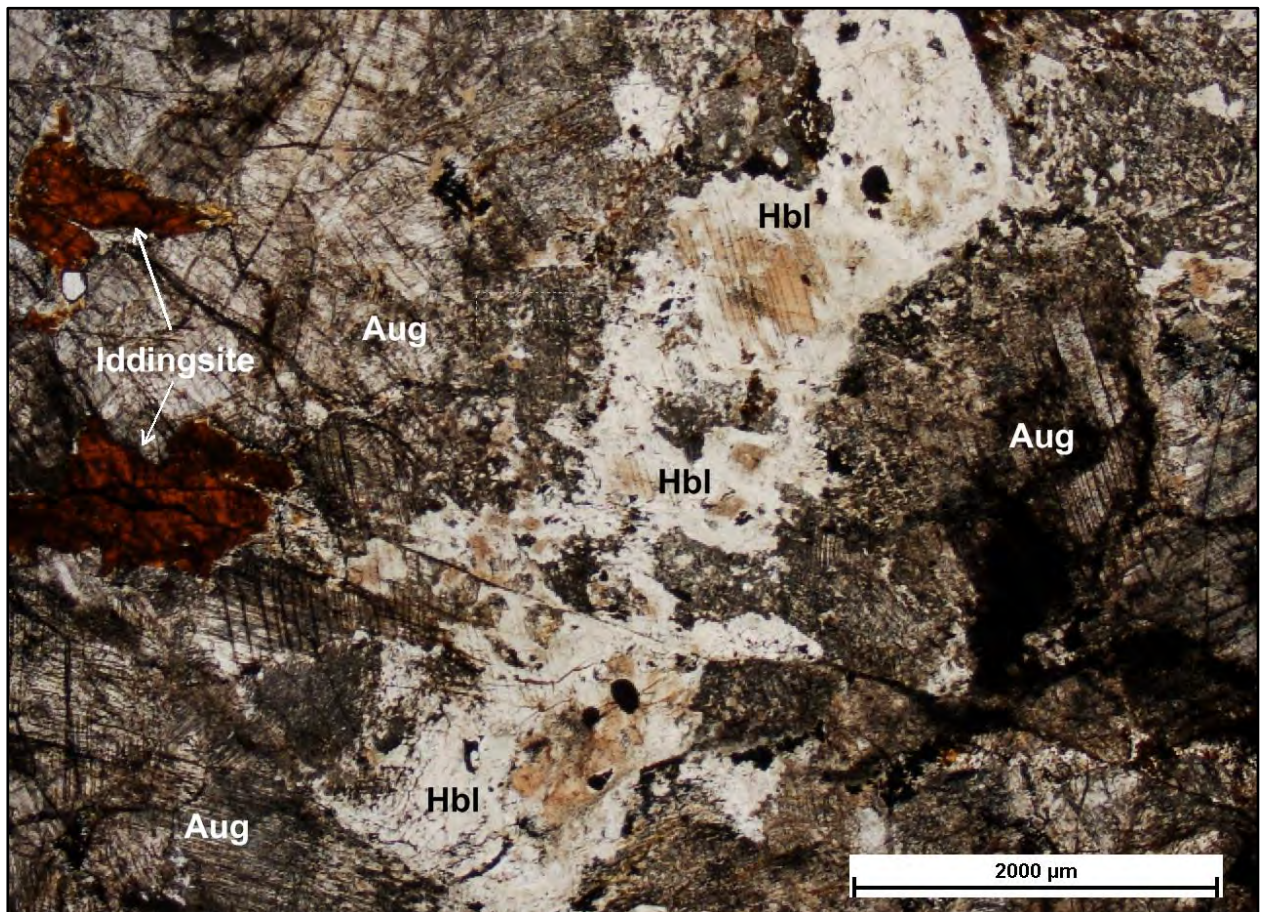


Figure 38: Photomicrograph of olivine clinopyroxenite (sample RB10) under PPL, displaying the serrated boundaries of the anhedral augite grains which are partly corroded where they border a vein of colourless to light brown secondary hornblende. Note how the isolated grains of olivine are fully altered to brown-red iddingsite.

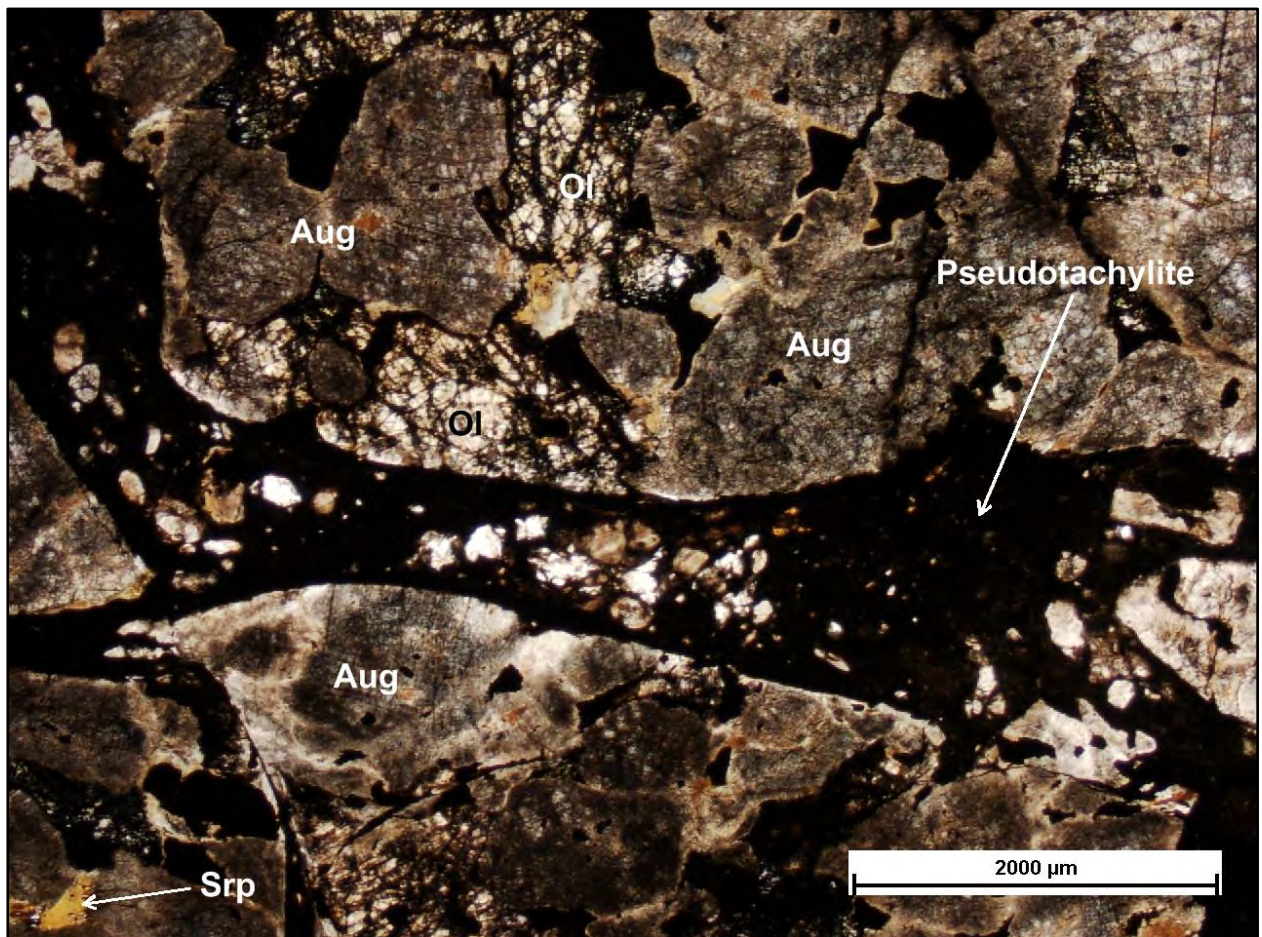


Figure 39: Photomicrograph of olivine clinopyroxenite (sample RB21) under PPL, illustrating the brecciated nature of the impact deformed sample, consisting of fractured allotriomorphic granular portions of augite and interstitial olivine, separated by dark veinlets of pseudotachylite. Note the partially rounded habit of the augite, as well as the small mineral clasts that are included in the pseudotachylite.

The sample of clinopyroxenite (RB31) that was found north of the main wehrlite body (Figure 9) progressed towards an adcumulate texture (Figure 40). This rock is composed primarily of larger (1000 to 3000 μm) anhedral grains of augite, which seems particularly fractured, but not brecciated such as in sample RB21 (Figure 39). Olivine is entirely absent, with orientated, euhedral apatite crystals filling the interstitial spaces. The apatite is poikilitically enclosed by a mass of crypto crystalline crystals, which are probably saussuritised plagioclase. Signs of serpentinisation and uralitisation, however, are generally scarce since the augite is predominantly unaltered, apart from several fractures.

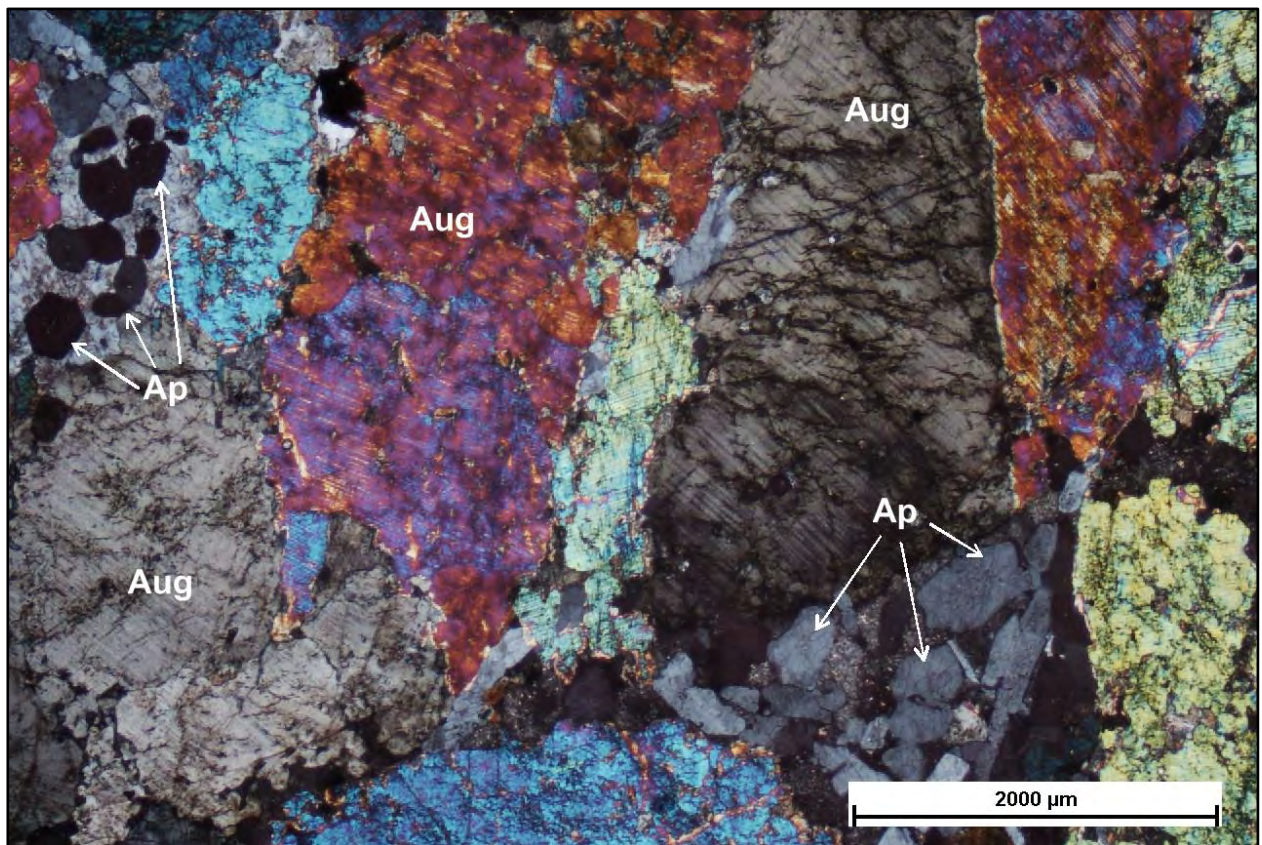


Figure 40: Photomicrograph of clinopyroxenite (sample RB31) under XPL, illustrating the accumulative texture that comprises of anhedral elongated, highly fractured augite with interstitial apatite, which seems to be poikilitically enclosed by a fine mass of saussurite. Note how the euhedral clusters of apatite are orientated crystallographically, with the right hand cluster being viewed perpendicular to [001] and the left-hand cluster parallel to [001].

4.4.4 Uralitite

The sample of uralitite (RB14) is mostly monomineralic, possessing a micro-porphyroblastic texture (Figure 41a & b) that is composed of relatively larger (500 to 2000 μm) porphyroblasts of hypidioblastic to xenoblastic Mg-hornblende (pseudomorphic after augite) located in a cryptocrystalline groundmass (<10 μm) of fibrous crystals of the same composition as Mg-hornblende, which is referred to as uralite. The porphyroblasts are thus remnants of augite grains which have been partly recrystallised when uralitisation occurred.

Mg-hornblende varies from almost colourless to pale yellow to light green. Twinning and prominent exsolution textures, apart from minor magnetite, is generally absent, yet the porphyroblasts display the typical 56° and 124° intersection of cleavage planes that is characteristic of amphiboles (Nesse, 2004).

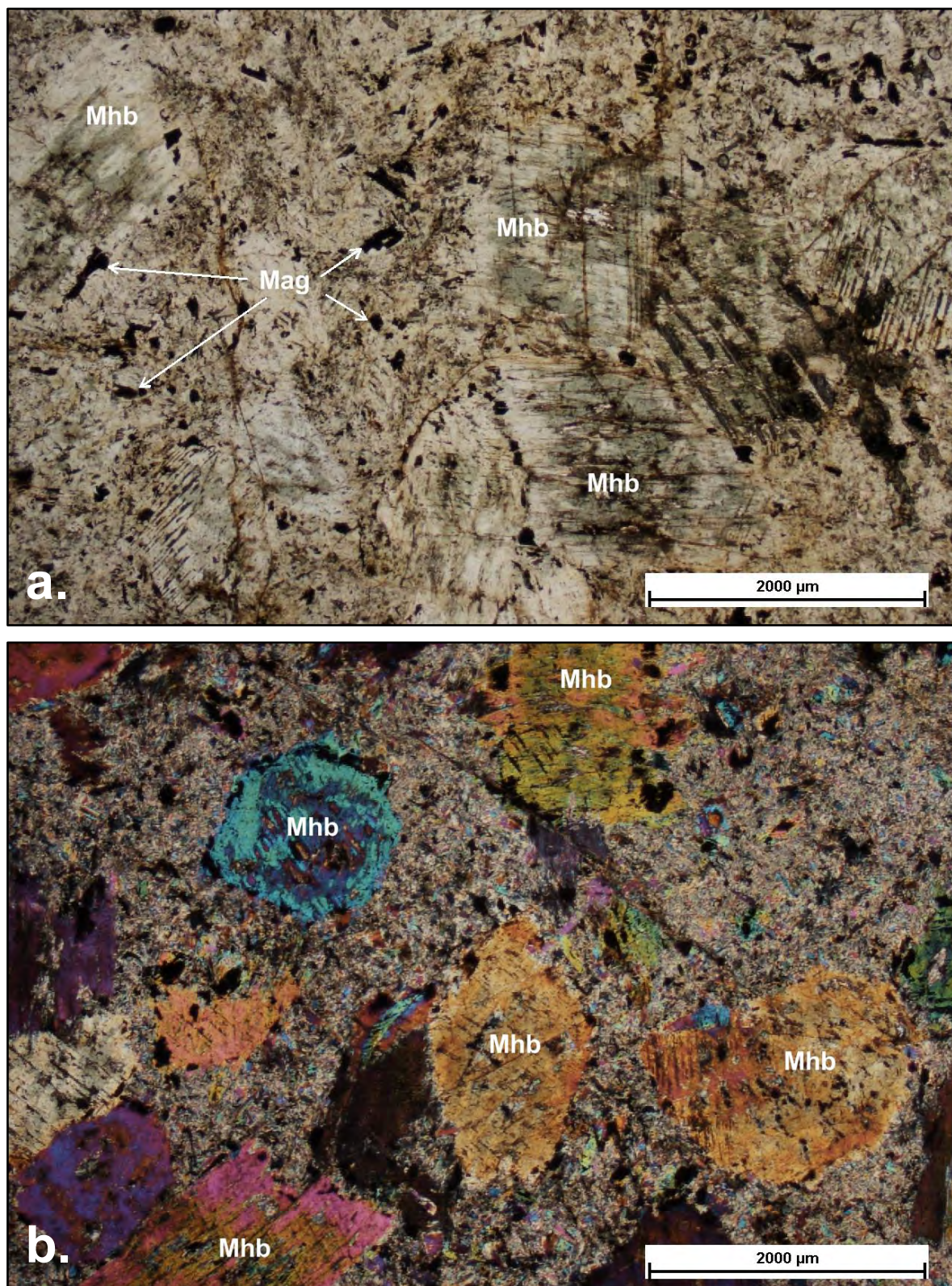


Figure 41: Photomicrographs of uraltite (sample RB14) under (a) PPL, displaying the rock's micro porphyroblastic texture that consists of colourless to light green Mg-hornblende forming larger xenoblastic to hypidioblastic porphyroblasts, which are set in a groundmass of fibrous uraltite (Mg-hornblende), and (b) the same plate under XPL, illustrating the first to second order interference colours, as well as zoning of the grains. Note in both images how the larger grains of accessory opaque magnetite are included along the edges of the porphyroblasts.

Xenoblastic magnetite is accessory and particularly associated with the porphyroblasts of Mg-hornblende, which enclose them poikiloblastically. The larger grains of magnetite (20 to 50 μm) can commonly be seen included along with the edge of the grain boundaries (Figure 41a & b).

4.4.5 Orthopyroxene Hornblende Gabbro

The sample of orthopyroxene hornblende gabbro (RB3) is medium- to fine-grained allotriomorphic-granular (Figure 42) and is composed of long irregular anhedral grains (100 to 1500 μm) of plagioclase, augite and accessory orthopyroxene, which is extensively altered to light brown hornblende.

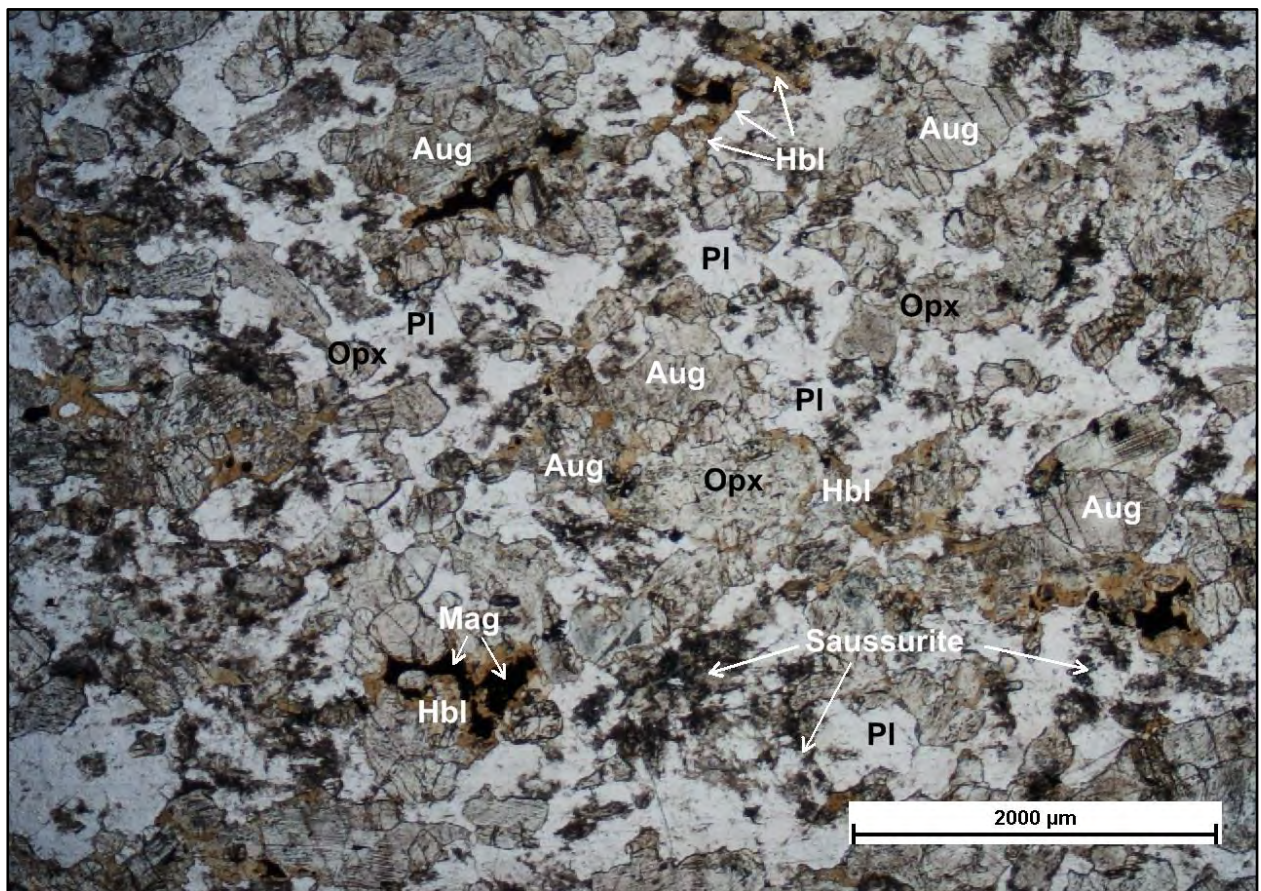


Figure 42: Photomicrograph of orthopyroxene hornblende gabbro (sample RB3) under PPL, displaying the rock's medium- to fine-grained allotriomorphic texture, which comprises of irregular elongated grains of plagioclase and augite with accessory orthopyroxene that have been extensively altered to brown hornblende. Uralitisation is particularly prominent where pyroxene encloses small magnetite grains, as well as plagioclase, appears partly discoloured due to saussuritisation.

Plagioclase generally forms the largest grains that are elongated, colourless and typically display Carlsbad and albite twinning. Dark grey, cloudy patches appear in the centre of the grains that are the result from severe saussuritisation. The plagioclase regularly encloses small

orthopyroxene grains, however, not the larger augite. This suggests that orthopyroxene was initially formed followed by the crystallisation of augite and plagioclase.

Augite grains can be distinguished from orthopyroxene, by their larger elongated habit and second order interference colours (Figure 43), whereas orthopyroxene is typically small and irregular to rounded while displaying a lower first order interference colour. Simple twinning is occasionally displayed by the augite, as well as lamellar exsolution lamellae. These features are not, however, readily seen for orthopyroxene, which is distinctly more fractured.

Hornblende occurs as small irregular patches surrounding the edges of the pyroxenes. The grains are strongly pleochroic varying from colourless to light brown and is particularly prominent where the pyroxenes have poikilitic inclusions of magnetite. No twinning or exsolution lamellae is displayed.

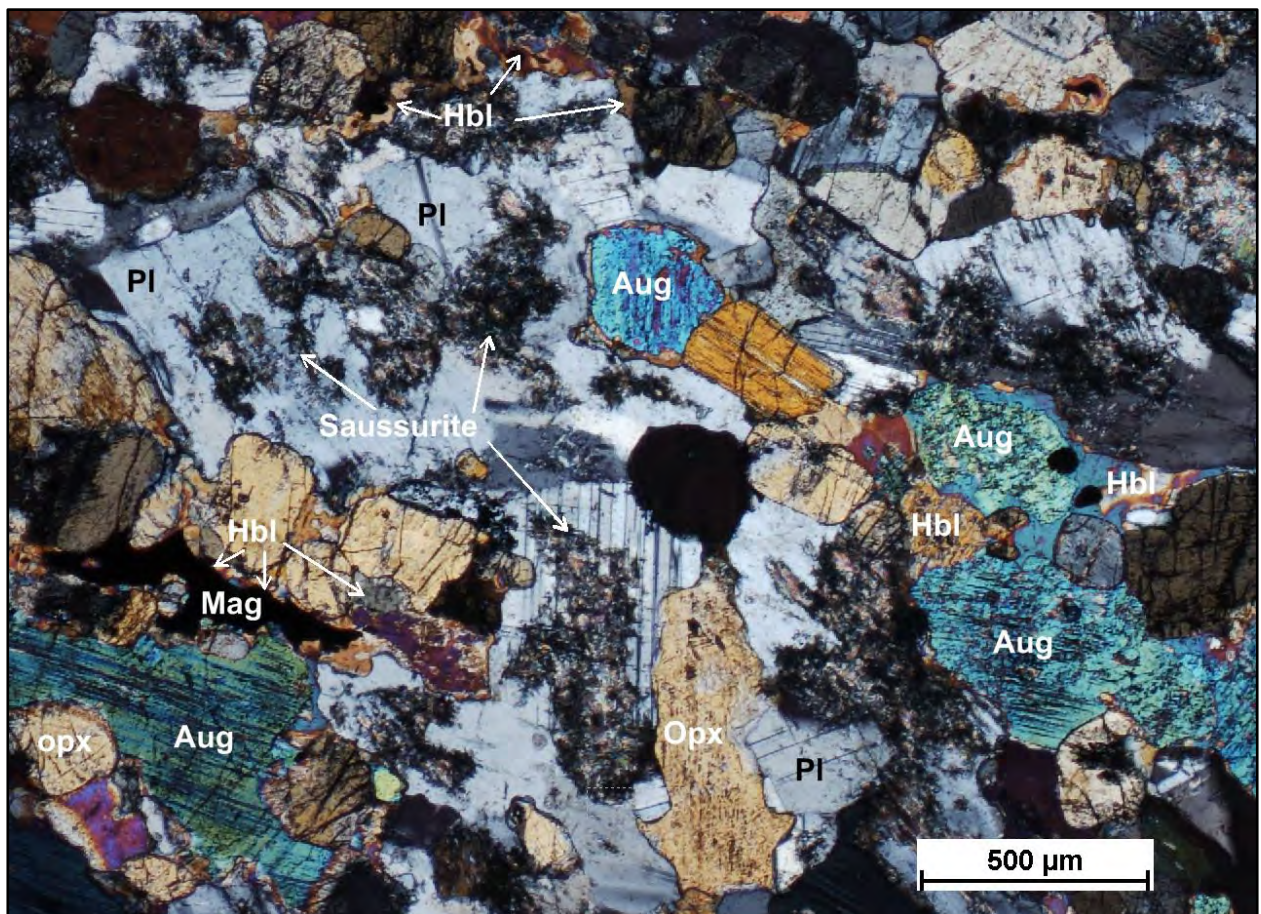


Figure 43: Photomicrograph of orthopyroxene hornblende gabbro (sample RB3) under XPL, illustrating the elongated and irregular habit of plagioclase and augite grains, as well as the partly rounded nature of accessory orthopyroxene. Irregular brown patches of hornblende can be seen partly replacing pyroxene grains, as well as plagioclase grains, which have been extensively saussuritised and contain small inclusions of epidote.

4.4.6 Hornblende & Mela-spessartite

The hornblende (RB6B) and mela-spessartite (RB6A) are principally composed of the same characteristic mineral assemblage, however, they vary quite broadly in terms of their texture and modal composition. The sample of hornblende is medium-grained hypidiomorphic-granular (Figure 44) and comprises primarily of subhedral to anhedral brown hornblende with green rims of Mg-hornblende (1000 to 3000 μm), as well as smaller grains of interstitial plagioclase (50 to 1500 μm). The mela-spessartite, in contrast, is hial porphyritic (Figure 45) and comprises of isolated grains of euhedral to subhedral hornblende that are comparable in size to that of the hornblende (1000 to 3000 μm), but are set in a considerably finer groundmass containing larger amounts of plagioclase and dark green Mg-hornblende (100 to 250 μm). Sporadic remnant augite grains, in addition to irregularly shaped magnetite, are also observed in both rock types.

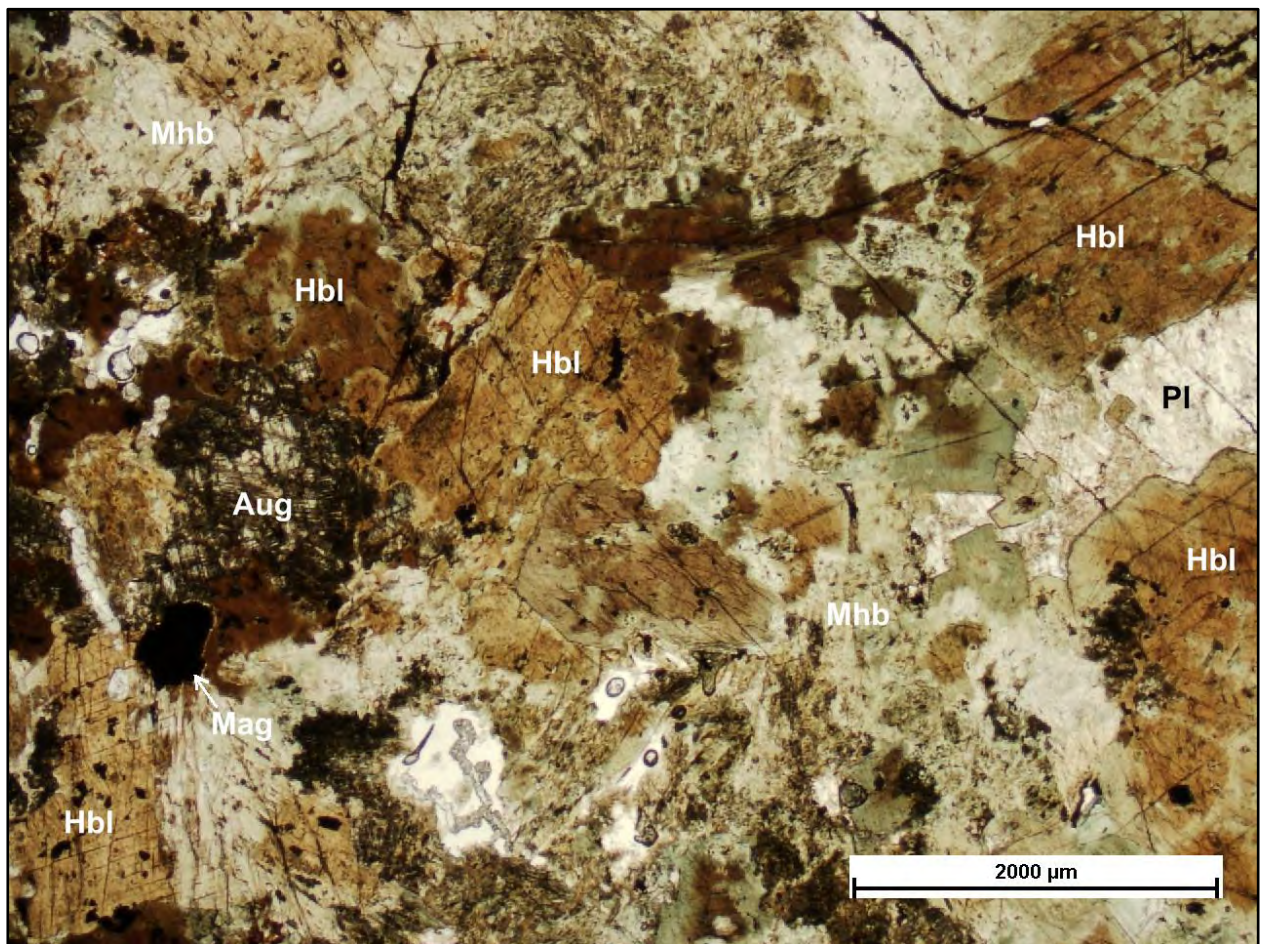


Figure 44: Photomicrograph of hornblende (sample RB6B) under PPL, displaying a hypidiomorphic-granular texture that comprises of subhedral to anhedral grains of brown to green zoned hornblende and interstitial plagioclase, which appears generally colourless and irregular. Note the isolated cluster of dark grey augite grains, which have been partly replaced by an outer rim of hornblende. Some isolated magnetite can also be observed, which is also enclosed by brown hornblende.

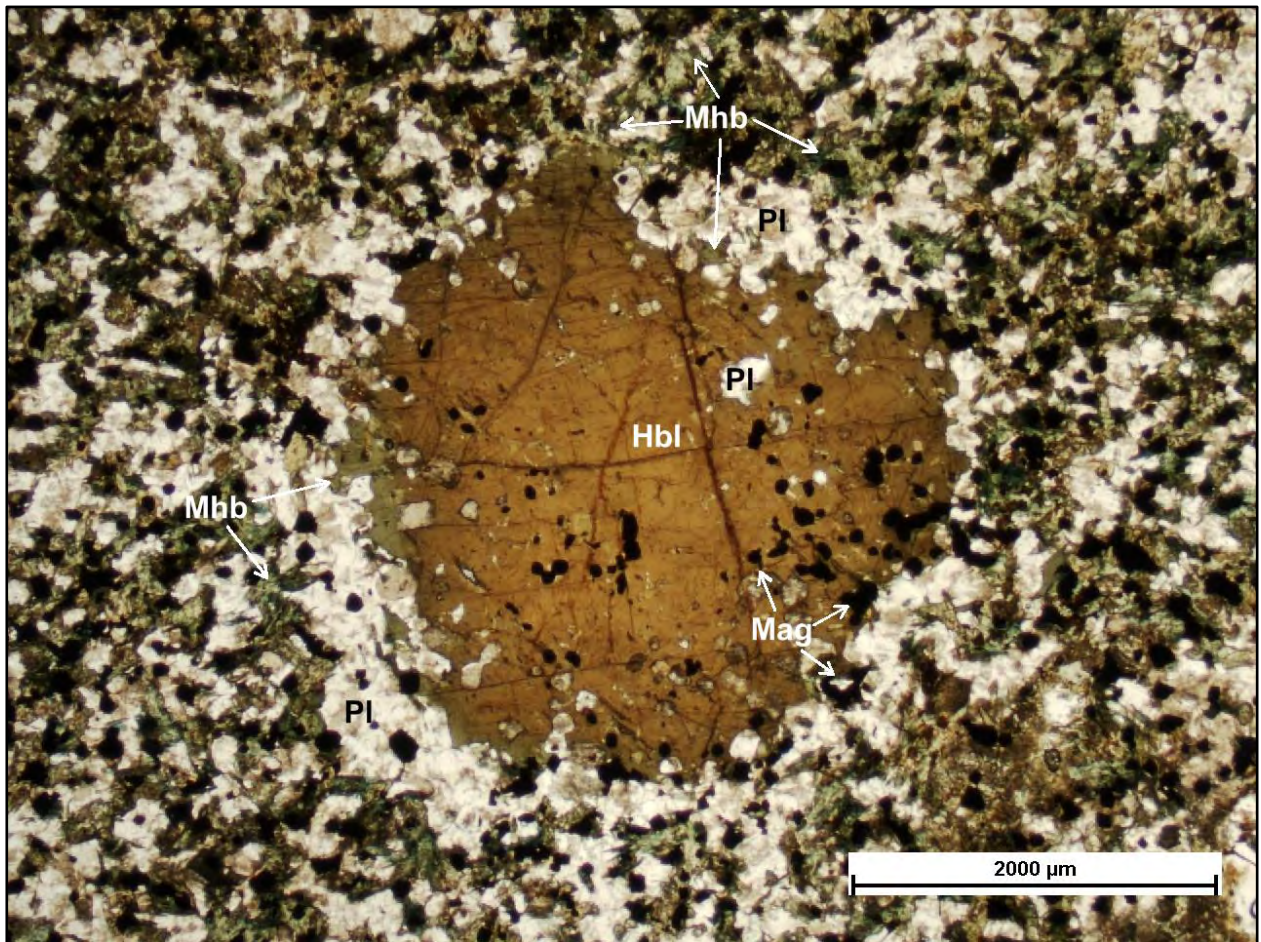


Figure 45: Photomicrograph of mela-spessartite (sample RB6A) under PPL, displaying the hial porphyritic texture of the rock that comprises of larger subhedral to euhedral phenocrysts of brown to green hornblende, occurring within a fine groundmass of rounded to blocky plagioclase grains and slender dark green crystals of Mg-hornblende. The phenocrysts can be seen enclosing plagioclase and magnetite grains, which also tend to cluster around the edges of the hornblende grain, forming partial coronas.

Similar as in the ultramafic and mafic rock types, hornblende is secondary after clinopyroxene and is observed to partly replace the augite, as well as penetrate or enclose smaller grains of plagioclase and magnetite. In the hornblendite and mela-spessartite, however, the grains are much larger and well formed, and in some samples display orthocumulate textures.

Hornblende grains are typically equant to irregular with light to dark brown cores, which grade toward green serrated boundaries of Mg-hornblende. The hornblende displays characteristic 56° and 124° cleavage intersections on $\{110\}$, but twinning is seldom developed. In the groundmass of the mela-spessartite, Mg-hornblende occurs as slender prismatic crystals, but not as small or fibrous as in the epidiorite (RB16 & RB17) or uraltite (RB14).

Plagioclase varies from andesine to oligoclase and generally occurs as colourless grains that appear enclosed between the larger expanded grains of hornblende. In the hornblendite, plagioclase grains are coarser and irregular, while in the mela-spessartite plagioclase occurs as small blocky to rounded grains that partially mantle the hornblende. Saussuritisation textures are less common and twinning is rarely seen amongst the small crystals occurring within the interstitial spaces of the hornblendite, as well as in the ground mass of the mela-spessartite.

Partly assimilated augite grains are seldom encountered in these rock types. In the hornblendite, augite occurs as random isolated clusters of smaller grey augite grains which are mantled by brown hornblende (Figure 45), and in the mela-spessartite, they occur as small poikilitic inclusions that are typically in the centre of the hornblende crystals.

4.4.7 Leuco-spessartite

The samples of leuco-spessartite (RB4, RB5 & AR4) are generally more even grained than the mela-spessartite, and comprise of a medium-grained poikilitic texture (Figure 46) composed of euhedral to subhedral oikocrysts of brown hornblende and green Mg-hornblende (500 to 2000 μm), which are entirely enclosed by subhedral to anhedral grains of plagioclase (500 to 1000 μm). Accessory augite is encountered occasionally, while irregular magnetite grains are more commonly found enclosed by hornblende as well as plagioclase.

Compared to the mela-spessartite, hornblende grains tend to be smaller and more subhedral while displaying thicker rims of green Mg-hornblende. Typical amphibole cleavage on {110} can be seen, while twinning is likewise absent.

Plagioclase grains are more Na-rich in the leuco-spessartites and classify predominantly as oligoclase. The plagioclase is larger and more euhedral and appears colourless under plain polarized light. Polysynthetic albite twinning is commonly displayed on the (010) composition plane, while 94° intersections of the {001} and {010} cleavage planes can at times also be observed.

Augite is less commonly encountered in the leuco-spessartites, its modal percentages vary quite abruptly between samples, of which only one (RB5) contained an accessory amount that could be identified by XRD analysis. The augite grains are small and rounded and occur within the centre of the hornblende. Due to their small grain size, no distinguishing features could be identified.

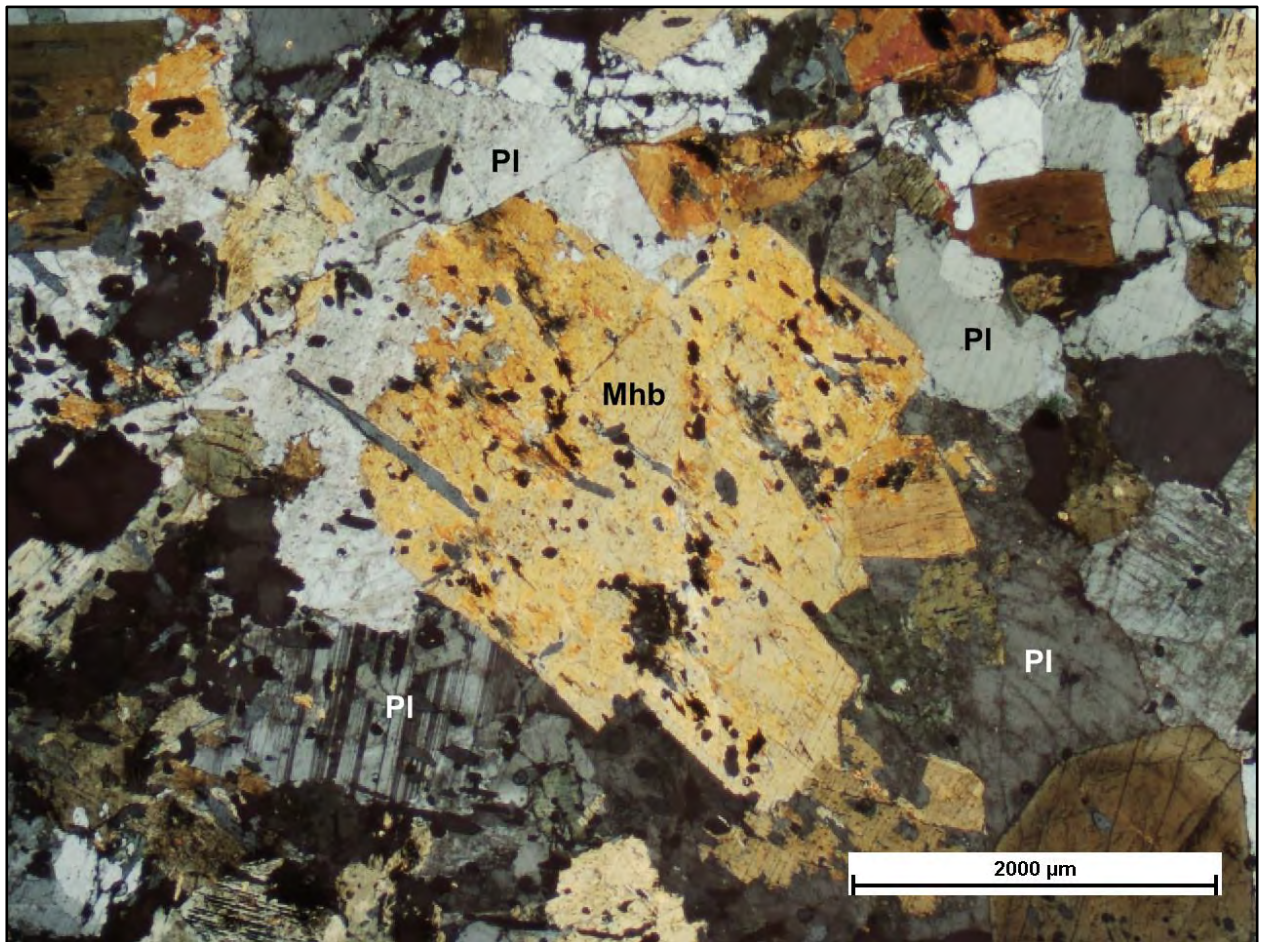


Figure 46: Photomicrograph of leuco-spessartite (sample RB4) under XPL, displaying a medium-grained poikilitic texture that comprises of slightly larger subhedral to anhedral oikocrysts of brown hornblende and green Mg-hornblende that are enclosed by medium-grained plagioclase crystals exhibiting albite twinning. Note the dark inclusions that occur within the hornblende and plagioclase, these are composed primarily of opaque magnetite with small fractions of dark grey augite.

4.4.8 Quartz Akali Syenite

The sample of quartz alkali syenite (RB13) is medium grained hypidiomorphic granular (Figure 47) and consists primarily of long, skeletal, subhedral grains of plagioclase (500 to 4000 μm) and smaller euhedral grains of microcline (200 to 1000 μm), quartz (50 to 250 μm) and dark brown Fe-hornblende (50 to 1000 μm).

The cores of the plagioclase are completely altered due to albitisation with rims of pure colourless albite. Albite and Carlsbad twinning is common amongst larger grains, and cleavage planes, $\{010\}$, are regularly observed parallel to the c-axis.

Microcline grains are more irregular than plagioclase and displays a darker first order grey interference colour. The microcline grains commonly occur along the borders of plagioclase but

are difficult to distinguish from quartz, since twinning or exsolution lamellae are not regularly displayed. Quartz occurs as small equant grains that are colourless and regularly enclosed by plagioclase.

Fe-hornblende grains are is euhedral and irregular and occur with small patches of muscovite within the interstitial spaces between the plagioclase and microcline. Some smaller fragments of Fe-hornblende are also poikilitically enclosed by plagioclase but are too small to be able to observe any twinning or cleavage. The Fe-hornblende grains do display minor alteration to dark green clinocllore at their grain boundaries and contain small opaque inclusions of magnetite.

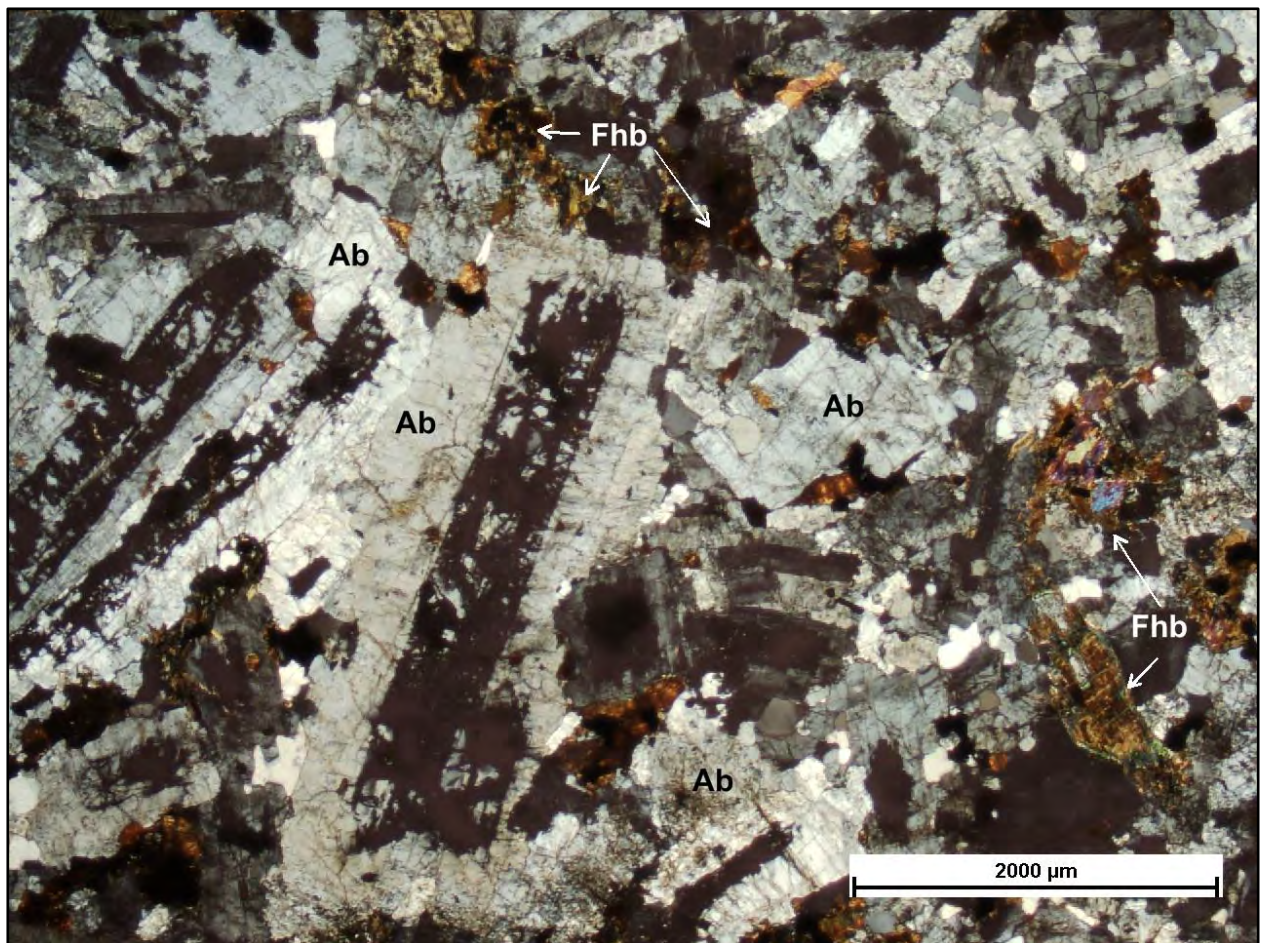


Figure 47: Photomicrograph of quartz alkali syenite (sample RB13) under XPL, illustrating the hypidiomorphic granular texture of the rock, which is primarily composed of long subhedral plagioclase laths and smaller euhedral grains of microcline, quartz, and Fe-hornblende. The plagioclase grains are completely albitised and are mantled by a rim of albite displaying a first order white interference colour. Within the interstitial spaces, partly altered dark brown grains of Fe-hornblende can be observed with thin rims of green clinocllore as well as small patches of light brown muscovite.

4.4.9 Alkali Granite

The samples of alkali granite (AB17 & RB15) are texturally similar to the quartz alkali syenite but display some distinct mineral characteristics (Figure 48). The rock is medium- to fine-grained hypidiomorphic-granular and comprise primarily of albite (500 to 4000 μm) and quartz (100 to 300 μm) with minor amounts of microcline perthite (500 to 1000 μm). Small irregular patches of dark brown Fe-hornblende, muscovite and clinocllore could also be identified but not as frequently.

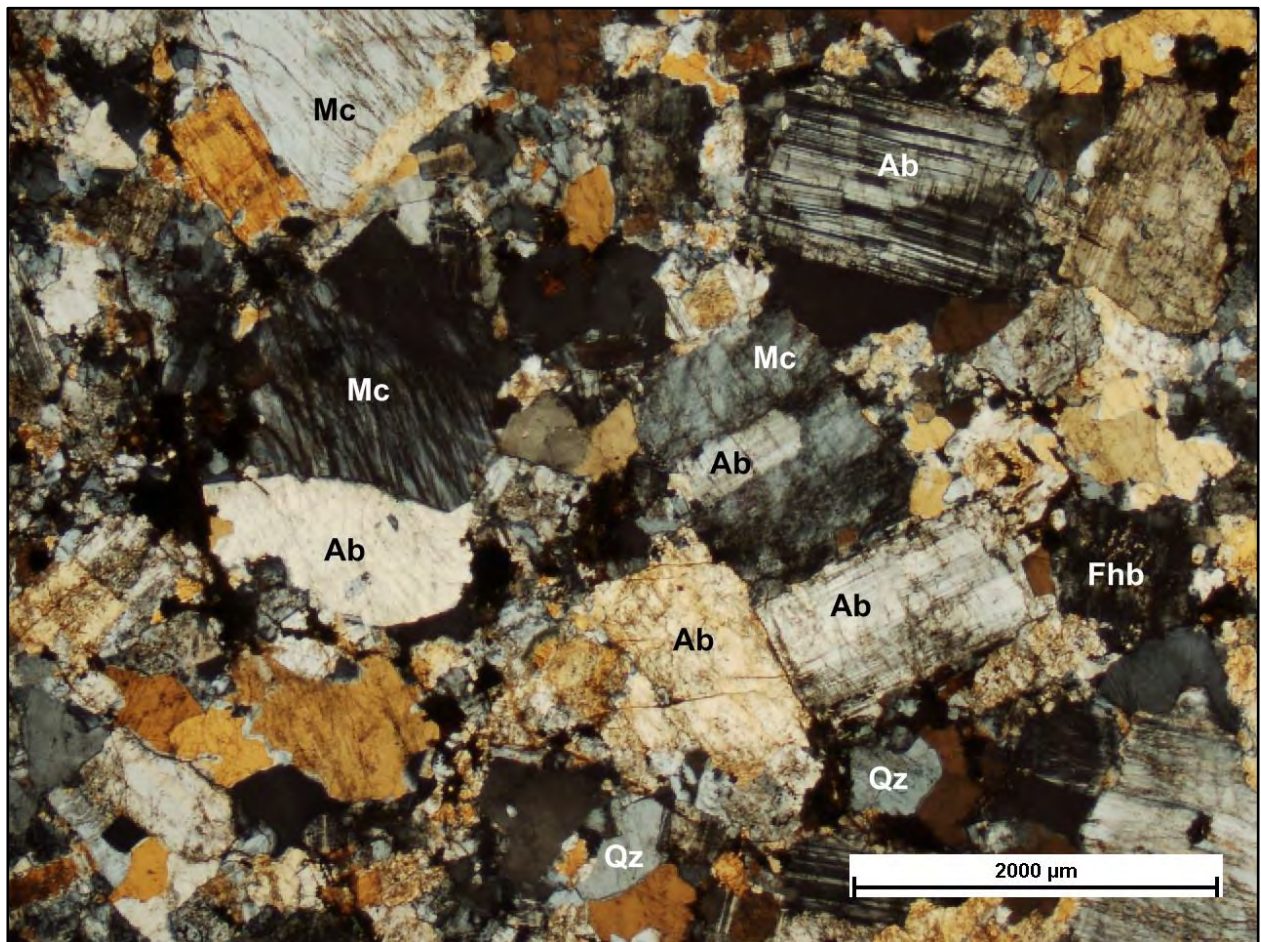


Figure 48: Photomicrograph of alkali granite (sample RB15) under XPL, displaying a medium- to fine-grained hypidiomorphic-granular texture, which is composed of tabular subhedral grains of albite, smaller rounded grains of quartz and minor amounts of microcline perthite that encloses some of the smaller plagioclase grains. The lamellar twins of the albite grains are partly dislocated and appear to have recrystallized, while in the microcline small exsolution lamellae of albite can be clearly seen within the host grains. Some accessory dark brown Fe-hornblende is also enclosed within the feldspars.

Albite is generally the most prominent mineral and occurs as larger subhedral grains, which are tabular and lack the eroded cores that are observed in the quartz alkali syenite. Polysynthetic lamellar twins are regularly seen on the (010) composition plane, and in some grains, the twins look deformed and recrystallised (Figure 48). The {010} cleavage plane is commonly displayed

parallel with the c-axis, while exsolution lamellae are absent within the plagioclase. Smaller grains of albite are regularly seen enclosed by the microcline perthite.

