The Bushveld Large Igneous Province

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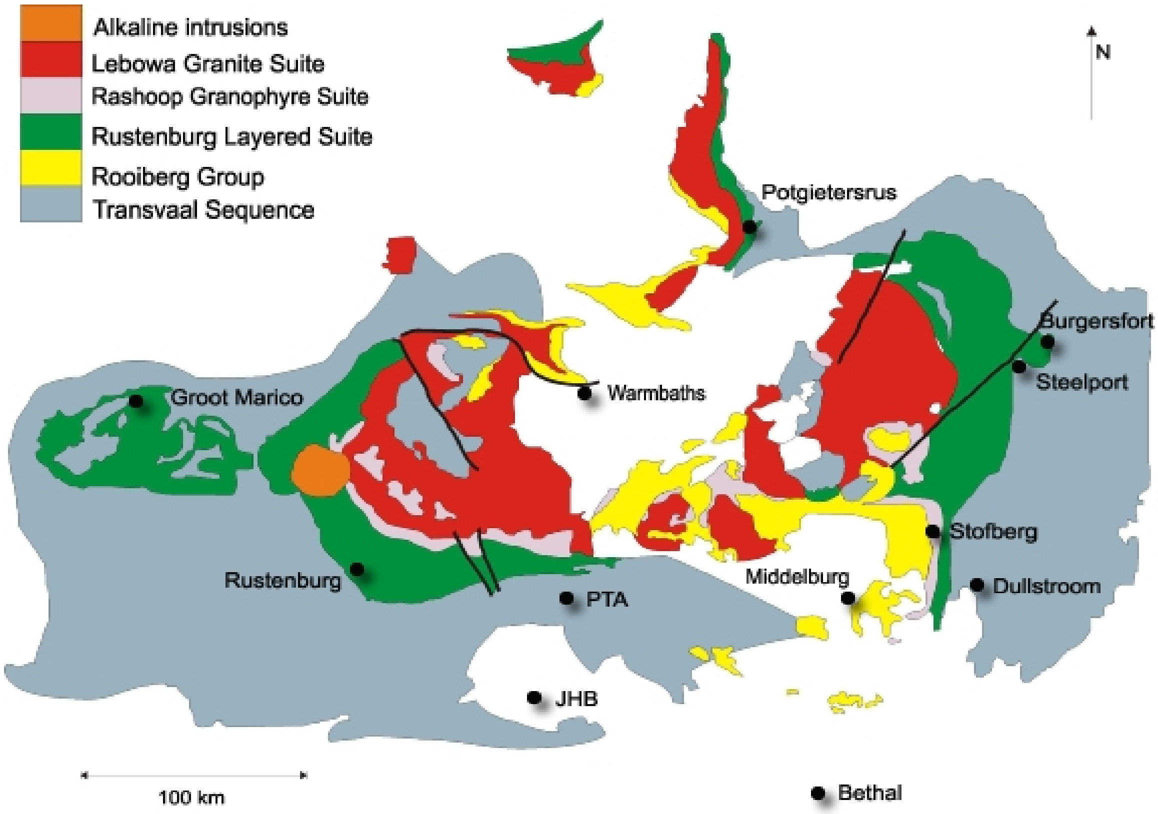


Figure 1 Simplified geological map of the Bushveld Large Igneous Province, which includes the Rustenburg Layered Suite, the Rooiberg Volcanics and the Lebowa Granite Suite

Overview of the Palaeoproterozoic Bushveld Igneous Province

The Palaeoproterozoic Bushveld Igneous Province in South Africa is comprised of:

- a suite of mafic sills which intruded the floor rocks of Transvaal Supergroup
- the bimodal but predominantly Rooiberg Group volcanic province: one of the largest pyroclastic provinces on Earth covering at least 50 000 km^2 and up to 3 km thick
- the Rustenburg Layered Suite, the largest and oldest mafic layered complex on Earth which covers an area of approximately 65,000 km^2 and comprises anorthosites, mafic and ultramafic cumulates
- the Lebowa Granite Suite
- the Rashoop Granophyre Suite developed at the contacts between the granites and Rustenburg Layered Suite which is comprised of metamorphosed sediments and intrusive acidic rocks.
- various satellite intrusions of similar age including the Molopo Farms and Nkomati - Uitkomst.

Introduction

Large Igneous Provinces (LIPs) have been defined by Coffin and Eldholm (1994) as 'massive crustal emplacements of predominantly mafic Mg and Fe rich extrusive and intrusive rock which originate via processes other then *normal* seafloor spreading and include continental flood basalts, volcanic passive margins, oceanic plateaus, submarine ridges, seamount groups and ocean basin flood basalts'. The Bushveld magmatic province is an unusual LIP. In spite of its Palaeoproterozoic age, it is undeformed and it comprises in part voluminous volcanics that were predominantly felsic rather than basaltic in composition and it lacks associated dyke swarms. Indeed the feeders to the province are a matter of debate. However, in common with other LIP's, magmatism occurred over a very short time interval, of less than 10 Ma, and was very voluminous. A conservative estimate of the volume would be around half a million cubic kilometres (Table 1) although the precise estimation of LIP size cannot be calculated because of erosion of the extrusive components in particular and it is difficult to estimate the amount of basaltic material that has been intruded as sills or indeed which underplated the crust. However, Harmer and Armstrong (2000) has suggested that between 0.7 and 1 million km^3 were produced within 1-3 Ma, which would require magma generation rates of between 1 and $0.3x10^6km^3$ per Ma respectively. If the estimates of magma volumes of 384 $x10^6km^3$ for the Rustenburg Layered Suite (Cawthorn and Walraven, 1997) and 200 $x10^6km^3$ for Molopo Farms (Reichardt, 1994) are included then a cumulative volume of magma in excess of 1 to 1.5 $x10^6km^3$ was generated which is comparable in volume to major flood basalt provinces such as the Deccan and North Atlantic Tertiay Provinces (Gibson **and** Stevens, 1998).

The Palaeoproterozoic Bushveld Complex has been studied in detail in the last century, primarily because of the richness of the ore deposits of platinum, palladium, rhodium, chromium and vanadium. In spite of the vast literature that has accumulated there has been little attention to the Large Igneous Province as a whole to which it belongs, and there is no consensus of opinion as to the tectonic setting for the magmatism, or whether it was plume-related.

	Areal extent	Maximum thickness	Conservative estimate of total volume $x10^3$ km^3
Rooiberg Group volcanic province	50 000 – 100 000 km^2 although it may have originally extended >200 000 km^2	Up to 3 km	350 (384, Cawthorn and Walraven, 1997)
Rustenburg Layered Suite	65,000 km^2	Up to 9 km	150^1-400^2
Bushveld Granite Suite and granophyres	30 000 km^2	Up to 3 km(*)	180
Satellite intrusions – Molopo Farms	12 000 km^2	Up to 3 km(*)	30
Total			710-1060

Table 1 General estimates for the areal distribution and volume of the various components of the Bushveld LIP. [1]No mafics below centre [2]Mafics continuous below centre. Estimated volumes from Harmer, (2000). Reichardt, 1994 gives an estimated volume for the Molopo Farms Complex of 200 000 km^3.

This paper aims to be an early contribution towards a summary of this Large Igneous Province although the author is by no means an expert and contributions from readers of this page will be appreciated. The enormous contribution of Johan (Moose) Kruger and Paul Nex to this compilation is gratefully acknowledged.

The host rocks

It is generally accepted that the Kapvaal Craton had already formed a deep root by 3.2-3.1 Ga. It was subsequently shaped and changed by tectonic processes which saw the amalgamation and accretion of a mosaic of discrete crustal blocks into stable continental crust (Figure 2). Today these multiple parts are recognised as discrete terranes, each with its own tectonic, metamorphic and mineralisation history. Following stabilisation of the craton, the Late Archaean to early Proterozoic history is characterised by the development of large volcano-sedimentary intracratonic basins on the stable platform. One of the most significant of these is the NE-trending Archaean Witwatersrand basin (Figure 3), the source of 40% of the world's gold. This is a thick sequence of more than 7000 m of gravels, sands and muds that were deposited in a foreland Basin.

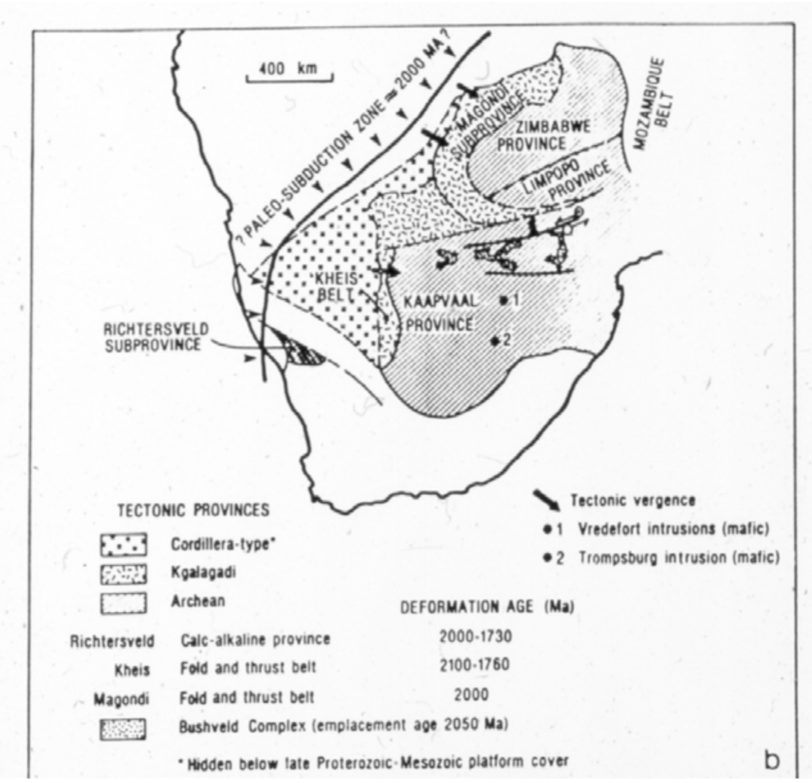


Figure 2 Map of southern Africa showing the location of the Kapvaal craton.

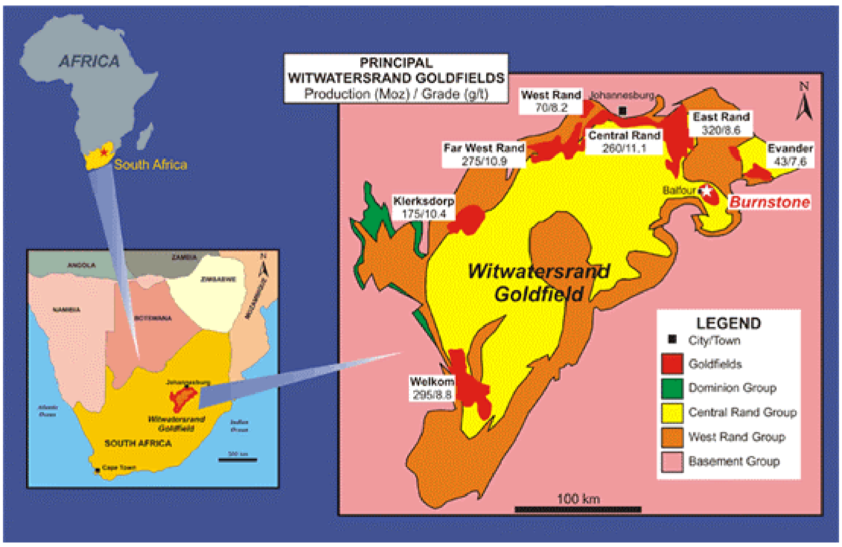
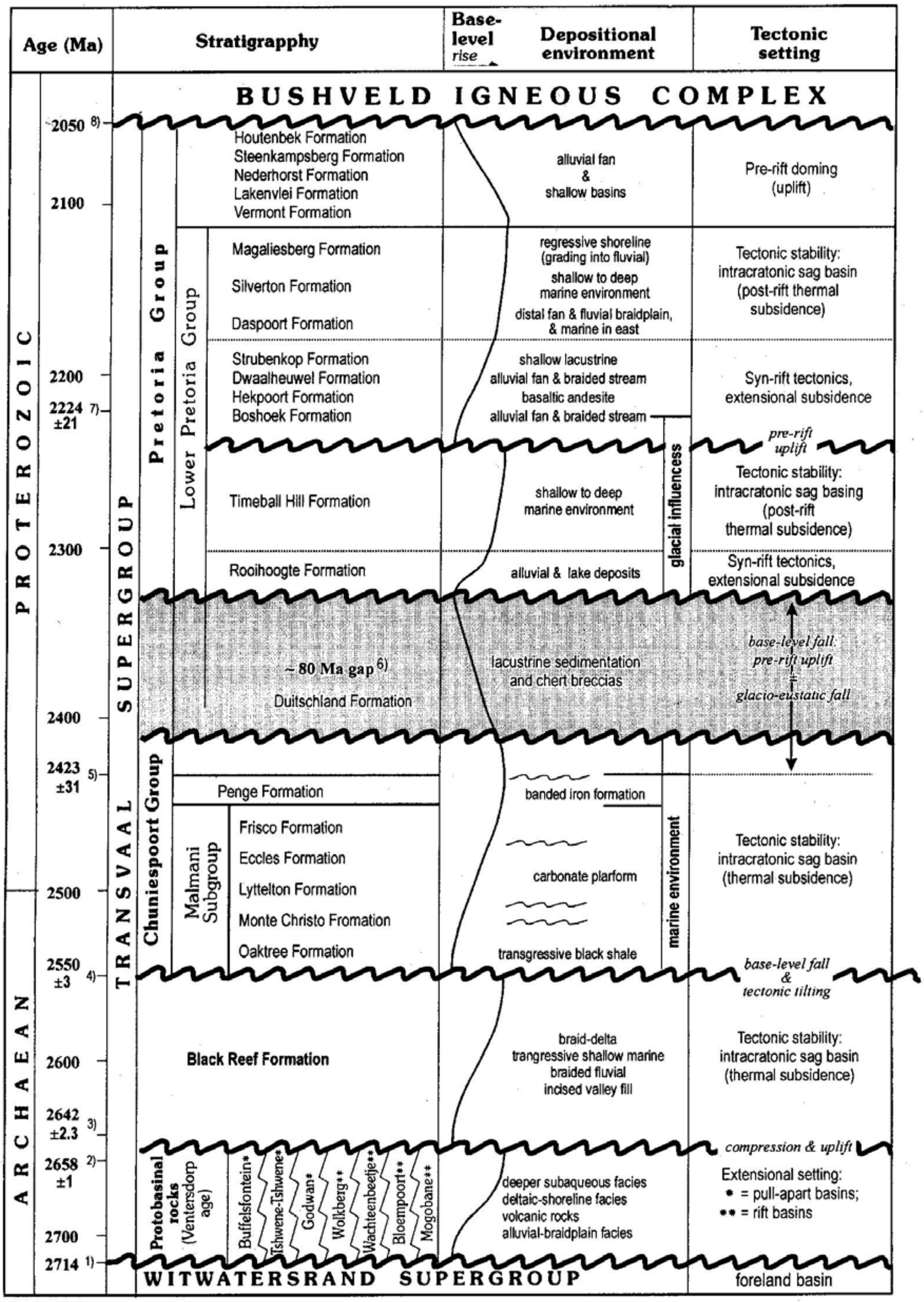


Figure 3 The Archaean Witwatersrand goldfield showing the principal production sites. Map from Great Basin Gold website.

Figure 4. Summary of the Transvaal Supergroup stratigraphy (Eriksson et al., 2001).



The Witwatersrand sequence was deposited either on the granite-greenstone basement or on older sediments and lavas of the Dominion Group. The underlying Dominion lavas have been dated at 3074 Ma, the intercalated lavas have an age of 2917-2780 Ma whilst the overlying Ventersdorp lavas began at 2714 ± 8 (Armstrong et al, 1991).

Ultra high temperature metamorphism in the lower crust is directly correlative with the rapid eruption of the Ventersdorp flood basalts at 2714 ± 8 Ma, and associated crustal melting, plutonism, and widespread extensional tectonics (Schmitz and Bowring (2003). Metamorphism and magmatism is postulated to have occurred in response to nonuniform intracratonic lithospheric thinning and superimposed magmatic heat advection (Schmitz and Bowring (2003). The Klipriviersburg lavas which were erupted rapidly, are up to 2 km thick and cover nearly 160,000 km^2 making them a strong contender for a LIP.

The deposition of the Transvaal Supergroup has been placed within 20 Ma of the end of Ventersdorp rifting (Walraven and Martini 1995). The Transvaal Supergroup, a circa 15 km thick package of sediments, were deposited on the Kaapvaal craton (Figure 4). These range in age from 2.714 (Ventersdorp lavas) to 2.100 Ma (Eriksson et al, 2001). The Black Reef Formation lies unconformably on Archaean basement granite-greenstone. It is composed of quartzites and conglomerates that are of considerable lateral extent, although only a few metres in thickness. The succeeding Chuniesport Group, which has a thickness <2km, comprises a succession of dolomites with intercalated chert and limestone layers and is locally capped by the Penge banded iron formation and the predominantly shaley Duitschland Formation.

An 80 million year gap in sedimentation is represented by a regional angular unconformity and the development of a karst surface. The overlying Pretoria Group, which is up to 2 km thick consists predominantly of shales and quartz arenites with subordinate carbonates and volcanic rocks. The Pretoria Group, which is only preserved on the northern and western part of the Kaapvaal craton, thickens towards the west and the north. For a more detailed overview see Eriksson et al (2001) and Eglinton and Armstrong 2002.

