

# **An Investigation of the carbonatites of the Kaapvaal Craton and their tectonic context**

by

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## **Abstract:**

Carbonatites are exotic rocks which usually occur in discrete intrusions. Considering the association of carbonatites with rifting environments, this dissertation proposes that: carbonatites occur in extensional tectonic settings and therefore carbonatite occurrence can be used to constrain an extensional setting for related occurrences. In order to give context in which to consider this hypothesis, the formation of carbonatites is reviewed. This work favours the direct mantle melting model as it is most relevant and consistent with observations of Kaapvaal Craton carbonatites. However the liquid immiscibility model cannot be entirely rejected with current knowledge. The hypothesis is applied to the Bushveld Igneous Complex after providing evidence of the rift-carbonatite association. The Bushveld Igneous Complex is considered to have been emplaced in the same tectonic setting as carbonatites. Therefore the Bushveld Igneous Complex was emplaced in an extensional tectonic setting. Finally the carbonatites which are part of the Pilanesberg Alkaline Province are considered in order to test the hypothesis. This work finds that the Pilanesberg carbonatites do occur with other rift related magmatism during the break-up of Nuna (Columbia) and therefore the hypothesis is not rejected. This work offers reviews of basic carbonatite background, formation models and carbonatite occurrences on the Kaapvaal Craton.

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## Chapter 1:

### Introduction

This dissertation examines carbonatites on the Kaapvaal Craton and the tectonic setting and geological history that these carbonatites reveal for their respective time periods. The carbonatites were emplaced in two distinct time periods, both in the Proterozoic. The older carbonatites, referred to as Bushveld carbonatites, include the Phalaborwa and Schiel (possibly) complexes that were emplaced around 2.06 Ga. The carbonatites included in the Pilanesberg Alkaline Province were emplaced between 1.43 Ga and 1.2 Ga (Harmer, 1992) and will be referred to as “Pilanesberg carbonatites”.

#### 1.1 Basic Background

##### *Melt properties*

Carbonatite magmas have low density and very low viscosity. Wolff (1994) estimated the viscosity and density of calcite-carbonatite magma from molten salt data to be 0.1 Pa s at 700-800 °C and 2.3-2.5 g.cm<sup>-3</sup> respectively. Low viscosities are due to low silica contents that result in little polymerization. The low viscosity results in the magma being unable to suspend crystals, which therefore grow on chamber walls, thus allowing significant crustal growth *in situ* (Treiman and Schedl, 1983).

##### *Occurrence*

Carbonatites occur almost exclusively on continents. The exceptions are found on the Kerguelen, Canary and Cape Verde Islands (Woolley and Kjarsgaard, 2008) and are collectively referred to as “oceanic carbonatites”. Carbonatites typically occur as pipe-like intrusions or dykes and are commonly part of alkaline silicate complexes. The relationship between carbonatite and associated silicate rocks is poorly understood; alkali silicate rocks and carbonatite may form from the same parental magma (involving liquid immiscibility) or independently (Bell *et al.*, 1999).

In addition to the alkaline silicate-carbonatite association, many carbonatites have also been associated (spatially and temporally) with Large Igneous Provinces (LIPs) (Ernst and Bell, 2009). LIPs are large volumes of mostly of flood basalts and their

“plumbing systems” and are emplaced quickly by intra-plate magmatism. Examples of LIPs associated with carbonatites include: the Deccan flood basalts; Paraná-Etendeka; the Siberian LIP; the Kola Alkaline province; the Central Iapetus Magmatic Province; the Keweenaw LIP; and, the Bushveld Igneous Complex (BIC) (Ernst and Bell, 2009).

As may be expected from the association of carbonatites with rift related occurrences like large igneous provinces (LIP's), a significant proportion of known carbonatites are associated with extensional environments (esp. the East African Rift Valley). There are, however, others associated with faults and domal swells (Ernst and Bell, 2009). Carbonatites are also known to be spatially related to group 1 kimberlites and ultramafic lamprophyres (Bell *et al.*, 1999). The rift-carbonatite link is considered in chapter 3.

### *Composition*

Carbonatites are defined by the recommended IUGS classification as igneous rocks containing more than 50% carbonate minerals by volume.

The above classification is lacking as it does not describe what is most often meant by the term carbonatite. The term carbonatite as often used (as in this work) denotes a group of carbonate dominated igneous rocks with characteristic high contents of incompatible elements including: REE's, Sr and Nd (Bell and Simonetti, 2009), the occurrence of noted accessory minerals (including: pyroxene, apatite, magnetite, olivine, biotite and hematite) in varied amounts and a similar (but debated) petrogenesis. As such, the term “carbonatite” often excludes carbonate-dominated rocks like carbonate-rich dykes associated with kimberlites.

Harmer and Gittins (1997) explain that carbonatites generally occur in a bimodal distribution where carbonatites are either predominantly dolomitic (dolomite-carbonatite) or calcitic (calcite-carbonatite). Calcite-carbonatite is more common and is usually associated with igneous complexes, whereas dolomite-carbonatite most often occurs in Archaean crust (Harmer and Gittins, 1997).



For reference: Zindler and Hart (1986) proposed that isotopic variability in the mantle is expressed by four end-members; HIMU (high  $\mu$ ; where  $\mu = {}^{238}\text{U}/{}^{204}\text{Pb}$ ), EM I (enriched mantle 1), EM II (enriched mantle 2) and DMM (depleted MORB mantle). These mantle end members are defined using radiogenic isotopes (esp. Sr, Nd and Pb). HIMU exhibits low  ${}^{87}\text{Sr}/{}^{86}\text{Sr}$  in addition to a high  $\mu$  value and indicates recycled oceanic crust (eclogite) (Chase, 1981). EM I is characterized by intermediate  ${}^{87}\text{Sr}/{}^{86}\text{Sr}$  and low  ${}^{206}\text{Pb}/{}^{204}\text{Pb}$  and indicates recycled ocean floor (Röhkamper and Hofmann, 1992) or mantle delamination (Mahoney *et al.*, 1991). EM II has high  ${}^{87}\text{Sr}/{}^{86}\text{Sr}$  intermediate  ${}^{206}\text{Pb}/{}^{204}\text{Pb}$  and indicates recycled continental crust (Hart, 1988). DMM represents the isotopic signature of mantle from which silicate melt has been extracted repeatedly (Debaille *et al.*, 2006). FOZO (focal zone) is characterised by moderately depleted Sr and Nd signatures, radiogenic Pb isotopes, and elevated  ${}^3\text{He}/{}^4\text{He}$  ratios (Hart *et al.*, 1992) and is thought to represent the isotopic composition of the lower mantle and is associated with plumes (Hauri *et al.*, 1994)

The isotopic characteristics of carbonatites are similar to those of ocean island basalts (OIBs) which further indicates a mantle source and a link to mantle upwellings or plumes (Bell and Simonetti, 2009). The presence of FOZO, HIMU, EM I (illustrated in figure 1.1) but not DMM mantle components in some carbonatites suggest a sub-lithospheric mantle source (Ernst and Bell, 2009; Bell and Simonetti, 2009). HIMU and EM1 are considered to be subducted lithosphere, due to HIMU and EM I signatures present in OIBs (Bell and Simonetti, 2009).

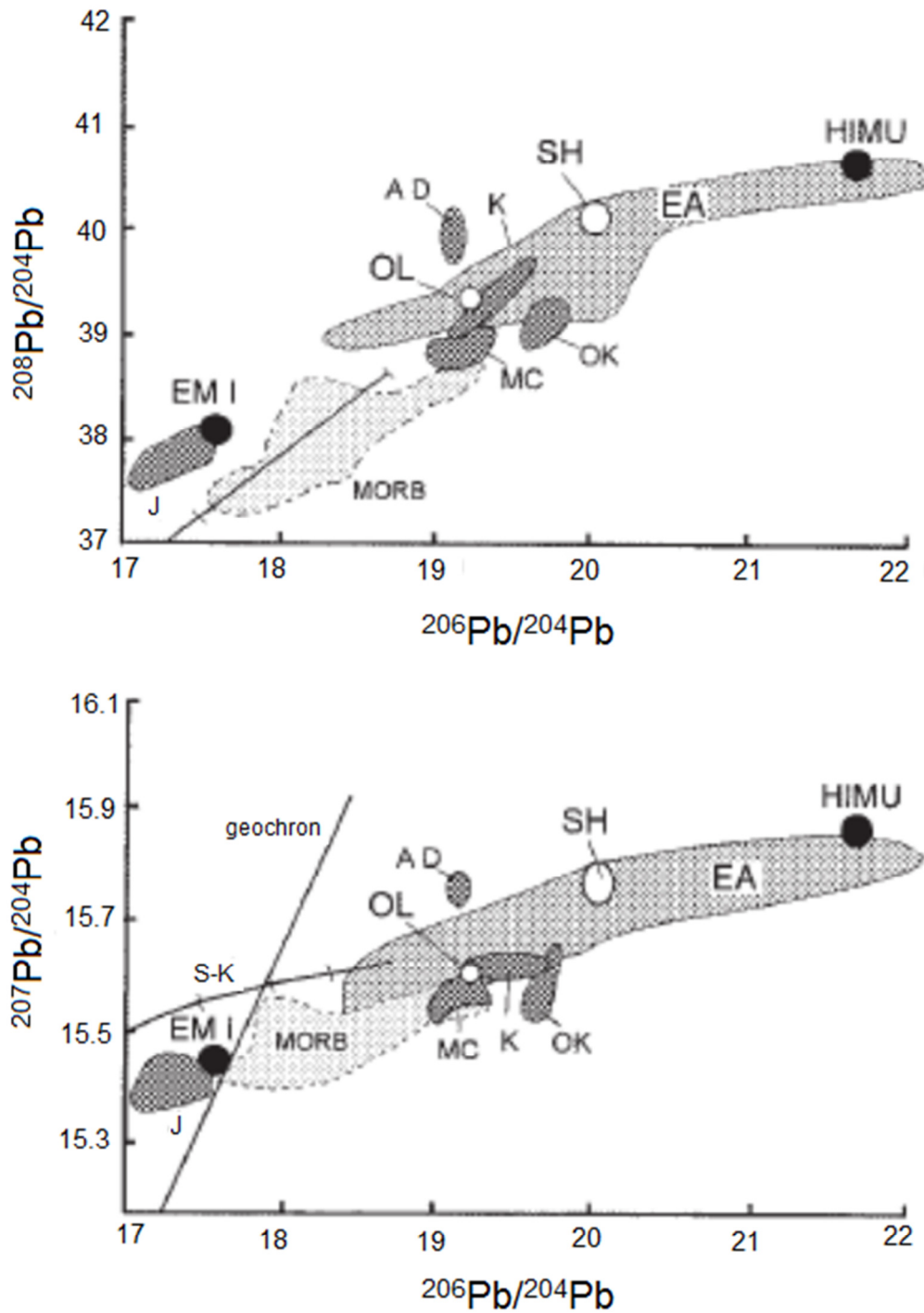


Figure 1.1: Pb isotopic plots of carbonatites illustrating the EM I and HIMU mantle sources. The figure includes data from the following carbonatites: Oldoinyo Lengai (OL), Shombole (SH), Amda Dongar (AD), East African carbonatites (EA), Kaiserstuhl (K), Oka (OK) and Magnet Cove (MC). After Bell (1998).

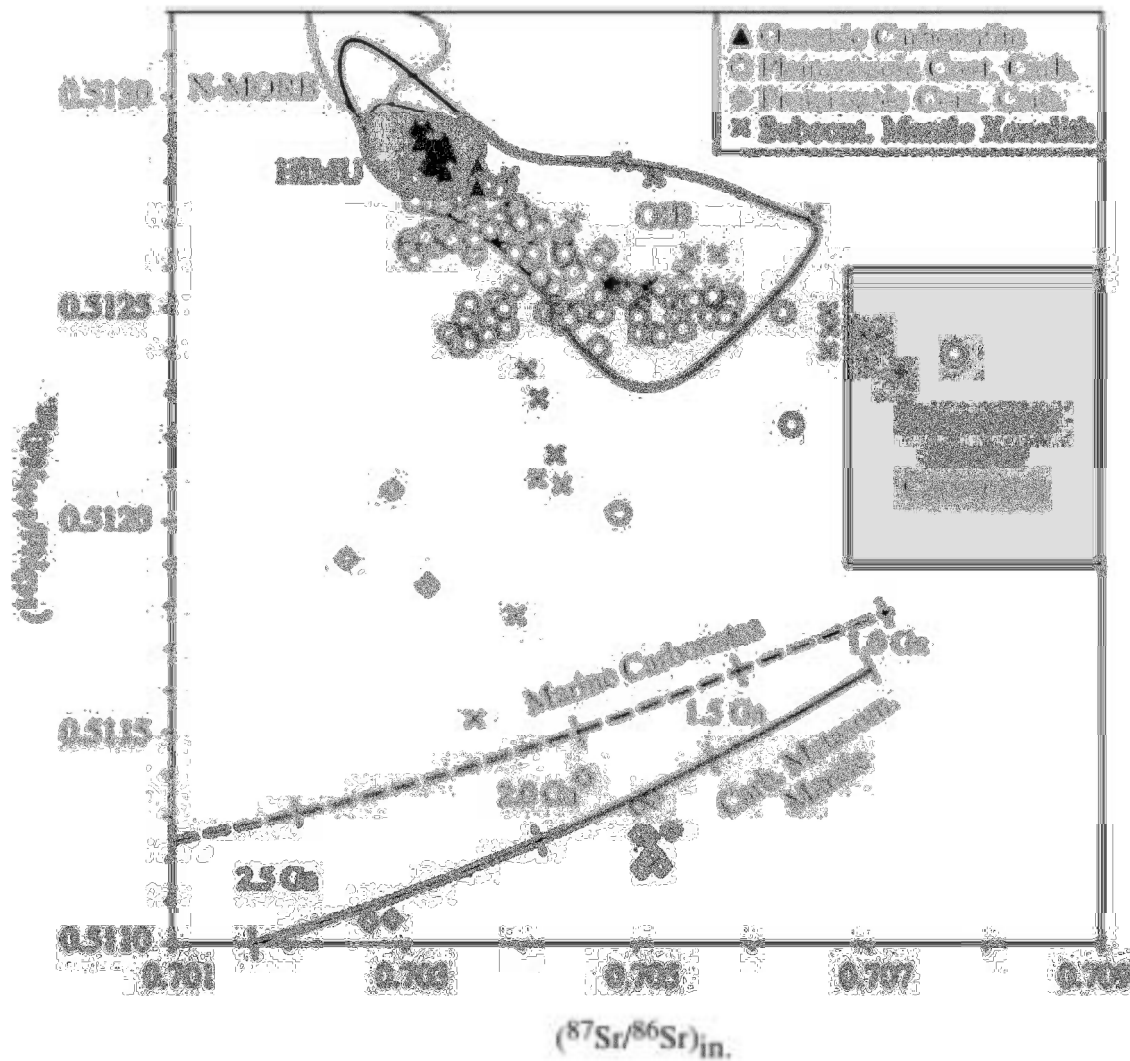


Figure 1.2: Initial Sr and Nd plots showing the similarity in isotopic compositions between OIB and carbonatite. After Hoernle *et al.* (2002).