Microcline perthite has a smaller grain size than the plagioclase feldspar and is generally more euhedral and irregularly shaped, with grain boundaries that appear serrated and often altered to sericite. The microcline host grains display a first order dark grey interference, while the finer albite lamellae can be distinguished by their lighter grey interference colour. The microcline also frequently display cross-hatched patterns that comprise of lamellar twins according to albite and pericline laws. This type of twinning is characteristic of microcline (Klein & Duttraw, 2008; Nesse, 2004).

Quartz grains are smaller than the perthite, but are much more prominent throughout the rock. Similar to the quartz alkali syenite, the quartz grains are equant to rounded and seldom display fracturing. Quartz occurs mostly along the boundaries of the feldspar and is often enclosed by some of the grains as well.

4.5 GEOCHEMISTRY

The whole rock geochemical composition of selected samples from the epidiorite sill, as well as the syn-Bushed aged intrusions, as determined with XRF analysis, is provided in Appendix E1. To test whether the different syn-Bushveld rock types show any chemical relation to one another, as field occurrences and mineralogical trends suggest, and to demonstrate the degree of chemical variation, which may have an influence on their related soil properties, bivariate plots were created for selected major oxides and trace element concentrations. An intra-element correlation matrix was constructed (Appendix E3), in order to select element and oxide pairs that best could represent associated chemical variations amongst the different rock types, as well as demonstrate, with respect to soil formation, if the ultramafic rock series possess the characteristics needed to possibly produce the “serpentine effect”.

The correlation of co-variant element pairs was calculated based on the Pearson correlation coefficient (r), which can be stated by the following equation:

$$r = \frac{\sum (x - \bar{x})(y - \bar{y})}{\sqrt{\sum (x - \bar{x})^2 \sum (y - \bar{y})^2}}$$

Correlation coefficients for the entire group of igneous and metamorphic rocks of the KC, revealed strong (>0.8) to moderate (0.7 to 0.8) correlations for some element assemblages. Table 7 summarises the correlated trends for these particular groups, from which specific pairs were

selected for the bivariate diagrams displayed in Figure 49a to g, Figure 50a to h and Figure 51a & b.

Table 7: Selected intra-element correlations for the syn-Bushveld rock series of the Koedoesfontein Complex, based on correlation coefficients (r) of $>\pm 0.7$ (Appendix E3).

POSITIVELY CORRELATED ($>+0.7$)		NEGATIVELY CORRELATED (>-0.7)
SiO ₂	Na ₂ O, K ₂ O, Zr	FeOt, Mn ₃ O ₄ , Zn, As, Se, Br
TiO ₂	FeOt, Zn, As, Se, Br, Y	
Al ₂ O ₃	Na ₂ O, Ga, Ba	MgO
FeOt	Mn ₂ O ₃ , Zn, As, Se, Br	
Mn ₃ O ₄	Zn, As, Se, Br	
MgO	Cr, Co, Ni	Na ₂ O, Ga, Ba
CaO	Sc	
Na ₂ O	K ₂ O, Ga, Ba	
K ₂ O	Rb, Zr, Ba	

Polynomial and exponential regression was applied to the selected element pairs to determine how closely each of the intrusive rock types fits within a differentiated trend. Samples of epidiorite were omitted in order to gain a better visual and statistical discrimination between samples that are thought to be related. Extraordinary samples including the diopside pegmatite (RB18) and uraltite (RB14) were, nevertheless, included in order to illustrate their relative association with the entire intrusive series. These are small, isolated, highly altered rock varieties that in some cases did not conform to the general trends observed for the major rock types. As a result, lower regression values were observed for some oxide and trace element pairs (Figure 49a & b, and Figure 50d, f, g & h).

A polynomial or curvilinear trend uses the following equation to calculate the least squares fit through points:

$$y = b + c_1x + c_2x^2 + c_3x^3 + \dots + c_6x^6$$

Where b and $c_1 - c_6$ are constants.

Whereas for an exponential trend the following equation is used:

$$y = ce^{bx}$$

Where c and b are constants and e is the base of the natural logarithm.

In accordance with the coherent major oxide and trace element pairs (Table 7), statistical regression similarly revealed strong (> 0.8 to < 0.9) to moderate (> 0.7 to < 0.8) associations between the different rock types, which generally display closely correlated trends ranging from ultramafic to felsic. On the basis of chemical variation, results correlate with the mineralogy of each rock type determined with XRD analysis (Appendix D2) and display distinct geochemical trends across the entire range of igneous and metamorphic rocks. The abundance and correlation between different major– and trace element pairs can be explained by the presence of major rock-forming silicates, as well as phosphate- and oxide minerals, which crystallised at different stages of magmatic evolution and related metamorphic alteration.

Bowen (1928) proposed a generalised reaction series for crystallising magmas, from which it can be deduced that mafic silicate minerals (olivine, pyroxene and amphibole) along with Ca-rich plagioclase are generally more stable at high temperatures and will therefore crystallise first, followed by felsic minerals such as Na-rich plagioclase, K-feldspar, muscovite and quartz, which will typically crystallise at lower temperatures. If these minerals and their comprising elements are removed by fractional crystallisation during progressive cooling of a magma, then corresponding geochemical differentiation trends will appear within the resulting rock types. (Albarède 2003, Best 2003; Gill, 2010; Winter, 2010). These subsequent trends can be clearly observed for the rock series of the KC, which display alternating concentrations of major oxides and trace elements (section 4.5.1 & section 4.5.2).

4.5.1 Major element variation

The largest concentrations for FeOt (9.11 to 20.70 wt. %), Mn_3O_4 (0.20 to 0.40 wt. %) and MgO (7.42 to 26.31 wt. %), are observed within the mela-dolerite, ultramafic rock types (wehrlite, Ol clinopyroxenite, uraltite and hornblendite), as well as the mela-spessartite (Appendix E1). The aforementioned elements illustrated a progressive decrease in concentration towards the rest of the mafic and dioritic rock varieties (Opx-Hbl gabbro, leuco-spessartite and epidiorite), which contain lower amounts of FeOt (5.43 to 9.2 wt. %), Mn_2O_3 (0.14 to 0.27 wt. %) and MgO (3.92 to 13.15 wt. %). The quartz alkali syenite and alkali granite have the lowest values for these oxides (1.4 to 6.78 wt. % FeOt; 0.01 to 0.11 wt. % Mn_2O_3 ; 0.14 to 1.41 wt. % MgO).

Fe, Mn and Mg are particularly compatible with mafic silicate- and oxide minerals such as olivine, pyroxene, amphibole and magnetite, and are removed early from a differentiating melt when fractional crystallisation of these minerals occur (Albarède, 2003; Best, 2003, Klein and Dutrow, 2007; Nesse, 2004). These elements correlated negatively with the SiO_2 content, which tend to increase within a melt, since oxides do not accommodate silicon (Si) and relatively less of it is removed by fractionating mafic silicates (Winter, 2010). The major oxide trends perceived for

these elements throughout the layered sequences, are therefore attributed to varying modal percentages of mafic silicate- and oxide minerals.

From the bivariate diagram FeOt vs SiO₂ (Figure 49a), it can be deduced that the FeOt content is strongly correlated with the occurrence of magnetite, since rock types containing larger amounts of this mineral (mela-dolerite, mela-spessartite and the deformed sample of Ol clinopyroxenite) grouped near the upper left end of the series trend (20.7 to 18.86 wt. % FeOt). The remaining ultramafic rock types (wehrlite, Ol clinopyroxenite, uralitite and hornblendite) plots further down the trend (9.11 to 13.75 wt. % FeOt) despite comprising larger amounts of olivine, augite and hornblende. These minerals thus generally contributes less Fe to the rocks bulk composition. A similar observation is also seen for the rest of the mafic to felsic rock types, which plots from the midpoint towards the lower end of the series (1.4 to 9.2 wt. % FeOt). At the midpoint, the slope of the trend decreases as a result of Fe that became more incompatible, marking the crystallisation of felsic minerals including muscovite, feldspar and quartz. The noticeable difference in FeOt, between the samples of quartz alkali syenite (6.78 wt. %) and alkali granite (1.41 to 1.81 wt. %), specifically characterises the absence of magnetite within the alkali granite as well as the decline in accessory hornblende, which was noted optically.

The Mn₃O₄ content (Figure 49b) indicated a strong association with both the presence of magnetite and augite, and was therefore positively correlated with the distribution of FeOt (Table 7). Particularly high concentrations of Mn₃O₄ were observed for the mela-dolerite (0.40 wt. %) and diopside pegmatite (0.40 wt. %), whereas the ultramafic rocks, Opx-hornblende gabbro, mela-spessartite, and leuco-spessartite (RB5) grouped within a distinctively lower cluster, just above the midpoint of the series trend (0.18 to 0.27 wt. % Mn₃O₄). From the midpoint downward (0.01 to 0.15 wt. % Mn₃O₄), the slope of the trend decreases due to the significant reduction in augite from the dioritic to tonalitic leuco-spessartite samples (RB5 & RB4), and absence of this mineral within the quartz alkali syenite and alkali granite. The relatively large gap in Mn₃O₄ content between the samples of quartz alkali syenite (0.11 wt. %) and alkali granite (0.01 to 0.02 wt. %) is also attributed to lower concentrations of accessory hornblende, as well as the absence of magnetite.

The MgO content (Figure 49c) is specifically correlated with fractionated forsterite, since Mg is strongly lithophilic (Albarède, 2003), and is the most concentrated within the wehrlite (22.25 to 26.31 wt. %) and olivine clinopyroxenite (15.60 to 17.27 wt. %). The chilled contact sample of mela-dolerite, which occurs at the base of the ultramafic rock series, illustrated a contrasting lower MgO content (7.42 wt. %), which was more comparable to that of the mela-spessartite (5.08 wt. %). These two rock types are particularly silica poor and grouped separately from the general series trend. This indicates that they are probably more primitive in terms of the general series

evolution, since their silica content was more diluted during crystallisation. In these rocks the majority of Mg is accommodated by both augite and hornblende. The same is observed for the uralitite, hornblendite, diopside pegmatite and Opx-Hbl gabbro, which have slightly higher concentrations of MgO (8.44 to 15.63 wt. %), and groups within the midpoint of the series trend. The leuco-spessartite, quartz alkali syenite and alkali granite, group at the lower end of the trend (0.14 to 5.99 wt. % MgO) signifying the gradual disappearance of augite, which is also marked by the distinct decrease in slope angle.

The TiO_2 content illustrated a particularly strong association with FeOt (Table 7 & Appendix E3) and similarly, the largest concentrations are observed within the mela-dolerite (2.73 wt. %) and mela-spessartite (2.83 wt. %). The deformed sample of Ol clinopyroxenite (1.19 wt. %), the uralitite (0.98 wt. %), hornblendite (1.75 wt. %) and quartz alkali syenite (1.26 wt. %) are more comparable and have slightly lower concentrations, while the rest of the rock types, wehlrite (0.24 to 0.34 wt. %), Ol clinopyroxenite (0.42 wt. %), Opx-Hbl gabbro (0.4 to 0.51 wt. %), leuco-spessartite (0.58 to 0.61 wt. %) and alkali granite (0.08 to 0.12 wt. %) have significantly lower amounts.

Titanium (Ti) is a high field strength (HFS) element that is generally incompatible with olivine and pyroxene in basaltic magmas (Best, 2003, Winter, 2010), and will rather concentrate within oxides such ilmenite and magnetite, as well as at later stages of differentiation within hornblende (Best, 2003). In alkali basalts, Ti may also concentrate within titanian augites (Gill, 2010).

The bivariate diagram for TiO_2 vs FeOt (Figure 49d) illustrates the close association between the TiO_2 content and the occurrence of magnetite as well as the degree of uralitisation. The mela-dolerite and mela-spessartite both contain larger amounts of granular magnetite and are significantly uralitised. These samples plot at the highest end of the trend line (2.73 to 2.83 wt. %) suggesting that both magnetite and hornblende are significant reservoirs for Ti. The deformed sample of Ol clinopyroxenite, the uralitite and hornblendite have higher degrees of alteration but contains fewer amounts of magnetite, and therefore plots lower along the trend line (1.75 to 1.91 wt. % TiO_2).

The quartz alkali syenite contains a similar amount of TiO_2 compared to the aforementioned rock types, but far less FeOt. This is ascribed to an even lower content of magnetite, while the remaining ultramafic to felsic rocks cluster more along the lower end of the trend (0.08 to 0.61 wt. % TiO_2), because these rocks contain fewer to no hornblende, as well as more feldspar that excludes Ti during crystallisation (Best, 2003).

The SiO_2 , Al_2O_3 , Na_2O and K_2O content strongly correlates amongst one another (Table 7 & Appendix E3), and display a progressive increase from the ultramafic to felsic rock types

(Appendix E1). In the ultramafic rocks as well as mela-dolerite, the lowest ranges for SiO₂ (35.47 to 47.50 wt. %), Al₂O₃ (0.76 to 8.14 wt. %), Na₂O (0.13 to 2.28 wt. %) and K₂O (0 to 0.76 wt. %) are observed, whereas the mafic and dioritic varieties (Opx-Hbl gabbro, mela-spessartite, leuco-spessartite and epidiorite), typically illustrate larger values (40.48 to 56.30 wt. % SiO₂; 5.66 to 16.05 wt. % Al₂O₃; 0.41 to 5.32 wt. % Na₂O; 0.37 to 1.31 wt. % K₂O). The largest concentrations for these specific elements occurred within the felsic rock types (52.81 to 63.65 wt. % SiO₂; 13.95 to 17.75 wt. % Al₂O₃; 4.85 to 7.56 wt. % Na₂O; 2.14 to 4.05 wt. % K₂O).

The SiO₂ concentration reflects the difference in abundance of mafic and felsic silicates that crystallised within each rock type. Mafic minerals accommodate lower ratios of Si compared to other structurally bonded cations and will therefore contribute less Si than felsic minerals to the rocks bulk composition. It is for this reason that a gradual increase in the SiO₂ content is observed from the ultramafic to felsic rock types (Figure 49a to g) since Si will increase within the melt during differentiation (winter, 2010).

The mela-dolerite (35.47 wt. %), wehrlite (40.80 to 43.40 wt. %), deformed sample of Ol clinopyroxenite (39.36 wt. %) and mela-spessartite (40.48 wt. %) contain the lowest concentrations of SiO₂ due to a higher abundance of magnetite and / or olivine. Larger concentrations are observed for the Ol clinopyroxenite (47.50 wt. %), uraltite (47.39 wt. %), Opx-Hbl gabbro (46.66 to 47.23 wt. %) and hornblendite (44.77 wt. %), which contain more augite and / or hornblende as well as plagioclase. The largest concentrations of SiO₂ was observed within the samples of leuco-spessartite (49.32 to 56.30 wt. %), quartz alkali syenite (52.81 wt. %) and alkali granite (62.33 to 63.65 wt. %), and are ascribed to the dominance of plagioclase, microcline and quartz, and lower amounts of hornblende and accessory magnetite.

When the rock samples of the series are classified based on their silica content, it appears that for some samples this does not correspond with their mineralogical classification. Similar as for the epidiorite, which is more basic (40.43 to 50.85 wt. % SiO₂) due to its original inherent chemistry, it would seem the samples that were altered, retained most of their silica content to some extent, although some enrichment can also be noticed.

The mela-dolerite, which classifies as dioritic based on its mineral composition, has a particularly low silica content and is classified as ultrabasic. This indicates that the rock must have been more ultramafic or mafic in terms of differentiation, but was later recrystallized at lower temperatures to a more dioritic mineralogy. A similar observation can be made for the leuco-spessartite which are more dioritic and tonalitic in terms of the mineralogy, but classifies as basic to intermediate based on silica content. The alkali granite is also particularly silica poor for a felsic rock and borders between intermediate to acid. This is probably because the alkali granite had evolved from a more syenitic magma, which became more enriched with silica as crystallisation progressed.

Aluminium (Al), Na and K are lithophile elements, and are mainly contained within in feldspars and to a lesser extent in hornblende (Klein and Dutrow, 2007; Nesse, 2004). Higher values are therefore encountered in rock types that primarily comprise of these minerals (Opx-Hbl gabbro, mela-spessartite, leuco-spessartite, epidiorite, quartz alkali syenite and alkali granite). For the mafic and dioritic rock types, some of the alkalis are accommodated by secondary hornblende, however, feldspars contain larger quantities of Al, Na and K compared to hornblende, and therefore these elements were particularly concentrated within the quartz alkali syenite and alkali granite, which is mainly composed of albite and microcline.

The bivariate diagram for Al_2O_3 vs MgO (Figure 49e) indicates a strong association between the different rock types, which form a decreasing trend indicative of the different concentrations of hornblende and feldspar. The wehrlite, Ol clinopyroxenite and diopside pegmatite group at the lower right end of the trend (0.76 to 2.74 wt. % Al_2O_3), signifying their low concentrations of hornblende and incompatibility of Al with olivine and augite, while the hornblendite and uraltite, group further left, just below the midpoint of the trend (5.66 to 6.11 wt. % Al_2O_3), because of their larger concentration of hornblende. From the midpoint upwards (8.14 to 17.75 wt. % Al_2O_3), the slope of the trend is increased due to the high compatibility of Al with feldspar within the mafic, dioritic end felsic rock types. These samples display a gradual increase in Al_2O_3 as their feldspar content increases.

The bivariate diagrams for Na_2O and K_2O vs SiO_2 (Figure 49f & g), display series trends that can be correlated the distribution of different phases of feldspar (plagioclase and microcline). With respect to the Na_2O content (Figure 50f), a strong association with plagioclase can be observed, while the K_2O content (Figure 50g) relates more to the distribution of microcline. The mela-dolerite and mela-spessartite, which plots at the silica poor ends of both series trends (35.47 to 40.40 wt. % SiO_2), have elevated levels of alkalis (2.28 to 3.32 wt. % Na_2O & 0.46 to 0.51 wt. % K_2O). This can be attributed to the presence of plagioclase, while the ultramafic rocks have distinctly lower concentrations (0.25 to 1.20 wt. % Na_2O & 0.01 to 0.37 wt. % K_2O) due to the absence thereof.

As a result of the high volatility of K and its higher incompatibility with the plagioclase solid solution series, a smaller difference in the K_2O concentration is observed between the ultramafic rocks and the mela-dolerite and mela-spessartite. The Opx-Hbl gabbro and leuco-spessartite plots further along the increasing trend lines (0.41 to 5.32 wt. % Na_2O & 0.56 to 1.31 wt. % K_2O), which correlate with a higher plagioclase content. In terms of their Na_2O concentration, the Opx-Hbl gabbro and leuco-spessartite are more dissimilar, as the plagioclase of the leuco-spessartite is more Na-rich. Whereas in terms of their K_2O concentration, the rock types are more comparable since little K was accommodated by the plagioclase of both rock types during crystallisation. The quartz alkali syenite and alkali granite plot at the upper part of the series trends (4.85 to 7.56 wt.

% Na₂O; 2.14 to 4.05 wt. % K₂O), but are separated from one another due to the particular difference in the microcline content, which is increased within the alkali granite.

CaO did not correlate with any major elements (Appendix E3), and displayed less variation amongst the intrusive rock types (Appendix E1). The highest values for CaO are observed within the ultramafic rocks (9.98 to 18.05 wt. %), nevertheless the mafic (10.97 to 11.67 wt. %) and dioritic rock types (8.48 to 11.25 wt. %) do not differ much from these values. The felsic rock types, in contrast, contained considerably lower concentrations of CaO (0.43 to 1.81 wt. %).

The high CaO values observed within the wehrlite (9.88 to 12.75 wt. %) and Ol clinopyroxenite (13.47 to 18.05 wt. %) can be attributed to augite, which is the primary source of Ca within the ultramafic rocks. Augite contains less Si compared to plagioclase or hornblende, as well as no Na or Al. Therefore a slightly higher relative abundance of Ca is observed, compared to the mafic and dioritic rock types that contain less or no augite. In these rocks hornblende and plagioclase are the primary sources of Ca. In the felsic rock types, due to absence of augite, most of the Ca was incorporated within plagioclase, however, as a result of the melt's strong sodic character the CaO concentration was particularly low and therefore Na-rich albite was formed.

The P₂O₅ concentration did not correlate with any major- or trace elements (Appendix E3), as well as displayed no clear trend across the different rock types (Appendix E1). The highest values for P₂O₅ were reported within the quartz alkali syenite (1.29 wt. %) while the mela-dolerite (0.61 wt. %) and both varieties of spessartite (0.68 to 0.91 wt. %) had slightly lower concentrations. All the other rock types displayed comparable lower values for P₂O₅ (0.01 to 0.1 wt. %).

The P₂O₅ concentration is primarily related to the distribution of apatite, which occurs as an accessory mineral within most of the rock series. Apatite found within igneous rock types can be particularly associated with magnetite as well as alkaline rock types (Klein and Dutrow, 2007). According to Best (2003), P is incompatible with mafic magmas and will usually crystallise within more dioritic and silicic melts where apatite is more stable. Thus, with respect to the syn-Bushveld series, it would seem that crystallisation of apatite was favoured within the dioritic spessartite group as well as the felsic quartz alkali syenite.

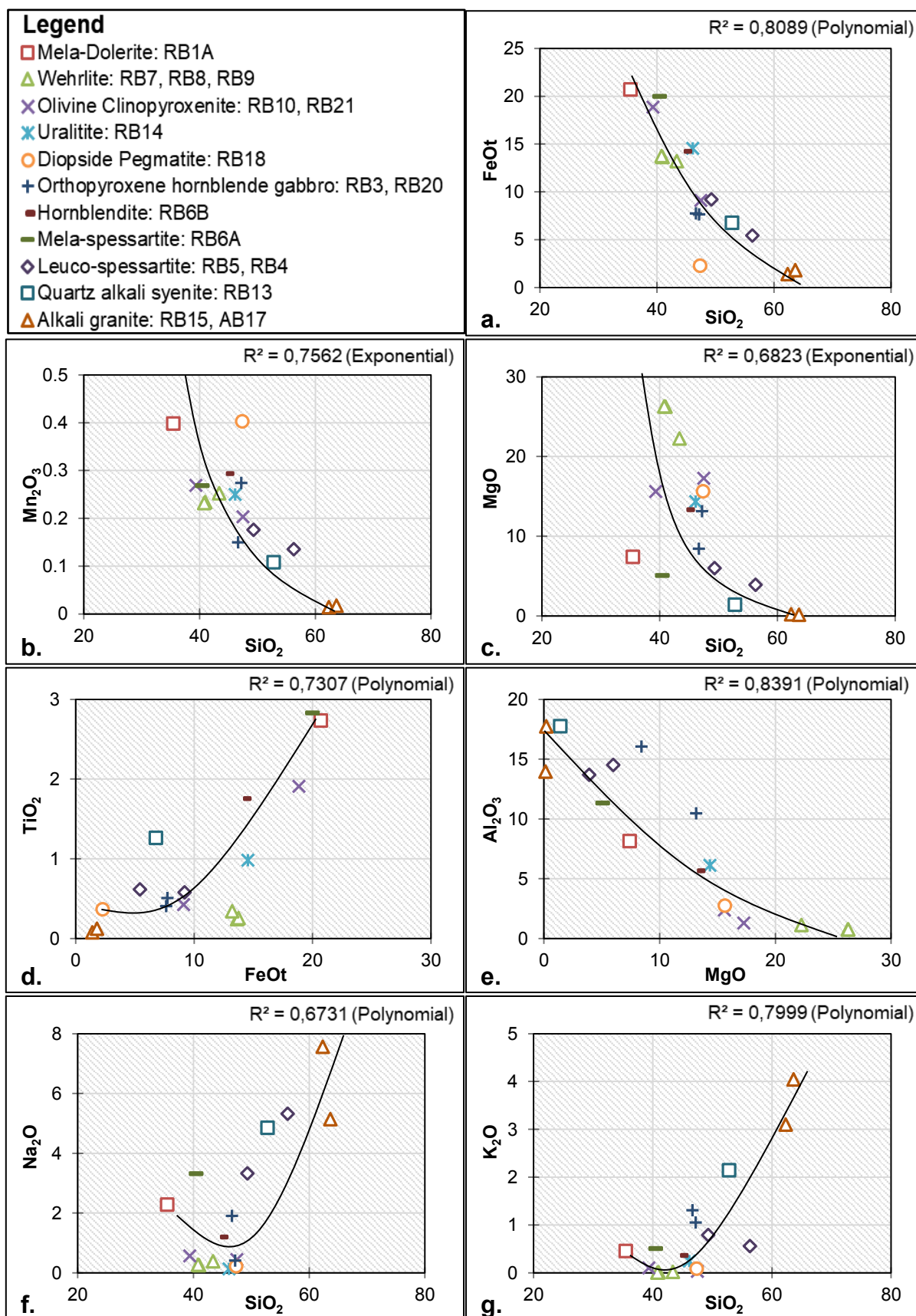


Figure 49: Bivariate diagrams for selected major oxides of the syn-Bushveld intrusive series, depicting (a, b & c) FeOt, Mn_2O_3 and MgO vs SiO_2 , (d) TiO_2 vs FeOt, (e) Al_2O_3 vs MgO and (f & g) Na_2O and K_2O vs SiO_2 .

4.5.2 Trace element variation

Trace elements are more selective for different mineral phases than major elements, and are therefore more selectively partitioned within the crystal lattice based on their compatibility (Winter, 2010). This property makes some trace elements strong petrogenetic indicators for magmatic processes such as fractional crystallisation (Albarède 2003; Best 2003; Winter, 2010), since they are commonly exchanged with corresponding major elements and can concentrate in specific rock types when removed from a crystallising melt. Some immobile trace elements such as yttrium (Y) and Co are also less affected by weathering and alteration, and may provide better discrimination between samples that are genetically related (Best, 2003; Floyd & Winchester, 1975, Winchester & Floyd, 1977, Hastie *et al.*, 2007).

Statistical analysis of the element composition for all the intrusive rock types of the KC, provided several trace element groups that correlated with the distribution of different major oxides (Table 7). From these groups, some of the strongest correlated pairs were observed to clearly discriminate between the different rock groups and their corresponding layers.

Zinc (Zn 17 to 197 ppm), Arsenic (44 As to 109 ppm), Selenium (Se 12 to 38 ppm) and Bromine (Br 10 to 37 ppm) illustrated a strong correlation with one another (Appendix E3), and specifically associated with larger values of TiO_2 , FeOt and Mn_2O_3 (Table 7). The largest concentrations of these elements (Zn 47 to 197 ppm; As 71 to 109 ppm; Se 21 to 38 ppm; Br 18 to 37 ppm) were observed within the ultramafic and mafic rocks as well as the dioritic mela-spessartite and epidiorite (Appendix E1). The leuco-spessartite has lower values for these elements (Zn 60 to 69 ppm; As 59 to 72 ppm; Se 16 to 23 ppm; Br 16 to 20 ppm), while the quartz alkali syenite and alkali granite contain the lowest concentrations for As (44 to 57 ppm), Se (12 to 15 ppm) and Br (10 to 15 ppm). The Zn content of the quartz alkali syenite, in contrast, is elevated (113 ppm) and comparable with the metamorphic uralite (124 ppm), whereas the alkali granite contains the lowest concentration of this element (17 to 38 ppm).

Zn, As and Se are chalcophile trace elements and will typically occur within sulphide minerals, while Br is a lithophile element and will rather be incorporated within silicates (Albarède 2003). To determine the mineralogical association for these elements, the strongest correlated pairs (Se and Br) were plotted against FeOt (Figure 50a & b). Results from the bivariate plots indicate a clear relationship between these trace elements and rock samples containing larger amounts of magnetite (Mela-dolerite, Ol clinopyroxenite (RB21) and mela-spessartite). It can, therefore, be deduced that Se and Br partitioned early during crystallisation, and occur as respective sulphides and bromides or bromine oxides that are rather associated with the opaque grains and veinlets of magnetite.

Yttrium (Y 2 to 23 ppm) has some association with the former trace elements (Zn, As, Se, Br; Table 7), but is rather positively correlated with TiO_2 (Appendix E3), since Y is an incompatible element and will strongly partition into amphibole or apatite at later stages of magmatic evolution (Best 2003; Winter, 2010).

The bivariate diagram for Y vs TiO_2 (Figure 50c), illustrates a low regression value ($R^2 = 0.5434$) due to the occurrence of two distinct trends within the intrusive rock series that are both indicative of original differentiation and the degree of secondary alteration. The hornblende within the ultramafic to dioritic intrusions is secondary after augite, but its abundance does not correspond with the Y content. This is because the element was far less affected by alteration compared to TiO_2 . It can, therefore, be argued that the Y content must have originally partitioned within apatite during fractional crystallisation, and the concentrations observed are thus inherent of the original differentiation process.

The mela-dolerite (23 ppm), mela-spessartite (16 ppm) and hornblendite (16 ppm) contain particularly high values of Y that correlates with the distribution of apatite. These rock types are also highly uraltised, and plot at the upper right end of the trend line, as a result from their higher TiO_2 content (Figure 50c). The wehrlite (2 to 3 ppm), Ol clinopyroxenite (5 to 9 ppm), Opx-Hbl gabbro (8 to 10 ppm) and leuco-spessartite (12 to 14 ppm) contain lower concentrations of Y and are less altered. As a result, these rock types form a separate differentiated trend that is indicative of the gradual increase in compatibility of Y towards the dioritic leuco-spessartite. The deformed sample of clinopyroxenite displays more alteration and contains larger amounts of TiO_2 . Some slight enrichment of Y (9 ppm) can be observed as this sample is more comparable to the Opx-Hbl gabbro. This is most likely due to its close occurrence to the felsic intrusions which contain higher amounts of Y (5 to 13 ppm), due to the presence of apatite, and are believed to have caused the late stage alteration during emplacement. This observation is even more noticeable and confirmed within the uraltite that contains a far higher Y content (13 ppm) compared to the wehrlite and Ol clinopyroxenite. The uraltite is particularly comparable to the felsic rock types which crosscut it. These rocks all contained Y concentrations that are similar to the leuco-spessartite.

Cr, Co and Ni are highly compatible with early crystallising mafic silicate and spinel minerals (olivine, clinopyroxene and magnetite), and can be readily incorporated into their crystalline structure by substituting for Fe^{2+} or Mg^{2+} . The intrusive rocks of the KC illustrate a strong relationship between the Cr, Co and Ni content of each rock type, and the presence of more Mg-rich olivine and augite, which relates to a higher MgO concentration (Table 7 & Appendix E3). A decline in concentrations of these trace elements were observed from the ultramafic to felsic rock

types (Figure 50d, e & f), which suggests removal by fractionation of olivine and augite (Best, 2003; Winter, 2010).

Very high concentrations of Cr (711 to 1752 ppm), Co (99 to 156 ppm) and Ni (277 to 608 ppm) were observed within the ultramafic wehrlite and Ol clinopyroxenite (Appendix E1), due to the dominance of olivine and clinopyroxene. The Cr content is particularly high due to its strong compatibility with augite, while Co and Ni are more compatible with olivine (Best 2003). During alteration, these trace elements were later incorporated into new crystalline structures of serpentine and hornblende. For this reason, the sample of uralitite has a particularly high Ni content (781 ppm) and can be related to its wehrlite photolith, which probably comprised of wehrlite containing a large amount of olivine.

The Opx-Hbl gabbro and hornblendite also contain fairly high amounts of Cr (590 to 2008 ppm) due to the presence of pyroxene and secondary hornblende. The Co (74 to 96 ppm) and Ni (134 to 386 ppm) content, however, are significantly lower within these rock types due to the absence of olivine and serpentine.

The mela-dolerite, both the varieties of the spessartite as well as the epidiorite contain less augite and more hornblende, consequently these rock types contain lower concentrations of Cr (246 to 27 ppm), Co (66 to 104ppm) and Ni (58 to 151 ppm), since these elements are generally less compatible with amphibole (Winter 2010).

The quartz alkali syenite and alkali granite crystallised after the ultramafic, mafic and dioritic rock types, and comprise the least amount of mafic minerals. Correspondingly these rock types contain the lowest concentration of Cr (33 to 26ppm), Co (24 to 80ppm) and Ni (12 to 34ppm), as these elements less compatible with hornblende, and incompatible with feldspars (Winter 2010).

Scandium (Sc) positively correlated with CaO (Table7 & Appendix E3), and displays a distinguishing trend for the different intrusive rock types that is indicative of the occurrence of augite (Figure 50g). This is because Sc concentrates within pyroxenes, and can be used as an indicator for pyroxene fractionation (Albarède, 2003; Winter, 2010).

The largest values for Sc (Appendix E1) are observed within the Ol clinopyroxene (55 to 57 ppm), and mela-dolerite (49 ppm), and can be attributed to the dominance and early crystallisation of augite within the ultramafic series. The wehrlite (29 to 41ppm) and uralitite 26 ppm), which form part of this series, contain lower values of Sc, since the presence of more olivine in the lower layers, dilutes the concentration of augite within the bulk rock composition. The correspondence in Sc content between the wehrlite and uralitite strengthens the notion that the uralitite must have originated from an olivine-rich wehrlite containing lower amounts of augite.

The wehrlite and uralitite are more comparable to the Opx-Hbl gabbro (38 to 43 ppm), hornblendite (44 ppm) and both varieties of spessartite (21 to 37ppm), which contain lower amounts of Sc. This can be attributed to the gradual increase in plagioclase, and the simultaneous decrease in augite. Because augite did not crystallize within the felsic series, the quartz alkali syenite (6 ppm) and alkali granite (2 to 3 ppm) contain particularly low concentrations of Sc, and plot at the lower left end of the series trend.

The sample of diopside pegmatite (RB18) does not conform to the general geochemistry of the series trend, due to a far lower Sc content (30 ppm) compared to the Ol clinopyroxenite. This observation suggest that the pegmatite must have formed at later stages of crystallization, since most of the Sc is concentrated within the clinopyroxene of the lower ultramafic series.

Gallium (Ga), Barium (Ba), Rubidium (Rb) and Zirconium (Zr) are positively correlated with one another and illustrate a strong relationship with the SiO₂, Al₂O₃, Na₂O and K₂O content (Table 7 & Appendix E3). The wehrlite and adjoining Ol-clinopyroxenite (Appendix E1) contain the lowest concentrations of these elements (1 to 4 ppm Ga; 17 to 54 ppm Ba; 0.5 to 0.8 ppm Rb; 3 to 6 ppm Zr), while a gradual increase is observed within the rest of the ultramafic, mafic and dioritic rock types (8 to 20ppm Ga; 96 to 478 ppm Ba; 2 to 37 ppm Rb; 17 to 88 ppm Zr). The largest values for these elements (20 to 24 ppm Ga; 605 to 741 ppm Ba; 17 to 62 ppm Rb; 135 to 221 ppm Zr) are observed within the quartz alkali syenite and alkali granite.

Ga, Ba, Rb and Zr are highly incompatible with early crystallising basaltic magmas and will therefore remain within the differentiating melt during early fractionation (Best, 2003). Ga, Ba and Rb can be later incorporated within amphiboles, feldspars and muscovite, which crystallise in more silicic melts (Best, 2003), while Zr tends to remain within the residual fraction due to a higher field strength (Albarède, 2003; Winter, 2010). In silicates, Ga particularly replaces Al, whereas Ba and Rb will be more likely to exchange for K in K-feldspar and muscovite (Best, 2003).

The bivariate diagrams for Ba and Rb vs K₂O (Figure 50h & Figure 51a) particularly demonstrates the fractionation trends of feldspar throughout the differentiated rock series. The wehrlite, Ol clinopyroxenite and diopside pegmatite, group at the lower left end of the series trends (17 to 108 ppm Ba & 0.5 to 2 ppm Rb), signifying the strong incompatibility of these elements with olivine and clinopyroxene. The rest of the ultramafic, mafic and dioritic rock types that contain predominantly amphibole and plagioclase illustrated a higher compatibility and grouped further along the series trend (92 to 478ppm Ba & 5 to 37 ppm Rb). The strongest association for these elements are observed during the crystallization of microcline within the quartz alkali syenite and alkali granite (605 to 741 ppm Ba & 17 to 62 ppm Rb). Ba was particularly strongly fractionated in microcline and thus a larger concentration gap is observed between the felsic and mafic to dioritic rock types (Figure 50h) compared to the Rb concentration (Figure 51a). Ba also illustrated

a greater mobility than Rb, and is particularly elevated within the uraltite (478 ppm), which is attributed to chemical exchange with the quartz alkali syenite during contact metamorphism.

The Zr content was plotted against SiO₂ (Figure 51b), which demonstrated its gradual evolution within the melt during differentiation. The wehrlite and Ol clinopyroxenite that fractionated first, contain the lowest amount of Zr (3 to 6 ppm) while the mela-dolerite and mela-spessartite that is more primitive in terms of silica content contains larger amounts (45 ppm). Little to no difference is observed for the Opx-Hbl gabbro (17 to 33 ppm) and leuco-spessartite (47 to 58 ppm), which are comparable to the mela-spessartite. Higher values (70 to 88 ppm) are, however, observed for the altered rock types including the hornblendite, uraltite and diopside pegmatite. The quartz alkali syenite and alkali granite contained the largest values for Zr (135 to 221 ppm) as these rock types were the last to crystallize during differentiation, and the Zr was therefore the most concentrated within the residual fluid that solidified within these rocks.

4.6 DISCUSSION

The KC comprises of a considerable record of structural, igneous, and metamorphic development, which can be confirmed by the orientation and distribution of its different rock types, as well as established textural, mineralogical and geochemical trends. In correspondence with the research aims and objectives, results were primarily focused on providing detailed documentation of the main geological features of the Complex, in order to determine if the different intrusive rocks indeed form a chemically differentiated series, and as a result, can cause measurable changes in the soil chemistry and related woody vegetation.

The following discussion will aim to interpret the findings of the petrological study, by describing the apparent geological history of the Complex, as well as its implication for soil development. A detailed classification of the magmatic affinity and related interpretation of the petrogenesis of the Complex falls beyond the scope of this study. Thus, a petrogenetic model will not be presented.

4.6.1 Field relations, relative age and emplacement sequence

The oldest exposed body of altered mafic - dioritic igneous rock is confirmed to be epidiorite and is accepted to have formed as a conduit for the 2714 Ma old Ventersdorp lavas (Anhaeusser, 2006; Bisschoff, 1999; Stepto 1990). The sill was emplaced prior to the formation of the younger intrusive series and is genetically unrelated. Field measurements suggest that the epidiorite sill of the KC is separated by a zone of impact-related, northwest trending thrust faults (F₁, F₂ & F₃; Figure 9), along which the displaced and partially rotated a group of younger ultramafic to dioritic intrusions occur.

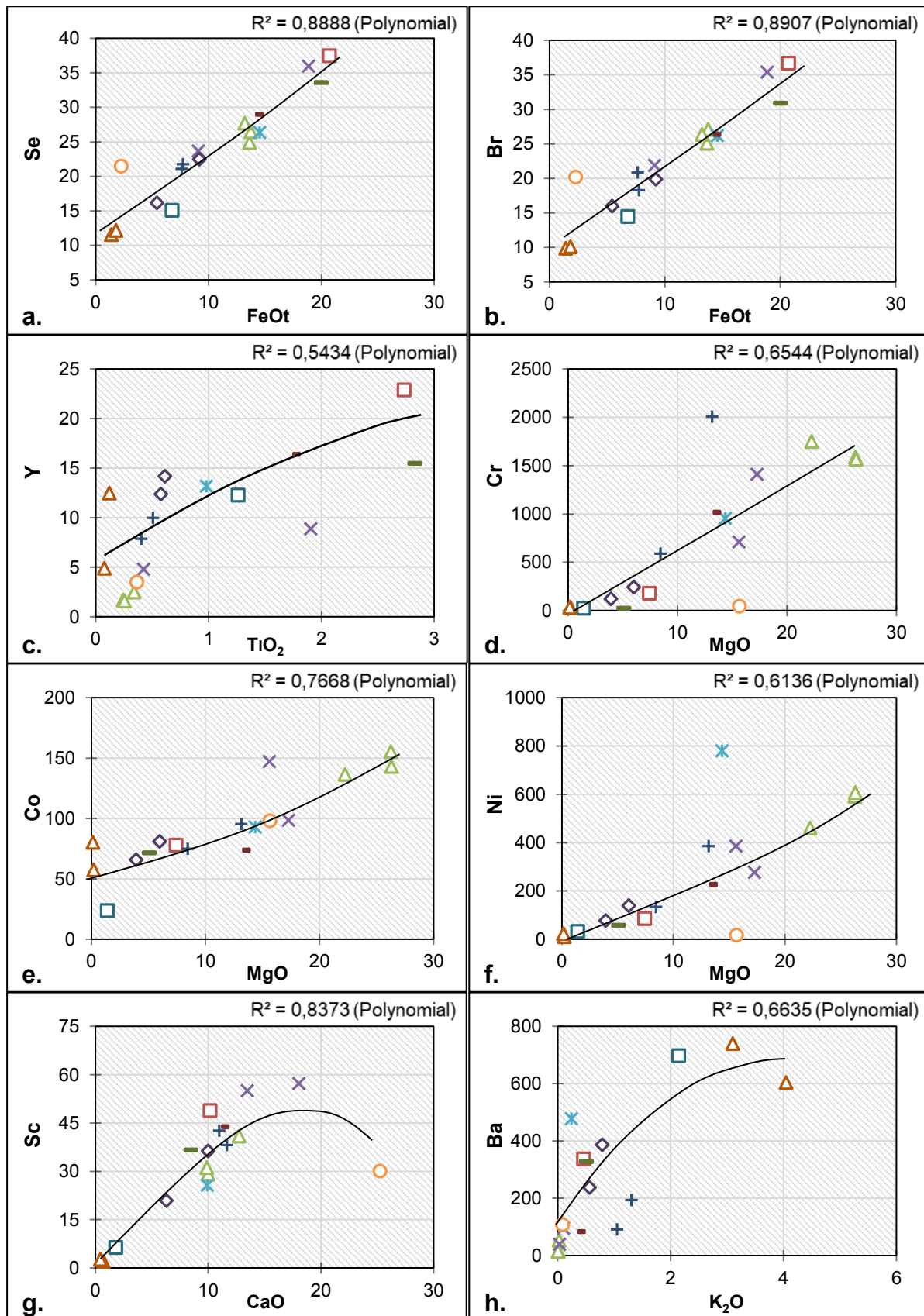


Figure 50: Bivariate diagrams for selected trace elements and major oxides of the syn-Bushveld intrusive series, depicting (a & b) Se and Br vs FeOt, (c) Y vs TiO₂, (d, e & f) Cr, Co and Ni vs MgO, (g) Sc vs Ca and (h) Ba vs K₂O. A sample legend is provided in Figure 49.

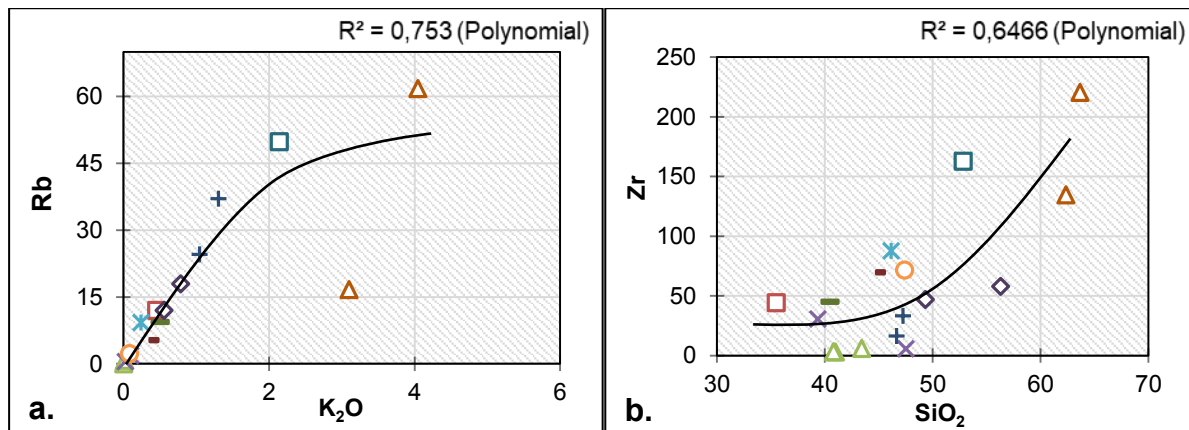


Figure 51: Bivariate diagrams for selected trace elements and major oxides of the syn-Bushveld intrusive series, depicting (a) Rb vs K₂O and (b) Zr vs SiO₂. A sample legend is provided in Figure 49.

The ultramafic, mafic and dioritic intrusions have been emplaced stratigraphically above the intrusion of epidiorite, prior to deformation, and are related to a later stage of magma emplacement. This notion was supported by their mineral textures (section 4.3) and geochemistry (section 4.5).

The latter intrusions are followed by a closely trending group of felsic alkaline intrusions that are primarily exposed on the north-eastern slope of the large quartzite ridge (Figure 9). Here, the felsic intrusions are observed to cut across portions of the ultramafic layer, which displays a corresponding degree of mineralogical and geochemical alteration (Uralitite - sections 4.4 & 4.5).

Macro- and microscopic veinlets of pseudotachylite and planar deformation features are well-known markers for impact-related deformation (Albat, 1988; Albat & Mayer, 1989; Elsenbroek, 1999; Gibson and Reimold, 1999; Reimold, 2006). These markers (Figure 15, 16, 23, 25 & Figure 39) confirm that the intrusions of the KC are indeed of pre-impact occurrence and are older than 2023 ± 4 Ma years (Kamo *et al.*, 1996).

Several authors (Anhaeusser, 2006; Bisschoff, 1999; de Waal *et al.*, 2006, 2008; Elsenbroek, 1999) indicated, based on petrographical observations, that strong connections exist between the younger intrusions of the KC and the surrounding intrusive complexes (section 2.2). Results from this study support this view, as the sill structured layering of the KC can be closely related to the intrusive bodies of the RTC, RC, LDC and SC as illustrated in Table 2, with several of the rock layers displaying very similar mineralogical and petrographical characteristics.

The modal layering of the wehrlite to Ol $\pm$ clinopyroxenite, the presence of a magnetite cumulate layer and the transition to a mafic/dioritic layer is similarly described for the RTC (Bisschoff, 1969, 1973, 1999).

The hornblende cumulate and porphyritic spessartite of the KC, are strikingly similar to dioritic samples found at RTC, RC and LDC (Anhaeusser, 2006; Bisschoff, 1969, 1972b, 1973, 1999; Clark, 1972; de Waal *et al.*, 2006, 2008).

The quartz alkali syenite of the KC is strongly comparable to the alkali-syenite or albite-syenite of the surrounding complexes (RTC, RC, LDC & SC). At the RTC and SC, the alkali-syenites are also observed to be closely associated with the alkali granite. (Bisschoff, 1969, 1972b, 1973, 1999; Elsenbroek, 1991, Graham *et al.*, 2005)

SHRIMP zircon dating of the diorite at the RC (2053.9 ± 9.2 Ma; de Waal *et al.*, 2006, 2008), the spessartite at the LDC (2054.8 ± 5.7 ; de Waal *et al.*, 2006, 2008) and alkali granite of the SC (2052 ± 14 Ma; Graham *et al.*, 2005), provided consistent ages with the Bushveld igneous event (2050-2060 Ma; Cawthorn, *et al.*, 2006), and due to their strong connection with the KC, this study regards a similar age constraint for the younger intrusions of this Complex.

4.6.2 Differentiation and alteration

Mineralogical analysis and petrographic observation indicated a strong genetic relationship between the entire series of syn-Bushveld aged ultramafic to felsic intrusions of the KC, which suggests that they were most likely derived from the same deeper seated source of parental magma that over time had evolved via differentiation. This notion is supported by their gradual transitions in mineralogy and associated geochemical trends, which illustrated strong correlation and regression values for major oxide and trace elements (Table 7, Figures 49 to 51 & Appendix E3).

A proposed lineage for the corresponding differentiated rock layers of the KC is displayed in Figure 52. Due to the cross-cutting relationship between the ultramafic and felsic sills, it is believed that emplacement of the syn-Bushveld age intrusions must have occurred in pulses. The mafic and dioritic rock layers display gradual transitions in mineralogy directly north (above in stratigraphy) of the lower layers of ultramafic rocks.

Although the transitional contact between the upper layers of $\pm$ Ol clinopyroxenite and the mafic layer of Opx- Hbl gabbro could not be verified (as with the RTC – Bisschoff, 1973), it can be deduced from their similar strike patterns, correlation in modal and phase layered sequences and close trending geochemistry, that the ultramafic, mafic and dioritic layers of the KC differentiated from a single pulse of derived magma, while the felsic intrusions must have crystallised from a second more evolved pulse. This deduction agrees with Bisschoff, (1973), which concluded that the alkali granite of the RTC must have crystallised separately, and most likely originated from a derived albite-rich alkali syenitic magma.

When the rock layers are rotated to their original horizontal position, as illustrated in Figure 52, the corresponding modal and phase layering sequences illustrate evidence for crystal settling and fractionation. This notion was confirmed by their cumulate textures (section 4.4) and geochemical trends (section 4.5).

The thin basal layer of mela-dolerite that occurs below the layer of wehrlite, appears to be the most primitive in terms of its SiO_2 content. The sharp contrast in grain-size between the mela-dolerite and wehrlite is an indication of the increased cooling rate, which must have dominated along the lower contact of the magma chamber. If the mela-dolerite is accepted to represent the lower chilled contact of the magma that had formed from the initial pulse, then its geochemistry can be assumed to be the closest to the initially derived magma (Pulse 1). With respect to the mela-dolerite's mineralogy, it can be deduced that the initial magma had a basaltic composition.

Magnetite, olivine and augite were of the first minerals to crystallise within ultramafic rocks, and are observed to have formed cumulate textures within the lower layers of wehrlite and $\text{Ol} \pm$ clinopyroxenite. The crystallisation of olivine appears to have gradually ceased within the upper division of $\text{Ol} \pm$ clinopyroxenite, while augite and accessory magnetite continued to crystallise throughout the rest of the mafic to dioritic rock layers. Fractionation of these phases corresponded with the observed trends for FeO, Mn_3O_4 , MgO, CaO, Se, Br, Cr, Co, Ni and Sc, which strongly partitioned within these minerals.

The accumulation of magnetite, olivine and augite must have caused isolation of the melt that produced the layer of mela-dolerite. This would explain why such contrasts in mineralogy and geochemistry are observed between the mela-dolerite and wehrlite layer. Higher concentrations of SiO_2 , MgO, CaO, Cr, Co, Ni and Sc were observed within the wehrlite due to the fractionation of mafic silicates, while incompatible components such as Al_2O_3 , Na_2O , K_2O , P_2O_5 , Y, Rb and Zr remained within the melt and therefore attained higher concentrations within the layer of mela-dolerite.

Elements that strongly partitioned into magnetite (Fe, Mn, Ti, Se & Ba) were also higher in the mela-dolerite than the wehrlite. This would suggest that the magnetite that had sunk to the bottom of the chamber, enriched the isolated portion of melt before crystallisation occurred.

Plagioclase crystallised within the middle to the upper parts of the magma chamber. This observation is consistent with the tendency of plagioclase crystals to be suspended or slowly rise in basaltic magmas, which will eventually cause them to accumulate above heavier mafic silicates (Gill, 2010).

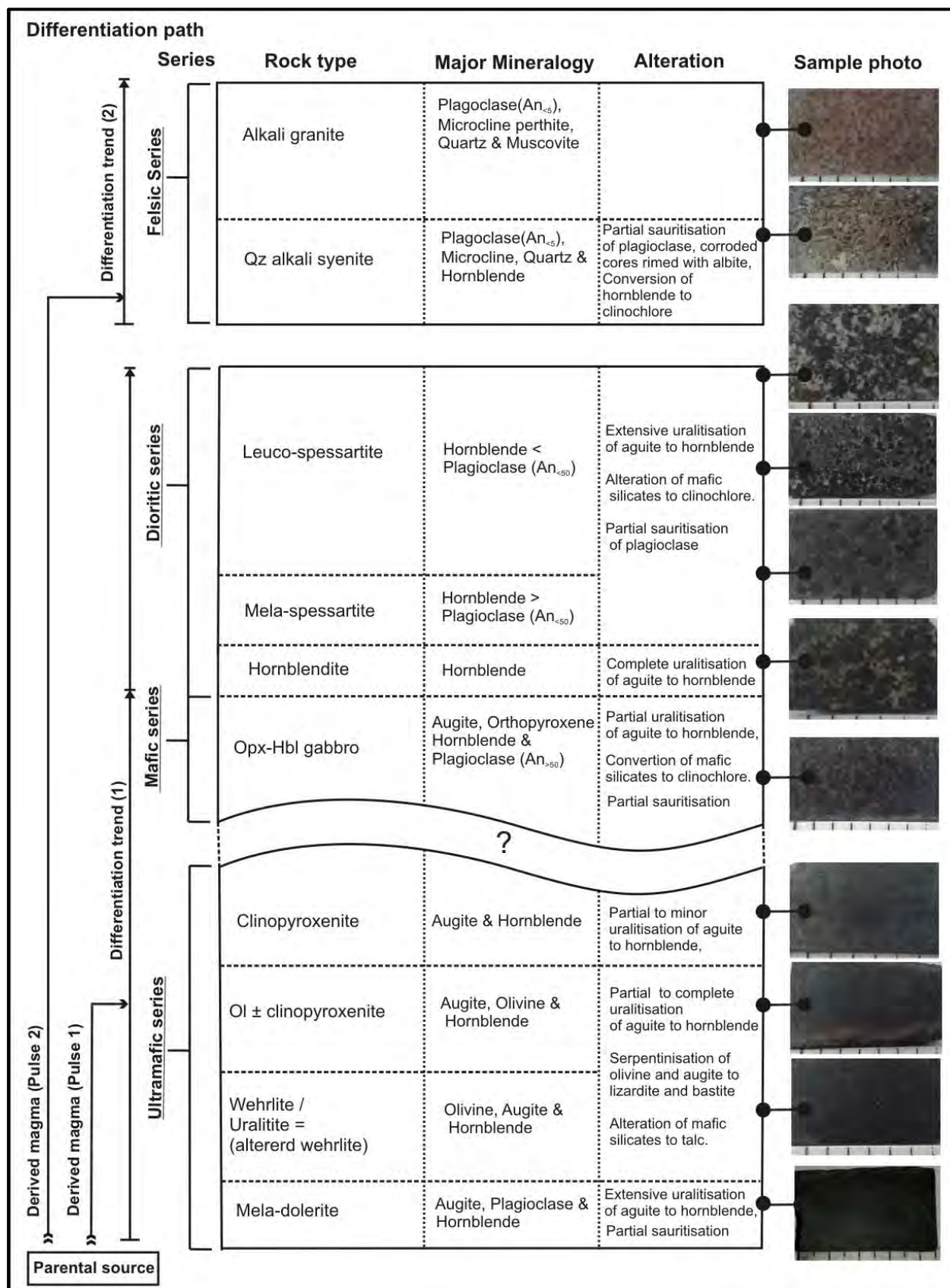


Figure 52: A proposed lineage and stratigraphical summary for the syn-Bushveld igneous rock series of the Koedoesfontein Complex. Layer thicknesses are not according to scale.

