

ANALYSIS OF PRE-IMPACT AND IMPACT-INDUCED GEOLOGICAL STRUCTURES IN THE NORTHERN COLLAR OF THE VREDEFORT DOME, SOUTH AFRICA

A Dissertation submitted to the Faculty of Science, University of the Witwatersrand; in fulfilment of
the requirements for the degree of Master of Science.



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DECLARATION

I declare that this Dissertation is my own, unaided work. It is being submitted for the Degree of Master of Science at the University of the Witwatersrand, Johannesburg. It has not been submitted before for any degree or examination at any other University.

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12th day of August 2016, in Johannesburg.

ABSTRACT

Rocks of the Neoarchaeon Witwatersrand Supergroup exposed in the collar of the impact-induced 2.02 Ga Vredefort Dome exhibit complex geological structures. These structures are generally considered to have been formed by the Vredefort impact event, through rapid deformations on time scales of seconds to minutes associated with the relatively brief impact processes. However, geological mapping of the structures and petrographic analysis from the northern collar of the dome show that the collar hosts at least three generations of pre-impact structures. In contrast to impact-induced structures, these pre-impact structures indicate slow and progressive deformations that are uncharacteristic of impacts.

The pre-impact deformations comprise: (a) an extensional D1 deformation characterised by listric faults up to kilometre-scale; (b) Syn-metamorphic ($M_{2(NC)}$) D2 ductile deformation characterised by regional S_2 foliation, which locally indicates northwest-directed vergence; and (c) D3 deformation that crenulated the pre-existing S_2 foliation (S_3). Pre-impact structures can be distinguished from impact-induced structures by: (1) difference in the geometry and sense of slip between D1 faults and D4 impact-induced faults; and (2) crosscutting relationships between impact-induced D4 features and D2 and D3 pre-impact features.

In their present (rotated) orientation, the D1 faults exhibit an apparent strike-slip separation, which translates to normal-slip fault geometries when impact-induced overturning of strata is undone. Displacement affects the Witwatersrand and Ventersdorp Supergroup rocks but no offset is observed of the base of the Transvaal Supergroup. The faults also exhibit a listric geometry, curving into parallelism with bedding in the lower West Rand Group. In their restored orientation, faults define half-graben and horst blocks, synthetic and antithetic faults, and rollover and drag folds, which are typical for extensional tectonics. These geometries and crosscutting relationships of the D1 faults are similar to that of the Neoarchaeon listric faults described in the Witwatersrand goldfields and the wider Kaapvaal craton, that exhibit a general west-side-down sense of slip (2.70-2.64 Ga Hlukana-Platberg extensional event).

Metamorphic grade in the study area decreases from amphibolite- to greenschist-facies away from the centre of the dome. These are largely $M_{2(NC)}$ metamorphic assemblages that are attributed to elevated regional heat flow related to 2.06 Ga Bushveld magmatism. There is some evidence that $M_{2(NC)}$ metamorphic mineral assemblages developed along the same stratigraphic units differ across the large D1 faults, indicating the pre-impact nature of the D1 faults and implying that the $M_{2(NC)}$ metamorphism occurred after the Hlukana-Platberg event. Also, $M_{2(NC)}$ assemblages are syn-tectonic to the S_2 foliation hosted in metapelite units of the West Rand Group and knotted quartzite horizons of the Central Rand Group. The S_2 foliation is attributed to the post-Transvaal Supergroup, compressional, Ukubambana Event. Crosscutting relationships in the study area indicate a deformational period of 2.06 Ga to no less than 2.02 Ga. The northwest-directed vergence exhibited

by the S_2 foliation is broadly consistent with the regional, general north-directed, vergence exhibited by post-Transvaal Supergroup foliation developed in the northeastern collar and the Johannesburg Dome. The S_2 foliation and $M_{2(NC)}$ mineral assemblages are crosscut by D4 pseudotachylitic breccia, micro-faults and kinks, and $M_{4(NC)}$ metamorphic features associated with the impact.

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1.0 INTRODUCTION

1.1. Regional Context

The Witwatersrand Basin in South Africa (Figure 1.1) hosts approximately 30% of all known gold reserves, making the basin the world's largest gold province (Beach and Smith, 2007). Studies of the Witwatersrand Basin by Pretorius (1975), Minter (1976), Robb and Meyer (1985), Beukes (1990), Barnicoat et al. (1997), and Phillips and Law (2000) have focused on understanding how the gold was concentrated, and proposed models for its formation. These studies have shown that in addition to sedimentary processes, the post-depositional deformational and thermal history of the basin were influential in the development and evolution of the gold deposits (Robb and Robb, 1998; Jolley et al., 2004; Dankert and Hein, 2010).

Pretorius (1975), Minter (1976), and Beukes (1990) proposed that the gold deposits were formed by alluvial placer processes. In contrast, Barnicoat et al. (1997), Phillips and Law (2000), and Jolley et al. (2004), amongst others, proposed formation by hydrothermal processes, whereas Robb and Meyer (1985) proposed a modified placer model for ore formation. The modified placer and hydrothermal models for ore formation were supported by structural data by Robb and Robb (1998) and Jolley et al. (2004). The analyses of structural data included studies by Barnicoat et al. (1997), Jolley et al. (1999, 2004) and Phillips and Law (2000), which proposed that the gold, uraninite and hydrocarbons were introduced into the basin by hydrothermal fluids associated with thrust-fracture networks that formed after basin sediments were lithified. The demonstration of structurally-hosted gold deposits eventually led to early structural geology reviews of the Witwatersrand Basin by Coward et al. (1995) and later Dankert and Hein (2010; and references therein), comparing structures described across goldfields. Reviews of the structural geology, combined with seismic interpretations, showed that rocks of the basin were subjected to more than one deformation event (Coward et al., 1995; Beach and Smith, 2007; Jolley et al., 2007; Dankert and Hein, 2010; Manzi et al., 2013). These included:

- 1) A regional compressional deformation at 2.90-2.72 Ga (McCarthy et al., 1982; Roering et al., 1991; Barnicoat et al., 1997; Brink et al., 2000), divided into the Asazi Event (2.90 Ga; Manzi et al., 2013), and the Umzawami Event (2.73 Ga; Dankert and Hein, 2010).
- 2) Extensional tectonics at 2.70-2.64 Ga (Gibson et al., 2000; Frimmel and Minter, 2002; Tinker et al., 2002; Dankert and Hein, 2010; Manzi et al., 2013), termed the Hlukana-Platberg Event by Manzi et al. (2013).