### Origin

It has been established that carbonatites are derived from the mantle and are products of low degrees of partial melting. The mantle origin is evident from isotope and trace element chemistry. Low degrees of partial melting are indicated by trace element chemistry. However the location of partial melting, which rocks melt, and the nature of and relationship with, parental magmas (if any are involved) are still debated.

## 1.2 Aims and objectives

The hypothesis presented by this work is that carbonatites occur in extensional tectonic settings and therefore carbonatite occurrence can be used to constrain an extensional setting for related occurrences. This work aims to consider the validity of the hypothesis by considering the formation of carbonatites and the occurrence of carbonatites and tectonic history of on the Kaapvaal Craton. As such this work also offers a review of carbonatite occurrence on the Kaapvaal craton

This work is presented in five chapters. The first chapter (this chapter) is an introduction, which presents some basic background of carbonatites.

*Chapter 2: Carbonatite formation* reviews models of carbonatite formation in order to provide insight into carbonatite formation. A petrogenetic model is useful in that it is a framework in which to consider hypotheses pertaining to this work. Chapter two reviews the fractional crystallization, liquid immiscibility and direct partial melting formation models.

*Chapter 3: An extensional setting for the Bushveld Igneous Complex, the carbonatite link* is intended for publication and therefore is referred to as a paper (rather than a chapter) in text and was written with extra supervisor involvement, in his capacity as co-author. Chapter 3 considers the tectonic setting of carbonatites in order to infer the tectonic setting during emplacement of the Bushveld Igneous Complex; via a temporal and spatial association with carbonatite complexes. Chapter 3 therefore includes a review of Bushveld age carbonatites on the Kaapvaal craton.

*Chapter 4: Carbonatites of the Pilanesberg Alkaline Province* is a brief review of the carbonatites included in the Pilanesberg Alkaline Province and considers other geological activity taking place in Southern Africa at that time. Chapter 4 discusses the relevant geological activity in order to test the hypothesis that carbonatites may be used to infer an extensional tectonic setting.

*Chapter 5: Summary* presents abstracts of chapters 2, 3 and 4.

## Chapter 2: Carbonatite Formation

### 2.1 Introduction

When considering carbonatites it is useful to have some insight regarding their genesis. A petrogenetic model provides a framework in which to construct and assess the plausibility of different hypotheses. Such hypotheses may have implications for carbonatites in stratigraphy; expected processes (e.g. enrichment of economically significant materials) and associations with other rock types. For example, Ernst and Bell (2009) note that it is plausible for large igneous provinces (LIP's) and carbonatites to be associated because both are modes of mantle melting.

This chapter will summarise models for carbonatite formation, by reviewing the three noted formation hypotheses. For the sake of presenting models in a coherent manner, models will be presented from a single source, but considered and discussed with reference to and considering other work.

### 2.2 Formation models

Mechanisms suggested for carbonatite melt generation are: primary mantle melting; liquid immiscibility; or, crystal fractionation (Bell *et al.*, 1999; Harmer and Gittins, 1997). .

#### *Fractional Crystallization*

Fractional crystallization occurs in carbonatite melts and is presented as a mechanism for the evolution of alkali-rich compositions, as well as in liquid immiscibility models (e.g.: Lee and Wyllie, 1998). Fractional crystallization would concentrate dissolved carbonates with increasing degrees of fractionation, resulting in carbonatite magma. Some carbonatites are reported to originate from fractional crystallization (e.g. Veksler *et al.*, 1998). However, fractional crystallization is unlikely to produce all carbonatites, because the process will not produce the concentrations of Nb, REE and other incompatible elements characteristic of carbonatites, and the

required concentrations of  $\text{CO}_2$  (for the volume of carbonatites observed) are not present in the proposed parental melts (Gittins, 1989). Another observation that does not support fractional crystallization is the bimodal distribution of Ca and Mg content in carbonatites, which are almost always dominated by either calcite or dolomite; this is referred to as the compositional dichotomy (Bailey, 1993). Fractional crystallization is not consistent with the compositional dichotomy because fractional crystallization would not produce a bimodal composition, but rather a continuum between calcite-carbonatite and dolomite-carbonatite “end-members”. Although not disproven, the hypothesis of carbonatite genesis by fractional crystallization is usually paid little attention, as there is a lack of supporting experimental data.

### *Liquid Immiscibility*

The formation of an immiscible carbonatite liquid from a parental silicate magma was described by Lee and Wyllie (1998) as a mechanism for carbonatite melt generation, based on experimental studies involving the  $\text{CaO-Na}_2\text{O-(MgO + FeO)-(SiO}_2\text{+ Al}_2\text{O}_3\text{)}$  (+  $\text{CO}_2$ ) system. The process involves the silicate-carbonate liquidus field boundary and a miscibility gap between silicates and carbonates. Carbonates are precipitated at the silicate-carbonate liquidus field boundary where carbonate cumulates may form. The parental magma most likely reaches the silicate-carbonate liquidus field boundary by traversing the miscibility gap, but direct crystallization is also possible.

The silicate-carbonate liquidus field boundary is shown as a bold curve on Figure 2.1 at mantle (3 GPa) and crustal (0.5 GPa) pressures in the  $\text{CaO-MgO-SiO}_2\text{-CO}_2$  system, projected from  $\text{CO}_2$ . Figure 2.1 is partly estimated on the basis of experimental data and on earlier work (Wyllie and Huang, 1975, 1976; Wyllie and Lee, 1998). Thermal barriers prevent crystallizing liquids from reaching the silicate-carbonate field boundary and coprecipitating carbonates.  $\text{CO}_2$  and increased pressure removes thermal barriers and allows coprecipitation of dolomitic carbonatites (Lee and Wyllie, 1998).

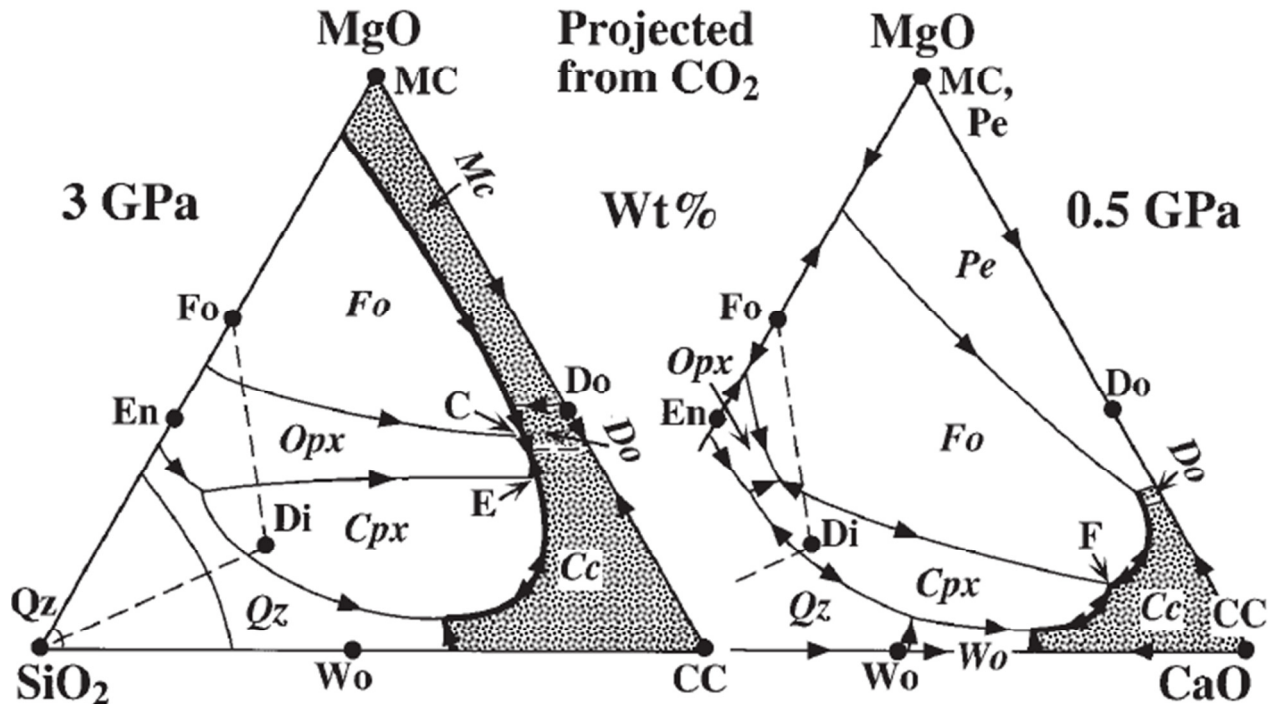


Figure 2.1: For the system  $\text{CaO-MgO-SiO}_2\text{-CO}_2$ , the vapour saturated liquidus surface is projected at mantle and crustal pressures. The bold curve between the carbonate liquidus (stippled) and the silicate liquidus is the silicate-carbonate liquidus field boundary. Dashed lines at low pressure represent thermal divides. Cc= calcite, Do= dolomite, Mc= magnesite, Fo=forsterite, Opx= orthopyroxene, Cpx= clinopyroxene, Qz= quartz, En= enstatite, Di= diopside, Wo= wollastonite, and Pe= periclase. After Lee and Wyllie (1998).

The silicate-carbonate liquid miscibility gap renders carbonatite and silicate magmas immiscible causing calcite-carbonatite magmas to separate from silicate parents. Figure 2.2 shows two Hamilton projections, which are projected compositional slices of complex systems showing liquid paths through multi-component space, rather than ternary diagrams. In Figure 2.2 phase fields are distorted to permit visibility; relative positions of projected field boundaries are a function of bulk composition and pressure, and projections are polythermal. Figure 2.2 shows situations where the miscibility gap field boundary does and does not reach the silicate-carbonate field boundary. Magmas derived from initial silicate- $\text{CO}_2$  liquids may only precipitate carbonates along silicate-carbonate field boundaries (Lee and Wyllie, 1998).

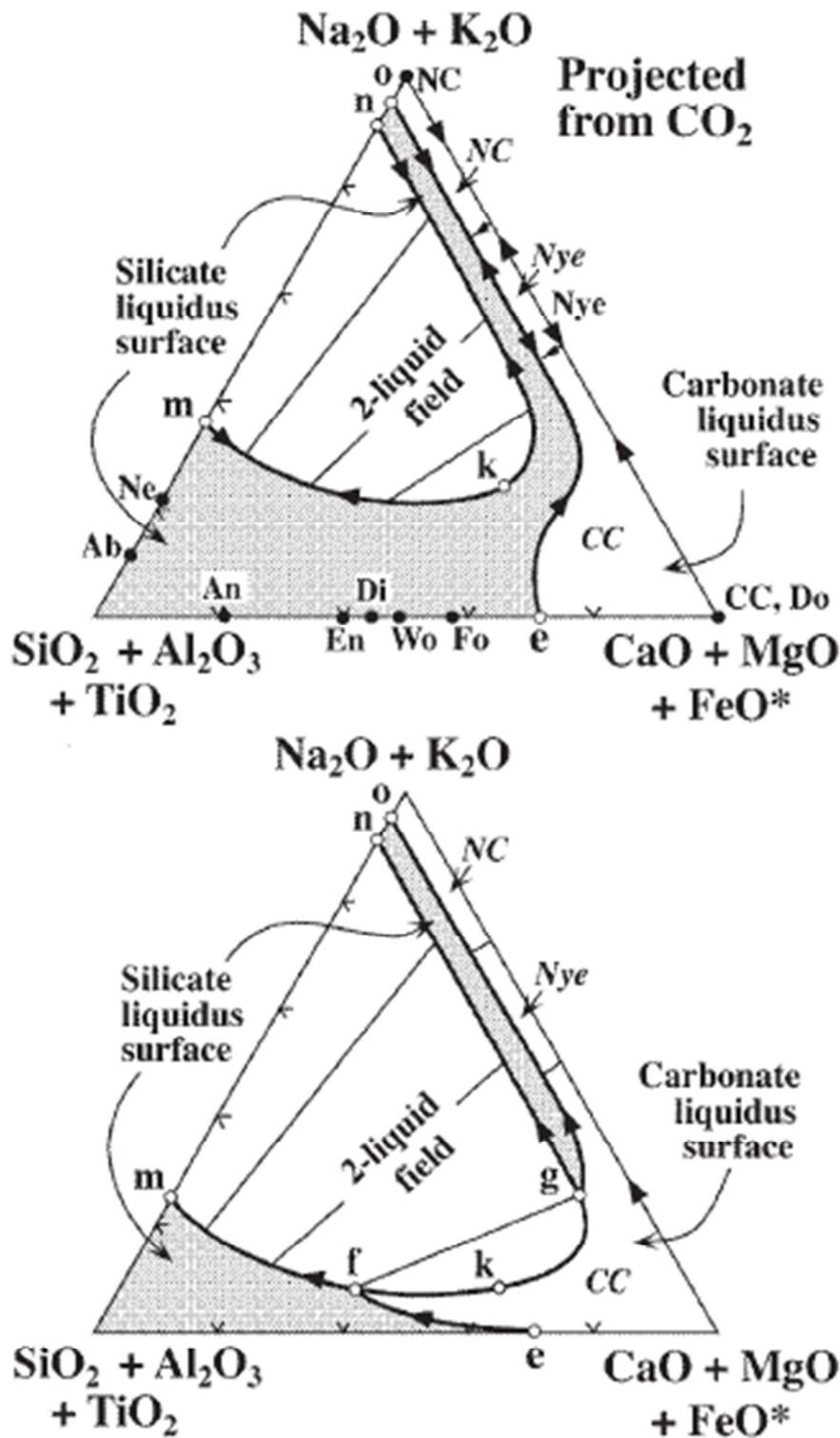


Figure 2.2: Hamilton projections distorted for clarity; showing the silicate-carbonate liquidus field boundary (e-o, e-f and g-o) and miscibility gap field boundary (n-k-m, n-g-k-f-m). FeO\* is total Fe expressed as oxide Cc= calcite, Do= dolomite, Mc= magnesite, Fo=forsterite, Opx= orthopyroxene, Cpx= clinopyroxene, Qz= quartz, En= enstatite, Di= diopside, Wo= wollastonite, Nye= nyerereite, NC= sodium carbonate, Ab= albite, Ne= nepheline, An= anorthite and Pe= periclase. After Lee and Wyllie (1998).



