CHAPTER TWO: OVERVIEW OF THE BUSHVELD COMPLEX

2.1 General

The Bushveld Complex is the largest known layered intrusion on Earth, with an outcrop area of over 29450 km^2 and further sub-outcrops of 36550 km^2 (Fig. 1.1, Von Gruenewaldt, 1977). It lies almost entirely within the bounds of the Transvaal sedimentary basin with both the intrusion and the sedimentary rocks possibly forming in response to intracratonic rifting (Eriksson et al. 1991). The Bushveld Complex contains approximately 80% of the world's resource of PGE (Morrissey, 1988), as well as the bulk of resources in Cr and V.

The Complex consists of three suites of plutonic rocks, namely the mafic-ultramafic Rustenburg Layered Suite (South African Committee for Stratigraphy, 1980), the Rashoop Granophyre Suite and the Lebowa Granite Suite (Von Gruenewaldt et al., 1985). The Rustenburg Layered Suite is an approximately 8 km thick succession of layered mafic and ultramafic rocks, exposed in 5 major lobes, i.e., the eastern-, western-, and far-western lobes, the northern or Potgietersrus - Villa Nora lobe, and the Bethal lobe. The latter, hidden below younger sedimentary cover, was identified on the basis of a gravity high and is only known from borehole core. The individual lobes differ in aerial extent, thickness and degree of exposure. There is still controversy as to whether the limbs are joined at depth. Connectivity of the lobes at depth was first proposed by Hall (1932). Later interpretations of the Bouger anomalies suggested that the mafic rocks were not continuous at depth (Cousins, 1959). Drilling

established extensions of the western limb at its northern end beneath the Bushveld granite, and of the eastern limb beneath the Karoo sedimentary cover to the west of the Wonderkop fault (Eales and Cawthorn, 1996). Cawthorn and Webb (2001) proposed that Hall's initial model was correct, in that the eastern and the western limb of the Bushveld Complex are linked and formed within a single lopolithic intrusion. Their proposition is largely based on the lithological and compositional similarity between the different lobes, and by considering the isostatic response of the crust after emplacement of the Bushveld magma.

The Complex intruded the 2550 – 2060 Ma (Nelson et al., 1999) Transvaal Supergroup largely along an unconformity between the Magaliesberg quartzite of the Pretoria Group and the overlying Rooiberg felsites (Cheney and Twist, 1992). The exceptions are (i) the eastern lobe south of the Steelpoort fault, where the Complex transgressed upwards through more than 2 km of sediments and (ii) the Potgietersrus limb where intrusion occurred at the level of the Magaliesberg quartzite in the south, but transgressing downwards towards the north, until the mafic rocks abut Achaean granitic gneiss (Eales and Cawthorn, 1996).

The Transvaal Supergroup is a supracrustal volcanosedimentary sequence which consists, from the base to the top, of (a) the protobasinal Wolkberg and Buffelsfontein Groups, (b) the carbonaceous Chuniespoort Group, and (c) the largely pelitic Pretoria Group. U/Pb dating of titanite yielded an age of 2058 ± 0.8 Ma for the Bushveld Complex (Buick et al., 2001), and this age is virtually identical to the age of the Rooiberg volcanics at the top of the Transvaal Supergroup (Walraven et al., 1990).

The last magmatic event of the Bushveld Complex is dated at 2052 ± 48 m.y (Rb-Sr whole rock age) (Walraven et al., 1990) and is represented by the Lebowa Granite, which was emplaced at, or close to, the contact between the Rustenburg Layered Suite and the acid roof rocks (Von Gruenewaldt et al., 1985).

The mode of emplacement of the layered suite was one of repeated injections of magma. This is suggested by distinct reversals in the initial Sr – isotopic ratio and in differentiation indexes such as Mg number and Cr contents, as well as textural evidence such as resorbed plagioclase inclusions in olivine and pyroxene (Eales et al. 1986).

2.2   The Rustenburg Layered Suite

The Rustenburg Layered Suite is generally sub-divided into five zones (Hall, 1932): at the base is the Marginal Zone which is overlain by the Lower Zone, Critical Zone, Main Zone, and the Upper Zone. A simplified stratigraphic column representative of the Bushveld Complex is shown in Fig. 2.1.