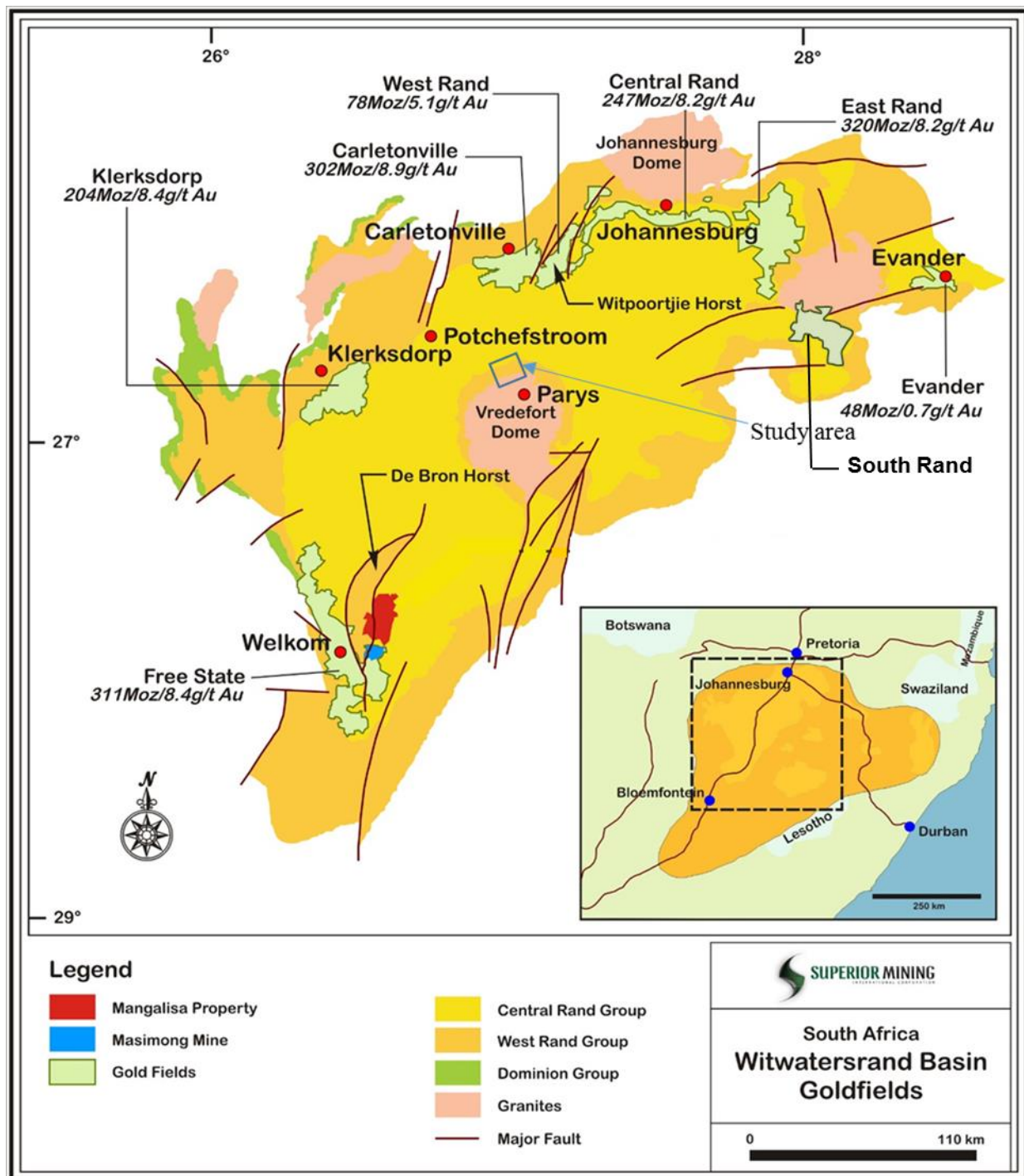


Figure 1.1. A regional geological map of the Witwatersrand basin, with younger overlying rocks excluded. The Vredefort Dome is situated within the basin, surrounded by the seven main goldfields in the basin (Map published with permission from Superior Mining International Corporation, with an addition of the South Rand goldfield).

The Witwatersrand Supergroup is exposed along the margins of the Witwatersrand basin (Figure 1.1) and in the collar (the outer ring of supracrustal strata surrounding the Archaean Basement Complex core) of the Vredefort Dome (Lana et al., 2003a, 2003b; Wieland et al., 2005). Producing goldfields are located along the margins of the basin at Welkom, Klerksdorp, Carletonville, West Rand, Central Rand, East Rand, Evander, and South Rand (Figure 1.1). According to McCarthy et al. (1986, 1990), Grieve and Masaitis (1994), and Reimold et al. (2005), down-faulting and down-folding associated with the 2.02 Ga Vredefort impact preserved large parts of the Witwatersrand Supergroup from erosion. The entire Witwatersrand Supergroup stratigraphy is exposed in the collar of the Vredefort Dome because the Mesoarchaeon to Paleoproterozoic geology was rotated by $\geq 90^\circ$ during the Vredefort impact (Lana et al., 2003b; Wieland et al., 2005; Gibson and Reimold, 2008). This exposure provides an opportunity to study the entire succession, together with the underlying Dominion Group and Archaean Basement Complex, and overlying Ventersdorp and Transvaal supergroups at surface. The surface exposure provides a different approach to studying the structure of the Witwatersrand Supergroup, compared to structural studies in the goldfields conducted through underground mine mapping, interpretations of borehole data, and 2-D and 3-D seismic modelling (e.g., Coward et al., 1995; Tinker et al., 2002; Jolley et al., 2004; Gibson, 2005; Beach and Smith, 2007; Manzi, 2010; Mambane et al., 2011; Mashabela, 2011; Manzi et al., 2013).

