CHAPTER 2

REGIONAL GEOLOGY: THE SIGNIFICANCE OF THE UMKONDO AND ASSOCIATED MAFIC INTRUSIVE ROCKS

2.0 Introduction

The Umkondo group comprises a sequence of metasedimentary strata that are sub-divided into the Zimbabwe facies and the Mozambique facies, which overlie Archean basement granites and gneisses of the Zimbabwe craton. In Zimbabwe, rocks of the Mozambique facies crop out near the Mozambiquen and have been thrust westwards and juxtaposed with rocks of the Zimbabwe facies (Payne, 1984; Munyanyiwa, 1999). The Zimbabwe facies comprises relatively undeformed, low grade metamorphic strata, while rocks of the Mozambique facies are intensely deformed and metamorphosed. Extensive dolerite sills and dykes have intruded the metasedimentary strata and the Archean basement granites. Dykes of possible Mashonaland- and Umkondo age predominantly occur in granite terrain and are scarce in the area covered by metasedimentary strata (Fig. 1.3)

In this chapter, Umkondo dolerites are correlated with other igneous rocks of similar ages in southern Africa. Findings from previous work are reviewed and briefly discussed. Available geochronological data for three mafic dyke swarms of Wilson *et al.* (1987), namely the Mutare, Guruve, and Kamativi dykes are given, and their possible age relationships to Umkondo dolerites are discussed. For convenience and avoiding confusion, the terminologies used for the dyke swarms by Wilson *et al.* (1987) are retained in this thesis.

2.1 Umkondo Group

The Umkondo group comprises metasedimentary rocks that crop out in two areas that are some 200 km apart, i.e. in the Chimanimani-Chipinge districts to the south-, and the Nyanga district to the north of Mutare town (Fig. 1.1). Deposition of the rocks in both areas possibly occurred in a single, north-south trending basin that had an average width of at least 100 km in the south, and extends into Mozambique to the east (Payne, 1984). In the Nyanga area, Stocklmayer (1978) sub-divided rocks of the Umkondo group into the relatively undeformed and weakly metamorphosed Inyanga facies that occurs to the west; and the intensely deformed and highly metamorphosed Gairezi facies (Tyndale-Biscoe, 1957) to the east. In the Chimanimani-Chipinge area, Watson (1969) subdivided the rocks of the Umkondo group into the low metamorphic grade, less deformed Calcareous-, Quartzite- and Argillaceous Series to the west; and the strongly deformed rocks of the Frontier Series that crops out near the Zimbabwean boarder to the east and adjacent parts of Mozambique. The classification by Tyndale-Biscoe (1957) and Watson (1969) was based on the variation in degrees of deformation and metamorphism.

Stocklmayer (1978) correlated rocks of the Umkondo group in the Chimanimani-Chipinge and Nyanga areas. The rocks were subsequently sub-divided into a western succession of weakly deformed, low metamorphic grade rocks, namely the Zimbabwe facies, and an eastern succession of intensely deformed and high metamorphic grade rocks, the Mozambique facies (Stocklmayer, 1978). The classification by Stocklmayer (1978) was based on broad lithological, metamorphic and structural similarities of rocks in Nyanga and Chimanimani.

The Zimbabwe facies, which makes up most of the exposures in Zimbabwe, consists of pebbly basal arkoses that grade upwards into limestones, cherts, shales and sandstones and capped by basalts (Watson, 1969; Stocklmayer, 1978; Munyanyiwa, 1999). These rocks overlie- and are flanked by the Archean basement granites to the west (Fig. 1.1). Detrital grains of the basal arkose include very angular pink feldspar, grey quartz and a little mica in places, which are the constituents of the underlying granite (Swift, 1962).

The arkose was probably derived from weathering of the granite by sub-aerial erosion (Swift, 1962; Watson, 1969). The Zimbabwe facies was possibly deposited nonconformably on a peneplained Archean basement and on gneisses of the North Marginal Zone and Central Zone of the Limpopo belt to the west (Tyndale-Biscoe, 1957; Watson, 1969; Stocklmayer, 1978; Hanson *et al.*, 1998b). The Mozambique facies consists of highly deformed and strongly metamorphosed, greenschist- to amphibolite metamorphic grade psammites, phyllites, metapelites, calc-silicate hornfelses, limestones and schists (Tyndale-Biscoe, 1957; Watson, 1969; Stocklmayer, 1978; Payne, 1984; Dirks, personal communication, 1999). There is a sharp change in grade of metamorphism and deformation between the Mozambique facies and the Zimbabwe facies (Dirks, personal communication, 1999).

The Mozambique facies was inferred to be older than the Zimbabwe facies because of the higher degrees of deformation and metamorphism and marked lithological differences along a sharp, tectonic contact (Bond, 1951; Tyndale-Biscoe, 1957). However, subsequent workers in the area (e.g. Slater, 1965, Stocklmayer, 1978) asserted that the Mozambique facies was deposited contemporaneously with the Zimbabwe facies and the difference in appearance was interpreted to reflect facies variations and degrees of deformation and metamorphism.

2.2 Regional Structure

The Mozambique facies overlies the Mozambique quartzofelspathic gneisses and gneisses of the Limpopo belt (Watson, 1969; Dirks, 1999). The gneisses grade upwards into an approximately 3m-wide cataclastic zone of lineated, sheared quartz-mica schists at the contact with the Mozambique facies (Tyndale-Biscoe, 1957; Watson, 1969; Payne, 1984) Evidence for thrusting includes the occurrence of north–south trending major faults with east-west plunging sub-horizontal lineations that cut the Mozambique facies at shallow angles as well as asymmetric shear sense indicators. These have been documented in the Nyanga and Chimanimani areas (Watson, 1969; Payne, 1984; Tennant, 1995). The occurrence of tight and recumbent folds with north-south trending

fold axes and sheath folds (Watson, 1969; Payne, 1984) suggests high strain that is compatible with thrusting. Thus the Mozambique facies is clearly allochtonous and was thrust westwards as shown by kinematic indicators (Tyndale Biscoe, 1957; Watson, 1969; Payne, 1984, Tennant, 1995).

The Zimbabwe facies has been considered by most workers to be an autochthonous sequence that rests non-conformably on the Zimbabwe craton and the Limpopo belt to the west (Tyndale Biscoe, 1957; Swift; 1962; Watson, 1969; Stocklmayer, 1978). However, Watson (1969) has documented evidence of local brecciation and mylonitisation at the contact zone of the Zimbabwe craton and the Zimbabwe facies. Swift (1962) reported a silicified fault that separates the Archean basement granites and the Zimbabwe facies some 3 km south of Mapari borehole (Grid Ref.: 418395; 7777225). Rocks of the Zimbabwe facies were, in places, folded in to low amplitude, gentle folds with fold axes that plunge shallowly to the south (Tyrwhitt, 1966; Watson, 1969; Stocklmayer, 1978). These observations raise the possibility that the contact between the Archean basement granites and the Zimbabwe facies could be tectonic.

2.3 Umkondo Dolerites and Basalts

Dolerites intrude the Archean basement granites and gneisses and the overlying Umkondo strata as both dykes and sills, although the sills are far more widespread (Munyanyiwa, 1999). Some sills were intruded along bedding planes, which are transgressed in places, and along the Archean basement granite/Zimbabwe facies contact (Watson, 1969; Mukwakwami *et al*, 2001). The dolerites are referred to as Umkondo dolerites and are common in the Zimbabwe facies where they are relatively undeformed and are believed to have intruded wet sediments (Munyanyiwa, 1997; Hanson *et al.*, 1998b). Rare dolerite sills occur in the Mozambique facies where they have been strongly deformed and are structurally conformable with rocks of the Mozambique facies (Watson, 1969; Tennant, 1995; Munyanyiwa, 1999). Basalts cap the Zimbabwe facies metasedimentary strata in the area around Hot Springs (Fig 1.5).