The Bushveld Large Igneous Province

Until the 1990's, it was envisaged that the magmatic rocks of the Bushveld Large Igneous Province evolved over a long period of time, possibly exceeding 100 Ma. However, research in the last decade has shown that the Rooiberg Group volcanics, the Rashoop Granophyre Suite, the intrusive Layered Suite and Granite Suite were synchronous (Table 2). Taken together, the new data indicate that much of the magmatic activity represented by the Rooiberg-Loskop-Bushveld Complex succession occurred within the attainable precision of SHRIMP within 3-5 Ma.. The emplacement ages for the satellite intrusions are also broadly synchronous (Table 2).

Lithostratigraphic unit		Age (Ma ± 95%)
Loskop Formation	rhyolite	2057.2 ± 3.8[1]
Lebowa granite Suite	Makhutso granite Nebo Granite Steelpoort Park Granite	2053.4 ± 3.9[1] 2054.2 ± 2.8[1] 2057.5 ± 4.2[1]
Rustenburg Layered Suite	Critical Zone (SHRIMP) Critical Zone (IDTIMS)	2054.4 ± 2.8[1] 2054.5 ± 1.5[1]
Rashoop Granophyre Suite Rooiberg Group	Rooikoppies porphyry Kwaggasnek Formation	2061.8 ± 5.5[1] 2057.3 ± 2.8[1]
Satellite intrusions	Molopo Farms Mashaneng Complex Uitkomst Complex	2044 ± 24[2] 2054 ± 2[3] 2044 ± 8[4] (2055 +45/-17)[4]

Table 2 Recent age dates for various component of the Bushveld LIP. [1]From Harmer and Armstrong (2000) [2]Kruger (1989) [3] Mapeo et al., (2004) [4]de Waal et al., (2001).

Rooiberg Group volcanics

The Rooiberg Group volcanics, which are up to 3.5 km thick in the Loskop area, are preserved over an area of > 50,000 km^2 (Schweitzer et al, 1995) and may even have exceeded 300 000 km^3 (Twist and French, 1983) although in many areas the succession is extensively thinned or removed by erosion (Buchanan et al, 2002). A postulated extent of the volcanics is shown in Figure 5. The Rooiberg Group, unconformably overlie the Transvaal Supergroup (Cheney and Twist, 1991) and the Rustenburg Layered Suite of the Bushveld Complex was generally emplaced along or above the unconformity between the volcanics forming the roof of the Complex and the underlying Pretoria Group of the Transvaal Supergroup with volcanics preserved in the floor and roof of the Rustenburg Layered Suite (RLS) of the Bushveld Complex Extrusion of the upper units of the Rooiberg Group may have been synchronous with late RLS or granite emplacement (Schweitzer et al.,1995) If so, then felsic units added to the top of the volcanic pile while sills intruded below. Isotopic ages for these extrusive units fall within the age range of 2061 +/- 2 and 2052 +/- 48 (Walraven 1987).

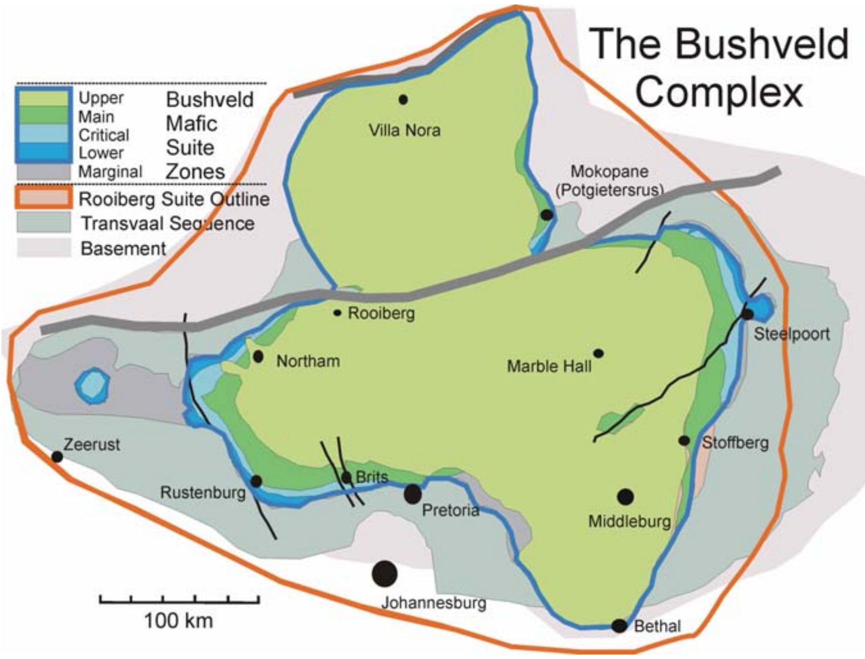


Figure 5. Postulated extent of the original Rooiberg Volcanic Suite (Kruger, 2004)

The Rooiberg Group has been subdivided into four Formations (Figure 6) on the basis of colour, texture, phenocryst content and internal structure:

Schrikkloof Formation
Kwaggasnek Formation
Damwal Formation
Dullstroom Formation

The Dullstroom Formation is the oldest stratigraphic unit of the LIP. It has been subdivided into at least three compositional groups: low-Ti mafic to intermediate units, high-Ti mafic to intermediate units and high-Mg felsic units (Buchanan et al, 1999). The top of the Dullstroom Formation is marked by the last extrusion of high-Mg felsite and the first sedimentary intercalations and pyroclastics of the Damwal Formation together with high-Fe, Ti, P volcanics in the lower part of the Formation. The Kwaggasnek and Schrikkloof Formations are dominated by dacitic pyroclastics, rare rhyolites flows and intercalated sedimentary horizons (Figure 6).

Major and rare earth element chemistry of the Rooiberg Group is shown in Figure 7.

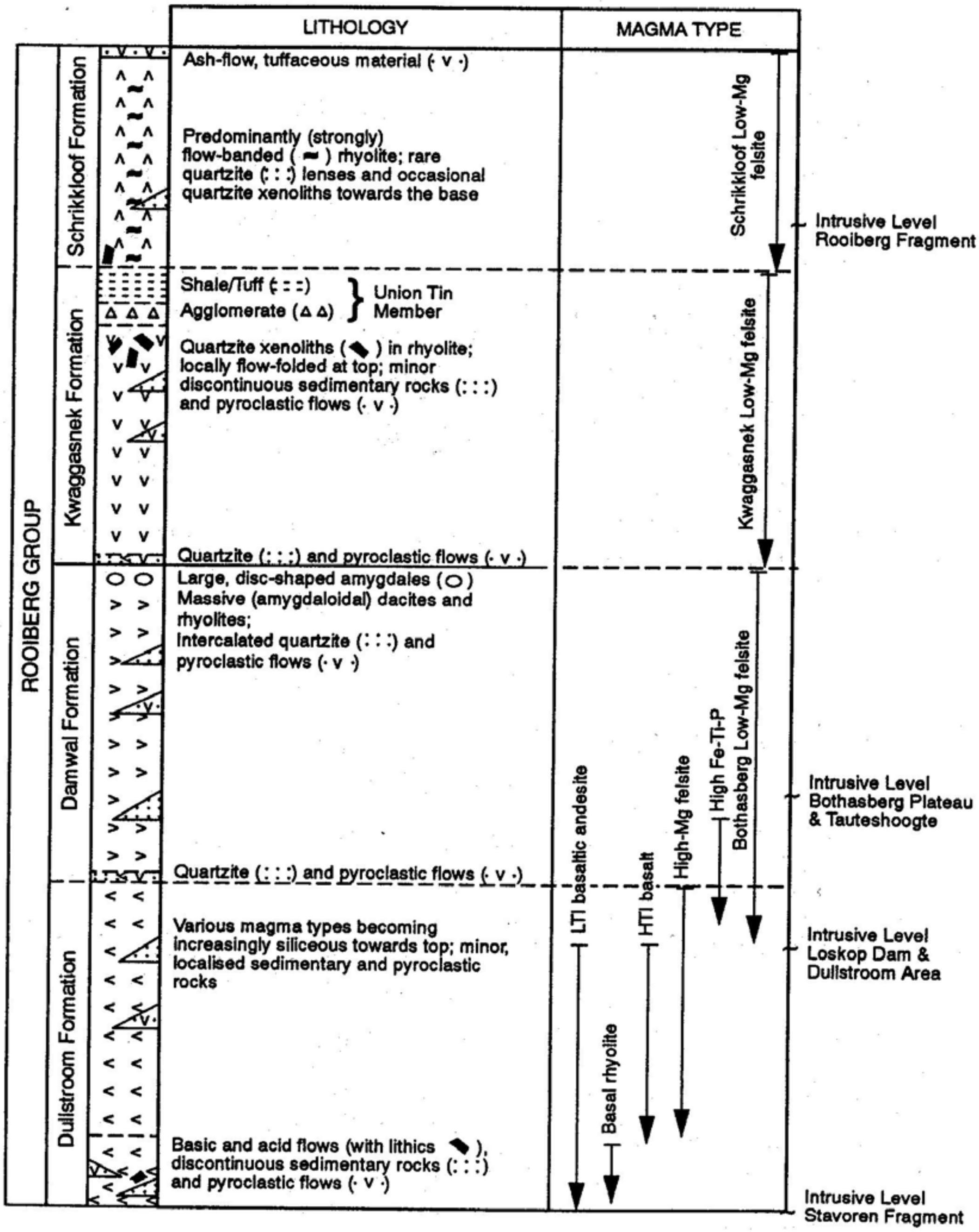


Figure 6. Regional stratigraphic subdivision of the Rooiberg Group (Schweitzer and Hatton, 1995)

The Loskop Formation is composed of clastic sediments with minor volcanic intercalations and, in the east, overlies the Rooiberg Group with no discernible unconformity (Harmer and Armstrong, 2000). Detritus derived from the lower portions of the RLS has been identified in the Loskop Formation sediments (Martini, 1998). The fact that the RLS components intrude into the Rooiberg Group and yet RLS-derived detritus is found within sediments of the overlying Loskop Formation argues that the Rooiberg-Bushveld magmatism must have occurred over a short period of geological time (Harmer and Armstrong, 2000).

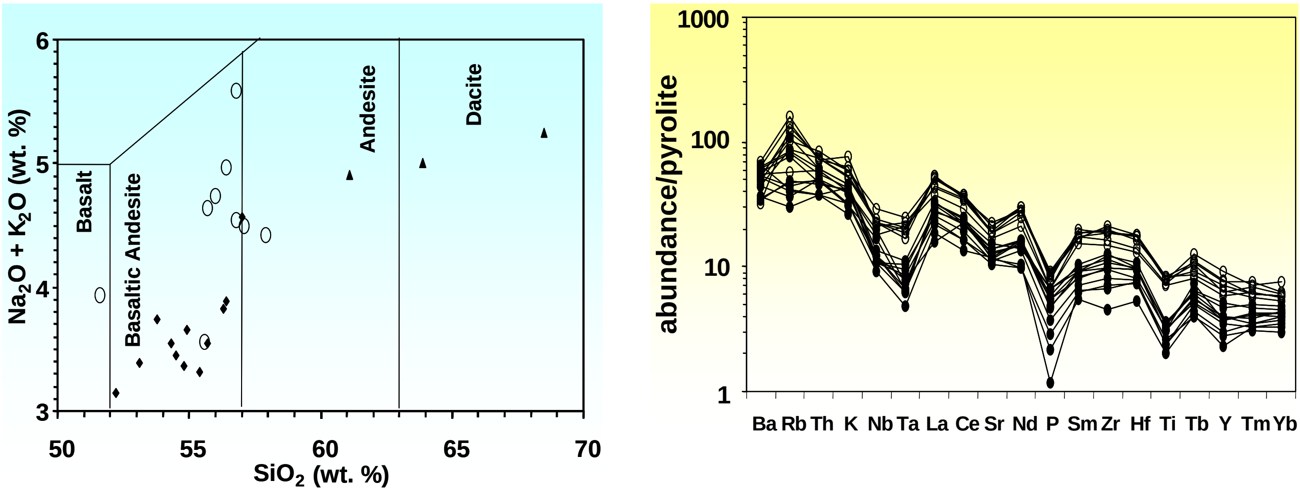


Figure 7. (a) Classification of the Rooiberg volcanics on the basis of chemistry (b) Rare earth chemistry of the Group (data from Buchanan et al, 1999)

Rashoop Granophyre Suite

Granophyric rocks of the Bushveld Complex occur widely between the RLS beneath and the Rooiberg Volcanics above although they are never voluminous. Little work has been completed on the granophyric rocks since a memoir by Walraven in 1987. It is unfortunate in some ways that these texturally similar rocks have been grouped together because of the diversity of origin within the group.

According to Walraven, (1987), the Stavoren granophyre, which is the predominant type, is a shallow intrusive facies of a magma which intruded below the rhyolite roof of the Rooiberg Group or Pretoria Group sediments and also extruded to form the volcanic pile. In contrast, other granophyres formed by the melting of the overlying volcanic roof rocks by the underlying RLS, by recrystallisation of Rooiberg volcanic rocks, or by metamorphism of sedimentary roof rocks.

Rustenburg Layered Suite

The Rustenburg Layered Suite (RLS) was emplaced at shallow crustal levels beneath the volcanic pile of Rooiberg felsites and Rashoop granophyres as sills in the Transvaal Supergroup. North of Burgersfort, emplacement occurred at the level of the Magelliesberg quartzite, but to the south it transgressed upwards through more than 2 km of sediments so that near Stoffberg basaltic rocks of the Dullstroom Formation (at the base of Rooiberg Group) are preserved in the floor. The crescentic outcrop pattern of the RLS is comprised of four exposed sectors, the eastern limb, the western limb, the far western limb and the northern limb, with a fifth limb, the south-eastern Bethal limb, obscured by younger sediments (Figure 8).

The main western and eastern lobes are disrupted by domes and diapirs of floor rocks, the largest of which are the Crocodile River, the Moos River and the Marble Hall fragments. Exposure is poor in the northern and western limbs, but the 200 km long eastern limb extending from Chuniespoort to Stoffberg underlies rugged terrain where surface exposures are far better. Figure 10 is a simplified geological map of the eastern lobe showing some of the localities that are traditionally visited on excursions. A number of these are described and illustrated on the Bushveld Group website at the University of the Witwatersrand. Spectacular views of the stratigraphy and layering of the Rustenburg Layered Suite can be seen from the Chuniespoort – Burgersfort Road near Atok (Figure 11).

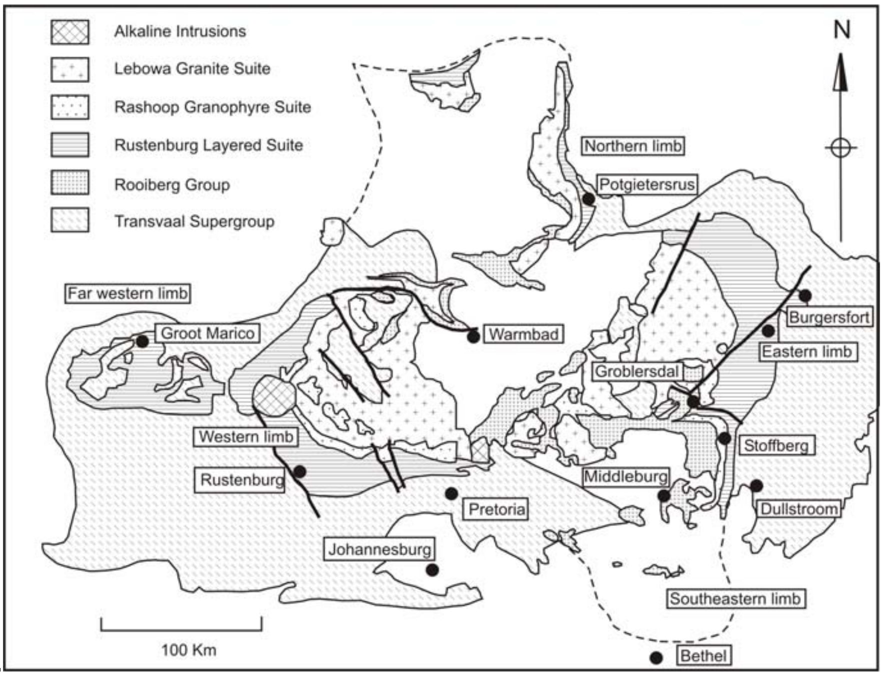


Figure 8. Simplified map of the Bushveld Complex showing the location of the various limbs: the eastern, western, far western and northern limbs and the south-eastern limb, which is obscured by younger cover (Kinnaird et al., 2004). The interpretation of the extent of the northern and south-eastern limbs is based on the gravity data shown in Figure 9.