2-D and 3-D seismic structural studies show that the Witwatersrand Supergroup in the goldfields hosts large-scale 2.90-2.72 Ga folds and thrust faults, as well as 2.70-2.64 Ga listric faults that displace the Witwatersrand Supergroup and the unconformably overlying Ventersdorp Supergroup (Tinker et al., 2002; Manzi et al., 2013). Because the Witwatersrand Supergroup is also exposed in the collar of the Vredefort Dome, it is logical to propose that geological structures that formed prior to the impact event should be exposed in the collar and that they may have been reactivated during the impact event. This means that regional structures identified in the goldfields using seismic profile interpretations could be validated by surface mapping of the Vredefort Dome, at the same time as providing a rare opportunity to study crosscutting relationships of impact-induced structures with the pre-impact structures, because the meteorite impact is associated with multi-stage deformations (French, 1998). Therefore, the main aim of this study has been to distinguish impact-induced structures from pre-impact structures in the Vredefort Dome, and compare the geometry of the pre-impact structures with that observed in the goldfields.

Additionally, the geological history of the Witwatersrand Supergroup involves metamorphism (Gibson et al., 1997b; Gibson and Reimold, 2008), which has been an essential aspect in differentiating impact-induced structures from pre-existing ones in the Vredefort Dome. Metamorphic studies have shown that the dome is a polymetamorphic terrain in which assorted rocks have undergone at least four episodes of metamorphism. The oldest metamorphism (M_1) affected the basement granite at approximately 3.10 Ga (Lana et al., 2003a, 2003b; Ogilvie, 2010). M_2 (2.72-2.69 Ga; zircon U-Pb geochronology; Schmitz and Bowring, 2003) is documented as an ultra-high-

temperature metamorphism that formed granulite facies lower crustal assemblages as evidenced in kimberlite xenoliths south of the dome (Dawson and Smith, 1987; Dawson et al, 1997; Schmitz and Bowring, 2003). M_3 metamorphism affected supracrustal rocks at approximately 2.06 Ga (Gibson and Stevens, 1995; Gibson and Wallmach, 1995); and M_4 metamorphism resulted from shock heating associated with the Vredefort impact (Gibson, 1993; Gibson and Wallmach, 1995; Gibson et al., 1998; Foya, 2002). The M_2 and M_3 metamorphic events are essential in crosscutting relationships exhibited in the study area because they occurred after deposition of the Neoarchaeon Witwatersrand Supergroup, and before the Vredefort impact at 2.02 Ga. The analyses of metamorphic assemblages across pre-impact structures, coupled with crosscutting relationships of structural features in the dome, have proved important in distinguishing pre-impact structures from impact-induced structures in this study.

1.2. Location of the study area

The Vredefort Dome is located in the southeastern part of the Kaapvaal Craton, South Africa. The dome is situated within the Witwatersrand Basin, some 100 km southwest of Johannesburg (Figure 1.1). It comprises a core of Mesoarchaeon basement composed of granitoid gneisses, granites and mafic gneisses; which are surrounded by a collar of Neoarchaeon to Palaeoproterozoic supracrustal rocks whose differential erosion has formed the semicircular ridges of the Vredefort mountainland (Henkel and Reimold, 2002; Lana et al., 2003a; Wieland et al., 2005; Reimold and Gibson, 2006). The study area ($10 \times 11 \text{ km}^2$) is located in the northern collar of the Vredefort Dome, and covers a number of privately-owned farms (Figure 1.2). Permission to access some farms was not granted, limiting access to some outcrops.

Farm boundaries in Figure 1.2 were constructed using information gathered from old maps, personal communication with the farmers, field mapping, and Google EarthTM images. Most of the important structural features referenced in this thesis are located on Berakah, Koedoesfontein, Donkervliet, Doringbos, Eldorado, Kopjeskraal, Mooiplaas, Oppie Plaas, Rondekop, and Buffelshoek farms (Figure 1.2).