Some initial plagioclase formed as interstitial grains in the upper parts of the Ol ± clinopyroxenite layer, while the bulk of the plagioclase is observed to have formed along with augite and hornblende within the finer-grained transitional layer of Opx-Hbl gabbro, and upper layers of spessartite. The plagioclase of the gabbro layer crystallised as labradorite, while in the upper layers of hornblendite and spessartite, the melt became richer in Na and formed andesine to oligoclase.

The secondary replacement of augite with late-stage magmatic hornblende (known as uralitisation - Buseck *et al.*, 1980), reflects the build-up of dissolved water and incompatible elements (Al, Na & K) during the evolution of the melt (Gill, 2010). This is regarded as deuteric alteration which is a type autometamorphic process, which involves the late stage hydration of minerals (Best, 2003; Winter, 2010).

Autometamorphic processes from a natural part of igneous cooling and are regarded separately from distinct later alteration events (Winter, 2010), such as the metamorphism caused by the younger felsic intrusions. The summary of mineral alterations included in Figure 52 regards both autometamorphic and metamorphic processes that will be discussed. Apart from uralitisation, evidence for deuteric alteration includes the partial to severe saussuritisation of the plagioclase grains (Figure 33, Figure 42 & Figure 43), as well as the presence of clinocllore (Appendix D2).

Lamprophyres such as spessartite, are characteristic to originate from alkaline melts containing particularly high volatile content of H₂O and have been known to develop from natural differentiation processes (Winter, 2010). Since anhydrous minerals were the first to crystallise and accumulate at the bottom of the magma chamber, the volatile H₂O content must have increased within the melt during fractionation. In this process, some of the H₂O content was also probably derived from heating and assimilation of the sedimentary host rocks.

When magmas reach a specific point of saturation, a hydrous vapour phase will develop that diffuses through the melt and concentrates near the top of the magma chamber (Winter, 2010). H₂O lowers the degree of SiO₄-tetrahedra polymerisation within a magma and will enhance diffusion rates causing coarse grained textures or pegmatites to develop (Best, 2003; Gill, 2010; Winter, 2010). This would explain why an increase in the hornblende's grain-size is observed towards the upper dioritic rock layers, and why coarse-grained schlieren is present within the leuco-spessartite.

During the formation of the Opx-Hbl gabbro, uralitisation of the augite commenced after the initial crystallisation of plagioclase, replacing the augite grains from the border inwards (Figure 42 & 43). In the spessartite layers, it seems that hornblende and plagioclase have crystallised

simultaneously since plagioclase is enclosed by expanded grains of hornblende, but also mantles these grains as well (Figure 45).

The coarser-grained hornblende which encloses the smaller grains of plagioclase indicates a difference in growth rate. This can be attributed to differences in melting point and element diffusibility, which in the spessartite layers appears to have favored the rapid growth of hornblende. The dominance of small plagioclase grains that surrounds these hornblende grains (Figure 45), can be regarded as the result of the surrounding interstitial melt being deleted in chemical elements contributing to the formation of hornblende. This caused plagioclase to form directly around the hornblende grains producing hornblende oikocrysts.

The larger size and higher density of hornblende grains, caused them to sink and form a cumulate layer on top of the Opx-Hbl gabbro. This explains why a rapid transition in grain size is observed. In the layers of spessartite, the hornblende grains appear euhedral (Figure 45 & Figure 46) while in the lower cumulate layers, the crystals appear jagged and skeletal (Figure 44). This can be attributed to a higher magma temperature within the centre of the chamber, which caused the accumulated hornblende to be partially reabsorbed within the melt.

The difference in crystal growth rate can partly explain why spessartite possesses a porphyritic texture but does not fully clarify why such a fine matrix of hornblende is observed within the lower layers (closer to the warm centre of the chamber). Rapid crystallisation of the remaining melt must have ensued during the sudden release of H₂O, which will create a rise in the liquids temperature (Winter, 2010), as well as an aplitic texture. This is commonly caused when the roof rocks on top of the intrusion are fractured during an increase in pressure (Winter, 2010).

The mela-spessartite and hornblendite are chemically more evolved than the mela-dolerite, but more primitive than the Opx-Hbl gabbro. The stratigraphical sequence and geochemistry of these layers can therefore not be solely contributed to fractionation. To explain this feature, it is suggested that spessartite formed part of an isolated H₂O-rich boundary layer, which became compositionally stratified within the upper parts of the magma chamber, as proposed by Hildreth (1979). Near the interior of the chamber, where the hornblendite and mela-spessartite crystallised, the magma was still mostly molten remaining more primitive than the differentiating portion below, while increasingly lighter and more evolved magma within the boundary layer crystallised near the upper portion of the chamber and formed leuco-spessartite.

From the borehole evidence provided by Bisschoff (1973), and the samples obtained from field investigation, it can be concluded that the second pulse of derived magma, was partly emplaced discordantly to the differentiated layers of the first pulse. This produced a corresponding degree of metamorphic alteration, which is particularly noticeable in rocks of the ultramafic series.

The sample of uraltite illustrated a similar concentration in compatible trace elements (Se, Br, Cr, Co, Ni & Sc) than the wehrlite, and is therefore regarded as a portion of the wehrlite sill which was extensively altered due to contact metamorphism. Evidence for chemical exchange with the second magma pulse is provided by the uraltite's geochemistry (Appendix E1). Apart from being enriched with early partitioned trace elements, the uraltite also contains a high amount of incompatible elements (Ti, Al, Y & Ba). These elements were the most concentrated within the more evolved rocks of quartz alkali syenite and alkali granite and had thus enriched the portion of wehrlite during alteration to hornblende.

Serpentinisation of the lower ultramafic layers also serves as evidence for a metamorphic event. The conversion of the olivine to lizardite and the corresponding formation of batiste within the augite grains indicates that the lower layers must have been hydrothermally altered (Best, 2003; Gill, 2010). This is regarded as the result of an underlying alkali granite pluton, which must have caused metasomatism of the older ultramafic layers.

In accordance with the observed stratigraphical sequence, the quartz alkali syenite was established to be more primitive than the overlying layer of alkali granite and was, therefore, the first to form during differentiation of the felsic series. Results correspond with the findings of Elsenbroek (1991) and Graham *et al.*, (2005) which concluded that the alkali syenite of the SC is more primitive than the alkali granite. It would thus appear that the alkali granite must have evolved from a more syenitic melt as Bisschoff (1973) originally suggested.

4.6.3 Structural deformation

As indicated in section 4.2, it could be deduced from the orientated portions of the KC's intrusive sills and their relationship with the surrounding quartzite layers, (Figure 1 and Figure 9), that the Complex underwent two separate stages of structural deformation.

The first stage is noted on a larger scale and is marked by the two major faults occurring to east and west of the Complex (Figure 1). Wieland *et al.*, (2005) proposed that these faults are of pre-impact age, and is believed to correspond with Ventersdorp age rifting. Research observations from this study support this view and suggest that a portion of the Main Quartzite Formation was already displaced discordantly within the layers of the lower West Rand Group, prior to the emplacement of the epidiorite and syn-Bushveld intrusions. Bisschoff (1973) suggested that these particular fault zones must have played a significant role in the emplacement of the intrusions of the KC, and also makes a similar conclusion for the SC, which is believed to have been emplaced similarly along a large strike-slip fault of pre-impact age. This notion was later supported by Elsenbroek (1991).

The second stage of deformation is related to the impact event and can be deduced from a smaller scale (Figure 9). The sills are overturned to near vertical, dipping to the south-west. If the outcrops are connected, the layers provide evidence for small scaled compressional folding and signs of related thrust faulting that are consistent with impact deformation (F_1 , F_2 & F_3 ; Figure 9). In addition, several minor faults were verified to cut across the emplaced intrusions of quartz alkali syenite (Figure 23). These faults were also interpreted as impact-related, and thus confirm that the quartz alkali syenite is of pre-impact occurrence.

Ultimately the folded nature of the KC's intrusions and their displaced occurrence, due thrust faulting, explains why they are so irregularly distributed within the field. This feature contests to the sure difficulty that must have posted previous attempts at mapping this complex without the advantages of GPS and high definition satellite imaging.

4.6.4 Implications for soil development

Field observations and analysis confirmed the presence of mineralogical and geochemical gradients that occur throughout the different geological substrates of the KC (Appendix D2 & Appendix E1). In correspondence with the topography and geological association, the topsoil of the study area was selectively divided into geobotanical units representing possible mineral-soil relationships (section 3.2; Figure 6). From the presented results, the following deductions can be made with regards to the geology and general soil association within each of these units:

The alluvial floodplain (Geobotanical Unit A) and exposed quartzite ridges (Unit C) are primarily composed of quartz-rich country rock, containing accessory amounts of hematite and muscovite. These units contain the least amount of mafic minerals, as well as feldspars, and thus, their derived soil products can be expected to illustrate lower nutritive capabilities since quartz does not contribute any nutritional elements and weathers far less rapidly compared to mafic silicates and feldspars (Boggs, 2010). Accessory hematite and muscovite within the quartzite bedrock will provide some source of K and Fe within the soils, however, these amounts are presumed to be less compared to units that in addition contain intrusive rock types (Units D1, D2 & D3).

The southern shaded quartzite (Unit B) is similarly dominant in quartz, nonetheless, minor outcrops of hornfels contribute minor sources of amphibole and Ca-plagioclase, which can relate to increased concentrations of Ca, Mg, Na, K and Fe. The southern aspects are also particularly steep compared to the northern quartzite slopes, and increased surface runoff and related erosion rates can cause a higher degree of mineral weathering and soil formation, although the deficiency of primary sources for the development of clay minerals would be apparent with the lack of soil structure and nutrient retention. (Proctor, 2003 and Spargo *et al.*, 2013).

The geobotanical units that demarcate the intrusive rock types (Units D1, D2 & D3) comprise a larger amount of mafic minerals, feldspars, apatite and Fe-Mn-oxides. These minerals provide abundant sources of nutritional elements including Ca, Mg, Na, K, P, Fe and Mn (McGrath & Smith, 1990; Mengel & Kirkby, 2001), although they are not evenly distributed as a result of fractional crystallisation, related layering and subsequent deformation.

Variance in the dominant rock types and the corresponding mineralogy are expected to cause specific element toxicities and/or deficiencies to occur. This characteristic has been particularly identified due to the presence of ultramafic intrusions (section 2.3), which are known and have deficiencies of Ca, K and P, and contain toxic values of Mg, Cr, Co and Ni (Brady *et al.*, 2005; Kabata-Pendias, 2011; McGrath & Smith, 1990; Mengel & Kirkby, 2001; O'Dell *et al.*, 2006; Robinson *et al.*, 1997).

The mafic zone (Unit D1), is established to contain ultramafic cumulates of wehrlite, Ol $\pm$ clinopyroxenite and hornblendite. These rocks comprise primarily of olivine, augite and hornblende as well as accessory magnetite, serpentine, talc and clinochlore. Chemical analyses indicate that these rocks will indeed provide very high amounts of Mg, Fe and Mn to soil substrates, and in addition contain high concentrations of Cr, Co and Ni (Appendix E1) that are considered as phytotoxic for vegetation (Kabata-Pendias, 2011, Mengel & Kirkby, 2001). Because of the rare occurrence of accessory felsic minerals, the wehrlite and Ol $\pm$ clinopyroxenite specifically contain the lowest concentrations of Na, K and P, and will also offer the least amount of these elements to soil substrates compared to the other intrusive rock types.

High Ca values were observed in the wehrlite and Ol $\pm$ clinopyroxenite and is attributed to augite that is the primary source of Ca within the ultramafic rocks. Augite contains less Si, Al and Na than does plagioclase or hornblende, and, therefore, a slightly higher relative abundance of Ca is observed than with the mafic and dioritic rock types. Nevertheless, due the presence of olivine and serpentine the Mg content is still relatively more abundant than Ca and thus, a lower soil Ca:Mg ratio could be expected (Hazelton & Murphy, 2007; Mengel & Kirkby, 2001).

With regards to their ultramafic character, the wehrlite and Ol $\pm$ clinopyroxenite certainly illustrate the properties that could cause the serpentine effect in soil (Brady *et al.*, 2005; O'Dell *et al.*, 2006), yet the close occurrence of the mafic and dioritic intrusions (also grouped in Unit D1) will most likely result in a collective contribution of soil characteristics, due to the intermixed nature of their weathered material. In addition to mafic minerals and magnetite, dolerite, gabbro and spessartite layers provide Ca-Na-plagioclase, muscovite and apatite that contributes additional sources of Ca, Na, K and P.

The dioritic zone (Unit D2) encompasses primarily the outcrops of epidiorite, which contain no olivine or serpentine, less clinopyroxene and more Mg-hornblende. As a result, this rock type provides significantly lower concentrations of Mg, Fe and Mn during soil formation, whereas the Ca content is more or less similar to the mafic rock types of Unit D1 (Appendix E1), due to the presence of Na–Ca–plagioclase. Additionally, because of the presence of Na–Ca–plagioclase, a higher amount of Na is available during weathering, while the scarcity of apatite and the absence of K-feldspar is noted by the lower rock concentrations of K and P.

In the felsic zone (Unit D3), the felsic intrusions of quartz alkali syenite and alkali granite are composed primarily of Na-plagioclase, K-feldspar and muscovite, which dominates over small quantities of Fe-hornblende (Appendix D2). These rock types will contribute the least amount of Mg, Fe and Mn to the soil, and is also more deficient in Ca, owing to the more albitic character of the plagioclase. Equally, owing to the dominance of Na-plagioclase, as well as higher amounts of K-feldspar and muscovite, these rock types are the most concentrated sources of Na and K. The P concentration seemed to be elevated only within the quartz alkali syenite, due to the presence of apatite, whereas the alkali granite provides P concentrations that are similar to the ultramafic intrusions. The small portion of altered uraltite that also occur in Unit D3 is composed primarily of amphibole and will contribute additional small sources of mainly Ca, Mg, Fe and Ni.

4.7 CONCLUSION

From the re-evaluation of the geology, it can be concluded that intrusions of the KC formed during the successive emplacement of three sill structured bodies of layered igneous rocks. The bodies are established to have been emplaced before the Vredefort impact event occurred (2023 ± 4 Ma years), and could be relatively dated by their the cross-cutting relationship, as well as the presence of oblique-thrust faults, pseudotachylite veinlets and prominent macroscopic planar deformation structures, which are established markers for impact-related deformation (Albat, 1988; Albat & Mayer, 1989; Elsenbroek, 1999; Gibson and Reimold, 1999; Reimold, 2006).

The epidiorite sill, based on previous studies (Anhaeusser, 2006; Bisschoff, 1999; Stepto 1990), is accepted to be related to the Ventersdorp lavas (2714 Ma). The succeeding intrusions are assumed to be of syn-Bushveld age (2050 - 2060 Ma). This was supported by mineralogical analysis and supporting petrographical observations, which indicate strong similarities with the layered intrusions of the surrounding complexes that date within this time constraint (Anhaeusser, 2006; de Waal *et al.*, 2006, 2008; Graham *et al.*, 2005).

XRD and XRF analysis of the syn-Bushveld age sills and their derived layers could confirm that they are genetically related and originated from differentiation. Based on their mineralogical and geochemical trends, the sills formed during two separate pulses of derived magma. Differentiation

trends for these pulses are ultimately attributed to a mixed process of fractional crystallisation, assimilation, volatile transport and the in situ formation of a compositionally stratified boundary layer.

Due to their large variance in chemical composition and mineral stability, it is believed that weathering of the rock layers of the KC will cause similar chemical gradients to occur within the related soil profile. Changes in the dominant rock types of each geobotanical unit will, therefore, be expected to have related affects on the abundance and availability of specific soil nutrients. With respect to plant nutrition, very high amounts of Fe, Mn, Mg, Cr, Co and Ni were observed for the ultramafic intrusions, and are expected to cause toxicity within the related soil of geobotanical Unit D1.

CHAPTER 5

SOIL CHEMISTRY OF THE KOEDOESFONTEIN COMPLEX

5.1 INTRODUCTION

The aim of this chapter is to convey and discuss particular associations identified amongst the differentiated igneous and metamorphic rocks of the KC and the chemistry of its distinctive topsoil. The chemical attributes of the soil layer (especially alluvial mineral soils) are largely derived from the geochemistry of the source material, which in turn, influences plant nutritional characteristics. These associations will help to recognise and possibly clarify why particular floristic distribution patterns are observed on the Complex (as discussed in Chapter 6).

Soil chemical parameters particularly indicative of ultramafic rock associations (as discussed in Chapter 2) are the main focus of this chapter. Variations in pH, CEC, SOC, nutrient status, exchangeable Ca:Mg ratio and trace element concentration are distinguished for the sampled geobotanical units (See Figure 6, Chapter 3 for localities and unit stratification) to test whether the abrupt transitions in the geochemistry of the parent material, are indeed expressed to some extent in the soil chemistry. Specific elements including Ca, Na, Fe, Cr, and Ni showed the strongest association with changes in specific lithologies, and are concluded to be as the main soil nutrients and trace elements that are primarily affected by changes in geology.

XRF and soil chemical analyses results of soil composite samples “KKSA” will be presented, hence with the exception of Fe, only total values are reported for the micronutrients. For raw data, refer to appendix E2 and F1 respectively. The total and soluble metal analysis results for composite samples “KSB” will be presented and discussed in Chapter 6.

5.2 pH AND BASE SATURATION

Results

Soil samples from the alluvial floodplain (Unit A), the upper parts of the shaded quartzite ridge (Unit B) and the exposed quartzite ridges (Unit C) had the lowest pH values, ranging between 4.0 and 5.8 (Appendix F). The calculated mean for each unit was ~5.5, ~5.3 and ~4.0 respectively (Figure 53). A strong correlation was seen between the soil pH and percentage base saturation (Figure 54), and as such, samples from Units A, B and C illustrated moderate to low values of 71.68 to 75.70%, 54.16%, and 21.36% base saturation respectively.

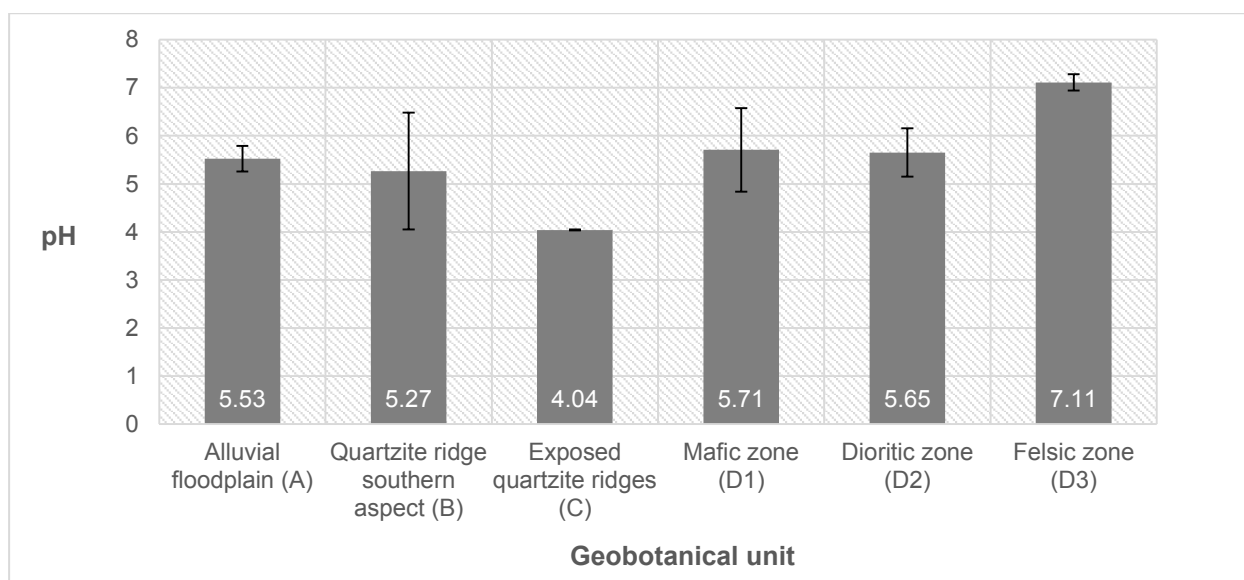


Figure 53: Mean soil pH for the geobotanical units. Lowest values (pH<5.5) are observed in Units A, B and C, situated on quartzite country rock, followed by slightly higher pH values (~5.5 and ~5.7) in Units D1 and D2 containing ultramafic-dioritic intrusions. Unit D3, containing felsic intrusions, shows the highest average pH (~7.1).

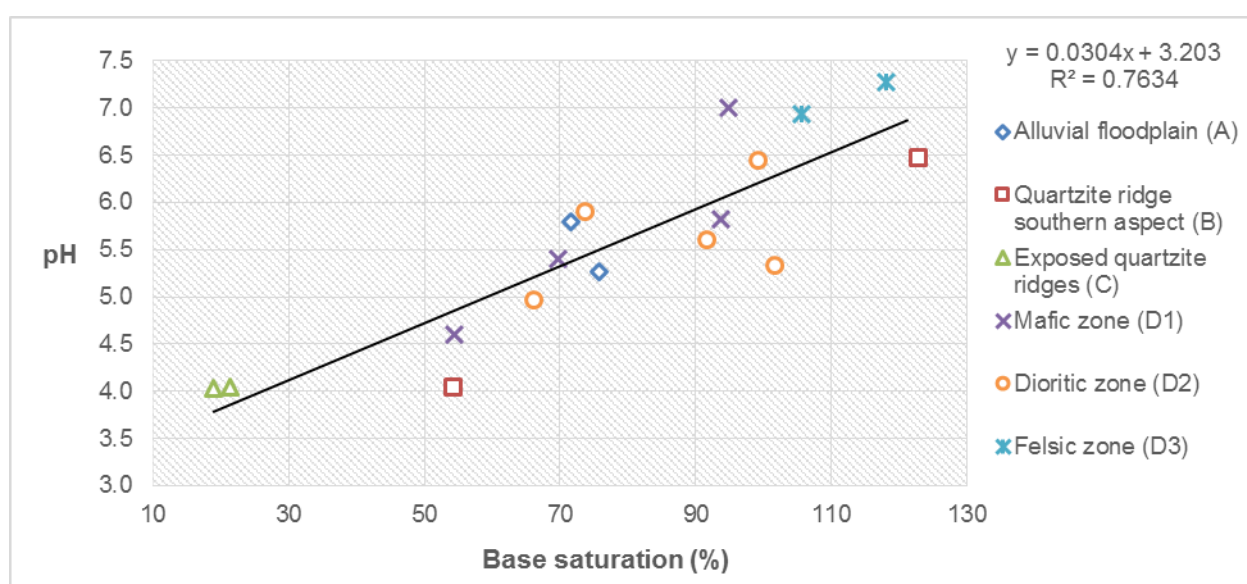


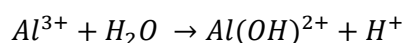
Figure 54: Positive linear correlation between soil pH and percentage exchangeable basic cations (Ca, Mg, K, and Na). Lowest base saturations (~20 to 60%) are seen in the upper parts of Units B, and C. Units A, D1, and D2 illustrate moderate to high values (~20 to 60%), whereas the highest occurrences of basic cations (100 to 123%) are observed in the lower parts of Units B and D3.

The mafic zone (Unit D1) and the dioritic zone (Unit D2) had higher pH values, ranging from 4.6 to 7.0, with the majority of the samples averaging ~5.7 and 5.6 within each unit. The range in base saturation of these units was accordingly higher, 54.44 to 94.92% and 66.21 to 101.69%.

Samples from the base of the shaded quartzite ridge (Unit B) and the felsic zone (Unit D3) had the highest pH values of ~6.5 and 7.3, respectively. A particularly high pH range (~6.9 to 7.3) was observed in Unit D3 with a sample mean of 7.11. These soils had a corresponding base saturation of >100% (Figure 54) and were over-saturated with basic cations (specifically Ca in Unit B - 122.68%, and Ca and Na in Unit D3 - 105.67 to 118.08%).

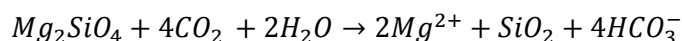
Discussion

Variations in soil pH and base saturation among the different geobotanical units are ascribed to differences in the mineral composition of the source material, and the related rate of chemical weathering. Soil pH or acidity is determined by the concentration of dissolved H^+ in the soil solution (pore water) and is responsible for the active acidity in soils (Mengel & Kirkby, 2001 and Hazelton & Murphy, 2007). The exchangeable acidity of a soil refers to exchangeable H^+ and Al^{3+} ions, which are adsorbed on the colloidal surface of soil particles. Al^{3+} is included as it produces H^+ during hydrolysis:



The active acidity and exchangeable acidity are in equilibrium, and interact through buffering systems (Sparks, 2003).

During weathering of silicate minerals, concentrations of bicarbonate ions along with different basic cations (Mg, Ca, Na and K) are increased within the soil solution. This can be illustrated, for example, by the following net reaction of olivine weathering (adapted from Eby, 2004):



Silicate minerals that generally weather more rapidly will, thus, produce higher amounts of bicarbonate ions, and the related soil will have a higher pH. During this reaction, some of the bicarbonate ions are neutralised by soil acids, and converted to carbon dioxide, whereas the remainder can precipitate *in situ* as magnesite, or be leached from terrestrial systems. In conjunction with the increase in pH, the concentration of basic cations (Mg, Ca, Na and K) will also be increased. Consequently, when the concentration of basic cations is increased, a large number of these cations will occupy the exchange sites of soil particles, through buffer reactions with the soil solution (Mengel & Kirkby, 2001; Sparks, 2003; Hazelton & Murphy, 2007). The number of exchange sites that are ultimately occupied by Ca, Mg, K and Na is expressed by a percentage value known as the base saturation. The base saturation is, thus, closely related to the pH and accordingly illustrates a linear correlation (Figure 54). Therefore, soils that originate from minerals that weather more rapidly, and contain supplementary sources of basic cations, will

have a higher pH and related base saturation than do soils that originate from more Si- and Al-rich minerals.

Within Unit A, the upper parts of Unit B, and Unit C, quartzite is the prevailing rock type. Quartzite contains relatively low basic cations within its mineralogy and generally weathers slower than the intrusive rock types. During weathering and soil formation, the majority of these cations are lost during leaching, and the related soil has a low base saturation and corresponding pH (Figure 54). Soils from these units will also have increased availability of Al, which can become toxic at very low pH values (<4.7 - Hazelton & Murphy, 2007).

Units D1 and D2 contain ultramafic, mafic and dioritic rock types, with increased amounts of basic cations (particularly Ca and Mg). The related soils correspondingly illustrated a higher pH and base saturation, as increased weathering and mineral sources of these cations can compensate for leaching. At these pH ranges (between 5.0 and <9.0) macronutrients are also generally more available. However, trace elements specifically Cr will mostly occur as insoluble compounds (pH >5.5 - McGrath & Smith, 1990).

Observations made from the base of Units B and D3 can be attributed to higher weathering rates, and greater soil CEC values (Appendix F). Because of higher moisture content and the presence of hornfels on the southern slope of the quartzite ridge (Unit B), and high concentrations of Na, K and P in the felsic intrusions within Unit D3, these units are associated with the highest vegetation density and related biomass production, ultimately resulting in increased soil organic material that increases the soil water retention and CEC, as well as additionally releasing nutrients and organic acids during decomposition (Sparks, 2003). This results in increased weathering and the release of more cations in the soil solution. In addition, because of the higher CEC, a greater amount of these cations are also retained against leaching. This also explains the higher base saturation (owing to Ca accumulation) at the base of the shaded quartzite ridge (Unit B), which acts as a sink for leached cations from the weathered hornfels, as well as rock types from the adjacent units (D1 and D2).

It is clear that the weathering rate of the igneous intrusions, as well as the soil water regime (related to the microclimate) has a major effect on the related soil pH and base saturation. Soils from units containing intrusive rock types are characterized by a higher base saturation and pH ranges from ~5 to 7, and will accordingly have a higher macronutrient availability, but lower trace element availability. Whereas soils from units containing only siliceous quartzite country rock generally have a lower base saturation and pH ranges from ~4 to 5. The shaded quartzite ridge (Unit B) however is the exception due to the presence of hornfels, higher weathering rates associated with the water regime, and the accumulation and retention of leached nutrients.

5.3 CATION EXCHANGE CAPACITY (CEC) AND SOIL ORGANIC CARBON (SOC)

Results

Soils from the alluvial floodplain (Unit A) had the lowest CEC values of 8.85 and 10.34 cmol(+)/kg (Appendix F), with an average of 9.6 cmol(+)/kg (Figure 55). A strong correlation was observed between the soil's CEC and SOC content (Figure 56), which generally indicated a moderate carbon content of 1.22 and 1.23% for Unit A.

The soils of the exposed quartzite ridges (Unit C), the mafic zone (Unit D1), as well as the dioritic zone (Unit D2) had slightly higher CEC values, ranging between 12 and 18 cmol(+)/kg, with respective means of 13.44, 12.87 and 14.89 cmol(+)/kg. The SOC content showed similarly increased values of 3.56 to 3.61%, 2.14 to 2.38% and 2.85 to 3.67%.

The southern shaded quartzite (Unit B) and the felsic zone (Unit D3) had the highest soil CEC values, ranging between 16 and 24 cmol(+)/kg. Sample averages for each unit were 21.42 and 20.05 cmol(+)/kg respectively and contained corresponding SOC values of 4.63 to 5.41% and 3.48 to 4.65%.

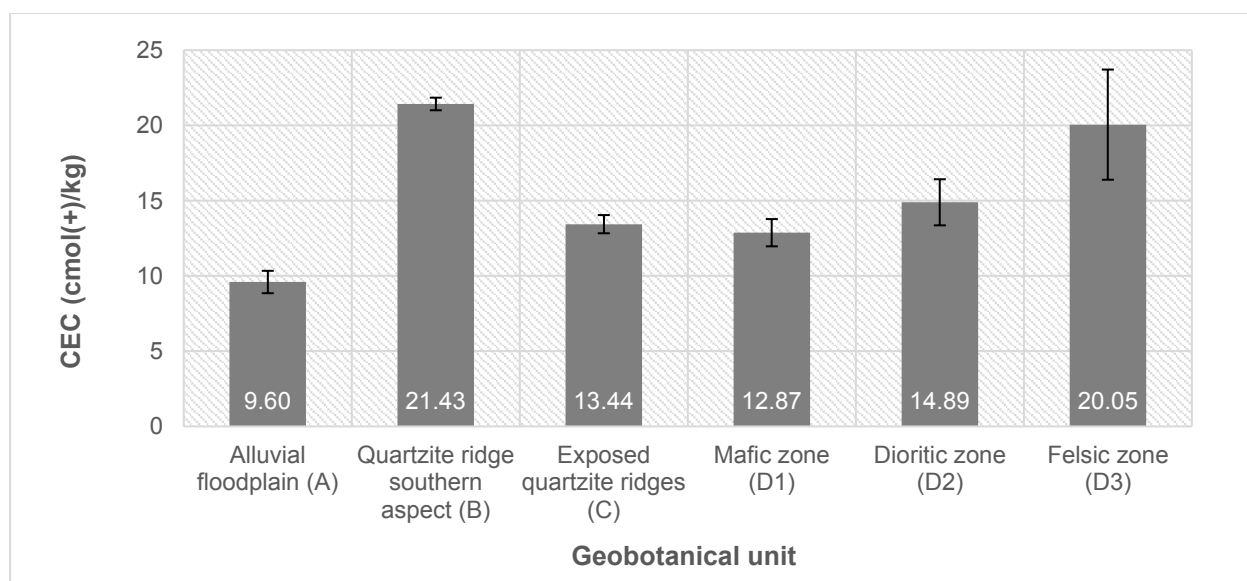


Figure 55: Mean soil CEC for the geobotanical units. Unit A illustrates the lowest value of <10 cmol(+)/kg, whereas Units C, D1, D2 were observed to have higher values ranging from ~12 to <15 cmol(+)/kg. The highest soil CEC occurs in Units B and D3, ~20 to <22 cmol(+)/kg.

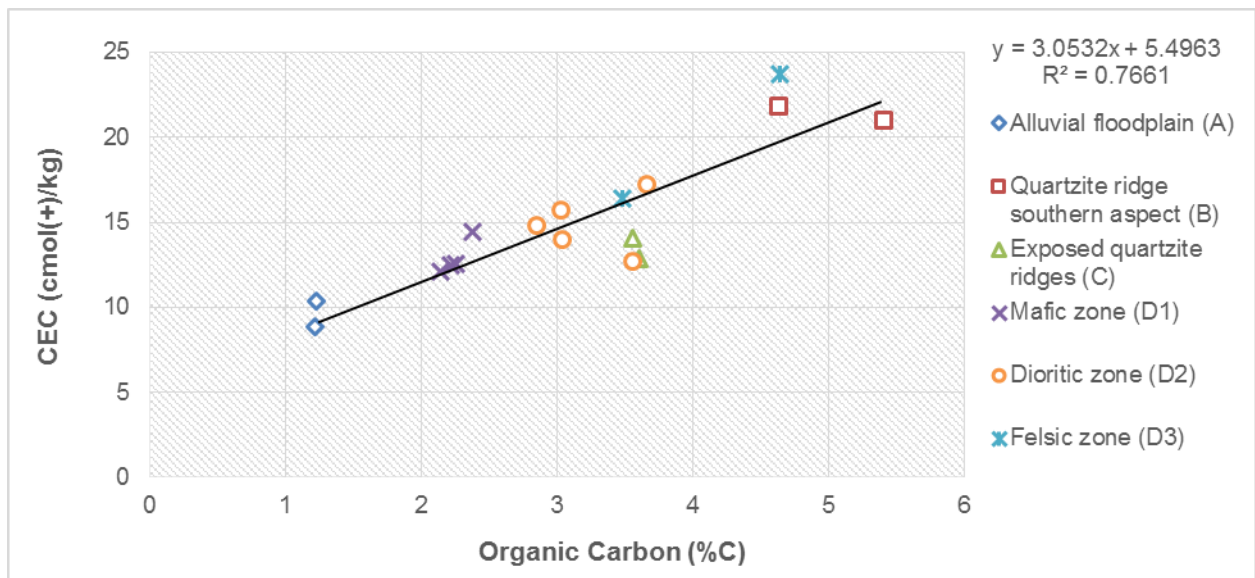


Figure 56: Positive linear correlation between soil cation exchange capacity (CEC) and soil organic carbon (SOC). The lowest SOC content of <1.3% is observed in Unit A, while Units C, D1, and D2 illustrates higher values between ~2 and 3.7. Units B and D3 have the highest indicated values ranging from ~3.4 to 5.4.

Discussion

As discussed previously (section 5.2), the CEC provides a buffering effect against pH changes and available nutrients (Hazelton & Murphy, 2007), and is thus a major controlling factor of soil fertility, and ultimately plant growth. The soil's CEC is a function of the number of negatively charged sites on soil particles, which originate primarily from clay minerals, as well as carboxylic- and phenolic hydroxylic groups in soil organic material (SOM) (Mengel & Kirkby, 2001).

Soils of the KC contain relatively low concentrations of clay minerals since large fractions of the soil are derived from the moderate- to very mature quartzites hosting the various thin igneous outcrops. Therefore, the majority of the soil CEC is contributed to SOM. This observation was confirmed by the linear relationship ($R^2=0.7661$) between the CEC of the soil samples and their percentage SOC (Figure 56).

The SOC gives an indication of the amount of SOM contained within the soil; thus, the CEC, in this case, is closely related to the density and biomass production of the vegetation supported by the soil, which in turn is controlled by the pH, availability of nutrients and microclimate. The CEC, therefore, gives a good indication of the overall soil fertility and vegetation productivity within each geobotanical unit.

The low CEC values observed in Unit A can be correlated with the lack of SOM, which is related to nutrient leaching (Section 5.4.1), and low density of woody vegetation (Appendix C1).

Higher CEC values observed in Unit C could mainly be due to lower weathering rates of the quartzite bedrock, resulting in higher accumulation rates of organic material, than for inorganic particles. A higher SOC concentration was observed; nonetheless, the nutrient status and vegetation density remained relatively low, similar to those in Unit A.

Higher CEC values in Units D1 and D2 can be ascribed to a higher nutrient status and related vegetation density, because intermediate to ultramafic rock types generally weather faster than quartzite and contain more nutrient-rich minerals. However, because of lower availability of Ca (Figure 57 and Figure 58) and N (Figure 59) and higher concentrations of total Fe (Figure 60), Cr and Ni (Figure 62), their supported vegetation productivity is still negatively affected, because the soils of the southern shaded quartzite (Unit B) and the felsic zone (Unit D3) had the highest vegetation density, SOC content, and correlated CEC values.

5.4 MACRONUTRIENTS (Ca, Mg, K, Na, P, N)

5.4.1 Nutrient availability

The mean macronutrient status of the soil from each geobotanical unit (Appendix F1) is displayed in Figure 57. Na is included as a beneficial element, since it is not generally considered essential to plant growth, but can be required by a few species for growth and development (Orcutt & Nilsen, 2000). Although the total macronutrient concentration (Appendix E2) is more related to the variance in geology, it is the plant-available portion of these elements that control the floristic variation of the vegetation. The availability of this portion was more affected by soil pH, CEC, and element mobility.

Results

Units A and C generally had lower available nutrients (Figure 57). This correlated with lower pH values associated with the quartzite country rock. The soil CEC was also accordingly lower and, thus, fewer nutrients are retained against leaching. Nonetheless, the available P content in both units was relatively high (~10 to 31 mg/kg) compared to the dioritic zone (Unit D2) (~4 to 25 mg/kg). The Mg availability in Unit A (~180 to 270 mg/kg) was also particularly high, given the dominant bedrock mineralogy (quartzite). Whereas on the quartzite ridge, Unit C, Mg availability was significantly lower (~54 to 62 mg/kg).

K availability for Units A and C had a similar trend to the extractable Mg content. A slightly higher range of extractable K (~108 to 132 mg/kg) was observed in Unit A compared to Unit C, and even most samples from Unit D2 (69 to 85 mg/kg) (with the exception of one outlier containing up to 186 mg/kg). The soils from Unit C contained the lowest amount of available K (~53 to 69 mg/kg).

The extractable Na content for Units A and C showed a smaller variation. Unit C contained the lowest concentration of available Na (1.5 to 2 mg/kg), whereas Unit A had a marginally higher range (1.5 to 3 mg/kg), which was comparable to the mafic zone (Unit D1).

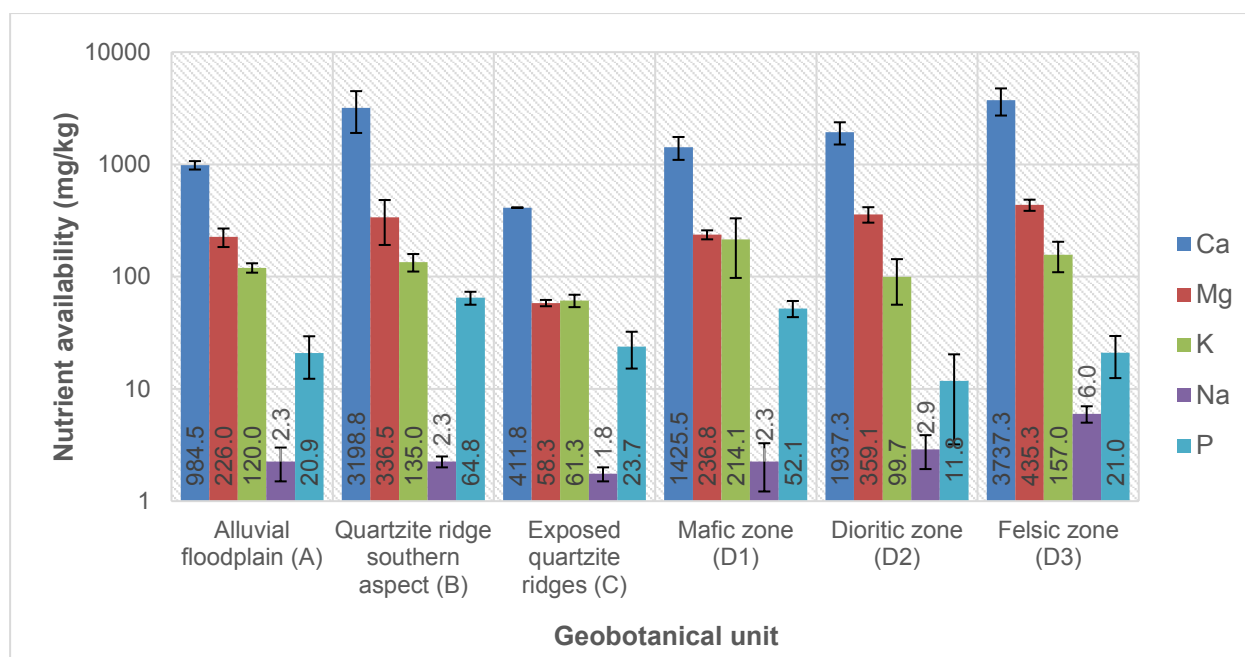


Figure 57: Mean macronutrient availability for the geobotanical units. Units A and C, situated on quartzite, have overall lower nutrient concentrations, corresponding to lower values of soil pH, CEC, and OC. Units D1, D2, and D3, containing ultramafic to felsic rock types, as well as Unit B, which is situated on the southern aspect of the quartzite ridge, have higher nutrient concentrations, with corresponding pH, CEC, and OC values. Units D1 and D2, containing ultramafic to intermediate rock types, have a lower availability of Ca and Na than do Units B and D3, containing felsic rock types, and correlating with higher abundances of pyroxene, and lower amounts of Ca–Na-feldspars. No substantial variation attributed to the geology is observed for the availability of Mg, K, and P.

Units B, D1, D2, and D3 had overall higher ranges of available nutrients than did Units A and C (Figure 57). This is associated with higher values of soil pH and CEC. Because of the difference in mafic and felsic mineral abundance and correlated geochemistry, these units differed primarily in available Ca and Na. A similar conclusion could not be drawn for the available Mg and K content, which did not correlate with the distribution of major mineral sources, - olivine, clinopyroxene, muscovite and microcline.

P availability did not have a strong relationship with the geochemistry of the bedrock materials either. However, some association could be seen with the occurrence of spessartite in Unit D1 (~69 to 31 mg/kg). Nonetheless, the highest concentration of extractable P was observed in Unit B (~70 to 59 mg/kg) containing high amounts of organic material but no major sources of apatite.

Units D1 and D2 had significantly lower Ca availabilities (~979 to 1826 mg/kg and ~1309 to 2653 mg/kg), compared to Unit B (~1900 to 4490 mg/kg) and Unit D3 (~2724 to 4750 mg/kg), which contained the highest amount of extractable Ca.

The available Na content for Units D1 and D2 had a broader but generally similar sample range (1.5 to 4 mg/kg), however, the mean value for Unit D1 was lower (2.3 mg/kg) and more comparable Unit B (2 to 2.5 mg/kg). Whereas the mean value for Unit D2 was slightly higher (2.9 mg/kg). The felsic zone (Unit D3) had the highest Na availability (5 to 7 mg/kg), which showed a significantly higher sample mean than any of the other units (6 mg/kg).

Discussion

It is reasoned that the overall availability of the aforementioned elements is controlled mainly by the interaction of four factors, specifically the pH, CEC, source material composition, and the relative mobility of specific ionic species. The degree to which each factor contributes towards nutrient availability, however, varies for each element. Some elements, therefore, could be more directly correlated with the geology than others.

Furthermore, when a large portion of the CEC is derived from SOM (humic acids), such as in this case, the number of exchange sites (carboxylic- and hydroxylic groups) is greatly dependent on the pH which governs protonation and deprotonation. Therefore, a decrease in pH is usually correlated with a decrease in CEC (Mengel & Kirkby, 2001). Subsequently, the pH plays a double role in element mobility. Not only does the prevailing pH determine the solubility of a particular ionic species, it also affects the number of adsorption sites in organic colloids, which in part controls the number of nutrients retained against leaching.

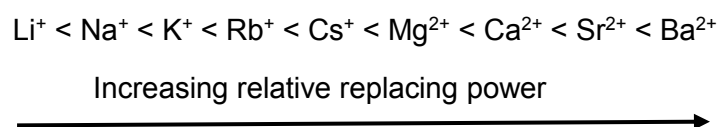
Mg and K generally have a higher mobility compared to Ca (Orcutt & Nilsen, 2000), therefore no significant variation directly attributable to the bedrock geology could be observed between their available concentrations in the topsoil (Figure 57). With respect to Mg, this equal distribution across the units is mainly achieved through leaching. In the case of K, which had less major variation partly due to leaching, the majority of this element is provided by feldspars, muscovite, and minor amounts of clay minerals within the abundant quartzite bedrock. Ultimately in terms of availability, these elements are largely controlled by the soil pH and CEC.

The relatively high Mg values observed in Units A and B are mainly attributed to Mg leaching from the adjacent Units D1 and D2, whereas the lower Mg availability in Unit C can be attributed to the absence of mafic minerals, lower pH ranges, as well as no leachate contamination due to it being located on top of the ridge.

The relatively high K availability in Units A and B can similarly be attributed to leaching from the spessartites (containing feldspars) in Unit D1. However, compared to Mg, a larger portion of the available K is probably derived *in situ* from the less mature quartzite bedrock. Typically lower values were observed for Unit C, which is contributable to lack of mineral sources. This is due to the general maturity of the quartzite, as well as the absence of igneous rocks specifically containing feldspars and muscovite.

The concentration of available P had a greater variation among units, but apart from Unit D1 did not correlate strongly with the variation in geology. This may be partly due to leaching because this element was concentrated mainly in lower-lying areas with higher CEC values. This observation relates to the highly mobile characteristic of the inorganic P anion, which has likely accumulated within these areas. On the other hand, a strong relationship could also be observed with the SOC in Units B and C, which would suggest that much of the available P is also derived from decomposing organic material. In mineral soils, organically derived P can comprise between 20-80% of the total soil P (Mengel & Kirkby, 2001). This would clarify why particularly high P values are observed in these units, despite having no identified igneous outcrops containing sources of apatite.

The concentrations of extractable Ca and Na had the greatest variation that could be correlated with the variance in geology. These elements (Ca in particular) seem to be held strongly in soil cation-exchange sites and are less affected by leaching compared to Mg, K and P. With respect to Ca this is due to higher binding strength and related replacing power as illustrated by Hofmeister's cation sequence (Mengel & Kirkby, 2001):



Ca can therefore easily displace other cations, and is largely retained near its mineral source. This is not typically the case for Na, which, as illustrated, is weakly adsorbed as a monovalent cation and easily replaced by K and Mg.

The fact then that Na seems to be more strongly adsorbed is rather related to the geology (source material) that affects its kinetic energy. The purely ultramafic outcrops of the KC are generally thin and sparsely distributed, while the majority of the mafic to felsic rocks (containing Na-rich plagioclase) are more common and occur as thicker outcrops. As a result, geobotanical units containing igneous rocks have a large source of Na during weathering. According to Mengel & Kirkby (2001) when a large amount of ions are present of a given species, a small proportion will occur as “high energy” cations following the Maxwell energy distribution curve. These “high

energy” ions allow weakly bound monovalent cations such as Na^+ to replace more strongly bound species such as Mg^{2+} or Ca^{2+} . Therefore units with rocks containing larger amounts feldspar will tend to have soils with greater amounts of more strongly bound Na, as well as Ca.

Lower values of Ca and Na observed in Unit D1 is attributed to the presence of ultramafic rocks containing olivine, serpentine, and pyroxene, while slightly lower values in Unit D2 is due to the epidiorite that contains lower concentrations of plagioclase feldspar.

Units B and D3 contained higher amounts of available Ca and Na, because of higher amounts of plagioclase and Ca–Na-rich amphiboles. In Unit B, most of the Ca originates from hornfels, as well as leaching from epidiorite in D2, whereas quartz alkali syenite, alkali granite and small sill of hornblendite are the primary sources of plagioclase and amphibole in Unit D3. The feldspars of the quartz alkali syenite and alkali granite are also more Na-rich. Therefore, Unit D3 also had a remarkably higher Na availability.

Ca and Na availability of the soil is, thus, significantly affected by the variation in dominant rock types, and these are the macronutrients with the greatest relevance, which can control the floristic composition on the basis of differences in geology.

5.4.2 Exchangeable Ca:Mg

Alternate trends in Ca and Mg availability could be established from their soil extracted concentrations (Figure 57). The effect on one another in terms of relative availability, is better illustrated by their exchangeable ratios (Figure 58). While the extractable Mg content was surprisingly less affected by the geology, its relative availability was indirectly affected by the amount of Ca supplied by the source material. This is because as was previously discussed, the concentration and related activity of either of these elements can also determine the ease with which they are exchanged between soil colloids and the soil solution (Mengel & Kirkby, 2001). The Ca:Mg ratio and thus relative Mg exchangeability can, therefore, be largely correlated with the bedrock geology.

Results

The lowest Ca:Mg ratios (2.41 to 2.97) were observed in the alluvial floodplain (Unit A) which had an indicated sample mean of 2.69 (Figure 58). The mafic zone (Unit D1) and dioritic zone (Unit D2) had similar, marginally higher ratios ranging between 2.6 and 4.7, of which Unit D1 had a slightly higher sample average (3.67) compared to that of Unit D2 (3.25).

Despite having the lowest available Ca content (Figure 57), the exposed quartzite ridges (Unit C) had higher ratios (4.01 to 4.6) than the former Units (A, D1, D2), averaging 4.3.

The southern aspect of the quartzite ridge (Unit B) and the felsic zone (Unit D3) had the highest ratios ranging from 4.28 to 6.03, with respective sample means of 5.84 and 5.11.

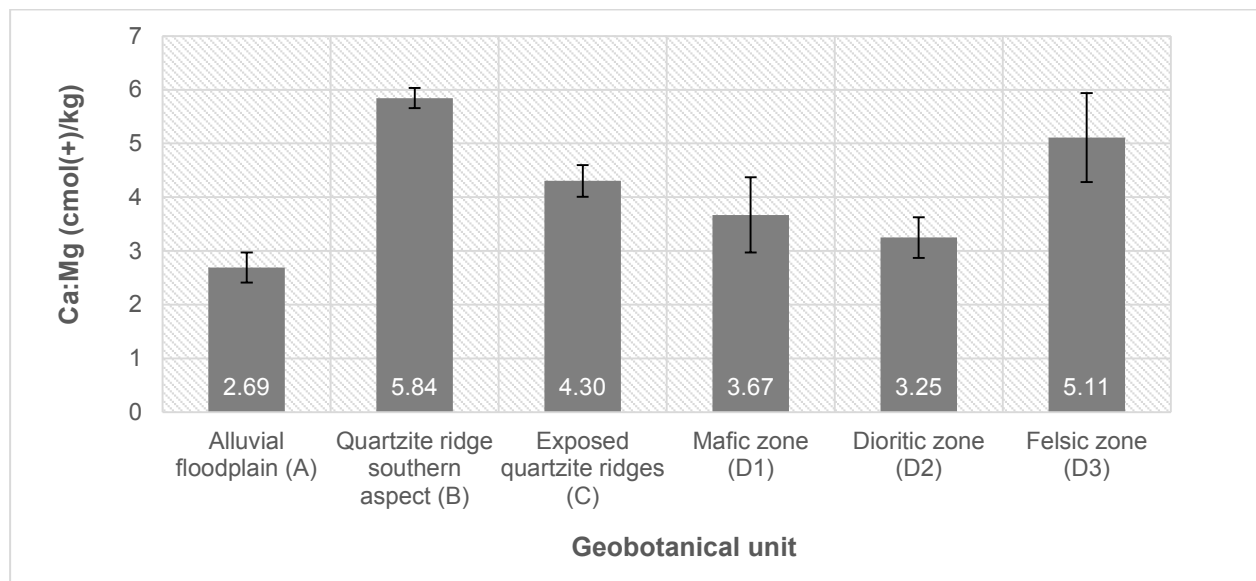


Figure 58: Mean exchangeable soil calcium:magnesium ratio (Ca:Mg) of each geobotanical unit. Ratios of <4 are observed for Units A, D1, and D2, which indicates decreased Ca availability. Low ratios within Unit A result from contamination by Mg leaching, whereas for Units D1 and D2, this can be directly ascribed to lack of Ca-rich feldspar sources within the ultramafic and mafic rock types, as well as less retention of Ca ions in the soil profile. As a result, their supported vegetation is exposed to lower Ca, and higher Mg availability, than for Units B, C, and D3, which are not affected by Mg leaching (Unit C), and contain sufficient sources of Ca (Units B and D3).