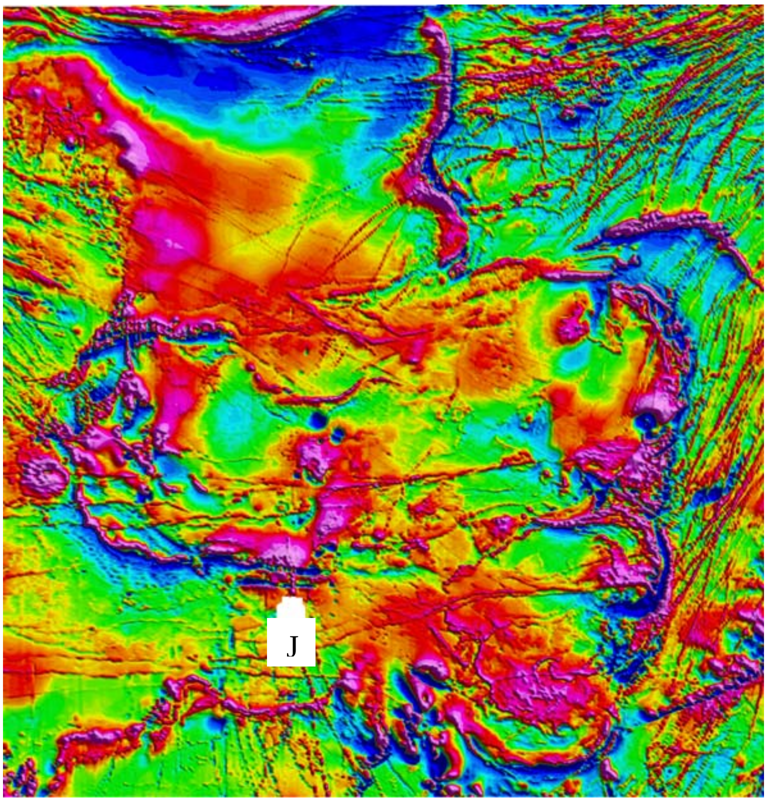


Figure 9. Gravity map of the Bushveld Complex. J marks approximate location of Johannesburg

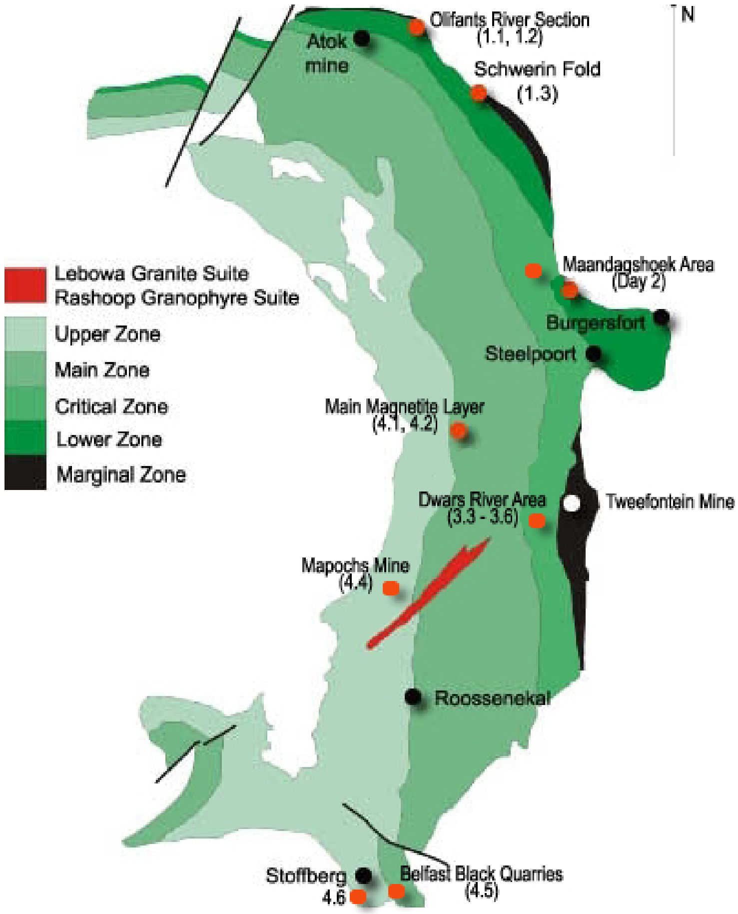


Figure 10. Simplified geological map of the eastern lobe of the Bushveld Igneous Complex.

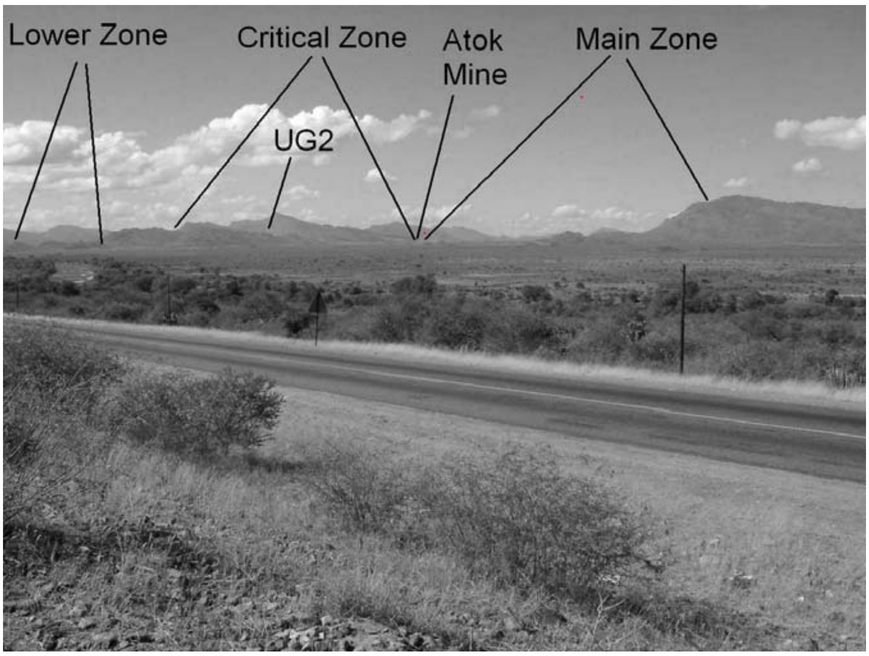


Figure 11. View from the Chuniespoort–Burgersfort road of the Rustenburg Layered Suite. Atok Mine indicates the Merensky Reef at the top of the Critical Zone. Flat-lying land to the right (west) is the lower part of the Main Zone with the range of hills to the right formed by the inverted pigeonite-bearing gabbronorites of the upper Main Zone. Near the crest of this hill is the transition to the Upper Zone in the vicinity of the Pyroxenite Marker.

The RLS has been subdivided into a number of zones, the Marginal, Lower Zone (LZ), Critical (CZ) Main (MZ) and Upper Zones (UZ), although their exact boundaries have been the subject of much debate (e.g. Kruger 1990). Lateral facies variations within the sequence are common.

The Marginal Zone

The Marginal Zone is not always present. Where it occurs it ranges in thickness from zero to hundreds of metres along the basal contact of the Complex. The rocks are most commonly norites with variable proportions of accessory clinopyroxene, quartz, biotite and hornblende, which reflect varying degrees of contamination from the underlying sediments. Generally, it is related to the immediately adjacent cumulate rocks but in places it has been disrupted and has been partly digested by later magma injections (see Eales, 2003 for an overview). However, where Marginal Zone occurs beneath the Lower Zone, it may represent an early magma which in the east occurs as the Shelter norite (SACS, 1980), a succession up to 400m thick around Burgersfort. For a discussion on magma lineages see Kruger, (2004).

The Lower Zone

The Lower Zone has the most limited lateral extent, and is best developed in the northern parts of both eastern and western limbs and in the southernmost part of the northern limb. The thickness of the Lower Zone has been influenced by floor topography and structure and is 1300 m at maximum (Cawthorn et al 2002). In the Oliphants River Trough, in the eastern limb (Figure 12) Cameron, 1978, subdivided the Lower Zone into 3 zones, a central harzburgite between an upper and lower pyroxenite:-

The lower pyroxenite is extremely uniform in composition, containing on average 98% and never less than 95% orthopyroxene with minor interstitial plagioclase and clinopyroxene. Chromitite is absent. The harzburgite unit consists of cyclic units of dunite, harzburgite and pyroxenite varying in thickness from a few to tens of metres. Dunite layers are distinctive, they weather more easily than pyroxenite to a dull greasy brown, they usually contain magnesite veins, and are covered in magnesite float. Little serpentinisation is apparent. Up sequence the orthopyroxene occurs as small oikocrysts, increasing in size up to 1-2 cm. As the modal proportion of orthopyroxene increases the texture changes, with harzburgites containing sub-equant grains of both minerals. In the olivine pyroxenites the olivine appears anhedral against pyroxene. However, in view of the extreme textural recrystallisation in these rocks the inference that the olivine is post-cumulus should be viewed with caution. Scattered chromite grains are present, green clinopyroxene and plagioclase are rare. The orthopyroxene changes in habit from granular to elongate with a range of grain sizes. Igneous lamination may be apparent with 1-3 cm elongate crystals lie in the pla

The upper pyroxenite of Cameron's Lower Zone is similar to the lower one except that variations in grain size produce recognisable layering. The orthopyroxene varies little in composition (En_{84-87}) throughout the entire Lower Zone with more magnesian compositions occurring in the harzburgite together with olivine (Fo_{85-87}).

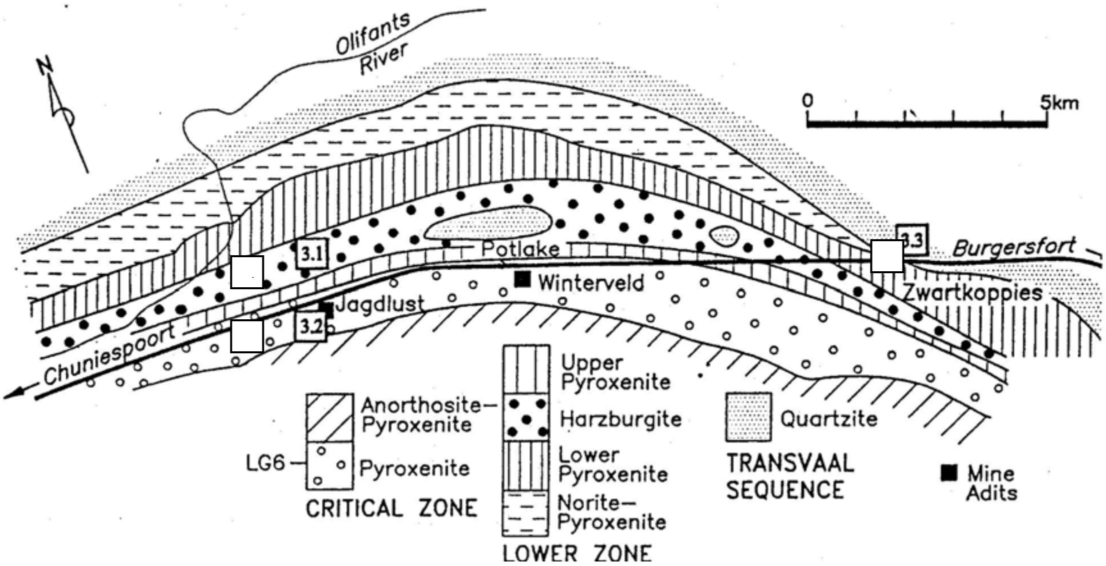


Figure 12 Map of the Oliphants River Trough from Cameron (1978). Location 1.1 in Figure 10

The Critical Zone

The Critical Zone, which is characterised by spectacular layering (Figure 13), hosts world-class chromite and platinum deposits in several different layers (termed reefs). The Critical Zone, which is up to 1500 m thick, is divided into a lower sub-zone (C_LZ) which is entirely ultramafic and is characterised by a thick succession of orthopyroxenitic cumulates and an upper sub zone (C_UZ) that comprises packages of chromitite, harzburgite, pyroxenite, through norite to anorthosite. Subdivision into magmatic cycles is somewhat subjective but nine cycles have been recognised in the C_LZ and eight cycles in the C_UZ consisting of partial or complete sequences from a base of ultramafic cumulates through norite to anorthosite.

Figure 13. Interlayering between chromitites and anorthosites, upper Critical Zone, Dwars River. Location 3.3-3.6 in Figure 10. Note the bifurcations in the chromitite layers.

The base of the upper Critical Zone is defined as the first appearance of cumulus plagioclase and is drawn at the base of the lowermost anorthositic layer of the RLS between two chromitite layers.

Two distinctive cyclic units, the Merensky and Bastard units were included within the CZ of the original classification, however a significant break in the initial Sr isotope ratio, and a major unconformity at the base of the Merensky Unit, led Kruger (1992), to draw the boundary between the CZ and MZ at the base of the Merensky Unit, where the major magma influx occurs, rather than at the top of the Giant Mottled Anorthosite, a distinctive layer characterised by large oikocrysts of pyroxene at the top of the Bastard Unit.

The Main Zone

The Main Zone, which is >3000 m in thickness, forms almost half the thickness of the entire RLS. It comprises a succession of gabbronorites with infrequent anorthosite and pyroxenite bands while olivine and chromite are absent. In addition to the Merensky Reef at its base it is economically important for numerous dimension stone quarries which exploit the Pyramid Gabbronorite; a dark-coloured inverted-pigeonite-bearing gabbronorite.

Although not as spectacularly layered as the Critical Zone discrete packages of modally layered rocks can be identified (Molyneaux, 1974; Mitchell, 1990; Nex et al., 1998, 2002), possibly associated with the influx of new magma. In the eastern Bushveld a modally layered succession of gabbronorites 10-20 m thick occurs some 60-70 m below the Pyroxenite Marker (Quadling and Cawthorn, 1994). This layered package is continuous for 80 km along strike. It has also been identified in the western Bushveld with a 20 km strike extent (Nex et al, 1998). All the layers have sharp bases and planar tops and are composed of orthopyroxene (inverted pigeonite) + clinopyroxene + plagioclase but the proportions vary so that the lighter layers are typically 70% plagioclase, whereas the darker layers are 30-40% plagioclase. Darker layers vary from 2-10 cm in thickness. The layering is considered to be due to mechanical re-distribution of crystals since none of the layers has typical cotectic proportions. In the eastern Bushveld geochemical studies suggest that compositional reversals in orthopyroxene and plagioclase occur slightly above this layered package reflecting the influx of new magma to form the Upper Zone (Nex et al., 2002).

Figure 14 View of layered package in Main Zone gabbronorites south of Steelpoort.

Upper Zone

The Upper Zone is characterized by sequences which are intensely banded with gabbros as the dominant rock type, There is no chill at the top contact with the metamorphosed felsite or granophyre, and the most differentiated rocks occur towards the top. The most striking feature of the Upper Zone is the presence of some 25 magnetitite layers in the eastern limb (Molyneaux, 1974) that cluster into four groups, each with up to seven layers. Magnetite layers typically have sharp bases, but gradational tops. The thickest is 6 m, while the Main Magnetite layer, near the base of the Upper Zone is 2 m thick and is mined for its vanadium content. The titaniferous magnetitite layers comprise a vast source of vanadium ore and hosts almost half of the world's vanadium reserves.

Figure 15. At the national monument locality at Magnet Heights, the contact between the anorthosite and overlying Main Magnetitite layer is well exposed (Location 4.1 on Figure 10). Layers here have the biggest density difference of any two layers above the core-mantle boundary.

Bushveld Granites

The Bushveld Granite Suite, locally termed the Lebowa Granite Suite, comprises a series of sheeted intrusion between 1.5 and 3.5 km thick (Molyneux and Klinkert, 1978; De Beer et al., 1987; Kleeman and Twist, 1989) with an areal extent of some 30 000 km^2 (Figure 1). The granites underlie the heterogeneous, predominantly felsic volcanics of the Rooiberg Group, and sills of the Rashoop Granophyre Suite. The granites post-date the 7-8 km thick layered mafic and ultramafic RLS (Figure 1), as shown by granite feeder dykes that cut the RLS in the eastern limb of the Complex (e.g. Hammerbeck, 1970; Walraven and Hattingh, 1993) and xenoliths of mafic rocks in a granite intrusion breccia (Kleeman and Twist, 1989). According to Wilson et al. (2000), magnetic foliations and lineations are horizontal, reflecting vertical host-rock compression and horizontal magma flow during emplacement, with space created for the granites by roof uplift and floor depression.

The LBS has been subdivided into seven facies (SACS, 1980) with finer-grained variants cutting through or grading into porphyritic types. The Nebo Granite is predominant; the aplitic Lease Granite and the coarse-grained red Bobbejaankop Granite are widespread facies defined largely on colour and texture, whereas the coarse-grained porphyritic Verena Granite, the porphyritic Balmoral Leucogranite, the coarse-grained, locally porphyritic biotite-rich Makutso Granite, and the medium-fine grained usually porphyritic Klipkloof Granite of the eastern Bushveld are geographically restricted facies.

Although early literature (Nicolaysen et al., 1958; Burger et al., 1967; Davies et al., 1970; Hamilton, 1977; Hunter and Hamilton, 1978 and work reviewed in Walraven et al., 1990a) indicated that the granites substantially post-dated the RLS, more recent papers have suggested that both extrusive and intrusive Bushveld magmatism occurred within a time span of a few million years (e.g. Kruger et al., 1987; Walraven et al., 1990a; Walraven and Hattingh, 1993; Schweitzer and Hatton, 1995; Walraven, 1997). This has been confirmed by recent zircon SHRIMP dating on samples from the mafic suite, the Loskop Formation rhyolites, Rashoop granophyres and the Lebowa Granite Suite (Harmer, 2000), which is consistent with the Pb-Pb evaporation dating of single zircons by Walraven and Hattingh (1993), and recrystallised metamorphic titanite of Buick et al. (2001), which indicates that the whole magmatic succession was a quasi-continuous event that happened between 2061 and 2054 million years ago, within the error of the data.

Compositionally, the granites of the Lebowa Suite are predominantly alkali feldspar granites with iron-rich ferromagnesian minerals and silica contents that generally fall in the range 71-77% SiO_2, with low CaO (0.35- >1%), K_2O/Na_2O ratios >1 and with an upward decrease from the base to the roof of the sheet in Ca, Mg, Ti, P, Sr, Ba and a concomitant increase in Si, F, Rb, La, Y and Hf. The granites, have many characteristics of A-type granites as defined by Loiselle and Wones (1979), Collins et al. (1982), Whalen et al. (1987, 1996) and Eby (1990, 1992) and have been categorized as A-type by Kleeman and Twist (1989). Petrographic evidence includes the occurrence of fayalite in the least-evolved facies, and biotite of near end-member annite composition, amphibole of near hastingsite composition (MacCaskie, 1983) and tin and fluorite mineralisation in the most evolved facies. Geochemical evidence includes the relatively low Al, Mg and Ca content, relatively high Fe, F, Cl and HFS elements (Ti, Zr, Hf, Nb and Ta) and low Cr, Co and Ni compared with non-A type granites having comparable SiO_2. Recent petrogenetic models for A-type granites have involved either extensive crystallisation from mantle-derived magmas (+ crustal assimilation) or partial melting of crustal protoliths and the merits of these models have been extensively discussed in the literature (e.g. Collins et al., 1982, Clemens et al., 1986; Whalen et al., 1987; Hill et al., 1996; Whalen et al., 1996). The $\delta^{18}O$ values for the Lebowa Suite range from +5.9 ‰ to +9.5 ‰, which precludes a predominant pelitic source for the Suite (Hill et al., 1996).