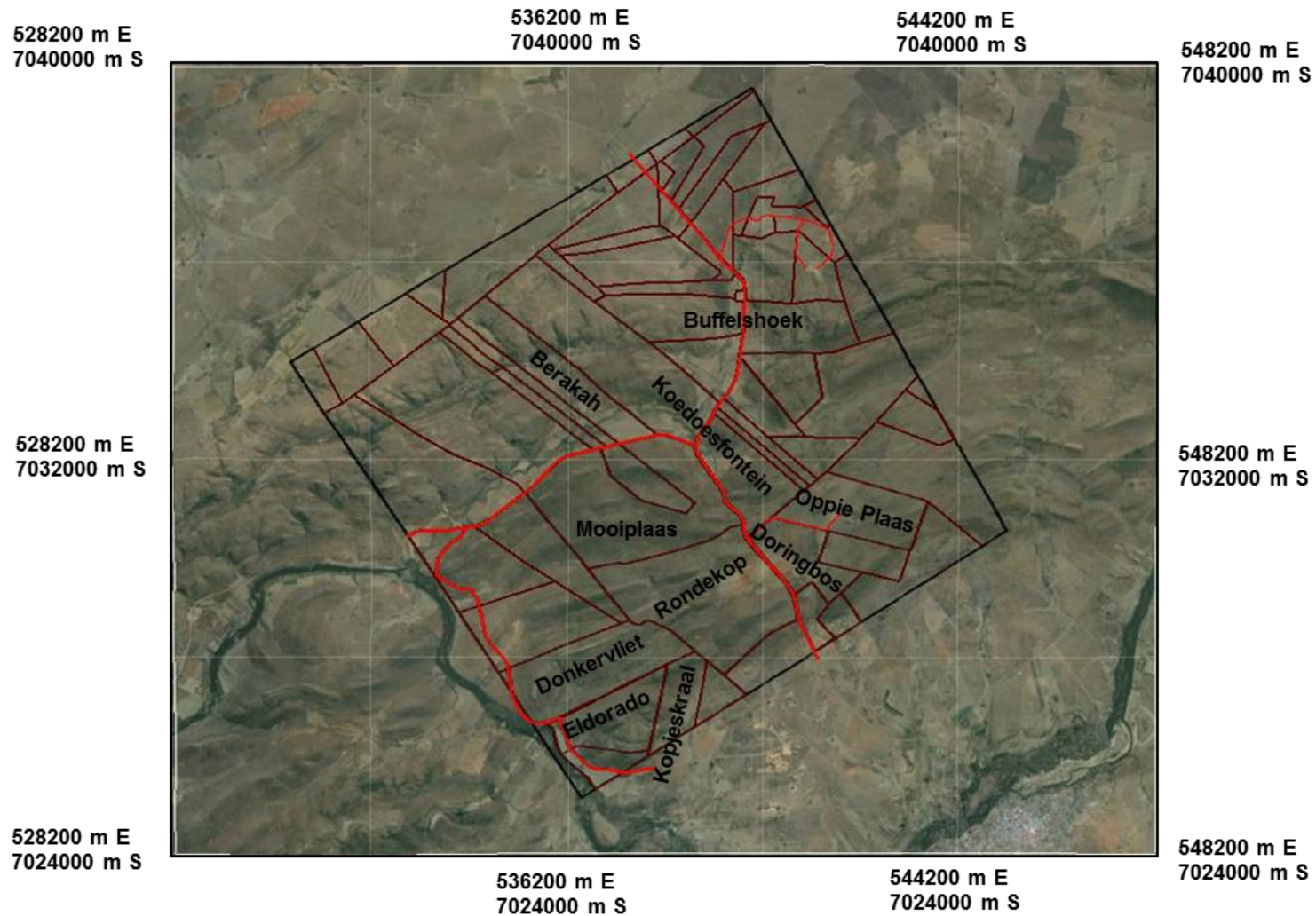


Figure 1.2. Google EarthTM image of the northern collar of the Vredefort Dome showing farm boundaries in the study area. This figure includes names of farms that are referenced in the thesis, within which some important structural features are hosted.

1.3. Aims and hypothesis

Previous studies by Nel (1927), Simpson (1977), Lilly (1978), Colliston and Reimold (1990), Corner et al. (1990), Martini (1992), Lana et al. (2003a, 2003b) and Wieland et al. (2005), and orthophotographs and satellite images indicate that the collar of the Vredefort Dome is structurally complex. Complications include: (1) brittle and ductile structures, and variable fault offsets and throws that cannot be attributed to a single event; (2) kilometre-scale tangential and radial fold geometries (Wieland et al., 2005); (3) contrasts in bedding orientation across the collar rocks (overturned versus right-side-up strata and tangential to locally radial strikes); and (4) palaeostress directions that are not consistent with the deformation model suggested for the orientation of the dome, as interpreted by Lilly (1978).

As shown by French (1998), meteorite impacts the size of the Vredefort event form highly variable structural features, ranging from micrometre-scale, planar deformation features (PDF), to centimetre-scale shatter cones and micro-faults, to dip-slip and strike-slip faults with displacements up to hundreds of metres (French; 1998; Gibson and Reimold, 2008; Wieland et al., 2005). Crosscutting relationships of these structural features are complicated by the multi-stage nature of impacts and by the similarity of some of the structures to structures formed by alternative geological processes (e.g., shatter cones/multiply-striated joint surfaces/striated slip surfaces; see discussion in Gibson and Reimold, 2008). Besides the fact that large meteorite impacts form structures that are complex on their own, earlier deformation events affecting the target rocks over the ~1 Ga between their deposition and the impact, would have further complicated the resultant structural architecture of the target rocks (e.g., the effect of the pre-impact Foch Thrust and Potchefstroom-Master Plane Fault on the formation of the Potchefstroom Synclinorium during impact; Brink et al., 2000).

Lana et al. (2003a), Lilly (1978) and Wieland et al. (2005) proposed that pre-impact structures and impact-induced rotation of fault-blocks locally affected the orientation of strata in the collar of the Vredefort Dome. Wieland et al. (2005) discussed how impact-induced rotation of fault-blocks affected the geometry of the collar; however the effects of pre-impact structures on the dome and their nature have not been extensively studied. Furthermore, Tinker et al. (2002) correlated pre-impact faults hosted in the western part of the Potchefstroom synclinorium (Chapter 2, Figure 2.2), to listric faults in the goldfields that were active during the deposition of the Ventersdorp Supergroup. If the impact-induced synclinorium hosts Neoarchaean structures, the collar of the dome possibly exhibits these structures.

In addition to the large-scale structural complexity of the collar, the formation of pseudotachylitic breccias in the dome and the goldfields has been attributed to the Vredefort impact by Trieloff et al. (1994), Kamo et al. (1996), and Reimold and Gibson (2006). Reimold and Gibson (2006) suggested that most of the breccias in the dome were formed by shock melting associated with

the impact, with or without frictional melting. However, Mambane et al. (2011) described pseudotachylite veins hosted in the listric West Rand Fault. They concluded that the veins were formed by frictional melting along the listric fault system, an interpretation consistent with that of Killick and Reimold (1990), and Berlenbach and Roering (1993). This means that pre-impact pseudotachylite-bearing listric faults may be hosted in the Witwatersrand Supergroup exposed in the collar of the Vredefort Dome. As a result, this study provided an opportunity to test whether pre-impact pseudotachylitic breccias are exposed in the study area.