To the west of the Umkondo group are dolerite sills/dykes dated at 1.8 Ga termed the Mashonaland sills (Compston and McElhinny, 1975; Wilson et al, 1987). Umkondo dolerites and Mashonaland dolerites cannot be distinguished in the field and some of the dykes mapped as Mashonaland dykes could in fact be of Umkondo age and vice-versa (Hanson *et al.*, 1998b). Most dykes and sills intruding the Archean basement are assumed to belong to the Mashonaland suite. However, Wilson *et al.* (1987) and Allsopp *et al.* (1989) obtained a palaeomagnetic pole and Rb-Sr isotopic data that are consistent with an Umkondo age for the Deweras dyke in northwestern Zimbabwe. Unpublished Rb-Sr work on the Deweras dyke constrains the age of the swarm to between 500 and 1300 Ma Wilson *et al.* (1987). The limited palaeomagnetic and geochronological data suggest an age of ~1000 Ma for the Deweras dykes, but do not rule out a correlation with the Umkondo dolerites (Wilson *et al.* 1987). If so then the Umkondo igneous event may have extended across most of the exposed Zimbabwe craton.

2.4 Previous Geochronology on Umkondo Dolerites

The age of Umkondo dolerites has been the subject of many publications (Table 2.1). Cheney *et al.* (1990) asserted that regression of 5 lava samples collected by Allsopp *et al.* (1989) yielded an age of ~1900 Ma. Cheney *et al.* (1990) subsequently correlated the Umkondo group with the Paleoproterozoic sedimentary basins of southern Africa. The maximum Mesoproterozoic age for the Umkondo Group, and hence the Umkondo dolerites, has been based on the assertion that the Zimbabwe facies of the Umkondo group unconformably overlies the North Marginal Zone and the Central Zone of the Limpopo belt (Swift, 1962; Watson, 1969). These parts of the Limpopo belt have suffered the 2.0 Ga high-grade granulite facies metamorphism, which did not affect the Umkondo group metasediments (Van Breeman, 1968; Kamber *et al.*, 1995). The Mashonaland ages are interpreted to reflect inheritance through contamination by older crustal rocks, while the younger ages (Vail and Dodson, 1969 and Table 2.1) reflect Ar-loss during subsequent tectono-thermal events and the significance of these younger ages is further discussed in chapter 6.

Available precise U-Pb zircon dating gave an emplacement age of 1105 ± 2 Ma for an Umkondo dolerite sill that intrudes the Umkondo group to the east of Chimanimani town (Hanson *et al.,* 1998b and Fig. 2.1). Wingate (2001) subsequently published a SHRIMP U-Pb zircon-baddeleyite age of 1099 ± 9 Ma for another Umkondo sill north of Nyanga town (Fig. 2.1). However, Wingate (2001) asserted that the best estimate for the Umkondo magmatic event is 1105 ± 5 Ma, which is a mean age obtained by combining the age and uncertainty of 1107 ± 7 Ma derived from the Hanson *et al.* (1998b) data and the U-Pb zircon-baddeleyite age of 1099 ± 9 Ma. Findings from this project show that Umkondo sills in eastern Zimbabwe were emplaced between 1105 and 1109 Ma (Chapter 6). The mean age of Wingate (2001) from the two sills that are at least 100 km apart is probably imprecise and may have no geological meaning since the two sills may have slightly different ages.

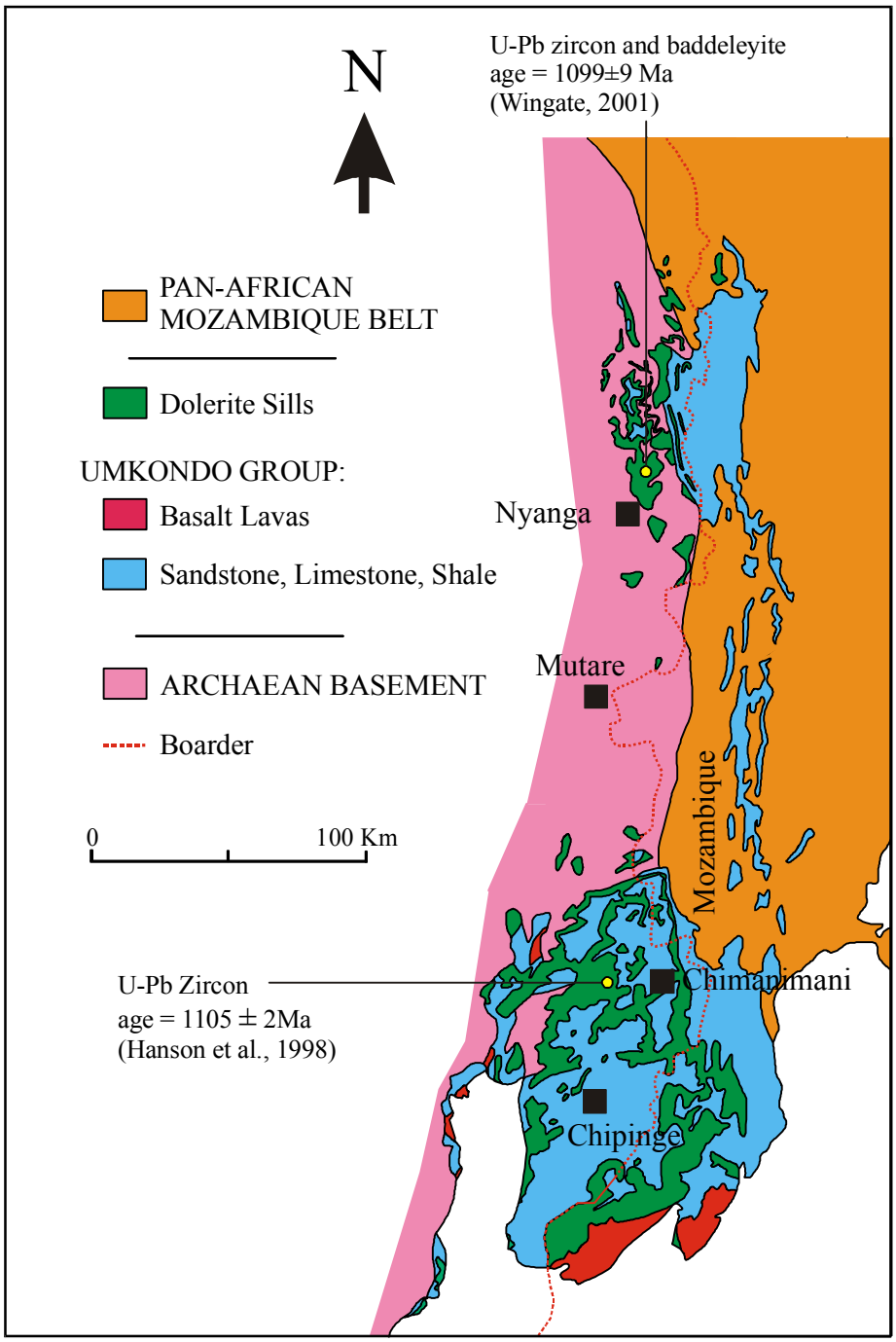

Fig. 2.1 Spatial distribution of the Umkondo group and associated dolerite sills that intruded the strata at about c. 1100 Ma in eastern Zimbabwe and adjacent parts of Mozambique. The sampling sites of U-Pb geochronological samples reported in Hanson *et al.* (1998b) and Wingate (2001) are shown. Modified after R. Hanson (unpublished manuscript).

2.5 Geochemistry of the Umkondo Dolerites

Munyanyiwa (1999) carried out geochemical studies on Umkondo dolerites and basalts. Although Umkondo dolerites and basalts were generally metamorphosed at low greenschist facies (chapter 4), Munyanyiwa (1999) concluded that these processes did not cause significant mobility on most major and trace elements.

Umkondo dolerites and basalts are low P-Ti (LPT) and have low *mg* numbers (*mg* = 100 x Mg/[Mg + Fe]). The LPT dolerites and basalts are inferred to have formed by crustal contamination of the mafic magma *en route* to the surface (Peate and Hawkesworth, 1996). Low *mg* numbers are typical of continental flood basalts (Wilson, 1989). The rocks are quartz normative and sub-alkaline and fall within the tholeiitic field of Irvine and Barager (1971). Major element data plotted against MgO show a general decrease in K_2O, Na_2O, SiO_2 and TiO_2 and an increase in CaO with increasing MgO, which suggest that the Umkondo magma underwent differentiation processes, possibly crystal fractionation and crustal contamination (Wilson, 1989).