Figure 16. Exposures of tin-bearing Bobbejaankop granite at Zaaiplaats

Kinnaird et al (2004) recognise two hydrothermal processes, a primary magmatic mineralisation that occurred at 2057±3 Ma and took place in a very short time span (<1 my) and a much later disseminated and vein-filling hydrothermal process c. 1950 possibly related to the marginal tectonism of the Kaapvaal craton in the Limpopo belt, which may have been a longer-lived event.

Satellite intrusions

In recent years, new geochronological data have not only shown that the Rooiberg Volcanics, the granophyres, Bushveld granites and Layered Suite were co-eval, but a number of other intrusions including the Phalaborwa Carbonatite Complex were also emplaced in the same time period. A discussion of these is beyond the scope of this review. However, a brief review is presented of a few of the satellite intrusions that have been linked to the Bushveld Magmatic Province. These include the Uitkomst Complex, 50 km to the southeast and the Molopo Farms and Mashaneng Complexes to the west in Botswana. The Trompsburg Complex to the south, a 25 000km^2 layered intrusion with up to 2km of gabbro-troctolite-anorthosite, has recently been shown to be 1915 ± 6 Ma which is younger than the Bushveld Magmatic Province (Maier et al., 2003).

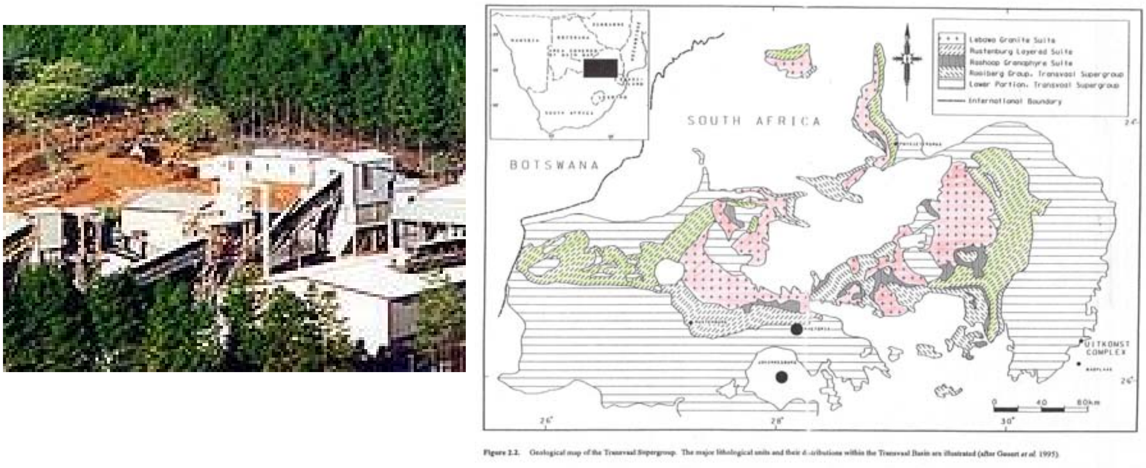


Figure 17. (a) The Nkomati mine is one of the world's lowest cost nickel producers. (b) location map showing the location of the complex 50 km southeast of the main Bushveld Complex.

Uitkomst (Nkomati) Complex

The Uitkomst Complex, a Ni–Cu–PGE–Cr-mineralised layered basic to ultrabasic intrusion hosted by sedimentary rocks of the lower part of the Transvaal Supergroup, lies approximately 200 km east of Pretoria, and 50km east of the eastern limb of the Bushveld Complex (Figure 17). The intrusion is approx. 10 km long and up to 850m thick (Figure 18). It is divided into 7 lithological units (from base to top) the Basal Gabbro, Lower Harzburgite, Chromitiferous Harzburgite, Main Harzburgite, Pyroxenite, Gabbronorite and Upper Gabbro units. The basal gabbro overlies a 2-3m quartzitic unit within the basal Oaktree Formation of the Malmani and the upper contact is within the upper Timeball Hill Formation.

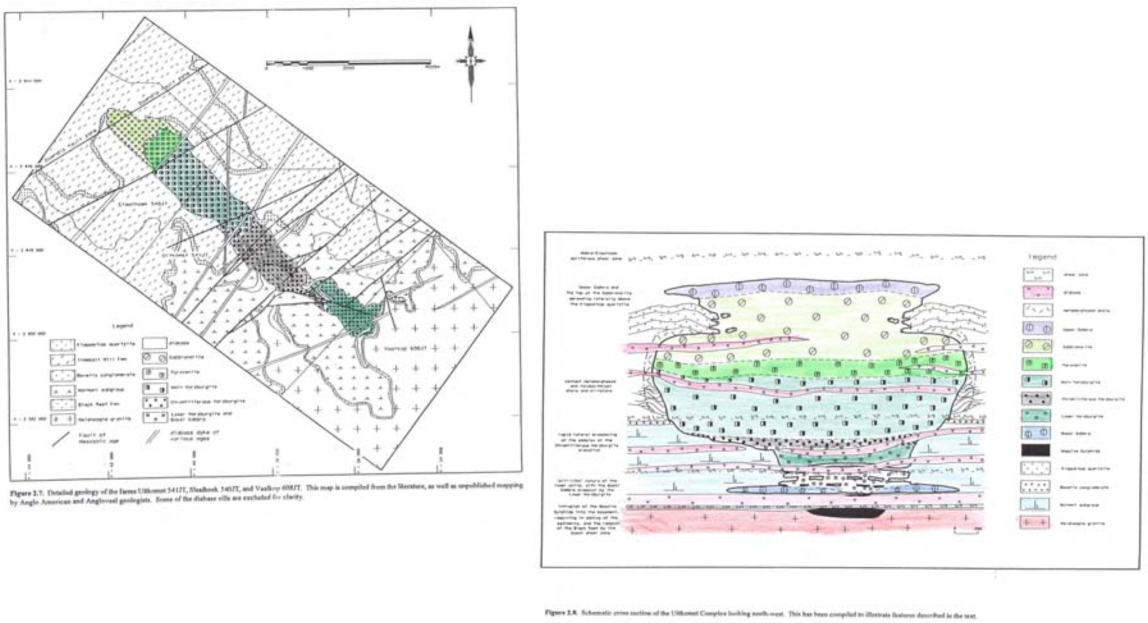


Figure 18. (a) Simplified map of the Uitkomst Complex (b) cross-section through the Complex

The Complex has a concordant $^{207}Pb/^{206}Pb$ zircon age of 2044 ± 8 Ma (de Waal et al, 2001). Some non-zero lead loss is indicated and a Monte Carlo simulation yields a discordia intercept at 2055(+45/-17) Ma. suggesting that it is coeval with the Rustenburg Layered Suite (RLS) of the Bushveld Complex(de Waal et al, 2001). They suggest that chemical modelling provides evidence that the boninitic Bushveld B1 magma is parental to both the lower ultrabasic and upper basic layered series of the Uitkomst Complex. The layered series crystallized in two stages, *i.e.*, a lower conduit and an upper closed-system stage. The tubular shape of the Uitkomst Complex is the result of the intersection of a near-horizontal bedding plane fault with an existing vertical fracture zone under tensional conditions. During the conduit stage, a combination of magma mixing, contamination and flow dynamics may have facilitated sulphide formation and segregation. The identification of Bushveld B1 magma as the major parental magma of the Uitkomst Complex has significance in the exploration for similarly mineralised sub-RLS intrusive bodies (de Waal et al, 2001).

There are 3 disseminated, sulphide-mineralised zones:

–the Basal Mineralised Zone (BMZ),

–Main Mineralised Zone (MMZ) and the –Chromititic Pyroxenite Mineralised Zone (PCMZ).

At the base of the Uitkomst complex, another sulphide orebody, the Massive Sulphide Body (MSB), was discovered which is the smallest and richest of the ore bodies.

Molopo Farms

The Molopo Farms Igneous Complex (MFIC) is located in southern Botswana on the border with South Africa (Figure 19). It is a large layered intrusion covering some 13,000 km^2 and up to 3 km thick (Reichardt, 1994) that is completely covered by Tertiary to Recent Kalahari Beds. Borehole and gravity data show the complex to be an elliptical saucer-shaped intrusion. Kruger (1989) obtained an age of 2044 +/- 24 for the Complex. The MFIC is bisected by a major ENE structure, the Kgomodikae Lineament. Where this structure passes into South Africa it is termed the Thabazimbi-Murchison Lineament.

Figure 19. Map showing the location of the Molopo Farms and Moshaneng satellite intrusions in southern Botswana (from Mapeo et al 2004). For the Bushveld Complex NL=northern limb, EL =eastern limb, WL = western limb and FWL=far western limb. TML = Thabazimbi-Murchison Lineament.

This lineament (Figure 19) is regarded as a fundamental feeder zone for Bushveld Province magmas and may well have provided a connecting channel in Bushveld times. Little new published work is available on this complex since Reichardt (1994), although there has been extensive recent drilling to investigate Ni Cu-Ni-PGE resources.

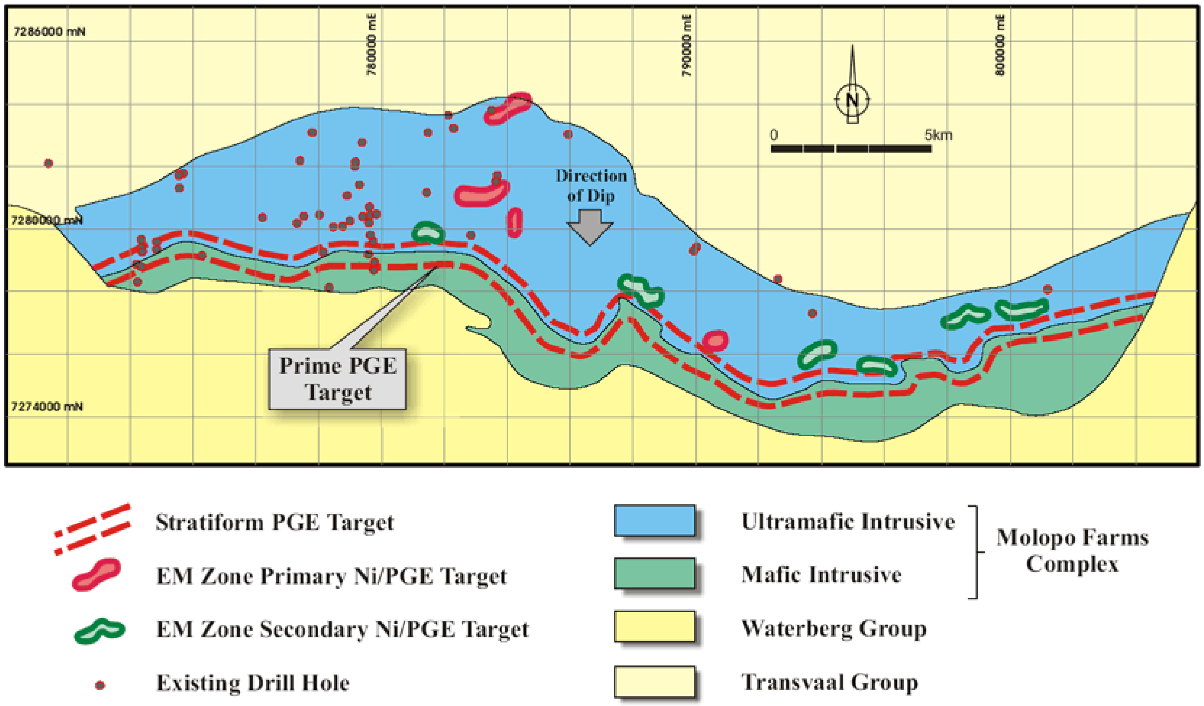


Figure 20. Simplified geological map of the Molopo Farms Complex from Tau Mining website

The Molopo Farms Complex can be sub-divided into three zones:

- An upper zone of homogenous quartz gabbronorite, quartz gabbro and quartz diorite which are often magnetite bearing, similar to the Upper Zone of the Bushveld Complex.
- A layered mafic zone, up to 1000 m thick, consisting of a cyclic sequence of pyroxenite (often feldspathic), norite and gabbronorite, shows an overall upward fractionation from norite and gabbronorite to quartz gabbronorite and quartz hornblende gabbro.
- The Ultramafic sequence (shown in blue on Figure 20) consists of cyclically layered pyroxenite, harzburgite and minor dunite. It can be subdivided into a lower 900 m thick Harzburgite Succession where olivine-rich cumulates dominate and an upper approx. 400 m thick Bronzitite Succession dominated by opx-rich rocks (Reichardt, 1994) The base of the ultramafic zone (base of the intrusive) has the potential to host massive sulphide deposits as implied by the shear controlled massive sulphide mineralisation intersected previously at Keng and disseminated sulphides at Tubane.

Strongly fractionated sill and dyke offshoots of the Complex are a common feature (Reichardt, 1994). Within the layered mafic zone, equated with the Critical Zone of the Bushveld Complex, Gold Fields obtained an intersection of 1.24m of 3.25 g/t PGE + Au.
By including the Molopo Farms into the LIP, the Province extends over 900 km long and 600 km wide.

Mashaneng Complex

Between the Molopo Farms Complex and the Bushveld Complex lies the Moshaneng Complex. It is a 35 km^2 oval-shaped pluton, with coarse-medium grained gabbros and diorites at the centre rimmed by granites and syenites that show evidence of mixing and mingling of co-existing mafic and felsic magmas (Mapeo et al, 2004). U-Pb zircon and titanite isotopic data indicate an emplacement age of 2054 +/- 2 Ma (Mapeo et al, 2004).

Structure of the Bushveld Complex

The overall structure of the Bushveld Complex is still controversial: is it a number of disconnected, arcuate, deep and narrow troughs corresponding to the current outcrop areas, or, is it a single, connected, wide and shallow, soup-dish shaped intrusion? The major intrusion appears to have exploited the contact between the low-density volcanic pile and the denser Pretoria Group. This contact is thought by some to be an unconformity (Cheney & Twist, 1991), or the result of discordant intrusion into a structurally tilted basin (Cawthorn 1998). Kruger (2004) suggests that this low-density carapace of felsite is why a layered intrusion rather than a volcanic province formed, and that the Bushveld Complex as a whole initially intruded as an elongate body beneath the Rooiberg carapace. He concludes that the Bushveld Complex is a lobate, interconnected, *wide and shallow*, sill-like intrusion with upturned margins; rather like a flat-bottomed soup-dish (Figure 21). This is in sharp contrast to the disconnected, *deep and narrow*, ring-like troughs, or, steeply dipping, wedge-shaped, cone-intrusions inferred from geophysical and other considerations (see Kruger, op. cit. for a more detailed discussion and other references). Kruger's model also implies that the first magmas intruded immediately south of the Thabazimbi-Murchison Lineament (TML) and that a half-graben developed that progressively deepened as more magma was added. This occurred as three to five major intrusive episodes of different magma types and probably exploiting different feeders. These magmas now dominate the Lower and Critical Zones, the Main Zone and the Upper Zone. This is clearly shown in the Sr-isotope (Figure 22) and mineral chemistry, and is confirmed by the major differentiation indexes of the dominant minerals. The final intrusion of the Bushveld granites largely exploited the contact between the dense Bushveld Complex rocks and the overlying felsite and granophyre.

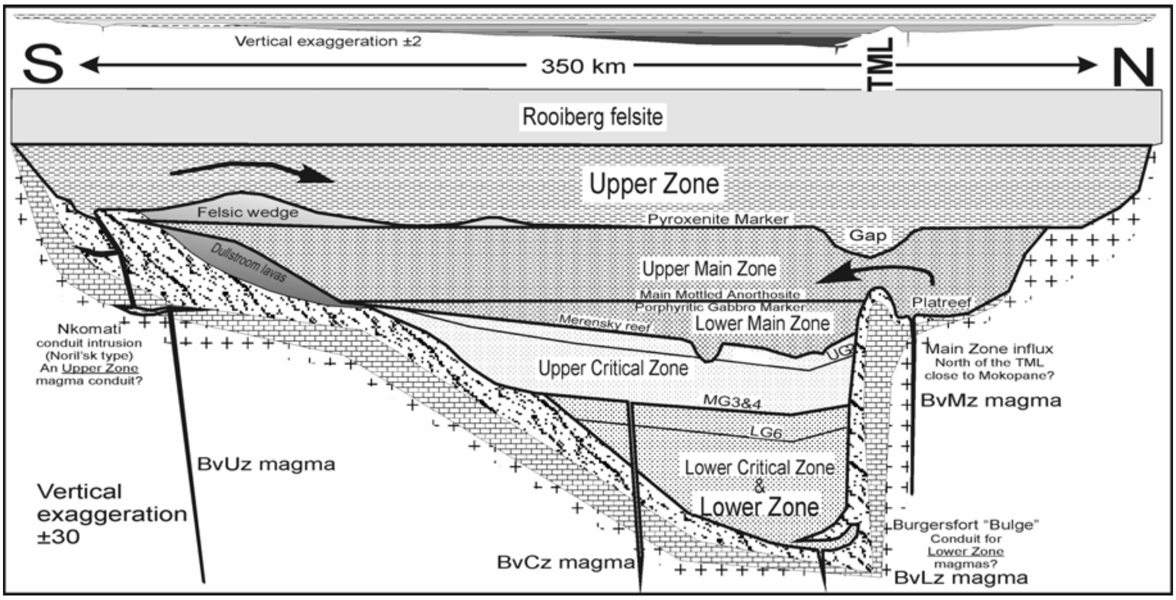


Figure 21 Schematic N-S section through the Bushveld Complex illustrating the "half-graben" geometry and the importance of the Thabazimbi-Murchison Lineament as a structural feature. Note the increasing lateral extent of the different zones from the bottom up (Kruger, 2004).