This structural analysis of the northern collar focusses on distinguishing pre-impact from impact-induced structures in the study area. In this way, the study aimed to establish the nature of both the pre-impact and impact-induced structures and the implications for pre-impact tectonics in the wider Witwatersrand basin. This was achieved through:

- a) Mapping the northern collar of the Vredefort Dome. This technique: (1) obtained lithological data for establishing the stratigraphic succession and identifying marker horizons that indicated the sense and magnitude of fault displacements; (2) established the character and magnitude of displacements associated with large-scale pre-impact and impact-induced structures (kinematic analysis of the structures was not possible because fault surfaces are poorly exposed); (3) constrained crosscutting relationships between cleavages and mesoscopic impact-induced structures; and (4) established the metamorphic grade across the field area.
- b) The construction of a geological map that describes the structural, metamorphic and lithological characteristics of the northern collar.
- c) Comparing the nature of pre-Vredefort impact faults in the northern collar to that of faults described in the West Rand and West Wits Line goldfields. This was done by comparing the geometry of pre-impact faults mapped in the study area to 3D seismic sections interpreted in the goldfields by Manzi et al. (2013), including scale and stratigraphic location of the faults within the Witwatersrand Supergroup, magnitude and sense of displacement.
- d) Investigating the timing and character of pseudotachylitic breccia in areas where pre-Vredefort impact structures were identified.
- e) Establishing the structural evolution of the northern collar of the Vredefort Dome that included possible deformation events that took place before the meteorite impact.
- f) Creating Microsoft Excel[®], Google Earth[™], and ArcGIS[™] databases of the lithological descriptions, bedding measurements, and metamorphic and structural data of the mapping area.

1.4. Abbreviations

VCR = Ventersdorp Contact Reef

BLR = Black Reef Formation

WRF = West Rand Fault

BF = Bank Fault

KF = Karolina Fault

LSA = Listric shift area (an area where listric faults flatten out to form a decollement/detachment zone)

PDF = Planar Deformation Features

BPP = Bedding-parallel pseudotachylitic breccia

FHP = Fracture-hosted pseudotachylitic breccia

1.5. Thesis organisation

The thesis is divided into a total of 9 chapters including the Introduction:

- Chapter Two presents a review of: (a) the regional geology, structure, and metamorphism of the Witwatersrand Basin; and (b) the structure, metamorphism, and impact-induced features in the Vredefort Dome.
- Chapter Three presents a breakdown of the methods used to address the aims of the research. The methods include aerial photograph lineament analyses, field mapping, sampling, and petrographic analyses. These methods were aided by the use of software that included Google EarthTM, CorelDRAW[®], ArcGISTM 10.1, GEOrient[®], CellsensTM, Olympus Stream EssentialsTM, and Oxford INCA x-act[®].
- Chapter Four presents the stratigraphy and lithologies hosted in the northern collar of the Vredefort Dome. Lithological descriptions include field observations and petrographic analysis.
- Chapter Five presents the structural geology of the northern collar of the Vredefort Dome. The structures presented range from kilometre-scale to microscopic.
- Chapter Six describes metamorphic assemblages from rocks in the study area. The presentation of these assemblages includes field descriptions and petrographic analysis of metapelite, meta-dolerite, meta-basalt, meta-ironstone, quartzite, and knotted quartzite horizons, amongst others.
- Chapter Seven presents a description of melt-rocks hosted in the study area. The chapter is focused on the geometry, and petrographic and compositional characteristics of the different occurrences of pseudotachylitic breccias, but also includes features of the impact-melt dykes.
- Chapter Eight presents the Discussion.
- Chapter Nine presents the Conclusions.

2.0 LITERATURE REVIEW

2.1. Regional geology of the Witwatersrand Basin

The Witwatersrand Supergroup conformably and/or unconformably overlies the Dominion Group, and is in turn conformably/unconformably overlain by the Ventersdorp and Transvaal supergroups (Figures 2.1, 2.2; Stanistreet and McCarthy, 1991; McCarthy, 2006; Dankert and Hein, 2010). These supracrustal strata were deposited on the Kaapvaal Craton, which was accreted as early as 3.7 Ga (De Wit et al., 1992). The stratigraphy of the Witwatersrand Supergroup is presented in Figure 2.3.