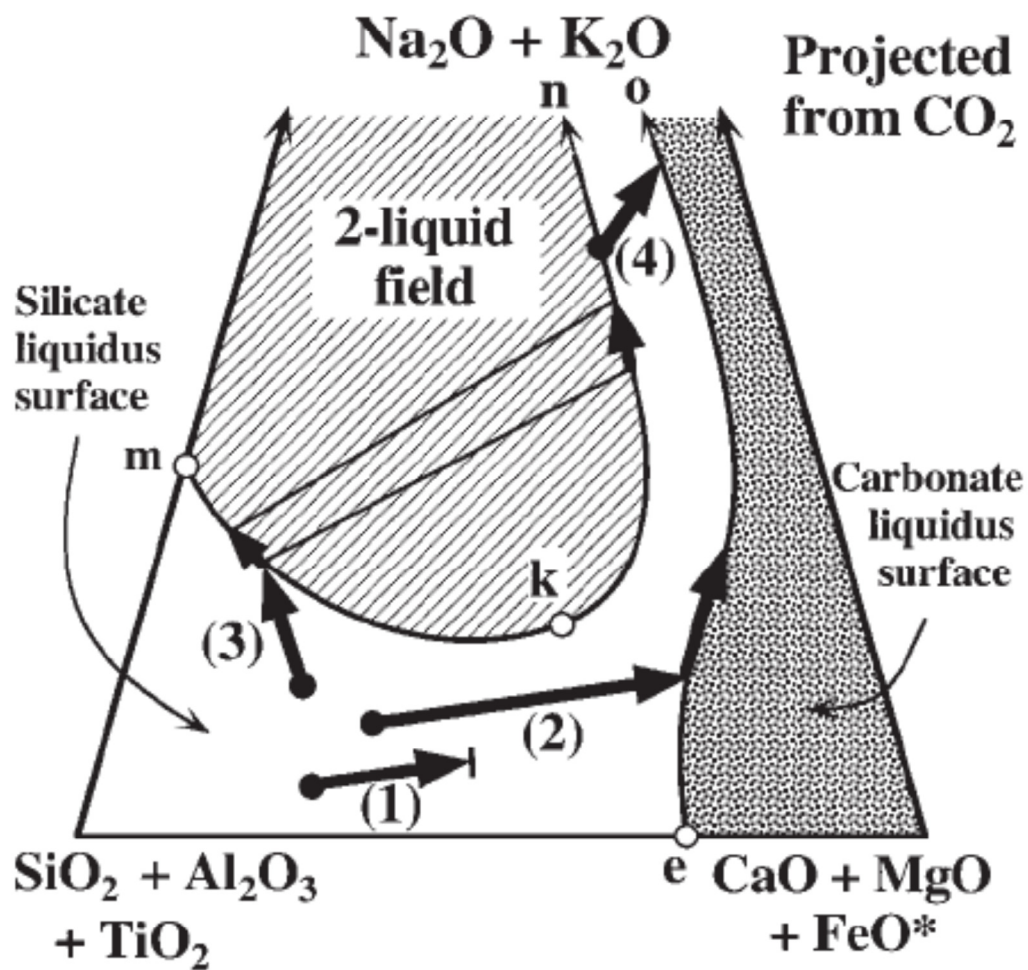


Figure 2.3: A distorted portion of Figure 2.2, vertically exaggerated by two. The figure shows possible crystallization paths in relation to the silicate-carbonatite miscibility gap (shaded) and the carbonate liquidus surface (stippled). Note that, despite similar interpretation this figure is not part of a ternary diagram. After Lee and Wyllie (1998).

Figure 2.3 illustrates phase fields involved in various crystallization paths for initial silicate-CO<sub>2</sub> liquids. Crystallization path 1 terminates and produces silicates and vapour. Path 2 reaches the silicate-carbonate liquidus boundary and carbonate precipitation is accompanied by enrichment in alkalis. Path 3 traverses the miscibility gap, producing two liquids coexisting with silicate minerals. Path 4 is only applicable if carbonate-rich liquid is physically separated from silicate liquid; carbonates are coprecipitated upon reaching the silicate-carbonate field boundary.

Carbonate-rich magma becomes carbonatite magmas only if the silicate-carbonate field boundary is reached, by fractional crystallization (path 2 on Figure 2.3) or liquid immiscibility (paths 3 or 4 on Figure 2.3).

### *Primary Mantle melting*

Gudfinnsson and Presnall (2005) performed melting experiments using a multi-anvil apparatus (calibrated using the transition of quartz to coesite at 3.22 GPa) in the system CaO-MgO-Al<sub>2</sub>O<sub>3</sub>-CO<sub>2</sub> (CMAS-CO<sub>2</sub>) at 3-8 GPa from 1340-1800 °C considering the garnet lherzolite phase assemblage. When in equilibrium with CO<sub>2</sub> bearing melts the CMAS-CO<sub>2</sub> system produces a continuous gradation of melt compositions between carbonatite, kimberlite, melilite, komatiite, picrite and basalt. In general, silica content decreases with temperature and “moving” away from the CO<sub>2</sub>-free lherzolite solidus (Figure 2.5). Along this path komatiitic melts through kimberlitic compositions and finally carbonatite are produced. Dolomite-carbonatite melts are produced by lower temperatures over a range of pressures >2.5 GPa and a small degree of partial melting (<0.5%). Kimberlite is produced at higher temperature, pressure (5-6 GPa, 10 GPa or higher) and degree of partial melting (<1%). Melilite or similar compositions are produced by similar degrees of partial melting to kimberlite, but with pressure <4 GPa and an elevated geotherm. Basalt and picrite are produced at pressures below 4 GPa and near a CO<sub>2</sub>-free lherzolite solidus (Figure 2.4). Isopleths support the experimental finding as oxide contents expected for carbonatite are high where carbonatite melt is stable (Figure 2.5). The CMAS-CO<sub>2</sub> system does not take water into account, which could depress melting temperatures considerably (as much as 150 °C). No experimental run products indicated liquid immiscibility.

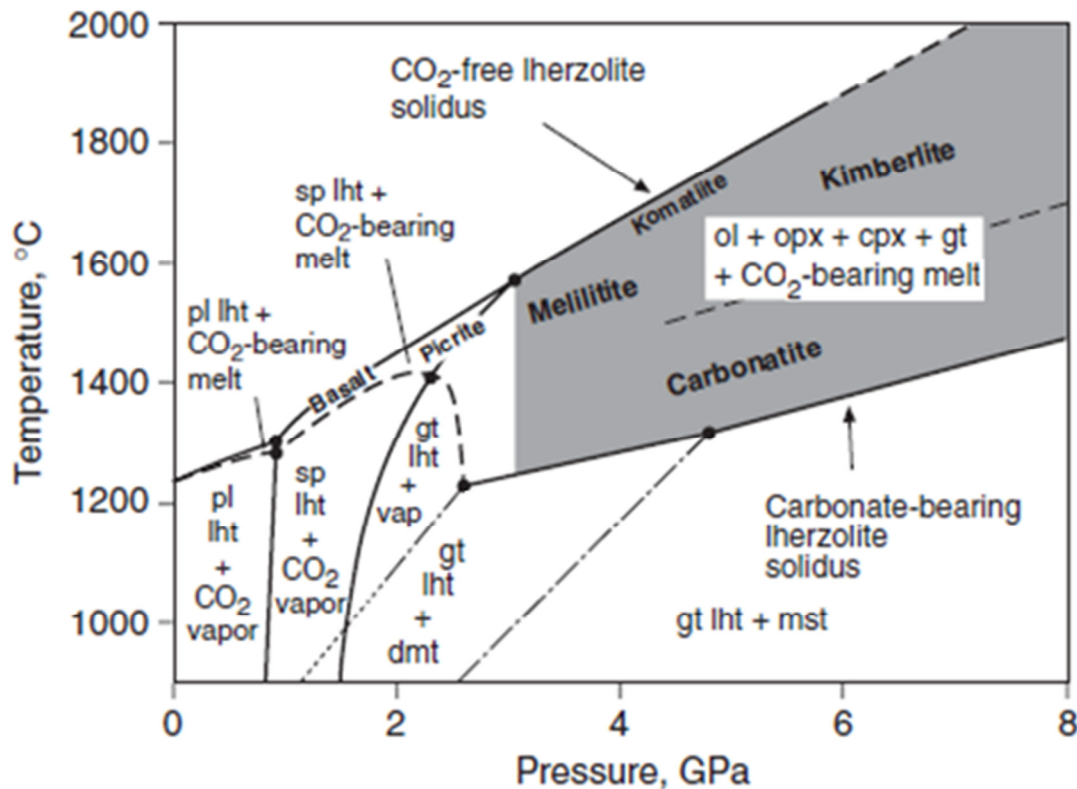


Figure 2.4: Pressure temperature diagram for the CMAS-CO<sub>2</sub> system. Division between carbonatites and kimberlites is placed where 50% modal carbonate can form (based on amount of CO<sub>2</sub> in the liquid). ol= olivine, opx= orthopyroxene, cpx= clinopyroxene, gt= garnet, dmt= dolomite, mst= magnesite, pl= plagioclase, sp= spinel, vap= CO<sub>2</sub>vapor, lht= lherzolite. After Gudfinnsson and Presnall (2005).

Experimental studies show carbonatite magma produced as above will have high Mg# and MgO (Wyllie and Huang, 1975). The model proposed by Harmer and Gittins (1997) explains the emplacement of such carbonatite based on surface expression of the rocks. The primitive carbonatite magma does not mix, assimilate and react with harzburgite and lherzolite because the conduit develops a lining of metasomatic wehrlite or avoiding equilibrium by moving rapidly. If the magma does react with wehrlite it will evolve to a more Ca rich composition (refer to Figure 2.6). Harmer and Gittins (1997) argue that the entire spectrum of carbonatite compositions may be produced by such a process.

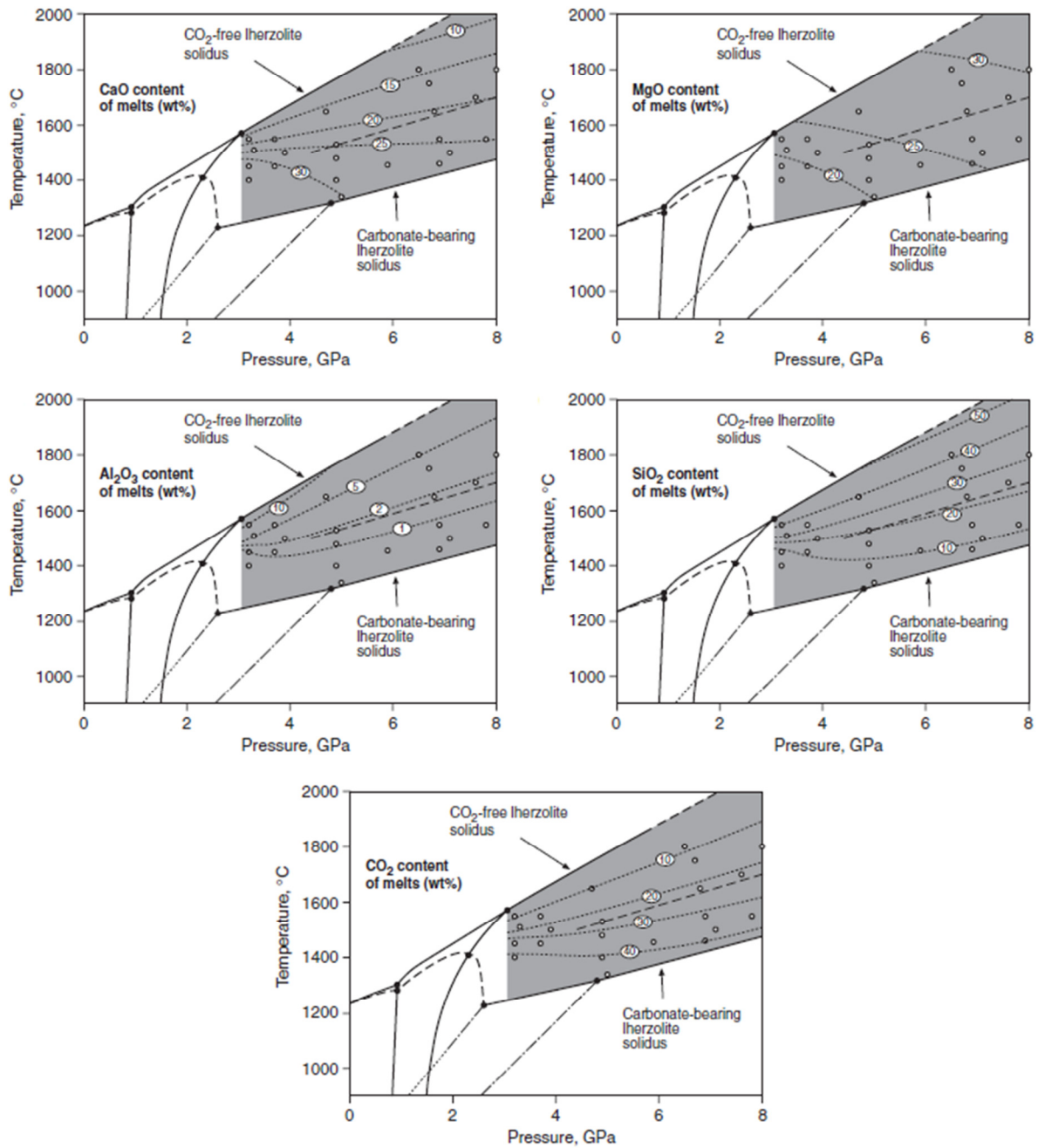


Figure 2.5: Isopleths for oxide contents of CaO, MgO, Al<sub>2</sub>O<sub>3</sub>, SiO<sub>2</sub> and CO<sub>2</sub> in weight percent (wt%) (values shown in ovals). This figure shows variation in composition of CO<sub>2</sub>-bearing melts in equilibrium with garnet lherzolite (CMAS-CO<sub>2</sub> system) for the shaded divariant surface. As would be expected; contents of CaO, MgO, and CO<sub>2</sub> are high and SiO<sub>2</sub> content is low under conditions where carbonatite melts form (refer to Figure 2.4). After Gudfinnsson and Presnall (2005).