Figure 22 Johan (Moose) Kruger demonstrating the shape of the Bushveld Complex!

The Rustenburg Layered Suite has been subjected to extensive palaeomagnetic investigations. A summary of the data is presented in Eales *et al.* (1993). The cumulates of the Critical and Main Zones acquired their remanent magnetization with the igneous layering in the horizontal position, with later down-warping of the crust to produce the present dips. This post emplacement down-buckling is seen as an isostatic response of the crust to the load created by the high density mafic rocks. The palaeomagnetic pole positions for the different zones of the Complex indicate age differences between the zones, which precludes the emplacement and crystallization of the complex from a single large volume of magma. The different pole positions between the zones, imply a series of magmatic pulses occurred. Isotopic evidence suggest new magma pulses associated with each chromitite layer in the Critical Zone (Kinnaird et al, 2002). Continuity of the limbs has been the subject of debate (Cousins, 1959, Meyer and De Beer, 1987), but Cawthorn and Webb (2001) argue that the gravity data is consistent with continuity of the eastern and western limbs and previous models failed to take into account the isostatic re-adjustment.

Evolution of the Bushveld Complex

The Bushveld Igneous Complex gives an overall impression of differentiation (with Fe-enrichment) as suggested for Skaergaard. The initial cumulates are ultramafic in nature (magnesian orthopyroxenites and harzburgites, followed by a noritic sequence, a gabbronoritic sequence and eventually magnetite-bearing gabbros and ferro-diorites. However, Early Sr-isotopic data were instrumental in demonstrating the multiple intrusive nature of the Bushveld Complex, and the close association of intrusion with mineralisation, particularly at the level of the Merensky Reef (Kruger and Marsh, 1982; Kruger 1992). Numerous magma influxes dominate the lower part of the stratigraphy, whilst differentiation dominated the upper parts of the stratigraphy. According to Kruger, 1994, the Bushveld Complex as a whole can be viewed as having three main magmatic lineages– the Lower and Critical Zone harzburgite to noritic lineage (with low Sr ratio from 0.705 – 0.7064), the Main Zone gabbronorite lineage (with high Sr ratio c, 0.7082) and the Upper Zone Fe-rich gabbronorite lineage (with Sr ratio c. 0.7075). The boundaries between these major magmatic episodes are major unconformities within the magma chamber coincident with the base of the Merensky Reef and the Pyroxenite Marker (Figure 23). However, isotopic evidence also suggests that during the formation of the Lower and Critical Zones there were repeated influxes of new magma, which expanded the chamber both upwards and outwards (Kinnaird et al, 2002).

During the accumulation of the Lower and Critical Zones, the chamber was continually fed in the eastern and western lobes by olivine- and orthopyroxene- crystallizing magmas that formed the Lower and Critical Zones. Progressive mixing of new and residual fractionated magma resulted in the slow evolution from a harzburgite/orthopyroxenite dominated Lower Zone, through a feldspathic orthopyroxenite dominated lower Critical Zone, to a norite/anorthosite dominated upper Critical Zone. More than one magma type was intruded during this time and may have varied over time from more ultramafic magmas in the initial stages to more noritic magmas in the upper Critical Zone. Interaction of the influxes of new melt with a roof melt, now represented by the granophyric rocks, resulted in variations in isotopic ratio and the production of major chromitite layers. This process is schematically illustrated in Figure 25.

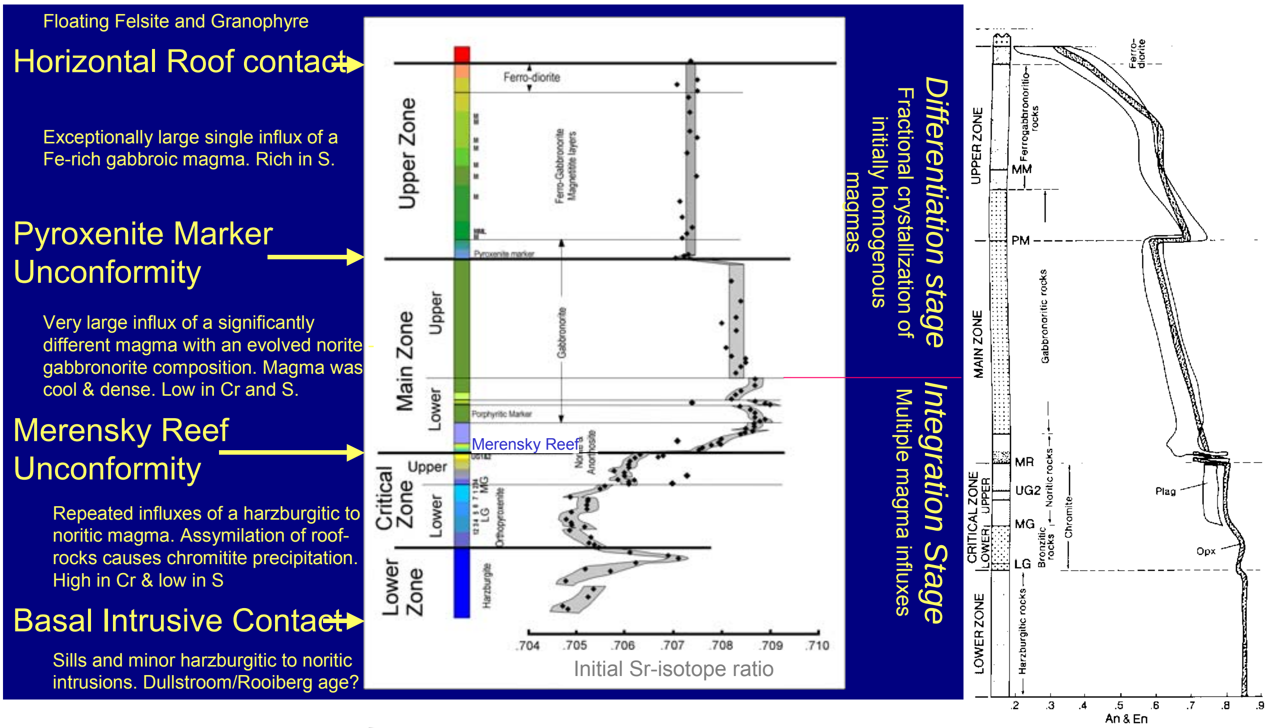


Figure 24 (a) Plot of initial $^{87}Sr/^{86}Sr$ ratio vs. height in the stratigraphy from the western limb of the Bushveld Complex (from Kruger, 1994). (b) Plot of En and An of the layered rocks from De Wit & Kruger (1990). The same general trends are present in the eastern and western lobes of the Bushveld Complex.

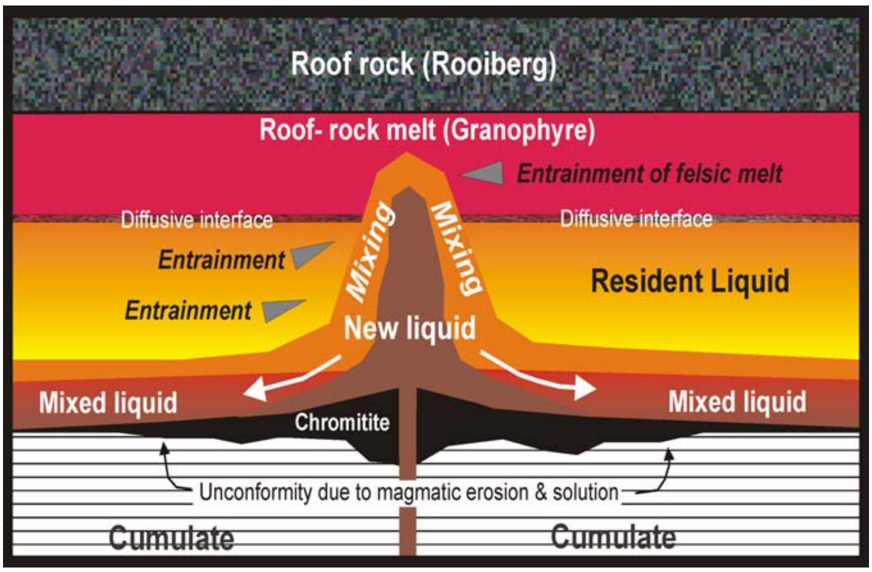


Figure 25 Schematic diagram of chromitite formation resulting from a fountain of magma into the chamber that partially melts roof rocks causing contamination and mixing (Kinnaird et al 2002)

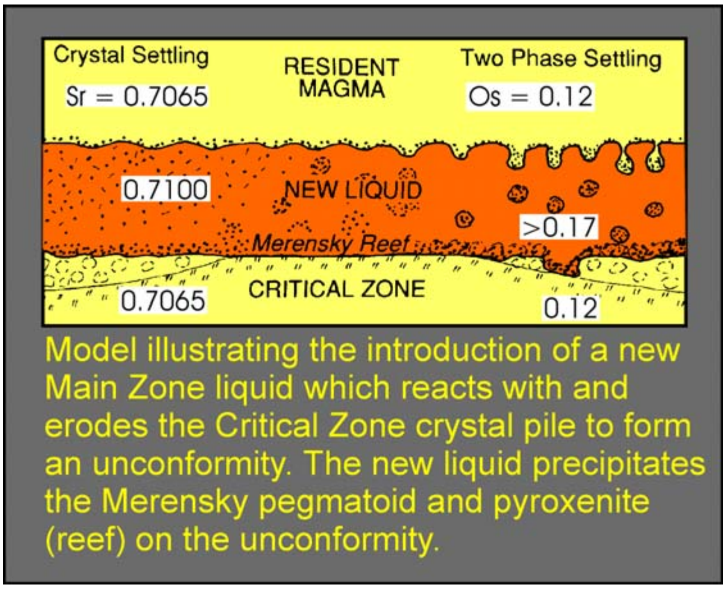


Figure 26 Schematic diagram showing the formation of the Merensky Reef by an influx of hot dense magma which reacted with the floor rocks (Kruger, 1992)

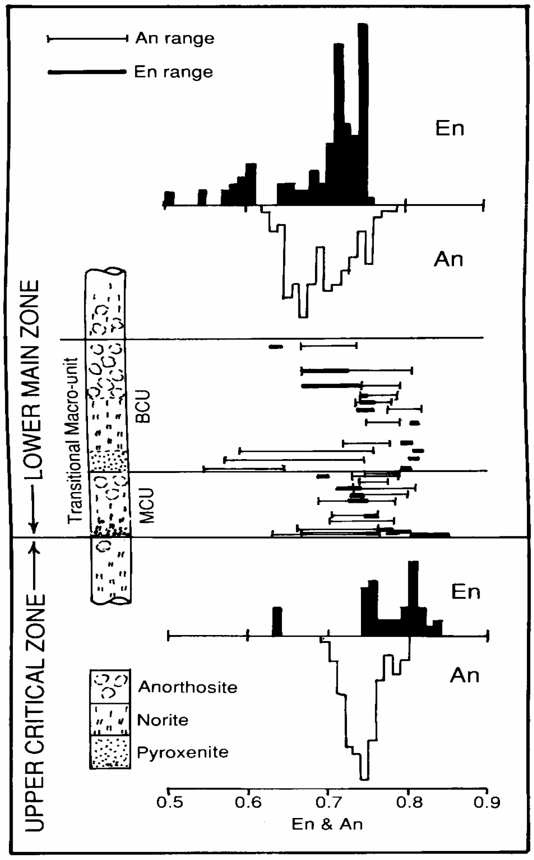


Figure 27 Orthopyroxene En and plagioclase An values for Bushveld Igneous Complex cumulates from the upper Critical Zone and lower Main Zone. Electron probe data from Kruger (1983) and Mitchell (1986).

Further evidence that this magma was cooler and denser than the resident magma is that this new magma was also more Fe-rich and Ab-rich as indicated in Figure 27.

After the precipitation of the Merensky and Bastard cyclic units, the new magma continued to flow into the chamber and concomitant crystallization produced the lower Main Zone. The magma then ceased to flow in and mixed thoroughly and fractional crystallization proceeded in the Upper Main Zone.

The final and largest influx of the Bushveld Complex was initiated just below the Pyroxenite Marker, which like the Merensky Reef represents a major unconformity in the magma chamber. Isotopic data (Figure 24a) indicate that the Upper Zone represents a single influx of a new magma type that mixed completely with the resident magma in a 6:4 proportion. This mixed magma then differentiated *in situ* without any further addition (Cawthorn *et al.*, 1991). The cryptic and modal layering evident in the Upper Zone (including the magnetitite layers) developed from this initially well mixed magma layer (Kruger *et al.* 1987).

The presence of magnetitite and anorthosite layers in the Upper Zone represents the largest magmatic density difference of any system. Crystallization of these two phases may well result in double diffusive convection (DDC.) in the chamber. A possible DDC model for the crystallization of the vanadium-bearing titaniferous magnetitite layers is presented below (from Kruger and Smart, 1987) and utilizes the data of Cawthorn and MacCarthy (1980).

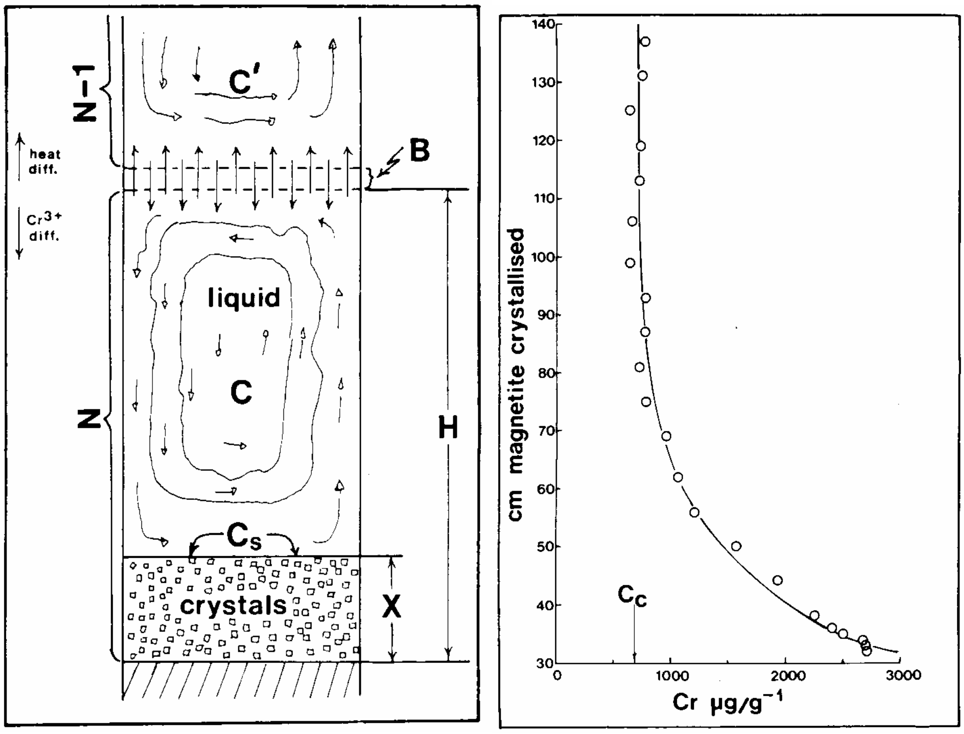


Figure 28 Schematic double diffusive convection model for the crystallization of the magnetitite layers of the Upper Zone (Kruger and Smart, 1987).

Given the sill-like nature of the Bushveld, new influxes of magma spread out laterally. For the Main Zone and Upper Zone magmas, this resulted in an increase in lateral extension as well as expansion upwards. However, during the Critical Zone development when chromitites formed, the pulses of magma were much smaller and the depth of magma and residual liquid is likely to have been much thinner. On the margins of the intrusion, the expansion of the chamber is visible as onlapping relationships (Fig. 1). Nevertheless, the BC is a sill-like lopolithic intrusion with a very large lateral to vertical aspect ratio of > 44:1, and the eastern and western "lobes" where the Lower and Critical Zones are present were interconnected from the start.
For a recent view of the filling of the chamber see http://www.wits.ac.za/geosciences/egri

<u>**Mineralisation in the Layered Suite**</u>

The Rustenburg Layered Suite contains enormous resources of chromite, platinum, palladium, rhodium and vanadium. The location of the major deposits is shown in Figure 29. For an overview of the PGE deposits of the Bushveld see Cawthorn et al, 2002a. In addition, tin and fluorite deposits associated with the Granite Suite have also been important in the South Africa economy. Increasingly, it is becoming apparent in the Bushveld complex that the PGE-bearing horizons such as the Merensky Reef, UG-2, and the Platreef are each the result of multiple introductions of mineralisation.