Umkondo dolerites and basalts have low contents of Ni, Cr, Cu and Co although the dolerites are slightly more enriched in these elements than the basalts (Munyanyiwa, 1999). These low values suggest that the parental magmas were not directly derived from a peridotitic mantle source and have undergone differentiation through crystal fractionation and crustal contamination at depth (Wilson, 1989). Munyanyiwa (1999) noted an increase in Ni and Cr with increasing MgO, while incompatible elements such as Zr, Y, Th and Nb decrease with increasing MgO, which he interpreted to imply differentiation of the magma on ascent to the surface. High Zr, Y and Nb content in one dolerite sample suggest crustal contamination (Munyanyiwa, 1999), which is also evidenced by a high $^{87}Sr/^{86}Sr$ initial ratio of 0.7110 ± 0.0025 for the Umkondo dolerites (Allsopp *et al.*, 1989). Most Umkondo mafic rocks plot in the intra-plate basalt field of Pearce and Norry (1979) on a Zr/Y versus Y diagram. Munyanyiwa (1999) showed that most data plot in the intra-plate field overlapping with the volcanic-arc basalt on the Zr-

Nb-Y diagram (Meschede, 1986) and within the subduction arc and MORB fields on discrimination diagrams of Pearce and Cann (1973) and Wood *et al.* (1979).

Although most trace element data plot within subduction-arc and MORB fields on discrimination diagrams of Pearce and Cann (1973) and Wood *et al.* (1979), there is no geological evidence for these tectonic settings (Munyanyiwa, 1999). Wang and Glover (1992) have shown that continental basaltic rocks may plot in fields for MORB and volcanic-arcs in some circumstances when Ti and Nb values are low. Umkondo dolerites and basalts plotted in the MORB fields possibly owing to the low Ti and Nb values (Munyanyiwa, 1999). The dolerites intruded the Umkondo strata, which have been interpreted as typical shallow marine platform-type sediments (Button, 1977; Stocklmayer, 1978). Munyanyiwa (1999) concluded that geological and geochemical evidence is compatible with a continental setting for the emplacement of Umkondo mafic rocks.

The chondrite-normalised REE patterns for Umkondo dolerites and basalts are similar (Munyanyiwa, 1999), which indicate that the dolerites may be feeders to the basalts as previously established from palaeomagnetic studies (McElhinny, 1966; Wilson *et al.*, 1987). The dolerites show high Ce/Yb values of 13-17 (Munyanyiwa, 1999), which are typical of continental tholeiites (Green, 1992). Munyanyiwa (1999) also showed that the dolerites and basalts are enriched in the incompatible LILEs, Ba, Th, Rb and K. He interpreted the LILE enrichment to possibly reflect crustal contamination that is farther supported by negative Nb anomalies (Munyanyiwa, 1999), which are typical of continental crust (Rollinson, 1993).

2.6 Mesoproterozoic Mafic Dykes in Zimbabwe – An Extension of Umkondo Magmatism Across Zimbabwe?

Wilson *et al.* (1987) classified mafic dykes in Zimbabwe into Archean-, Proterozoic- and Phanerozoic dyke swarms on the basis of scant Rb-Sr geochronology, palaeomagnetic data, trends and field relationships (Fig. 2.2). Archean dyke swarms, which occur within the south-central part of the country, are interpreted as possible feeders to the ~2700 Ma

Upper Bulawayan greenstone belts (Wilson *et al.*, 1987). Major Proterozoic mafic dyke intrusions include the ~ 1800 Ma Mashonaland dykes and the ~ 1100 Ma Umkondo dykes and sills. The Guruve, Kamativi and Mutare dykes have been considered as possible correlatives of the Umkondo dolerites in eastern Zimbabwe. Major Phanerozoic dyke swarms include the Botswana, Limpopo and Rushinga (also known as Rukore Suite) dyke swarms that oocur to the south, southeast and northeast of Zimbabwe respectively. Emplacement of these Phanerozoic dykes was associated with events that led to the break-up of Gondwanaland.

In the next sections three presumed Proterozoic dyke swarms, namely the Kamativi, Guruve and Mutare dykes are explored and the basis for their possible age relationships with the Umkondo dolerites is shown.

2.6.1 Kamativi Dykes

The Kamativi dykes intruded the Dett Inlier of folded Paleoproterozoic metasedimentary rocks in the Dett area, western Zimbabwe (Fig. 2.2). The dykes were emplaced along east northeast-trending fractures (Figs. 1.9 and 2.2). The precise ages of these dykes are not known. Some of the dykes are cut and displaced by lithium-cassiterite-bearing pegmatites (Lockett, 1979; Wilson *et al.*, 1987) that have a Rb-Sr muscovite-plagioclase isochron age of 1025 ± 15 Ma (Priem *et al.,* 1971), which suggests a minimum age of ~ 1000 Ma for the Kamativi dykes. Wilson *et al.* (1987) correlated the Kamativi dykes with the Guruve dykes on the basis of similarity in trends (Fig. 2.2).

The Kamativi dykes show heterogeneity in petrography and four varieties were distinguished in the area by Lockett (1979). These include “fine grained” dolerites consisting of clinopyroxene, plagioclase and ilmenite and “medium grained”, sub-ophitic rocks composed of labradorite, twinned pigeonite and varying amounts of hornblende, biotite, chlorite and titanite. Some altered dolerites containing pseudomorphs of fibrous amphibole after clinopyroxene, clouded plagioclase and minor biotite and chlorite, and

nepheline-bearing dykes with nepheline crystals in a matrix of hornblende, epidote and sodic plagioclase occur in places (Lockett, 1979). Ward and Hall (2001) distinguished five subgroups from the Kamativi dykes on the basis of mineralogical and textural grounds. The five subgroups show distinct geochemical signatures, which suggest different magma sources for the subgroups (section 2.6.5).