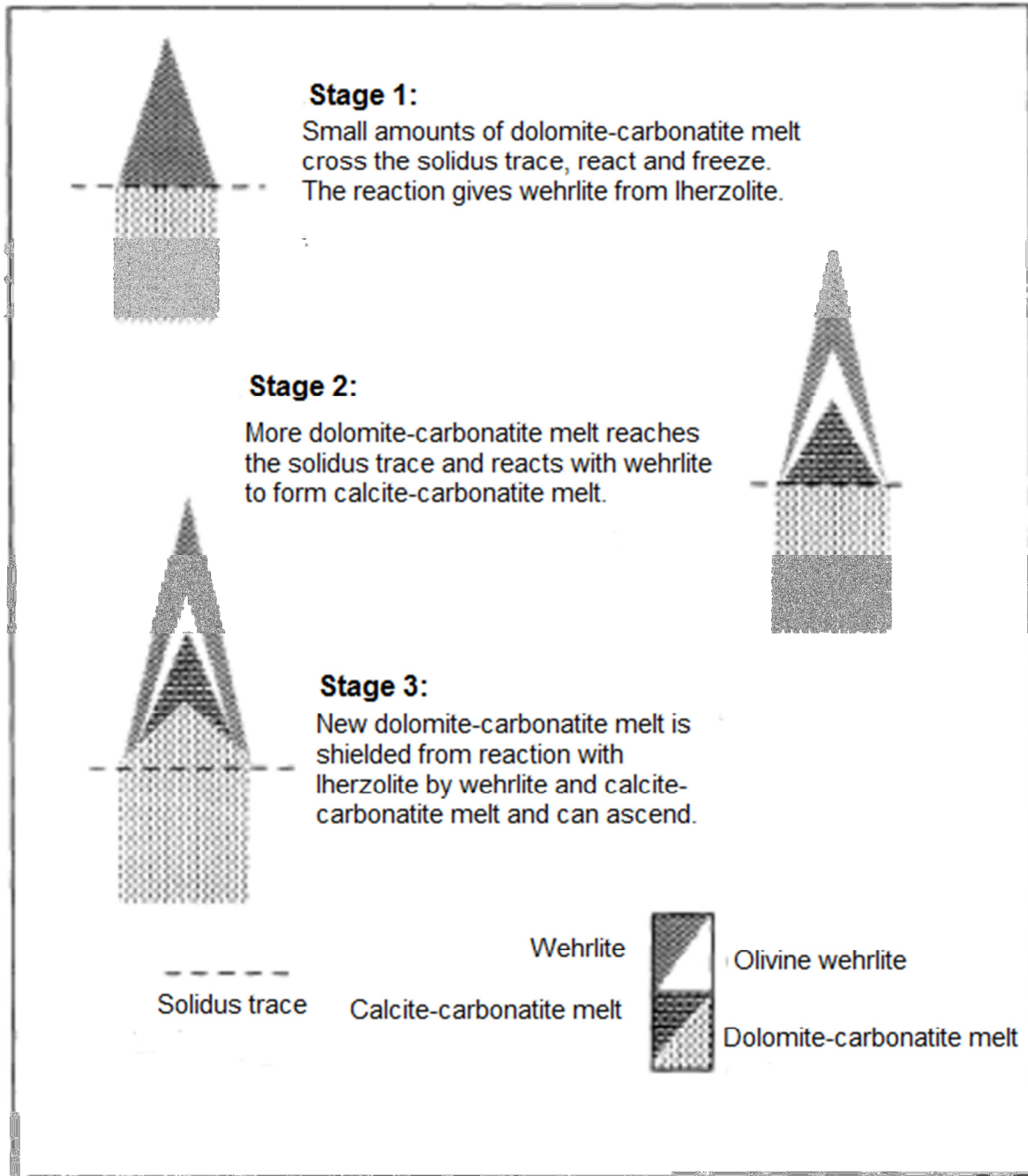


Figure 2.6: Summary of the process described by Harmer and Gittins (1998).

Harmer and Gittins (1997) refer to field, petrographic and compositional data to justify considering dolomite-carbonatite a magmatic rock, rather than a product of replacement occurring within calcite-carbonatite. Experimental studies demonstrating melting of carbonated mantle lherzolite to produce carbonatite liquids all give essentially dolomite-carbonatite melt, which is alkalic (Harmer and Gittins, 1997). Harmer and Gittins (1997) argue that because primary mantle carbonatites should be dolomite-carbonatite, it follows that calcite and dolomite-carbonatites are both

derived from Mg-rich carbonate liquid. With regard to carbonatite melts escaping from the mantle, carbonatite melts are probably not able to rise adiabatically, because high mobility at very low melt volumes makes aggregation into large volumes unlikely and therefore carbonatite melts will most likely ascend along ambient geotherms. In addition carbonatite melts should be consumed by metasomatic reactions that replace enstatitic orthopyroxene with clinopyroxene, olivine and CO<sub>2</sub> fluid. However enstatite is also consumed leaving wehrlite, which can achieve equilibrium with calcite-carbonatite melt. The mixture of calcite-carbonatite melt and olivine wehrlite can then form a conduit by enabling dolomite-carbonatite melt to avoid reacting with lherzolite. In support of this mechanism it is noted that dolomite-carbonatites usually post-date calcite-carbonatites within the same complex.

The problem of compositional dichotomy in carbonatite raised by Bailey (1993) is addressed when considering differentiation of carbonatite melts at crustal levels. Once dolomite-carbonatite magma reaches crustal levels the low pressure would cause a dolomite-carbonatite liquid to crystallize calcite, rather than dolomite (Harmer and Gittins, 1997). Only when additional components which influence the stability of dolomite (volatiles and alkalis) are present is direct crystallization of dolomite possible. In the event that such conditions are not met, the dolomite-carbonatite liquid can produce a cumulate mush of calcite carbonatite, producing a calcite-carbonatite with the associated Mg-rich silicate minerals (Harmer and Gittins, 1997). The implication of the above mechanism is that the bimodal distribution of carbonatites can be produced by variation of factors controlling dolomite stability, rather than requiring different magma types. Harmer and Gittins (1997) make a point of noting that the possibility of calcite-carbonatite liquid is far from excluded and dolomite-carbonatite magma can evolve to a more Ca-rich composition by interacting with upper mantle wall rocks at a depth between 70 km and 40 km (Wyllie and Lee, 1998). In the event that calcite-carbonatite magma reaches the crust, it will precipitate calcite-carbonatite (Wyllie and Lee, 1998) and the compositional dichotomy obviously need not be addressed.

## 2.3 The Source of Carbonatite Magma

Based on isotopic evidence Bell *et al.* (1999) consider parental melts of carbonatite to be derived from asthenosphere-lithosphere interactions. The cause of such interaction is unknown, but may be mantle plumes or lithospheric delamination. More recent work favours a sub-lithospheric location for the generation of carbonatite melt as argued by Bell and Simonetti (2009).

Radiogenic and stable isotope constraints are consistent with a sub-lithospheric source of parental carbonatite melts, produced by asthenospheric upwelling or plume-related processes. Evidence includes: associations with LIPs, primitive noble gas signatures and OIB-like radiogenic isotope ratios (HIMU, EM1, and FOZO components). Despite the sub-lithospheric source, the lithosphere still plays an important role in carbonatite formation, involving concentration of volatiles (Bell and Simonetti, 2009).

From this evidence Bell and Simonetti (2009) propose that carbonatite melt generation involves partial melts derived from and initiated by an asthenospheric source and mantle upwellings respectively. In addition to the observed radiogenic isotope evolution in carbonatite sources, the results from laboratory experiments on the conductivity of the upper mantle are also supportive (Gaillard *et al.* 2008). The similarity of isotopic signatures between carbonatites and OIB implies that both melt types acquired isotopic signatures from the same mantle source during formation. Carbonatites occur on continents because low volume partial melts or fluids derived from mantle upwelling get trapped by overlying lithosphere. Such trapped melts rise through mechanically weak regions, resulting in repeated localised occurrence of carbonatite over billions of years.

The frequency of the LIP-carbonatite association suggests that LIPs and carbonatites may be produced by the same magmatic process or system. Such a link appears plausible because LIPs and carbonatites are both modes of mantle melting and associated with rifting. LIPs and carbonatites may differ with regard to isotope chemistry, indicating a heterogeneous source, differing degrees of partial

melting or different evolution, and always differ with regard to degrees of partial melting required for formation (Ernst and Bell, 2009).

As part of a paper which establishes an association between LIP's and carbonatites by Ernst and Bell (2009), a model for carbonatite origin, consistent with the asthenospheric upwelling or plume origin proposed for many LIP's is proposed. The model involves a mantle plume or asthenospheric upwelling that generates both carbonatite and a LIP. According to Ernst and Bell (2009) carbonatite melts are generated ahead of and at the periphery of a rising heat source, which only causes the low degrees of partial melting required for carbonatites. The model requires carbonatites to precede LIP's and explains the small volume of carbonatite occurrences (Ernst and Bell, 2009).

## **2.4 Discussion**

Carbonatites produced by fractional crystallization may exist, but the hypothesis is not widely applicable as it is in conflict with observations in most cases. Proposing that a particular carbonatite originates by fractional crystallization would require strong evidence and is certainly not currently a popular or widely accepted hypothesis.

Harmer and Gittins (1998) do not exclude the possibility of a model involving liquid immiscibility in the mantle, but experimental data and relevant models only support liquid immiscibility producing carbonatites in the crust (Harmer and Gittins, 1998). Wyllie and Lee (1998) state that the formation of immiscible carbonate rich and silicate magmas in mantle conditions is unlikely due to a large separation of silicate-carbonate immiscibility volume and the liquidus field boundary. Brooker and Kjarsgaard (2011) argue that high CO<sub>2</sub> required at high pressure indicates that the process of liquid immiscibility is unlikely in the mantle, but also demonstrate that silicate-carbonate liquid immiscibility is possible at crustal pressures. Therefore, if carbonatites are produced by liquid immiscibility it will most likely be a process that takes place in the crust.



Harmer and Gittins (1998) cite isotopic evidence to conclude that carbonatites must have been discrete magmas in the mantle. In addition, the inclusion of mantle debris in extrusive calcite-carbonatites, despite low viscosity, indicates liquid originating in the mantle (Harmer and Gittins, 1998). Radiogenic and stable isotope constraints presented by Bell and Simonetti (2009), associations with LIPs, primitive noble gas signatures, and OIB-like radiogenic isotope ratios further support mantle sources for carbonatites. As such, though there may be exceptions, crustal processes like liquid immiscibility are probably not responsible for carbonatite genesis in most cases.

Immiscible carbonate-rich liquids tend to have the composition of calcite carbonatite in simulations (Lee and Wyllie, 1998). As such the liquid immiscibility model explains the usual occurrence of calcite-carbonatites with alkaline silicate complexes (Lee and Wyllie, 1998). The alkaline silicate-carbonatite association is not universal and may represent final magma expulsion from the source region (Bailey, 1993). Therefore dolomite-carbonatites and carbonatites not associated with silicate complexes are not explained by this model and not all carbonatites can be produced by liquid immiscibility. Direct mantle melting, especially if considered with the crustal level magma differentiation process proposed by Harmer and Gittins (1998), is more consistent with observation (e.g. compositional dichotomy and isotope constraints) and explains all but a handful of carbonatites which do not conform to the compositional dichotomy.

The hypothesis of this dissertation works well with the direct mantle melting model. The emplacement of a mantle melt does not rely on material to be subducted or orogeny to form (unlike the emplacement silica-rich melt derived directly from a subducting slab, for example). Magma derived directly from the mantle may require an extensional tectonic setting to escape the mantle and rise through continental crust; therefore a mantle origin favours the hypothesis. Note that the other formation models do not disprove the hypothesis.

It must be noted that it is not reasonable to declare that all carbonatites share a common genesis. Bailey (1993) states that carbonatites have multiple origins and attempts to produce a single hypothesis “may prove futile”. However it is unlikely that different processes have indistinguishable products and as such an attempt to

classify carbonatites based on genesis warrants further research as this will give further insight into the validity of formation models.

## 2.5 Conclusion

Fractional crystallization is disregarded because it is inconsistent with many observations, despite not being considered impossible. Liquid immiscibility is considered plausible, but has some limitation with regard to applicability (only applicable to calcite-carbonatite related to a silicate complex) and leaves some observations unaccounted for (esp. mantle origin and compositional dichotomy). However allowance must be made for any particular carbonatite to have originated by liquid immiscibility.

Direct mantle melting is favoured by this work, due to consistency with observation and wide applicability when considering carbonatites in general. The direct mantle melting model is consistent with the mantle origin and not only explains the compositional dichotomy, but allows for the occurrence of dolomite-carbonatite and carbonatites which are not associated with silicate complexes. However, other models cannot be disregarded entirely with current knowledge.

The favoured model (direct mantle melting) supports the hypothesis of this dissertation as the model is consistent with an extensional tectonic setting.

## Chapter 3:

# An extensional setting for the Bushveld Igneous Complex, the carbonatite link

## 3.1 Introduction

In the global tectonic paradigm, magmatic processes are intimately linked with the tectonic settings in which they occur, a hypothesis that is readily provable in Quaternary rocks where the tectonic setting is clear. In older rocks, and especially in Archaean and Proterozoic terranes that have experienced a long tectonomagmatic history, the tectonic regime in which a specific igneous event occurred may be obscured by deformation or erosion. In this case, the tectonic paradigm may be inverted, and the type and style of magmatism present may be used to infer the tectonic setting.