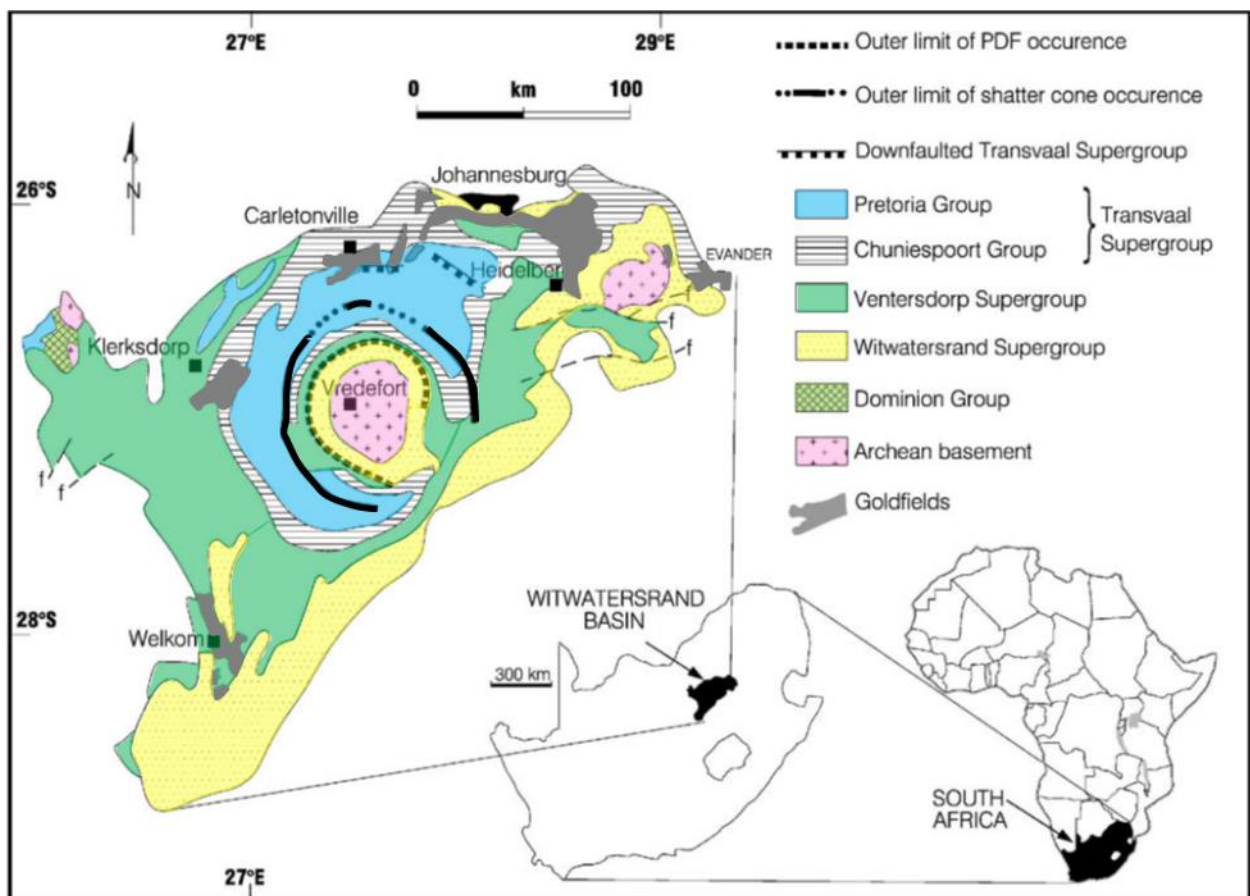


Figure 2.1. Regional geological map of the Witwatersrand Basin with the Karoo Supergroup removed. The Vredefort Dome is situated in the centre of the basin; also indicated are the outer limits for shock-induced shatter cones and PDF in quartzite linked to the impact event (Modified after Reimold and Gibson, 2006).