Discussion

The exchangeable Ca:Mg ratio of the soil from each geobotanical unit (Figure 58) was consistent with observations from the soil nutrient status (Figure 57). With the exception of the alluvial floodplain (Unit A), owing to higher Mg availability, units containing rock types with no mafic minerals (Unit C) or higher amounts of plagioclase and amphiboles (Units B and D3) illustrated higher Ca:Mg ratios (>4). Whereas units containing rock types with more olivine and pyroxene (Units D1 and D2) had soil Ca:Mg ratios of <4, at which the exchangeable Ca concentration on soil colloids was reduced, and the Mg content was relatively more available compared to Ca availability (Hazelton & Murphy, 2007).

Soil from Unit D1 had a higher Ca:Mg ratio than typically expected from ultramafic soils (<1, Mengel and Kirkby, 2001). This is primarily due to the high Ca content of augite, and the presence of diorite and spessartite, which contributes more Ca than Mg. These factors result in a slightly higher Ca:Mg ratio, which is similar to that in the dioritic zone (Unit D2). Nonetheless, in both of these units, the supported vegetation will be affected by lower Ca availability.

The lower Ca:Mg ratio observed in Unit A is attributed to Mg accumulation and absence of plagioclase minerals providing Ca. Although this unit does not contain ultramafic to intermediate rock types, the associated vegetation will suffer similarly from lower Ca availability and higher concentrations of available Mg due to leaching.

5.4.3 Nitrogen (N)

The most important plant available forms of N in soil are nitrate (NO_3^-) and ammonium (NH_4^+), of which the majority is produced during the mineralization of organic N compounds, and the related ammonification and nitrification process, that is catalyzed by variety of heterotrophic and autotrophic microorganisms (Mengel & Kirkby, 2001; Hazelton & Murphy, 2007). Ammonium is generally very low in aerobic soils since it is rapidly nitrified, and thus, plays no major role with respect to the available pool of N (Mengel & Kirkby, 2001). This was particularly observed for the topsoils of the KC, which is not prone to waterlogging due to a semi-arid climate, the inclined nature of the topography, and overall low clay content. As a result, samples generally illustrated a very low ammonium concentration (0.01 to 0.14 mmol/L; Appendix F1) and were subsequently omitted as a possible cause for major floristic variation, since no significant differences could be observed between geobotanical units. The nitrate concentration was far larger and showed a meaningful variation (Figure 59). The difference between units corresponded primarily to the amount of organic material present in the soil (as indicated by the concentration of SOC; section 5.3), the pH, and to some extent the concentration of Ni provided by the source material.

Results (NO_3^-)

The lowest concentration of NO_3^- was observed in the alluvial floodplain (Unit A) which had a range of 0.17 to 0.21 mmol/L, and mean value of 0.19 mmol/L (Figure 59). The exposed quartzite ridges (Unit C) had a slightly higher range of 0.52 to 0.75 mmol/L which amounted to mean value 0.64 mmol/L.

The mafic zone (Units D1) and dioritic zone (D2) illustrated higher NO_3^- values compared to Units A and C. These units were more analogous, with samples generally ranging between 0.46 and 1.18 mmol/L, however, both units also contained particularly high outlier values of 1.57 mmol/L and 2.25 mmol/L respectively. As a result, the mean value for Unit D1 (0.95 mmol/L) was slightly lower than for Unit D2 (1.14 mmol/L).

Units B and D3 on average had the highest NO_3^- content that ranged between 1.4 and 1.96 mmol/L. Unit B had a lower calculated mean (1.47 mmol/L) compared to unit D3 (1.85 mmol/L).

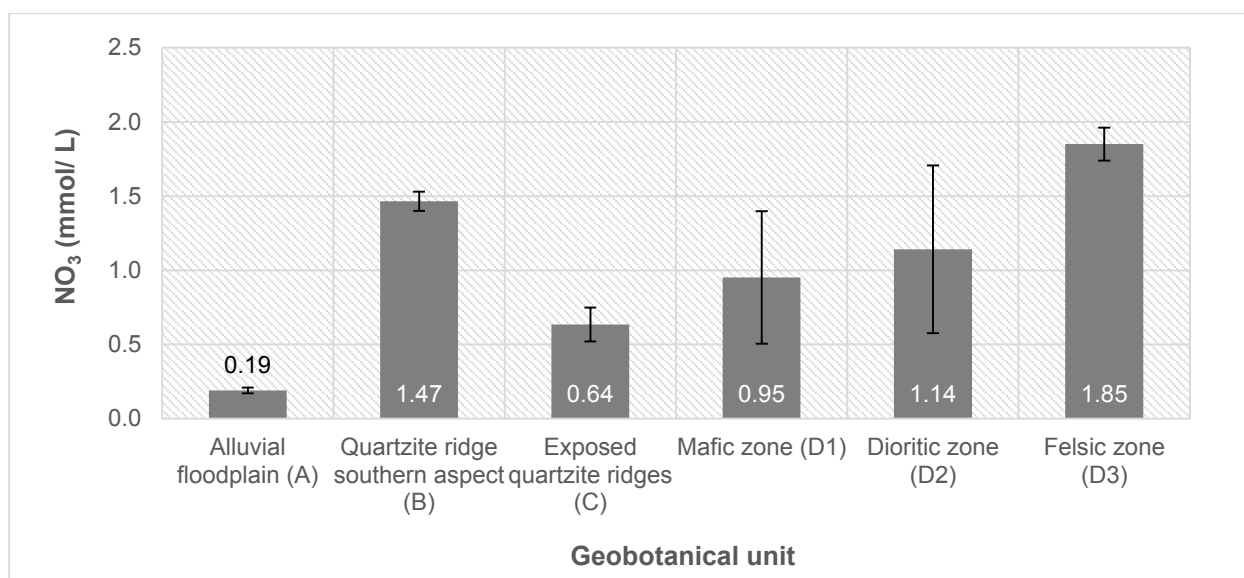


Figure 59: Mean soil nitrate (NO₃⁻) content for the geobotanical units. Lowest values observed in Units A and C are separately attributed to low amounts of SOM (Unit A) and strongly acidic pH (Unit C). Moderately higher values observed for Units D1 and D2 can be correlated with increased pH values and SOM. The lower mean value for Unit D1 compared to D2 is mainly due to a slightly lower SOM content but was probably also affected by higher concentrations of Ni. The highest values are observed in Units B and D3 and are attributable to a very high SOM content and a neutral to slightly acidic pH, which was the most optimal for N fixation and nitrification.

Discussion

The majority soil N in the form of NO₃⁻ is produced during nitrification that involves the biological oxidation of ammonia (NH₃). This is a multi-stage process which is facilitated by two specialized groups of autotrophic bacteria that oxidize NH₄⁺ to nitrite (NO₂⁻) and NO₂⁻ to NO₃⁻ (Mengel & Kirkby, 2001). Subsequently, the amount of NO₃⁻ produced then is a function of the rate of nitrification and availability of NH₃, which in turn is derived from heteromorphous bacteria which assimilate atmospheric N₂ (Nitrogen fixation) or catalyze the hydrolysis of organic N compounds (Nitrogen mineralization). The major source of NO₃⁻ therefore, is mainly biologically controlled rather than geochemically but can be influenced by the geology to some extent.

The concentration of soil NO₃⁻ was observed to be strongly related to the presence of organic material, although a statistically significant correlation could not be directly drawn between the SOC and NO₃⁻ content. Firstly, this is because the percentage of organic carbon derived from microbial mass and organic plant litter cannot be differentiated in the SOC value. Therefore the ratio of organic nitrogen compounds and bacteria able to mineralize and convert these compounds are not taken into consideration. This will drastically affect the amount of NH₃ provided for nitrification, as well as the speed at which nitrification occurs. Nonetheless, if this

ratio is assumed to be constant the SOC value can be used as a broad indication of organic material, and the related NO_3^- content as an overall indication of microbial activity.

Secondly, deviations were seen for Unit C and some samples from Units D1 and D2. This is attributed indirectly to the geology which affected the soil pH (section 5.2). N fixation and nitrification are greatly affected by low soil pH, as the bacteria mitigating these processes are pH sensitive and prefer neutral to slightly acidic soil conditions. Thus, in acid soils, despite a high SOM content, these processes are substantially depressed or even inhibited (Mengel & Kirkby, 2001). This was particularly seen for samples that had a pH ranging from 4 to 5 and could be verified by their slightly increased NH_4^+ content, indicating depression in nitrification. These findings correspond with Barber (1984) and Harris *et al.* (1996) stating that at low pH values the available pool of N remains in the NH_4^+ form and is not microbially converted to more plant-available NO_3^- as nitrification is inhibited.

From the results, it is difficult to determine the extent which the ultramafic and mafic intrusions specifically affected soil N availability. Generally, in terms of soil pH, weathering of these rock types will tend to promote nitrification. However, in terms of nutrient and trace element content, vegetation and microbial activity will be negatively affected, decreasing the amount and decomposition rate of organic N compounds. Overall it seems that the SOM content and pH was the strongest influencing factors compared to the Ni content, which rather than directly affecting nitrification, affected the amount of organic plant material accumulated in the soil. This observation explains why a higher NO_3^- content was observed in the dioritic zone (Unit D2) compared to Mafic Zone Unit D1 despite having a lower average pH.

5.5 MICRONUTRIENTS (Fe, Mn, Ni, Cr, Co)

5.5.1 Iron (Fe)

Due to the large mafic mineral content of the ultramafic and mafic intrusions, as well as the presence of magnetite (section 4.4 & 4.5), a large portion of Fe is transferred to the soil via weathering. Fe however, is not considered a limiting factor in ultramafic soils, since iron toxicity generally occurs at very low pH ranges (<4) when higher amounts of Fe become soluble in the form of Fe^{2+} (Brady and Weil 2008). Its effect on the associated vegetation is, thus, generally small and highly dependent on the soil pH and oxidation potential (Eh). Nevertheless, in cases such as the KC where parent materials differ considerably in iron content (Figure 49) and related soil pH (Figure 53), changes in geological substrates also become an important contributing factor. This observation is illustrated in Figure 60, displaying the total and soluble iron content of the soil in each geobotanical unit.

Results

Soil from the exposed quartzite ridges (Unit C) had overall the lowest Fe content, with an average total concentration of ~22000 mg/kg and soluble concentration of ~12 $\mu\text{mol/L}$ (Figure 60). The alluvial floodplain (Unit A), dioritic zone (Unit D2) and felsic zone (Unit D3) had higher mean values of total Fe which ranged from ~49000 to 59000 mg/kg. While these units were comparable in terms of the total Fe, the soluble Fe concentration for Units A and D2 was higher (~20 to 21 $\mu\text{mol/L}$) compared to Unit D3 (~15 $\mu\text{mol/L}$).

The southern slope of the quartzite ridge (Unit B) and mafic zone (Unit D1) contained the highest amount of total iron. The sample mean for each unit was ~81000 mg/kg and 70000 mg/kg⁻¹ respectively. Although the total Fe was higher for Unit B, it had a remarkably lower Fe availability (~17 $\mu\text{mol/L}$) compared to Unit D1, which had the highest soluble iron content (~27 $\mu\text{mol/L}$).

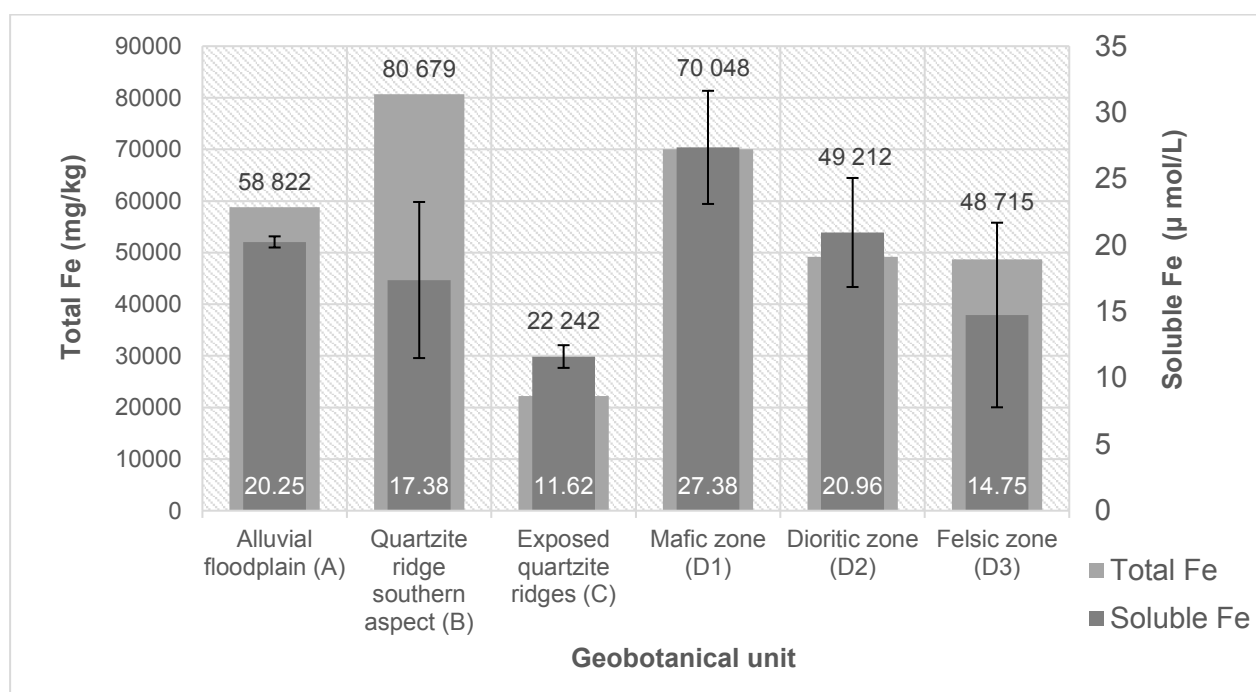


Figure 60: Mean concentration of total and soluble soil iron (Fe) in each geobotanical unit. Units B, C, and D3 contain lower concentrations of soluble Fe than do Units A, D1 and D2. For Unit C, this corresponds to the absence of mafic silicate minerals, whereas for Units B and D3, this is due to higher values of soil pH, which are strongly related to the base saturation. The higher concentrations of soluble Fe observed in Units A, D1 and D2 can be simultaneously correlated with source material that contains a high amount of magnetite or Fe-rich mafic minerals (wehrlite, olivine pyroxenite, diorite, and epidiorite), and moderately low pH values. The high concentration of Fe-rich source material in Unit A is however not residual and was probably derived from Units D1 and D2.

Discussion

Despite having the lowest pH values (~4), Unit C contained the lowest concentration of soluble Fe, because the dominant rock type (quartzite) contains no mafic minerals. Hence, a low total Fe content was observed, which indicated that the geology is also a major controlling factor. For this reason Units B and D3, which contain rock types with iron-rich amphiboles, and therefore a higher total Fe content would be expected to have higher amounts of soluble Fe. Nonetheless, owing to higher soil basicity, similar concentrations to Unit C was observed, confirming that the soil pH also affects Fe solubility to some extent.

Units A, D1, and D2 contained the highest concentrations of soluble Fe regardless of higher pH values. This observation is again related to the geology. Units D1 and D2 contain rock types with the highest Fe concentration (mela-dolerite, wehrlite, gabbro and epidiorite; Appendix E1) which directly contributed to higher soil Fe values. In Unit A, the relatively large amounts of Fe could not have solely originated from the quartzite bedrock and were most likely partly derived from weathered ultramafic and mafic materials, which were transported to the alluvial floodplain. This also explains why particularly high Mg values were observed in Unit A.

Ultimately, because the soil soluble Fe content is also controlled by changes in parent material, it can, therefore, be regarded as an important micronutrient related to the geology, which can affect species distribution. In addition to the direct effects on vegetation, Fe solubility will also affect the availability of trace elements such as Ni and Co. As discussed in chapter 2, this is ascribed to the specific tendency of hydrated Fe- and Mn-oxides to adsorb and co-precipitate with these trace elements, and if related changes in pH or redox potential cause dissolution of these oxides, the co-precipitated metals will be released accordingly (Alloway, 1990). For this reason, the Ni and Co concentration is expected to show a strong association with the Fe and Mn content, however, this was not particularly observed for the KC (Section 5.5.3).

5.5.2 Manganese (Mn)

Mn is able to isomorphically substitute Fe and Mg in mafic silicates and oxides, as a result, high concentrations of Mn are also common on ultramafic soils. Although on average the total Mn content of soil is generally lower than Fe (Hazelton and Murphy, 2007, Kabata-Pendias, 2011), Mn oxides are more easily dissolved at lower pH ranges, and harder to precipitate in more alkaline conditions (Alloway, 1990), hence, Mn toxicity will tend to be more common than for Fe. The total soil Mn content for the different geobotanical units is illustrated in Figure 61.

Results

The lowest concentration of Mn was observed in the topsoil of the exposed quartzite ridges (Unit C), which ranged between ~340 and 430 mg/kg, and averaged ~390 mg/kg (Figure 61).

Higher, more comparable ranges of ~770 to 1080 mg/kg were observed for the alluvial floodplain (Unit A) and the dioritic zone (Unit D2). Each unit had a respective average of ~810 mg/kg and ~960 mg/kg.

The southern aspect of the quartzite ridge (Unit B), the mafic zone (Unit D1) and the felsic zone (Unit D3), had the highest concentrations of Mn, which ranged between ~910 mg/kg and 1490 mg/kg.

Unit B had a particularly high average of ~1310 mg/kg, while Units D1 and D3 had slightly lower corresponding averages of ~1150 mg/kg and ~1180 mg/kg.

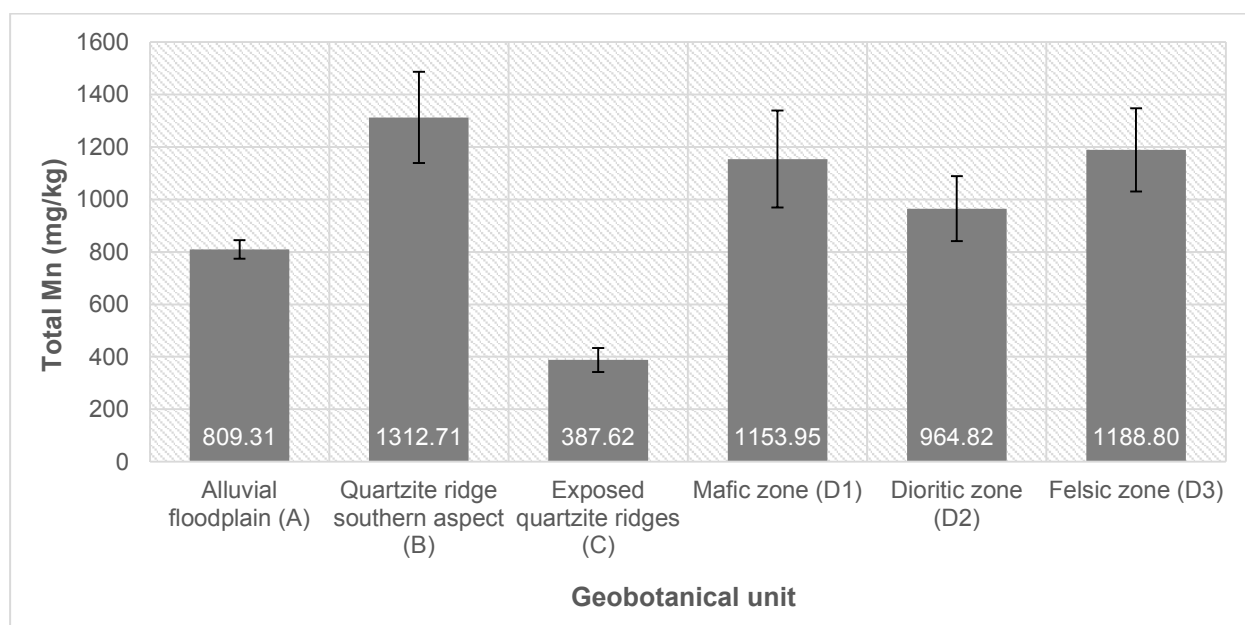


Figure 61 Mean concentration of total soil manganese (Mn) in each geobotanical unit. Unit C illustrates the lowest mean value for Mn, which is ascribed to lack of mafic minerals and element leaching. Moderately higher values are observed for unit A and D2. For Unit A, this is due to leachate and mafic mineral contamination from adjacent units, while for Unit D2, higher values are correlated to the presence of epidiorite containing amphibole and pyroxene. Units B, D1, and D3 have the highest mean values for Mn. The large amount of Mn in unit D1 is directly derived from ultramafic and mafic parent material, while in units B and D3, the remarkably high values are due to a combination of amphibole sources (hornfels, hornblendite, and alkali syenite), and high soil pH and CEC values.

Discussion

The soil Mn content could to some measure be correlated with the geology, however, the association with the bedrock was generally weaker than for Fe. This is attributed to the greater mobility of Mn^{2+} which is more prone to leaching in acid soils (Mengel & Kirkby, 2001), therefore, less variation in total Mn was observed across the geobotanical units. Apart from the source material, the Mn content seemed to be strongly affected by both the soil pH and CEC. Larger concentrations were observed to be retained in units that had a moderately low to neutral pH and higher CEC values, irrespective of major differences in geology.

The low soil Mn values observed on the exposed quartzite ridges (Unit C), is partly related to the lack of mafic minerals and Mn/Fe oxides, but also due to the acid pH and low soil CEC, which resulted in less Mn retained against leaching.

The alluvial floodplain (unit A) and dioritic zone (Unit D2) illustrated comparably increased concentrations of Mn, however, for Unit A this is attributed to leaching and accumulation of physically transported source material rather than being directly related to the bedrock. Alternatively, in Unit D2, the Mn is primarily derived from weathered epidiorite, containing amphibole and pyroxene.

The largest concentrations of Mn were observed in the lower shaded aspect of the quartzite ridge (Unit B), the mafic zone (Unit D1) and the felsic zone (Unit D3). In Unit D1 the high concentration is directly related to the ultramafic and mafic parent materials, which supply large amounts of Mn despite to the soil's leaching characteristics. Whereas in units B and D3, comparable to even larger values were observed regardless of fewer sources of amphibole. This is attributed to the relatively high pH and CEC. At greater pH ranges larger amounts of Mn will remain in an insoluble form, and are less likely to be leached, while the higher soil CEC will retain more of the mobile fraction.

Unit B specifically receives high amounts leached Mn from adjacent units, which becomes concentrated in insoluble forms due to more alkaline conditions. This explicates why this unit has particular high mean concentration compared to Units D1 and D3.

All the geobotanical illustrated total Mn concentrations below the general phytotoxic value of 1500-3000 mg/kg, as reported by Kabata-Pendias (2011). However, based on the toxicity values of 80 to 5000 mg/kg, published earlier by Alloway (1990), It can be suggested that species more sensitive to Mn will be the most likely affected in Units B, D1, and D3.

5.5.3 Trace elements (Ni, Cr, Co)

The total Ni, Cr and Co content of the soil from each geobotanical unit, as determined by XRF analysis, is displayed in Figure 62. Concentrations of Ni and Cr were strongly associated with the presence of wehrlite, olivine $\pm$ clinopyroxenite and orthopyroxene hornblende gabbro, and confirmed the observation made from the whole rock geochemistry (Section 4.5.2). However, the concentration of Co did not show a similar association. Apart from direct associations with the geology, all three elements did not show any significant relationship with the total soil Fe or Mn concentration (Figure 63 & Figure 64).

Results

The soil Ni content illustrated the lowest variation amongst geobotanical units, with a general sample range of 0 to 204 mg/kg. The highest reported sample concentration was observed in the mafic zone (Unit D1), which had sample average of ~128 mg/kg (Figure 62).

The shaded quartzite ridge (Unit B) and felsic zone (Unit D3) illustrated almost similar sample averages of ~121 and ~126 mg/kg. However, unlike unit D1 the highest reported Ni content for these units were 165 and 133 mg/kg respectively. The lowest concentration of Ni was observed in the alluvial floodplain (Unit A) and the exposed quartzite ridges (Unit C), which had a corresponding sample mean of ~81 and ~28 mg/kg.

The soil Cr content demonstrated a similar trend to Ni, although generally higher sample values of 152 to 567 mg/kg were observed across the geobotanical units. The largest concentration of Cr was reported in unit D1, which had a sample average of ~430 mg/kg. The latter units (A, B, C, D2, and D3) all illustrated considerably lower averages ranging from ~180 to ~235 mg/kg.

Compared to Ni and Cr, the soil Co content followed a remarkably different trend with respect to its distribution. The highest sample values were obtained from Units A, B and C. Each unit had a respective sample mean of ~200, ~237 and ~163 mg/kg. Units D1, D2, and D3, on the contrary, had considerably lower mean sample values of ~48, ~63, ~58 mg/kg.

Statistical analysis of the distribution of Ni, Cr or Co in relation Fe and Mn, did not reveal significant correlations (Figure 63 & Figure 64). Both the Fe and Mn content had the strongest association with Ni, followed by Cr and then Co.

Discussion

Elevated levels of Ni, Cr, and Co in the topsoil definitely expresses the presence of ultramafic and mafic intrusions. Nonetheless, compared to typical soil values from larger ultramafic bodies

(Mengel & Kirkby 2001), high concentrations were less prominent and sporadically distributed, resulting in lower sample averages. This is an aftereffect of the thin structural occurrence of the intrusions, which over long periods of erosion became patchily distributed and mixed with a large amount of weathered quartzite material. Not only does the higher amounts of quartz result in lower pH values and increased element mobility, the soil trace element concentration is also effectively diluted per unit mass. Therefore, a greater difference between the rock- and soil concentration was observed. Despite lower sample averages, all three elements were reported at concentrations above the indicated phytotoxic level for surface soils (100 mg/kg for Ni, 75-100 mg/kg for Cr and 25-50 mg/kg for Co (Kabata-Pendias & Pendias, 1992; Orcutt & Nilsen, 2000; Zayed & Terry, 2003).

Soil samples of the mafic zone (Unit D1) containing ultramafic and mafic intrusions, had the highest concentrations of Ni and Cr. These concentrations will typically put more stress on the vegetation than in the other geobotanical units (A, B, C, D2 and D3) that contained significantly lower concentrations. Both the Cr and Ni was strongly associated with the geology, although Ni was more mobile than Cr, and seemed to be also retained in areas with higher CEC values (Units B and D3).

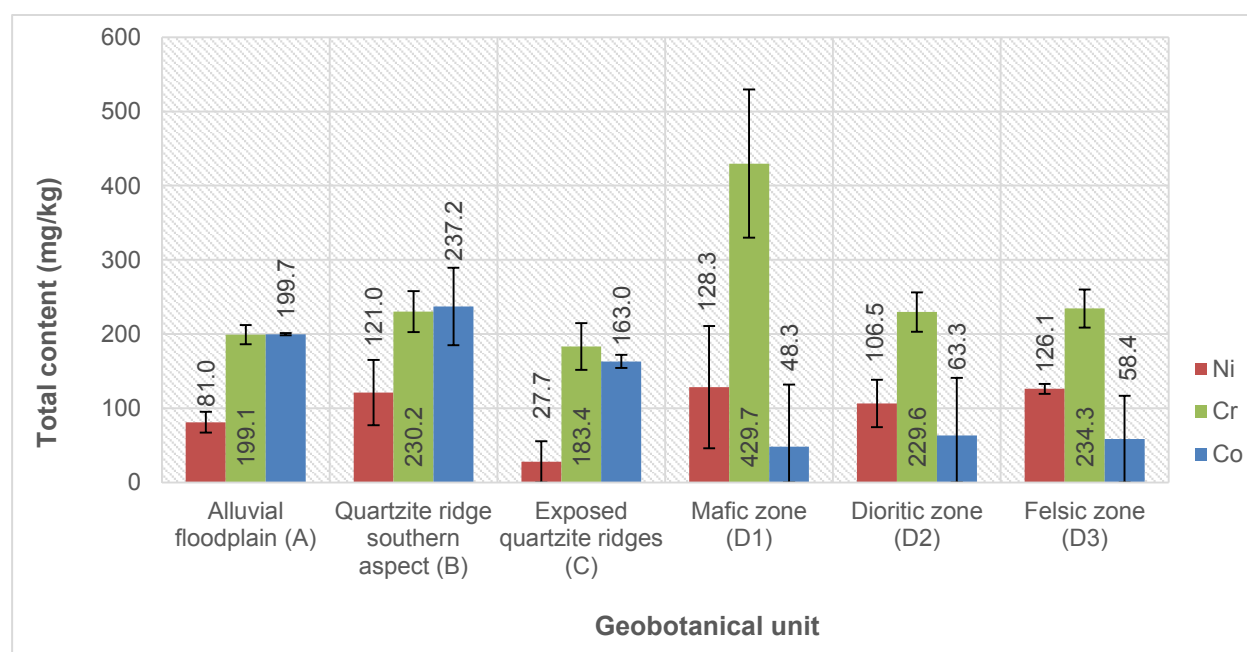


Figure 62: Mean concentration of soil nickel (Ni), chromium (Cr) and cobalt (Co) in each geobotanical unit. Cr and Ni are associated with olivine and pyroxene and are concentrated within the soil in Unit D1, which contains large amounts of ultramafic wehrlite, pyroxenite, and mafic diorite. The other units contain no ultramafic or mafic rocks and have lower and more analogous concentrations of Cr and Ni, the lowest values of which occur within Unit C. In contrast to Ni and Cr, Co is concentrated in Units A, B, and C, and, accordingly, does not correspond with the mineralogy of the rock substrate.

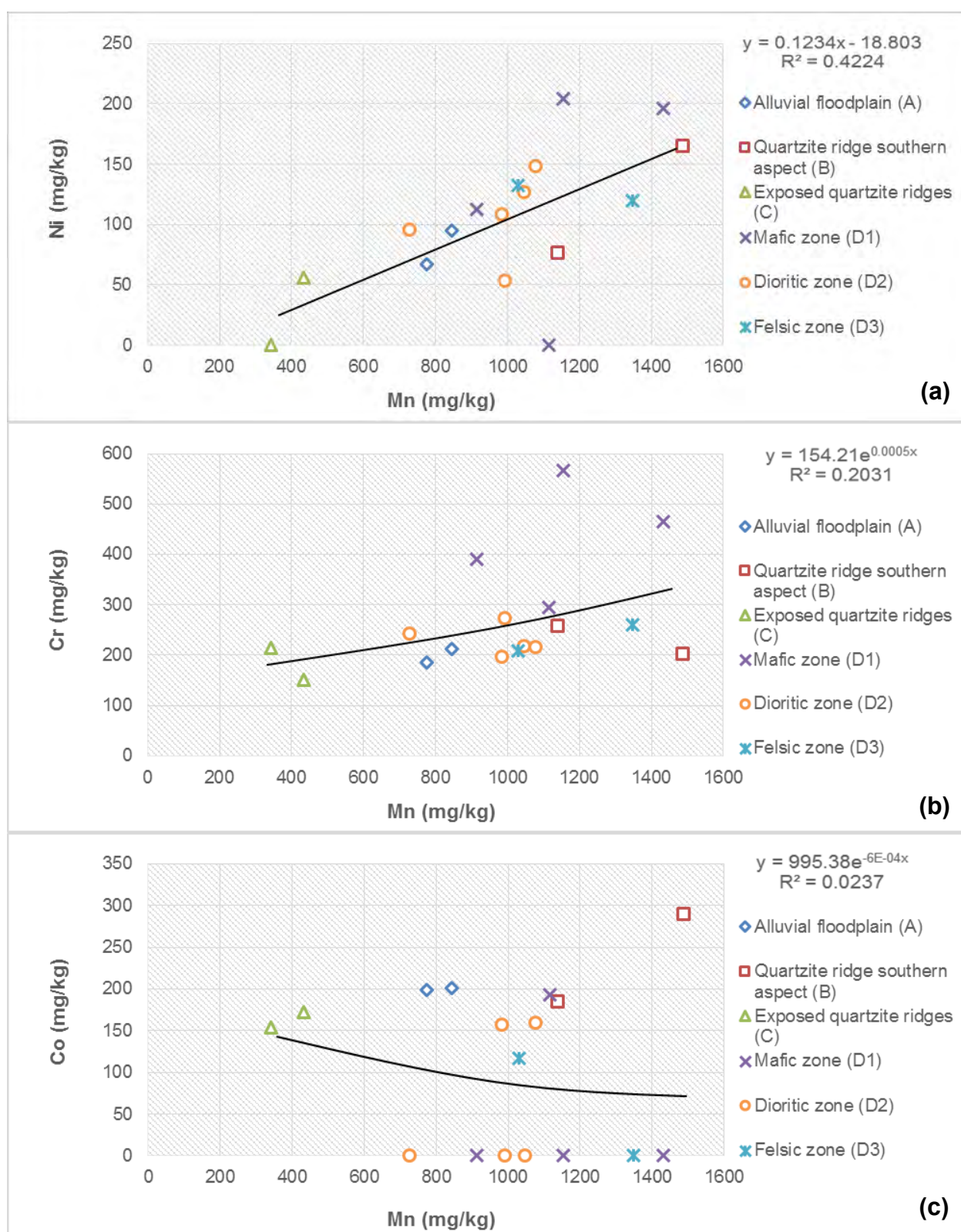


Figure 63: Sample scatter Plots for Mn vs (a) Ni, (b) Cr and (c) Co. A positive trend can be observed for all three elements in relation to Mn, however, due to the number of sample outliers, specifically from the Alluvial floodplain (Unit A), the base of the shaded quartzite ridge (Unit B) and mafic zone (Unit D1), no statistically significant association can be observed. Ni has the strongest linear trend in relation to Mn, while the Cr and Co content illustrates weaker exponential trends.

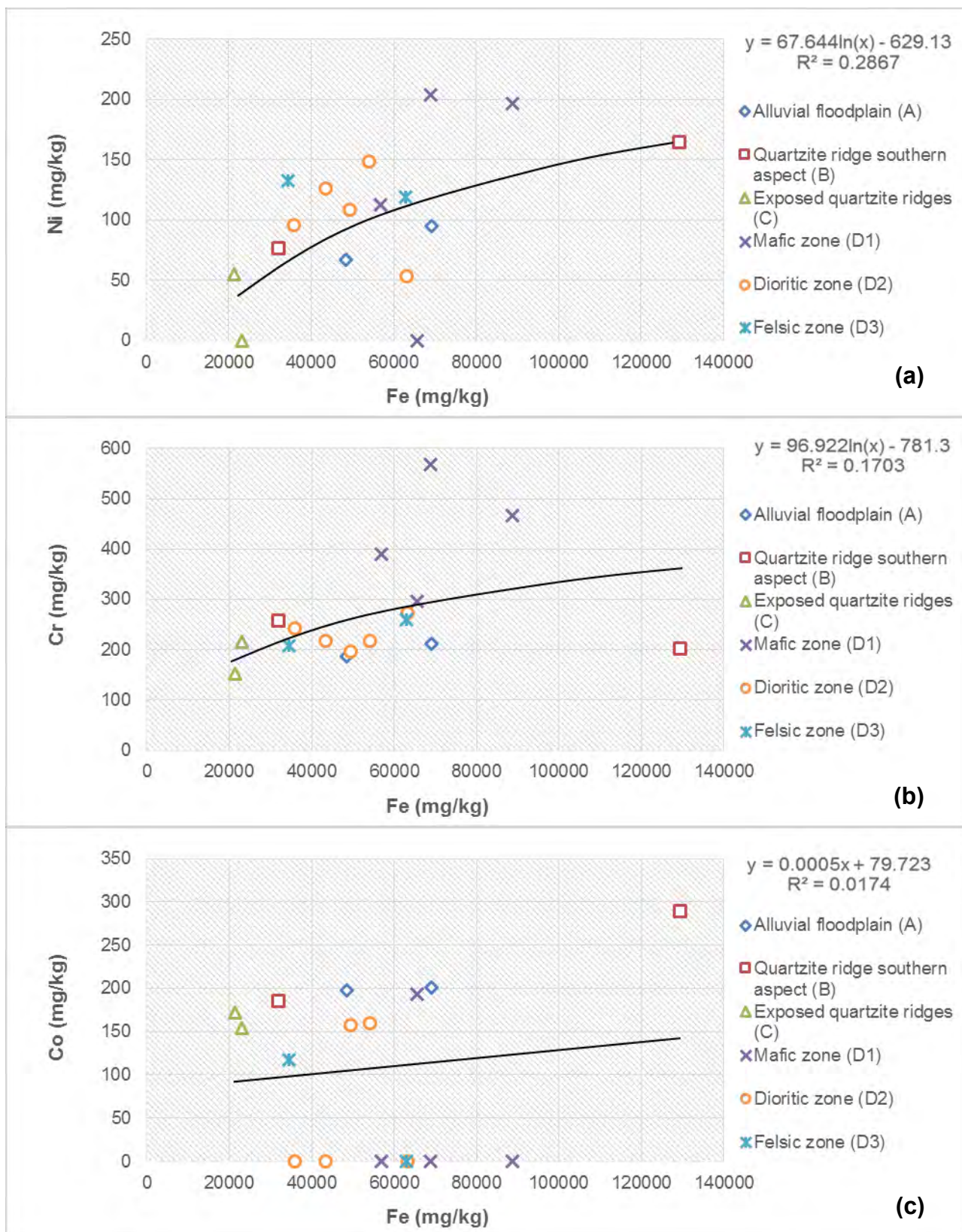


Figure 64: Sample scatter Plots for Fe vs (a) Ni, (b) Cr and (c) Co. All three elements illustrate a weaker association with Fe compared to Mn. Despite no significant correlation, a stronger logarithmic trend can be observed for Ni, as well as Cr in relation to Fe, while the Co content illustrates a weaker linear trend.

The Co concentration did not correlate with the distribution of wehrlite, pyroxenite, and diorite, and was rather high within soils containing higher amounts of quartzite and hornfels (Units A, B, and C). Although similarly high values for Co were observed in units D1 and D2, these concentrations were less pronounced. The concentration of Co, in this case, is most probably affected by small shale lenses (hornfels) that occur between the layers of quartzite. Sedimentary rocks such as shale are also known to contain elevated concentrations of Co (McGrath & Smith, 1990; Mengel & Kirkby, 2001), which may explain the higher Co concentrations observed in Units A, B, and C.

The soil Fe and Mn concentration did not have a meaningful association with the general distribution of Ni, Cr, and Co, although, stronger positive trends were observed for Ni (Figure 63 & Figure 64). This indicates that the aforementioned trace elements are not intensively adsorbed nor absorbed by Fe and Mn oxides, which is possibly attributed to differences in mobility based on pH variations, and adsorption strength on cation exchange complexes.

Ultimately, Ni and Cr indicated the greatest variance related to the geology, and are regarded as the phytotoxic trace elements of the highest importance that can affect floristic variation on the basis of changes in dominant rock types.

5.6 CONCLUSION

Based on the presented results, it can be concluded that the presence of igneous and related metamorphic rocks of the KC does have a measurable influence on the topsoil chemistry. Although the intrusions are relatively small, and their remnants sporadically distributed to some degree, relatable variances in chemical parameters could be identified in the soil profile.

Chemical characteristics including the pH, available -Ca, -Na and -Fe, and total Cr and Ni content, were more directly relatable to the mineralogy and geochemistry of the bedrock. While the CEC, available Mg, N, K, and P and total Mn and Co, were also greatly influenced by the topography, element mobility, SOC, and associated vegetation and soil microbial productivity. Thus, the overall variation in soil chemistry is contributed to a combination of both abiotic and biotic factors.

The greatest differences in soil chemistry were observed between sampling units situated primarily on quartzite bedrock, (Units A, B & C) and units containing altered- and unaltered igneous rock types (Units D1, D2 & D3). The quartzite dominated plots, in a general sense, illustrated low to very low- pH ranges, CEC values and macronutrient availabilities, lower concentrations of total Fe, Mn, Cr, and Ni, and high amounts of Co. Whereas the units containing igneous rock types overall had a moderate to neutral soil pH, higher CEC values, and available macronutrients, and large amounts of total Fe, Mn, Ni, and Cr, but less prominent Co values.

Some exceptions, however, were observed for samples from the base of the shaded quartzite ridge (Unit B), which, due to the topography and presence of hornfels on the upper slope, received large quantities of leached elements specifically Ca, Mg, P, Fe, Mn, Ni and Co. This resulted in an increased soil pH and higher nutrient availability that was comparable to Unit D3.

Specific differences could also be established across the individual units containing igneous rocks of varying geochemical affinities. Soil from Unit D1, situated ultramafic-dioritic rocks, had the highest Fe availability and total Ni and Cr content, and lower available Ca, Na and N. Unit D2, containing only dioritic rocks, hosted soils quite similar to D1 apart from slightly lower Fe, Mn, Ni and Cr concentrations, and higher Ca, Na and N availability. Unit D3, containing felsic rocks, and some small uraltite outcrops, had the highest Ca, Na and N availability and lower concentrations of Fe, Cr, and Ni.

To conclude with regards to the “serpentine effect”, observations made from the soil chemical data indicate the presence of ultramafic and mafic intrusions does produce ultramafic soil characteristics. This includes lower Ca and N availability, higher extractable concentrations of Mg and Fe, and toxic amounts of total Cr and Ni. However, due to the close structural occurrence of dioritic and felsic rocks, as well as the abundance of quartzite, much of this effect is greatly reduced through mineral dilution, and the addition of more nutritive mineral sources of Ca, K, Na, and P.

CHAPTER 6

WOODY VEGETATION OF THE KOEDOESFONTEIN COMPLEX

6.1 INTRODUCTION

Woody vegetation patterns are known to be associated with specific felsic peralkaline outcrops of the SC in the Vredefort dome (Bezuidenhout & Bredenkamp, 1988). This study investigated whether the same holds for the unique geology and associated soils identified for the KC. The geological and pedochemical data (Chapter 4 & 5) revealed that abrupt environmental gradients do occur across the identified geobotanical units of the KC. The aim of this chapter is to demonstrate the degree of floristic-, physionomical- and biochemical association of woody plant species in relation to these gradients, and to identify and discuss the significance of species occurrences, particularly that of *Vachellia* and *Senegalia* species, with respect to their prominence and performance on the ultramafic to felsic intrusions.

Specific objectives for the botanical facet of this study were to determine whether (i) woody species assemblages (floristics) are correlated with different rock substrates (section 6.2), (ii) individual species are associated with specific soil chemical variables related to the geology (section 6.3), (iii) species on ultramafic rock perform better than others according to performance index values (Appendix H2) calculated from abundances and stem circumferences (Appendix C2) (section 6.4) and, (iv) geological occurrence and soil chemical association are related to differences in tolerance of specific major and trace elements (section 6.5).

Firstly, based on differences in species response and resistance to changes in soil pH (Hazelton & Murphy, 2007), nutrient availability (Mengel & Kirkby, 2001; O'Dell *et al.*, 2006), and trace element toxicity (Kabata-Pendias & Pendias, 1992; Kabata-Pendias, 2011; McGrath & Smith, 1990; Orcutt & Nilsen 2000), it is hypothesised that variation in floristic patterns of the associated woody vegetation will correspond with changes in geology, and therefore, should relate to the identified geobotanical units. Secondly, it is hypothesised that amongst the woody flora, *V. karroo* will be more prevalent on the ultramafic lithologies due to its general robustness and adaptability (Hayward, 2004; Nair, 1993; Nyakundya *et al.*, 2011; Spicer *et al.*, 2005), and strong pioneering attributes specifically towards these rock types (Mshumi, 2006; Nyakundya *et al.*, 2011). Thus, this species is expected to be predominantly encountered in the Mafic zone (Unit D1). Prominence of *V. karroo* on ultramafic rocks is ascribed to a higher tolerance by means of element exclusion, and will be tested for a higher Ca:Mg ratio and lower concentration of Cr, Co and Ni compared to co-occurring woody vegetation.

6.2 FLORISTICS

The MNDS ordination of the normalised species data produced an improved visual segregation in clustered groups (Figure 66) compared to the PCA ordination which illustrated some non-conforming plots (16, 19, 20, 25, 26 and 31; Figure 65). This is ascribed to dominant woody species which are not limited by a single geological substrate, and readily occur in more than one geobotanical unit. Overall, species assemblages could be distinguished from one another (Figure 66) based on relative abundances. The majority of species in a cluster had a preference for a specific rock chemistry (Table 8). Similar conclusions from both ordination techniques could be drawn in terms species variation across the soil and geological gradients.

Six plant assemblages corresponded with the proposed geobotanical units (Figure 66 & Table 8), and similarly with the soil chemistry. A clear distinction can be made between plant assemblages of quartzite dominated plots (Group A, B, and C) and ultramafic to felsic intrusions (Group D1, D2, and D3). It is evident that the presence of the intrusions does affect the floristic composition and several species associations could be identified that illustrated a significant dissimilarity across different substrate types (r -value = 0.6991; Table 9). No ultramafic endemic specialists or known hyperaccumulating taxa were encountered and the absence thereof is ascribed to reduced occurrence of ultramafic soil properties, owing to Ni dilution by colluvial quartzite weathering, leaching of available Mg and P, and elevation of Ca and K concentrations from weathering of the dioritic and felsic intrusions (spessartite, quartz alkali syenite and alkali granite).

The groups of quartzite-dominated plots (A, B, and C) illustrated significant species dissimilarities from one another ($P = <0.05$). This can be attributed to established variances in their soil chemistry, which is related to the topography, microclimate, and minor substrate variations as discussed in section 3.2.1.

Group A (Figure 66) corresponded with the alluvial floodplain and was dominated by *Vachellia karroo*, *Pyracantha angustifolia* (invasive species), *P. crenulata* (invasive species) and *Ziziphus mucronata*, as well as woody shrub species *Asparagus laricinus* and *Gymnosporia buxifolia*.

Group B corresponded with the southern aspect of the quartzite ridge and was dominated by *Celtis africana*, *Dombeya rotundifolia* and *Senegalia caffra*. Dominant shrubs included *Asparagus suaveolens*, *Diospyros lycioides* and *Grewia occidentalis*.

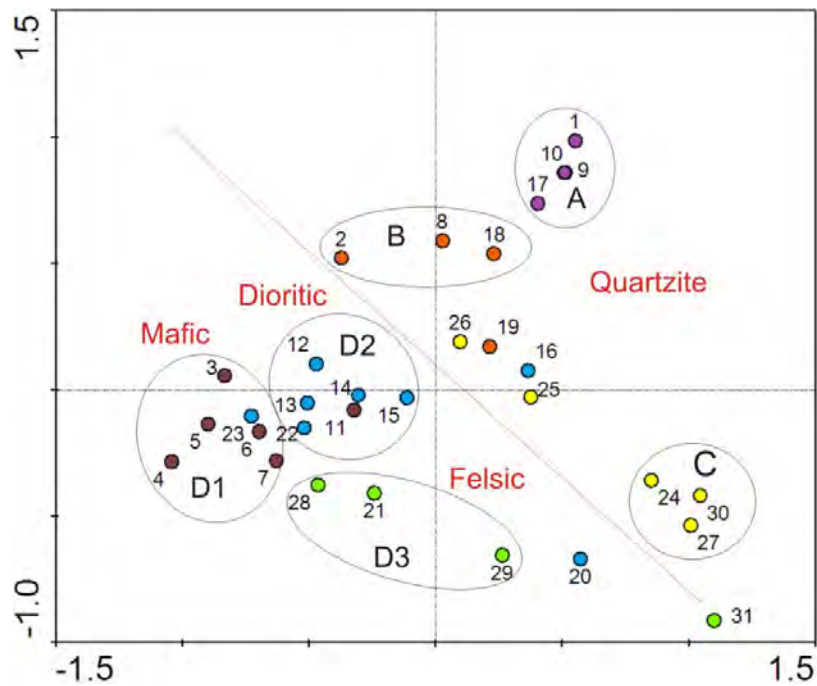


Figure 65: Principle correspondence analysis (PCA) of non-transformed floristic plot data of the Koedoesfontein Complex (colors correspond with geobotanical units; Figure 6). Clustering of sampling plots indicates correspondence in floristic composition. Major species differences are observed between quartzite dominated plots (Group A, B, and C) and felsic-ultramafic intrusions (Group D1, D2, and D3). Individual groupings correspond strongly with each geobotanical unit. Eigenvalues for axes one and two are 0.361 and 0.221 respectively.

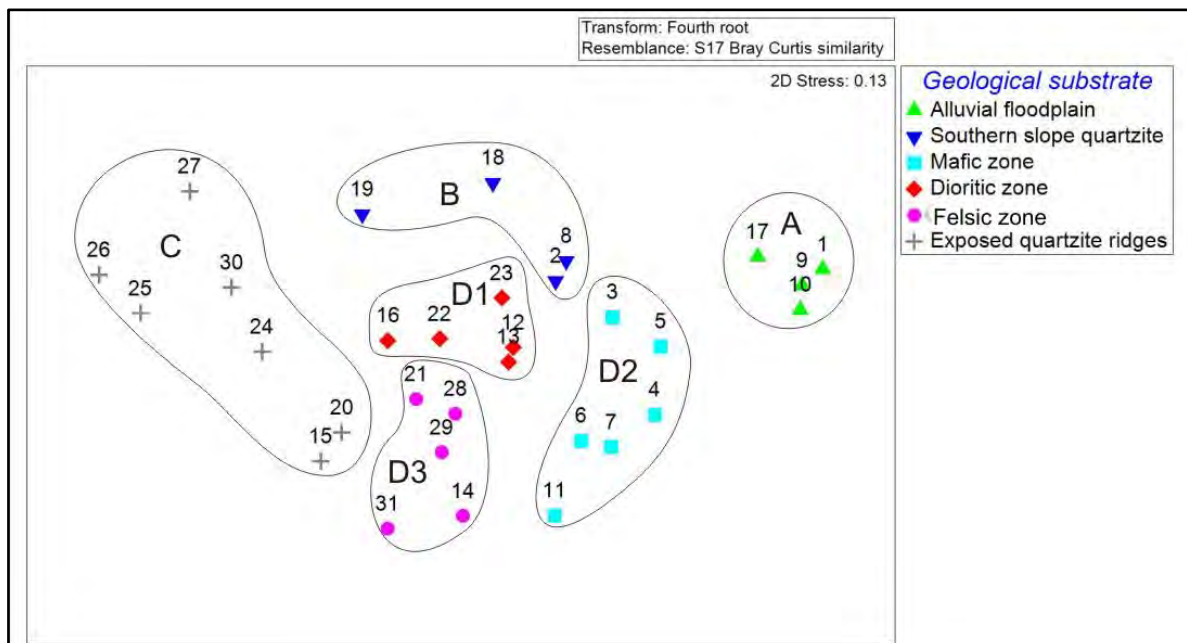


Figure 66: Non-metric multidimensional scaling (NMDS) analysis of transformed floristic plot data of the Koedoesfontein Complex. Similar to Figure 65, plots clustered according to their relevant geobotanical unit, signifying dissimilarity in woody species assemblages across different geological substrates.

Table 8: Average relative species abundance (%) for the identified geobotanical units of the Koedoesfontein Complex. Woody genera are ranked according to their largest relative abundance.

Woody genera encountered	Geobotanical Unit					
	A	B	C	D1	D2	D3
Growth form: Tree						
<i>Vachellia robusta</i>				10	14	72
<i>Combretum molle</i>			63		1	1
<i>Vachellia karroo</i>		13		56	9	3
<i>Pyracantha angustifolia</i> (invasive species)	39					
<i>Ziziphus mucronata</i>	33	18		23	9	1
<i>Senegalia caffra</i>		16	23		30	1
<i>Pyracantha crenulata</i> (invasive species)	28					
<i>Celtis africana</i>		21		4	1	1
<i>Dombeya rotundifolia</i>		6	6		10	1
<i>Pappea capensis</i>				3	9	15
<i>Searsia leptodictya</i>		2	6	1	7	13
<i>Buddleja saligna</i>		6				
<i>Euclea crispa</i>		5		2	6	13
<i>Mystroxydon aethiopicum</i>		4				
<i>Searsia pyroies</i>		3		2	1	
<i>Gymnosporia buxifolia</i>		3			1	
<i>Zanthoxylum capensis</i>					1	
<i>Olea europaea</i> subsp. <i>africana</i>			1			
<i>Brachylaena rotundata</i>			1			
<i>Pavetta zeyheri</i>				1	1	
<i>Melia azederach</i> (invasive species)						
Growth form: Shrub						
<i>Asparagus laricinus</i>	73	8		7	1	3
<i>Asparagus suaveolens</i>	7	48	5	56	39	50
<i>Pavetta zeyheri</i>			47	1	1	2
<i>Ehretia rigida</i>		4		15	23	31
<i>Vangueria parvifolia</i>			24		1	
<i>Grewia occidentalis</i>		14		4	20	1
<i>Nuxia congesta</i>		3	16		1	
<i>Grewia flava</i>		1		5	10	14
<i>Diospyros lycioides</i>		14	7	1	1	
<i>Gymnosporia tenuispina</i>				13	2	
<i>Gymnosporia buxifolia</i>	12	6	1		2	
<i>Cestrum laevigatum</i> (invasive species)	8					
<i>Searsia rigida</i>		3			1	

Table 9: Statistical significance ($P < 0.05$) of species dissimilarity between geobotanical units, based on square-root transformed data (ANOSIM). Mean rank within plots = 94.23, Mean rank between plots = 256.3, and R-value = 0.6991.

	A	B	C	D1	D2
B	0.0294				
C	0.003	0.0026			
D1	0.0043	0.004	0.0009		
D2	0.0082	0.032	0.0019	0.0014	
D3	0.0072	0.0068	0.0051	0.0059	0.024

Group C was mainly characterized by *Combretum molle*, *Olea europaea* subsp. *africana*, and *S. caffra* as the dominant trees, and *Nuxia congesta*, *Pavetta zeyheri* and *Vangueria parvifolia* as the prevailing shrubs.

Greater dissimilarities in species composition were observed between intrusive rock types (D1, D2 and D3) and quartzite rocks ($P = < 0.05$), than among the different intrusive rock types themselves ($P = < 0.05$). This was also indicated by less variation among their clustered plots in Figure 65 and Figure 66.

Assemblages of intrusive rock types were characterized by the tree species *S. caffra*, *V. karroo*, and *V. robusta*, which dominated these areas. Woody shrubs that were also predominantly encountered included *A. suaveolens*, *Ehretia rigida*, *Grewia flava*, *G. occidentalis* and *Gymnosporia tenuispina*.