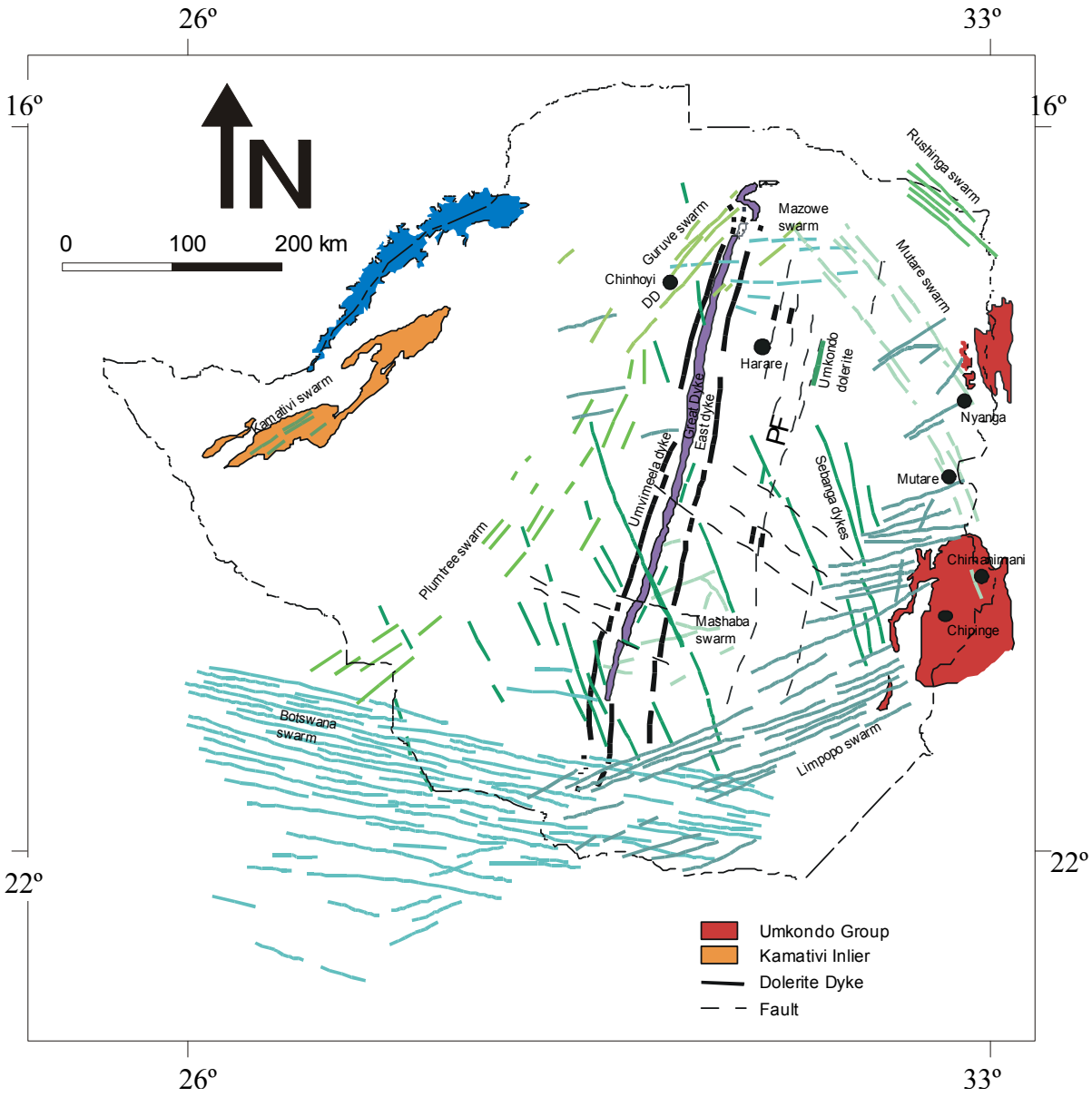

Fig. 2.2. Simplified map showing the distribution of mafic dyke swarms in Zimbabwe, modified after Wilson *et al.* (1987). DD = Deweras dyke, PF = Popoteke fault.

2.6.2 Mutare Dykes

The Mutare dykes form an arcuate group of dykes that extend for a distance of over 400 km from near the north end of the Great Dyke to south of Mutare and Mozambique (Figs. 1.3, 1.4, 1.5, 2.4). These dykes intrude the Archean granites and Umkondo metasedimentary rocks at the southern end, and strike north-northwest parallel to the prominent Mashonaland-Sebanga dykes (Fig. 2.2). In the north, the Mutare dykes swing to strike northwest and cut some Sebanga dykes and Mashonaland sills.

The age for the Mutare dykes has been poorly constrained and very little geochronological and palaeomagnetic work has been done. Palaeomagnetic work by McElhinny *et al.* (1968) gave directions that can be associated with either the Mashonaland dolerites or rocks of Pan African age emplaced at c. 500-600 Ma. The dykes predate metamorphism and tectonism of the Zambezi belt (Leitner and Phaup, 1974) because they were affected by metamorphism that also affected the Zambezi belt. Wilson *et al.* (1985) initially speculated that the Mutare dykes were possibly emplaced contemporaneously with Umkondo dolerites. The Rb-Sr isotopic data obtained from two members of the Mutare dykes are scattered on an isochron diagram and have been interpreted to preclude an age greater than approximately 700 Ma (Wilson *et al.*, 1987). Data from the two techniques when considered together tentatively suggest a Pan African age of 500-600 Ma for the two dykes, and hence the age of the Mutare dykes. However, it is plausible to speculate that some of the Mutare dykes may be of Umkondo age since some of the dykes intrude the Umkondo group, occur in the same geographical area as the Umkondo sills (Fig. 1.3) and share petrographical and textural similarities with the Umkondo sills (Chapter 3). The Mutare dykes were preliminarily subdivided into three subgroups on the basis of geochemical data (Ward *et al.*, 2000). Ward (2002) subsequently reclassified the dykes into six subgroups after acquiring additional geochemical data.

2.6.3 Guruve Dykes

The northeasterly trending Guruve dykes occur in the northern part of the Zimbabwe craton (Figs. 1.8, 2.4). The Guruve dykes predate the Pan African deformation of the Zambezi belt since some of the dykes are folded in the area of the Zambezi belt (Wilson *et al.*, 1987). The Guruve dykes include the Deweras dyke (Figs. 1.8; 2.4) that is up to 100 m wide and has a near-continuous strike length of at least 120 km with some overlapping sections and local bifurcations (Stagman, 1961; Tennick, 1976). Relatively smaller, parallel dykes occur to the north of the Deweras dyke (Wiles, 1972) and also to the east, beyond the Great Dyke (Wilson *et al.*, 1987; Fig. 2.2). The Deweras dyke intruded metasedimentary rocks of the Deweras Group (Tennick, 1976) near its southern end. Wilson *et al.* (1987) inferred that the Deweras dyke postdates overthrusting of the Lomagundi metasediments onto the Zimbabwe craton, suggesting that the emplacement of the Guruve dykes postdate the Magondi orogeny (Wilson *et al.*, 1987).

Dykes with a typical Guruve trend follow regional fold axes and foliation trends of the Shamvaian metasediments (Hahn and Steiner, 2001). The dykes vary petrographically from metamorphosed metadolerites to partly altered and unaltered dolerites (Wiles, 1972). These include north northeast-trending gabbroic dykes, amphibolite dykes and doleritic dykes. Northeast trending, foliated amphibolite dykes that were deformed with the Shamvaian metasediments are possibly the oldest set of dykes that pre-date the emplacement of the Guruve dykes. It is possible that some of these dykes were feeders to the Shamvaian and Magondi lava flows. Hahn and Steiner (2001) presumed that these dykes predate the Magondi orogeny dated at 1997.5 ± 2.6 Ma (McCourt *et al.*, 2001) and possibly represent the oldest generation of dolerite dykes in the area. Metamorphism of these dykes was possibly related to the Magondi orogeny. In the area around Mhangura mine, Master (1991) noted some pyrite-bearing dykes that post-date M_1 and M_2 metamorphic events as well as D_1 and D_2 deformation events of the Magondi orogeny. Localised sigmoidal cleavages at Mhangura mine produced by shearing during D_3 reactivation of the thrust constrains the age of these dykes to between 1800 Ma and 1200 Ma (Master, 1991). The Deweras dyke and associated dykes are considered by Hahn and

Steiner (2001) to be the youngest set of dykes and belong to a magmatic event that is different from the Guruve dykes although they both have the same trend. Ward *et al.* (2000) initially noted three subgroups on the basis of geochemical data, but her further work has shown that the dykes can be classified into six subgroups (section 2.6.5).

Rb-Sr work on the Deweras dyke constrains the age of the Guruve dykes to between 500 and 1300 Ma (Wilson *et al.*, 1987). Wilson *et al.* (1987) obtained a pole direction opposite to that observed in the Umkondo dolerites and concluded that available palaeomagnetic and geochronological data suggest an age of approximately 1000 Ma for the Deweras dyke, and hence the Guruve dykes. However, Wilson *et al.* (1987) asserted that these data do not rule out a possible correlation with the Umkondo dolerites. Höehndorf *et al.* (1990) obtained a Sm-Nd age of 1100 ± 280 Ma for the Deweras dyke. Hahn *et al.* (1991) subsequently published a Rb-Sr errochron date of 1100 ± 270 Ma for the Deweras dyke and a magnetization direction similar to the pole for the Umkondo dolerites, but with reversed polarity.