The focus of this paper is on the possible tectonic regime operating on the Kaapvaal Craton during the Late Proterozoic (2100-2000 Ma). During this period, the world famous Bushveld Complex was emplaced into the Kaapvaal Craton, but several smaller carbonatite complexes were also formed, notably the Phalaborwa carbonatite complex on the eastern edge of the craton. If all these intrusive events are contemporaneous, then it may be possible to infer the tectonic setting for craton at the time.

Wu *et al.* (2011) note that carbonatite complexes are typically found in anorogenic settings and state that no evidence of carbonatite formation in subduction zones is known. Ernst and Bell (2009) support this statement by stating that an association between carbonatites and subduction related environments is rare if such an association exists at all, and note that the Deccan carbonatites and carbonatites in the East African Rift Valley provide examples of a clear carbonatite-rift association. In addition to rifting environments Ernst and Bell (2009) state that carbonatites tend to be opportunistic and can be emplaced along mechanically weak regions in the lithosphere (rifts and other faults) and therefore it may be possible to use carbonatites to mark sites of lithospheric disruption (Ernst and Bell, 2009). Burke *et al.* (2003) propose that deformed alkaline rocks and carbonatites (DARC's) may be

used to identify ancient rift zones after being obscured by convergent tectonic processes. The hypothesis that DARC's are concentrated in suture zones and mark Wilson cycles was tested by Burke and Khan (2006), who provisionally accepted the hypothesis pending a global scale investigation.

If carbonatites are restricted to extensional or transtensional settings, then it will follow that other igneous rocks emplaced at the same time are also emplaced within such a setting. Such a hypothesis may hold for the Bushveld Igneous Complex (BIC), which has a temporal and spatial association with the Phalaborwa carbonatite complex.

### **3.2 Geological background:**

#### *The Kaapvaal Craton and the Transvaal Supergroup*

The Kaapvaal Craton is of Archaean age and due to recent uplift of East Africa is well exposed (esp. in the Northeast), but also well preserved. In addition, isostatic rebound during the Vredefort impact event, around 2 Ga (Moser, 1997), turned a near-complete section of cratonic crust on-edge, and the Vredefort impact structure includes 3.5 to 3.3 Ga old mantle lithologies at the centre of the structure (Tredoux *et al.*, 1999). Mapping and geochronology of the Vredefort impact structure by Moser *et al.* (2001) reveals that the deep crust of the Kaapvaal Craton was last subject to high temperature metamorphism (assumed to be a herald of stabilization) 0.12 billion years after crust assembly at  $3.08 \pm 0.01$  Ga. Moser *et al.* (2001) note that the last recognised episode of voluminous granite magmatism in the Vredefort section took place at 3.09 Ga (maximum age of stabilization) and the age of basal rocks of the Witwatersrand basin gives a minimum stabilization age of 3.07 Ga. Once the Kaapvaal Craton had stabilized; the Witwatersrand and Ventersdorp supergroups were deposited and tectonic extension occurred. Following these events are the Limpopo orogeny and the deposition of the Transvaal Supergroup, which formed the rocks that the igneous complexes relevant to this paper intruded into and reveal the tectonic conditions preceding intrusion.

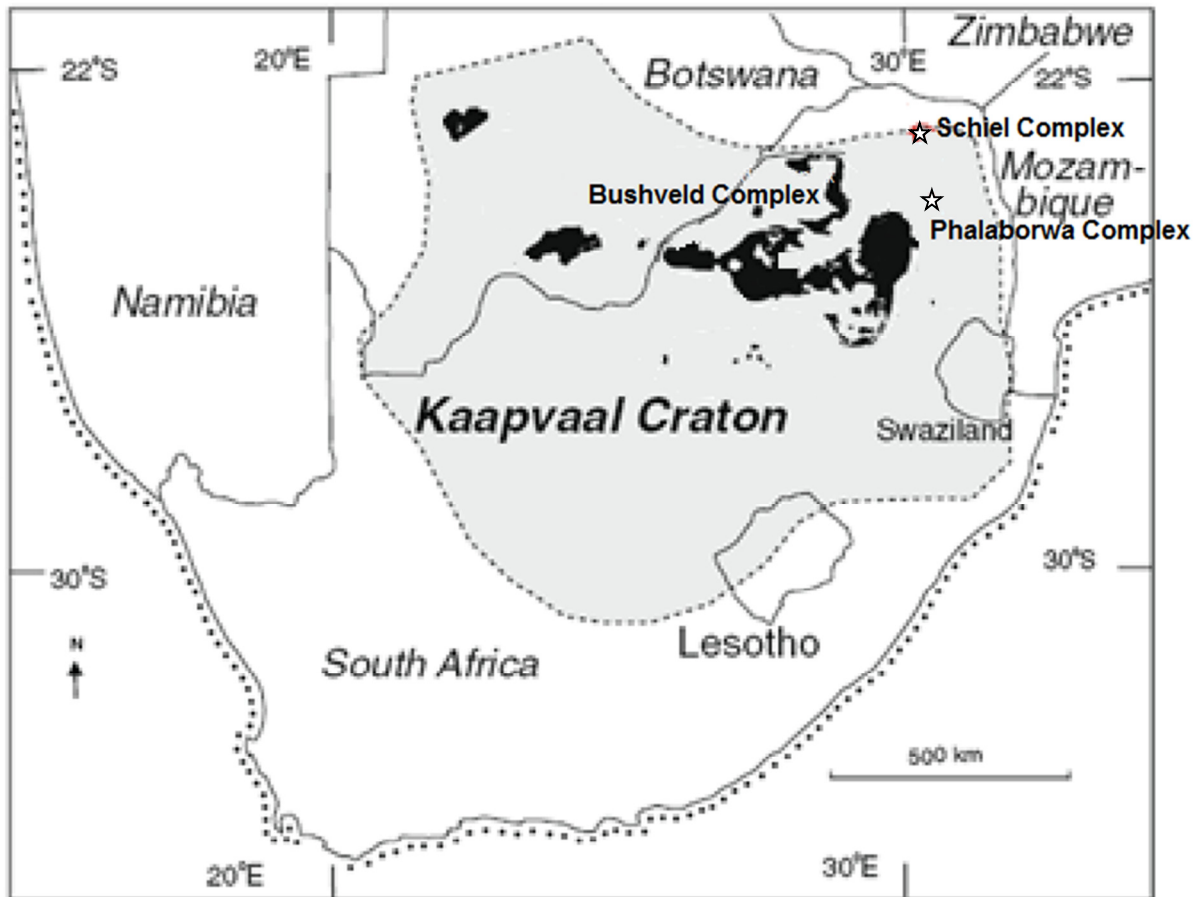


Figure 3.1: Locations of the Schiel, Phalaborwa and Bushveld Complexes in addition to the extent of the Kaapvaal craton. Locations of the Phalaborwa and Schiel complexes are marked by small stars. After Ernst and Bell (2009).

The Limpopo Belt is located between the Kaapvaal and Zimbabwe cratons and was formed because of the two cratons colliding, resulting in the Kalahari Craton. This high grade metamorphic province is separated by major shear zones into three distinct domains; the Northern Marginal Zone (NMZ), Limpopo Central Zone (LCZ) and the Southern Marginal Zone (SMZ). The SMZ which is adjacent to the Kaapvaal Craton consists mainly of gneisses and experienced a single metamorphic event at 2.72-2.65 Ga (Khoza *et al.*, 2013). Kreissig *et al.* (2000) showed that the isotopic and trace element characteristics of the Kaapvaal Craton and SMZ resemble each other and overlap. Such a resemblance suggests that the SMZ (granulite facies) is a high grade equivalent of the Kaapvaal Craton (greenschist to amphibolite facies).

The Transvaal Supergroup is a predominantly sedimentary sequence, deposited from the late Archaean to the early Proterozoic (Eriksson *et al.*, 2006). The

sedimentary sequence is well preserved considering its age and is often studied in order to understand Archaean processes. The Transvaal Supergroup overlies the Archaean basement and the Witwatersrand and Ventersdorp Supergroups (Eriksson *et al*, 2006). The Transvaal Supergroup is preserved in three basins on the Kaapvaal Craton: the Transvaal basin is located in northern South Africa; the Griqualand West basin is located in central South Africa; finally the Kanye Basin is in southern Botswana. The three basins have similar stratigraphy, correlating best for the lower chemical sediments and worst for the upper clastic and volcanic deposits. The stratigraphies of the Transvaal Supergroup dip towards the Bushveld Igneous Complex in most areas (Eriksson *et al*, 2006).

Sedimentation began with the Black Reef and Vryburg formations, which were emplaced mostly in fluvial settings with sea level rising relative to the craton. After deposition of the Black Reef and Vryburg Formation, thermal subsidence occurred on the Kaapvaal Craton, which is reflected by the deposition of chemical sediments in marine settings. Chemical sediments make up the Chuniespoort, Ghaap and Taupnoe Groups. Erosion resumed and clastic sediments were deposited in three sedimentary basins which deposited the Pretoria, Postmansburg and Segwagwa Groups. The mostly volcanic Rooiberg Group and Loskop Formation are the uppermost succession in the Transvaal basin (Eriksson *et al.*, 1995).

The Black Reef Formation consists predominantly of quartz arenites, which are accompanied by conglomerate (some of which is basal) and mudrock. The Vryburg formation consists of basal conglomerate, quartzites, shales, subordinate carbonates and in some places basaltic to andesitic lavas. The chemical sediments of the Chuniespoort, Ghaap and Taupnoe Groups are mostly marine shelf carbonates and banded iron formations (BIF). The Pretoria Group consists mostly of mudrocks alternating with quartzitic sandstones. Interbedded basaltic-andesitic lavas, conglomerates, diamictites and carbonates make up the remainder of the Pretoria Group. All rocks in the Pretoria Group have been subjected to low grade metamorphism (Eriksson *et al*, 2006).

With regard to economic significance, the Transvaal Supergroup hosts a range of deposits in various lithologies. Placer gold has been mined from the Black Reef

formation (e.g., Söhnge, 1986). In addition hydrothermal gold has been mined in Mpumalanga (province in eastern South Africa; Meyer, 1988). Metallurgical dolomite is mined in the Pretoria and Potchefstroom areas (Eriksson *et al.*, 2006). Mississippi Valley deposits are known and likely formed during a weathering period directly preceding deposition of later clastic sediments (Martini *et al.*, 1995). BIF deposits have been exploited for iron (Eriksson *et al.*, 2006). Significant manganese fields are located in the Northern Cape (province in western South Africa), (Eriksson *et al.*, 2006). Metamorphic ores include asbestos in BIF (Beukes and Dreyer, 1986) and andalusite in the aureole of the Bushveld Complex (Hammerbeck, 1986).

### *The Bushveld Complex*

The Bushveld Complex is located in northern South Africa, on the Kaapvaal Craton, and has been studied extensively. The Bushveld Complex includes the Rooiberg Group (basalt-rhyolite lavas), the Lebowa Granite Suite, the Rashoop Granophyre Suite and the Rustenburg Layered Suite (often referred to as the mafic component). The granitic rocks form the central roof of the Bushveld Complex and are surrounded by lobes of mafic rocks. The mafic rocks are part of the Rustenburg Layered Suite (Eriksson *et al.*, 1995).

The Rooiberg Group is divided into four formations, with the Dullstroom being the lowermost formation, followed by the Damwal and Kwasgasnek formations, with the Schrikloof being the uppermost formation (Schweitzer *et al.* 1995). The Rooiberg Group is associated with and is considered the uppermost unit the Transvaal Supergroup (i.e. the lava flows terminated sedimentation), but it also petrogenetically linked to and preceded emplacement of the BIC (Eriksson *et al.*, 1995). The Rooiberg group consists of basaltic to rhyolitic lava flows, minor pyroclastic rocks and subordinate sedimentary interbeds (Lenhardt and Eriksson, 2012). Note that Lenhardt and Eriksson (2012) present a thorough review of the Rooiberg Group.

The Rustenburg Layered Suite is 7-9 km thick and consists mostly of a variety of rocks in layers including pyroxenites, norite, gabbro, anorthosite, magnetite and olivine diorite. In addition there are marker reefs present including the Merensky reef, Thornhill pyroxenite layers, magnetite layers and chromitite layers and stringers. The Rustenburg Layerd suite is divided into the Marginal (norite), Lower (pyroxenite),

Critical (pyroxenite, norite and chromitites), Main (Norite, gabbronorite and anorthosites) and Upper Zones (gabbronorite, diorite and magnetite) (Eales and Cawthorn, 1996). Occurrences of pipe like bodies within the BIC (esp. within the mafic component) have been reported (e.g. Cawthorn *et al.*, 2000).

Transvaal Supergroup inliers occur within the BIC, and are divided into two categories. The attached structures, consisting of lower Transvaal strata, include Crocodile River, Marble Hall and Dennilton domes and were likely formed by interference folding. The attached structures are characterized by low to medium grade metamorphism and are highly deformed. The detached structures, consisting of upper Transvaal strata, including the Rooiberg and Stavoren fragments, are less deformed than the other group of inliers and show only low grade metamorphism. In addition the highly metamorphosed Transvaal xenoliths like the Lezwete and Parys structures are also considered inliers (Hartzer, 1995).

Interest in the Bushveld Complex is generated by unusual characteristics (esp. the size and layering) and economic potential. The economic potential includes deposits associated with three different lithologies. The Rustenburg Layered Suite, the mafic component of the Complex hosts iron, nickel, chromium, vanadium and major occurrences of PGE ore, and the roof rocks host tin and fluorine and the metamorphic aureole hosts andalusite (Eriksson *et al.*, 1995). However, despite the volume of information gathered on the economic potential of the Bushveld Complex, the origin of the complex remains controversial.

Earlier models for the origin of the BIC involved a bolide impact have been largely discounted (e.g. Hatton 1995). Back arc rifting during the collision between the Zimbabwe and Kaapvaal cratons has been considered, but does not explain the heat source required to produce the volume of magma required (Ernst and Bell, 2009) and the Limpopo orogeny predates the emplacement of the BIC. Considering the heat required, a plume beneath the Kaapvaal Craton has been proposed (Hatton, 1995). Clarke *et al.* (2009) present relevant structural and compositional data and conclude that the BIC is a sill-like intrusion with a feeder dyke which utilised the Thabazimbi-Murchison Lineament.