With respect to succulent taxa, it is interesting to note that *Aloe greatheadii* (Figure 67) had a very high frequency within these units (D1, D2 and D3). This species, in particular, is endemic to the Witwatersrand (Reddy et al. 2001) and known to occur on ultramafic rock types. Although the species is not restricted to intrusive rock types, it does show prevalence on units containing mafic-dioritic intrusions specifically gabbro (D1) and epidiorite (D2).

Species groups of Units D1, D2 and D3 could be distinguished from one another on the basis of the dominance of specific *Senegalia* or *Vachellia* species, as well as the relative abundance between *Pappea capensis* and *Ziziphus mucronata*. Encountered shrub species also illustrated minor variations in relative abundance (Table 8).

Unit D1, which was located on ultramafic, mafic and dioritic intrusions, was mainly characterized by *V. karroo* and *Z. mucronata*, as well as the absence of *S. caffra*. In areas where dioritic spessartite was more abundant, however, *V. robusta* became more prevalent (Plots 6 and 11; Figure 65 and Figure 66). Dominant shrubs included *A. suaveolens*, *E. rigida* and *G. tenuispina*.

Unit D2, which contains only dioritic epidiorite, had a greater similarity in species composition to plots containing felsic rock substrates (D3) ($P = 0.024$) than to plots containing ultramafic and mafic rock types ($P = 0.0014$). The epidiorite intrusion was mainly inhabited by *S. caffra* and *V. robusta*, whereas *V. karroo* and *Z. mucronata* were generally absent or sparsely distributed. Shrub species including *A. suaveolens* and *G. tenuispina* were relatively less abundant (Table 8), whereas *E. rigida*, *G. flava* and *G. occidentalis* had higher relative abundances.

Unit D3 (on felsic rock), was dominated by *P. capensis* and *V. robusta*, which exhibited particularly high densities. *S. caffra* and *V. karroo* were also present, although in lower abundances. *G. tenuispina* was completely absent among the shrub species, whereas *A. suaveolens*, *E. rigida* and *G. flava* had higher relative abundances compared to Unit D2.



Figure 67: A photo of the succulent species *Aloe greatheadii*, occurring on the exposed sill of epidiorite on top of the eastern quartzite ridge. Here, the species can be seen acclimated to the bare patches of rocky colluvial soils, along with some smaller hemicryptophytes and grass species such as *Eragrostis curvula*. Note that the small tree in the background is *Vachelia karroo*.

6.3 PLANT-SOIL RELATIONS

CCA ordinations of the adult-species data disclosed several groups of species that correlated with specific soil chemical variables related to the geology. In addition, the evaluation of performance-index values for *Senegalia* and *Vachellia* species reflected the effect of these associations, specifically with respect to ultramafic rocks, on the general productivity of trees (section 6.4). These associations explain why certain floristic changes were observed, and can aid in identifying certain geological- and soil chemical- environments (Table 10).

The CCA ordinations (Figure 68) show that several species are correlated with low soil pH, overall nutrient status and low concentrations of Fe, Mn, Cr, and Ni. These include *B. rotundata*, *C. molle*, *N. congesta*, *O. europaea* subsp. *africana*, *P. zeyheri* and *V. parvifolia*. These species were indicative of warm quartzite environments with slow weathering rates, and high soil leaching capabilities. This observation is consistent with the findings of Bezuidenhout and Bredenkamp (1988).

Species mainly restricted to the alluvial flood plain were correlated with higher Mg availability and total Co content, including *A. laricinus*, *Gymnosporia buxifolia*, *P. angustifolia* and *P. crenulata*. These species were indicative of deeper, highly weathered, mixed alluvial quartzite soils, with higher moisture content due to higher amounts of clay-sized particles.

Apart from the alluvial floodplain, *Diospyros lycioides* subsp. *guerkei* and *Euclea crispa* subsp. *crispa* were correlated with Co in the shallower, more organic, shaded quartzite environment.

Species that correlated with low Ca and Na availability, and high values of Fe, Mn, Cr and Ni include *A. suaveolens*, *Searsia pyroides*, *V. karroo* and *Z. mucronata*. These species were indicative of ultramafic and mafic rock types and more adapted to environments with high Mg availability and high concentrations of phytotoxic trace elements.

S. caffra and *G. occidentalis* correlated with the availability of Ca and are mainly restricted to areas with higher soil Ca:Mg ratios. These include shaded quartzite environments, as well as areas containing hornfels, felsic and dioritic rock types.

Species that correlated with the availability of Na include *D. rotundifolia*, *E. rigida*, *G. flava*, *P. capensis* and *V. robusta*. These species are indicative of alkaline felsic and dioritic intrusive rock types specifically containing higher amounts of Na- rich feldspars.

Table 10: Summary of the woody species associated with specific geological and soil chemical environments of the Koedoesfontein Complex.

GEOLOGY		SOIL CHEMISTRY							CORRELATED WOODY SPECIES
Chemical type	Lithology	pH	Nutrient availability	Mean metal content based on XRF analysis (mg/kg)* ²					
			(Mean base saturation)* ¹	Fe	Mn	Cr	Ni	Co	
Acid/Felsic/Quartzose	Quartzite (Exposed)	Very acid (4 - <5)	Very Low (20%)	Moderate (~22 000)	Moderate (~400)	Moderate (~200)	Moderate (~30)	Very high (~160)	<i>Brachylaena rotundata</i> , <i>Combretum molle</i> , <i>Nuxia congesta</i> , <i>Olea europaea</i> , <i>Pavetta zeyheri</i> , <i>Vangueria parvifolia</i> , <i>Diospyros lycioides</i> , <i>Euclea crispa</i> .
Ultramafic - Mafic	Wehrlite/ Ol ± Clinopyroxenite, Hornblendite, Opx-Hbl Gabbro,	Moderately acid-slightly acid (5 -<6)	Moderate-High (80%)	Very High (~70 000)	High (~1200)	High (~400)	High (~130)	Moderate (~50)	<i>Asparagus suaveolens</i> , <i>Saersia pyroides</i> , <i>Vachellia. karroo</i> , <i>Ziziphus mucronata</i>
Dioritic	Epidiorite, Spessartite		Very high (90%)	High (~49 000)	High (~1000)	Moderate (~200)	High (~110)	Moderate (~60)	<i>Senegalia caffra</i> , <i>Dombeya rotundifolia</i> , <i>Ehretia rigida</i> , <i>Grewia flava</i> , <i>Grewia occidentalis</i> , <i>Pappea capensis</i> , <i>Vachellia robusta</i> .
Felsic	Alkali granite, Quartz alkali syenite	Slightly acid - Slightly alkaline (6-7.2)	Very High (110%)	High (~49 000)	High (~1200)	Moderate (~200)	High (~130)	Moderate (~60)	

*1: The mean base saturation (%) refers to the general availability of Ca, Mg, K, and Na. P and N values are not considered, as they were primarily related to the soil organic material.

*2: Moderate values rank between the world average, and maximum allowable value (MAV) for agricultural soils (Hazelton and Murphy, 2007; Kabata-Pendias, 2011). High values rank above the MAV towards what is considered a minimum/low trigger action value (TAV; Kabata-Pendias, 2011), Very high values are considered moderate to severe TAV's.

6.4 SPECIES PERFORMANCE

The abundance of *Senegalia* and *Vachellia* species and their strong association with the geology (Table 8 and Table 10) makes them ideal indicators to detect changes in geological substrates. It was observed that these species illustrate contrasting associations with Mg and Ca availability as well as the soil Ni, Cr and Co concentrations. These different preferences may reflect differences in evolutionary adaption to specific geological and pedological environments. The majority of the woody species (including *S. caffra* and *V. robusta*) are mainly restricted to Na-Ca-rich dioritic and felsic rock types. However, *V. karroo* tends to dominate the sparsely vegetated Mg-, Ni-, Cr- and Co-rich ultramafic outcrops and seems unaffected.

The species performance indices of the different *Senegalia* and *Vachellia* species (Figure 69 and Figure 70) depicted progressive or regressive performance to changes in the Ca:Mg ratio, and abundance of Cr, Co, and Ni across the different geological substrates. The index value is derived from the abundance and stem circumference of each species, and a higher value indicates higher species performance and, thus, tolerance of greater Mg availability and high concentrations of the relevant soil trace elements.

Statistical regression did not reveal any strong conclusive trends ($R^2 < 0.4$) across all sampled individuals of the related species. This is because each individual is affected by a unique combination of biotic and abiotic factors, which vary quite broadly along the study area. Despite this, it was observed that each species showed a particular tolerance range or niche that corresponded to a specific geology.

With respect to the Ca:Mg (Figure 69), *V. karroo* had the highest index values at ratios lower than 3.5 and, thus, showed the highest productivity at high soil Mg availability. When the ratio increased, a decline in productivity was observed for *V. karroo*. This is ascribed to increased interspecific competition with *S. caffra* and *V. robusta*, as higher ratios between 3.5 and 6 indicated higher productivity for the latter two in areas with higher Ca availability. *S. caffra*'s niche specificity occurred around the 3.5 ratio, which was the transitional point between that of *V. karroo* and *V. robusta*. This partly explains why *S. caffra* is particularly prevalent on dioritic rock types, whereas *V. robusta* was more abundant on the felsic substrates. *V. robusta* particularly had the highest performance indices (>14) of all three species above the 3.5 ratio, with a niche specificity at ratios from 4 to 4.5.

Senegalia caffra and *V. robusta* also exhibited higher index values at lower Cr (<250 mg/kg; Figure 70a), Co (<125 mg/kg; Figure 70b) and Ni (<140 mg/kg; Figure 70c) concentrations, whereas lower index values were observed for *V. karroo* at these sites despite more fertile soil conditions (higher K, Na and P). This indicated that *V. robusta* and *S. caffra* are able to

outcompete *V. karroo* in soils originating from more felsic and dioritic rock types. At higher metal concentrations, the index values for *S. caffra* and *V. robusta* decreased, and a clear threshold value was reached. Conversely, *V. karroo* became more productive at higher concentrations of Cr (>400 mg/kg), Co (>150 mg/kg) and Ni (>100 mg/kg), and was thus able to outcompete the other species on ultramafic and mafic rock types. The concentration of Cr, Co and Ni associated with the abundance of mafic minerals (olivine, serpentine and pyroxene) could, therefore, determine the dominant species within the landscape. This relationship proved extremely useful as an indicator tool to determine the relative extent and general orientation of the partly unexposed intrusions of the KC (Figure 9; Section 4.2).

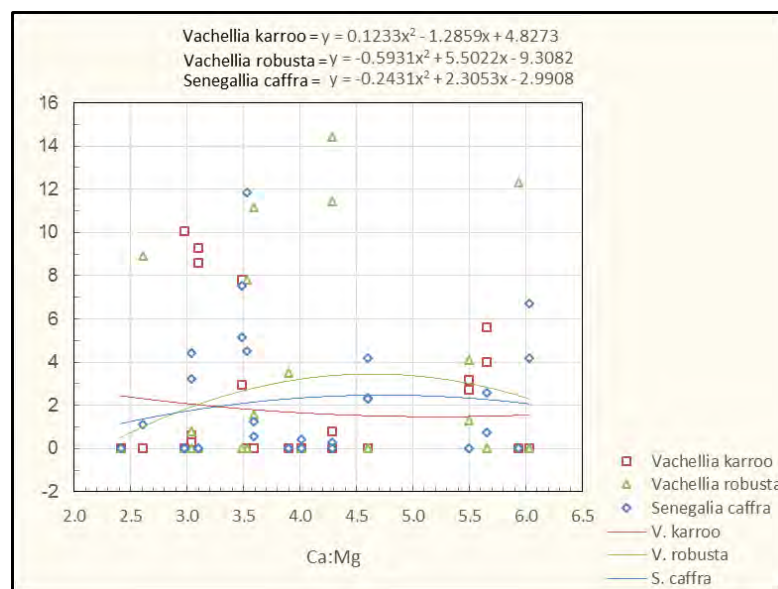


Figure 69: Scatterplots of the performance index values of *Senegalia* and *Vachellia* species versus the exchangeable soil Ca:Mg ratio. *V. karroo* displays the highest index values (based on abundance and stem circumference) at low ratios (2.4-3.5), which tend to decrease along the gradient ($R^2 = 0.008$) as Ca becomes relatively more available. *V. robusta* ($R^2 = 0.03$) and *S. caffra* ($R^2 = 0.01$) alternatively display lower index values at low ratios, which tend to increase from 3.5. R^2 values were calculated based on polynomial trends.

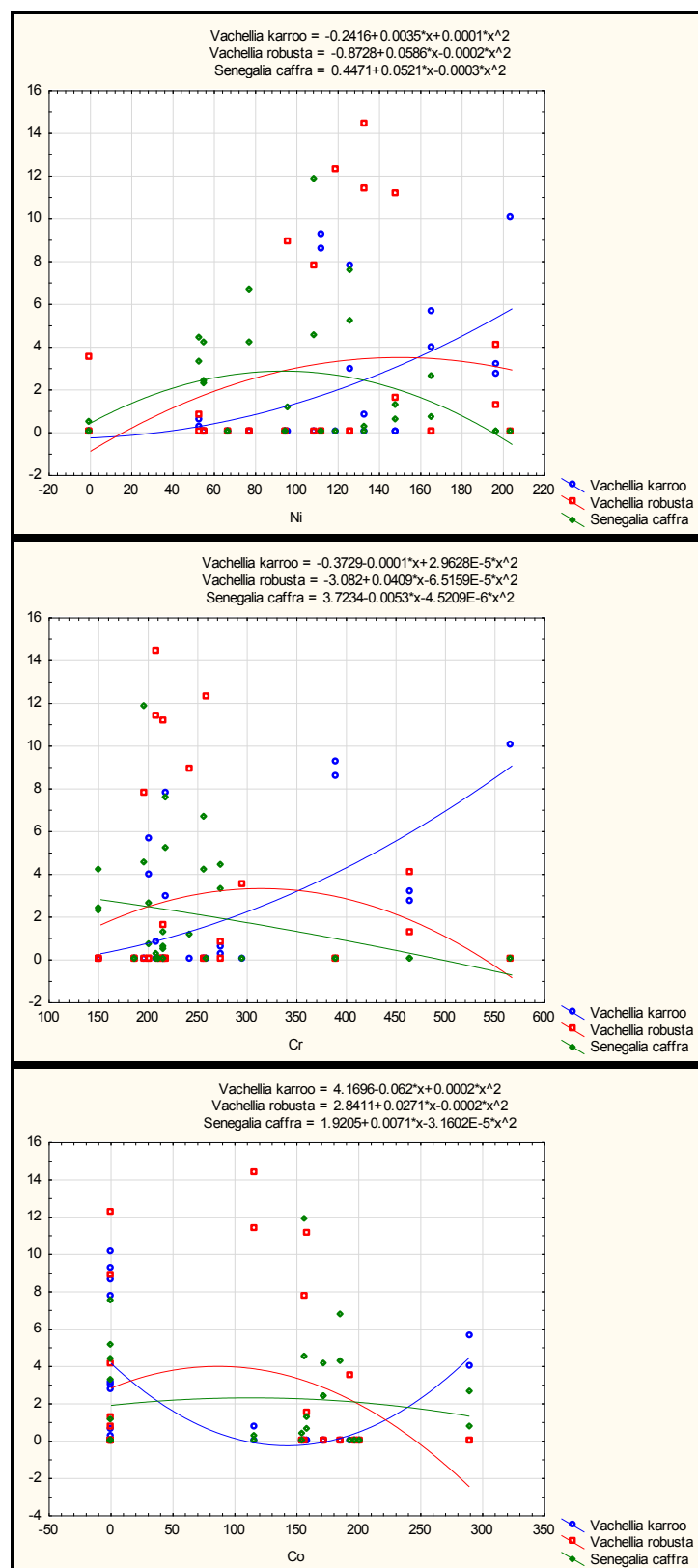


Figure 70: Scatterplots of the performance index values of *Senegalia* and *Vachellia* species versus total concentrations (mg/kg) of (a) chromium (Cr), (b) cobalt (Co) and (c) nickel (Ni). Index values for *V. karroo* tend to increase along the Cr ($R^2 = 0.40$), Co ($R^2 = 0.44$) and Ni ($R^2 = 0.28$) gradient, whereas index values for *V. robusta* (Cr, $R^2 = 0.31$, Co, $R^2 = 0.03$, Ni, $R^2 = 0.08$) and *S. caffra* (Cr, $R^2 = 0.08$, Co, $R^2 = 0.007$, Ni, $R^2 = 0.13$) tend to decrease, indicating lower trace-element tolerances. R^2 values were calculated based on cubic trends

6.5 PLANT COMPOSITION AND ELEMENT ASSOCIATION

It is presumed, due to the presence of ultramafic soil characteristics, that the distribution of some of the dominant woody species, principally the *Senegalia* and *Vachellia* species, are strongly governed by their ability to tolerate less favourable soil conditions to avoid competition or to exploit more optimal ones provided by the contrasting igneous lithologies. To test this assumption, the leaf and stem tissue of selected dominant woody species (*G. flava*, *S. caffra*, *V. karroo* and *V. robusta*) were analysed to reveal specific biochemical trends of the nutritive and phytotoxic elements. The main focus of this study is placed on the major elements Ca and Mg, and the trace elements Cr, Co and Ni. These elements were selected in order to determine how the dominant species react with regards to the "serpentine effect", which is caused by the presence of weathering ultramafic outcrops.

For a more detailed summary of the total and soluble element compositions, refer to Appendix F2 for soil (KSB) samples and Appendix G1 for plant tissue samples. The soil composite (KSB) samples, were analysed with the same method as for the plant material and serves only as a reference to determine element source and uptake trends. These samples are therefore more representative of the point specific area where the vegetation was sampled, and not necessarily the entire geobotanical unit as was interpreted from a larger analysed data set and discussed in Chapter 5.

6.5.1 Major elements (Ca & Mg)

Ca and Mg variations of the selected species are illustrated across geobotanical units associated with the main groupings of igneous rocks (Figure 71, Figure 72 & Figure 73). In general, the concentration of major elements, in the leaves and stems of all the selected species, was found to be typically closer to the total values that were reported for the corresponding soils.

The Ca content (Figure 71) was particularly higher for all four species (~6700 to 20800 mg/kg) than their corresponding total soil values (~3200 to 8200mg/kg). *V. robusta* accumulated the highest concentration of Ca throughout the different geobotanical units (~12800 to 20800 mg/kg), followed by *V. karroo* (~10100 to 16000 mg/kg), *S. caffra* (~8200 to 12800 mg/kg) and *G. flava* (~6200 to 13100 mg/kg).

Vachellia karroo and *V. robusta* tend to have less variable distributions of Ca between their leaves and stems, whereas *G. flava* and *S. caffra* had higher amounts of Ca in their leaves. The leaf Ca content of *G. flava* (10500 to 13100 mg/kg) was specifically almost double that of its stems (~6200 to 8300mg/kg).

All tree species of the mafic zone in geobotanical unit D1 (*G. flava*, *V. karroo* and *V. robusta*) had suppressed Mg concentrations (~1400 to 3100 mg/kg) below the total soil value (~3800 to 4000 mg/kg). Nevertheless, in the dioritic and felsic units (units D2 and D3), similar concentrations to the mafic zone was maintained (~1300 to 3000 mg/kg) despite containing lower amounts of total Mg (~1200 to 1500 mg/kg). This observation might suggest that the Mg content of the plants is more affected by the available portion in the soil, which was more uniform throughout the different geobotanical units (~300 mg/kg).

In the dioritic and felsic zones, Mg was concentrated above the total soil value in the stems and leaves of most species, especially *S. caffra* (~3700 to 2500mg/kg). However, not so in the stems of *G. flava* (~1400 mg/kg), and *V. karroo* (~1300 mg/kg). This is because all four species tended to allocate more Mg to their leaves, and *V. karroo* seems to be the most efficient in this process since it contained the lowest stem concentration of Mg (~1300 to 1500 mg/kg), followed by *G. flava* (~1400 to 1600 mg/kg), *V. robusta* (~1700 to 2600 mg/kg) and then *S. caffra* (~2500 to 2600 mg/kg). From this observation it can be deduced that all the species to some extent control their Mg content via leaf litter, however, in the felsic zone where Ca was more available, this mechanism was less apparent amongst the species apart from *V. karroo*.

Considerable variation was observed across the Ca:Mg ratios due to the different management of Ca and Mg (Figure 73). The tissue ratios were more similar to the soluble ratio of soil, which suggests that the species accumulate these elements at the ratios at which they are usually available, and over time concentrate them differently based on the soil conditions. Species that were more dominant in the mafic zone (*V. karroo* and *V. robusta*) for example, displayed ratios ranging from similar to well above the soil soluble fraction, and are thus able to increase Ca uptake and suppress the Mg accumulation. Species more dominant in the dioritic and felsic zone (*G. flava* and *S. caffra*) on the other hand, had ratios that were similar to far below the soil soluble fraction, indicating a lower uptake of Ca and higher accumulation of Mg.

Vachellia robusta contained the highest Ca:Mg ratio in both its leaves (~6 to 9) and stems (~7 to 11), within the mafic zone and dioritic zone, and second highest, after *V. karroo*, in the felsic zone (leaves ~5 and stems ~6). This high Ca:Mg ratio, can explain why it occurs along with *V. karroo* within the mafic zone, as it is able to have high Ca uptake despite high concentrations of soil Mg.

Apart from excluding Mg below the total soil value, the mechanism of *V. robusta* to maintain a high ratio is specifically focused more on the Ca content, since it is accumulated in larger values compared to the co-occurring genera. This is probably why this species is more prone to occur on the dioritic and felsic rocks as it needs supplemental sources of Ca to maintain a higher ratio. This was confirmed by the species performance index values, which indicated a better performance for *V. robusta* at higher soil Ca:Mg ratios.

Since *V. robusta* distributed Ca evenly between its leaves and stems and moderately allocated more Mg to its leaves, a slightly lower ratio was observed for the leaves compared to its stems. This is, however, the opposite in the felsic zone where the species contained higher amounts of Mg in its stems when larger amounts of soil Ca was available.

The results for *V. karroo* remained more constant throughout the different geobotanical units. In the mafic zone, this species contained the second highest Ca:Mg ratio in its stems (~9), but the lowest ratio in its leaves (~5). Unlike *V. robusta*, the manner with which the Ca:Mg ratio is maintained in *V. karroo*, is focused on both the Ca and Mg content. With respect to Ca the species has a higher elemental accumulation compared to *G. flava* and *S. caffra*, but couples this with suppression of Mg and a higher tendency to allocate larger amounts of this element to the leaf tissue, probably in order to remove excess amounts as leaf litter. As a result, in the mafic zone, a lower ratio is maintained in the leaf tissue (~5) compared to the co-occurring vegetation (~7 to 9). This ability allows *V. Karroo* it to lower its Mg content while simultaneously conserving Ca within its stems. This is an ideal ability for soils where the high Mg concentration may suppress Ca availability as a result of the source material's geochemistry (Hazelton & Murphy, 2007), and thus clarifies why this species is particularly dominant on the ultramafic outcrops, unlike *V. robusta* that cannot conserve Ca but requires it from substrates (rock types) containing higher Ca:Mg ratios.

In the felsic zone, *V. karroo* was able to maintain an even lower Mg content, but allocated more of this element along with Ca to its leaves. This resulted in a similar leaf ratio (~5) but at the cost of a slightly lower stem ratio (~8). Both of these values were still higher than the leaves (~3 to 5) and stems (~3 to 6) of the co-occurring vegetation, and therefore *V. karroo* is also easily acclimated to the felsic intrusions.

In the mafic zone, *G. flava* contained the second highest Ca:Mg in its leaves (~9), but the lowest ratio in its stems (~5), whereas in the felsic zone this species had the second lowest ratio in its leaves (~4) and stems (~4). Similarly to *V. karroo* and *V. robusta*, this species also tends to suppress its stem's Mg content and allocate most of it to its leaves. Although this is done more effectively than in *V. robusta*, the Ca content is not accumulated as high as in the former species and thus significantly lower ratios are typically observed for its stems. *G. flava* is therefore much less capable of maintaining a high Ca: Mg ratio in the mafic zone than the co-occurring genera. This partly explains why this species is much less abundant in this unit compared to dioritic and felsic zones.

In the mafic and dioritic zones, a secondary mechanism is observed whereby *G. flava* also allocates almost double its Ca content to its leaves, possibly to increase the Ca:Mg ratio to a suitable level. When this mechanism is compared with that of *V. karroo*, the additional Ca seems

to be wasted when the Mg is then removed by leaf fall. However, when the phenology of the two species is considered this process makes more sense.

Species such as *V. karroo*, *V. robusta*, and *S. caffra* have small compound leaves, which, individually, may require far less nutritional expenditure to be replaced compared to the fewer large simple leaves of *G. flava*. *G. flava* can therefore be expected to be much less likely to drop its leaves due to element toxicity, and therefore rather corrects the ratio in the expense of its stem's Ca content.

Senegalia caffra had the lowest Ca:Mg ratios of all the species in both its leaves (~3 to 4) and stems (~3 to 4). This can be particularly ascribed to its higher Mg content, since this species was found to suppress the least amount of this element compared to the total soil value, and does not allocate it to its leaves as effectively as the co-occurring species. This partly explains why *S. caffra* avoids Mg-rich ultramafic rocks, and is not readily encountered in the mafic zone since it most likely cannot maintain a high ratio such as *V. karroo* and *V. robusta*.

Compared to *G. flava*, however, *S. caffra* does accumulate more Ca, but similar to the former species most of it is allocated to its leaves, thus, greater amounts are lost via leaf fall. Like *V. robusta*, this species is more abundant on dioritic rocks, and based on the results, this association might be ascribed to a greater requirement for Ca, and avoidance of high Mg sources such as the ultramafic and mafic rocks.

6.5.2 Trace elements (Cr, Co, Ni)

Variations in the phytotoxic trace element composition (Cr, Co, & Ni), are depicted for the selected species and their related soils across the different geobotanical units (Figure 74, Figure 75 & Figure 76).

It was observed that all the dominant species have a tendency to accumulate the respective trace elements above their soil soluble values, but far below the total concentrations. This indicates that all the tested genera tended to exclude or suppress the uptake these specific trace elements.

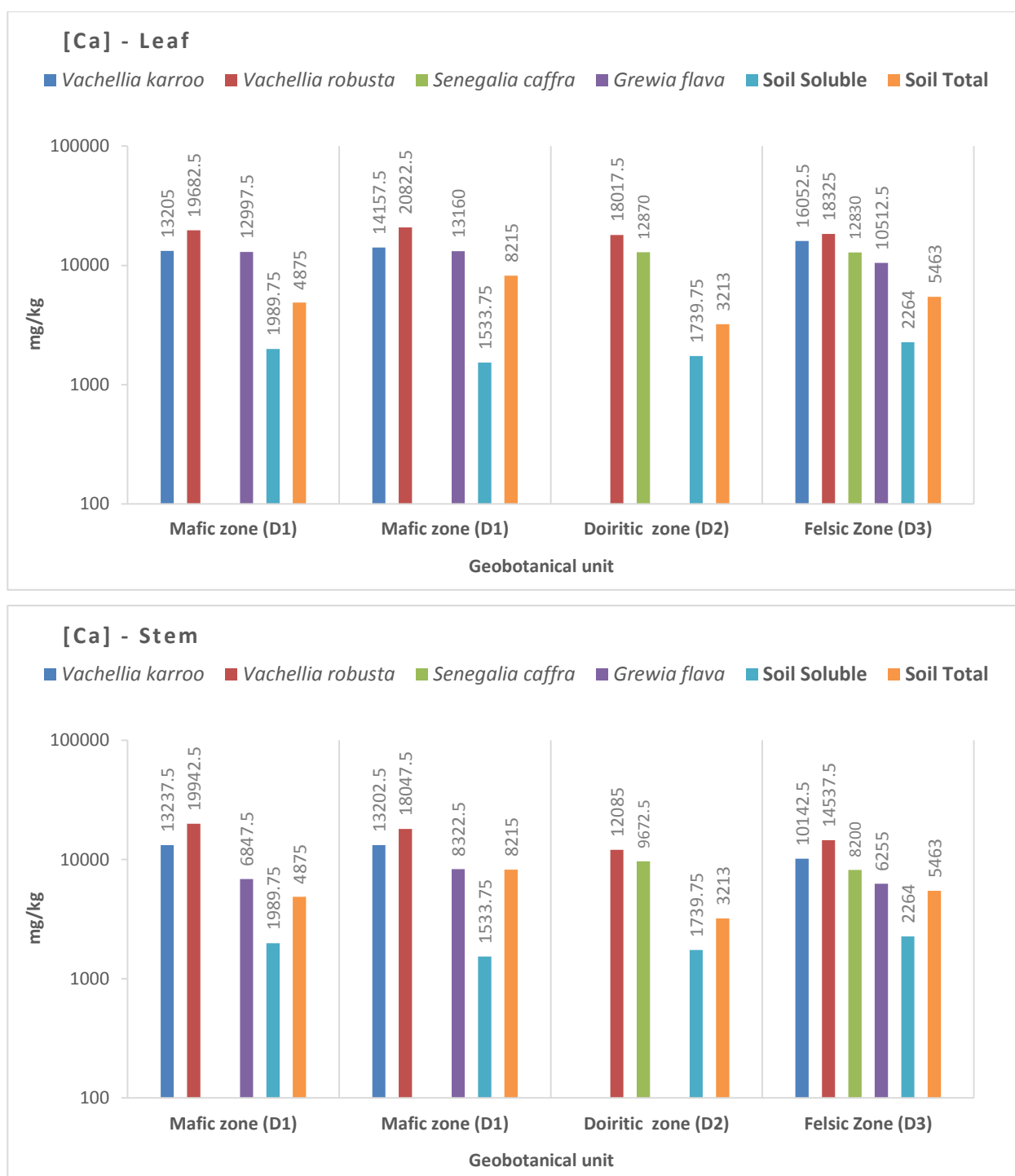


Figure 71: Total Ca content for the leaves and stems of the dominant woody species in each geobotanical unit, and related total and soluble fraction of their corresponding soil composite sample (KSB).

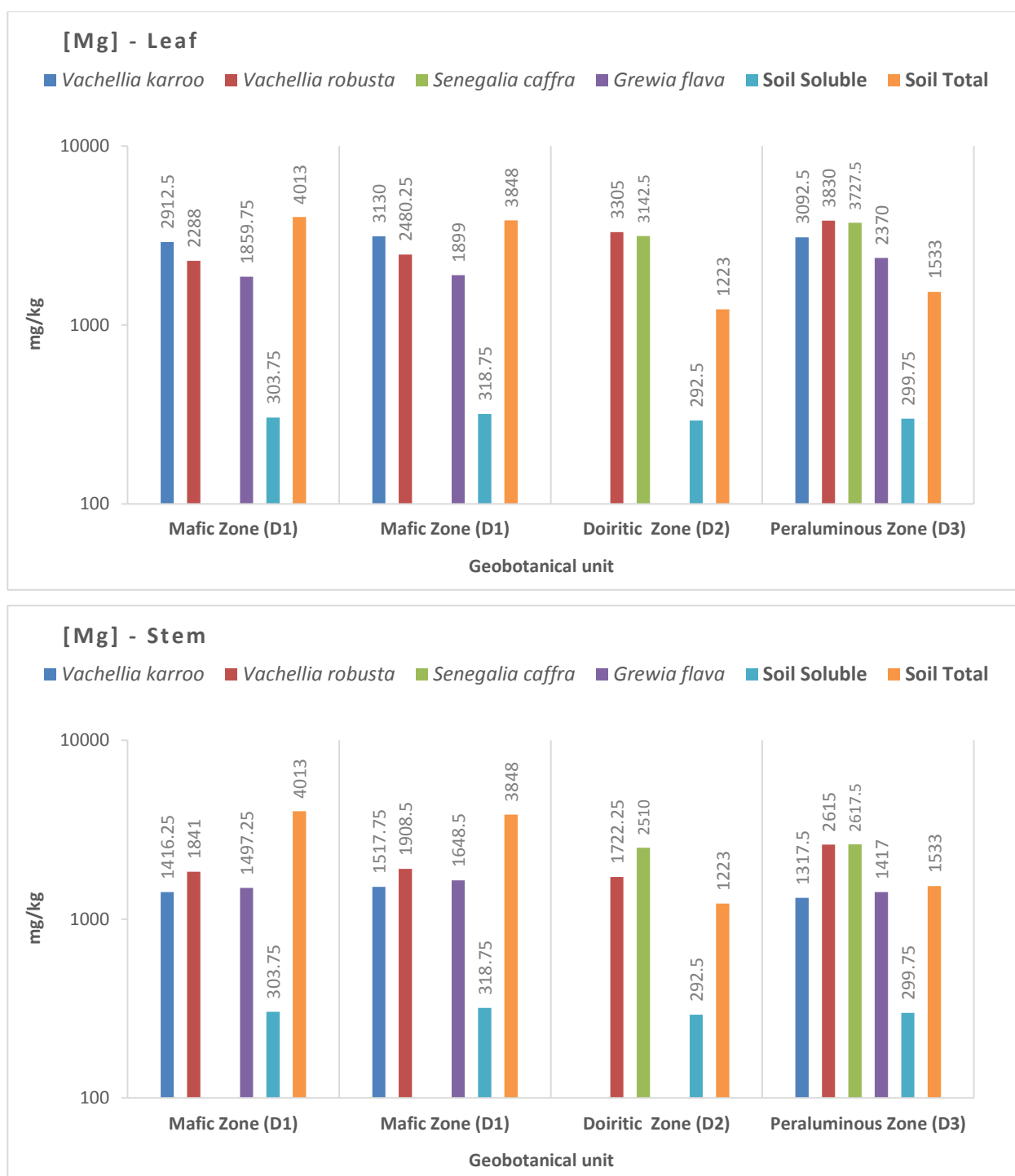


Figure 72: Total Mg content for the leaves and stems of the dominant woody species in each geobotanical unit, and related total and soluble fraction of their corresponding soil composite sample (KSB).

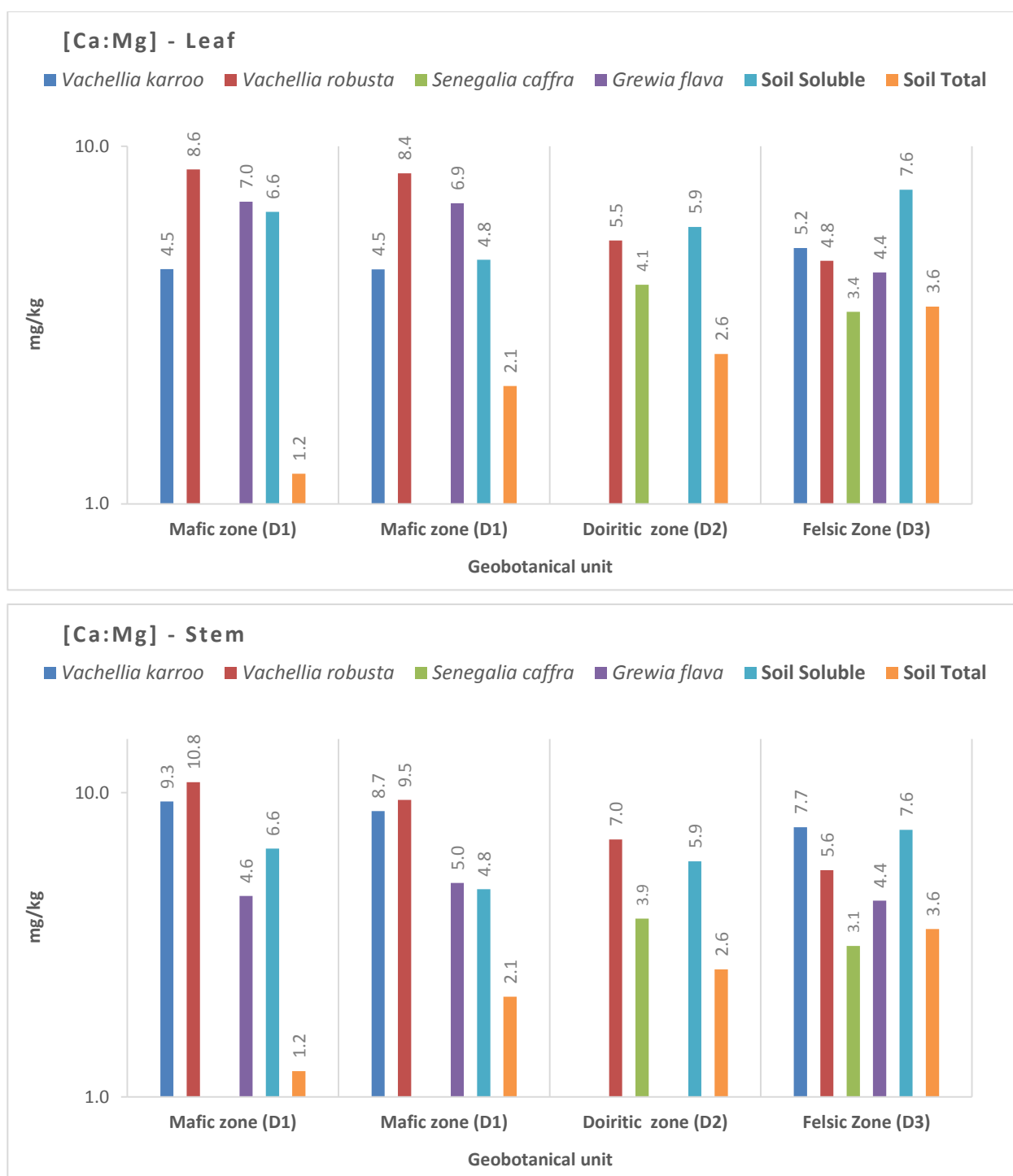


Figure 73: Total Ca:Mg for the leaves and stems of the dominant woody species in each geobotanical unit, and related total and soluble ratios of their corresponding soil composite sample (KSB).

Cr (Figure 74) typically occurred in all four species above and below the minimum toxicity value indicated for plant tissues (5 mg/kg- Orcutt & Nilsen, 2000; Kabata-Pendias, 2011). In the mafic zone, *V. karroo* contained more Cr in its stems (~5 to 10 mg/kg) compared to its leaves (~2 to 3 mg/kg), which was the lowest concentration of all three species. In the felsic zone, *V. karroo* contained a similar stem concentration (~5 mg/kg) but had a far higher leaf concentration (~11 mg/kg) above the toxicity value.

Vachellia robusta had a more evenly distributed Cr content between its leaves (~4 to 9 mg/kg) and stems (~4 to 11 mg/kg). In all the geobotanical units (apart from the areas with more dioritic spessartite-mafic zone) the stem concentrations ranged over the minimum toxicity value, this similar pattern was observed for the leaf content, however in the felsic zone the concentration was also below the minimum value (~4mg/kg).

Grewia flava illustrated rather contrasting concentrations between the mafic and felsic zones. In the mafic zone, *G. flava* maintained the highest Cr values of all three species in both its leaves (~6 mg/kg) and stems (~10mg/kg), while in the felsic zone this was exactly the opposite. The species had the lowest reported values for Cr that remained evenly distributed throughout the tissue samples (~3 mg/kg).

Senegalia caffra did not display much variation across the dioritic zone and felsic zone. Its leaf Cr content (~3 mg/kg) was particularly low and comparable to *G. flava*, while its stem concentration (~8 to 10 mg/kg) was the highest compared to all four species.

The Co content for all four species (Figure 75) was well below the indicated minimum toxicity value (15 mg/kg) (Orcutt & Nilsen, 2000; Kabata-Pendias, 2011). *Vachellia karroo* (~1 to 7 mg/kg) and *V. robusta* (~3 to 6 mg/kg) were more or less comparable, although generally *V. karroo* had the lowest leaf concentrations (~1 to 3 mg/kg), while *V. robusta* had lower concentrations in its stems (~3 to 6 mg/kg).

Grewia flava contained the highest concentration of Co in its leaves (~6 to 12 mg/kg) and stems (~7 to 10 mg/kg) throughout the mafic and felsic zones, while *S. caffra* had lower leaf (~2 to 4 mg/kg) and stem (~2 to 3 mg/kg) concentrations within the dioritic zone as well as the felsic zone. In these units the compositions were roughly comparable to *V. robusta* (~2 to 3 mg/kg).

All the species, except for *S. caffra* in the felsic zone (Figure 76), accumulated Ni below the minimum toxicity value of 10 mg/kg (Orcutt & Nilsen, 2000; Kabata-Pendias, 2011). *V.karroo* had the second lowest Ni content in its stems throughout the geobotanical units (~2 to 3 mg/kg), and in the mafic zone had the lowest leaf content (~2 to 3 mg/kg), followed by the second lowest (~2 mg/kg) in the felsic zone.

Vachellia robusta constantly had slightly higher values for Ni than *V. karroo*. It maintained the highest stem concentration (~4 to 5 mg/kg) and highest leaf concentration (~2 to 3 mg/kg) in the mafic zone, whereas in the dioritic zone it had a lower leaf content (~3 mg/kg) than *S. caffra* (~4 mg/kg), but contained a higher stem value (~3 mg/kg) than the former species (~2 mg/kg). In the felsic zone, *V. robusta* had the second highest leaf (~2 mg/kg) and stem values (~6 mg/kg) after *S. caffra* (~8 mg/kg and ~10 mg/kg respectively).

Grewia flava contained the lowest stem concentrations (~1 to 2 mg/kg) throughout the geobotanical units. Larger amounts of Ni were typically observed in the leaves, and as a result, this species contained one of the highest leaf concentrations in the mafic zone (~3 mg/kg). In the felsic zone the leaf content was far lower (~1 mg/kg) and similar to its stem concentration.

In terms of the ability to exclude phytotoxic trace elements, it was observed that *V. karroo* individuals were best adapted to the mafic zone where they contained lower amounts of Cr, Co and Ni in either the stems or leaves compared to *V. robusta* and *G. flava*. In the felsic zone, however, Cr and Co were increased above that of the co-occurring vegetation. Although this observation is still unexplained, it may be attributed to stress from increased interspecific competition. The higher trace element values can, nevertheless, clarify why this species has a lower performance index in the felsic zone, and is thus better suited to the mafic zone.

Vachellia robusta and *G. flava* were less effective in suppressing their trace element concentrations in the mafic zone, but seems to improve near dioritic and felsic rocks. *V. robusta* kept most of its elements in its stems, while *G. flava* allocated more of these elements to its leaves. In the felsic zone both these species were able to maintain lower concentrations in both their leaves and stems. These species are therefore more adapted to dioritic and felsic rocks, as suggested by their floristic patterns, and performance values.

Based on *S. caffra*'s trace element concentration, it was especially more adapted to the dioritic zone, where it maintained roughly a similar concentration than *V. robusta*. In the felsic zone, however, Cr and Ni were not suppressed as efficiently as the co-occurring species (*V. karroo*, *V. robusta* and *G. flava*).

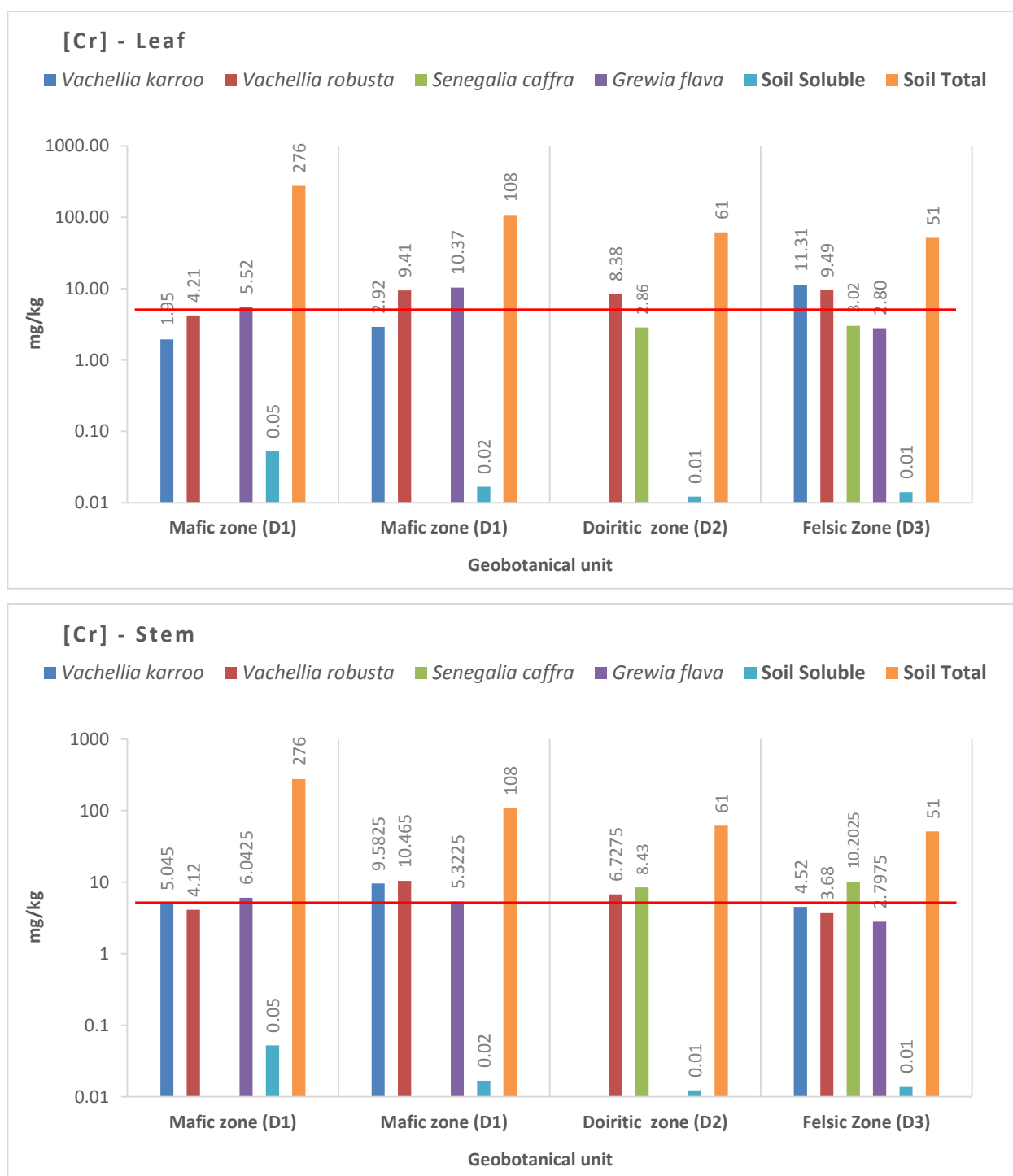


Figure 74: Total Cr content for the leaves and stems of the dominant woody species in each geobotanical unit, and related total and soluble fraction of their corresponding soil composite sample (KSB). The minimum tissue value for Cr toxicity (red line) is indicated at 5 mg/kg (Orcutt & Nilsen, 2000; Kabata-Pendias, 2011).

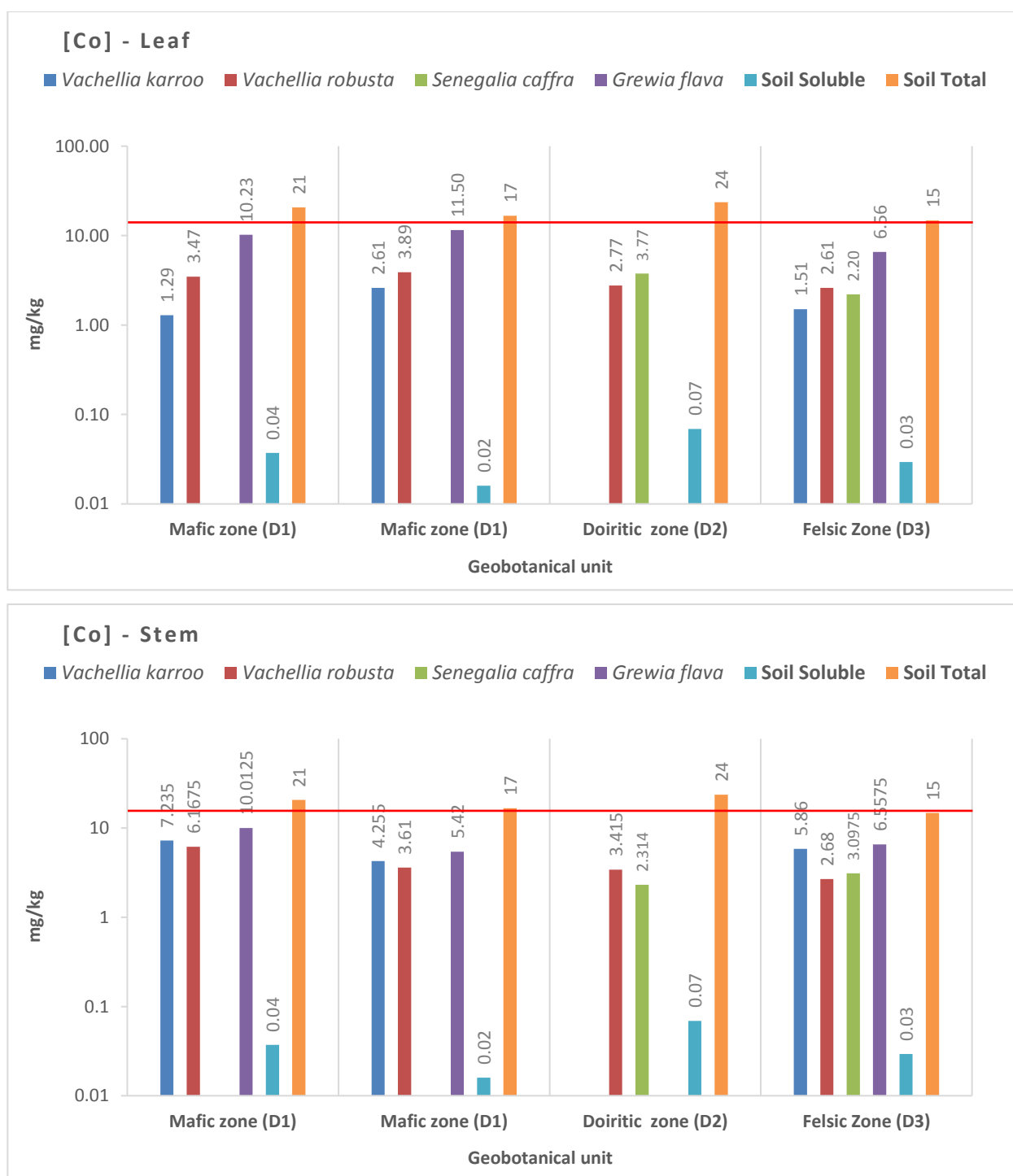


Figure 75: Total Co content for the leaves and stems of the dominant woody species in each geobotanical unit, and related total and soluble fraction of their corresponding soil composite sample (KSB). The minimum tissue value for Co toxicity (red line) is indicated at 15 mg/kg (Orcutt & Nilsen, 2000; Kabata-Pendias, 2011).

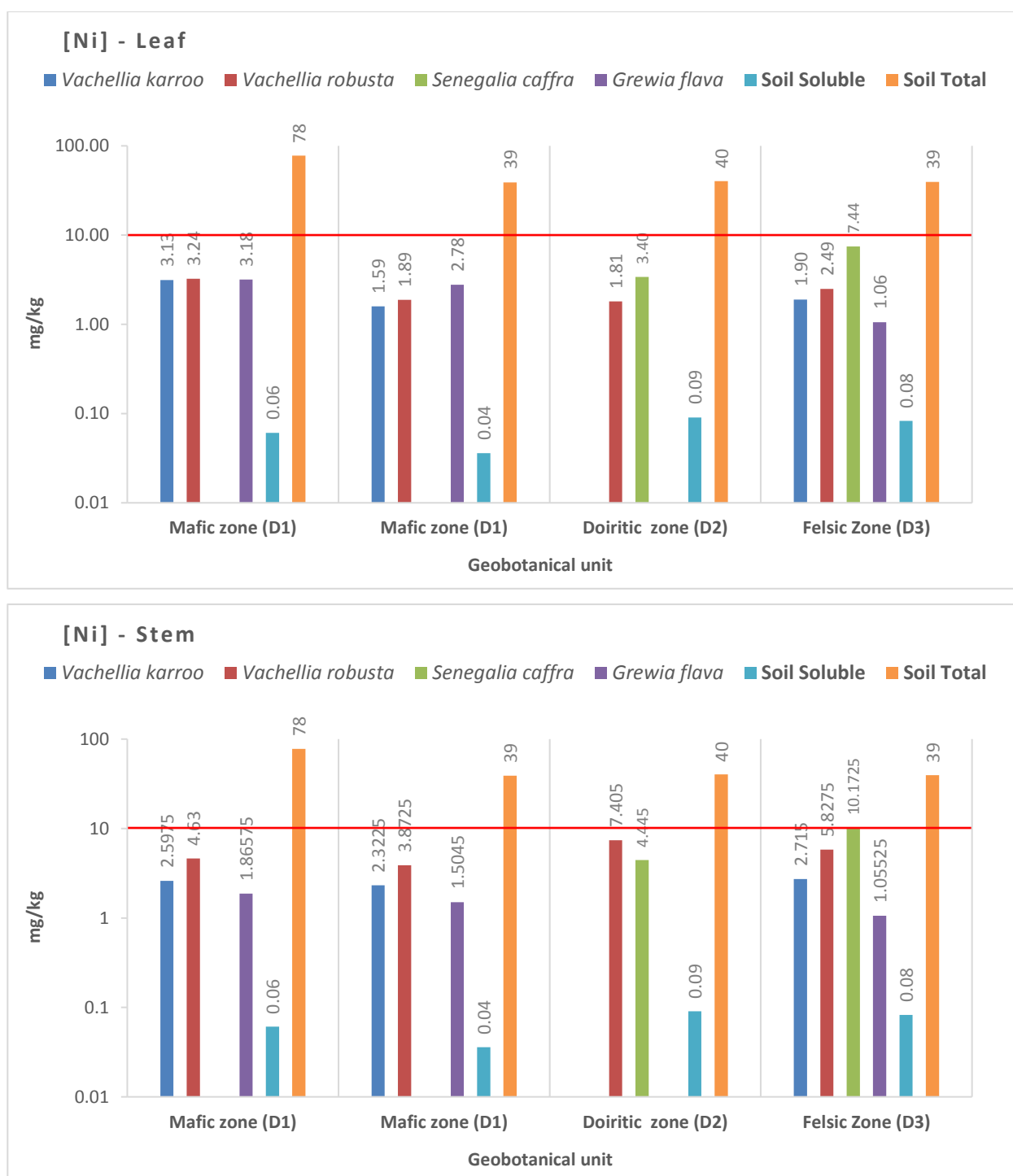


Figure 76: Total Ni content for the leaves and stems of the dominant woody species in each geobotanical unit, and related total and soluble fraction of their corresponding soil composite sample (KSB). The minimum tissue value for Ni toxicity (red line) is indicated at 10 mg/kg (Orcutt & Nilsen, 2000; Kabata-Pendias, 2011).