2.6.4 Chimbadzi Mafic/Ultramafic Dyke

The Chimbadzi mafic/ultramafic dyke occurs some 25 km north northeast of Chinhoyi (Fig. 1.8). The dyke is sub-vertical and has a thickness of up to 800 m and a minimum length of 11.5 km and trends parallel to primary bedding and foliation of the Shamvaian metasediments (Manyeruke, 2001). Stagman (1959; 1961) asserted that the Chimbadzi mafic/ultramafic dyke, which comprises troctolite and magnetite-ilmenite-olivine rock units (Manyeruke, 2001), is genetically similar to the Great Dyke and the Bushveld Igneous Complex of South Africa. Recent isotopic U-Pb badelleyite dating by shows that the emplacement age of the Chimbadzi dyke is 2262.6 ± 1.5 Ma (Manyeruke, 2001; Manyeruke *et al.,* 2001) and is younger than the Great Dyke by ~ 300 Ma. The age of the Chimbadzi dyke does not correspond to any major documented tectonothermal event in the Zimbabwe craton or adjacent orogenic belts. Kamber *et al.* (1995a) have documented an apparent bulk sphene/plagioclase Pb-Pb age of 2209 ± 58 Ma for granulites in the

Northern Marginal Zone of the Limpopo belt, which is within the error of margins of the age of the Chimbadzi dyke. A conventional U-Pb zircon age of 2201± 36 Ma (Holzer *et al.*, 1995) from strongly deformed granites and a feldspar-whole rock-garnet Pb-Pb isochron of 2216 ± 10 Ma (Kamber *et al.*, 1995b) have also been reported in the Northern Marginal Zone. However, the 2.2 – 2.4 Ga ages in the Northern Marginal Zone have been interpreted as mixing ages that reflect partial resetting of mineral chronometers during the low-grade retrogression of late Archean granulites at c. 1.97 Ga (Kamber *et al.*, 1995a). There is need to revisit these 2.2-2.4 Ga age data in the light of new age data from the Chimbadzi dyke.

Manyeruke (2001) also documented the occurrence of amphibolite schist dykes, gabbroic dykes and pyrite-bearing dykes that trend parallel to the Guruve dykes in the vicinity of the Chimbadzi mafic/ultramafic dyke. North and east-northeast-trending quartz-bearing metadolerite dykes that are garnetiferous in places have been reported in the area around Guruve (Wiles, 1972). Quartz occurs in quartz-bearing dykes as interstitial mineral and as granophyric intergrowth with alkali-feldspar (Manyeruke, 2001).

2.6.5 Geochemistry of the Mutare-, Guruve- and Kamativi Dykes: A Review

Geochemical work based on major elements-, incompatible trace elements- and REE data by Ward (2002) have established six subgroups within both the Mutare and Guruve dyke suites. The geochemical data from each of the suites are complex and cannot be explained by simple fractionation, crystal accumulation, partial melting or crustal contamination. Available data permit 3 magma sources for the 6 geochemical subgroups within the Mutare and Guruve dykes.

Crustal contamination was insignificant in the formation of type I dykes, which were derived from low-degree partial melting of a relatively deep, garnet-bearing source. Some type II dykes show geochemical signatures that are indicative of a depleted source, while others point to an undepleted spinel Iherzolite source. Type III dykes were also derived

from low-degree partial melting of a depleted source that was significantly influenced by subduction zone metasomatism of mantle reservoirs. Type III dykes have geochemically signatures that are similar to the Umkondo dolerites in eastern Zimbabwe.

The Kamativi dykes have been classified into 4 geochemical subgroups (K_1-K_4) by Ward and Hall (2001). Their work show that K_2 and K_3 partly overlap geochemically, but the K_2 dykes have higher *Mg* content than the K_3 dykes. K_1 dykes have low silica content, but high Ti, Fe, P, Nb, Sr and Zr contents. K_4 dykes have high *Mg*, Ni, Cr and the ratios of their incompatible trace elements are relatively high and do not fall on a simple fractionation/accumulation trend of the other dykes. Ward and Hall (2001) concluded that the Sr-Nd isotope data suggest at least 2 different magma sources for the Kamativi dykes.

2.7 Umkondo Igneous Province

Jones and McElhinny (1966) have shown that some dolerites in Botswana and South Africa have a magnetic pole that corresponds to the Umkondo pole. On the basis of geochronological and palaeomagnetic evidence, Hanson *et al.* (1998b) subsequently proposed that the Umkondo dolerites in eastern Zimbabwe are part of a much bigger igneous province, i.e. the Umkondo Igneous Province (Fig. 2.3). The province encompasses much of the Kalahari craton, including the original continuation of that craton into East Antarctica (Fig. 2.3). The province comprises huge volumes of igneous rocks emplaced throughout this broad area at about 1100 Ma (Table 2.1). Rb-Sr and K-Ar isotopic ages that are older than the Umkondo age of 1100 Ma (Table 2.1) have been interpreted to have no geological meaning and reflect crustal contamination (Walraven, 1986; Moyes *et al.*, 1995), while the younger ages (Table 2.1) record argon loss during subsequent tectonothermal events (Carney *et al.*, 1994; Hanson et al, 2004).

Umkondo dolerites can be correlated with other igneous bodies in southern Africa that include Mesoproterozoic bimodal rhyolites of the Kgwebe-Ghanzi rift in western Botswana (Fig. 2.3 and Table 2.1), which have U-Pb ages of 1106 ± 2 Ma (Schwartz *et al.*, 1996) and 1106 ± 4 Ma (Singletary, 1999). Correlatives of the Kgwebe bimodal

rhyolites that occur along strike in Namibia (Fig. 2.3) have U-Pb zircon ages of 1094 ± 20 Ma (Hegenberger and Burger, 1985) and 1102 ± 7 Ma (Pfurr *et al.*, 1991). Similar volcanic rocks occur in the Koras Group of the Namaqua belt in South Africa (Grobler *et al.*, 1977; Moen, 1987; SACS, 1980 and Fig. 2.3), where a convectional U-Pb zircon age of 1180 ± 74 Ma (Botha *et al.*, 1979) was obtained. Singletary (2003) obtained U-Pb zircon ages of 1108 ± 1 and 1107 ± 2 Ma for granites that are inferred to be the plutonic equivalent of the Kgwebe rhyolites in northern Botswana (Hanson *et al.*, in press). Post-Waterberg dolerites, i.e. the Dibete-Shoshong and Kanye-Mochudi dolerite suites that intrude the Palapye group and the Waterberg group in Botswana and adjacent parts of South Africa (Fig. 2.3), have recently yielded U-Pb baddeleyite ages that range from 1108 ± 1 to 1110 ± 1 and palaeomagnetic data that are similar to that of the Umkondo dolerites (Jones and McElhinny, 1966; Pancake, 2001 and Table 2.1). The Tshane and the Xade complexes in Botswana (Fig. 2.3) were emplaced during the Umkondo igneous event (Table 2.1). The Timbavati gabbro, a dolerite sill that intrudes the Vredefort Dome impact structure and a dolerite sill that intrudes the Archean basement to the east of the Vredefort Dome in South Africa (Fig. 2.3) have yielded Mesoproterozoic crystallisation ages (Table 2.1) and an Umkondo magnetization direction (Hargraves *et al.*, 1994; Allsopp *et al.*, 1989; Jones and McElhinny, 1966).

The Grunehogna province in Western Dronning Maud Land (Fig. 2.3) represents a detached segment of the Kalahari craton (Martin and Hartnady, 1986; Groenewald *et al.*, 1991, 1995). Structural, palaeomagnetic and geochronology data permit correlation of the Grunehogna province with the Umkondo Igneous Province in southern Africa (Fig. 2.3; table 2.1). The Namaqua-Natal belt that wraps the southern margin of the Kaapvaal craton is interpreted to have been originally continuous with the Dronning-Maud Belt in East Antarctica (Fig. 2.3; Jacobs *et al.*, 1993; Groenewald *et al.*, 1995). The Umkondo group exhibits some stratigraphic and lithological similarities with the Proterozoic strata of the Ritscherflya Supergroup that occurs in the Grunehogna Province (Groenewald *et al.*, 1991). The Ritscherflya Supergroup overlies the Archean basement and is intruded by the ~ 1100 Ma the mafic/ultramafic Borgmassivet tholeiitic intrusions and capped by tholeiitic basalts (Tingey, 1991; Groenewald *et al.*, 1995). The overlying basalts are

geochemically and palaeomagnetically similar to the Borgmassivet intrusions suggesting that the intrusions are the intrusive equivalents of the Ritscherflya basalts (Krynauw *et al.*, 1991; Peters *et al.*, 1991). Some of the Borgmassivet intrusions were emplaced while the host Ritscherflya sediments were still poorly lithified (Krynauw *et al.*, 1988; 1994) as with the case with the Umkondo dolerites (section 3.8).