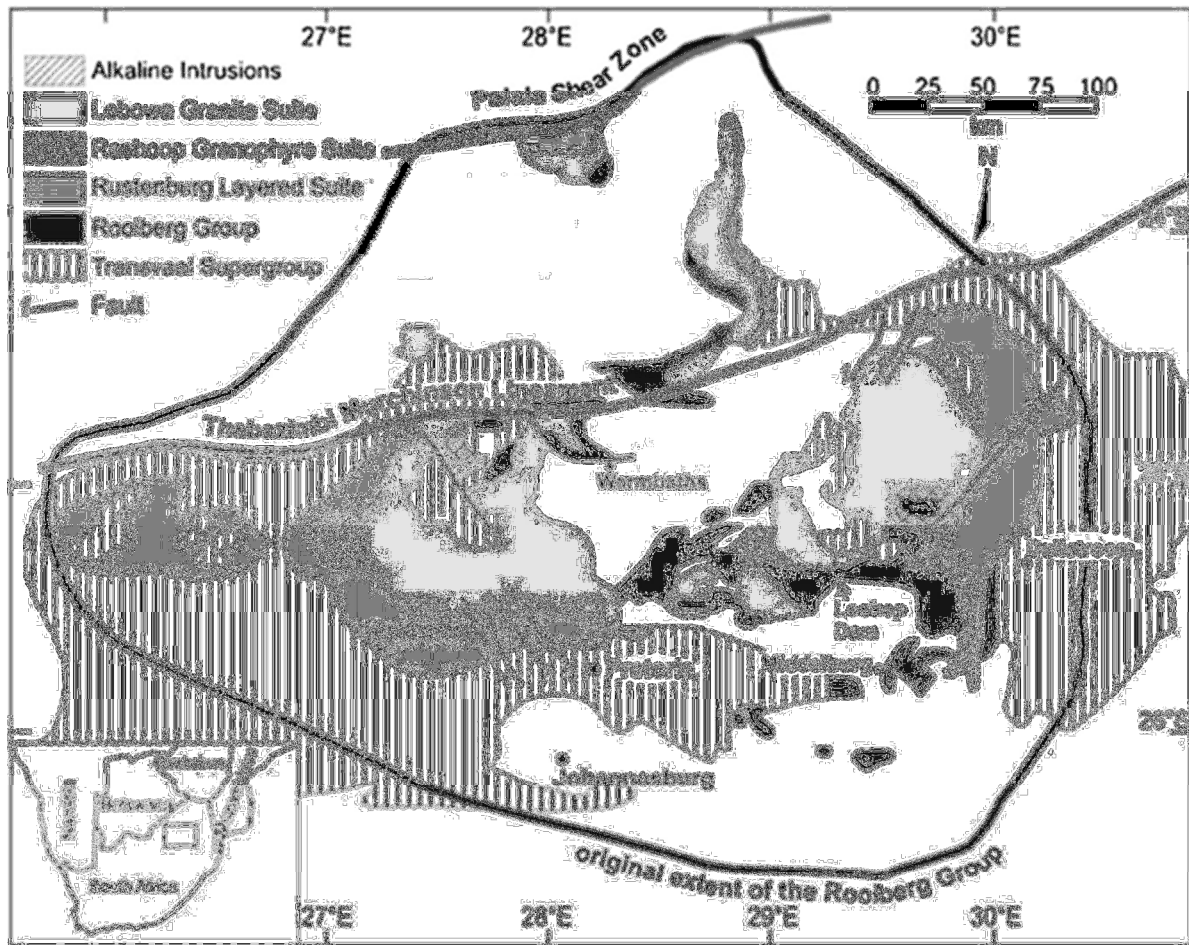


Figure 3.2: A simplified geological map of the Bushveld Igneous Complex; showing the occurrence of the relevant groups and suites. In addition an estimated original extent of the Rooiberg Group is shown to illustrate the size of the complex that has been obscured by erosion and overlying rock. After Lenhardt and Eriksson (2012).

### *The Phalaborwa Carbonatite Complex*

The Phalaborwa Complex is located in the Limpopo Province (formerly North Transvaal) of South Africa, on the Kaapvaal Craton. The Phalaborwa Complex is characterised by Cu and Zr mineralisation and includes mainly pyroxenites, phoscorite and carbonatite. Unusually for carbonatite, the Phalaborwa Complex displays high initial Sr and low initial Nd and Hf isotopic compositions. The complex is associated with satellite syenite and minor granites (Wu *et al.*, 2011; Ernst and Bell, 2009).

The Phalaborwa Complex has been studied extensively, due to economic interest. Such interest is focused on copper, phosphate (from apatite) and vermiculite (phlogopite altered by weathering processes) as well as the secondary products of magnetite, zirconia, uranium, nickel, gold and PGE's (Verwoerd and du Toit, 2006). Carbonatite has been mined underground for copper by Phalaborwa Mining Company since the open pit was closed in 2002. Foskor operates two open pits from which phoscorite and pyroxenite are mined for phosphate and vermiculite respectively.

A vast amount of work has been done on the Phalaborwa Complex. Reports by mining company staff; the review by Verwoerd and du Toit (2006) and petrology, geochemistry isotopic work reported by Eriksson (1989) are notable. The shape of the Phalaborwa Complex (excluding satellite granites and syenites) reflects three coalesced intrusions (Verwoerd and du Toit, 2006), intruded over a short time span (Eriksson, 1989). The northern lobe has a centre of serpentine and vermiculite (altered olivine and phlogopite) pegmatiod (described by Wu *et al.* (2011) as pegmatitic pyroxenite as in figure 3.3). The centre of the southern lobe is similar, but rich in pyroxene and apatite. The copper-bearing Loolekop carbonatite pipe and surrounding phoscorite is located in the centre of the Phalaborwa Complex. The other 70% of the Phalaborwa Complex is micaceous clinopyroxenite, with variable phlogopite and apatite contents. The pyroxenite is surrounded by a narrow marginal zone of feldspathic pyroxenite and some fenitization of country rock (granite-gneiss) has been observed (Verwoerd and du Toit, 2006).

The trace element geochemistry of carbonatite and clinopyroxenite in the Phalaborwa Complex (low Nb, but high Sr and light rare earth elements (LREE)) is a result of cumulus apatite. Zr concentration is high in phoscorite, but not silicate rocks. Differentiation of silicate rocks and carbonatite could not have been due to crystal fractionation or liquid immiscibility, because initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios are too different. By process of elimination the direct mantle melting model explains the genesis of carbonatite at Phalaborwa. Other isotopic systems indicate magma from mantle sources mixing as the origin of the Phalaborwa Complex (Eriksson, 1989).

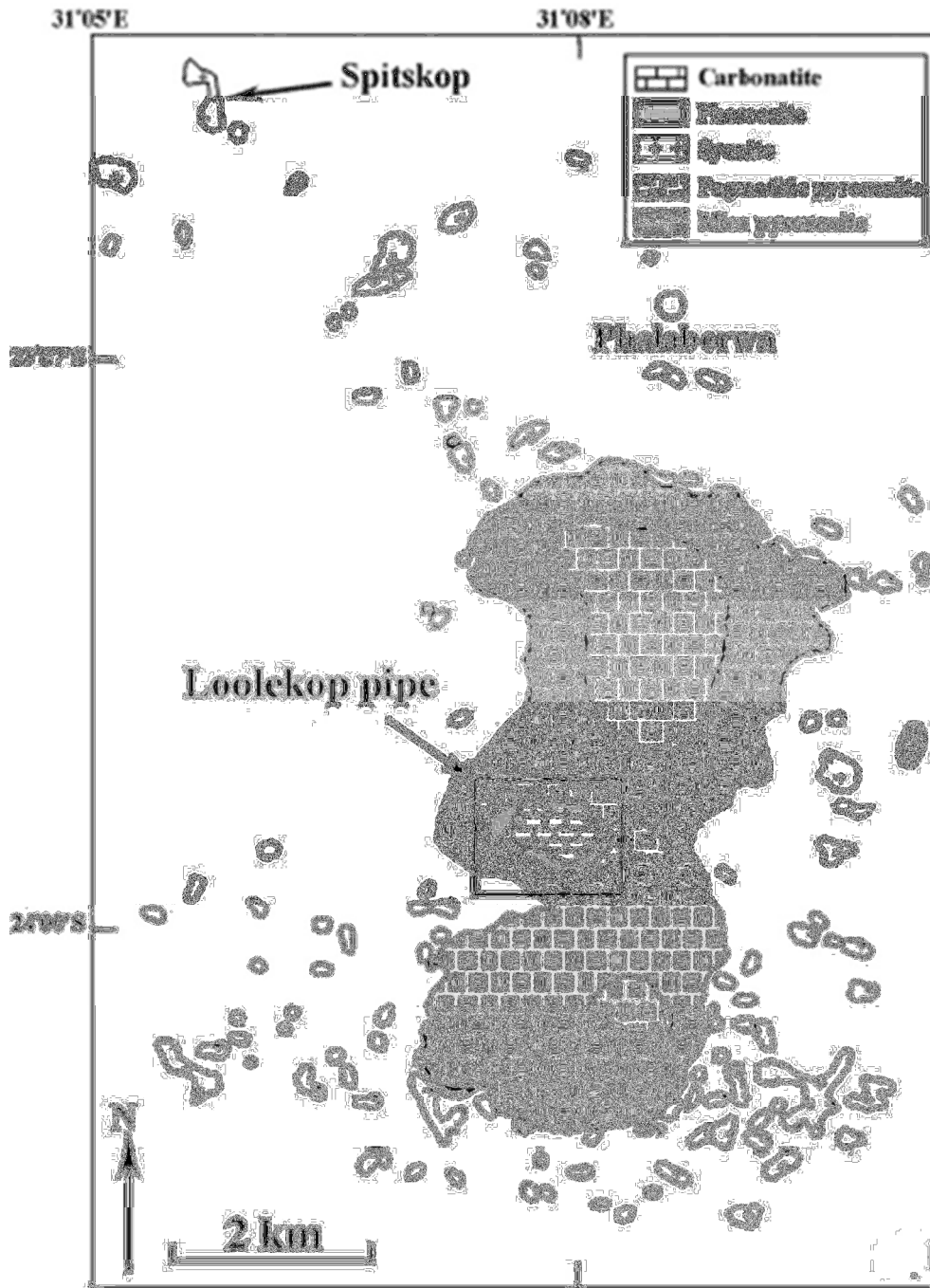


Figure 3.3: Generalised geology of the Phalaborwa complex. After Wu *et al.* (2011).

### *The Schiel Complex*

The Schiel complex has been associated with the Phalaborwa Complex due to petrological similarities and dating by Walraven *et al.* (1992) (Verwoerd and du Toit, 2006); therefore the Schiel Complex is included in this paper. However the age of the Schiel complex is uncertain as conflicting ages have been presented by Walraven *et al.* (1992) and Barton *et al.* (1996). Fortunately the Schiel Complex is not “required” for the purpose of this paper and is included only for completeness sake.

The Schiel Complex is a large alkaline plutonic intrusion in the Limpopo Province of South Africa, consisting of two units (eastern and western parts) (Stettler *et al.*, 1993). The Schiel complex was emplaced into the Southern Marginal Zone (SMZ) of the Limpopo folded belt and is related to the evolution of the metamorphic rocks, but was emplaced in an extensional period after orogeny (Lubala *et al.*, 1994). The Schiel Complex includes pyroxenite, carbonatite, dolerite, syeno-granite, syenite and granite. The eastern unit appears to have an annular structure emplaced during at least two episodes at low pressure (shallow depth). The western unit appears to be a sheet-like intrusion (Stettler *et al.*, 1993).

Despite geochemical similarities, the Schiel complex does not share the economic potential of the Phalaborwa Complex. The economic potential of Schiel was investigated by Foskor and reported by Viljoen (1966). Viljoen (1966) reported phosphate ore, but found vermiculite, magnetite, copper ore and syenite not to be of commercial or minable quality. As a result the Schiel Complex has not been studied as extensively as the Phalaborwa Complex. A review of the economic potential was published by Verwoerd (1986).

After the investigation reported by Viljoen, Du Toit (1979) mapped an area including the Schiel Complex. More detailed work has been done by Rodgers (1989) and Stettler *et al.* (1993) interpreted geophysical data of the Schiel Complex. The Schiel Complex was dated by Walraven (1992) and Barton *et al.* (1996).

The Schiel Complex is situated near Louis Trichardt, on the Kaapvaal Craton, and is situated directly north of the Kudus River Lineament. The Schiel Complex is not

metamorphosed nor deformed (Verwoerd and du Toit, 2006). Verwoerd and du Toit (2006) explain that the eastern part of the complex was emplaced in four intrusive phases. The first phase emplaced rocks similar to the apatite bearing rocks in the Phalaborwa Complex. These rocks include a centre of carbonatite and pyroxenite, glimmerite, phoscorite and later veins of carbonatite. The second intrusive phase emplaced a plug-like syenogabbro body. The third intrusive phase represents most of the complex in plan and emplaced quartz syenite. The fourth intrusive phase is represented by pyroxene and amphibole bearing granite (Verwoerd and du Toit, 2006).