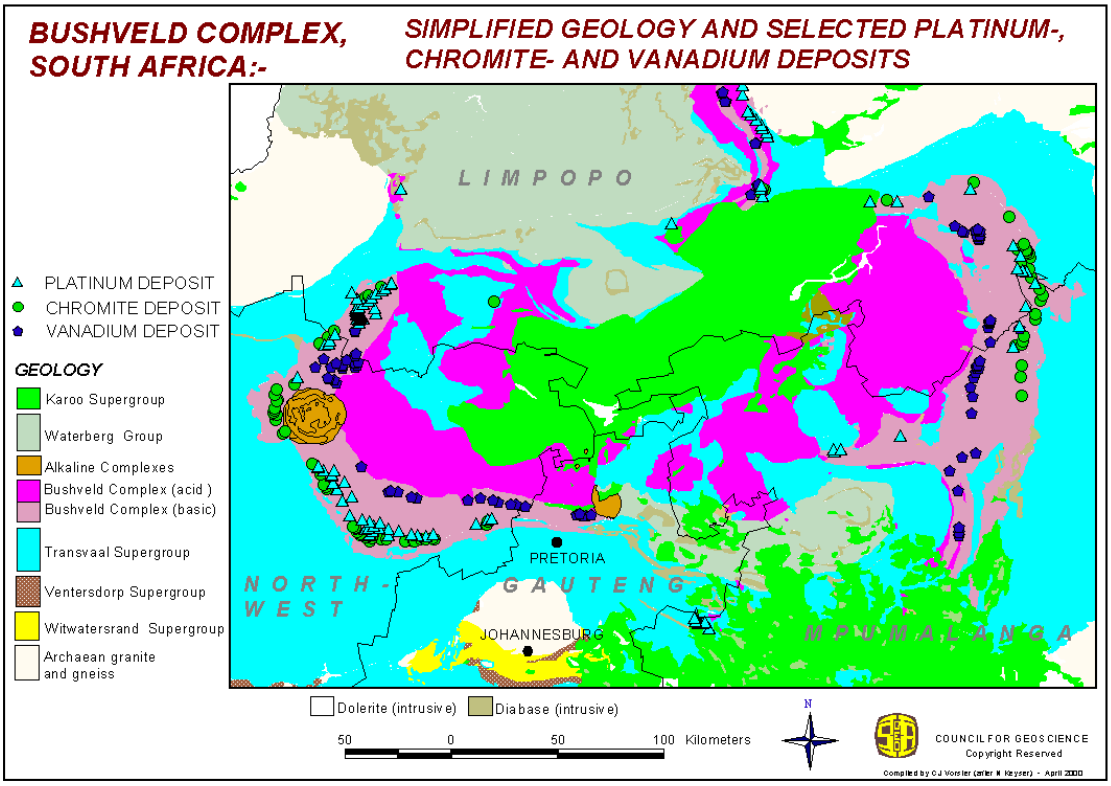


Figure 29 Simplified geological map of the Bushveld Complex showing the location of the major platinum, chromite and vanadium mines.

Chromitites

Chromitite layers occur in three stratigraphically delineated groups (Cousins and Feringa, 1964) each composed of several layers and numbered from the base upwards.

The Lower Group (LG) consists of <7 layers hosted in feldspathic pyroxenite, the thickest being the LG6 layer, also known as the Steelpoort seam (Schürmann et al., 1998). The LG1 to LG4 chromitites are often associated with olivine, a feature that distinguishes them from overlying chromitites, which are invariably olivine-free. The Middle Group comprises 4 chromitites, although multiple layers of each may develop, The MG group straddles the boundary between the lower Critical Zone and the Upper Critical Zone and by convention, MG1 and MG2 lie below the first anorthosite, and MG3 and MG4 are above. The Upper Group usually consists of two chromitites (UG1 and UG2), although in the eastern Bushveld, UG3 and UG3a layers are also recognised. The UG1 is world renowned for the bifurcations of individual chromitite layers although locally the UG2 and LG5 may also show significant development of bifurcations (Nex, 2004). In any given locality, there may be a significant variation in number of chromitite layers within a vertical package, and in one borehole core, 30 individual chromitite layers can be distinguished between LG5 and MG4 (Kinnaird et al., 2002).

There is a considerable lateral variation in the occurrence and thickness of the chromitite layers in the different lobes and also between the northern and southern portions of the eastern and western lobes. Cawthorn and Webb (2001) imply continuity of individual layers over more than 300 km. Although single layers can be traced for more than 70 km in both eastern and western lobes, it may be the package of chromitite that is laterally continuous over long distances, rather than an individual layer (Kinnaird et al., 2002).

Throughout the Bushveld Complex, the chromium content of the chromitite layers decreases upwards. The LG6 has a Cr_2O_3 content of 46–47%, MG chromitites have 44–46% Cr_2O_3 and the UG2 layer has around 43% Cr_2O_3 (Schűrmann et al., 1998). This is reflected in the upward decrease in Cr:Fe ratio; the LG6 layer has a Cr:Fe ratio of between 1.56 and 1.6, MG chromitites are between 1.35 and 1.5 whilst the UG2 layer has a Cr:Fe ratio of between 1.26 and 1.4. Chromite grains, which vary in size from <50 microns to > 2mm, exceed 50% of the mineral assemblage in a chromitite, while the interstitial minerals change from mainly orthopyroxene in the lowest LG group to orthopyroxene and plagioclase in the MG's to dominantly plagioclase with minor orthopyroxene in the uppermost UG chromitites. A poikilitic texture is frequently developed in the chromitites, where oikocrysts of pyroxene or plagioclase enclose chadacrysts of very fine-grained chromites. Accessory minerals include clinopyroxene, biotite/phlogopite, chlorite, talc, quartz, carbonates, sulphides and platinum group minerals.

A number of different models have been put forward for the formation of thick chromitite seams, based on evidence not only from the Bushveld Complex but also from Stillwater in particular. Previous models for chromitite formation include:-

(i) Gravity-induced separation, crystal sorting and settling, (Wager and Brown, 1968) has been discounted both on textural evidence (Eales and Reynolds, 1986), on the basis of co-tectic proportions (Eales and Cawthorn, 1996), and on the physics of processes in non-Newtonian magmas;

(ii) immiscibility of Cr-rich liquid (Sampson, 1932) which has largely been discounted because of the high temperature (c. 1700°C) at which Cr_2O_3-SiO_2 immiscibility occurs;

(iii) increases in oxygen fugacity by country rock degassing (Cameron and Desborough, 1969) seems unlikely because of the difficulty of controlling such changes over the area of the Bushveld and because oxygen fugacity appears to increase systematically from the lowest LG chromitite layer to the uppermost chromitite layers (Teigler and Eales, 1993).

(iv) contamination by a siliceous component (Irvine, 1975);

(v) mixing between resident and new magma (Irvine, 1977),

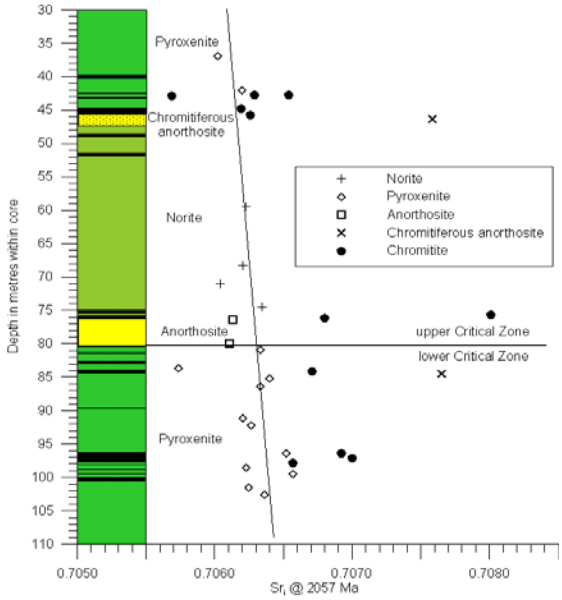
(vi) lateral growth within a stratified magma column (Irvine et al, 1983)

(vii) pressure changes; Cameron (1977) noted that changes in total pressure within a crystallising magma chamber could change the equilibrium liquidus assemblage. The fields of spinel and orthopyroxene expand with increasing pressure over a range of 1 to 10kbars at the expense of the olivine and plagioclase fields, so a pressure increase within a magma chamber could result in chromite, magnetite or orthopyroxene-rich layers, whereas pressure decreases could result in anorthositic or dunite formation. The attraction of this model is that the effects of a pressure change would be felt nearly simultaneously over the whole magma chamber although the magnitude of the pressure change necessary to shift the magma composition from the cotectic into the field of chromite alone is not clear. A pressure change in the order of >> 1kbar would be needed according to Hatton and von Gruenewaldt (1989) and the general effect of pressure change on mineralogy has been shown to be trivial (Hatton, 1984);

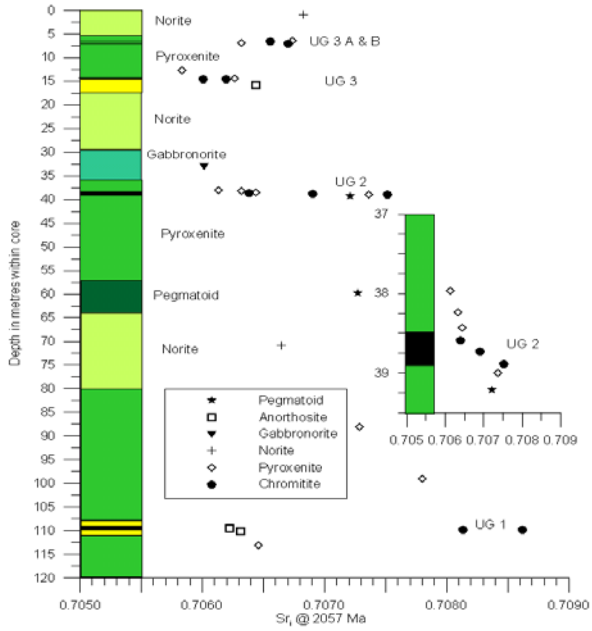
(viii) injection of a chromite-phyric magma (Eales et al., 1990) still requires that chromite is precipitated somewhere else at greater depth in order to be entrained in the ascending magma.

The various merits of these individual models have been extensively discussed in the literature. However, the sharp contacts and remarkable continuity of some of the chromitite layers require that whatever the process, it must have operated at a sufficient scale to affect the whole chamber at certain periods.

Strontium isotope data presented by Kinnaird et al, 2002 (Figure 30), indicate that each chromitite layer is associated with a new influx of magma. Mixing between a primitive and evolved ultrabasic liquid may have resulted in the formation of chromitite layers associated with olivine (LG1-LG4) but for thicker layers associated with orthopyroxene (LG5-MG1) or with orthopyroxene and plagioclase (MG2 and above) mixing between two magmas of very different compositions has been suggested (Irvine, 1977, Irvine et al., 1983). The different magmas proposed for this process have been termed U-type for a magma with a crystallisation sequence of olivine, orthopyroxene, plagioclase, clinopyroxene and A-type for a magma crystallising plagioclase, olivine, clinopyroxene, then orthopyroxene (Sharpe, 1981, 1982; Harmer and Sharpe, 1986) although recent work shows that there is little evidence for an A-type magma (Teigler and Eales, 1996).



(a)



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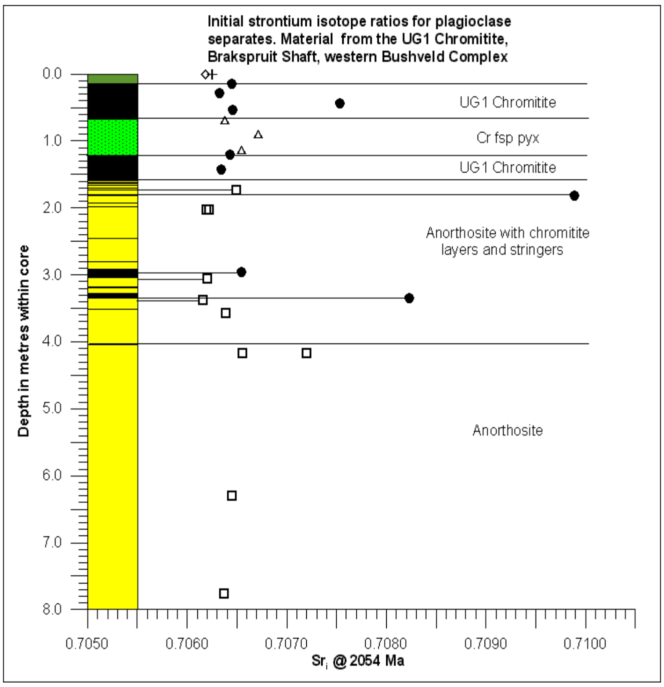


(c)

Figure 30. Initial strontium isotope ratios demonstrating higher Sr$_i$ ratios for plagioclases within most chromitites compared with the host silicates. (a) profile in lower-upper Critical Zone silicates and chromitites from LG5 to MG4. Isotopic trend line illustrates slight upward decrease in initial strontium ratio of silicate rocks from lower Critical Zone to upper Critical Zone. (b) Profile in upper Critical Zone silicates from 10m below UG1 to 14m above UG3. Expanded inset illustrates upward decrease in initial ratios from bottom to top of UG2. (c) Profile through the UG1 chromitite package, footwall anorthosite and hanging wall pyroxenite. (Kinnaird et al., 2002).

PGE mineralisation

Platinum Group element (PGE) mineralisation occurs in well-defined layers in the Merensky Reef and UG2. In addition, the Platreef in the northern limb, which is being mined at Sandsloot for PGE's, is currently the focus for extensive exploration. In the Merensky Reef and UG2 the grade is typically around 7 g/t whereas in the Platreef of the northern limb the grade is generally 4 g/t or less in the mineralized zone (Figure 31).

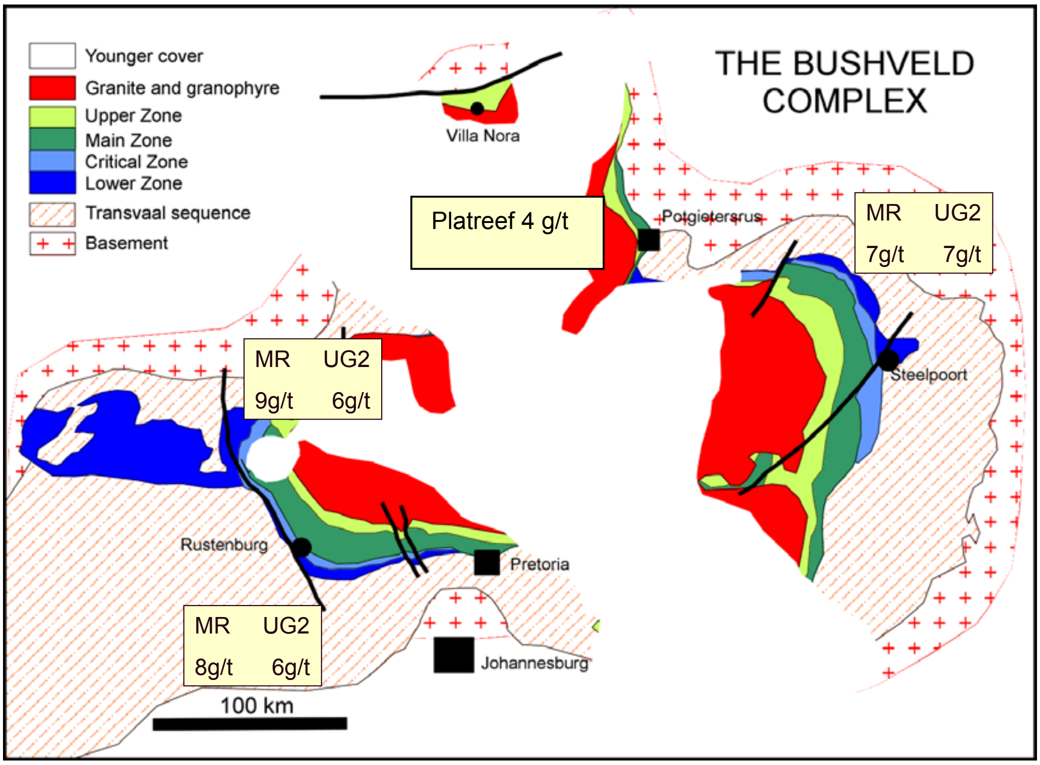


Figure 31. Simplified map of the Bushveld Complex showing generalized PGE grades for the Merensky Reef, UG2 chromitite layer and Platreef.

UG2

The UG2 (Upper Group 2) chromitite layer in the upper Critical Zone is probably the largest PGE resource on Earth although all the chromitite layers contain elevated levels of PGE's. The UG2 occurs 15-400m below the Merensky Reef, with the smallest vertical separation in the western and greatest in the eastern Bushveld (Lee, 1996). The layer is 0.5 –1 m thick generally with a pegmatoidal feldspathic pyroxenite footwall, and more rarely anorthosite. Potholes are a common feature of the UG2. Two to four minor chromitite leaders occur in the hanging wall. The chromite content is 60-90%, with an average Cr/Fe ratio between 1.26-1.4 with 43.5% Cr_2O_3. The PGE are interstitial to the chromite grains and the only PGM commonly enclosed by chromite is laurite. PGE contents are up to 10 ppm PGE+Au (3.6 ppm Pt, 3.81 ppm Pd, 0.3 ppm Rh) Cu and Ni are low generally less than 0.05% and the amount of accessory base metal sulphides is low (Lee, 1996). There are frequently two peaks in the PGE distribution (Hiemstra, 1985). The Pt:Pd ratio varies with geographic location.