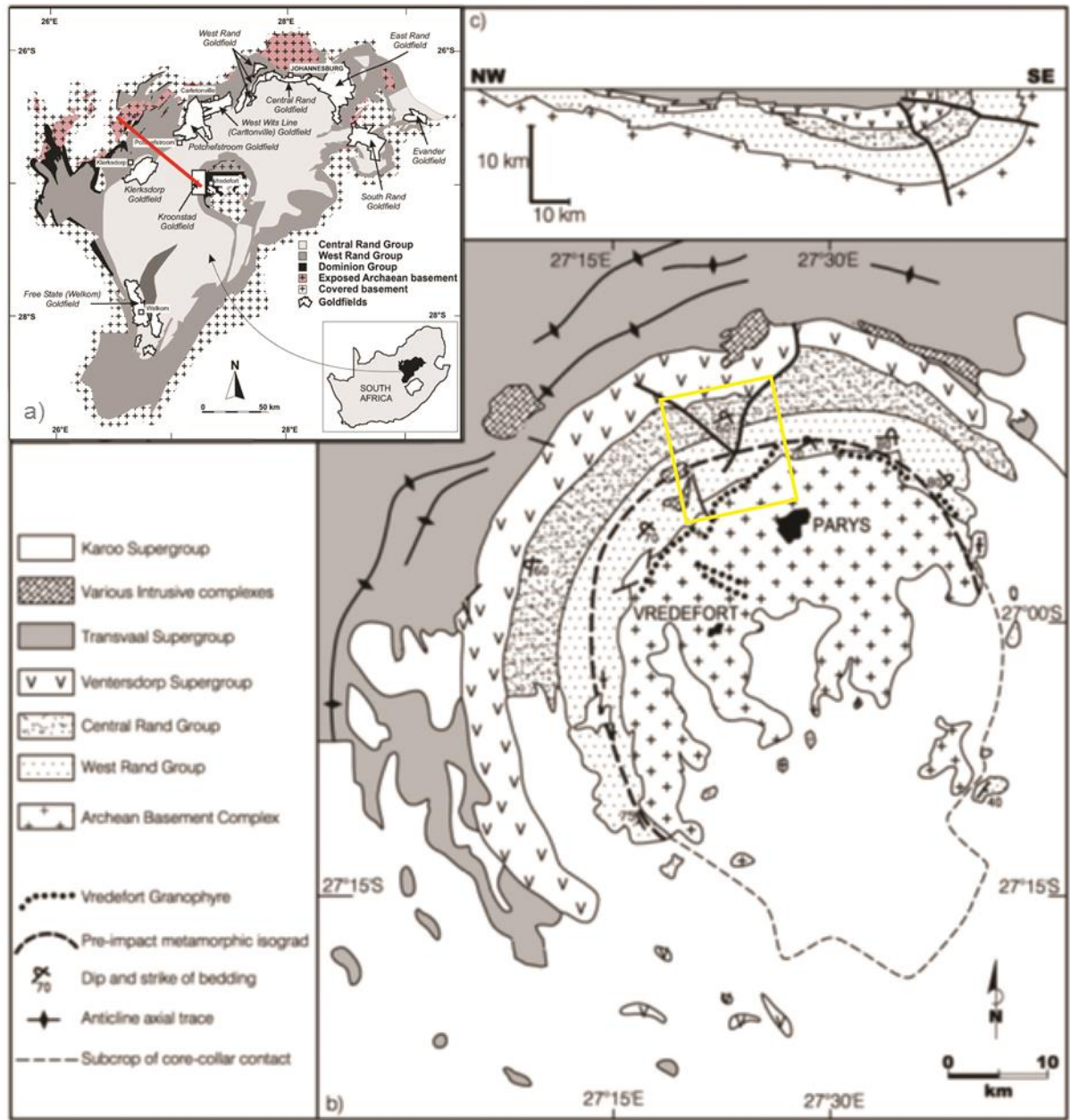


Figure 2.2. (a) Geological map of the Witwatersrand basin, with the Karoo Supergroup strata removed, showing the location of the Vredefort Dome (after Dankert and Hein, 2010). (b) Geological map of the Vredefort Dome showing major geological structures and the fold hinge of Potchefstroom Synclinorium (after Wieland et al., 2005). The yellow box indicates the study area which straddles two major faults. (c) NW-SE cross-section across the Witwatersrand Basin, from the Vredefort Dome to the goldfields (section line shown in (a); after Wieland et al., 2005).

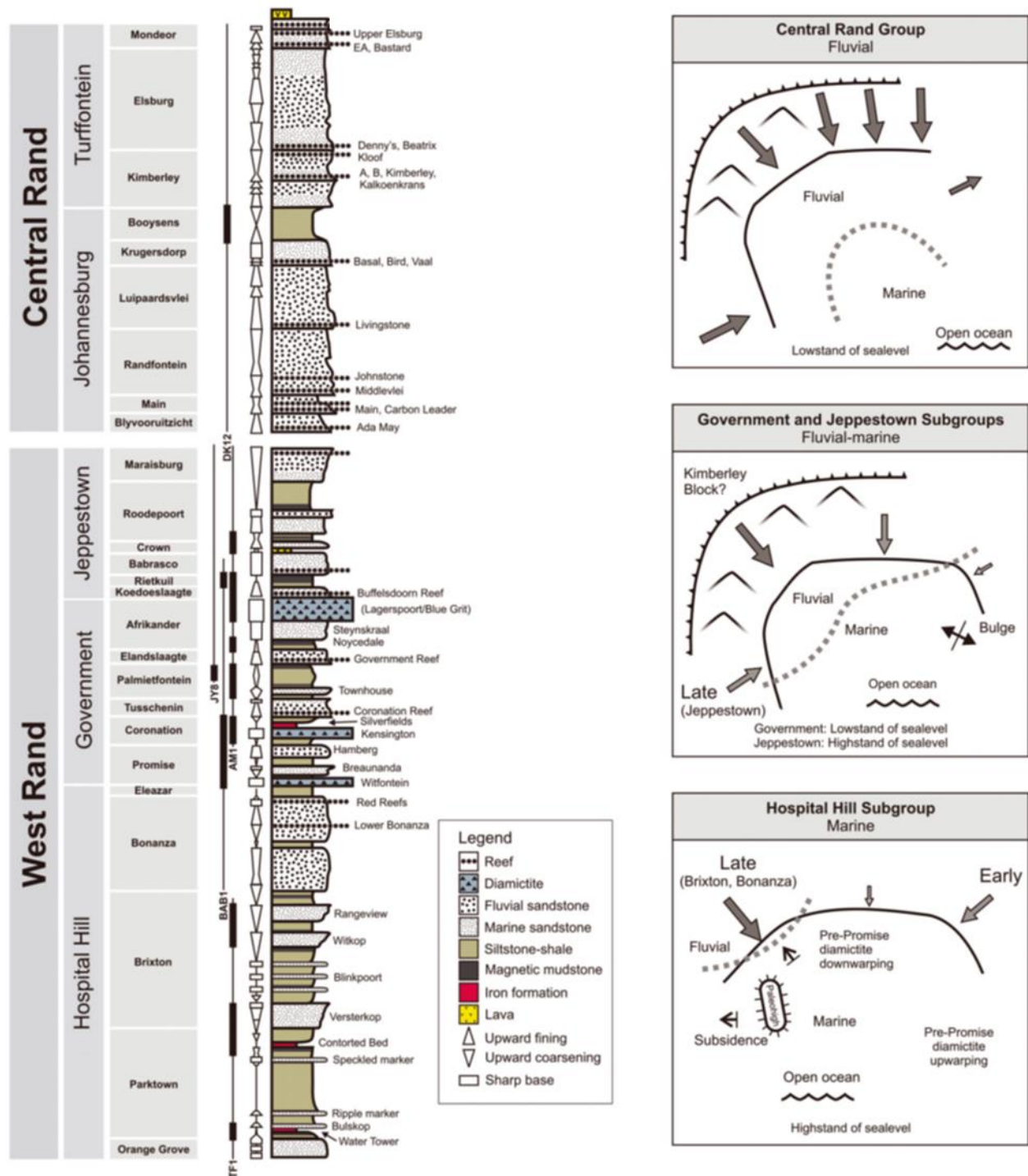


Figure 2.3. Generalised stratigraphic profile of the Witwatersrand Supergroup. Sketches of interpreted depositional environments in the Witwatersrand Basin (from Guy et al., 2010).

2.1.1. *Dominion Group*

Armstrong et al. (1991) estimated the age of the Dominion Group to be 3074 ± 6 Ma, dated from authigenic zircons in the basalts (zircon U-Pb SHRIMP). The Dominion Group is divided into three formations (Bisschoff, 2000; Marsh, 2006). The basal Rhenosterspruit Formation is composed of quartzite and meta-conglomerate units up to 120 m in thickness. It is overlain by the Rhenostershoek Formation that is made up of mafic to intermediate lavas (200-1500 m in thickness). The Rhenostershoek Formation is overlain by the Syferfontein Formation, composed of felsic volcanic and porphyritic rocks, which range from a few hundred metres to >1500 m. The maximum thickness of the Dominion Group is reached in the Klerksdorp area situated northwest of the Dome; however the thickness does not exceed 200 m in the study area, and appears to be absent in the southern parts of the Dome (Bisschoff, 2000; Jackson, 1994; Lana et al., 2006). Marsh (2006) proposed that the Dominion Group represents a protobasinal phase of the wider Witwatersrand Basin, possibly defining the maximum age of the Witwatersrand Supergroup. The alternative hypothesis by Burke et al. (1986) is that the Witwatersrand Basin developed as a foreland basin rather than being part of a continental rift.