The basal Marginal Zone consists of unlayered fine- to medium grained heterogeneous gabbronoritic rocks and varies in thickness between 0 and 250 m (western Bushveld Complex, Coertze, 1974). The rocks of the Marginal Zone are generally
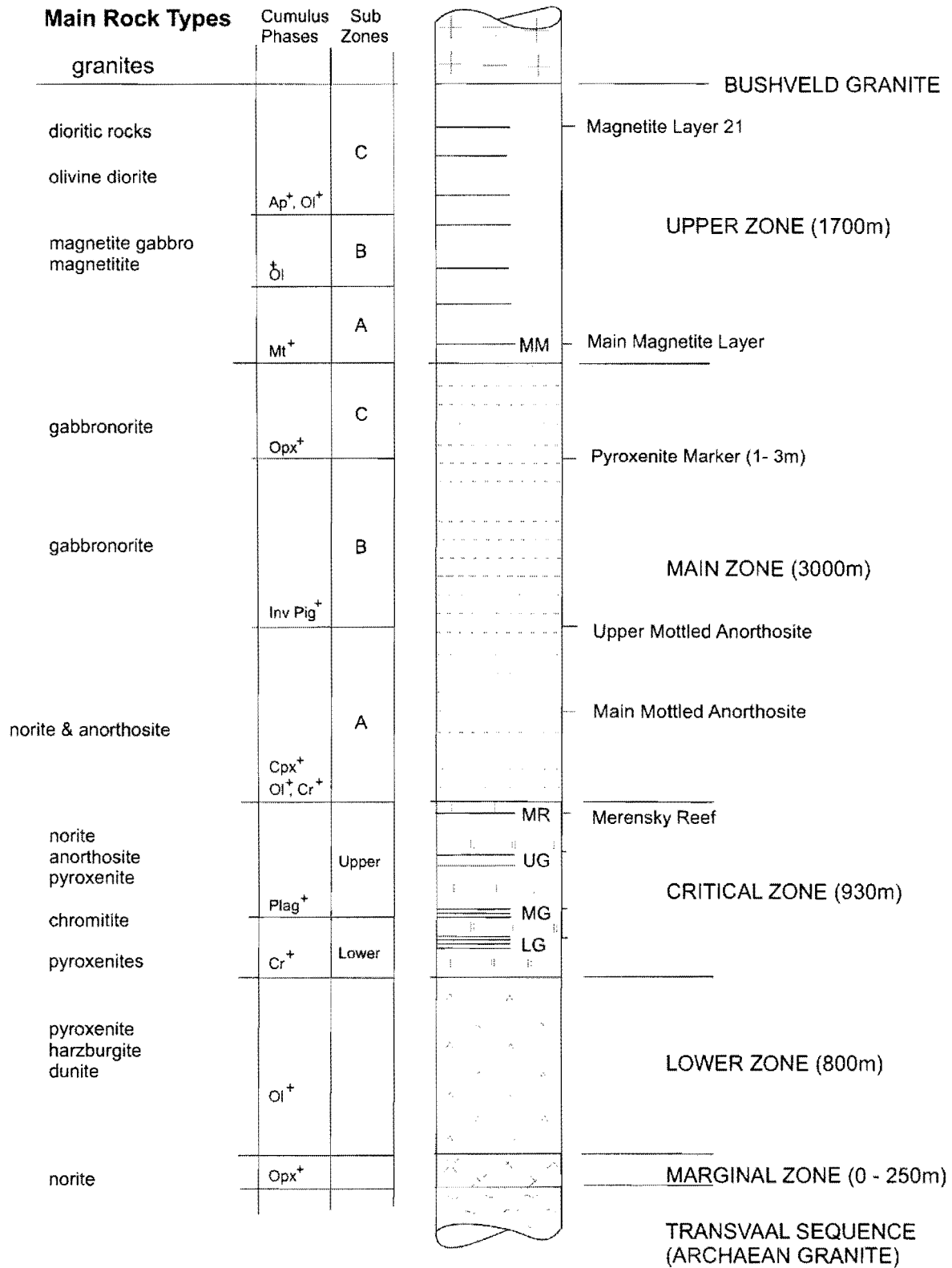
Fig. 2.1: Stratigraphy of the Rustenburg Layered Suite, Bushveld Complex (from Mitchell 1990).