6.5.3 Element association

Major- and trace element variation patterns of the analysed species, displayed a clear difference in the concentration and distribution of specific elements. To determine if any of the elements show a particular association with one another, as well as other nutritive elements (K, Na, and P), a corresponding correlation matrix was constructed (Table 11). In all the tested species, the Cr, Co and Ni concentrations did not show any meaningful association with one another, but did however, correlate with some of the other major elements.

In samples of *V. karroo* there was a very strong association between Ni and Na ($r = 0.903$), as well as some association with Ca ($r = 0.723$) and K ($r = 0.706$). These elements tend to be retained together within the stems. Co showed some association with P, while Cr did not have any strong correlations with major elements.

In samples of *V. robusta*, Ni was negatively correlated with Ca ($r = -0.815$) and K ($r = -0.813$) because they were allocated more to the leaves, while larger amounts of Ni was retained in the stems. No association could be seen for Cr and Co.

In samples of *S. caffra*, Cr showed the strong positive association with K ($r = 0.986$) which was retained more in the stems. Ni correlated with P ($r = 0.859$), and Co did not show any association.

In samples of *G. flava*, only Ni showed a strong association with Mg ($r = 0.948$), which was both allocated to the leaves, but similar to *V. robusta*, no correlation could be seen for Cr and Co.

Table 11: Element correlation (r) matrix of analysed woody species. (a) *Vachellia karroo*, (b) *Vachellia robusta*, (c) *Senegalia caffra*, (d) *Grewia flava*. Positive correlations are shaded in green and negative correlations in red.

(a)	Ca	Mg	K	Na	P	Ni	Cr
Ni	0.723836	0.499088	0.706895	0.903036	0.655398	1	
Cr	0.453021	0.17359	0.140875	0.213458	0.313171	0.104057	1
Co	0.367567	-0.08799	0.367678	0.607316	0.406837	0.593168	0.1336
(b)	Ca	Mg	K	Na	P	Ni	Cr
Ni	-0.81561	-0.61707	-0.81359	0.27765	-0.19128	1	
Cr	0.189332	0.324123	0.487088	-0.36321	0.290126	-0.49661	1
Co	0.399092	-0.60651	-0.49678	-0.41742	0.156461	0.131659	-0.30586
(c)	Ca	Mg	K	Na	P	Ni	Cr
Ni	-0.54402	-0.03011	0.611462	0.18945	0.859197	1	
Cr	-0.99777	-0.86118	0.986124	-0.71327	0.13346	0.505492	1
Co	0.10509	-0.14172	-0.06489	-0.28526	0.195791	-0.21457	-0.14504
(d)	Ca	Mg	K	Na	P	Ni	Cr
Ni	0.4245	0.948573	0.029511	-0.1491	0.517973	1	
Cr	0.50699	-0.10455	-0.05389	0.371214	0.524808	-0.33577	1
Co	0.650726	0.320375	0.239192	0.415895	0.518165	0.25245	0.612346

6.6 CONCLUSION

Analysis of the floristic data, physiognomic features- and element composition, indicate that broad variation in geological substrates and related soil chemistry, does have a measurable effect on the woody plant population. Overall, the most abundant species, particularly *Senegalia* and *Vachellia* genera, illustrated a very strong degree of correlation with soil chemical elements that could be directly related to the geology. This explains why there is such a close association between the outcrop patterns and woody species assemblages.

Multivariate analysis of the botanical survey data (Appendix C1), revealed six significantly distinguishable woody species assemblages, which correlated with different groupings of rock substrates (section 6.2). A clear difference in the relative abundance of the different tree and shrub species could be observed, which illustrated some preferences amongst the different geobotanical units (Table 8). The plots dominated by quartzite bedrock (units A, B, and C), and the plots that were grouped based on their dominant igneous rock types (units D1, D2 and D3) could therefore be distinguished based on dominant woody genera.

Vachellia karoo showed a preference for the mafic zone (unit D1) which contained ultramafic, mafic and dioritic rocks, while *S. caffra* preferred the dioritic zone (unit D2) containing primarily epidiorite. *Vachellia robusta* was encountered throughout most of the igneous rocks but was particularly more dominant in the felsic zone (unit D3) containing quartz alkali syenite and alkali granite.

Direct gradient analyses illustrated specific associations between individual species and soil chemical variables related to the geology (section 6.3). In these ordinations, one could particularly observe different associations for both *Vachellia* and *Senegalia* species.

Vachellia karroo correlated with low Ca and Na availability, and high values of Fe, Mn, Cr and Ni. This species is thus indicative of ultramafic and mafic rock types. *Senegalia caffra* correlated with the availability of Ca and is mainly restricted to areas with higher soil Ca:Mg ratios, which are associated with dioritic and felsic rocks. *Vachellia robusta* correlated with the availability of Na and is indicative of felsic and dioritic intrusive rocks specifically containing higher amounts of Na-rich feldspars.

Evaluation of the related species performance index values (Appendix H2) indicated different trends for *Senegalia* and *Vachellia* genera. It was observed that *V. karroo* performed better at high concentrations of Mg, Cr, Co and Ni, while *V. robusta* and *S. caffra* performed better when these elements were less abundant. Thus *V. karroo* was able to outperform the co-occurring species on ultramafic soils and visa versa.

Assessment of specific element concentrations amongst the dominant woody genera, including the shrub species *G. flava*, revealed that each species illustrated different tolerance mechanisms that was observed by differences in their Ca:Mg ratio and trace element concentrations (Cr, Co and Ni).

Vachellia karroo was able to manage both its Mg and Ca content more effectively than the co-occurring vegetation, and generally had a higher Ca:Mg ratio than *G. flava* and *S.caffra*. The species tends to accumulate more calcium in its stems and is believed to suppress the Mg content the most effectively, via leaf litter, since it was able to maintain a low ratio in its leaves. *V. karroo* also had the lowest trace element content compared to the co-analysed species.

Vachellia robusta accumulated more Ca than *V. karroo*, but illustrated similar Ca:Mg ratios. This is because this species did not regulate its Mg as effectively and was more dependent on external sources of Ca. The species therefore did not perform well in the mafic zone and a similar trend could be seen for its trace element concentration, which was elevated in this area.

Grewia flava had a similar tendency than *V. robusta*, but was not able to accumulate Ca as effectively and thus, in the mafic zone, lower ratios was generally observed for this species together with higher amounts of trace elements. This species was far more acclimated to the felsic zone where it maintained lower concentrations of trace elements compared to *V. robusta* and *S. caffra*.

Senegalia caffra was the least effective in managing its Ca:Mg ratio and trace element concentration. The species did not accumulate Ca nor suppressed its Mg content as greatly, which resulted in far lower ratios than the co-occurring genera. Furthermore, individuals situated in the felsic zone contained the highest concentrations of trace elements, and therefore this species was best suited for the dioritic zone where it accumulated less trace elements and maintained a slightly higher Ca:Mg ratio.

Ultimately, based on the presented results, the specific changes in the dominance of *Vachellia* and *Senegalia* genera can now be clearly correlated with variations in their elemental concentration. This observation is hypothesised to be due to their different tolerance and avoidance mechanisms for the various soils and geological substrates of the KC.

CHAPTER 7

CONCLUSIONS AND RECOMMENDATIONS

A petrological and geobotanical study of the KC was undertaken to address a specific shortage of information regarding the Complex's geological features, as well as their related ecological implications for the Vredefort dome's hosted woody vegetation. With regards to the study aims (section 1.4), this chapter will now conclude on the findings by summarising the results of the respective chapters and validating the study hypotheses. Recommendations for further research and closing notes are discussed at the end of this chapter.

7.1 GEOLOGY OF THE KC

From mapping and analysis of the geology, it could be established that the igneous and metamorphic outcrops of the KC form part of three consecutive bodies of layered intrusive sills that had been overturned during the Vredefort impact event. The sills are severely deformed by impact-related folding and thrust-faulting (Figure 9 & Appendix I3), and as a result have, been displaced discordantly with respect to one another.

The oldest sill of mafic – dioritic rock is established to be metamorphosed epidiorite and is proposed to have been related to the 2714 Ma old Ventersdorp lavas. The two younger bodies of ultramafic to dioritic and felsic layered sills have been shown to correspond in mineralogy and stratigraphy with the surrounding complexes of syn-Bushveld age, and are regarded within a similar age constraint (2050 – 2060 Ma).

The un-mapped sills of hornblendite, Opx-Hbl gabbro and quartz alkali syenite, could be established to be related to the known syn-Bushveld age intrusions of the KC (as previously described by Bisschoff, 1969), and are confirmed to collectively form a coherent stratigraphical, mineralogical and geochemical sequence that supports the notion for differentiation from a parental source. To explain their cross-cutting occurrence it is proposed that the two bodies of layered sills differentiated separately and were the result of two successive emplacements of derived magma.

The results provide support for the first hypothesis regarding the geology, and considers the KC to be a differentiated layered intrusive structure, similar to the RTC. In consideration of the amount of faulting and proposed displacement patterns of the intrusive series comprising the KC, the lithodemic unit term "Structural Complex" as suggested by Owen (2009) may be fittingly applied to this particular group of rocks, as it denotes their specific intermixed nature due to deformation.

7.2 SOIL CHEMISTRY OF THE KC

Analysis of soil within the demarcated geobotanical units (Figure 6), revealed that the presence of the intrusive sills amongst the quartzite dominated landscape does have measurable influences on the topsoil chemistry. As a result of the multiple emplacements of the layered bodies and their associated differentiation trends, chemical gradients were identified within the soil profile that correlated with the variance in substrate geology.

Variation in soil chemical characteristics was ultimately attributed to both abiotic and biotic factors. The availability of less mobile or highly adsorbed elements such as Ca, Na, Fe, Ni and Cr were strongly correlated with mineral weathering *in situ*. While the availability of more mobile elements, namely Mg, K, Mn and Co, or those additionally sourced from microbial matter, N and P, were more affected by the topography, CEC, SOC content and vegetation productivity.

It was noted that the intrusions weathered more rapidly than the quartzite country rock, and provided more abundant sources of nutritional elements. Soils originating from geobotanical units that contained intrusive rock types (Units D1, D2 & D3) displayed higher base saturations, less acid to more neutral pH ranges, higher CEC values and macronutrient availabilities, and larger Fe, Mn, Ni and Cr concentrations.

The difference in mineralogy and geochemistry between the groups of intrusive rock types could also be related to the soil chemistry observed within their associated geobotanical units.

The mafic zone (Unit D1) displayed lower availabilities for Ca, Na and N while attaining the largest concentrations for Fe, Cr and Ni. This was related to the presence of ultramafic and mafic rocks, which contributed olivine, serpentine, pyroxene, hornblende and magnetite. While the results do confirm that the ultramafic intrusions contribute towards the occurrence of the “serpentine effect”, the presence of mafic and dioritic intrusions lessened this effect by the addition of feldspar and apatite.

The absence of ultramafic and mafic intrusions within the dioritic zone (Unit D2) was particularly noted by the lower concentrations of Fe, Mn, Cr and Ni. While higher availabilities for Ca, Na and N were attributed to larger amounts of feldspar and hornblende, as well as improved soil conditions and vegetation productivity.

The felsic zone (Unit D3) displayed the largest availability of Ca, Na and N. Higher amounts of Ca and Na directly related to the abundance of feldspar, which is provided by the quartz alkali syenite and alkali granite. Whereas higher N values were attributed to better soil conditions and higher vegetation productivity. The scarcity of ultramafic rocks in Unit D3 was apparent in the

lower concentrations of Fe, Cr and Ni. Although total values for Ni were lower in this unit, the available portion was still elevated due to the presence of uralitite and a higher soil CEC.

7.3 WOODY VEGETATION OF THE KC

Botanical survey results confirmed that the gradients in soil chemistry, caused by the geology of the KC, have significant influences on the floristics, physiognomy and biochemistry of the associated woody vegetation.

Dominant woody species, in particular members of *Senegalia* and *Vachellia*, were strongly correlated with soil chemical elements directly related to the geology. Ultimately, based on floristic variation six unique woody species assemblages could be distinguished for the different groupings of rock substrates (Table 8 & Appendix I3). Since patterns in woody vegetation are associated with changes in dominant rock types, the second hypothesis was supported.

Evaluation of the plant-soil associations and species performance index values supported the third hypothesis. *V. karroo* showed a preference for soil chemical characteristics associated with ultramafic rock types (Table 10) at this locality, while *S. caffra* and *V. robusta* preferred soil characteristics related to the dioritic and felsic intrusions. The ability of *V. karroo* to outcompete *S. caffra* and *V. robusta* on ultramafic intrusions was confirmed by higher species performance index values with increasing Mg, Cr, Co and Ni, and thus this species is accepted as an indicator for ultramafic rocks types at the KC.

The dominance of *V. karroo* on ultramafic substrates was hypothesised to be due to a higher tolerance of Mg, Cr, Co and Ni by means of element exclusion. When the leaf and stem metal content of this species was compared to the co-occurring vegetation (*Grewia flava*, *S. caffra* and *V. robusta*), results were in support of this assumption. All the tested species contained Mg concentrations below the total soil value. However, *V. karroo*, as well as *V. robusta*, were more effective in maintaining lower concentrations of Mg. In addition, on ultramafic substrates *V. karroo* maintained the highest Ca:Mg ratio within its stems and was able to suppress Cr, Co and Ni content below the indicated toxicity value, unlike the co-occurring woody species. The findings are consistent with the conclusions of O'Dell *et al.*, (2006), which indicated that species adapted to serpentine environments maintain a higher Ca:Mg ratio than non-serpentine adapted species.

7.4 STUDY RECOMMENDATIONS AND CLOSING NOTES

The study of the KC served a dual purpose with regards to investigative multidisciplinary research. Firstly, re-evaluation of the geology certainly poses new insight towards our understanding of the Complex and how it is associated with the surrounding intrusive complexes of the Vredefort dome. This information may prove useful for future studies to determine an absolute age for these

intrusions and to describe their petrogenesis. Since only a broad oversight of the geochemistry was presented, it is recommended that more detailed analysis of the KC's trace element chemistry should be performed. Rare earth elements in particular, such as Y, were less affected by alteration and should be fully analysed to provide a complete compositional spectrum for these intrusions. Furthermore, the geology of the areas directly surrounding the mapped portion of the KC should also be re-assessed. These areas might similarly harbor un-documented outcrops that may further aid to determine if the KC and the RTC is or were structurally connected.

The second purpose of this study was to draw attention to soil and plant characteristics that are associated with intrusions of the Vredefort dome. By studying plant relationships with soil and rock chemistry, several woody indicator species could be identified for the KC. These species can serve as visual indicators for specific soil and rock types, and must be further tested to verify if they can indeed be utilised for detection of similar outcrops within the surrounding areas. Since specific patterns in woody species could be identified, it is believed that similar characteristics may also be recognised for the succulent, grass and herbaceous species of the KC, and are recommended for future study. The succulent species *Aloe greatheadii* specifically illustrated a strong association with the mafic intrusions and should be further examined for its metal accumulation characteristics.

With regard to metal accumulation and related soil–plant toxicity, this study specifically focused on Cr, Co and Ni concentrations. These elements are characteristic of the serpentine effect caused by ultramafic intrusions, although high values for zinc (Zn), copper (Cu) and vanadium (V) were also observed. An additional study should be considered for these elements, since their effect on the dominant woody vegetation may further explain the specific differences in substrate association for *Vachellia* and *Senegalia* species.

Lastly, attempts should be made as far as possible to not only preserve the cultural heritage of the KC but also to safeguard its geological significance. Due to the thin structural occurrences of its intrusions and their related weathering rates, much of the Complex has already been eroded over time. Rock types such as the wehrlites and mela-dolerite are, therefore, extremely scarce and removal or disturbance of these outcrops, during field visits, should be avoided.

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APPENDIX A1: GEOLOGICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Lithology	Structure	Comments	Locality		Strike	Dip
Pseudotachylite	Large vein	Pseudotachylite occurs within the Main Bird Quartzites	26°49.412'S	27°23.734'E	N20°E	90°
Alkali Granite	Dyke	Coarse-grained, felsic plutonic rocks	26°49.816'S	27°24.072'E	N120°E	90°
Alkali Granite	Sill		26°49.689'S	27°23.953'E	?	?
Alkali Granite	(Weathered) sill	Highly weathered, coarse-grained saprolitic rocks	26°49.850'S	27°24.050'E	?	0°
Alkali Granite	(W) sill		26°49.639'S	27°23.921'E	?	0°
Alkali Granite	(W) sill		26°49.646'S	27°23.926'E	?	0°
Quartz alkali Syenite	(W) sill	Highly weathered, leucocratic, porphyritic rocks	26°49.715'S	27°23.982'E	?	?
Quartz alkali Syenite	(W) sill		26°49.698'S	27°23.929'E	?	?
Quartz alkali Syenite	(W) sill		26°49.710'S	27°23.926'E	?	?
Spessartite	(W) dyke-rocks	Medium grained, leucocratic, with a equigranular to poikilitic texture	26°49.562'S	27°23.730'E	?	?
Spessartite	(W) dyke-rocks		26°49.619'S	27°23.708'E	?	?
Spessartite	(W) dyke-rocks		26°49.550'S	27°23.775'E	?	?
Spessartite	(W) dyke-rocks		26°49.559'S	27°23.794'E	?	?
Spessartite	(W) dyke-rocks		26°49.585'S	27°23.722'E	?	?
Spessartite	Dyke	Medium grained, melanocratic, varies between equigranular to porphyritic-poikilitic textures	26°49.588'S	27°23.764'E	N70°E	80°
Spessartite / Hornblendite	(W) dyke-rocks		26°49.582'S	27°23.754'E	?	?
Spessartite / Hornblendite	(W) dyke-rocks		26°49.571'S	27°23.738'E	?	?
Spessartite/ Hornblendite	(W) dyke-rocks		26°49.559'S	27°23.805'E	?	?
Spessartite/ Hornblendite	(W) dyke-rocks		26°49.585'S	27°23.741'E	?	?
Gabbro	(W) dyke-rocks		26°49.615'S	27°23.730'E	?	?
Gabbro	(W) dyke-rocks		26°49.612'S	27°23.698'E	?	?
Gabbro	(W) dyke-rocks		26°49.625'S	27°23.723'E	?	?
Gabbro	(W) dyke-rocks		26°49.588'S	27°23.828'E	?	?
Uralite	(W) sill	Thin, medium-grained, highly weathered sill of altered Clinopyroxenite	26°49.734'S	27°23.915'E	?	0°
Uralite	(W) sill		26°49.717'S	27°23.921'E	?	0°
Uralite	(W) sill		26°49.726'S	27°23.917'E	?	10°
Uralite	(W) sill		26°49.720'S	27°23.923'E	?	0°
Wehrlite	Dyke	Northern dyke	26°49.589'S	27°23.769'E	N130°E	30°
Wehrlite	(W) Dyke		26°49.597'S	27°23.765'E	?	?

APPENDIX A1 (CONTINUED): GEOLOGICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Lithology	Structure	Comments	Locality		Strike	Dip
Wehrlite	(W) dyke-rocks	Southern dyke	26°49.588'S	27°23.747'E	?	*?
Wehrlite	(W) dyke-rocks		26°49.603'S	27°23.773'E	?	*?
Wehrlite	Dyke		26°49.616'S	27°23.776'E	N90°E	50°
Wehrlite	Dyke		26°49.616'S	27°23.783'E	N110°E	40°
Wehrlite	Dyke		26°49.610'S	27°23.799'E	N90°E	50°
Wehrlite	(W) dyke-rocks	Very fine grained, massive andesitic dyke,	26°49.611'S	27°23.785'E	?	*?
Epidiorite	W. Dyke		26°49.677'S	27°23.821'E	?	*?
Epidiorite	W. Dyke		26°49.736'S	27°23.900'E	?	*?
Epidiorite	W. Dyke		26°49.688'S	27°23.860'E	?	*?
Epidiorite	Dyke		26°49.425'S	27°23.524'E	N198°E	90°
Epidiorite	(W) dyke-rocks	Weathered dyke-rocks near contact-zones	26°49.625'S	27°23.723'E	?	*?
Spessartite-Epidiorite	(W) contact-rocks		26°49.603'S	27°23.725'E	?	*?
Spessartite-Wehrlite	(W) contact-rocks		26°49.592'S	27°23.758'E	?	*?
Spessartite-Wehrlite	(W) contact-rocks		26°49.589'S	27°23.746'E	?	*?
Spessartite-Gabbro	(W) contact-rocks		26°49.609'S	27°23.727'E	?	*?
Wehrlite-Gabbro	(W) contact-rocks		26°49.604'S	27°23.755'E	?	*?
Wehrlite-Hornblendite	(W) contact-rocks		26°49.589'S	27°23.769'E	?	*?
Amphibole-Hornfels	Rock	Very fine grained, massive rock, sometimes porphyroblastic with small visible grains of Anthophyllite	26°49.646'S	27°23.777'E	?	*?
Amphibole-Hornfels	Rock		26°49.609'S	27°23.727'E	?	*?
Quartzite	Rock-bedding	Fine- very fine-grained, siliceous quartzites, with minor inclusions of hematite, and muscovite	26°49.650'S	27°23.849'E	N150°E	90°
Quartzite	Rock-bedding		26°49.721'S	27°23.844'E	N143°E	90°
Quartzite	Rock-bedding		26°49.650'S	27°23.849'E	N76°E	85°
Quartzite	Rock-bedding		26°49.711'S	27°23.895'E	N140°E	30°
Quartzite	Rock bedding	Fine-very fine-grained, siliceous quartzite, with minor inclusions of dark mica (possibly biotite) and/or magnetite	26°49.548'S	27°23.683'E	N144°E	90°
Quartzite	Rock bedding		26°49.554'S	27°23.674'E	N124°E	75°
Quartzite	Rock bedding		26°49.590'S	27°23.644'E	N104°E	80°
Quartzite	Rock bedding		26°49.592'S	27°23.652'E	N112°E	75°
Quartzite	(W) Rock bedding	Fine-very fine-grained, siliceous quartzites, with minor inclusions of hematite, and Pseudotachylite	26°49.408'S	27°23.717'E	?	?
Quartzite	Rock bedding		26° 49.402'S	27° 23.728'E	N118°E	90°
Quartzite	Rock bedding	Very fine-grained, siliceous quartzites, with visible signs of recrystallized quartz	26°49.797'S	27°24.081'E	N20°E	50°
Quartzite	Rock bedding		26°49.741'S	27°24.170'E	N180°E	80°

APPENDIX A2: ROCK SAMPLES OBTAINED FROM THE KOEDOESFONTEIN COMPLEX, (*ARCHIVE SAMPLES).

Rock type	Rock sample no.	Comments	Locality		Thin section	XRD Analysis	XRF Analysis
Epidiorite	RB16	Main south-eastern outcrop	26°49,679'S	27°23,838'E	X	X	X
	RB17		26°49,736'S	27°23,900'E	X	X	X
Wehrlite	RB7	Northern outcrop of wehrlite, (olivine > augite)	26°49,604'S	27°23,777'E	X	X	X
	RB8		26°49,612'S	27°23,786'E	X	X	X
Olivine-clinopyroxentite	RB9	Southern outcrop of wehrlite / olivine clinopyroxenite, (olivine < augite)	26°49,616'S	27°23,777'E	X	X	X
	RB10		26°49,588'S	27°23,828'E	X	X	X
Clinopyroxentite	RB21	Shock-deformed outcrop on the northern quartzite slope	26° 49.603'S	27° 23.859'E	X	X	X
	RB31	Northern outcrop of clinopyroxenite	26° 49.558'S	27° 23.822'E	X		
Uralite	RB14	Altered ultramafic outcrop on the north-eastern quartzite slope	26°49,717'S	27°23,921'E	X	X	X
Mela-dolerite	RB1A	Very fine-grained magnetite-rich gabbroic rocks	26° 49.625'S	27° 23.723'E	X	X	X
	RB1B				X	X	X
Px Hbl Gabbro	RB2	Medium-grained altered gabbroic rocks	26°49,625'S	27°23,723'E	X		
	RB3		26°49,625'S	27°23,723'E	X	X	X
	RB20		26°49.622'S	27° 23.727'E	X		X
Diopside pegmatite	RB18	Pegmatite schlieren	26°49.577'S	27° 23.772'E	X	X	X
Mela-spessartite	RB6A	Lower plagioclase content, with larger grained oicocrusts of hornblende enclosed by a thin rim of fine-grained albitic feldspar	26°49,588'S	27°23,746'E	X	X	X
Hornblendite	RB6B		26°49,588'S	27°23,746'E	X	X	X
Leuco-spessartite	RB4	Higher plagioclase content, with smaller grained oicocrusts of hornblende enclosed by coarse-grained albitic feldspar	26°49,571'S	27°23,738'E	X	X	X
	RB5		26°49,571'S	27°23,738'E	X	X	X
	AR4* ¹	Archive samples * ¹ Boneschans, (2012) & * ² Bisschoff, (1996)	?	?	X		
	AB14* ²		?	?	X		
Quartz alkali syenite	RB13	Highly weathered sample	26°49,710'S	27°23,926'E	X	X	X
Alkali Granite	RB15	Fresh unweathered sample	26°49,953'S	27°23,689'E	X	X	X
	AB17* ²	Archive sample * ² Bisschoff, (1996)	?	?	X		X

APPENDIX B1: SAMPLING DATA FOR SOIL COMPOSITE SAMPLES (KSA & KKSA). (*1,2 Refer to figure 6, Chapter 3)

Geobotanical unit*1	Sampling plot*2	Soil sample no. (KSA)	Geology	Locality		Composite sample no. (KKSA)
A	KV1	KSA1	Quartzite	26°49'47.0" S	27°23'38.3" E	KKSA1
A	KV9	KSA9	Quartzite	26°49'28.1" S	27°23'40.4" E	KKSA1
A	KV10	KSA10	Quartzite	26°49'51.3" S	27°23'44.4" E	KKSA2
A	KV17	KSA17	Quartzite	26°49'52.0" S	27°23'48.5" E	KKSA2
B	KV2	KSA2	Quartzite	26°49'35.0" S	27°23'48.5" E	KKSA3
B	KV8	KSA8	Quartzite	26°49'49.4" S	27°23'48.8" E	KKSA3
B	KV18	KSA18	Quartzite	26°49'48.2" S	27°23'51.7" E	KKSA4
B	KV19	KSA19	Quartzite	26°49'45.1" S	27°23'49.6" E	KKSA4
C	KV24	KSA24	Quartzite	26°49'42.9" S	27°23'50.4" E	KKSA13
C	KV25	KSA25	Quartzite	26°49'44.3" S	27°23'52.2" E	KKSA13
C	KV26	KSA26	Quartzite	26°49'46.4" S	27°23'53.0" E	KKSA13
C	KV27	KSA27	Quartzite	26°49'41.7" S	27°23'53.5" E	KKSA11
C	KV30	KSA30	Quartzite	26°49'39.1" S	27°23'51.8" E	KKSA11
D1	KV3	KSA3	Wehrlite, gabbro & spessartite	26°49'38.3" S	27°23'40.6" E	KKSA6
D1	KV4	KSA4	Wehrlite, gabbro & spessartite	26°49'32.9" S	27°23'47.3" E	KKSA10
D1	KV5	KSA5	Wehrlite, gabbro & spessartite	26°49'38.1" S	27°23'44.0" E	KKSA6
D1	KV6	KSA6	Wehrlite, gabbro & spessartite	26°49'38.8" S	27°23'44.8" E	KKSA9
D1	KV7	KSA7	Wehrlite, gabbro & spessartite	26°49'35.0" S	27°23'48.6" E	KKSA9
D1	KV11	KSA11	Wehrlite, gabbro & spessartite	26°49'36.7" S	27°23'46.3" E	KKSA8
D2	KV12	KSA12	Epidiorite	26°49'43.5" S	27°23'44.2" E	KKSA5
D2	KV13	KSA13	Epidiorite	26°49'40.3" S	27°23'44.8" E	KKSA5
D2	KV14	KSA14	Epidiorite	26°49'41.0" S	27°23'49.9" E	KKSA12
D2	KV15	KSA15	Epidiorite	26°49'42.2" S	27°23'52.1" E	KKSA12
D2	KV16	KSA16	Epidiorite	26°49'43.5" S	27°23'53.1" E	KKSA14
D2	KV20	KSA20	Epidiorite	26°49'42.3" S	27°23'48.4" E	KKSA9
D2	KV21	KSA21	Epidiorite	26°49'44.3" S	27°23'54.0" E	KKSA14
D2	KV22	KSA22	Epidiorite	26°49'45.8" S	27°23'55.0" E	KKSA15
D2	KV23	KSA23	Epidiorite	26°49'44.6" S	27°23'56.8" E	KKSA15
D3	KV28	KSA28	Alkali granite and quartz.-alkali syenite	26°49'42.2" S	27°23'55.9" E	KKSA16
D3	KV29	KSA29	Alkali granite and quartz.-alkali syenite	26°49'39.0" S	27°23'54.5" E	KKSA16
D3	KV31	KSA31	Alkali granite and quartz.-alkali syenite	26°49'37.9" S	27°23'50.1" E	KKSA17

APPENDIX B2: SAMPLING DATA FOR SOIL COMPOSITE SAMPLES (KSB).

Intrusive Complex	Sampling area (r = 20 m)	Locality		Intrusive rock type(s)	Rock chemical type	Composite sample no. (KSB)
Koedoesfontein Complex	A01	26°49'34.4" S	27°23'47.2" E	Spessartite, gabbro & minor amounts of wehrlite	Mafic-Dioritic	KSB1
	A02	26°49'37.3" S	27°23'46.2" E	Wehrlite, gabbro & minor amounts of spessartite	Ultramafic-Mafic	KSB2
	A03	26°49'42.4" S	27°23'55.9" E	Alkali granite, quartz alkali syenite & fragments of hornblendite(uralite)	Felsic	KSB3
	A04	26°49'43.0" S	27°23'52.4" E	Epidiorite	Dioritic	KSB4

APPENDIX C1: BOTANICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Survey plot no.		KV1	KV2	KV3	KV4	KV5	KV6	KV7
Geobotanical unit		A	B	D1	D1	D1	D1	D1
Locality		26° 49'47.0"S	26° 49'35.0"S	26° 49'38.3"S	26° 49'32.9"S	26° 49'38.1"S	26° 49'38.8"S	26° 49'35.0"S
		27° 23' 38.3"E	27° 23'48.5"E	27° 23'40.6"E	27° 23'47.3"E	27° 23'44.0"E	27° 23'44.8"E	27° 23'48.6"E
Altitude (m)		1394	1421	1430	1418	1437	1453	1452
Diversity (species per plot)		10	17	16	11	15	13	12
<i>Aloe greatheadii</i>	Succulent	0	4	52	76	66	66	74
<i>Asparagus laricinus</i>	Shrub	58	6	14	0	4	0	2
<i>Asparagus suaveolens</i>	Shrub	2	58	40	46	40	20	18
<i>Brachylaena rotundata</i>	Tree	0	0	0	0	0	0	0
<i>Buddleja saligna</i>	Tree	0	0	0	0	0	0	0
<i>Celtis africana</i>	Tree	0	22	6	0	0	0	2
<i>Cestrum laevigatum</i>	Shrub	20	0	0	0	0	0	0
<i>Combretum molle</i>	Tree	0	0	0	0	0	0	0
<i>Diospyros lycioides</i>	Shrub	0	0	2	0	0	0	0
<i>Dombeya rotundifolia</i>	Tree	0	0	0	0	0	0	0
<i>Ehretia rigida</i>	Shrub	0	0	12	6	2	8	14
<i>Euclea crispa</i>	Tree	0	6	2	0	2	0	0
<i>Grewia flavescens</i>	Shrub	0	2	0	6	0	6	2
<i>Grewia occidentalis</i>	Shrub	0	6	6	2	4	0	0
<i>Gymnosporia buxifolia</i>	Tree	0	4	0	0	0	0	0
<i>Gymnosporia buxifolia</i>	Shrub	0	10	0	0	0	0	0
<i>Gymnosporia tenuispina</i>	Shrub	0	0	0	0	0	0	0
<i>Melia azederach</i>	?	0	0	0	0	0	0	0
<i>Mystroxydon aethiopicum</i>	Tree	0	0	0	0	0	0	0
<i>Nuxia congesta</i>	Shrub	0	0	0	0	0	0	0
<i>Olea europaea subsp. africana</i>	Tree	0	0	0	0	0	0	0
<i>Opuntia ficus-indica</i>	Succulent	0	28	10	0	0	0	2
<i>Opuntia humifusa</i>	Succulent	0	50	38	18	26	22	0
<i>Pappia capensis</i>	Tree	0	0	2	0	0	2	0

APPENDIX C1 (CONTINUED): BOTANICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Survey plot no.		KV1	KV2	KV3	KV4	KV5	KV6	KV7
<i>Pavetta zeyheri</i>	Tree	0	0	0	0	2	0	0
<i>Pavetta zeyheri</i>	Shrub	0	0	0	0	0	0	0
<i>Pyracantha angustifolia</i>	Tree	10	0	0	0	0	0	0
<i>Pyracantha crenelata</i>	Tree	4	0	0	0	0	0	0
<i>Saersia leptodictya</i>	Tree	0	2	0	0	0	2	0
<i>Saersia pyroies</i>	Tree	0	6	2	2	0	0	0
<i>Saersia rigida</i>	Shrub	0	0	0	0	0	0	0
<i>Sarcostemma viminalis</i>	Succulent	0	0	0	0	0	0	0
<i>Senegalia caffra</i>	Tree	0	4	0	0	0	0	0
<i>Vachellia karroo</i>	Tree	0	10	42	32	30	12	10
<i>Vachellia robusta</i>	Tree	0	0	0	0	0	2	12
<i>Vangueria parvifolia</i>	Shrub	0	0	0	0	0	0	0
<i>Zanthoxylum capensis</i>	Tree	0	0	0	0	0	0	0
<i>Ziziphus mucronata</i>	Tree	0	22	2	10	10	20	10
zz No Hit	Tree	86	24	44	56	56	62	66
zz No Hit	Shrub	20	18	26	40	50	66	64
zz No Hit	Succulent	100	18	0	6	8	12	24

APPENDIX C1 (CONTINUED): BOTANICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Survey plot no.		KV8	KV9	KV10	KV11	KV12	KV13	KV14
Geobotanical unit		B	A	A	D1	D2	D2	D2
Locality		26° 49'49.4"S	26° 49'28.1"S	26° 49'51.3"S	26° 49'36.7"S	26° 49'43.5"S	26° 49'40.3"S	26° 49'41.0"S
		27° 23'48.8"E	27° 23'40.4"E	27° 23'44.4"E	27° 23'46.3"E	27° 23'44.2"E	27° 23'44.8"E	27° 23'49.9"E
Altitude (m)		1397	1396	1402	1446	1407	1452	1494
Diversity (species per plot)		17	8	7	12	19	18	13
<i>Aloe greatheadii</i>	Succulent	0	0	0	52	32	54	52
<i>Asparagus laricinus</i>	Shrub	8	46	46	0	4	0	0
<i>Asparagus suaveolens</i>	Shrub	32	2	2	16	20	22	8
<i>Brachylaena rotundata</i>	Tree	0	0	0	0	0	0	0
<i>Buddleja saligna</i>	Tree	0	0	0	0	0	0	0
<i>Celtis africana</i>	Tree	10	0	0	0	0	2	0
<i>Cestrum laevigatum</i>	Shrub	0	0	0	0	0	0	0
<i>Combretum molle</i>	Tree	0	0	0	0	0	0	0
<i>Diospyros lycioides</i>	Shrub	28	0	0	0	0	0	2
<i>Dombeya rotundifolia</i>	Tree	0	0	0	0	6	4	4
<i>Ehretia rigida</i>	Shrub	2	0	0	6	26	6	0
<i>Euclea crispa</i>	Tree	2	0	0	0	2	4	0
<i>Grewia flavescens</i>	Shrub	0	0	0	2	0	2	2
<i>Grewia occidentalis</i>	Shrub	4	0	0	0	16	8	2
<i>Gymnosporia buxifolia</i>	Tree	2	0	0	0	0	0	0
<i>Gymnosporia buxifolia</i>	Shrub	2	0	0	0	0	0	0
<i>Gymnosporia tenuispina</i>	Shrub	0	0	0	42	0	2	0
<i>Melia azederach</i>	?	0	0	0	0	0	0	0
<i>Mystroxydon aethiopicum</i>	Tree	8	0	0	0	0	0	0
<i>Nuxia congesta</i>	Shrub	0	0	0	0	0	0	0
<i>Olea europaea subsp. africana</i>	Tree	0	0	0	0	0	0	0
<i>Opuntia ficus-indica</i>	Succulent	24	0	0	0	4	2	0
<i>Opuntia humifusa</i>	Succulent	18	0	0	2	62	20	10
<i>Pappea capensis</i>	Tree	0	0	0	2	4	6	0

APPENDIX C1 (CONTINUED): BOTANICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Survey plot no.		KV8	KV9	KV10	KV11	KV12	KV13	KV14
<i>Pavetta zeyheri</i>	Tree	0	0	0	0	0	0	0
<i>Pavetta zeyheri</i>	Shrub	0	0	0	2	0	0	0
<i>Pyracantha angstifolia</i>	Tree	0	2	0	0	0	0	0
<i>Pyracantha crenelata</i>	Tree	0	4	2	0	0	0	0
<i>Saersia leptodictya</i>	Tree	0	0	0	0	0	0	0
<i>Saersia pyroies</i>	Tree	0	0	0	0	0	2	0
<i>Saersia rigida</i>	Shrub	0	0	0	0	2	0	0
<i>Sarcostemma viminale</i>	Succulent	0	0	0	0	0	0	0
<i>Senegalia caffra</i>	Tree	4	0	0	0	8	20	18
<i>Vachellia karroo</i>	Tree	0	0	0	8	4	0	6
<i>Vachellia robusta</i>	Tree	14	0	0	0	4	2	0
<i>Vangueria parvifolia</i>	Shrub	0	0	0	0	0	0	0
<i>Zanthoxylum capensis</i>	Tree	0	0	0	0	0	0	0
<i>Ziziphus mucronata</i>	Tree	14	2	4	0	12	4	0
zz No Hit	Tree	46	92	94	90	58	56	72
zz No Hit	Shrub	24	52	52	32	34	60	86
zz No Hit	Succulent	58	100	100	46	2	24	38

APPENDIX C1 (CONTINUED): BOTANICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Survey plot no.		KV15	KV16	KV17	KV18	KV19	KV20	KV21
Geobotanical unit		D2	D2	A	B	B	D2	D2
Locality		26° 49'42.2"S	26° 49'43.5"S	26° 49'52.0"S	26° 49'48.2"S	26° 49'45.1"S	26° 49'42.3"S	26° 49'44.3"S
		27° 23'52.1"E	27° 23'53.1"E	27° 23'48.5"E	27° 23'51.7"E	27° 23'49.6"E	27° 23'48.4"E	27° 23'54.0"E
Altitude (m)		1498	1488	1399	1444	1464	1484	1491
Diversity (species per plot)		13	18	8	15	15	15	18
<i>Aloe greatheadii</i>	Succulent	44	10	0	0	18	26	44
<i>Asparagus laricinus</i>	Shrub	0	0	28	4	0	0	0
<i>Asparagus suaveolens</i>	Shrub	2	12	12	20	2	8	32
<i>Brachylaena rotundata</i>	Tree	0	0	0	0	0	0	0
<i>Buddleja saligna</i>	Tree	0	0	0	12	0	0	0
<i>Celtis africana</i>	Tree	0	2	0	4	4	0	0
<i>Cestrum laevigatum</i>	Shrub	0	0	0	0	0	0	0
<i>Combretum molle</i>	Tree	0	0	0	0	0	0	0
<i>Diospyros lycioides</i>	Shrub	0	0	0	2	2	0	0
<i>Dombeya rotundifolia</i>	Tree	4	14	0	6	6	4	8
<i>Ehretia rigida</i>	Shrub	0	2	0	4	4	16	24
<i>Euclea crispa</i>	Tree	0	0	0	0	2	6	4
<i>Grewia flavescens</i>	Shrub	2	0	0	0	0	6	6
<i>Grewia occidentalis</i>	Shrub	2	12	0	10	14	22	10
<i>Gymnosporia buxifolia</i>	Tree	0	2	0	0	0	0	0
<i>Gymnosporia buxifolia</i>	Shrub	0	4	28	0	2	0	0
<i>Gymnosporia tenuispina</i>	Shrub	8	0	0	0	0	0	0
<i>Melia azederach</i>	?	0	0	0	0	0	0	0
<i>Mystroxydon aethiopicum</i>	Tree	0	0	0	0	0	0	0
<i>Nuxia congesta</i>	Shrub	0	0	0	0	6	2	0
<i>Olea europaea subsp. africana</i>	Tree	0	0	0	0	0	0	0
<i>Opuntia ficus-indica</i>	Succulent	2	6	0	30	16	2	0
<i>Opuntia humifusa</i>	Succulent	14	18	0	8	20	12	16
<i>Pappea capensis</i>	Tree	0	2	0	0	0	14	8

APPENDIX C1 (CONTINUED): BOTANICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Survey plot no.		KV15	KV16	KV17	KV18	KV19	KV20	KV21
<i>Pavetta zeyheri</i>	Tree	0	0	0	0	0	0	0
<i>Pavetta zeyheri</i>	Shrub	0	0	0	0	0	0	0
<i>Pyracantha angustifolia</i>	Tree	0	0	2	0	0	0	0
<i>Pyracantha crenelata</i>	Tree	0	0	0	0	0	0	0
<i>Saersia leptodictya</i>	Tree	10	24	0	0	14	58	18
<i>Saersia pyroies</i>	Tree	6	14	0	2	0	4	2
<i>Saersia rigida</i>	Shrub	0	0	0	0	0	0	0
<i>Sarcostemma viminalis</i>	Succulent	0	0	0	8	0	0	0
<i>Senegalia caffra</i>	Tree	32	2	0	12	12	4	8
<i>Vachellia karroo</i>	Tree	0	2	0	0	0	22	26
<i>Vachellia robusta</i>	Tree	0	0	0	0	0	0	0
<i>Vangueria parvifolia</i>	Shrub	0	0	0	0	0	0	2
<i>Zanthoxylum capensis</i>	Tree	0	2	0	0	0	4	0
<i>Ziziphus mucronata</i>	Tree	2	4	6	0	0	0	10
zz No Hit	Tree	56	56	92	64	76	42	34
zz No Hit	Shrub	86	70	32	52	70	46	26
zz No Hit	Succulent	30	42	100	62	32	2	22

APPENDIX C1 (CONTINUED): BOTANICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Survey plot no.		KV22	KV23	KV24	KV25	KV26	KV27	KV28
Geobotanical unit		D2	D2	C	C	C	C	D3
Locality		26° 49'45.8"S	26° 49'44.6"S	26° 49'42.9"S	26° 49'44.3"S	26° 49'46.4"S	26° 49'41.7"S	26° 49'42.2"S
		27° 23'55.0"E	27° 23'56.8"E	27° 23'50.4"E	27° 23'52.2"E	27° 23'53.0"E	27° 23'53.5"E	27° 23'55.9"E
Altitude (m)		1455	1442	1478	1508	1467	1499	1457
Diversity (species per plot)		16	13	11	8	8	11	18
<i>Aloe greatheadii</i>	Succulent	48	64	14	26	30	14	44
<i>Asparagus laricinus</i>	Shrub	0	2	0	0	0	0	0
<i>Asparagus suaveolens</i>	Shrub	34	24	4	0	0	4	52
<i>Brachylaena rotundata</i>	Tree	0	0	0	0	0	0	0
<i>Buddleja saligna</i>	Tree	0	0	0	0	0	0	0
<i>Celtis africana</i>	Tree	2	0	0	0	0	0	0
<i>Cestrum laevigatum</i>	Shrub	0	0	0	0	0	0	0
<i>Combretum molle</i>	Tree	4	0	24	18	8	20	0
<i>Diospyros lycioides</i>	Shrub	0	0	0	2	8	0	0
<i>Dombeya rotundifolia</i>	Tree	0	0	4	4	2	0	2
<i>Ehretia rigida</i>	Shrub	20	0	0	0	0	0	16
<i>Euclea crispa</i>	Tree	6	2	0	0	0	0	2
<i>Grewia flavescens</i>	Shrub	18	4	0	0	0	0	10
<i>Grewia occidentalis</i>	Shrub	10	0	0	0	0	0	2
<i>Gymnosporia buxifolia</i>	Tree	0	0	0	0	0	0	0
<i>Gymnosporia buxifolia</i>	Shrub	2	2	0	0	0	0	0
<i>Gymnosporia tenuispina</i>	Shrub	0	0	0	0	0	0	0
<i>Melia azederach</i>	?	0	0	0	0	0	0	0
<i>Mystroxylon aethiopicum</i>	Tree	0	0	0	0	0	0	0
<i>Nuxia congesta</i>	Shrub	2	0	4	6	8	4	0
<i>Olea europaea subsp. africana</i>	Tree	0	0	0	0	0	2	0
<i>Opuntia ficus-indica</i>	Succulent	0	6	0	0	0	0	2
<i>Opuntia humifusa</i>	Succulent	22	22	0	0	0	0	20
<i>Pappea capensis</i>	Tree	4	2	0	0	0	0	6

APPENDIX C1 (CONTINUED): BOTANICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Survey plot no.		KV22	KV23	KV24	KV25	KV26	KV27	KV28
<i>Pavetta zeyheri</i>	Tree	2	2	0	0	0	0	0
<i>Pavetta zeyheri</i>	Shrub	0	4	18	16	10	12	0
<i>Pyracantha angustifolia</i>	Tree	0	0	0	0	0	0	0
<i>Pyracantha crenelata</i>	Tree	0	0	0	0	0	0	0
<i>Saersia leptodictya</i>	Tree	4	0	52	18	0	66	10
<i>Saersia pyroies</i>	Tree	2	2	6	0	0	0	0
<i>Saersia rigida</i>	Shrub	0	2	0	0	0	0	2
<i>Sarcostemma viminalis</i>	Succulent	2	0	0	0	0	0	0
<i>Senegalia caffra</i>	Tree	24	12	8	8	22	0	2
<i>Vachellia karroo</i>	Tree	0	0	0	0	0	0	38
<i>Vachellia robusta</i>	Tree	8	26	0	0	0	0	4
<i>Vangueria parvifolia</i>	Shrub	0	0	2	2	0	28	0
<i>Zanthoxylum capensis</i>	Tree	0	0	0	0	0	0	0
<i>Ziziphus mucronata</i>	Tree	2	4	0	0	0	0	0
zz No Hit	Tree	46	48	58	70	68	78	44
zz No Hit	Shrub	12	64	72	74	74	52	20
zz No Hit	Succulent	26	8	34	56	70	20	24

APPENDIX C1 (CONTINUED): BOTANICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Survey plot no.		KV29	KV30	KV31
Geobotanical unit		D3	C	D3
Locality		26° 49'39.0"S	26° 49'39.1"S	26° 49'37.9"S
		27° 23'54.5"E	27° 23'51.8"E	27° 23'50.1"E
Altitude (m)		1474	1479	1471
Diversity (species per plot)		13	12	8
<i>Aloe greatheadii</i>	Succulent	32	10	8
<i>Asparagus laricinus</i>	Shrub	6	0	0
<i>Asparagus suaveolens</i>	Shrub	36	0	6
<i>Brachylaena rotundata</i>	Tree	0	2	0
<i>Buddleja saligna</i>	Tree	0	0	0
<i>Celtis africana</i>	Tree	2	0	0
<i>Cestrum laevigatum</i>	Shrub	0	0	0
<i>Combretum molle</i>	Tree	0	38	2
<i>Diospyros lycioides</i>	Shrub	0	2	0
<i>Dombeya rotundifolia</i>	Tree	0	0	0
<i>Ehretia rigida</i>	Shrub	10	0	32
<i>Euclea crispa</i>	Tree	0	0	0
<i>Grewia flavescens</i>	Shrub	8	0	8
<i>Grewia occidentalis</i>	Shrub	0	0	0
<i>Gymnosporia buxifolia</i>	Tree	0	0	0
<i>Gymnosporia buxifolia</i>	Shrub	0	2	0
<i>Gymnosporia tenuispina</i>	Shrub	0	0	0
<i>Melia azederach</i>	?	0	0	0
<i>Mystroxylon aethiopicum</i>	Tree	0	0	0
<i>Nuxia congesta</i>	Shrub	0	4	0
<i>Olea europaea subsp. africana</i>	Tree	0	0	0
<i>Opuntia ficus-indica</i>	Succulent	0	2	0
<i>Opuntia humifusa</i>	Succulent	6	0	6
<i>Pappea capensis</i>	Tree	12	0	6

APPENDIX C1 (CONTINUED): BOTANICAL SURVEY DATA FOR THE KOEDOESFONTEIN COMPLEX.

Survey plot no.		KV29	KV30	KV31
<i>Pavetta zeyheri</i>	Tree	0	0	0
<i>Pavetta zeyheri</i>	Shrub	2	22	2
<i>Pyracantha angustifolia</i>	Tree	0	0	0
<i>Pyracantha crenelata</i>	Tree	0	0	0
<i>Saersia leptodictya</i>	Tree	46	58	84
<i>Saersia pyroies</i>	Tree	2	4	0
<i>Saersia rigida</i>	Shrub	0	0	0
<i>Sarcostemma viminalis</i>	Succulent	0	0	0
<i>Senegalia caffra</i>	Tree	0	2	0
<i>Vachellia karroo</i>	Tree	44	0	30
<i>Vachellia robusta</i>	Tree	0	0	0
<i>Vangueria parvifolia</i>	Shrub	0	8	0
<i>Zanthoxylum capensis</i>	Tree	0	0	0
<i>Ziziphus mucronata</i>	Tree	2	0	0
zz No Hit	Tree	38	54	62
zz No Hit	Shrub	38	62	52
zz No Hit	Succulent	16	30	2

APPENDIX C2: MEDIAN STEM CIRCUMFERENCE VALUES (cm) FOR WOODY TREE SPECIES OF THE KOEDOESFONTEIN COMPLEX.

Species	Plot number																
	KV1	KV2	KV3	KV4	KV5	KV6	KV7	KV8	KV9	KV10	KV11	KV12	KV13	KV14	KV15	KV16	KV17
<i>Brachylaena rotundata</i>																	
<i>Buddleja saligna</i>																	
<i>Combretum molle</i>																	
<i>Dombeya rotundifolia</i>												26	21.5	48.5	22	21	
<i>Euclea crispa</i>		15	22		17			35				15	29				
<i>Gymnosporia buxifolia</i>		26.5						60								19	
<i>Mystroxydon aethiopicum</i>								30.5									
<i>Olea europaea subsp. africana</i>																	
<i>Pappea capensis</i>			17			15					68	33.5	18			15	
<i>Pavetta zeyheri</i>					12												
<i>Pyracantha angustifolia</i>	17.5								17								19
<i>Pyracantha crenelata</i>	12.5								12	15							
<i>Rhus leptodictya</i>		33				30							10		35	26.5	
<i>Rus pyroies</i>		32.5	18	22								23					
<i>Senegalia caffra</i>		18.5						64.5				40.5	22	25	37	29	
<i>Vachellia karroo</i>		40	20.5	31.5	31	26.5	27	40				15	14				
<i>Vachellia robusta</i>						64	34				43.5	19.5		130		78	
<i>Zanthoxylum capensis</i>																11	
<i>Ziziphus mucronata</i>		30	26	34	28	30	20	22	13	13		17	28		80	16	11

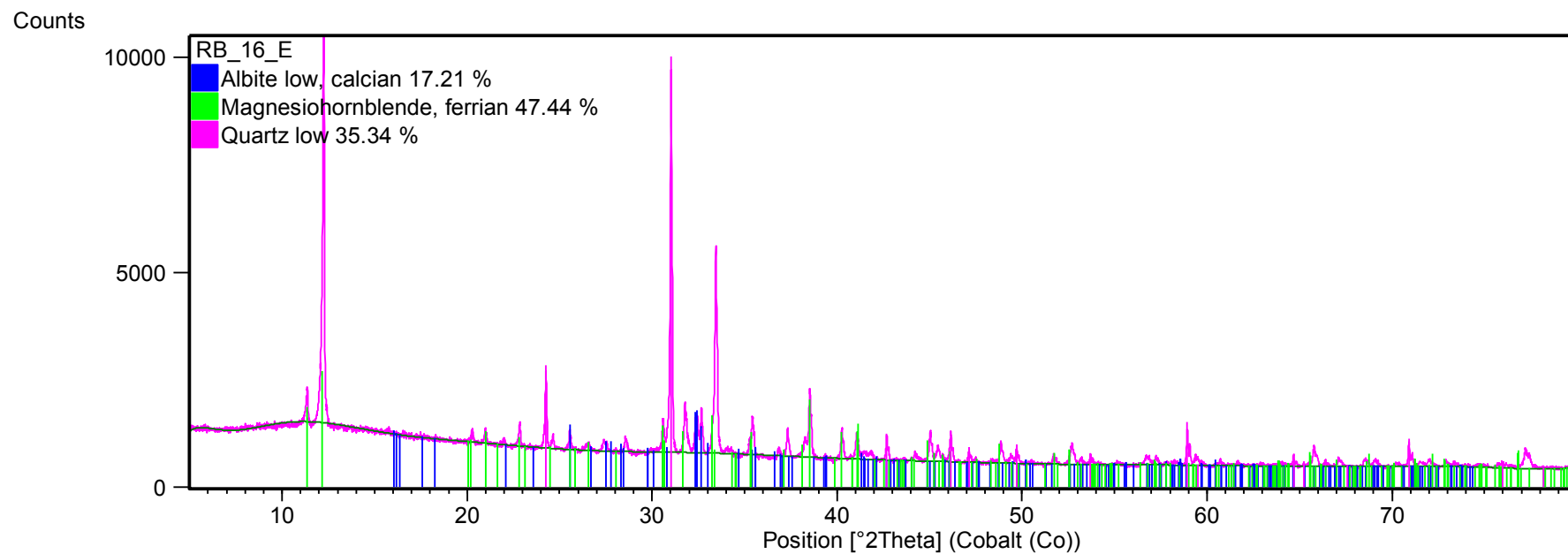
APPENDIX C2 (CONTIUED): MEDIAN STEM CIRCUMFERENCE VALUES (cm) FOR WOODY TREE SPECIES OF THE KOEDOESFONTEIN COMPLEX.