Table 2.1. A summary of geochronological data for Umkondo dolerites in Eastern Zimbabwe and associated Mesoproterozoic igneous rocks in southern Africa and the Grunehogna Province. The Mashonaland ages for some Umkondo sills reflect inheritance, and the post Umkondo ages reflect Ar-loss during subsequent tectonothermal events. From Hanson *et al.* (In Press), with minor modifications.

Rock Unit	**Symbol (Fig. 2.3)**	**Geochronological data (age in Ma)**	**Reference**
Zimbabwe	UG		
Umkondo dolerite sills		1105 ± 2 (U-Pb zircon)	Hanson *et al.* (1998b)
		1099 ± 9 (SHRIMP U-Pb zircon + baddeleyite)	Wingate (2001)
		1146 ± 9; 767 ± 20; 650 ± 20 (K-Ar whole rock)	Vail and Dodson (1969)
		725 ± 70 (K-Ar pyroxene)	Vail and Dodson (1969)
		1120 ± 50, 1785 ±80 (K-Ar whole rock [hornfels])	Snelling (1966)
		1780 ± 80 (K-Ar biotite hornfels)	Van Breeman *et al.* (1966)
		~ 1136 (K-Ar whole rock)	McElhinny and Opdyke (1964)
		715 ± 70 (K-Ar pyroxene)	Snelling *et al.* (1964)
		1136 ± 9, 757 ± 20, 640 ± 20 (K-Ar whole rock)	(Vail and Dodson, 1969)
		1376 ± 92 (Rb-Sr whole rock)	Allsopp *et al.* (1973)
		1787 ± 50 (K-Ar whole rock [hornfels])	Cahen *et al.* (1984)
		1080 + 140/-25 (Rb-Sr whole rock + biotite isochron)	Allsopp *et al.* (1989)
		1075 ± 25 (Rb-Sr biotite)	Allsopp *et al.* (1989)
South Africa			
Timbavati Gabbro	TG	1123 ± 5; 1072 ± 6, 1072 ± 5, 696 ± 4 ($^{40}Ar/^{39}Ar$ whole rock)	Burger and Walraven (1979; 1980)
		1454 ± 59 (Rb-Sr whole rock isochron)	Bristow *et al.*, (1982)
Post-Waterberg dolerites	PW	1111 ± to 1109 ± 1 (U-Pb baddeleyite, 4 different sills)	J. Ramezani, J. Crowley, S. Bowring, and J. Mukwakwami (*Unpublished data)*
		881 ± 35, 876 ± 35 (K-Ar plagioclase-pyroxene)	Jones and McElhinny (1966)
Dolerite sill intruding Premier Kimberlite	PK	1092 ± 15 (Rb-Sr biotite)	Allsopp *et al.* (1967)
Dolerite sill intruding		1060 ± 42 (K-Ar whole rock)	McDougall (1963)

Archean Basement	❋ AB		
Dolerite sill intruding Vredefort Dome	❋ VD	1052 ± 11 (Rb-Sr pyroxene-plagioclase-biotite-isochron) 1034 ± 8, 948 ± 5 ($^{40}Ar/^{39}Ar$ plagioclase)	Reimold *et al.* (2000) Reimold *et al.* (2000)
Botswana			
Mogkware Sill	PG	1110 ± 1 (U-Pb baddeleyite)	Pancake (2001)
Shoshong and related sills	PG	1110 ± 1, 1109 ± 1 (U-Pb baddeleyite)	Pancake (2001) + unpublished data
Kanye-Mochudi suite	KM	1108 ± 1 (U-Pb baddeleyite) 1264 ± 50, 1110 ± 44, 951 ± 38 (K-Ar plagioclase-pyroxene) 1110 ± 44 (K-Ar pagioclase-pyroxene) 911 ± 36 (K-Ar plagioclase)	Pancake (2001) + unpublished data Jones and McElhinny (1966) Jones and McElhinny (1966) McDougall (1963)
Tshane gabbroic complex	TGC	1105 ± 11 (Rb-Sr plagioclase –biotite isochron) 1071 ± 11 ($^{40}Ar/^{39}Ar$ whole rock) 1021 ± 86 (Rb-Sr whole rock isochron)	Key and Mapeo (1999) Key and Mapeo (1999) J. Barton, in Carney et al (1994)
Xade complex	XC	c. 1106 (U-Pb baddeleyite)	J. Ramezani and S. Bowring (unpublished data)
Kgwebe rhyolites	KR (See legend)	1106 ± 2 (U-Pb zircon) 1106 ± 4 (U-Pb zircon) 1104 ± 16 (SHRIMP U-Pb zircon)	Schwartz et al (1996) Singletary *et al.* (2003) Kampunzu *et al.* (2000a)
Granites related to Kgwebe rhyolites	Not Shown	1107 ± 2, 1108 ± 1 (U-Pb zircon)	Singletary *et al.* (2003)
Gabbro west of NW Botswana Rift	Not Shown	1107 ± 1 (U-Pb zircon + baddeleyite)	Singletary *et al.* (2003)
Namibia			
Rhyolites related to Kgwebe rhyolites	See legend	1102 ± 7 (U-Pb zircon) 1094 ± 20 (U-Pb zircon)	Pfurr *et al.* (1991) Hegenberger and Burger (1985)
Metavolcanic rocks in Ekuja Dome, Damara Belt		1115 ± 13, 1084 ± 7 (U-Pb zircon)	Steven *et al.* (2000)
Grunehogna Province			
Borgmassivet	GP	1107 ± 2 (U-Pb baddeleyite + zircon)	R. Tucker, in Hanson *et al.* (In

intrusions			review)
		1083 ± 9 (K-Ar, mean of six phlogopite from one intrusion)	Moyes *et al.* (1995)
		1802 ± 100 to 767 ± 49 (Rb-Sr whole rock errorchron or isochron)	Eastin *et al.* (1970), Barton and Copperthwaite (1983), Tingey (1991), Moyes et al (1995)
		1183 ± 33, 1109 ± 33 (K-Ar biotite)	
Ritscherflya basalts	GP	842 ± 30 to 666 ± 26 (K-Ar plagioclase)	Peters *et al.* (1991)
		1115 ± 37 to 666 ± 22 (K-Ar plagioclase)	Peters *et al.* (1991)
		699 ± 14 to 460 ± 16 (K-Ar groundmass)	
		876 ± 86 (Rb-Sr whole rock)	Peters *et al.* (1991)
		821 ± 58 (Rb-Sr whole rock)	Peters *et al.* (1991)
			Watters *et al.* (1991)
			Eastin *et al.* (1970)