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resources of chromite within the Middle Group and Upper Group chromitite layers (Hatton and Von Gruenewaldt, 1987).

According to the most widely accepted subdivision, the base of the Main Zone may be placed at the top of the Bastard unit, although the exact position is somewhat controversial (Kruger, 1990; Mitchell and Scoon, 1991). The Main Zone is a ca. 3000 m thick sequence consisting mainly of norite in the basal and uppermost portions, but gabbronorite in the intervening central portion (Mitchell, 1990). Anorthosite constitutes some 5 % of the rocks, while pyroxenite is rare. The Main Zone is characterised by the absence of olivine and chromian spinel, coupled with the general lack of the fine scale layering that is typical of the Critical Zone.

Also a subject of debate is the position of the Main Zone – Upper Zone boundary. Based on a reversal in Sr isotopic ratio and in the trend of iron enrichment (Von Gruenewaldt, 1973; Klemm et al., 1985), Kruger (1990) placed the boundary at the level of the Pyroxenite Marker, a prominent pyroxenite layer some 2.5 km above the base of the Main Zone. The more commonly used subdivision is by Wager and Brown (1968) who defined the base of the Upper Zone by the first occurrence of cumulus magnetite, some 660 m above the Pyroxenite Marker.

2.3 PGE Mineralisation

The PGE mineralisation in the Rustenburg Layered Suite occurs as:

a)   stratiform sulphide bearing horizons including the Merensky reef (Lee, 1983; Naldrett et al., 1986; Barnes and Maier, 2002), Platreef (Gain and Mostert,

1982; Van der Merwe, 1976), Bastard Reef (Lee, 1983), Pseudo Reef and Tarentaal layers (Naldrett et al., 1986), and the footwall of the Lower Magnetite Layer 2 (Von Gruenewaldt, 1976).

b)   chromitites (Gain, 1985; Von Gruenewaldt et al., 1986; Hiemstra, 1986; Lee and Parry, 1988; Teigler, 1990 a, b; Scoon and Teigler, 1994) and

c)   discordant PGE-enriched pipes of mafic-ultramafic pegmatite and magnesian dunite in the Critical Zone of the eastern Bushveld Complex, at Mooihoek, Onverwacht and Driekop (Scoon and Mitchell, 1994)

For the sulphide bearing layers, two main genetic models are generally considered. Most workers believe that the PGE were concentrated by a sulphide liquid that segregated from the silicate magma after S supersaturation was achieved. In the past, it was widely believed that sulphide supersaturation can be achieved by mixing of compositionally contrasting magmas, one being primitive, the other being more evolved (Naldrett and von Gruenewaldt, 1989; Irvine, 1986; Li et al., 2001a). It has recently been shown (Li et al., 2001b; Cawthorn, 2002) that this model is probably incorrect and that magma mixing cannot trigger S supersaturation. Therefore, other mechanisms to cause S supersaturation have to be considered, such as contamination of the silicate magma with the country rocks. Mechanical concentration of sulphides after initial precipitation appears to be a further pre-requisite to produce economic sulphide deposits.

An alternative model (Boudreau et al. 1986) proposes that the PGE are concentrated by percolating late-magmatic fluids, but this model has been criticized by several

authors, based, in part, on the occurrence of PGE-rich sulphide near the base of the Complex in the Platreef.

For the S-poor chromitites, some workers have in the past proposed that the PGE were initially concentrated by sulphides, but that much of the S was subsequently lost, with the PGE remaining behind (e.g. Naldrett and Lehmann, 1988; von Gruenewaldt et al., 1989). More recently, it is also considered whether PGM can precipitate directly from the silicate magma, perhaps in response to destabilization of atomic metal clusters (Tredoux et al. 1995).

The PGE mineralised discordant pipes contain a very unusual PGE assemblage, dominated by Pt, suggesting that the mineralisation is not of primary magmatic origin. Stumpfl (1993) has suggested that the PGE are instead the result of hydrothermal remobilization from cumulates. The origin of the pipes and the mineralisation remain enigmatic.