Species	Plot number													
	KV18	KV19	KV20	KV21	KV22	KV23	KV24	KV25	KV26	KV27	KV28	KV29	KV30	KV31
<i>Brachylaena rotundata</i>													13	
<i>Buddleja saligna</i>	30													
<i>Combretum molle</i>					26		41.5	34	12	19			27	28
<i>Dombeya rotundifolia</i>	47	55	20	29.5			35	30	13		10			
<i>Euclea crispa</i>		20	10	13	7	14					14			
<i>Gymnosporia buxifolia</i>														
<i>Mystroxyton aethiopicum</i>														
<i>Olea europaea subsp. africana</i>										60				
<i>Pappea capensis</i>			32.5	37.5	39	62					46	36.5		60.5
<i>Pavetta zeyheri</i>					19	21								
<i>Pyracantha angustifolia</i>														
<i>Pyracantha crenelata</i>														
<i>Rhus leptodictya</i>	70		12.5	34	30	64	35.5					12	52.5	
<i>Rus pyroies</i>						14					8			
<i>Senegalia caffra</i>	56	35	27.5	15.5	31.5	43	29.5	29	19		14		22	
<i>Vachellia karroo</i>					37	30					20			
<i>Vachellia robusta</i>			40.5	43							38	26		41
<i>Zanthoxylum capensis</i>			32.5											
<i>Ziziphus mucronata</i>				27	12	18.5						12		

APPENDIX C3: SAMPLING DATA FOR PLANT TISSUE ANALYSIS.

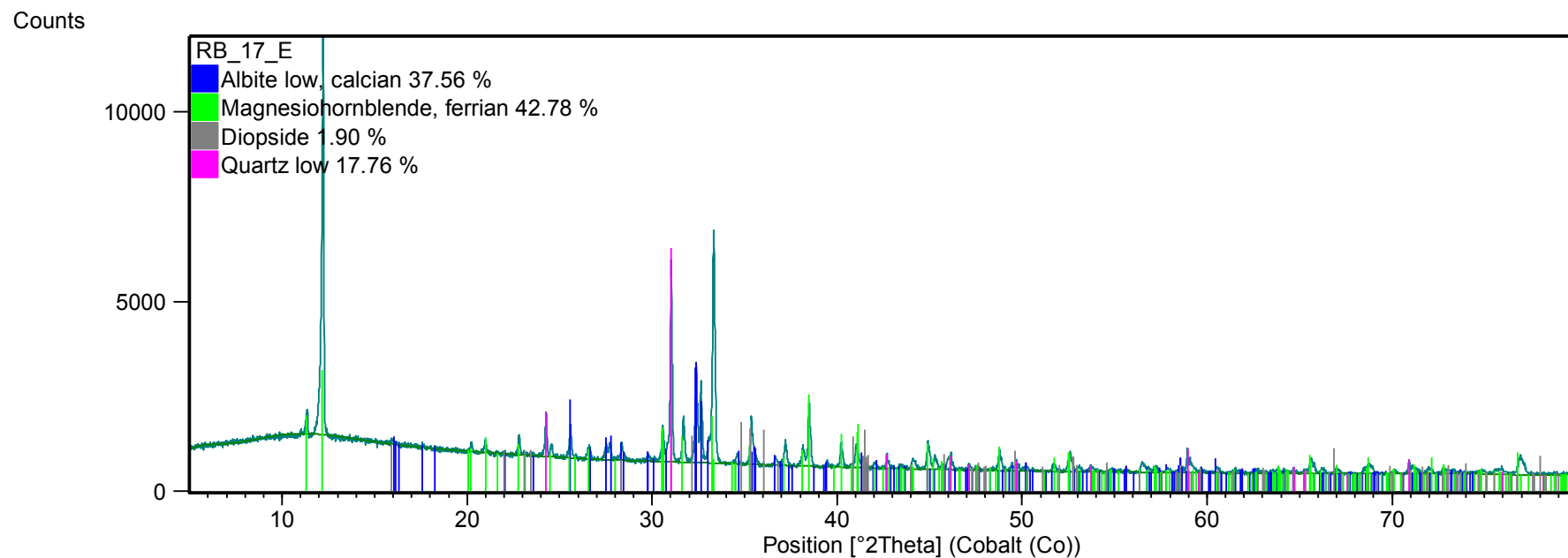
Intrusive Complex	Sampling area (r = 20 m)	Locality		Intrusive rock type(s)	Woody species sampled (Sample No.)							
					<i>S. caffra</i>		<i>V. Karoo</i>		<i>V. Robusta</i>		<i>Grevea Flava</i>	
					Leaf	Stem	Leaf	Stem	Leaf	Stem	Leaf	Stem
Koedoesfontein Complex	A01	26°49'34.4"S	27°23'47.2"E	Spessartite, gabbro & minor amounts of wehrlite/ Ol clinopyroxenite	X	X	01AK LS	01AK SS	01AR LS	01AR SS	01GF LS	01GF SS
	A02	26°49'37.3"S	27°23'46.2"E	Wehrlite / Ol clinopyroxenite, gabbro & minor amounts of spessartite	X	X	02AK LS	02AK SS	02AR LS	02AR SS	01GF LS	02GF SS
	A03	26°49'42.4"S	27°23'55.9"E	Alkali granite, quartz alkali syenite & fragments of hornblendite	03AC LS	03AC SS	03AK LS	03AK SS	03AR LS	03AR SS	03GF LS	03GF SS
	A04	26°49'43.0"S	27°23'52.4"E	Epidiorite	04AC LS	04AC SS	X	X	04AR LS	04AR SS	X	X

APPENDIX D1: XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB16



Peak List	
Albite low, calcian; Al1.26 Ca0.25 Na0.75 O8 Si2.74	
Magnesiohornblende, ferrian; H2 Al1.4 Ca1.8 Fe1.7 K0.1 Mg3.1 Na0.4 O24 Si7	
Quartz low; O2 Si1	

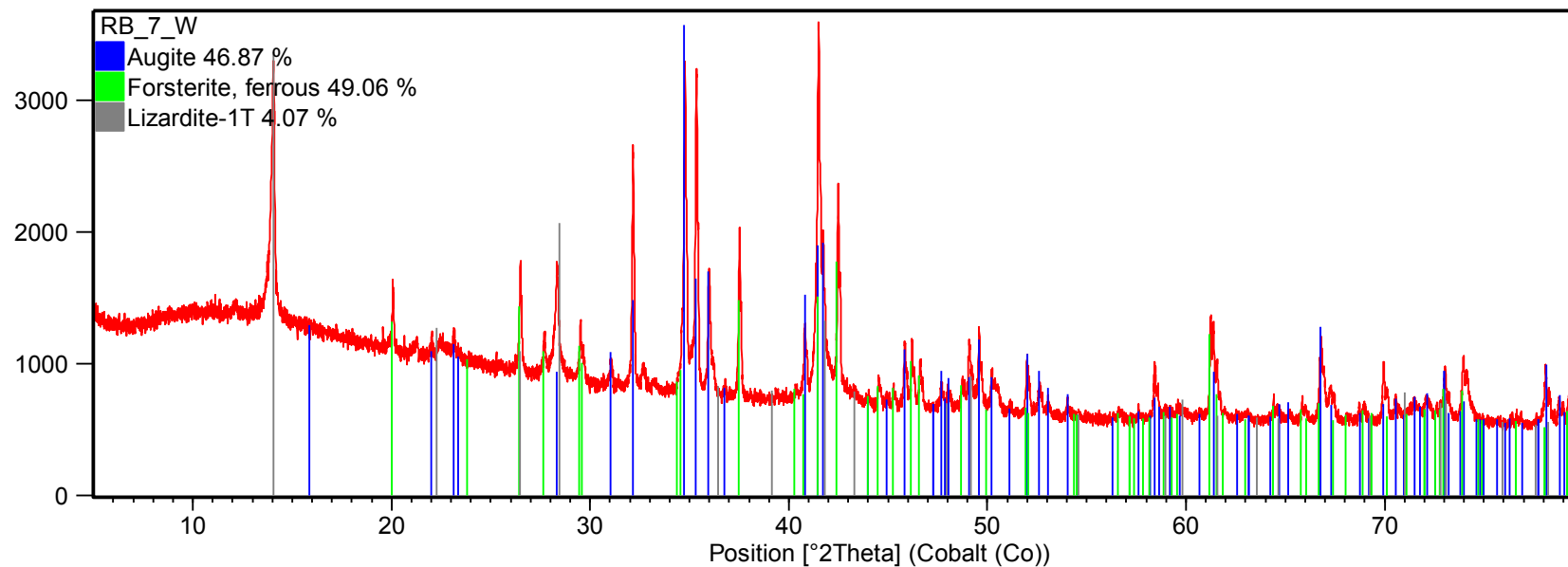
APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCKSAMPLE RB17



Peak List
Albite low, calcian; Al1.26 Ca0.25 Na0.75 O8 Si2.74
Magnesiohornblende, ferrian; H2 Al1.4 Ca1.8 Fe1.7 K0.1 Mg3.1 Na0.4 O24 Si7
Diopside; Ca1 Mg1 O6 Si2
Quartz low; O2 Si1

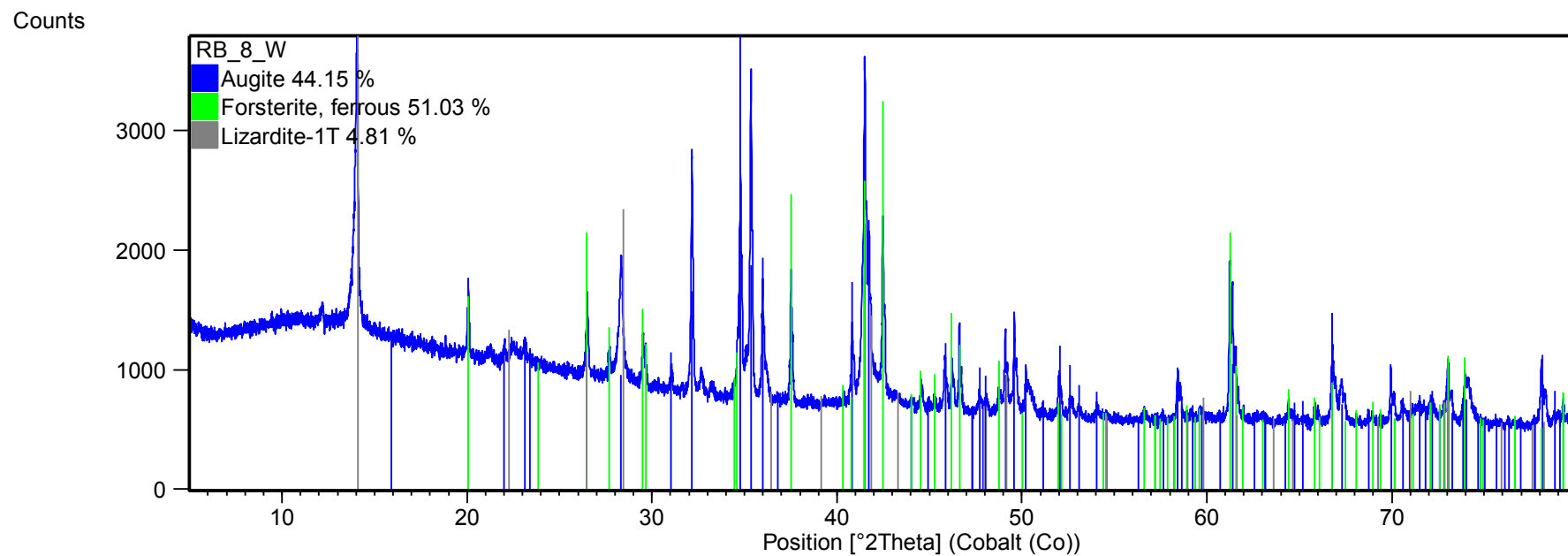
APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB7

Counts



Peak List	
Augite;	(Mg _{0.86} Fe ₁ Al _{0.03} Cr _{0.01}) (Ca _{0.9} Na _{0.01} Mg _{0.06} Fe _{0.03}) (Si _{1.96} Al _{0.04} O ₆)
Forsterite, ferrous;	Mg _{1.624} Fe _{0.376} Si O ₄
Lizardite-1T;	(Mg , Al) ₃ ((Si , Fe) ₂ O ₅) (O H) ₄

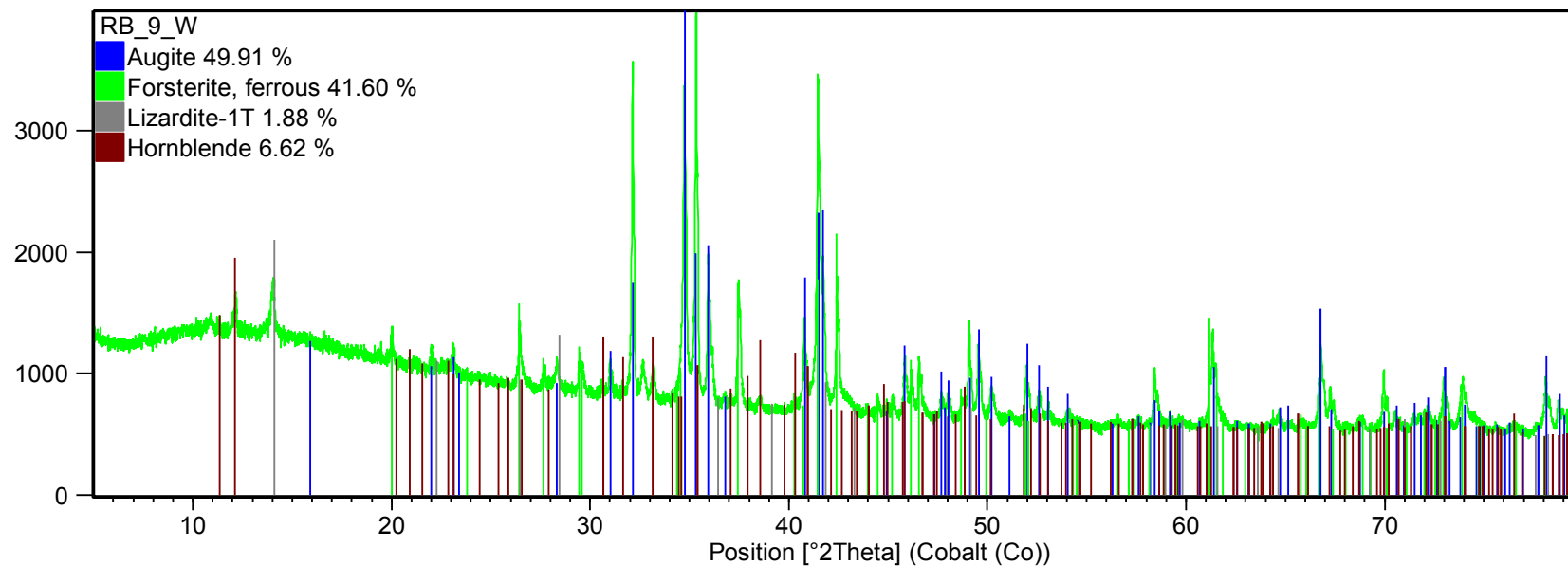
APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB8



Peak List
Augite; (Mg _{0.86} Fe ₁ Al _{0.03} Cr _{0.01}) (Ca _{0.9} Na _{0.01} Mg _{0.06} Fe _{0.03}) (Si _{1.96} Al _{0.04} O ₆)
Forsterite, ferrous; Mg _{1.624} Fe _{0.376} Si O ₄
Lizardite-1T; (Mg , Al) ₃ ((Si , Fe) ₂ O ₅) (O H) ₄

APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB9

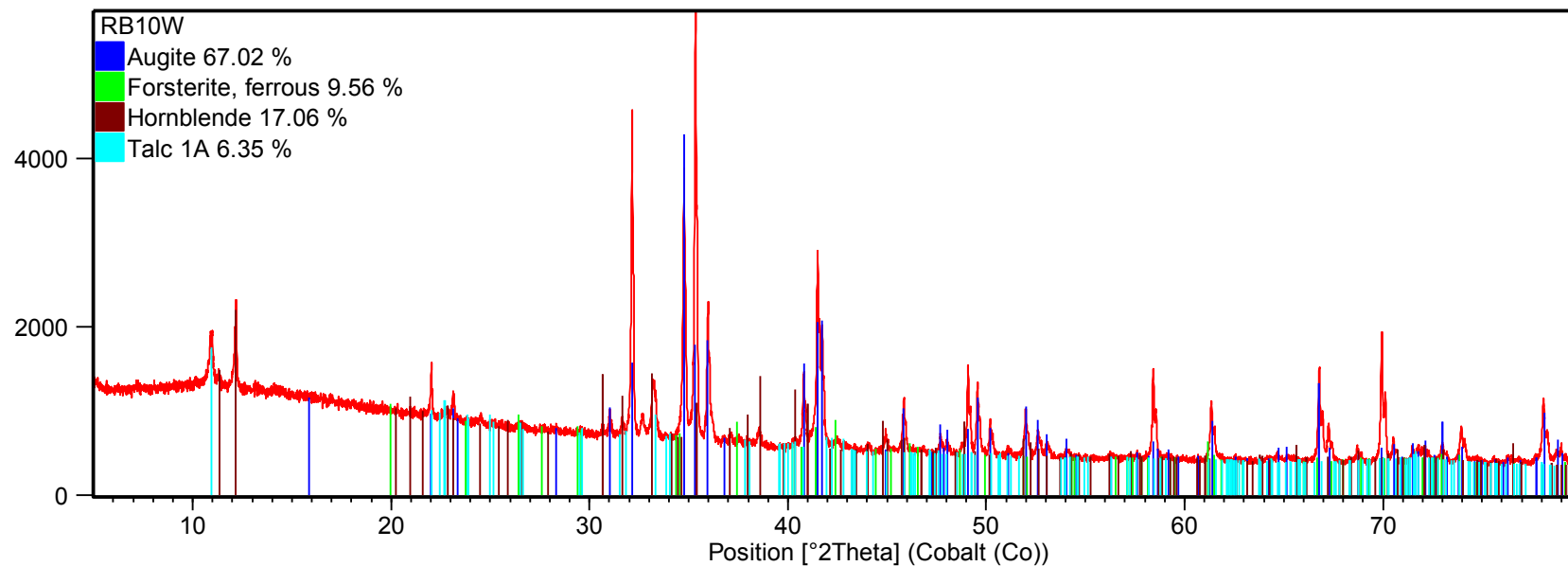
Counts



Peak List	
Augite;	(Mg _{0.86} Fe ₁ Al _{0.03} Cr _{0.01}) (Ca _{0.9} Na _{0.01} Mg _{0.06} Fe _{0.03}) (Si _{1.96} Al _{0.04} O ₆)
Forsterite, ferrous;	Mg _{1.624} Fe _{0.376} Si O ₄
Lizardite-1T;	(Mg , Al) ₃ ((Si , Fe) ₂ O ₅) (O H) ₄
Hornblende;	(K ₃ Na ₆) (Ca _{1.7} Mg ₃) (Mg ₃ Fe Fe ₅ Al ₃ Ti ₂) Al _{1.6} Si _{6.4} O _{22.5} (O H) _{1.5}

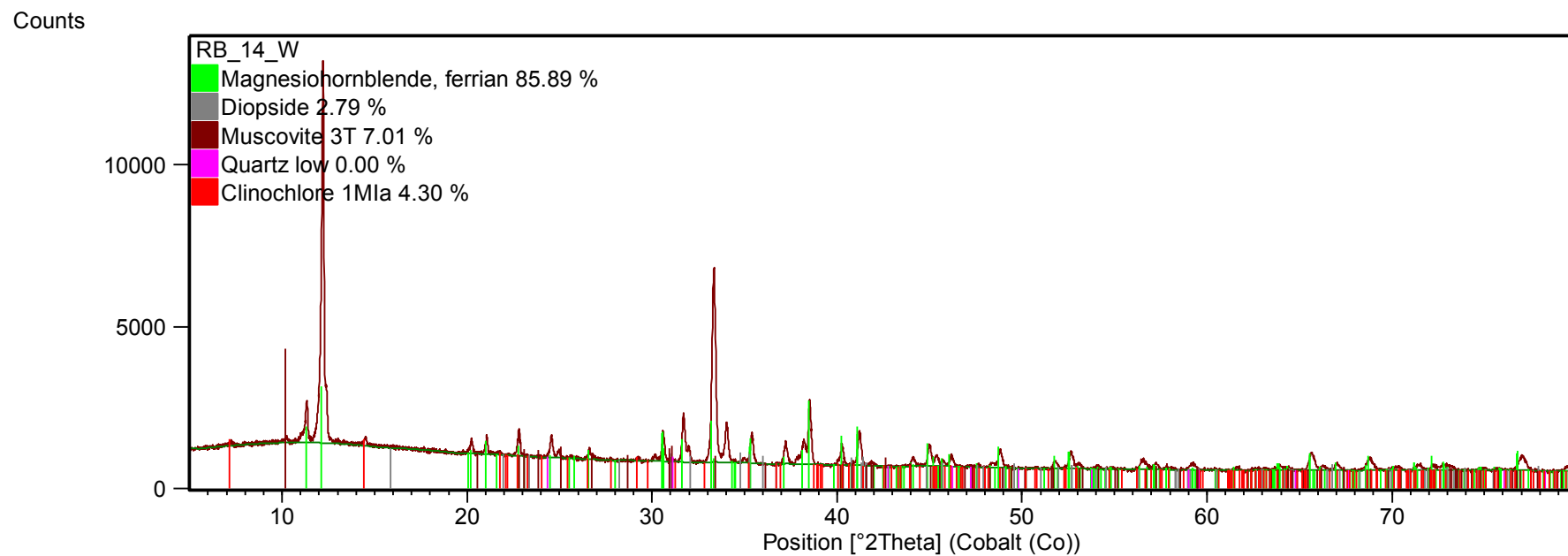
APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB10

Counts



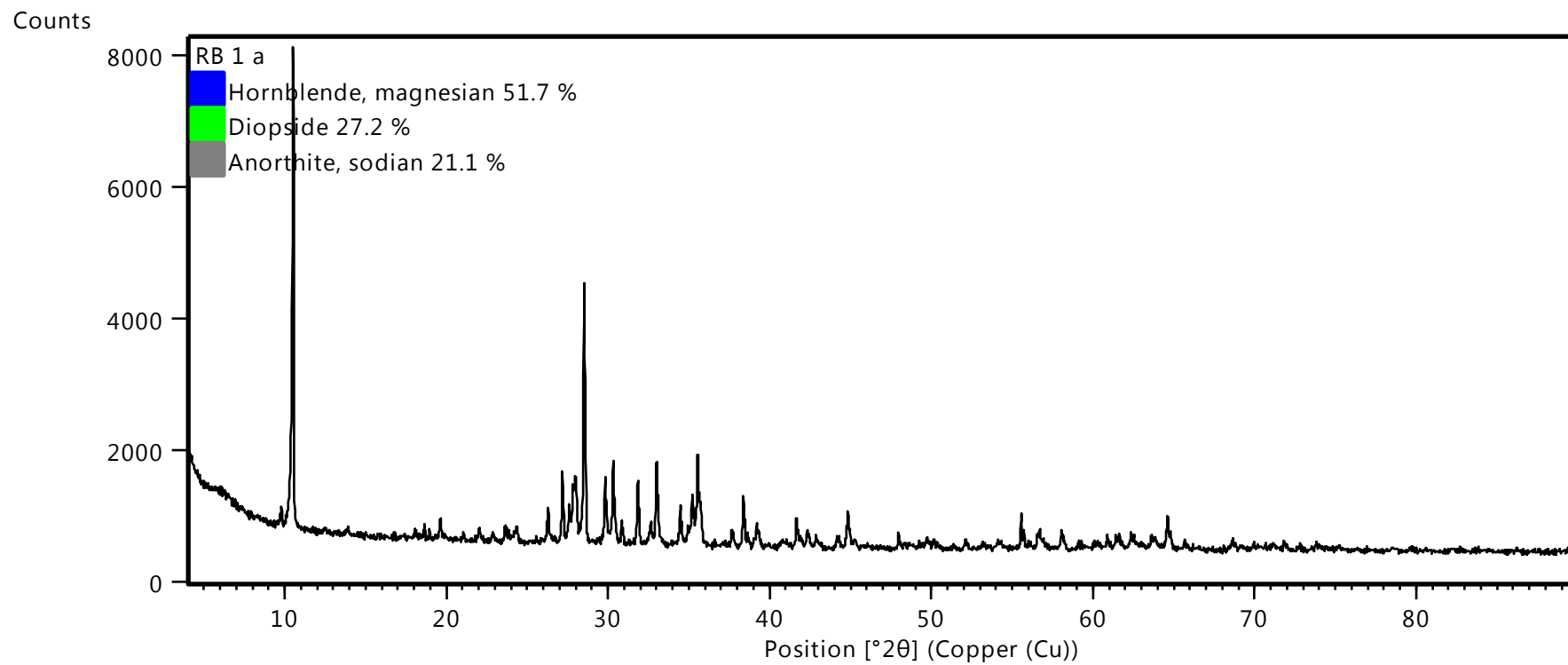
Peak List	
Augite; (Mg0.86 Fe.1 Al0.03 Cr0.01) (Ca0.9 Na0.01 Mg0.06 Fe0.03) (Si1.96 Al0.04 O6)	
Forsterite, ferrous; Mg1.624 Fe.376 Si O4	
Hornblende; (K.3 Na.6) (Ca1.7 Mg.3) (Mg3 Fe Fe.5 Al.3 Ti.2) Al1.6 Si6.4 O22.5 (O H)1.5	
Talc 1A; H2 Mg3 O12 Si4	

APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB14



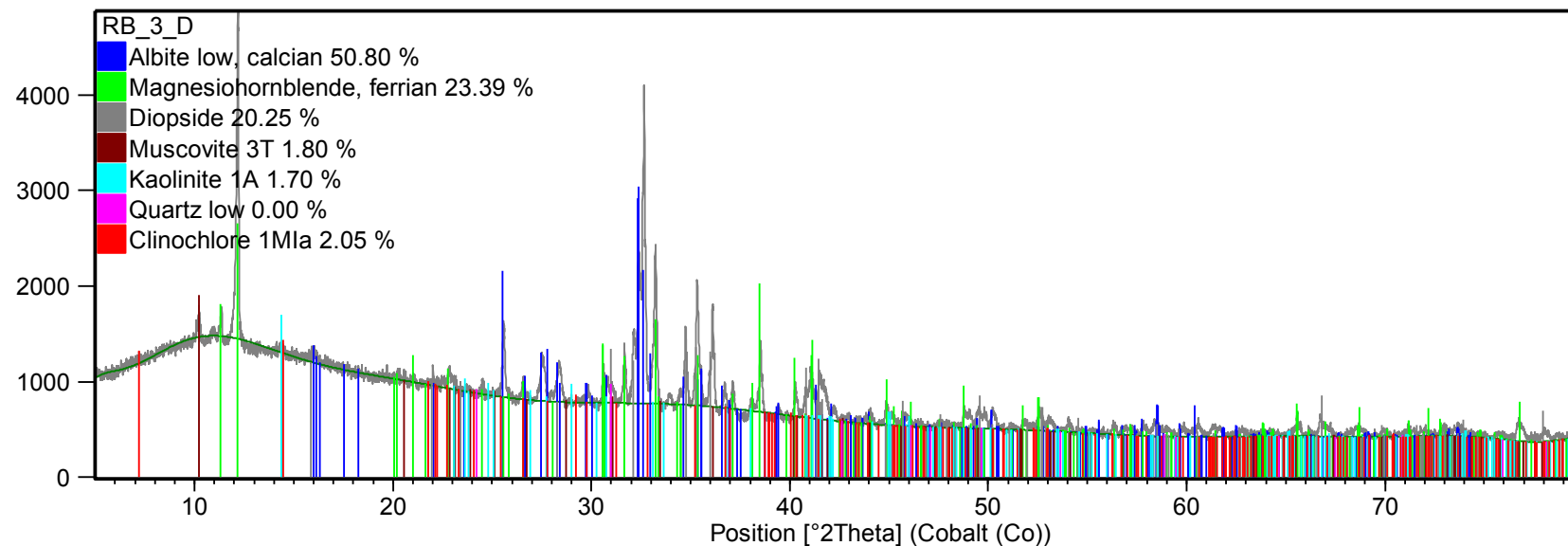
Peak List
Magnesiohornblende, ferrian; H2 Al1.4 Ca1.8 Fe1.7 K0.1 Mg3.1 Na0.4 O24 Si7
Diopside; Ca1 Mg1 O6 Si2
Muscovite 3T; H2 Al3 K1 O12 Si3
Quartz low; O2 Si1
Clinocllore 1Mla; H8 Al3.3 Fe1.65 Mg2.5 O18 Si2.2

APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB1a



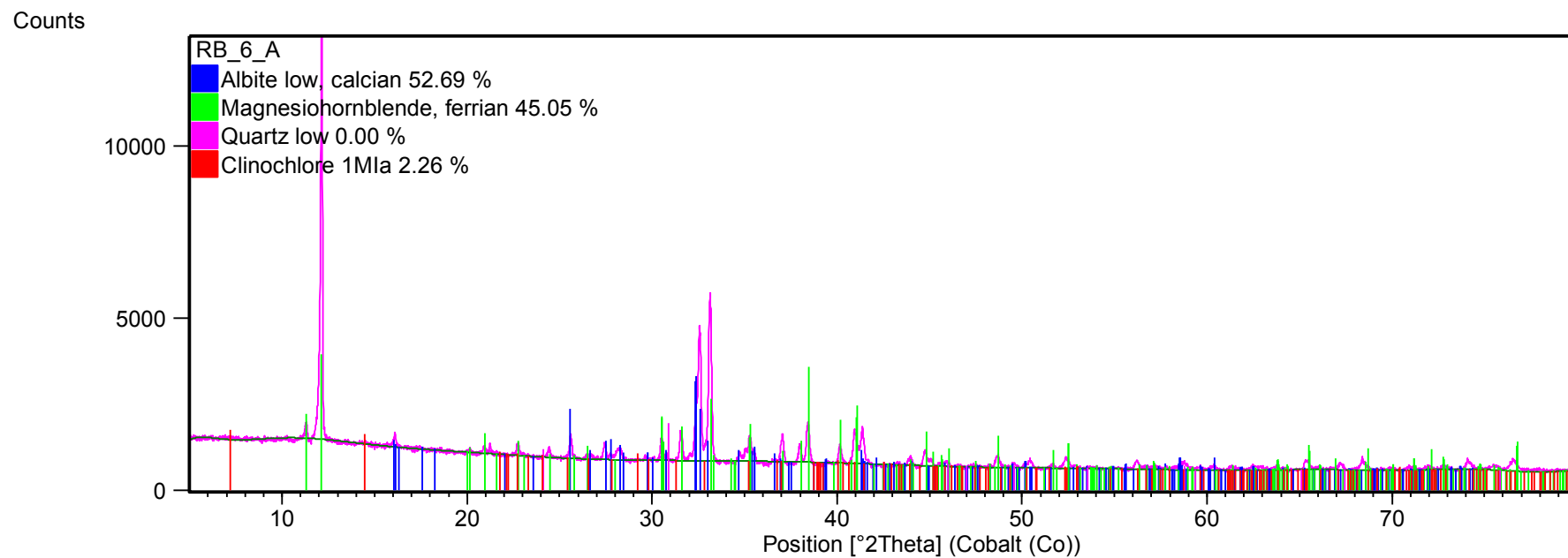
APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB3

Counts



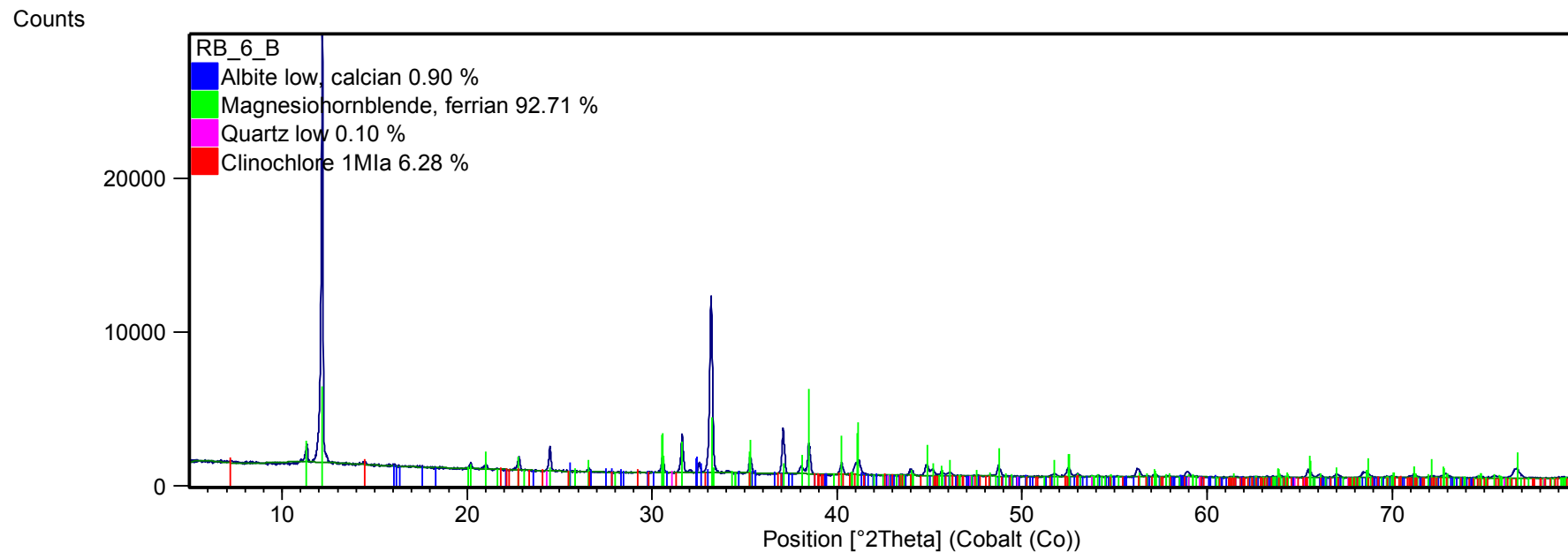
Peak List
Albite low, calcian; Al1.26 Ca0.25 Na0.75 O8 Si2.74
Magnesiohornblende, ferrian; H2 Al1.4 Ca1.8 Fe1.7 K0.1 Mg3.1 Na0.4 Q24 Si7
Diopside; Ca1 Mg1 O6 Si2
Muscovite 3T; H2 Al3 K1 O12 Si3
Kaolinite 1A; H4 Al2 O9 Si2
Quartz low; O2 Si1
Clinocllore 1Mla; H8 Al3.3 Fe1.65 Mg2.5 Q18 Si2.2

APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB6a



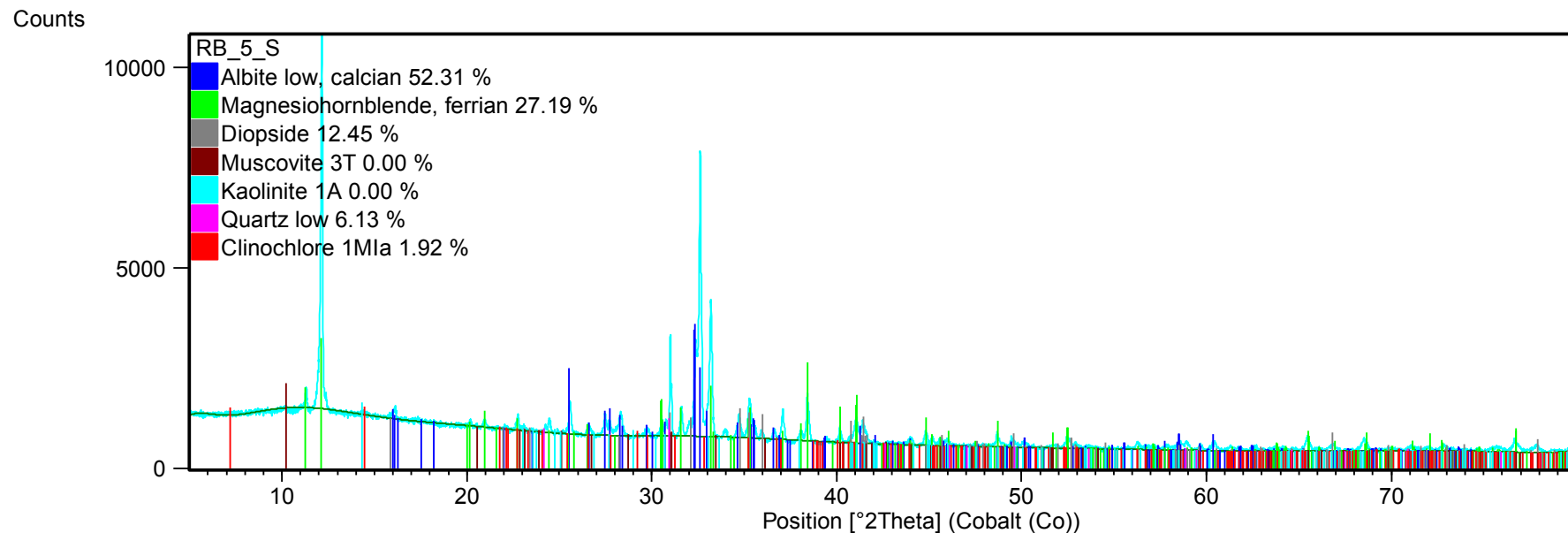
Peak List
Albite low, calcian; Al1.26 Ca0.25 Na0.75 O8 Si2.74
Magnesiohornblende, ferrian; H2 Al1.4 Ca1.8 Fe1.7 K0.1 Mg3.1 Na0.4 O24 Si7
Quartz low; O2 Si1
Clinocllore 1Mla; H8 Al3.3 Fe1.65 Mg2.5 O18 Si2.2

APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB6b



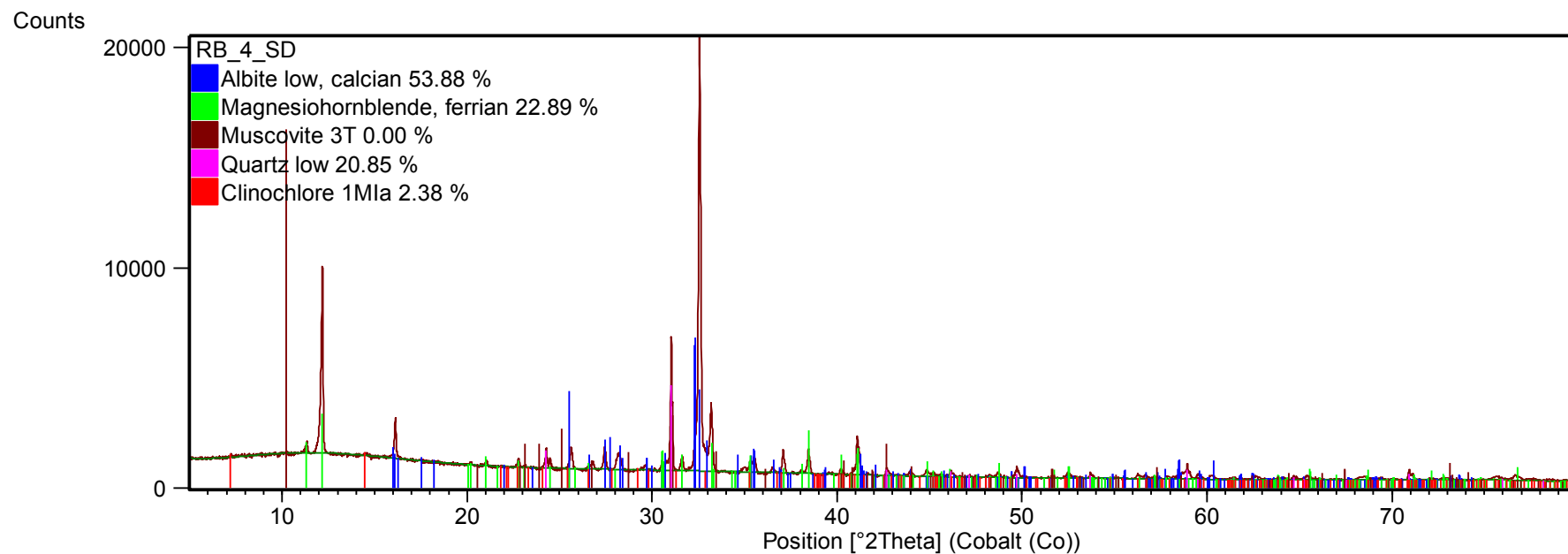
Peak List
Albite low, calcian; Al1.26 Ca0.25 Na0.75 O8 Si2.74
Magnesiohornblende, ferrian; H2 Al1.4 Ca1.8 Fe1.7 K0.1 Mg3.1 Na0.4 O24 Si7
Quartz low; O2 Si1
Clinocllore 1Mla; H8 Al3.3 Fe1.65 Mg2.5 O18 Si2.2

APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB5



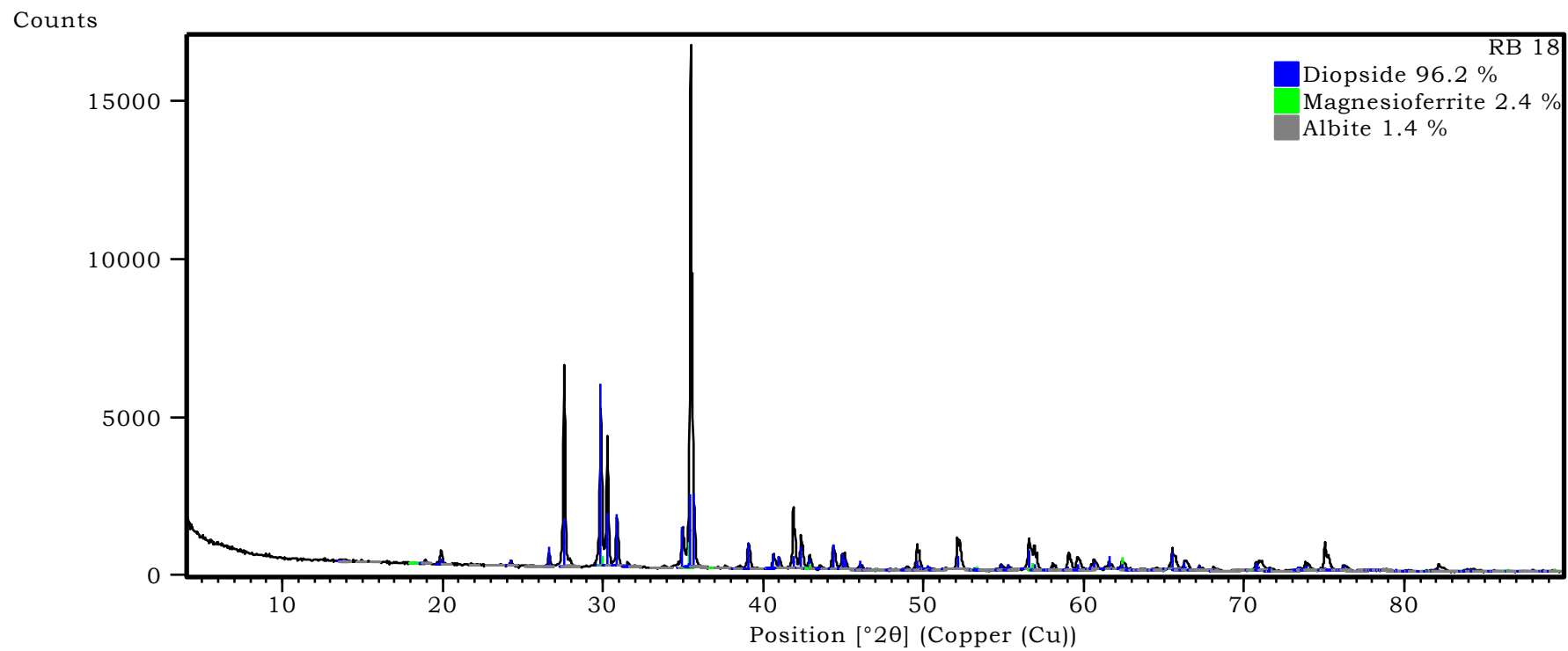
Peak List
Albite low, calcian; Al1.26 Ca0.25 Na0.75 O8 Si2.74
Magnesiohornblende, ferrian; H2 Al1.4 Ca1.8 Fe1.7 K0.1 Mg3.1 Na0.4 Q24 Si7
Diopside; Ca1 Mg1 O6 Si2
Muscovite 3T; H2 Al3 K1 O12 Si3
Kaolinite 1A; H4 Al2 O9 Si2
Quartz low; O2 Si1
Clinocllore 1Mla; H8 Al3.3 Fe1.65 Mg2.5 Q18 Si2.2

APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB4

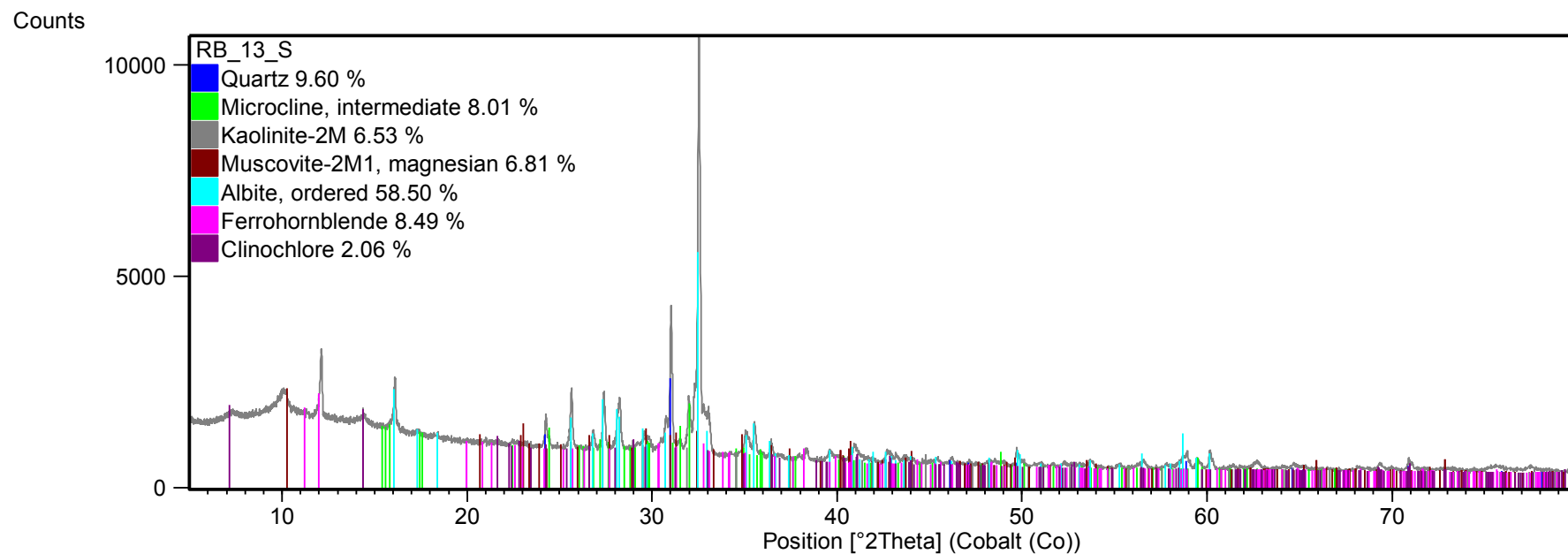


Peak List
Albite low, calcian; Al1.26 Ca0.25 Na0.75 O8 Si2.74
Magnesiohornblende, ferrian; H2 Al1.4 Ca1.8 Fe1.7 K0.1 Mg3.1 Na0.4 O24 Si7
Muscovite 3T; H2 Al3 K1 O12 Si3
Quartz low; O2 Si1
Clinocllore 1M1a; H8 Al3.3 Fe1.65 Mg2.5 O18 Si2.2

APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB18

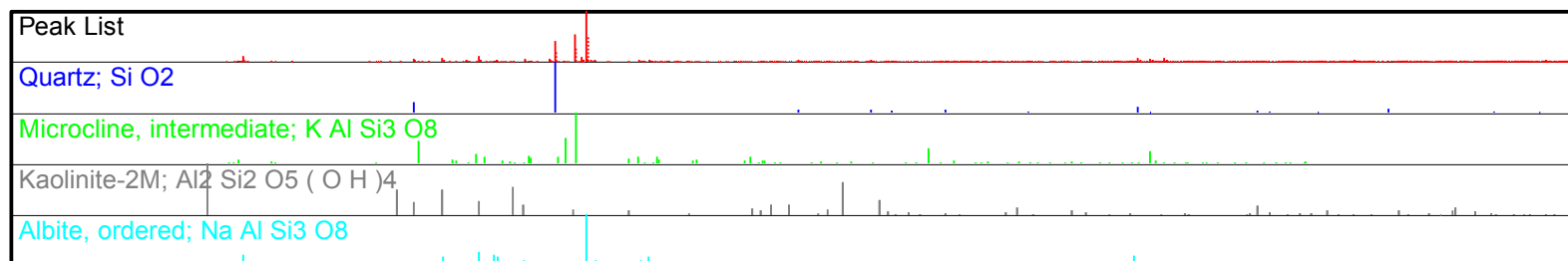
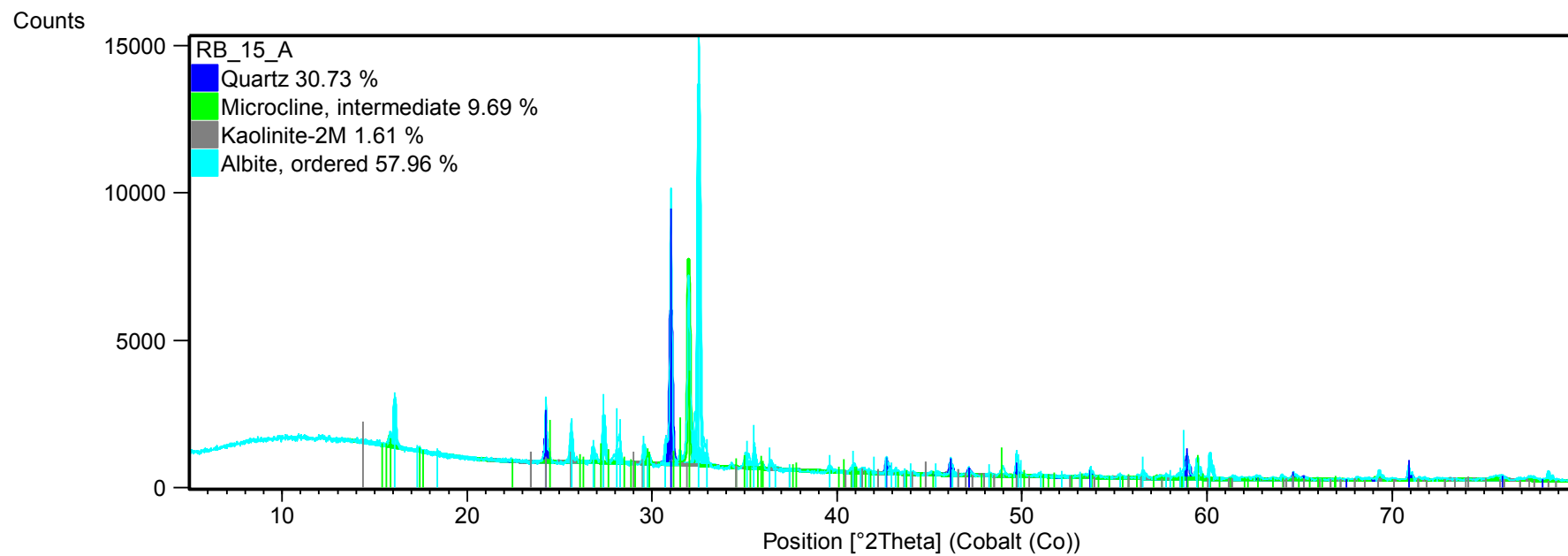


APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB13



Peak List
Quartz; Si O2
Microcline, intermediate; K Al Si3 O8
Kaolinite-2M; Al2 Si2 O5 (O H)4
Muscovite-2M1, magnesian; K0.87 Na0.07 Ca0.02 Mg0.50 Ti0.01 Fe0.14 Al2.04 Si3.39 O10.08 (O H)1.92
Albite, ordered; Na Al Si3 O8
Ferrohornblende; H2 Ca2 Fe5 Na1 O24 Si8
Clinocllore; H16 Al3.01 Fe0.852 Mg11.148 O36 Si4.99

APPENDIX D1 (CONTINUED): XRD-DIFFRACTOGRAM FOR ROCK SAMPLE RB15



APPENDIX D2: QUANTITATIVE SUMMARY OF THE IDENTIFIED MINERAL PHASES (wt. %) IN ANALYSED ROCK SAMPLES FROM THE KOEDOESFONTEIN COMPLEX, DETERMINED WITH XRD ANALYSIS. ANALYST: DR. S. VERYN, XRD ANALYTICAL & CONSULTING CC., LYNNWOOD GLEN, PRETORIA 0081.