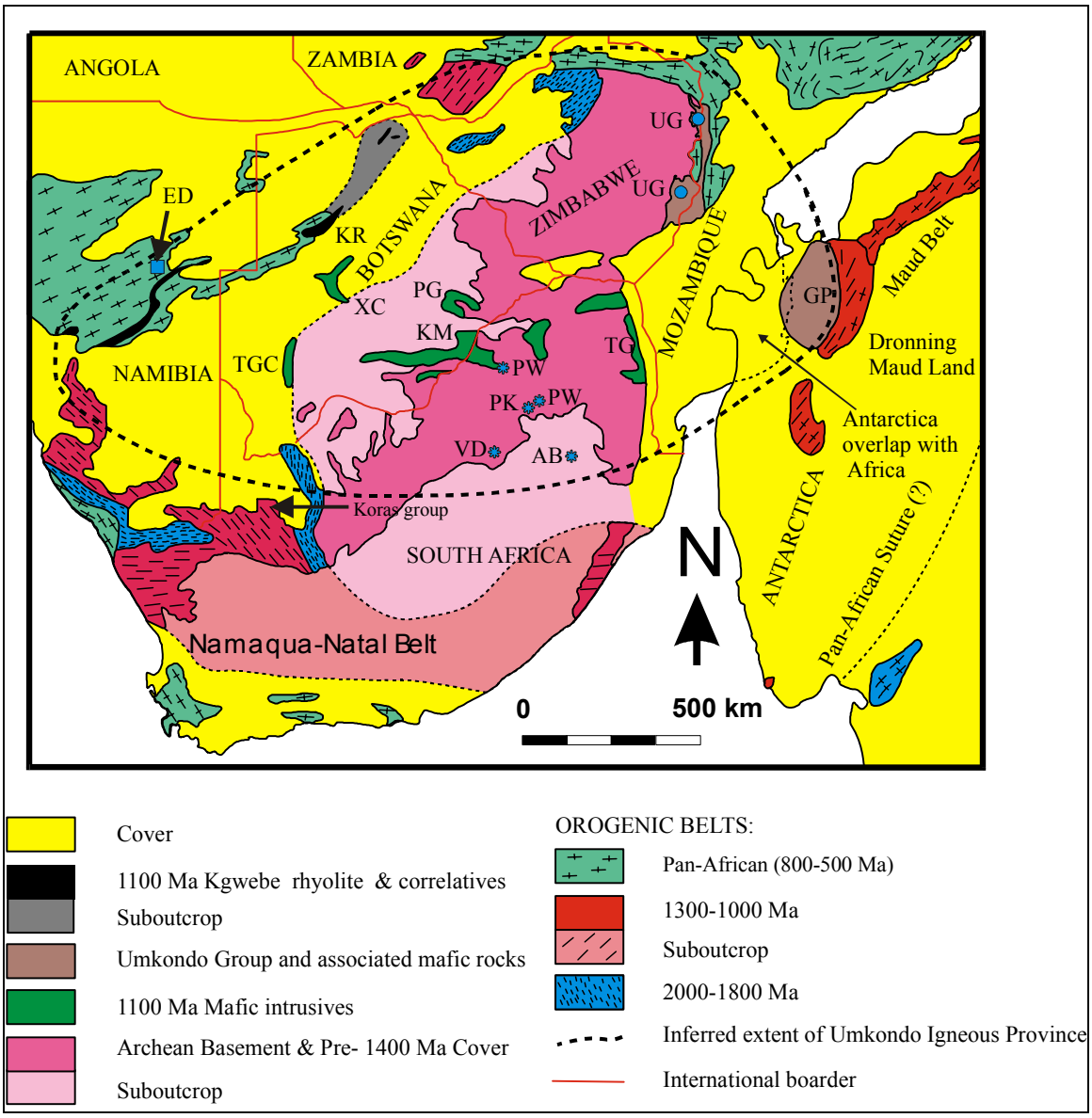

Fig. 2.3. Spatial distribution of mafic and felsic igneous rocks emplaced at ~ c. 1100 Ma across the Kalahari craton (KC) in southern Africa and Dronning Maud Land, Antarctica. The Umkondo Igneous Province of Hanson *et al.* (1998b) is inferred to encompass much of the Kalahari craton including the original continuation of this craton into East Arntarctica. The possible members of the Umkondo Igneous Province are: UG: Umkondo dolerites, TG: Timbavati Gabbro, PW: Post Waterberg dolerites, PK: Dolerite sill intruding the Premier Kimberlite, AB: Dolerite sill intruding Archean Basement, VD: Dolerite sill intruding Vredefort Dome, PG: Mogkware and Dibete-Shoshong, KM: Kanye Mochudi suite, TGC: Tshane gabbroic complex, XC: Xade complex, KG: Kgwebe rhyolites, GP: Grunehogna Province. Modified after R. Hanson (*Unpublished manuscript*) and Hanson *et al.* (In Press).

2.8 Supercontinent Assembly - Rodinia

The Rodinia supercontinent, which is the precursor to Gondwana, begun assembling at ~1300 Ma. All existing continental crust had probably been incorporated into Rodinia by c. 1100 Ma, and began to disperse at ~ 750 Ma (Hoffman, 1989). The Umkondo Igneous Province in the Kalahari craton has similarities in age and tectonic setting to the Keeweenawan mafic province in Laurentia inboard of the Grenville Belt (Fig. 2.4; Hanson *et al.*, 1998b; Powell *et al.*, 2001). Mesoproterozoic belts (Fig. 2.4) that record orogenesis from ~ c. 1150 to 950 Ma give evidence of Rodinia assembly. The c. 1100 Ma magmatic event was contemporaneous with extensive intraplate magmatism in the Keeweenawan mafic province (Hanson *et al.*, 1998b). The two igneous provinces are therefore possibly an expression of a single, extremely large magmatic event within a Mesoproterozoic supercontinent, Rodinia, (Hanson *et al.*, 1998b). Hanson *et al.* (1998b) suggest that the igneous provinces may provide critical constraints on positions of the Kalahari and Laurentia within the Rodinia cratons at c. 1100 Ma.

The exact position of the Kalahari craton within Rodinia configuration at c. 1100 Ma (Fig.2.4) has been a subject of many publications and many of the reconstructions have placed the Kalahari craton near the southern margin of Laurentia (Dalziel 1991, 1997; Hartnady, 1991; Hoffman, 1991; Hanson *et al.*, 1998b; Weil *et al.*, 1998). Three possible positions for the Kalahari craton have been proposed for Rodinia. Dalziel (1991, 1992, 1995) and Hoffman (1991) positioned the Kalahari craton adjacent to East Antarctica as part of East Gondwana and the southern tip of Laurentia (Fig. 2.4, Position I). This configuration permits alignment of the Umkondo and Laurentian igneous provinces and the coeval Mesoproterozoic orogenic belts, i.e. the Namaqua-Natal and the Grenville belts that partially wrap the two igneous provinces (Fig. 2.4). However, a palaeomagnetic test by Gose *et al.* (1997) has shown that available data do not permit positioning the Kalahari craton on the southwestern margin of Laurentia at ~ 1100 Ma since the palaeomagnetic data require a substantial separation between the two cratons (Hanson *et al.*, 2004). The c. 1105 Ma palaeopoles for the Kalahari and Laurentia cratons locates the Kalahari craton to the east or southeast of Laurentia. A contrasting model positions the

Kalahari craton in West Gondwana adjacent to the Rio de la Plata craton of South America, with the Kalahari craton faced in the opposite direction (Fig. 2.4; Position II). The model is compatible with palaeomagnetic data, but the solution is not unique (Dalziel *et al.*, 2000).

Palaeomagnetic data also permit placing the Kalahari craton in an isolated position directly opposed to the southern margin of Laurentia (Dalziel *et al.*, 2000, Fig. 2.4, Position III). The Kalahari craton is rotated such that the Namaqua-Natal-Maud belt faces the Grenville belt (Fig. 2.4). The Llano belt, which is the southern segment of the Grenville belt, is interpreted to have developed as a result of a "southern continent" colliding with the southern margin Laurentia in the period 1150 to 980 Ma (Mosher, 1998; Grimes, 1999). In this model (Fig. 2.4, Position III), the Kalahari craton is inferred to represent the indentor to the Laurentia craton, which is compatible with the shape of the Llano orogenic belt and kinematics of the contractional deformation along the belt (Dalziel *et al.*, 2000). The Llano orogenic and the Namaqua-Natal belts are interpreted to reflect craton-craton convergence along the present southern margin of the Laurentia craton and southwestern margin of the Kalahari craton during the assembly of Rodinia. Orogenesis occurred simultaneously between 1150 and 1090 Ma at the centre of the Llano and Namaqua-Natal collisional belts. The Himalayan-type collision of the two cratons generated extensional stresses within the cratons resulting in formation of rifts in the Kalahari and Laurentia cratons (Dalziel *et al.*, 2000). In the Kalahari craton, the Ghanzi-Chobe belt of northwestern Botswana has been inferred to represent a failed intracontinental rift that has bimodal volcanism dated c. 1106 ± 2 Ma (Modie, 1996). The north-northwest trending Pacos mafic intrusive suite formed in a rift basin, within the Laurentia craton, perpendicular to the collisional margin (Keller *et al.*, 1989; Adams and Keller, 1994, 1996). Although the structure in the Namaqua-Natal and Llano belts are compatible with the Dalziel (2000) model, the late Mesoproterozoic apparent polar wander paths for the Kalahari and Laurentia cratons do not support collision of the two cratons between c. 1150 and 1120 Ma (Powell *et al.*, 2001). Thus the position of the Kalahari craton remains uncertain.