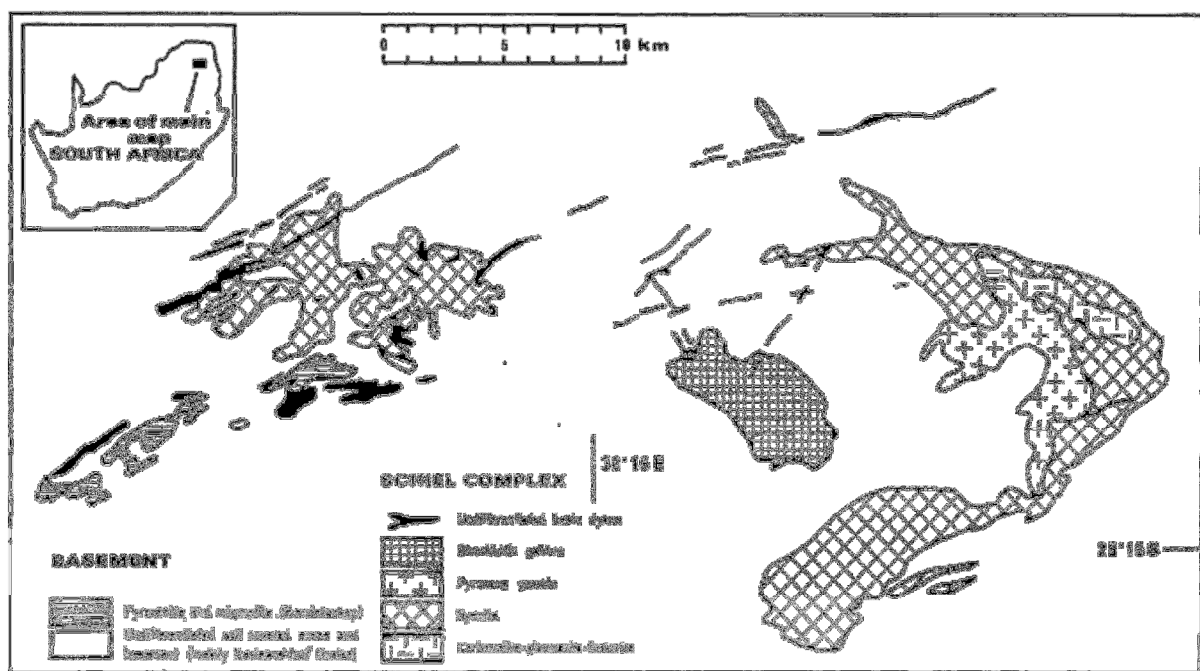


Figure 3.4: Generalized geology and location of the Schiel Complex. After Lubala *et al.* (1994).

### 3.3 Geochronology of Kaapvaal igneous occurrences

Table 1 presents previously determined ages for the Kaapvaal occurrences. Eriksson (1989) used Pb and Rb-Sr to date the Phalaborwa complex. Ages of 2060 Ma and 2047 +11/-8 Ma with Pb and 2012 ± 19 Ma with Rb-Sr were determined. The age of emplacement (2060.6 ± 0.5 Ma) was determined for the Phalaborwa complex using U-Pb on a baddelyite crystal (Reischmann, 1995). Wu *et al.* (2011) used U-Pb to date zircon and baddelyite in the Phalaborwa complex in situ. Wu *et al.* (2011) found an age of 2060 ± 4 Ma for the outer pegmatitic pyroxenite (at the Loolekop

pipe), an age of  $2062 \pm 2$  Ma for the main phoscorite and the ages for transgressive and banded carbonatite to be  $2060 \pm 1$  Ma and  $2060 \pm 2$  Ma respectively.

Walraven *et al.* (1992) presented Pb-isotopic data from various lithologies which indicated an age of  $2059 + 35 -36$  Ma for the Schiel Alkaline Complex. Barton *et al.* (1996) favour an age of  $\sim 1850$  Ma for the Schiel Complex. Barton *et al.* (1996) used U-Pb ion microprobe data for zircon in syenitic phases ( $\sim 1850$  Ma). Rb-Sr and Pb-Pb whole rock data ( $\sim 2100$  Ma) and Sm-Nd whole-rock isotopic data ( $\sim 2300$  Ma) were also reported by Barton *et al.* (1996). However the latter two ages are not trusted by the authors because the ages may represent Pb mixing, rather than the true age of the complex. Similarly, the 2300 Ma age is interpreted to indicate a LREE enriched source with different initial Pb ratios.

Table 1: Summary of dating referenced for the Bushveld, Phalaborwa and Schiel complexes. Ages marked with a \* are originally given in Ga.

Complex	Lithology	Mineral	Technique	Reference	Age
Phalaborwa			Pb	Eriksson (1989)	$2047 +11/-8$ Ma
Phalaborwa			Rb-Sr	Eriksson (1989)	$2012 \pm 19$ Ma
Phalaborwa		baddelyite crystal	U-Pb	Reischmann (1995)	$2060.6 \pm 0.5$ Ma
Phalaborwa	pegmatitic pyroxenite	zircon and baddelyite	U-Pb	Wu <i>et al.</i> (2011)	$2060 \pm 4$ Ma
Phalaborwa	Phoscorite	zircon and baddelyite	U-Pb	Wu <i>et al.</i> (2011)	$2062 \pm 2$ Ma
Phalaborwa	Carbonatite	zircon and baddelyite	U-Pb	Wu <i>et al.</i> (2011)	$2060 \pm 1$ Ma
Phalaborwa	banded carbonatite	zircon and baddelyite	U-Pb	Wu <i>et al.</i> (2011)	$2060 \pm 2$ Ma
Schiel	various lithologies		Pb-isotopic	Walraven <i>et al.</i> (1992)	$2059 + 35/-36$ Ma
Schiel	Syenite	Zircon	U-Pb	Barton <i>et al.</i> (1996)	$\sim 1850$ Ma *
Schiel			Rb-Sr	Barton <i>et al.</i> (1996)	$\sim 2100$ Ma*
Schiel			Pb-Pb	Barton <i>et al.</i> (1996)	$\sim 2100$ Ma*
Schiel			Sm-Nd	Barton <i>et al.</i> (1996)	$\sim 2300$ Ma*
Bushveld	retrogressed xenolith	Titanite	U-Pb	Buick <i>et al.</i> (2001)	$2058.9 \pm 0.8$ Ma
Bushveld	Merensky Reef	Zircon		Scoates and Friedman (2008)	$2054.4 \pm 1.3$ Ma
Bushveld	Merensky Reef	Rutile		Scoates and Friedman (2008)	$2055.0 \pm 3.9$ Ma
Bushveld	Merensky Reef	Zircon	U-Pb	Yudovskaya <i>et al.</i> (2013)	$2051 \pm 9$ Ma
Bushveld	Platreef	Zircon	U-Pb	Yudovskaya <i>et al.</i> (2013)	$2056 \pm 5$ Ma

The Bushveld Complex was emplaced at 2060 Ma, according to Walraven *et al.* (1990). Buick *et al.* (2001) used U-Pb isotope data from newly grown titanite in a completely retrogressed xenolith. Buick *et al.* (2001) found that retrogression took place at  $2058.9 \pm 0.8$  Ma which places a constraint on the age of the Rustenburg Layered Suite of the Bushveld Complex. Scoates and Friedman (2008) separated minerals from a sample obtained from the Merensky reef (of the Bushveld Complex) and found a U-Pb crystallization age of  $2054.4 \pm 1.3$  Ma for zircon and a U-Pb cooling age of  $2055.0 \pm 3.9$  Ma for rutile. Scoates and Friedman (2008) used single-crystal chemical abrasion ID-TIMS for zircon and multigrain fractions of rutile to produce data. Integration of their results with previous work led Scoates and Friedman (2008) to suggest that Bushveld Complex magmatism occurred at ca. 2054 Ma. Yudovskaya *et al.* (2013) used U-Pb SHRIMP dating of zircons from the Merensky Reef and Platreef (located near the top of the Critical Zone in the Bushveld Complex). Yudovskaya *et al.* (2013) found ages of  $2051 \pm 9$  Ma and  $2056 \pm 5$  Ma for the Merensky Reef and Platreef respectively.

### 3.4 Discussion

Ernst and Bell (2009) link carbonatites to LIP's including: the Deccan flood basalt province, Paraná-Etendeka, the Siberian LIP, widespread basaltic magmatism in the East European Craton, the Central Iapetus Magmatic Province, the Keweenaw LIP and the pan-Superior mafic-ultramafic magmatism. On reviewing the spatial and temporal links between carbonatites and LIP's, Ernst and Bell (2009) determine that the "LIP-carbonatite association" is as robust as the association between carbonatites and rifts. Ernst and Bell (2009) note in passing that the BIC can be regarded as the intrusive equivalent of flood basalt, and that there was a spatial and temporal link with Phalaborwa. As such, the LIP-carbonatite link may apply to the Bushveld.

#### *The spatial and temporal links between the Kaapvaal intrusions*

The age of ~2060 Ma is readily acceptable for the Phalaborwa Complex, as multiple authors using multiple labs, samples and minerals produce consistent results. In addition Nebel *et al.* (2011) suggest strongly that the U-Pb system in baddelyite has been not significantly compromised since the emplacement of carbonatite.

On the other hand, the little information available gives two conflicting ages of the Schiel Complex. The older age of 2059 + 35 -36 Ma includes the Schiel Complex in a group of widespread magmatic occurrences in the Kaapvaal Craton at that time, and overlaps the Bushveld Complex and the Phalaborwa Complex (Walraven *et al.*, 1992). Considering that the Phalaborwa complex was emplaced at that time it is plausible that the Schiel Complex (or any other carbonatite or alkaline complex) was emplaced then. If the age of 2059 + 35 -36 Ma is accurate the Schiel Complex supports the temporal association of carbonatite and the Bushveld Complex. The younger age of ~1850 Ma (Barton *et al.*, 1996) does not support a temporal association. The latter age may relate to later events in the evolution of the craton and cannot be dismissed; it may be related to rifting (Barton *et al.*, 1996) and may indicate another magmatic event at that time. Resolving the conflicting ages of the Schiel Complex is beyond the scope of this work and therefore the Schiel Complex will be considered to be only “possibly” related to the Phalaborwa and Bushveld Complexes.

Various ages are given for the Bushveld Complex, between 2055 Ma and 2060 Ma. Considering the size and complexity of the Bushveld Complex this is understandable. Fortunately this paper does not require an exact age, because the age of the Bushveld Complex can be considered similar (if slightly younger) to the age of the Phalaborwa complex. Wu *et al.*, (2011) similarly state that the Bushveld Complex is slightly younger than the Phalaborwa Complex at ~2055 Ma. The Schiel Complex may also be associated temporally with Phalaborwa and the Bushveld Complex (Walraven *et al.*, 1992; Ernst and Bell, 2009) and occurs on the same craton (refer to figure 2.1).

#### *Tectonic activity on the Kaapvaal Craton around 2060 Ma.*

The Limpopo orogeny shaped the craton and formed some of the rocks that the relevant igneous complexes intruded into. However, the collision between the Kaapvaal and Zimbabwe cratons predates the magmatism discussed in this paper and it is evident that the SMZ was not active during or after emplacement of Schiel and Phalaborwa, because no evidence of deformation is displayed by either



complex. In addition rocks of the Rooiberg Group (which are of similar age or slightly younger) are also not deformed (Lenhardt and Eriksson, 2012).

Eriksson *et al.* (2006) note that sedimentary deposition preceding the emplacement of the BIC and Phalaborwa Complex is controlled by extension. Verwoerd and du Toit (2006) state that the Bushveld shares a tensile environment with the Phalaborwa and Schiel complexes. Clarke *et al.* (2009) remark on a feeder dyke for the BIC utilizing the Thabazimbi-Murchison lineament (TML) and a stress field which allowed for dilation of the TML, suggesting NW-SE extension at the time of BIC emplacement. In addition it is unlikely that there was space for such a large complex, of dense mafic minerals, to be emplaced into anything but an extensional setting. The carbonatite- LIP link adds significant weight to this hypothesis, and it is reasonable to argue that the Kaapvaal Craton was subject to extension when the BIC was emplaced.

A clear extensional setting for the Kaapvaal craton lends credence to the mantle plume hypothesis proposed as the origin of the BIC by Hatton (1995). In this model, extension is caused by the thermal expansion created by the plume head, leading to decompression in the lower portion of the craton and subsequently the intrusion of first carbonatites and then mafic magma. A mantle plume hypothesis is also often proposed for the origin of carbonatites (Ernst and Bell, 2009; Bell and Simonetti, 2009), lending further support to the model. It is possible that the setting may have been transtensional rather than extensional; however, other than the Limpopo belt, the other margins of the Kaapvaal craton are obscure at 2060 Ma, and it is difficult to either refute or support such a model.

### 3.5 Conclusion

The Phalaborwa, maybe Schiel and Bushveld Complexes were emplaced in close spatial proximity on the same craton. The Bushveld Complex is slightly younger, but the Phalaborwa, Bushveld and possibly Schiel complexes are of similar age.

Therefore the complexes formed under the same tectonic setting, which is known to be extensional for the Phalaborwa and Schiel complexes. Therefore the hypothesis is not rejected and the Bushveld Complex can be considered to have been emplaced into an extensional tectonic setting. This lends support for a mantle plume origin for the Bushveld.

## Chapter 4:

# Carbonatites of the Pilanesberg Alkaline Province

### 4.1 Introduction

It was shown in chapter 3 that carbonatites can be considered rift related and used to infer extensional tectonic setting at the time and location of carbonatite emplacement. In this chapter, the Pilanesberg carbonatites of the Kaapvaal Craton will be reviewed and other magmatism at that time will be considered in order to test the hypothesis that carbonatites may be used indicators of extensional tectonic settings.

This chapter will review a younger “generation” of carbonatites on the Kaapvaal Craton, collectively referred to as the Pilanesberg Alkaline Province (Ferguson, 1973). The Pilanesberg Alkaline Province was emplaced between 1450 Ma and 1200 Ma (Harmer, 1992), during a time of relative tectonic stability. The period of stability occurred between the deposition of the ~1750 Ma old Waterberg Group and the ~324 Ma Dwyka glaciations (Visser, 1990) and included some tensional faulting, mafic intrusions (Verwoerd, 2006) and kimberlites (Skinner and Truswell, 2006).

This chapter will also serve as a brief review of the carbonatites that occur as part of the Pilanesberg Alkaline Province, but will exclude complexes which do not include carbonatites (with the exception of the Pilanesberg Complex itself). Note that by the definition used for this work, “carbonatites” with a metasomatic origin (e.g. the Goudini Complex) are not considered to be carbonatites.

#### *The Pilanesberg Complex*

The Pilanesberg Complex has a concentric ring structure and a 28 km diameter (East-West). The complex hosts volcanic rocks including phonolitic and trachytic lava flows, volcanic and lucastrine sediments and tuff. The plutonic rocks of the Pilanesberg Complex occur mainly as five rings which are composed of nepheline syenite and syenite (one intrusion) and several smaller arc shaped intrusions. No carbonatites have been reported to occur within the Pilanesberg Complex (Verwoerd, 2006).