Mineral Group	Mineral	Rock sample no. (Weight %)															
		RB16	RB17	RB7	RB8	RB9	RB10	RB14	RB1A	RB3	RB18	RB6A	RB6B	RB5	RB4	RB13	RB15
Silica	Quartz	35,34	17,76										0,10	6,13	20,85	9,60	30,73
K-Feldspar	Microcline															8,01	9,69
	Albite _{An0-10}															58,50	57,96
Plagioclase Feldspar	Oligoclase _{An10-30}	17,21	37,56									52,69	0,90				
	Andesine _{An30-50}										1,40			52,31	53,88		
	Labradorite _{An50-70}								21,10	50,80							
Mica	Muscovite							7,01		1,80						6,81	
	Fe-Hornblende															8,49	
Amphibole	Mg-Hornblende	47,44	42,78					85,89		23,39		45,05	92,71	27,19	22,89		
	Hornblende					6,62	17,06		51,70								
Pyroxene	Diopside		1,90					2,70	27,20	20,25	96,20			12,45			
	Augite			46,87	44,15	49,90	67,02										
Olivine	Forsterite			49,06	51,03	41,60	9,56										
Spinel	Magnesioferrite										2,40						
Serpentine	Lizardite			4,07	4,81	1,88											
Chlorite	Clinochlore							4,30		2,05		2,25	6,28	1,92	2,38	2,06	
Clay minerals	Talc						6,35										
	Kaolinite									1,70						6,53	1,61

APPENDIX D3: CALCULATED MODAL COMPOSITION (Vol. %) OF ANALYSED ROCK SAMPLES, BASED ON XRD DATA (APPENDIX D2).

Mineral Group	Mineral	Rock sample no. (Volume %)															
		RB16	RB17	RB7	RB8	RB9	RB10	RB14	RB1A	RB3	RB18	RB6A	RB6B	RB5	RB4	RB13	RB15
Silica	Quartz	38,48	19,24										0,12	6,59	21,70	9,67	30,41
K-Feldspar	Microcline															8,38	9,96
	Albite _{An0-10}															59,60	58,01
Plagioclase	Oligoclase _{An10-30}	18,74	40,69								1,72		1,07	56,23	56,09		
Feldspar	Andesine _{An30-50}									55,12		57,12					
	Labradorite _{An50-70}								24,10								
Mica	Muscovite							7,87		1,85						6,49	
	Fe-Hornblende															7,08	
Amphibole	Mg-Hornblende	42,78	38,38					84,41		21,02		40,44	91,34	24,20	19,73		
	Hornblende					6,94	17,38		50,00								
Pyroxene	Diopside		1,68					2,61	25,90	17,91	96,56			10,91			
	Augite			47,64	44,83	50,72	66,19										
Olivine	Forsterite			47,01	48,85	39,87	8,90										
Spinel	Magnesioferrite										1,72						
Serpentine	Lizardite			5,35	6,32	2,47											
Chlorite	Clinocllore							5,10		2,22		2,44	7,47	2,06	2,48	2,07	
Clay minerals	Talc						7,53										
	Kaolinite									1,88						6,70	1,62

APPENDIX E1: WHOLE ROCK GEOCHEMICAL COMPOSITION (wt. %) OF ANALYSED ROCK SAMPLES, DETERMINED WITH XRF ANALYSIS.
ANALYST: B.VENTER, XRD & XRF LABORATORY, NWU, POTCHEFSTROOM. (SEE APPENDIX A2 FOR SAMPLE INFORMATION)

Major oxides (wt. %)	Sample no.																			
	RB 16	RB 17	RB1A	RB1B	RB 7	RB 8	RB 9	RB 10	RB21	RB 14	RB18	RB 3	RB20	RB 6A	RB 6 B	RB 5	RB 4	RB 13	AB 14	RB 15
SiO ₂	40,43	50,85	35,47	22,91	40,99	40,80	43,40	47,50	39,36	46,15	47,39	46,66	47,23	40,48	44,77	49,32	56,30	52,81	62,33	63,65
TiO ₂	0,62	0,77	2,73	2,73	0,24	0,26	0,34	0,42	1,91	0,98	0,37	0,51	0,40	2,83	1,75	0,58	0,61	1,26	0,08	0,12
Al ₂ O ₃	10,93	14,28	8,14	1,46	0,76	0,76	1,12	1,29	2,38	6,11	2,74	16,05	10,46	11,34	5,66	14,52	13,69	17,75	17,74	13,95
FeOt	8,35	10,16	20,70	47,11	13,64	13,75	13,21	9,11	18,86	14,54	2,27	7,76	7,64	20,00	14,23	9,20	5,43	6,78	1,40	1,81
Mn ₃ O ₄	0,13	0,16	0,40	0,87	0,23	0,23	0,25	0,20	0,27	0,25	0,40	0,15	0,27	0,27	0,29	0,18	0,14	0,11	0,01	0,02
MgO	4,14	5,37	7,42	8,98	26,25	26,31	22,25	17,27	15,60	14,36	15,65	8,44	13,15	5,08	13,32	5,99	3,92	1,41	0,20	0,14
CaO	7,69	9,01	10,17	2,69	10,00	9,88	12,75	18,05	13,47	9,92	25,25	10,97	11,67	8,48	11,25	10,00	6,29	1,81	0,70	0,43
Na ₂ O	0,43	2,01	2,28	0,02	0,25	0,27	0,39	0,44	0,56	0,13	0,22	1,90	0,41	3,32	1,20	3,33	5,32	4,85	7,56	5,14
K ₂ O	0,03	0,11	0,46	0,14	0,01	0,00	0,02	0,03	0,10	0,24	0,08	1,31	1,05	0,51	0,37	0,79	0,56	2,14	3,10	4,05
P ₂ O ₅	0,07	0,08	0,61	1,29	0,02	0,02	0,02	0,01	0,03	0,13	0,06	0,10	0,05	0,68	0,23	0,08	0,91	1,29	0,03	0,02
LOI*	1,62	2,22	1,26	1,44	3,48	1,64	1,90	1,68	0,35	2,87	0,87	1,90	2,05	2,09	2,88	1,77	1,36	6,54	0,81	0,96
Total	75,38	96,16	91,96	94,90	97,39	95,44	97,12	97,02	94,99	97,30	95,56	96,61	95,23	97,29	97,53	96,77	95,14	97,52	94,12	90,49

* Total Fe is reported as FeOt, LOI = Loss on ignition

Trace elements (ppm)	Sample no.																		
	RB 16	RB 17	RB1A	RB 7	RB 8	RB 9	RB 10	RB21	RB 14	RB18	RB 3	RB20	RB 6A	RB 6 B	RB 5	RB 4	RB 13	AB 14	RB 15
Sc	36	35	49	29	31	41	57	55	26	30	43	38	37	44	36	21	6	2	3
V	186	181	563	94	95	133	174	713	113	31	216	168	388	275	171	83	68	7	6
Cr	222	208	181	1588	1565	1752	1412	711	956	47	590	2008	27	1019	246	123	26	33	33
Co	104	66	78	156	143	137	99	147	93	98	75	96	72	74	81	66	24	58	80
Ni	148	151	87	593	608	462	277	386	781	18	134	386	58	229	140	78	34	12	24
Cu	73	81	571	6	9	15	10	34	107	23	37	8	26	92	39	10	86	16	15
Zn	81	79	154	79	80	84	47	131	124	53	68	52	197	146	69	60	113	17	38

APPENDIX E1 (CONTINUED): WHOLE ROCK GEOCHEMICAL COMPOSITION (wt. %) OF ANALYSED ROCK SAMPLES, DETERMINED WITH XRF ANALYSIS. ANALYST: B. VENTER, XRD & XRF LABORATORY, NWU, POTCHEFSTROOM. (SEE APPENDIX A2 FOR SAMPLE INFORMATION)

Trace elements (ppm)	Sample No.																		
	RB 16	RB 17	RB1A	RB 7	RB 8	RB 9	RB 10	RB21	RB 14	RB18	RB 3	RB20	RB 6A	RB 6 B	RB 5	RB 4	RB 13	AB 14	RB 15
Ga	15	15	20	1,1	bdl	4	3	8	8	5	14	12	21	11	13	14	24	23	20
As	78	64	109	86	85	94	71	108	86	81	72	74	99	91	72	59	57	44	54
Se	22	20	38	25	27	28	24	36	26	22	22	21	34	29	23	16	15	12	12
Br	20	20	37	25	27	26	22	35	26	20	18	21	31	26	20	16	15	10	10
Rb	5	4	12	0,5	bdl	0,8	0,5	2	9	2	37	25	9	5	18	12	50	17	62
Sr	913	477	398	26	27	41	47	59	169	49	215	143	787	111	460	642	387	217	89
Y	16	15	23	2	2	3	5	9	13	4	10	8	16	16	12	14	12	5	13
Zr	76	81	45	3	3	6	6	31	88	72	17	33	45	70	47	58	163	135	221
Nb	3	3	11	bdl	bdl	bdl	bdl	0,9	11	4	1,4	0,5	6	14	2	5	18	6	15
Cd	9	9	8	14	12	11	9	2	9	6	5	9	8	9	13	12	5	6	6
Sn	15	9	11	11	8	8	5	3	10	5	8	5	13	13	15	19	10	11	15
Ba	140	233	338	17	15	54	40	96	478	108	194	92	328	84	388	239	698	741	605
Th	20	14	11	5	4	5	6	3	12	7	6	6	14	8	9	13	10	11	14

*bdl = below detection limit.

APPENDIX E2: CHEMICAL COMPOSITION (wt. %) OF ANALYSED SOIL COMPOSITE SAMPLES (KKSA), DETERMINED WITH XRF ANALYSIS.
ANALYST: B.VENTER, XRD & XRF LABORATORY, NWU, POTCHEFSTROOM.

Element oxides (wt. %)	Soil composite no.																
	KKS 1	KKS 2	KKS 3	KKS 4	KKS 5	KKS 6	KKS 7	KKS 8	KKS 9	KKS 10	KKS 11	KKS 12	KKS 13	KKS 14	KKS 15	KKS 16	KKS 17
O	0,00	0,00	0,00	0,00	8,60	7,84	7,44	8,09	7,93	8,99	5,02	9,35	4,55	9,51	9,74	9,36	8,73
SiO₂	62,60	56,90	38,30	48,10	47,50	48,10	48,70	47,10	43,90	47,60	52,40	48,00	56,80	46,30	47,80	48,20	41,60
TiO₂	1,55	2,76	3,01	0,65	1,45	1,23	0,73	1,45	1,92	1,27	0,53	0,71	0,51	0,72	0,84	0,70	1,24
Al₂O₃	12,90	12,80	15,70	12,50	15,70	14,00	16,80	13,10	12,20	15,00	14,60	17,00	12,50	18,00	17,90	17,40	16,00
Fe₂O₃*	6,93	9,89	18,50	4,57	9,03	8,13	5,13	9,38	12,70	9,85	3,30	7,07	3,06	7,74	6,21	4,93	9,00
MnO	0,10	0,11	0,19	0,15	0,13	0,12	0,09	0,14	0,19	0,15	0,04	0,13	0,06	0,14	0,14	0,13	0,17
MgO	0,65	0,43	0,74	0,51	1,39	2,94	0,74	3,04	3,83	4,05	0,41	0,89	0,35	1,02	0,79	0,97	1,53
CaO	0,51	0,57	3,58	1,62	1,81	2,45	0,82	2,99	4,09	2,62	0,32	1,54	0,30	2,00	0,98	0,47	0,47
Na₂O	0,98	0,16	0,12	0,22	1,81	1,72	0,29	1,77	1,49	1,87	0,30	0,23	0,26	0,26	1,22	1,43	3,08
K₂O	0,86	0,63	0,67	1,45	0,81	1,11	1,81	0,96	0,79	0,65	1,84	0,95	1,35	0,93	1,43	1,78	1,36
P₂O₅	0,20	0,70	1,59	0,66	0,58	0,70	0,42	0,90	0,93	0,47	0,32	0,24	0,31	0,22	0,25	0,38	0,77
SO₃	0,07	0,11	0,39	0,52	0,21	0,18	0,26	0,19	0,18	0,15	0,33	0,25	0,27	0,21	0,22	0,28	0,35
Cr₂O₃	0,03	0,03	0,03	0,04	0,04	0,06	0,04	0,04	0,07	0,08	0,03	0,03	0,02	0,03	0,03	0,03	0,04
Co₃O₄	0,03	0,03	0,04	0,03	0,00	0,00	0,00	0,03	0,00	0,00	0,02	0,02	0,02	0,02	0,00	0,02	0,00
NiO	0,01	0,01	0,02	0,01	0,01	0,01	0,01	0,00	0,03	0,03	0,00	0,01	0,01	0,02	0,02	0,02	0,02
CuO	0,01	0,01	0,04	0,01	0,02	0,00	0,01	0,02	0,02	0,01	0,01	0,01	0,00	0,01	0,01	0,01	0,02
ZnO	0,00	0,01	0,02	0,01	0,01	0,01	0,01	0,01	0,01	0,01	0,02	0,01	0,01	0,01	0,01	0,01	0,02
SrO	0,00	0,00	0,02	0,01	0,03	0,03	0,01	0,04	0,03	0,04	0,01	0,01	0,00	0,01	0,02	0,01	0,02
ZrO₂	0,03	0,02	0,03	0,03	0,03	0,02	0,03	0,02	0,02	0,02	0,03	0,03	0,03	0,03	0,03	0,03	0,03
Total	87,46	85,18	82,99	71,07	89,14	88,65	83,33	89,28	90,32	92,87	79,52	86,48	80,40	87,17	87,63	86,15	84,45

* Total Fe is reported as Fe₂O₃

APPENDIX E3: INTRA-ELEMENT CORRELATION MATRIX (r) FOR THE SYN-BUSHVELD IGNEOUS ROCK SERIES.

	SiO ₂	TiO ₂	Al ₂ O ₃	FeOt	Mn ₃ O ₄	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Sc	V
TiO ₂	-0,64382	1										
Al ₂ O ₃	0,631303	-0,12571	1									
FeOt	-0,86505	0,755455	-0,47069	1								
Mn ₃ O ₄	-0,83499	0,617067	-0,57466	0,867916	1							
MgO	-0,42716	-0,1854	-0,85628	0,178096	0,268399	1						
CaO	-0,22596	-0,11318	-0,54978	-0,13167	0,190804	0,587362	1					
Na ₂ O	0,714322	-0,10639	0,789504	-0,42275	-0,54807	-0,74721	-0,63181	1				
K ₂ O	0,706102	-0,26424	0,661171	-0,43436	-0,50069	-0,62015	-0,6226	0,778123	1			
P ₂ O ₅	-0,28093	0,617605	0,187284	0,49892	0,435139	-0,40073	-0,44204	0,260885	0,035642	1		
Sc	-0,35417	0,152338	-0,39212	0,033684	0,046168	0,457774	0,712487	-0,56595	-0,64289	-0,41518	1	
V	-0,4676	0,668937	-0,18246	0,371631	0,208861	0,055642	0,218703	-0,23089	-0,34734	0,020489	0,665633	1
Cr	-0,18408	-0,33426	-0,56804	-0,00771	0,011936	0,812564	0,368503	-0,58779	-0,3692	-0,46738	0,44851	-0,01685
Co	-0,20614	-0,32368	-0,65529	-0,08488	-0,08818	0,802538	0,493181	-0,56265	-0,43546	-0,67433	0,511119	0,220127
Ni	-0,28448	-0,19452	-0,60014	0,159523	0,072053	0,785468	0,234782	-0,61675	-0,45748	-0,38811	0,284467	0,028558
Cu	-0,31978	0,543805	0,025584	0,273238	0,254335	-0,17058	-0,02964	-0,00855	-0,09813	0,262328	0,212058	0,492046
Zn	-0,78504	0,853691	-0,30905	0,947994	0,834634	-0,01782	-0,20041	-0,29078	-0,32742	0,657161	-0,08436	0,338585
Ga	0,393829	0,290127	0,85308	-0,17165	-0,32125	-0,92519	-0,63792	0,777341	0,660626	0,41574	-0,41359	0,074987
As	-0,92133	0,719903	-0,60703	0,953931	0,929097	0,295464	0,087857	-0,56599	-0,51877	0,383689	0,147824	0,417425
Se	-0,91765	0,764519	-0,56811	0,970664	0,911071	0,2563	0,068456	-0,52688	-0,52226	0,39765	0,189793	0,471352
Br	-0,91256	0,736543	-0,58331	0,977307	0,924229	0,274623	0,042139	-0,52874	-0,52279	0,42165	0,136879	0,425429
Rb	0,56388	-0,17389	0,68523	-0,3852	-0,4358	-0,58817	-0,51716	0,579508	0,853657	0,184107	-0,48385	-0,25817
Sr	0,071842	0,230995	0,568675	-0,11712	-0,26515	-0,60778	-0,26791	0,34354	-0,04027	0,292423	0,001809	0,138895
Y	-0,29118	0,726429	0,317301	0,471532	0,342236	-0,61081	-0,38907	0,144746	0,011864	0,586197	-0,07805	0,363104
Zr	0,646467	-0,12679	0,606311	-0,39351	-0,41867	-0,70253	-0,55874	0,660779	0,810244	0,158534	-0,68679	-0,34719
Nb	0,136008	0,405031	0,309524	0,129217	0,084915	-0,51696	-0,50286	0,386448	0,481738	0,567696	-0,52924	-0,08708
Cd	-0,1494	-0,22888	-0,23116	0,125721	0,117136	0,357116	-0,01803	-0,1871	-0,36791	-0,02953	-0,01059	-0,39176
Sn	0,285675	0,018263	0,478286	-0,12689	-0,29184	-0,51901	-0,55658	0,54544	0,274333	0,264936	-0,36018	-0,22223
Ba	0,588687	-0,00646	0,734892	-0,30073	-0,4275	-0,73917	-0,64093	0,810812	0,788732	0,266198	-0,6466	-0,23721

APPENDIX E3 (CONTINUED): INTRA-ELEMENT CORRELATION MATRIX (*r*) FOR THE SYN-BUSHVELD IGNEOUS ROCK SERIES.

	Cr	Co	Ni	Cu	Zn	Ga	As	Se	Br	Rb	Sr	Y
Co	0,668719	1										
Ni	0,785974	0,687859	1									
Cu	-0,23593	-0,19428	-0,14667	1								
Zn	-0,2091	-0,2938	-0,0203	0,285016	1							
Ga	-0,70645	-0,73402	-0,70463	0,303212	0,033323	1						
As	0,044128	0,060783	0,174796	0,281274	0,900088	-0,29763	1					
Se	0,022873	0,016235	0,150265	0,315064	0,91131	-0,25613	0,989789	1				
Br	0,039473	0,019204	0,182327	0,301618	0,909453	-0,27851	0,988104	0,994328	1			
Rb	-0,3185	-0,47072	-0,40763	-0,01836	-0,24219	0,628085	-0,45337	-0,46704	-0,47324	1		
Sr	-0,52668	-0,35248	-0,43519	0,170698	0,02434	0,584291	-0,1954	-0,15773	-0,18274	0,048498	1	
Y	-0,57277	-0,62346	-0,4119	0,599046	0,609482	0,575674	0,384696	0,426262	0,405434	0,15274	0,510777	1
Zr	-0,5797	-0,52928	-0,46673	0,005571	-0,19632	0,687729	-0,46336	-0,483	-0,47154	0,71446	0,152624	0,240411
Nb	-0,49645	-0,65601	-0,33012	0,387289	0,351146	0,577633	0,053759	0,04698	0,049887	0,529581	0,049073	0,588271
Cd	0,374706	0,198033	0,352967	-0,10163	-0,00083	-0,40766	0,062615	0,056387	0,085801	-0,36064	0,064574	-0,1007
Sn	-0,44102	-0,33691	-0,32646	0,097938	-0,01824	0,464017	-0,22924	-0,22263	-0,23474	0,240337	0,608622	0,455617
Ba	-0,5988	-0,61369	-0,41178	0,149148	-0,13771	0,790262	-0,43857	-0,41383	-0,41089	0,669132	0,245757	0,251704
	Zr	Nb	Cd	Sn								
Nb	0,721237	1										
Cd	-0,40196	-0,29236	1									
Sn	0,357277	0,338385	0,397321	1								
Ba	0,820664	0,637269	-0,33415	0,369346								

APPENDIX F1: CHEMICAL PARAMETERS OF SOIL COMPOSITE SAMPLES (KKSA), DETERMINED WITH WET CHEMICAL ANALYSES. ECO ANALYTICA LABORATORY, NWU, POTCHEFSTROOM.

Geobotanical Unit	Sample	pH	OC	Nutrient availability							Exchangeable cations						
				Ca	Mg	K	Na	P	NO ₃	Fe	CEC	Ca	Mg	K	Na	Ca:Mg	Base saturation
			(% C)	(mg/kg ⁻¹)					mmo l/L ⁻¹	μmol/ L ⁻¹	(cmol(+)/kg ⁻¹)						(%)
Alluvial floodplain (A)	KKSA1	5.26	1.23	1069.0	268.5	108.5	3.0	10.3	0.17	19.83	10.34	5.33	2.21	0.28	0.01	2.41	75.78
	KKSA2	5.79	1.22	900.0	183.5	131.5	1.5	31.4	0.21	20.67	8.85	4.49	1.51	0.34	0.01	2.97	71.68
Quartzite ridge southern aspect (B)	KKSA3	6.48	4.63	4492.5	481.5	159.0	2.0	59.4	1.53	23.26	21.84	22.42	3.96	0.41	0.01	5.66	122.68
	KKSA4	4.05	5.41	1905.0	191.5	111.0	2.5	70.1	1.40	11.50	21.01	9.51	1.58	0.28	0.01	6.03	54.16
Exposed Quartzite Ridges (C)	KKSA11	4.05	3.61	410.0	62.0	69.0	2.0	19.5	0.52	12.47	12.83	2.05	0.51	0.18	0.01	4.01	21.36
	KKSA13	4.03	3.56	413.5	54.5	53.5	1.5	28.0	0.75	10.76	14.04	2.06	0.45	0.14	0.01	4.60	18.91
Mafic Zone (D1)	KKSA6	5.40	2.21	1260.5	246.5	139.5	1.5	61.8	0.60	20.30	12.45	6.29	2.03	0.36	0.01	3.10	69.73
	KKSA8	5.82	2.14	1636.0	254.5	410.0	1.5	69.2	1.57	27.76	12.07	8.16	2.09	1.05	0.01	3.90	93.76
	KKSA9	7.01	2.38	1826.5	246.5	193.5	2.0	46.8	1.18	30.48	14.41	9.56	2.03	0.50	0.01	4.71	94.92
	KKSA10	4.60	2.25	979.0	199.5	113.5	4.0	30.8	0.46	30.96	12.56	4.89	1.64	0.29	0.02	2.98	54.44
Dioritic Zone (D2)	KKSA5	5.61	2.85	2014.5	402.0	80.0	4.0	25.0	0.99	16.83	14.83	10.05	3.31	0.21	0.02	3.04	91.57
	KKSA7	4.97	3.04	1309.5	305.0	78.5	1.5	17.7	0.66	23.06	13.97	6.53	2.51	0.20	0.01	2.60	66.21
	KKSA12	5.33	3.56	1944.0	334.0	186.0	2.5	4.7	2.25	24.16	12.72	9.70	2.75	0.48	0.01	3.53	101.69
	KKSA14	6.44	3.67	2653.0	447.5	69.0	4.0	3.7	0.92	15.28	17.24	13.24	3.68	0.18	0.02	3.59	99.26
	KKSA15	5.91	3.03	1765.5	307.0	85.0	2.5	7.7	0.89	25.48	15.69	8.81	2.53	0.22	0.01	3.49	73.73
Peralkaline zone (D3)	KKSA16	6.94	3.48	2724.0	385.5	204.5	7.0	20.4	1.74	21.7	16.39	13.59	3.17	0.52	0.03	4.28	105.67
	KKSA17	7.28	4.65	4750.5	485.0	109.5	5.0	21.7	1.96	7.79	23.71	23.71	3.99	0.28	0.02	5.94	118.08

APPENDIX F2: ELEMENT COMPOSITION (mg/kg) OF ANALYSED SOIL COMPOSITE SAMPLES (KSB). (ICP-MS ANALYSIS) ECO ANALYTICA LABORATORY, NWU, POTCHEFSTROOM. (*SEE APPENDIX B2 FOR SAMPLE INFORMATION).

	KSB1	KSB1	KSB2	KSB2	KSB3	KSB3	KSB4	KSB4
	Total	Soluble	Total	Soluble	Total	Soluble	Total	Soluble
	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg
Be 9	0,41925	1,8645E-05	0,37325	1,95875E-05	0,71725	1,97775E-05	0,42775	0,000017325
B 11	0,02845	0,0031275	0,028925	0,0030775	0,028975	0,0031125	0,029	0,0031175
Na 23	441	6,6675	493,5	4,9925	361	22,985	346,5	6,825
Mg 24	4012,5	303,75	3847,5	318,75	1533	299,75	1223	292,5
Al 27	15735	2,0475	12105	0,29525	12490	1,48325	14112,5	1,1775
P 31	1231,5	1,577	2655	3,445	1051,5	2,23025	627	1,337
K 39	1511,25	119,8	2462,75	440	1397,25	176,675	843,5	126,625
Ca 43	4875	1989,75	8215	1533,75	5462,5	2264	3212,5	1739,75
Ti 47	683	0,093425	624,75	0,018955	273,75	0,041175	349,5	0,013885
V 51	116,6	2,49275E-05	98,95	0,000079325	41,7	0,0000306	49,425	0,00005615
Cr 53	276	0,0526	107,825	0,0167925	51,25	0,0140675	61,425	0,012255
Mn 55	394,25	6,8875	411,5	2,665	437,5	4,2775	530	15,2775
Fe 57	30075	17,68	26375	12,8375	16017,5	18,75	18247,5	14,365
Co 59	20,6525	0,0372	16,6975	0,0159925	14,7675	0,02945	23,6125	0,0691
Ni 60	77,5	0,0609	39	0,0359	39,45	0,08265	40,3	0,090325
Cu 63	34,075	0,0123375	62,875	0,021455	36,275	0,0743	35,325	0,0236775
Zn 66	34,2	0,000319	41,4	0,0010115	37,175	0,00096225	26,225	0,05985
As 75	1,62675	3,993E-05	1,24175	0,000047225	2,40975	0,0000297	3,9675	0,0000432
Se 82	0,131775	0,00030725	0,247875	0,00031975	0,33225	0,0002545	0,34275	0,000284
Rb 85	13,545	0,503	14,185	0,82825	19,96	0,7015	12,385	0,6385
Sr 88	34,825	19,1625	42,375	13,585	66,525	19,135	28,125	15,6425
Mo 95	0,36	0,0056275	0,37475	0,006215	0,5845	0,0061825	0,6165	0,0054225
Pd 105	0,409	0,06525	0,5185	0,047325	0,52225	0,0661	0,3885	0,053375
Ag 107	0,000090475	0,0000501	0,0069925	0,00004845	0,01473	0,00004685	0,0198375	0,0000473
Cd 111	0,056575	0,0000253	0,052975	0,000027875	0,058475	0,0000289	0,057775	2,30625E-05
Sb 121	0,1975	0,01222	0,171075	0,0120325	0,200525	0,0121525	0,226225	0,012165
Ba 137	80,875	23,75	88,325	18,7925	129,95	15,565	88	18,9175

APPENDIX F2 (CONTINUED): ELEMENT COMPOSITION (mg/kg) OF ANALYSED SOIL COMPOSITE SAMPLES (KSB). (ICP-MS ANALYSIS) ECO ANALYTICA LABORATORY, NWU, POTCHEFSTROOM. (*SEE APPENDIX B2 FOR SAMPLE INFORMATION).

	KSB1	KSB1	KSB2	KSB2	KSB3	KSB3	KSB4	KSB4
	Total	Soluble	Total	Soluble	Total	Soluble	Total	Soluble
	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg
Pt 195	0,0000552	0,000005305	0,000076075	0,00000588	0,000076775	9,8375E-06	0,0000696	9,8375E-06
Au 197	0,281	0,02438	0,2625	0,024245	0,386	0,0244125	0,26375	0,0244325
Hg 202	0,0000552	0,000005895	0,00005515	5,8775E-06	0,00005075	5,8425E-06	0,0000545	5,8925E-06
Tl 205	0,04915	0,0000118	0,027325	2,12625E-05	0,10115	9,6425E-06	0,0717	0,002615
Pb 208	5,645	0,000128875	3,6775	0,0001151	7,535	0,000151425	7,2775	0,000138325
Bi 209	0,35825	0,0281	0,3375	0,027025	0,379	0,027	0,384	0,027025
Th 232	1,50925	0,019855	1,33225	0,019285	2,30175	0,0200325	2,1995	0,0192825
U 238	0,278	0,000007295	0,2184	0,000007765	0,573	0,000004545	0,38825	4,5275E-06

**APPENDIX G1: ELEMENT COMPOSITION (mg/kg) OF ANALYSED PLANT TISSUE SAMPLES FROM SELECTED WOODY SPECIES. (ICP-MS ANALYSIS)
ECO ANALYTICA LABORATORY, NWU, POTCHEFSTROOM. (*SEE APPENDIX C3 FOR SAMPLE INFORMATION).**

	AO1 AK LS	AO1 AK SS	AO1 AR LS	AO1 AR SS	AO1 GF LS	AO1 GF SS	AO2 AK LS	AO2 AK SS	AO2 AR LS	AO2 AR SS
	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg
Be 9	0,000188325	0,00017135	0,0001921	0,000152475	0,00014305	0,0001657	0,0002091	0,00017135	0,00019965	0,0001638
B 11	0,17985	0,19465	0,19715	0,220375	0,19045	0,212525	0,202875	0,221375	0,2056	0,236175
Na 23	383	381,5	387,5	405,25	437,5	394,75	367,25	366	378,25	359,5
Mg 24	2912,5	1416,25	2288	1841	1859,75	1497,25	3130	1517,75	2480,25	1908,5
Al 27	37,975	104,125	42,05	90,7	174,55	55,725	32,225	94,125	45,225	70,4
P 31	1494,25	1255,5	1412,75	1518	1528,75	1165,25	1703,25	2185,5	1489,5	2223
K 39	7142,5	5702,5	8930	5422,5	4535	5947,5	8392,5	7192,5	8010	7727,5
Ca 43	13205	13237,5	19682,5	19942,5	12997,5	6847,5	14157,5	13202,5	20822,5	18047,5
Ti 47	3,3075	4,085	3,1025	4,44	6,4725	4,6875	3,7475	5,815	3,27	5,8625
V 51	0,0898	0,5835	0,37025	0,43625	0,747	0,59975	0,18115	1,043	0,98125	1,16725
Cr 53	1,95175	5,045	4,2075	4,12	5,52	6,0425	2,915	9,5825	9,4075	10,465
Mn 55	30,4	18,0575	32,75	17,26	48,65	26,9	27,75	18,1925	41,6	22,1925
Fe 57	154,8	225,95	236,875	268	396,25	148,05	149	202,175	240,2	226,025
Co 59	1,28775	7,235	3,465	6,1675	10,2275	10,0125	2,6125	4,255	3,8925	3,61
Ni 60	3,1325	2,5975	3,235	4,63	3,1825	1,86575	1,5905	2,3225	1,88625	3,8725
Cu 63	4,0525	4,875	7,6	5,8675	5,3325	4,2925	4,2175	4,8575	7,1275	6,695
Zn 66	11,8275	16,88	12,9925	17,0475	17,0425	20,7975	13,6975	16,81	13,5425	20,925
As 75	0,000122225	0,067275	0,078925	0,054875	0,100075	0,072375	0,0151225	0,1719	0,208	0,225875
Se 82	0,00089875	0,00127875	0,000371	0,001511	0,00121525	0,00203875	0,0014265	0,001532	0,00089875	0,00121525
Rb 85	10,92	6,845	11,47	8,8175	1,69375	1,71325	10,4525	7,1075	8,1725	7,105
Sr 88	94,95	149,1	126,4	170,75	67,175	50,225	86,575	120,05	106,875	125
Mo 95	0,114825	0,120075	0,109475	0,154025	0,1892	0,1102	0,1996	0,2515	0,173975	0,33725
Pd 105	0,32375	0,51875	0,4355	0,54875	0,2454	0,185925	0,297	0,43725	0,37825	0,439
Ag 107	0,0004665	0,00036625	0,000287	0,00033225	0,00036725	0,000389	0,000436	0,00032775	0,0004075	0,00039925
Cd 111	0,0220025	0,03335	0,02183	0,0164975	0,04865	0,035075	0,0214875	0,025625	0,0206275	0,017875
Sb 121	0,13285	0,138	0,137275	0,14145	0,147425	0,131725	0,13075	0,137325	0,1351	0,1351
Ba 137	44,6	56,55	65,075	63,775	15,82	14,6025	34,4	36,95	54,15	40,65

APPENDIX G1 (CONTINUED): ELEMENT COMPOSITION (mg/kg) OF ANALYSED PLANT TISSUE SAMPLES FROM SELECTED WOODY SPECIES. (ICP-MS ANALYSIS) ECO ANALYTICA LABORATORY, NWU, POTCHEFSTROOM. (*SEE APPENDIX C3 FOR SAMPLE INFORMATION).

	AO1 AK LS	AO1 AK SS	AO1 AR LS	AO1 AR SS	AO1 GF LS	AO1 GF SS	AO2 AK LS	AO2 AK SS	AO2 AR LS	AO2 AR SS
	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg
Pt 195	0,0000847	0,00008325	0,000085425	0,000091175	0,000087575	0,0000962	0,00009045	0,000081825	0,000094775	0,000087575
Au 197	0,26275	0,28175	0,2725	0,2625	0,269	0,2595	0,25925	0,3175	0,26075	0,26475
Hg 202	0,0000354	2,43825E-05	0,00208625	0,0000323	0,00002825	0,000032575	0,00004345	0,000043225	0,0000442	0,000049025
Tl 205	0,00025325	0,000158075	0,00025875	0,000184375	0,00032825	0,00032325	0,000262	0,000151625	0,0002605	0,00025925
Pb 208	0,37925	0,1847	0,54475	0,141775	0,93575	0,06195	0,41375	0,1049	0,304	0,031925
Bi 209	0,307	0,30775	0,28175	0,281	0,29	0,28075	0,28075	0,2825	0,27925	0,2765
Th 232	0,278	0,272	0,26775	0,27925	0,28325	0,2515	0,26975	0,265	0,271	0,25525
U 238	0,00006635	0,00115375	0,00005775	2,2905E-06	0,010985	0,0000505	0,000068375	1,81425E-05	0,00005945	0,000028925

APPENDIX G1 (CONTINUED): ELEMENT COMPOSITION (mg/kg) OF ANALYSED PLANT TISSUE SAMPLES FROM SELECTED WOODY SPECIES. (ICP-MS ANALYSIS) ECO ANALYTICA LABORATORY, NWU, POTCHEFSTROOM. (*SEE APPENDIX C3 FOR SAMPLE INFORMATION).

	AO2 GF LS	AO2 GF SS	AO3 AC LS	AO3 AC SS	AO3 AK LS	AO3 AK SS	AO3 AR LS	AO3 AR SS	AO3 GF LS	AO3 GF SS
	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg
Be 9	0,000139275	0,000188325	0,0001789	0,000175125	0,000175125	0,000169475	0,0001223	0,000160025	0,00014305	0,00017135
B 11	0,2115	0,236875	0,23555	0,23925	0,24925	0,025325	0,025425	0,026325	0,242925	0,026225
Na 23	380,75	387,25	366,5	351,5	321	354,5	1060,75	2098,5	359,25	349
Mg 24	1899	1648,5	3727,5	2617,5	3092,5	1317,5	3830	2615	2370	1417
Al 27	105,15	51,25	46,075	71,625	26,85	83,225	39,725	68,45	143,15	42,925
P 31	1738,75	1418,75	1391,75	1509,75	1163	1232,5	1376	1551,25	1561,25	946,25
K 39	3367,5	6142,5	5642,5	8997,5	6200	6565	8182,5	5942,5	5550	5152,5
Ca 43	13160	8322,5	12830	8200	16052,5	10142,5	18325	14537,5	10512,5	6255
Ti 47	5,525	3,9075	3,4375	4,4425	2,8575	3,91	3,4375	4,175	4,835	2,62
V 51	1,241	0,5225	0,2645	1,21775	1,235	0,472	1,0175	0,3485	0,226225	0,19915
Cr 53	10,3725	5,3225	3,015	10,2025	11,3125	4,52	9,4875	3,68	2,17875	2,7975
Mn 55	37,675	23,3	71,5	22,4725	31	20,675	65,075	29,95	31,65	19,9875
Fe 57	254,75	118,025	178,75	144,075	171,425	167,825	202,825	186,975	224,3	93,2
Co 59	11,5	5,42	2,201	3,0975	1,5065	5,86	2,61	2,68	8,7	6,5575
Ni 60	2,7825	1,5045	7,4425	10,1725	1,8995	2,715	2,4885	5,8275	7,7825	1,05525
Cu 63	6,1025	6,61	4,9775	6,375	3,4975	4,77	5,3	4,79	5,2675	4,69
Zn 66	17,455	23,2025	12,485	26,225	8,2425	13,2525	12,245	13,9925	13,645	16,8175
As 75	0,2164	0,04575	0,01403	0,171175	0,232425	0,0202275	0,1905	0,0249675	0,00126775	0,00007845
Se 82	0,001173	0,0016375	0,0004975	0,00072975	0,0198925	0,001532	0,0006665	0,00203875	0,0006875	0,0022075
Rb 85	1,60425	2,575	6,8175	8,885	12,41	8,15	11,3075	7,98	5,445	4,7725
Sr 88	59,875	50,375	86,05	84,1	111,2	106,9	122,15	128,15	63,975	50,275
Mo 95	0,18745	0,122225	0,10515	0,191675	0,119025	0,244675	0,44	1,328	0,17995	0,112775
Pd 105	0,236875	0,192425	0,3205	0,3105	0,415	0,38575	0,4365	0,43575	0,2316	0,205475
Ag 107	0,0003395	0,00034925	0,00041675	0,0003915	0,00034925	0,00044475	0,000126075	0,000451	0,00044925	0,0004175
Cd 111	0,04575	0,049525	0,00463	0,003425	0,002565	0,0121975	0,0044575	0,0066925	0,016325	0,02595
Sb 121	0,1409	0,132625	0,135675	0,1359	0,13345	0,13355	0,133	0,128825	0,13415	0,1296
Ba 137	9,9175	10,7775	33,775	20,95	58,1	42,775	89,475	57,025	11,035	10,7475

APPENDIX G1 (CONTINUED): ELEMENT COMPOSITION (mg/kg) OF ANALYSED PLANT TISSUE SAMPLES FROM SELECTED WOODY SPECIES. ECO ANALYTICA LABORATORY, NWU, POTCHEFSTROOM. (*SEE APPENDIX C3 FOR SAMPLE INFORMATION).

	AO2 GF LS	AO2 GF SS	AO3 AC LS	AO3 AC SS	AO3 AK LS	AO3 AK SS	AO3 AR LS	AO3 AR SS	AO3 GF LS	AO3 GF SS
	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg
Pt 195	0,00009405	2,35525E-05	0,00008685	0,000089025	0,0000847	0,000101975	0,0000955	0,000093325	0,000089725	0,000083975
Au 197	0,27975	0,255	0,2705	0,2665	0,26875	0,25325	0,26325	0,25175	0,25025	0,26375
Hg 202	0,00002995	0,0000517	0,000039225	0,000047	0,0000545	0,0000471	0,000049025	0,000050175	0,00003075	0,000046025
Tl 205	0,000321	0,00031925	0,00031125	0,000237175	0,0002615	0,00023245	0,00028975	0,000268	0,00031525	0,0003065
Pb 208	0,954	0,029325	0,31825	0,05145	0,60475	0,066675	0,22995	0,0066175	0,77725	0,0030375
Bi 209	0,282	0,27925	0,27975	0,2825	0,2765	0,276	0,2875	0,278	0,2935	0,2815
Th 232	0,254	0,249225	0,25075	0,25425	0,25625	0,2605	0,249075	0,26325	0,26525	0,25175
U 238	2,06725E-05	0,000038875	0,000056925	0,000042575	0,00006435	0,00002925	0,0000635	0,000046625	0,0000279	0,000059625

APPENDIX G1 (CONTINUED): ELEMENT COMPOSITION (mg/kg) OF ANALYSED PLANT TISSUE SAMPLES FROM SELECTED WOODY SPECIES. ECO ANALYTICA LABORATORY, NWU, POTCHEFSTROOM. (*SEE APPENDIX C3 FOR SAMPLE INFORMATION).

AO4 AC LS		AO4 AC SS	AO4 AR LS	AO4 AR SS	AO4 AC LS		AO4 AC SS	AO4 AR LS	AO4 AR SS
	mg/kg	mg/kg	mg/kg	mg/kg		mg/kg	mg/kg	mg/kg	mg/kg
Be 9	0,0001789	0,000190225	0,00012985	0,000146825	Pt 195	0,000034325	0,00008685	0,00009405	0,000101975
B 11	0,02735	0,04755	0,02605	0,02775	Au 197	0,33525	0,26275	0,256	0,252
Na 23	354,5	348	339,5	361	Hg 202	0,000033475	0,000052225	2,02225E-05	0,000052525
Mg 24	3142,5	2510	3305	1722,25	Tl 205	0,00031125	0,00026275	0,000113675	0,00020765
Al 27	46,725	45,025	29,275	46,975	Pb 208	0,3315	0,016075	0,27475	0,006465
P 31	1286,5	1137	1238,5	976,75	Bi 209	0,301	0,28175	0,28	0,27725
K 39	5575	7570	9380	5992,5	Th 232	0,256	0,27125	0,26875	0,25125
Ca 43	12870	9672,5	18017,5	12085	U 238	0,00101875	0,00005085	0,00007495	0,0000446
Ti 47	3,3225	3,5775	2,985	2,915					
V 51	0,233925	0,89675	0,826	0,66475					
Cr 53	2,86	8,43	8,3775	6,7275					
Mn 55	63,125	18,7875	75,45	29,1					
Fe 57	171,275	131,925	193,7	159,125					
Co 59	3,765	2,314	2,765	3,415					
Ni 60	3,3975	4,445	1,80825	7,405					
Cu 63	5,455	5,62	5,935	4,9125					
Zn 66	13,0625	24,0675	10,2075	14,7575					
As 75	0,0096525	0,133975	0,1577	0,085125					
Se 82	0,00121525	0,0006875	0,053675	0,001511					
Rb 85	6,1525	6,935	13,255	7,76					
Sr 88	76,725	82,8	115,8	105					
Mo 95	0,114925	0,1531	0,0995	0,182825					
Pd 105	0,30575	0,325	0,427	0,375					
Ag 107	0,0003725	0,0003525	0,0001204	0,000144125					
Cd 111	0,01323	0,01065	0,0194225	0,0066925					
Sb 121	0,135975	0,132725	0,17085	0,1292					
Ba 137	25,4	17,2425	71,25	45,475					

APPENDIX H1: STATISTICAL SUMMARY OF THE CANONICAL CORRESPONDENCE ANALYSES (CCA) PRESENTED IN FIGURE 68.

All data (Figure 68a)						Selected data (Figure 68b)					
No samples omitted						Ignored outliers 1, 9, 10, 17, 24, 25, 26, 27, 30					
Number of samples		31				Number of samples		22			
Number of species		35				Number of species		30			
Number of occurrences		353				Number of occurrences		289			
No transformation of species data						No transformation of species data					
No species-weights specified						No species-weights specified					
No sample-weights specified						No sample-weights specified					
No down-weighting of rare species						No down-weighting of rare species					
**** Summary ****						**** Summary ****					
Axes	1	2	3	4	Total inertia	Axes	1	2	3	4	Total inertia
Eigenvalues:	0.657	0.480	0.290	0.240	3.188	Eigenvalues:	0.373	0.308	0.213	0.163	1.732
Species-environment correlations:	0.955	0.952	0.953	0.947		Species-environment correlations:	0.973	0.977	0.948	0.960	
Cumulative percentage variance-						Cumulative percentage variance-					
-of species data:	20.6	35.7	44.8	52.3		-of species data:	21.5	39.3	51.6	61.0	
-of species-environment relation:	30.2	52.3	65.7	76.7		-of species-environment relation:	27.9	51.0	66.9	79.1	
Sum of all eigenvalues	3.188					Sum of all eigenvalues	1.732				
Sum of all canonical eigenvalues	2.173					Sum of all canonical eigenvalues	1.336				

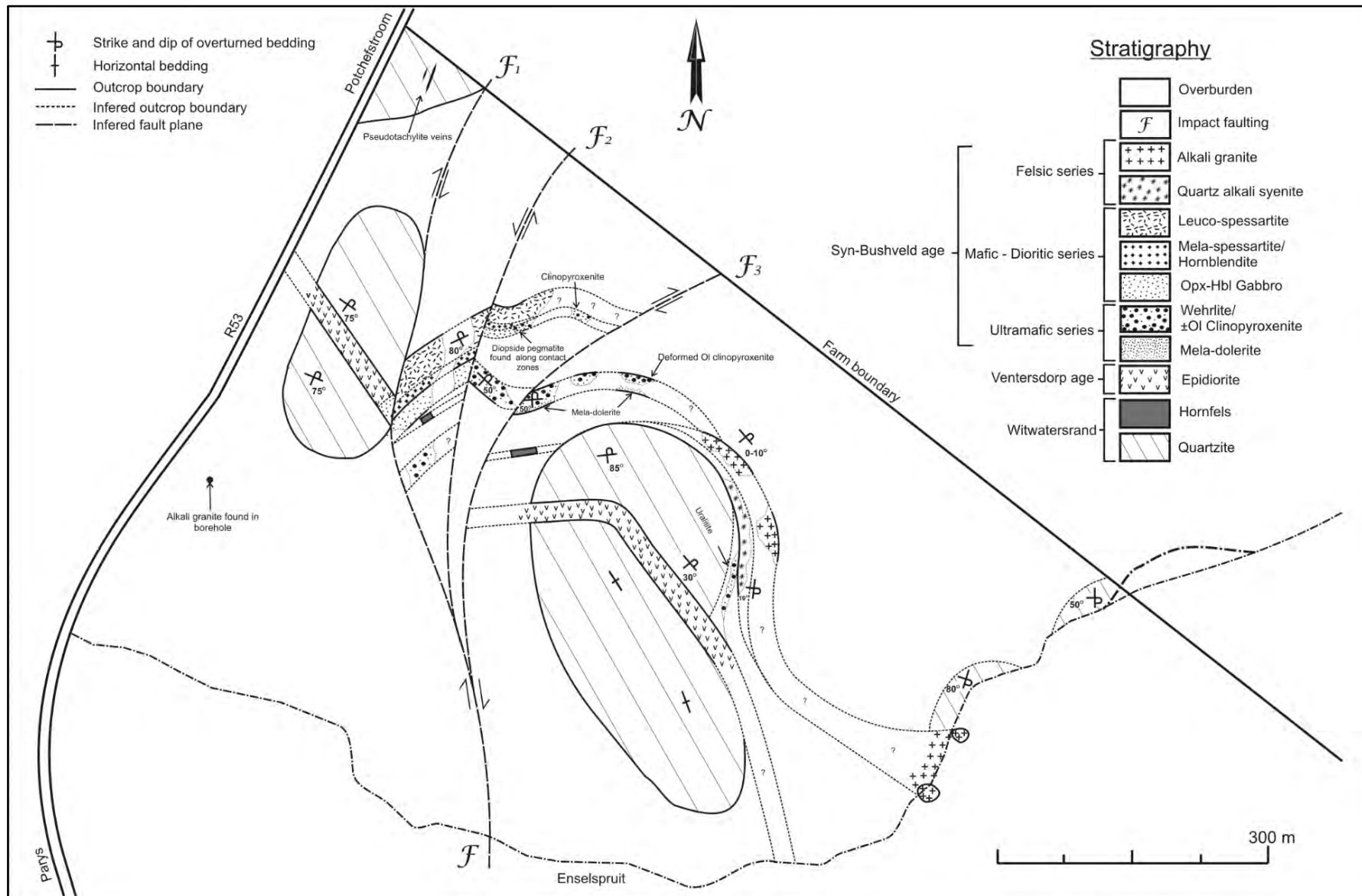
APPENDIX H2: SPECIES PERFORMANCE INDEX VALUES (IV) FOR WOODY TREE SPECIES OF THE KOEDOESFONTEIN COMPLEX.

Tree species	Survey plot no.															
	KV1	KV2	KV3	KV4	KV5	KV6	KV7	KV8	KV9	KV10	KV11	KV12	KV13	KV14	KV15	KV16
<i>Brachylaena rotundata</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Buddleja saligna</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Celtis africana</i>	0	5.5	1.2	0	0	0	0.1	2	0	0	0	0	0.4	0	0	0.4
<i>Combretum molle</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Dombeya rotundifolia</i>	0	0	0	0	0	0	0	0	0	0	0	1.56	0.86	1.94	0.88	2.94
<i>Euclea crispa</i>	0	0.9	0.44	0	0.34	0	0	0.7	0	0	0	0.3	1.16	0	0	0
<i>Gymnosporia buxifolia</i>	0	1.06	0	0	0	0	0	1.2	0	0	0	0	0	0	0	0.38
<i>Mystroxydon aethiopicum</i>	0	0	0	0	0	0	0	2.44	0	0	0	0	0	0	0	0
<i>Olea europaea subsp. africana</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Pappea capensis</i>	0	0	0.34	0	0	0.3	0	0	0	0	1.36	1.34	1.08	0	0	0.3
<i>Pavetta zeyheri</i>	0	0	0	0	0.24	0	0	0	0	0	0	0	0	0	0	0
<i>Pyracantha angustifolia</i>	1.8	0	0	0	0	0	0	0	0.3	0	0	0	0	0	0	0
<i>Searsia leptodictya</i>	0	0.66	0	0	0	0.6	0	0	0	0	0	0	0.2	0	2.1	3.71
<i>Searsia pyroies</i>	0	1.95	0.36	0.44	0	0	0	0	0	0	0	0.46	0	0	0	0
<i>Senegalia caffra</i>	0	0.74	0	0	0	0	0	2.58	0	0	0	3.24	4.4	4.5	11.8	0.58
<i>Vachellia karroo</i>	0	4	8.61	10.1	9.3	3.2	2.7	5.6	0	0	0	0.6	0.28	0	0	0
<i>Vachellia robusta</i>	0	0	0	0	0	1.3	4.08	0	0	0	3.48	0.78	0	7.8	0	1.56
<i>Zanthoxylum capensis</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0.22
<i>Ziziphus mucronata</i>	0	6.6	0.52	3.4	2.8	6	2	3.08	0.3	0.52	0	2.04	1.12	0	1.6	0.64

APPENDIX H2 (CONTINUED): SPECIES PERFORMANCE INDEX VALUES (IV) FOR WOODY TREE SPECIES OF THE KOEDOESFONTEIN COMPLEX.

Tree species	Survey plot no.														
	KV17	KV18	KV19	KV20	KV21	KV22	KV23	KV24	KV25	KV26	KV27	KV28	KV29	KV30	KV31
<i>Brachylaena rotundata</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0.26	0
<i>Buddleja saligna</i>	0	7.2	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Celtis africana</i>	0	1.54	2.24	0	0	0.2	0	0	0	0	0	0	0.2	0	0
<i>Combretum molle</i>	0	0	0	0	0	1	0	9.96	6.1	0.96	0.04	0	0	10.3	0.56
<i>Dombeya rotundifolia</i>	0	2.82	3.3	0.8	2.36	0	0	1.4	1.2	0.26	0	0.2	0	0	0
<i>Euclea crispa</i>	0	0	0.4	0.6	0.52	0.4	0.28	0	0	0	0	0.28	0	0	0
<i>Gymnosporia buxifolia</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Mystroxydon aethiopicum</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Olea europaea subsp. africana</i>	0	0	0	0	0	0	0	0	0	0	1.2	0	0	0	0
<i>Pappea capensis</i>	0	0	0	4.55	3	1.6	1.24	0	0	0	0	2.76	4.38	0	3.63
<i>Pavetta zeyheri</i>	0	0	0	0	0	0.4	0.42	0	0	0	0	0	0	0	0
<i>Pyracantha angustifolia</i>	0.4	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Searsia leptodictya</i>	0	1.4	0	0.5	0.68	0.6	1.28	2.13	0	0	0	0	0.24	2.1	0
<i>Searsia pyroies</i>	0	0	0	0	0	0	0.28	0	0	0	0	0.16	0	0	0
<i>Senegalia caffra</i>	0	6.72	4.2	1.1	1.24	7.6	5.16	2.36	2.3	4.18	0	0.28	0	0.44	0
<i>Vachellia karroo</i>	0	0	0	0	0	3	7.8	0	0	0	0	0.8	0	0	0
<i>Vachellia robusta</i>	0	0	0	8.91	11.18	0	0	0	0	0	0	14.4	11.4	0	12.3
<i>Zanthoxylum capensis</i>	0	0	0	1.3	0	0	0	0	0	0	0	0	0	0	0
<i>Ziziphus mucronata</i>	0.7	0	0	0	2.7	0.2	0.74	0	0	0	0	0	0.24	0	0

APPENDIX I1: DETAILED GEOLOGICAL MAP OF THE KOEDOESFONTEIN COMPLEX.



APPENDIX I2: ORTHOPHOTO AND GEOLOGICAL OVERLAY OF THE KOEDOESFONTEIN COMPLEX.



APPENDIX I3: GEOBOTANICAL MAP OF THE KOEDOESFONTEIN COMPLEX.