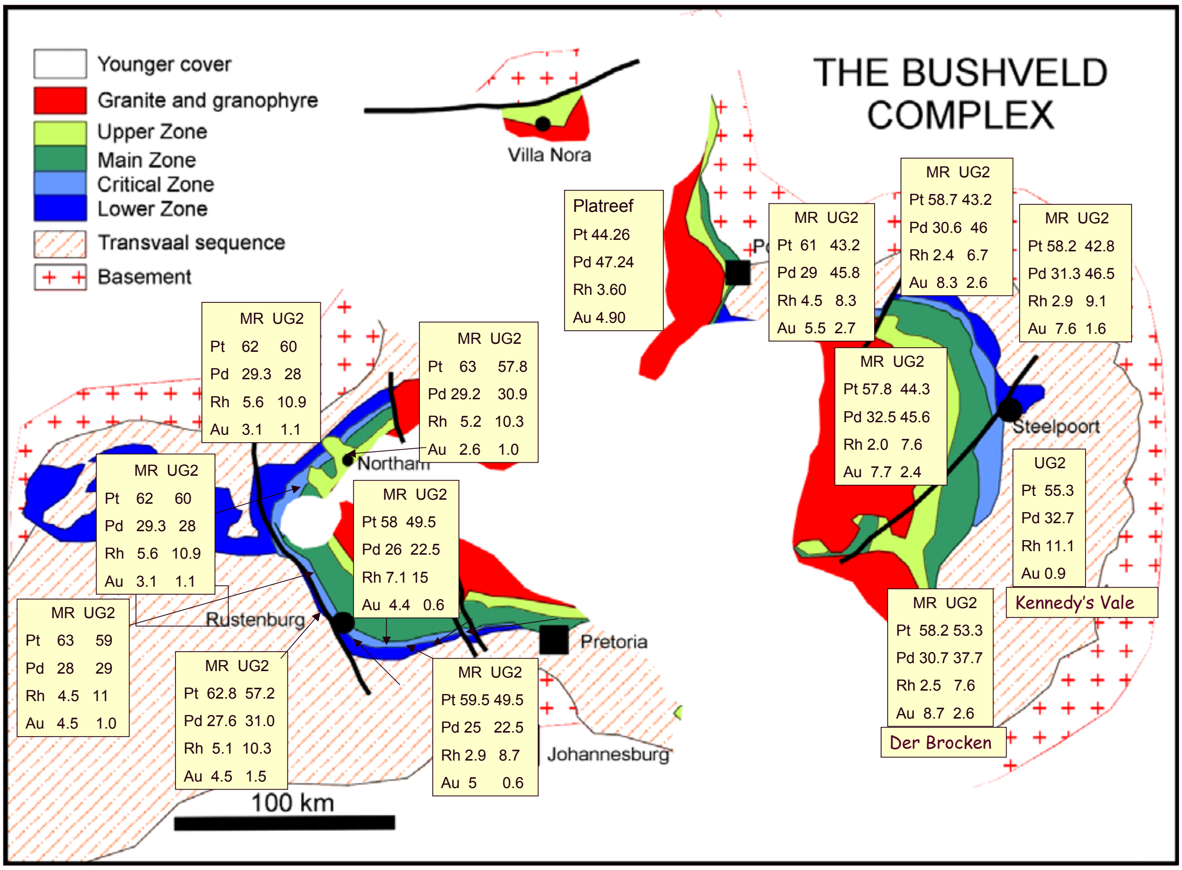
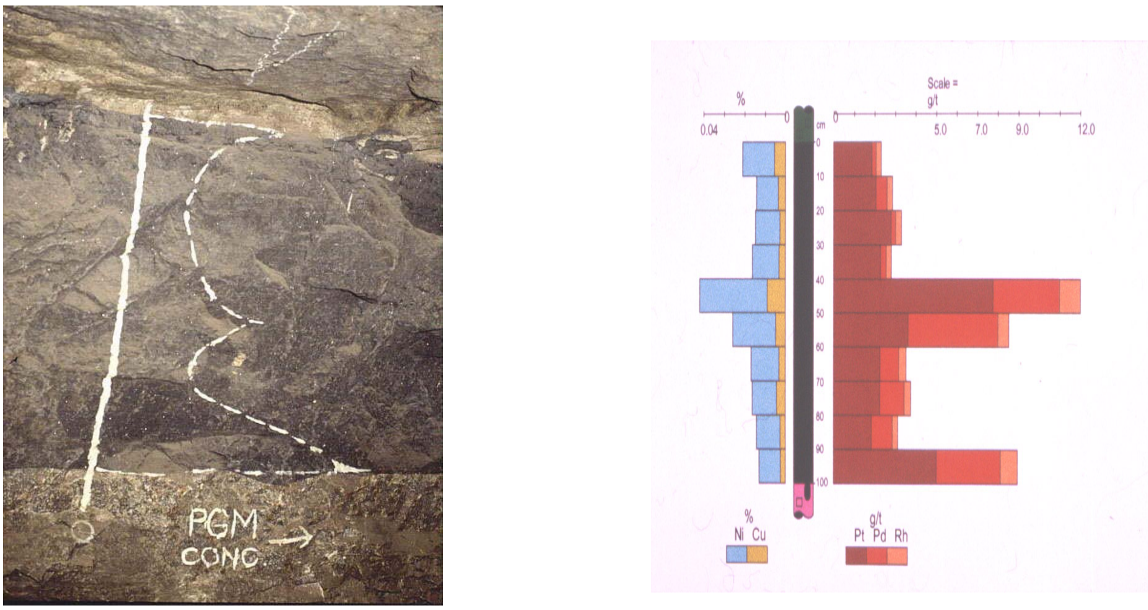


Figure 32 Variations in percentage of 3PGE+Au splits around the Bushveld Complex. Note that the Pt:Pd proportion for the Merensky Reef is approx. 2:1. For the UG2, the Pt:Pd proportion is also approx. 2:1. except for north of the Steelpoort Fault where it is approx. 1:1, Source: Platinum Map of southern Africa Barker and Associates, 1st Edition 2002.



(a) (b)

Figure 33 (a) photograph of the UG2 underground at Lonmin, east of Rustenburg in the Marikana section. (b) Grade of Ni and copper (left) and PGE (right) through this UG2.

Cawthorn et al. (2004b) drew attention to Hiemstra's (1985) section through the UG2 at Western Platinum Mine (Figure 3), pointing out that this can be interpreted in terms of 3 sequences of mineralisation (A, B and C in figure 3), in which the grade decreases upward in each cycle, and in which sequence 3 has a much higher Pt/Pd ratio than A and B. The overall Pt/Pd ratio for the UG-2 is 2.1, but the Pt/Pd for sequences A and B alone is 1.6. Cawthorn et al. (2004b) noted that in the part of the Eastern Bushveld north of the Steelpoort fault, the UG-2 is thinner (60-80 cm instead of the normal 100-120 cm) and has a low Pt/Pd ration of about 1.4. However, in this area, an additional upper group chromitite layer is present, the UG-3, which is about 40 cm thick and has a very high Pt/Pd ratio. South of the Steelpoort fault, the UG-3 is missing, and the UG-2 has a thickness (100-120 cm) and Pt/Pd ratio (≅ 2) similar to that in the west. They suggest that south of the fault and throughout the west, the interval of rock separating the UG-2 and UG-3 north of the fault is missing (either due to non-deposition or erosion) and that the UG-3 lies directly on top of the UG-2, forming that part represented by sequence C in Hiemstra's section. Hornsey (2004) described a "split reef facies" in part of the UG-2 of in the Two Rivers area (Dwars river bridge area south of the Steelpoort fault) in which the lower 2/3 of the UG-2 has a Pt/Pd ratio ≅ 1.3 and is separated by up to 6 m of pyroxenite or norite from the upper part which has a Pt/Pd ratio ≅ 3. [4]. Note the three distinct cycles, starting with high total PGE at the base, decreasing upward, and that each cycle has its own discrete Pt/Pd value, with the uppermost cycle having the highest value. (adapted by Cawthorn et al., 2004b from Hiemstra 1985).

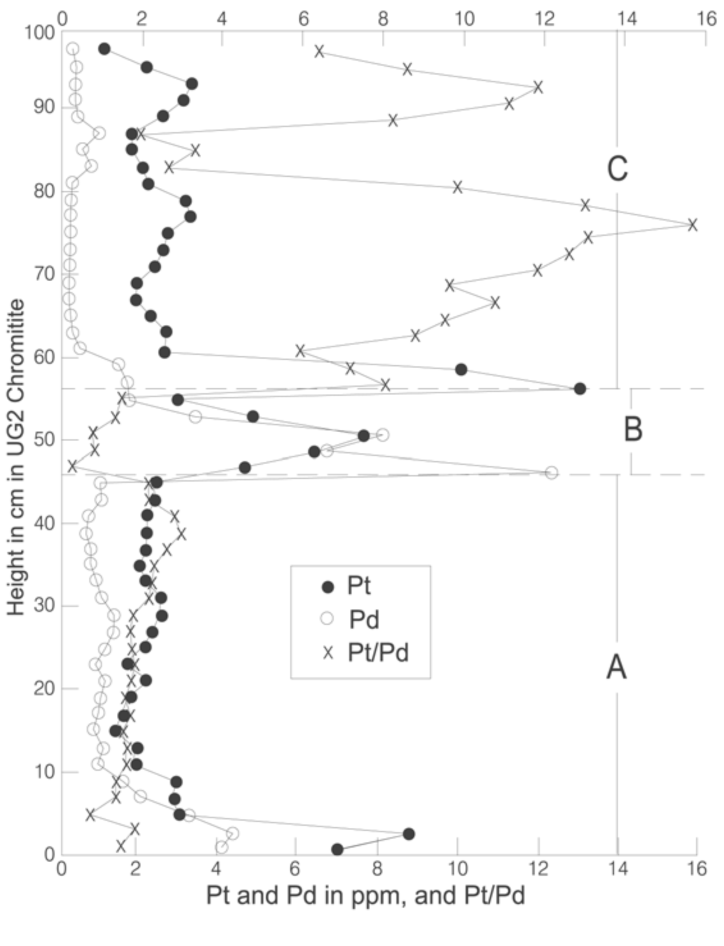


Figure 34 Data for Pt, Pd and Pt/Pd in a very detailed profile through the UG2 chromitite from Western Platinum Mine (Cawthorn 2004b after Hiemstra, 1985).

Merensky Reef

The Merensky Reef has been the world's most important source of platinum since exploitation commenced in 1928 although the UG2 will become the major resource for the future. The term Merensky Reef, is a mining term, used to designate the best cut of an economically mineralised package of rock (Lee, 1996). The Merensky Reef can be traced for >280 km around the complex, about 140 km in both the eastern and western limb (Cawthorn et al, 2002b). It ranges in thickness from 4 cm to 4m, although commonly around 1 m and in general the reef shows an inward dip towards the centre of the Complex between 8° and 27° although in a small area in the extreme northern portion of the eastern limb it dips up to 65°. Seismic surveys show reflectors correlated with the position of the Merensky Reef that can be traced as far as 50km down-dip of outcrop (Lee, 1996). Extraction is currently taking place along 100 km of strike.

In general, according to Lee (1996), the reef is composed of a texturally heterogeneous pegmatoidal feldspathic pyroxenite (Figure 35), partially pegmatoidal feldspathic pyroxenite or feldspathic pyroxenite. The rock is an orthocumulate consisting of a framework of 70-90% very coarse-grained subhedral to euhedral orthopyroxene and up to 30% intercumulus plagioclase. Clinopyroxene oikocrysts up to 3 cm long occur and mica is a common accessory (Lee, 1996). Two to four thin chromitite layers (1-2cm) commonly define the upper and lower limits of mineralisation and the highest grades are associated with these chromitites. The footwall is either anorthositic, or less commonly feldspathic pyroxenite or harzburgite. A thin anorthosite usually occurs below the lower chromitite when plagioclase cumulate is the footwall lithology (Lee, 1996). Variations in the profile through the Merensky Reef are shown in Figure 36. The hanging wall is generally a norite that grades upwards into anorthosite of the overlying Bastard Cyclic Unit, so called because it is similar to the Merensky Reef but lacks mineralisation.

Figure 35. Pegmatoidal Merensky Reef with anorthosite footwall (left) and hanging wall of norite. The chromitites, less than 1 mm thick mark the hangingwall and footwall contacts in this sample.

Up to 3% base metal sulphides (pyrrhotite, pentlandite, pyrite, cubanite and rare sulpharsenides, galena and sphalerite) and accessory PGE-minerals are interstitial to the silicates. An average grade of the Merensky Reef is typically 5-7g/t. The proportions of the precious metals are 4.82 ppm Pt, 2.04 ppm Pd, 0.66 pp Ru, 0.24 ppm Rh, 0.08 ppm Ir, 0.26 ppm Au, and the Cu:Ni ratio is 0.61. (Lee, 1996). The extent and relative amount of PGE and base metal sulphides appears to be a function of reef thickness with the highest grades occurring where the reef is thin. Whereas the grade of the reef is remarkably constant over extensive strike distances, the composition of the actual platinum group mineralogy is extremely variable even from mine to mine (Cawthorn et al., 2002b).

Where the Merensky Reef abruptly transgresses footwall rocks, potholes are developed. These may interrupt the normal mining of the reef. A range of Merensky Reef types has been documented by Kinloch and Peyerl (1990) based on reef thickness, whether pot-holed or pegmatite-replaced, the composition of the footwall rocks and the PGM assemblages.

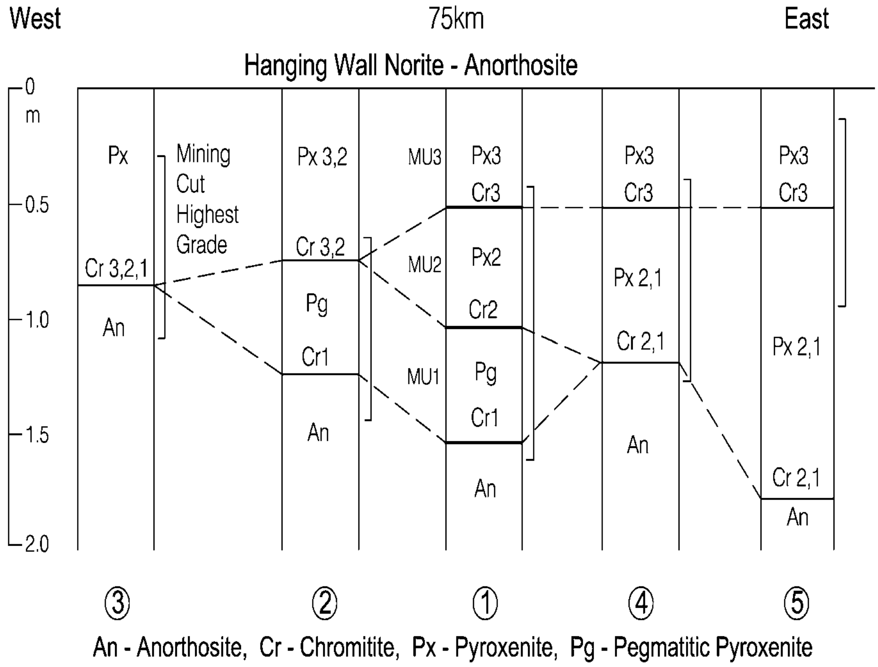


Figure 36. Section through Merensky Reef from west to east from Impala (3), Rustenburg (2), Karee (1), Western (4) and Eastern (5) Mines. (after Cawthorn et al., 2004a)

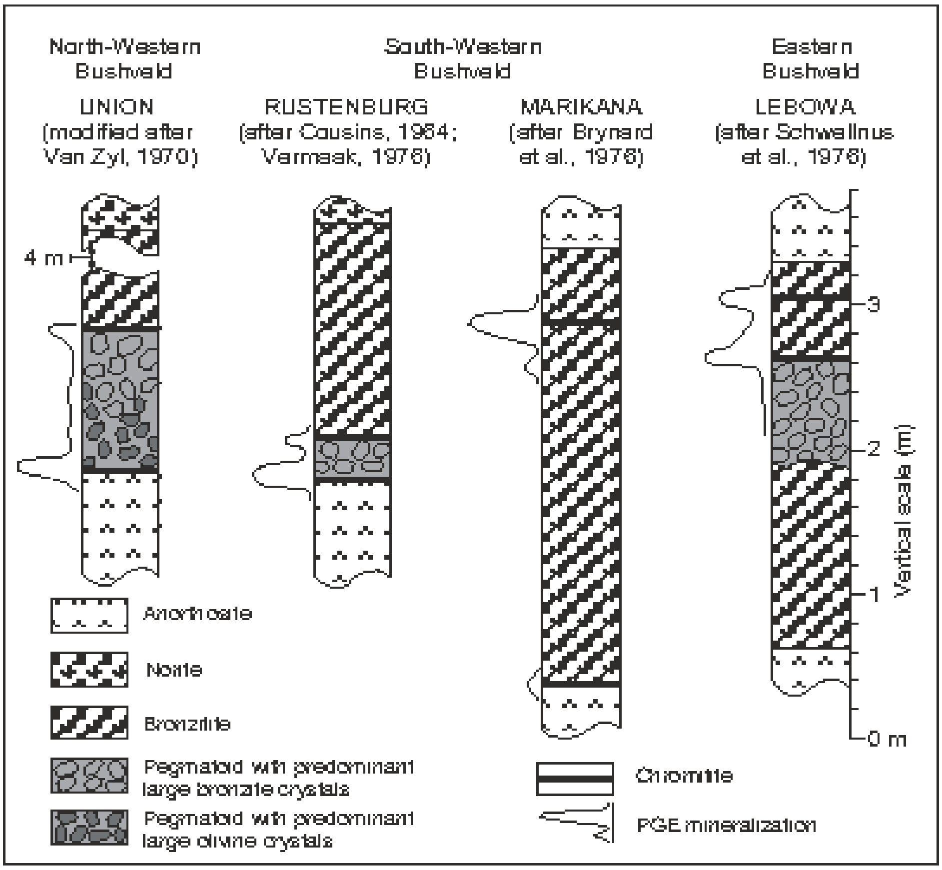


Figure 37. The nature of the Merensky Reef at different locations within the Bushveld Complex (from Naldrett, 1989)

Recently, Cawthorn et al. (2004a) proposed that the Merensky cyclic unit formed as the result of the introduction of 3 pulses of Main Zone magma (MZm) which displaced the resident Critical Zone magma (CZm) upward. After each introduction, plagioclase was deposited from the new MZm, chromite and pyroxene settled from the displaced Critical Zone magma. In different areas, cumulates from the earlier 2 pulses were either partially or completely eroded by the final pulse. It was after the introduction of this pulse that PGE-enriched sulphides settled for about 100 cm through the underlying cumulates from the overlying CZm to form the principal mineralisation. While Cawthorn et al. (2004a) considered that the bulk of the mineralisation accumulated after introduction of the third pulse, the PGE distribution within thicker sections of the Merensky Reef (see the Marikana and Lebowa columns in Figure 37) indicates that some PGE accumulated in conjunction with chromite layers that formed from earlier magma pulses.