With regard to geological setting, the Pilanesberg Complex was intruded into rocks of the Bushveld igneous Complex and most likely the Waterberg Group. The Pilanesberg Complex is in contact with Rustenburg Layered Suite gabbro and the Lebowa Granite Suite on the eastern side of the complex. The Pilanesberg Complex is not described as being in contact with the Waterberg Group at present, but should have been intruded into or overlain by the older Waterberg Group during emplacement (Verwoerd, 2006).

In addition to the Pilanesberg Complex; alkaline and carbonatite complexes include the Spitskop Complex, the Goudini Complex (figure 4.1) and two clusters, referred to as sub-provinces.

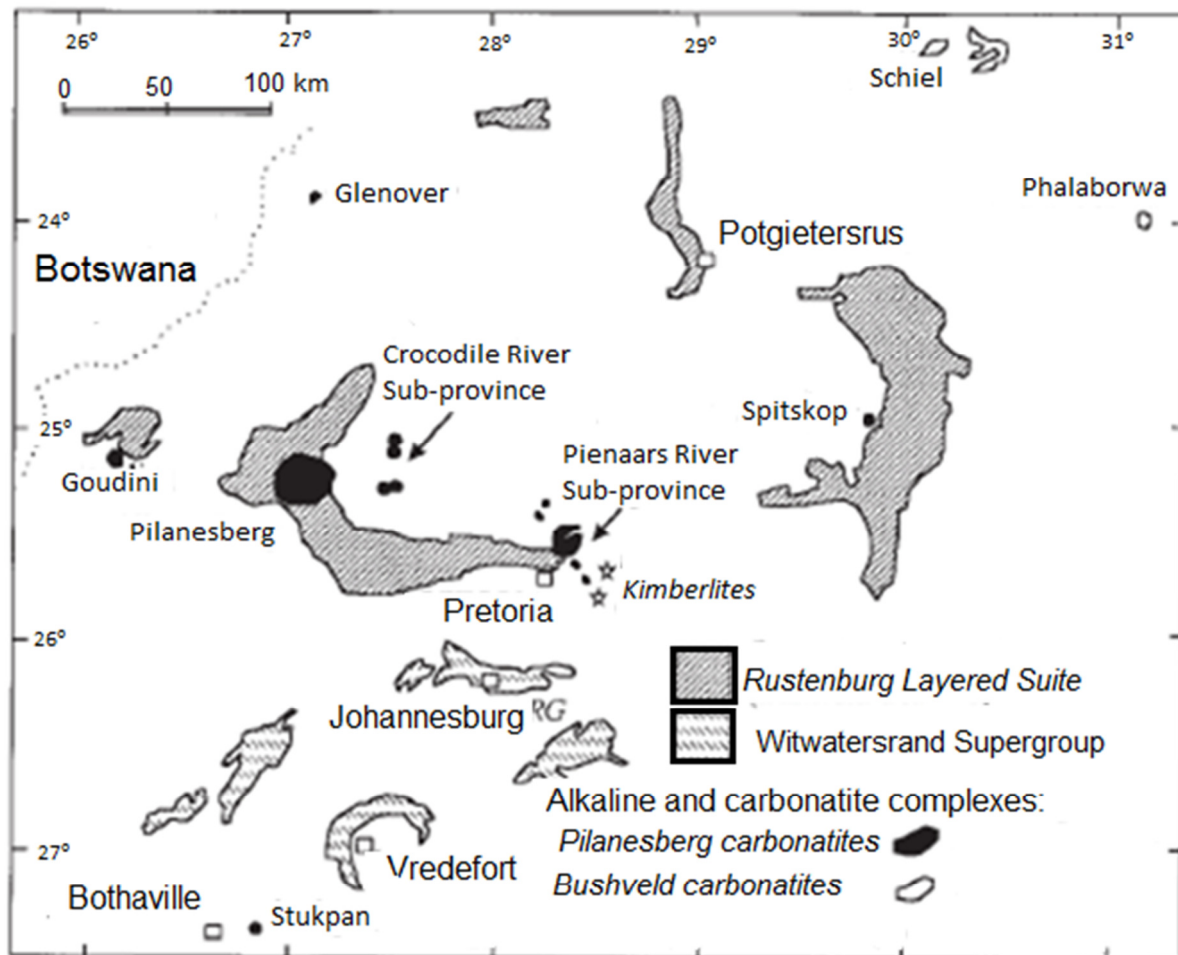


Figure 4.1: The locations of carbonatites on the Kaapvaal Craton and the Rustenburg Layered Suite (for reference) (after Harmer, 1999).

## 4.2 Carbonatites

### *Spitskop*

The Spitskop Complex is located 50 km northeast of Groblersdal (Verwoerd, 2006) and is the eastern-most complex of the Pilanesberg Alkaline Province (Harmer and Gittins, 1997). The Spitskop Complex is 1341 Ma old (Harmer, 1999) and was intruded into the Lebowa Granite Suite, a suite of granites contemporaneous with and generally included in the Bushveld Complex.

Initially the Spitskop Complex formed by the intrusion of a pyroxenite pipe, which was then intruded by ijolite sheets, nepheline syenite bodies and a plug like composite carbonatite intrusion with a diameter of ~1.5 km (Harmer and Gittins, 1997) and, along with intrusions, notable fenites formed (Verwoerd, 2006). The pyroxenite does not contain a considerable amount of mica in general, but is micaceous in some places and some glimmerite patches occur (Verwoerd, 2006). Harmer (1992) classifies ijolite into two groups, one which appears to be magmatic and another with of possible metasomatic origin.

Carbonatites at the Spitskop Complex are mostly dolomite-carbonatite, but also include calcite-carbonatite. The carbonatite plug in the Spitskop Complex consists of an early discontinuous outer zone of calcite-carbonatite around an intermediate zone of dolomite and calcite-carbonatite and a centre of dolomite-carbonatite. The intermediate zone was interpreted by Verwoerd (1967) to have been formed by replacement of earlier calcite-carbonatite by magnesian fluids. Harmer and Gittins (1997) note that textural relationships are not consistent with replacement; rather textural relationships are similar to other magmatic rock and represent phenocrysts (calcite) and groundmass (dolomite). With reference to the formation model proposed by (Harmer and Gittins, 1998), the intermediate carbonatite may represent a special case where conditions changed in magma during crystallization, favouring stability of dolomite, to such an extent that dolomite became stable and started precipitating around calcite phenocrysts.

### *Crocodile River Subprovince*

The Kruidfontein Complex is located within the Crocodile River Fragment (a Transvaal Supergroup inlier within the Bushveld Complex). The Kruidfontein Complex is exposed at the volcanic level and is made up of mostly of volcanic sediments, likely of alkaline composition, deposited in a caldera or crater lake. Most of the carbonate rich rocks in the Kruidfontein complex are of metasomatic origin and therefore not considered carbonatites by the definition used for this work. There are small intrusive bodies of calcite-carbonatite in the form of dykes or plugs (Verwoerd, 2006).

The Tweerivier Complex is located where the Elands and Crocodile rivers meet and consists of two coalesced carbonatite pipes (Verwoerd, 1967). The southern pipe consists of carbonatite dykes in blocks of Rustenburg Layered Suite gabbro and anorthosite. The northern pipe consists of calcite-carbonatite and dolomite-carbonatite, sedimentary (Transvaal Supergroup) xenoliths and a significant fenite aureole derived from the Lebowa Granite Suite (Verwoerd, 2006).

The Bulhoek Carbonatite is composed of only dolomite carbonatite and is not associated with any magmatic silicate rocks. Xenoliths from the Rustenburg Layered Suite and fragments of granite (country rock) occur. The country rocks have been disaggregated; metasomatised by carbonatite magma and been subject to fenitization (Harmer and Gittins, 1997).

### *Piensaars River Sub-province.*

The Glenover complex is located 85 km north of Thabazimbi and was emplaced into the Waterberg Group. The Glenover Complex is poorly exposed and probably consists mostly of pyroxenite. Dolomite–carbonatite with subordinate calcite-carbonatite traverses the pyroxenite.

The Stukpan Complex is ~ 1354 Ma old (Harmer, 1992), located in the Free State goldfield, ~180 km southwest of Johannesburg is indicated by geophysics to be oval and 3.1 by 1.7 km across. The Stukpan Complex intruded through the Witwatersrand and Ventersdorp supergroups and was discovered while exploring for gold. Drill

cores include calcite-carbonatite, some dolomite-carbonatite and fenite derived from Ventersdorp Supergroup volcanic rocks (Verwoerd, 2006).

### 4.3 Associated magmatic and tectonic activity

The Pilanesberg Dyke Swarm cuts through the north-eastern half of Pilanesberg and extends from Botswana to the Vaal River. The dyke swarm is considered to be part of the Pilanesberg Province on the basis of petrology and similar age. The rocks comprising the Pilanesberg Dyke Swarm have negative magnetic anomalies which causes a fortunate contrast with the Karoo dolerites (Verwoerd, 2006). Ernst *et al.* (2008) consider this dyke swarm a large igneous province (LIP) which coincides with the break-up of the proposed supercontinent Nuna (Columbia).

Kimberlites of similar age to the Pilanesberg Alkaline Province occur in the Cullinan Province; dated at ~1200 Ma (Kramers and Smith, 1983). The Cullinan Province is located east of Pretoria and was intruded through the basement into The Transvaal Supergroup and Waterberg group. Harmer (1992) considers it likely that the Cullinan Province was emplaced at the end of magmatic activity producing the Pienaars River Subprovince (Skinner and Truswell, 2006).

The Namaqua-Natal metamorphic belt was formed by orogeny along the southern and western margins of the Kaapvaal Craton. This orogeny was initiated at a similar time to the Pilanesberg Alkaline Province (Eglington *et al.* 1989) and formed between 1200 and 1000 Ma (Skinner and Truswell, 2006). However, the interior of the craton would have been stable, according to Verwoerd (2006), or experiencing an extensional tectonic setting, as indicated by the occurrence of the Pilanesberg Dyke Swarm and kimberlites during magmatic activity involving the Pilanesberg Alkaline Province. In addition, from the ages presented such magmatism appears to have preceded the orogeny.

#### **4.4 Discussion**

Information on the Pilanesberg Alkaline Province is sparse and the various complexes and sub-provinces are only linked by a loose spatial and temporal similarity. However, the occurrence of a dyke swarm and kimberlites and the break-up of an early supercontinent (Nuna) indicate that the Kaapvaal Craton was involved in wide scale extensional tectonic activity during the relevant time interval (which may have included multiple rifting events). Therefore the carbonatites of the Pilanesberg Alkaline Province can be associated with extensional tectonic settings, even if each carbonatite has an unclear association with the rest of the Province.

Considering that carbonatites are easy to identify, owing to their small size (which limits uncertainty due to partial exposure) and distinctive mineralogy (compared to the far more common silicates), carbonatites may be a useful indicator of extensional tectonic setting. In the case of the Pilanesberg Alkaline Province the hypothesis of this dissertation is upheld as other rift-related magmatism occurred at the time and an the time interval appears to have been during an extensional phase of a Wilson cycle.

It is likely that more magmatic occurrences of similar age exist on the Kaapvaal Craton but have not been discovered because of being obscured by overlying rocks (esp. the Karoo Supergroup) and the sands of the Kalahari. The period of relative stability ascribed to the Kaapvaal Craton which includes the above events may be considered relatively stable, but not inactive. The craton may appear stable due to what could have been a central and therefore protected location in a continent.

#### **4.5 Conclusion**

The carbonatites considered are spatially and temporally associated with extensional magmatic occurrences, including a dyke swarm and kimberlites. Carbonatite is associated with rifting, involving the break-up of the proposed supercontinent Nuna. Therefore the hypothesis that carbonatites may be used to infer an extensional tectonic setting at the time of emplacement is not rejected.



## Chapter 5: Summary

*Chapter 2* considers three mechanisms for carbonatite melt generation: primary mantle melting; liquid immiscibility; and, crystal fractionation. Even though fractional crystallization can and does occur in carbonatite magma, fractional crystallization is discounted as a mechanism for the origin of carbonatite because of a lack of support from experimental studies and inconsistency with observations. Liquid immiscibility is not ruled out as a possible mechanism for the origin of some carbonatites, but is not considered the process responsible for the formation of most carbonatites. The liquid immiscibility model is found lacking because it does not accommodate a mantle origin for carbonatites or address the compositional dichotomy. The direct mantle melting model is applicable to all but a few carbonatites and presents a more comprehensive explanation of observations than the liquid immiscibility model. The direct mantle melting model involves dolomite-carbonatite magma produced from low degrees of partial melting in the mantle. The dolomite-carbonatite melt then reacts with harzburgite and lherzolite to produce metasomatic wehrlite and become more Ca-rich. In so doing the magma creates a conduit for new melt to avoid such a reaction. Once carbonatite magma reaches the crust it simply crystallizes if Ca-rich or crystallizes calcite or dolomite (depending on if conditions allow the stability of dolomite) if the magma remains Mg-rich.

*Chapter 3* considers the Bushveld age carbonatites with a focus on inferring tectonic setting. The paper considers and supports that carbonatites are restricted to emplacement in extensional tectonic settings. In addition the ages of the relevant igneous complexes are examined and the conflicting ages for the Schiel Complex are not resolved in this work. An age of ~2.06 Ga is accepted for the Phalaborwa complex and the BIC is regarded as having an age between 2.06 and 2.055 Ga. These ages give a temporal association in addition to the spatial association between the Phalaborwa and Bushveld complexes, which restricts the BIC to the same tectonic conditions during emplacement as carbonatite, Therefore it is concluded that the BIC was emplaced into an extensional tectonic setting.

*Chapter 4* considers the carbonatites of the Pilanesberg Alkaline Province in an attempt to test if carbonatites can be used as pointers an extensional tectonic setting at the time of emplacement. Due to the emplacement of a dyke swarm, kimberlites and the break-up of Nena at the same time interval as the carbonatites of the Pilanesberg Alkaline Province; the hypothesis is not rejected.

## **Acknowledgements**

As above it must be acknowledged that the carbonatite-LIP association (esp. between Phalaborwa and the Bushveld Igneous Complex) shown by Ernst and Bell (2009) lead to the concept of inferring tectonic setting from the occurrence of carbonatites.

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Finally I would like to dedicate this dissertation to my grandfather, Petrus (Piet) Erasmus.

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