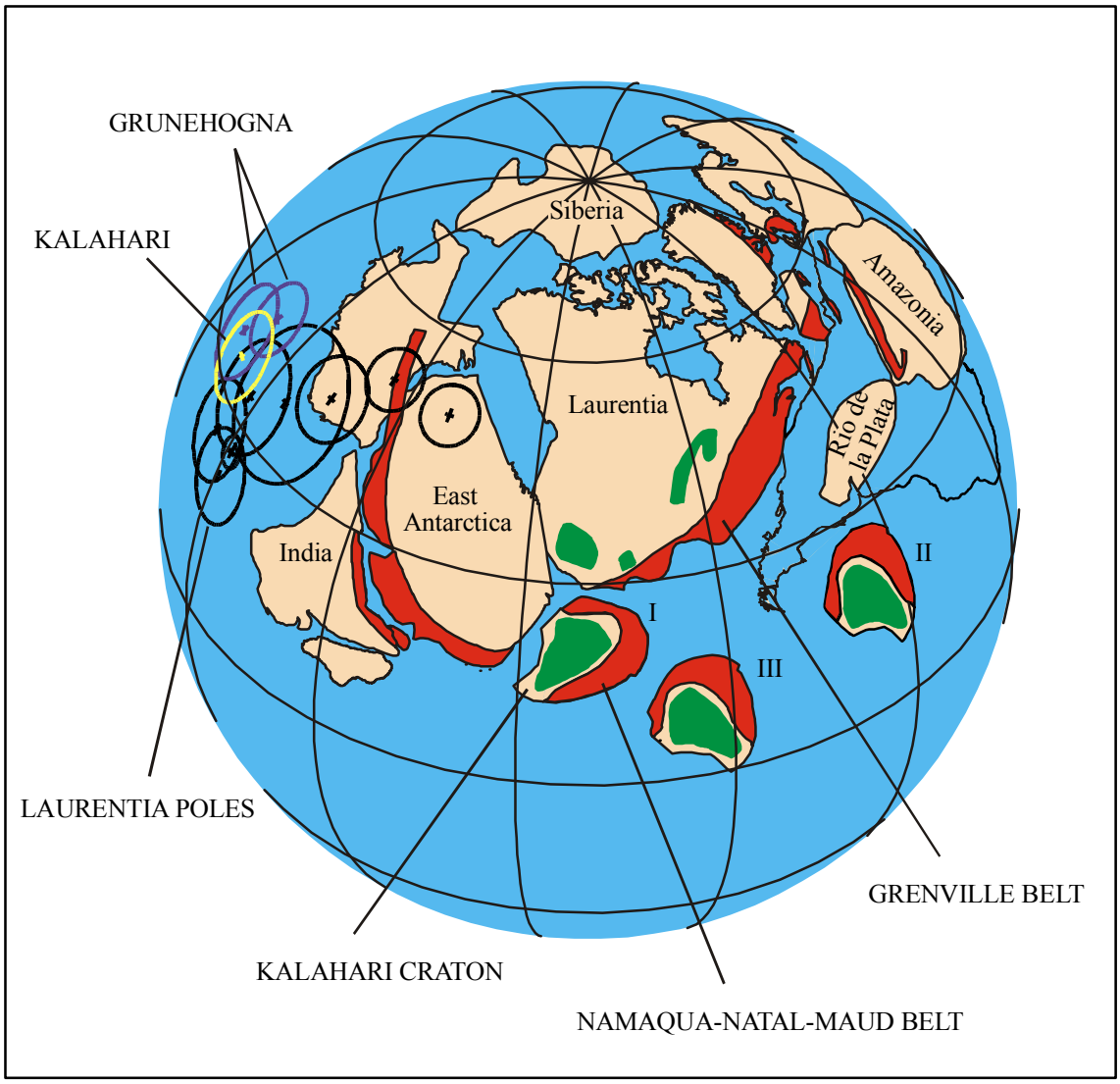

Fig. 2.4. Hypothetical configuration of Rodinia supercontinent of ~ 1000 to 750 Ma. Three possible positions are shown for the Kalahari craton: I, in East Gondwana adjacent to the East Antarctic craton (Hoffman, 1991; Dalziel, 1991, 1992); II, in West Gondwana adjacent to the Rio de la Plata craton of South America (Dalziel, 1997; Gose *et al.,* 1997); and III, in an isolated position directly opposed to the southern margin of Laurentia (Dalziel *et al.*, 2000). Mesoproterozoic orogenic belts are shown in brown colour.

2.9 Discussion and Conclusions

Geochemical work by Ward *et al.* (1999, 2000) and Ward (2002) have shown that dykes defining the Kamativi, Guruve and Mutare dykes can be classified into subgroups that have complex distinct signatures. Simple fractional crystallisation, partial melting, or crustal contamination models cannot explain the geochemical heterogeneity of these subgroups within each of the swarms. Therefore the swarms of Wilson et al (1987) are strictly not swarms since a dyke swarm should comprise dykes that are contiguous, and contemporaneous (Ward *et al.*, 1999, 2000). Major element data, trace and REE geochemistry and isotopic characteristics suggest three possible magma sources for the Guruve and Mutare dykes. Type III dykes are geochemically similar to Umkondo sills (Hanson *et al.*, 2004*).* Thus the possibility of the occurrence of dykes of Umkondo age amongst the Kamativi and Guruve dykes cannot be ruled out and their presence would give evidence for the occurrence of the Umkondo magmatic event locally across the whole of Zimbabwe. The heterogeneity in petrography, degrees of deformation and geochemistry in dykes that have previously been grouped in the dyke swarms of Wilson *et al.* (1987), but trend in the same direction, are interpreted to reflect structural control on magma emplacement at different ages along fractures that were possibly propagated during a single tectonic regime. Different subgroups of dykes were possibly generated from different magma sources and were emplaced at different stages of the assembly and breakup of Rodinia and Gondwanaland. Precise dating of these dykes is therefore necessary for re-defining dyke swarms in Zimbabwe as well as supercontinent assembly and dispersal.

The Chimbadzi dyke dated at c. 2262.6 ± 1.5 Ma (Manyeruke 2001), although parallel to the Guruve dykes, belongs to a unique mafic igneous event that is not related to any documented major magmatic events in Zimbabwe. In view of the age of the Chimbadzi dyke, there may be need to revisit the approximately similar 2.2 – 2.4 Ga ages (Kamber *et al.*, 1995) documented in the Northern Marginal Zone. These ages have been interpreted as geologically meaningless mixing ages, but in fact they may represent

cooling ages that were associated with a tectonothermal event across the Zimbabwe craton that was related to the emplacement of the Chimbadzi dyke.

The position of the Kalahari craton in Rodinia at c. 1100 – 750 Ma is poorly defined owing to insufficient precisely dated palaeomagnetic poles from that craton. Most of the available key palaeopoles for Rodinia models have been derived from a single craton, Laurentia. A lot more key palaeopoles from precisely dated rocks from other cratons other than Laurentia are needed to produce a perfect model of Rodinia using palaeomagnetism.

Although the Umkondo group in eastern Zimbabwe has only a relatively limited distribution in southern Africa, it is significant in that it is the type area for the larger Umkondo igneous province, and it may in itself also have implications for Rodinia reconstructions (Hanson *et al.*, 1998b). The Umkondo basin is located on the edge of the Kalahari craton. Understanding the palaeogeography of the Umkondo group may therefore provide useful constraints on Rodinia reconstructions. If this edge of the Kalahari craton was originally adjacent to another craton within Rodinia, then a perfect model of Rodinia at c.1100 Ma should show the Umkondo sedimentary basin on the Kalahari craton matching with a coeval basin on the opposite craton.