Platreef

The Platreef in the northern limb of the Bushveld Complex is predominantly a pyroxenitic PGE-Cu-Ni-bearing package with a hanging wall of Main Zone gabbronorite and a footwall of Transvaal Supergroup in the south and Archaean granite and gneiss in the north (Figure 38). The Platreef varies from 400 m thick in the S to <50 m in the north (Kinnaird et. al 2005). Although the overall strike is NW or N, with dips 40-45°W at surface, shallowing down dip, the overall geometry appears to have been controlled by irregular floor topography. There are basement highs on southern Macalacaskop and on southern Turfspruit with thinned Platreef on the flanks and thick reef in the intervening basins. In the southern basin, it is 400 m thick, dips are 32°NE on the S side and 47°SW on the N side. In the Turfspruit basin, the Platreef is 250 m thick with 40°inward dips (Kinnaird et al 2005). Faults offset the strike of the Platreef: a N-S, steeply dipping set is predominant with secondary ENE and ESE sets dipping 50-70°S. The fault architecture was pre-Bushveld and locally controlled thickening and thinning of the succession, especially in the south.

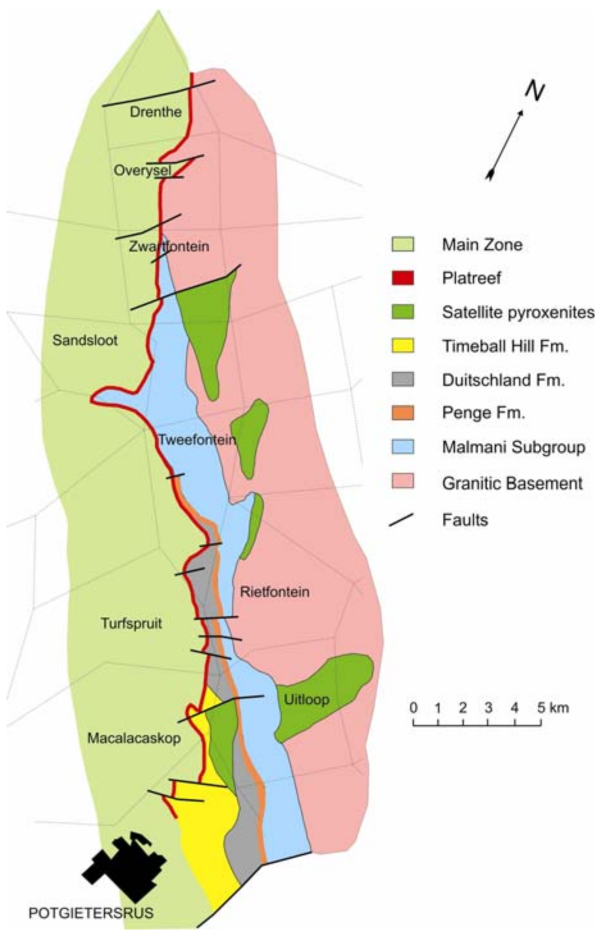


Figure 38. Geological map of part of the northern limb, showing farm boundaries.

The Platreef is heterogeneous and although predominantly pyroxenitic includes peridotites and norite cycles with anorthosite in the mid to upper portion of the Platreef in the south (Kinnaird et al, 2005). Zones of intense serpentinisation may occur throughout the Platreef. Country rock xenoliths <1500 m long, are common. In the south, these are typically hornfelsed banded ironstones, shales, mudstones and siltstones whereas further north the xenoliths are typically dolomitic or calc-silicates.

The Platreef generally shows lower PGE grade than the Merensky Reef and UG-2 chromitite. Overall grade is typically 4g/t where footwall is dolomite, but commonly 1-2 g/t elsewhere with intersections sometimes >10g/t. Grade may be bottom loaded e.g. Tweefontein, top-loaded e.g. Drenthe or be evenly distributed e.g. Overysel. Different styles of mineralisation occur within different sectors of the Platreef. For example, immiscible droplets accumulated in some structural traps or footwall depressions, good PGE grade is associated with dolomitic xenolithic rafts and zones of serpentinisation, and skarn-type mineralisation occurs in the central sector where dolomite forms the footwall (Armitage et al, 2002). Sulphides may reach 20% in some intersections, with overall grades of 0.1-0.6% Cu and Ni. Massive sulphides are localised, commonly, but not exclusively towards the contact with footwall metasedimentary rocks. Magmatic sulphides are disseminated or net-textured ranging from a few microns to 2 cm grains of pyrrhotite and pentlandite with chalcopyrite and minor pyrite. Much of the sulphide is associated with intergranular plagioclase, or quartz-feldspar symplectites, along the margins of rounded cumulus orthopyroxenes. Composite grains of pyrrhotite, pentlandite and chalcopyrite are rimmed by, or associated with other trace sulphides including galena and sphalerite. The PGE's occur as PtFe, Pt_3Sn and variable Pd or Pt-tellurides, bismuthides, arsenides, antimonides, bismuthoantimonides and complex bismuthotellurides. Pt:Pd ratio is ~1. Mantle-normalised metal patterns are shown in Fig. 1. PGM are rarely included in the sulphides. They occur as micron-sized satellite grains around interstitial sulphides and are common in serpentinised zones. PGM's also occur within xenoliths and in lenses in the Main Zone hanging wall.

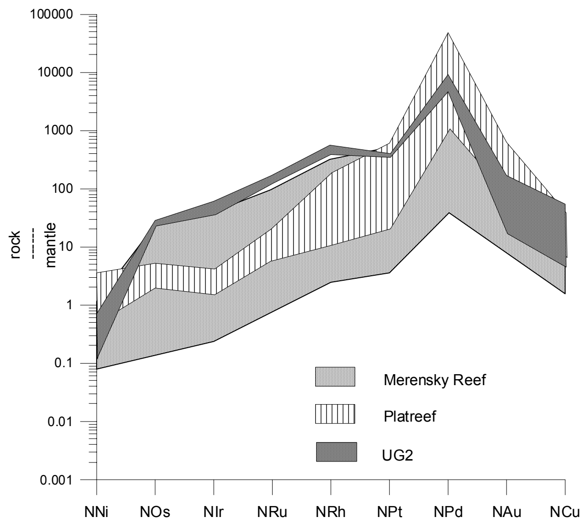


Figure 39. Mantle-normalised metal patterns for the Merensky Reef, UG2 chromitite and Platreef (Kinnaird et al, 2005)

Most of the previous literature regards the Platreef as a single body. Recent geochemical and lithological work in the southern Platreef indicates that it resulted from several pulses of magma, each of which is characterized by a package of rocks with distinctive geochemical characteristics and Pd:Pt ratios, with differing sulphide textures and proportions. In addition, there have been a series of processes that have modified the original magmatic PGM distribution and chemistry. Often, the different intrusive pulses are separated from each other by a slab of hornfels country rock, or by serpentinite or parapyroxenite or pegmatitic norite (Kinnaird et al., 2005).

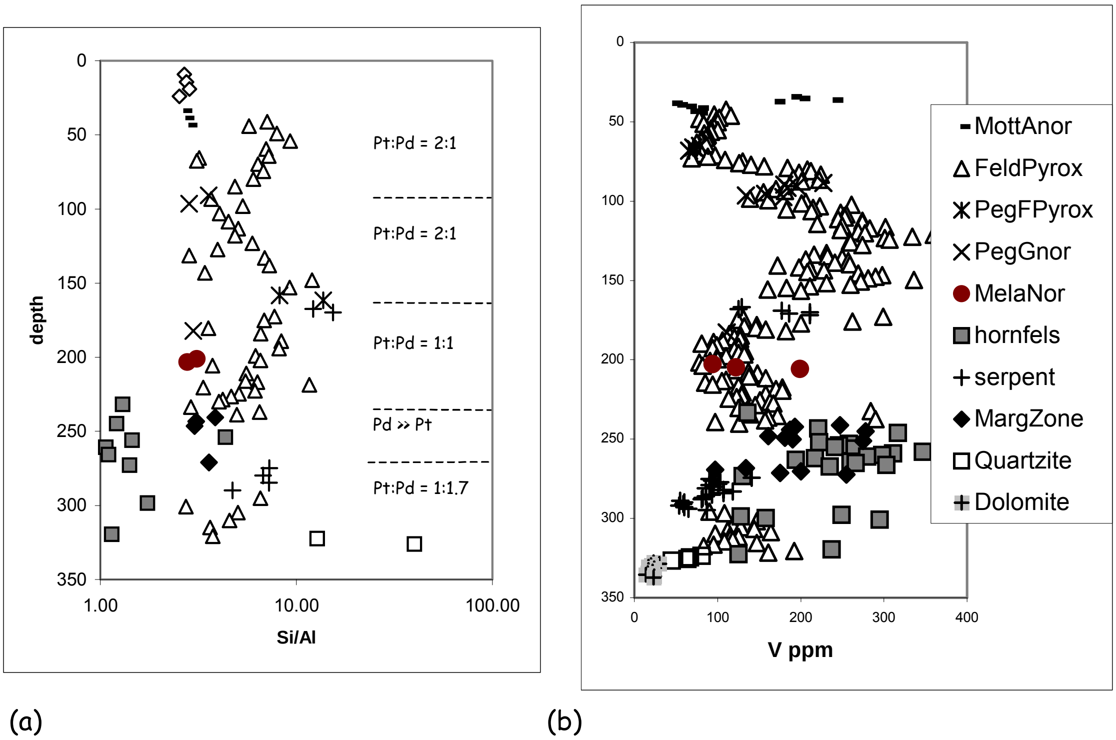


Figure 40. (a) Section through the southern Platreef showing various intrusive packages that are characterised by differing Si/Al and Pt:Pd ratios. (b) Variations in vanadium content with depth (Kinnaird et al, 2005)

Core ATS57 from Turfspruit on the southern Platreef shows clearly the lithological and geochemical variations of several intrusive pulses. A 'raft' of cordierite spinel hornfels, which may still be partly attached to the floor, occurs *c.* 250 m depth (Figure 40). A fine-grained Marginal Zone facies, above and below the hornfels raft is regarded as an early intrusive phase that is distinguished from the Platreef by higher Al and lower Cr. The Marginal Zone is not chilled Main Zone as it has much lower Al and Ca, higher Si and Mg but lower Mg# than Main Zone lithologies. Elsewhere this Marginal Zone may occur as a contact facies with footwall rocks. Serpentinised harzburgite, with up to 2250 ppm Cr occurs below the hornfels raft. This is regarded as equivalent to Lower Zone or Lower Critical Zone of the Bushveld Complex further south. In contrast, the serpentinite at approx. 170m depth is graphite-bearing, has shale fragments within it and is associated with pyroxenite, parapyroxenite and lenses of calc-silicate. This is a highly altered xenolith of footwall dolomite. It contrasts with the serpentinised harzburgite at 270-280 m depth by lower Cr content, higher Ca/Al and Si/Al ratios.

Several feldspathic pyroxenite packages occur. The lower pyroxenitic package beneath the hornfels raft is compositionally different from the pyroxenite above the raft. This lower pyroxenite carries the highest PGE and sulphide abundances (pyrrhotite, pentlandite and chalcopyrite) with coarse sulphides at the top of the unit grading downwards into disseminated sulphides. In other cores this lower pyroxenite unit becomes more noritic upwards.

The pyroxenite above the hornfels has lower Mg# with PGE and sulphide enrichment only at the base. Sulphides are coarser than in the unit below and PGE grades are slightly lower. The succeeding pyroxenite package has the highest vanadium, and lowest PGE content of the all the pyroxenites with Pt:Pd ratios of approximately 1. The uppermost pyroxenite is characterized by a higher Cr content, higher Mg#, and the highest Pt:Pd ratio.

There are two mineralised zones in core ATS57, one at the base of the lowest pyroxenite package, the other at the base of the pyroxenite above the hornfels raft, with slightly enhanced values of Pt at the base of the uppermost pyroxenite. In the mineralised zones, sulphides may locally reach 20%. Pt-Pd abundance may be decoupled from sulphur and base metal sulphides. Although PGM may occur as micron-sized grains around sulphides they occur rarely in the sulphides themselves. Instead, PGM's are common as discrete grains within talc, tremolite and serpentine or associated with intergranular plagioclase, or quartz-feldspar symplectites, along the margins of rounded cumulus orthopyroxenes. The PGE's occur as Pd or Pt-tellurides, bismuthides, arsenides, antimonides, bismuthoantimonides and complex bismuthotellurides.

It is envisaged that the earliest intrusive phase was equivalent to Lower or Lower Critical Zone magmas. This intruded within the Transvaal metasedimentary sequence as sills. As further pulses followed, some metasedimentary material became detached from the floor and later magma flowed under or over these layers. Some of the later pulses also interfingered with the early Platreef pulses. These intrusive pulses are regarded as equivalent to Critical Zone elsewhere in the Bushveld Complex. Although the well-developed chromitite layering of the Critical Zone does not occur, Cr/MgO ratios of the Platreef rocks and Critical Zone of the eastern Bushveld are similar. Later Main Zone norites/ gabbronorites also intruded the Platreef pyroxenitic packages as sills, sometimes incorporating metasedimentary rocks from the top of the Platreef.

The unit between the Main Zone gabbronorites is thus a complex zone of inter-fingered lithologies, including some Main Zone. This whole package is collectively called the Platreef. To add to the complexity, primary sulphides and PGE's were then re-distributed by several later processes.

Figure 41. Anglo Platinum's Sandsloot pit. The Platreef is being exploited for PGE, Cu and Ni. Footwall Transvaal metasedimentary rocks form the footwall (right of picture) while the Main Zone forms the hanging wall (left of pit).

For more details on the Platreef see Armitage et al 2002 and Kinnaird et al 2005. A special issue of the TransIMM on the Platreef will be published in late 2005.

Concluding Comments

In spite of the extensive data available for the Bushveld Magmatic Province, there is no consensus of opinion on several key issues relating to the number, nature, volume and source of the different magma types and the plate setting for magmatism.

Some authors have suggested a link between meteorite impacts and LIPs although only one magmatic province, the 1850 Ma Sudbury Complex in Canada has been conclusively linked with a meteoritic impact, in part based on the evidence of the occurrence of shock features (e.g. Lightfoot and Naldrett, 1994) although this province is of sub-LIP scale (Ernst et al, 2005). The Bushveld Complex has been cited as having been generated by rapid decompression melting at the leading edge of a mantle plume, triggered by the impact of a large (d $\geq$20km, v $\geq$10 k/sec) iron bolide (e.g. Rhodes, 1975, Elston, 1996). Evidence presented for an initial catastrophe is of a high-energy high-temperature debris flow at the base of the Rooiberg Group and from intense deformation bracketed between the end of the Transvaal sedimentation and the basal Rooiberg debris flows although Buchanan and Reimold (1998) refute this evidence. In contrast, Kruger (2002) suggests a back-arc, subduction-related setting for magmatism, whereas Gibson and Stevens (1998) suggest that during the Bushveld event the Kaapvaal Craton experienced extensional and strike-slip reactivation of Archaean structures, consistent with a NE-SW directed extension. They state that the lack of significant pre-Bushveld deformation of the Transvaal Supergroup, together with their preservation over large parts of the craton, indicates a lack of significant erosional exhumation which is a normal consequence of crustal thickening. The occurrence of A-type granites, which are generally associated with crustal extension, is consistent with this hypothesis. The preservation of the volcanic and shallow-level intrusive rocks of the Bushveld Complex indicates that the significant magmatic thickening related to the Bushveld event must have been compensated by concomitant crustal thinning (Gibson and Stevens, 1998).

With regard to the origin of the magmas Kruger (2000) suggests that the Lower and Critical Zone magmas were derived from a mantle source enriched in a seawater-rich subducted component, and that this dehydrated component was itself melted to form the Main Zone magmas. The whole of the Rustenburg Layered Suite is enriched in Si, K and Rb relative to many mafic magmas and Sr^{87}/Sr^{86} and Re-Os isotope are too radiogenic for a purely mantle-derivation of the magmas. This suggests that there was contamination of the magmas by a crustal source. McCandless et al (1999) suggest assimilation of 5% granulitic lower crust to account for the Re-Os radiogenic values. However, according to interpretations of lead isotope data by Kruger (2000), there may be a significant component of upper crustal source, especially for the Main Zone with little evidence that the lower crust contributed to the Bushveld magmas. The remarkably uniform chemistry of individual magma pulses across the Bushveld Complex indicates that whatever the contaminant was, the magma spent sufficient time at deep crustal levels to achieve thorough mixing with the crustal components.

The volume of magma estimated to have been involved in the formation of the Bushveld LIP suggest that optimum conditions for such an event involve the interaction of a mantle plume with lithosphere that has been thinned to between 110 and 50 km (Gibson and Stevens, 1998). Hatton (1995) was the first to suggest a plume-related origin. He envisaged hot Lower Zone magma derived from a mantle diapir which halted in the lower crust, flattening of the diapir led to the melting of the lower crust and the formation of the lower Critical Zone magma. However, according to Kruger (pers. comm) this would preferentially incorporate the lower crust whereas the lead isotope evidence implies an upper crustal component. In addition, evidence of Late Proterozoic to Cretaceous diamondiferous Kimberlites, which contain a c. 3.1 Ga diamond population (Richardson et al, 184, Shirey et al, 2003). indicates that a lithosphere root in excess of 140 km must have existed beneath the craton in the Archaean and that it survived the Bushveld event (Gibson and Stevens, 1998). Gibson and Stevens envisage a model for the Bushveld magmatothermal event, in which a hot juvenile plume reached the base of the Kaapvaal lithosphere at c. 2.06, leading to partial melting within the plume head and in asthenosphere beneath thinned regions of the lithosphere. Once these mantle melts formed, they rose to levels immediately below, or within, the crust where partial fractional crystallisation occurred. The heat released from these magmas resulted in an elevated crustal geotherm to values approaching 40-50°C km^{-1}, and regional metamorphism of the adjacent crust. Crustal anatectic magmas rose to form the felsic volcanic succession of the Rooiberg Group and shallow-level intrusions of the Rashoop Granophyre and Lebowa Granite Suite. At the same time, the partially fractionated and contaminated mafic magmas were remobilised and rose to intrude shallow crustal level beneath a carapace of volcanics and in the extensions in the Molopo Farms Complex.
Clearly, there is scope for a lot more research and discussion on this fascinating LIP.

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