2.1.2. *Witwatersrand Supergroup*

The Mesoarchaeon Witwatersrand Supergroup is approximately 7 km thick, and is divided into the West Rand and Central Rand groups (Figure 2.3; c.f., McCarthy, 2006), which are well exposed in numerous outcrops in the study area (Figure 2.2). Kositcin and Krapež (2004) dated the Witwatersrand Supergroup at 2.985-2.849 Ga (U-Pb detrital zircon SHRIMP), a smaller age period than the 2.970-2.710 Ga originally suggested by Robb and Robb (1998). The West Rand and Central Rand groups are reported to be conformable (c.f., McCarthy, 2006), however Manzi et al. (2013) concluded that an angular unconformity exists between the two groups, based on 3-D seismic analysis in the West Rand and Carletonville goldfields.

The basal ~5 km thick West Rand Group conformably and/or unconformably overlies the Dominion Group, and is divided into the Hospital Hill, Government, and Jeppestown subgroups. The younger, ~2 km and auriferous, Central Rand Group conformably/unconformably overlies the West Rand Group and is divided into the Johannesburg and Turffontein subgroups (Figure 2.3.; Holland, 1990; McCarthy, 2006).

2.1.2.1. *West Rand Group*

The West Rand Group (2.985-2.902 Ga; Kositcin and Krapež, 2004; U-Pb detrital zircon SHRIMP) is composed of approximately equal proportions of quartzite and slate units (Figure 2.3; Pretorius, 1976). The quartzite units host rare meta-conglomerate reefs and are interpreted to represent continental shelf sand bodies (Eriksson et al., 1981; Winter and Brink, 1991; Beukes and Nelson, 1995). The slate units are generally iron-rich and host up to 8 magnetite-bearing iron-formations. The most prominent meta-ironstone is a banded unit within the Hospital Hill Subgroup (the Contorted Bed). The West Rand Group also contains meta-diamictite units that consist of an argillaceous matrix and clasts of quartzite and granite in the Government Subgroup (McCarthy, 2006).

The Hospital Hill Subgroup is interpreted to represent shoreline, shoreface, and subtidal facies deposited during a major transgression (Eriksson et al., 1981; Watchorn, 1981; Stanistreet and McCarthy, 1991; McCarthy, 2006). It is characterised by upward-coarsening parasequences (Stanistreet and McCarthy, 1991), but shows an overall upward-fining package indicative of a transgressive sequence (Stanistreet and McCarthy, 1991; McCarthy, 2006). The suggested drivers for this transgression are thermal subsidence as a result of extension of the earlier Dominion Basin (Clendenin et al., 1988) or the development of the Witwatersrand Basin as a foreland basin (c.f., Catuneanu, 2001).

The Government Subgroup hosts the most diverse lithologies, ranging from meta-conglomerate to quartzite, to iron-formation, and also meta-diamictite units (Bisschoff, 2000; McCarthy, 2006). These rocks indicate diverse depositional environments attributed to extreme instability of the craton (McCarthy, 2006). The meta-diamictite units contain striated clasts and are unconformable to their footwall at some locations (Camden-Smith, 1980). Tankard et al. (1982) suggested that the meta-diamictite units represent early glacial activity based on their unconformable nature.

The Government Subgroup represents rapid transgression and regression cycles. Transgression cycles are represented by fine-grained sedimentary rocks (e.g., Coronation and Palmietfontein formations), which are directly overlain by rapid regression deposits characterised by a disconformity (e.g., the base of the Tusschenin Formation; Figure 2.3). The disconformity is overlain by conglomeratic reefs (e.g., Coronation and Government reefs) and quartzite units, which were deposited in a braid-plain environment (McCarthy, 2006).

The Jeppestown Subgroup generally represents marine shelf sands at the base, which coarsen upwards into Maraisburg Formation that represents progradation of a fluvial braid-plain (Beukes and Nelson, 1995; McCarthy, 2006). The base of the Jeppestown Subgroup is a disconformity marked by the Koedoeslaagte Formation (Winter and Brink, 1991), which is overlain by upward-fining quartzite units of the Babrasco Formation, indicative of a transgressive sequence (McCarthy, 2006). The Babrasco Formation is overlain by a series of basaltic andesite flows of the Crown Formation that are up to 250 m thick (Myers, 1991; McCarthy, 2006).

The West Rand Group represents an upward-coarsening sequence initiated by marine transgression across the Kaapvaal Craton, which is well represented by the lower Hospital Hill Subgroup. The West Rand Group coarsens upwards to fluvial braid-plain deposits (Maraiburg Formation), setting up a platform for the high energy environments that characterise the Central Rand Group (McCarthy, 2006).

2.1.2.2. Central Rand Group

The 2.902-2.849 Ga (Kositcin and Krapež, 2004; U-Pb detrital zircon SHRIMP) metasedimentary units of the Central Rand Group are well exposed in the northern part of the study area (Figures 1.2, 2.2). The Central Rand Group has a quartzite/slate ratio of approximately 12.6 whereas the ratio is 1.0 in the West Rand Group (Figure 2.3; Pretorius, 1976). It conformably and/or unconformably overlies the West Rand Group, and is dominated by quartzite interbedded with slate and gold-bearing meta-conglomerate reefs (c.f., McCarthy, 2006). The group is divided into the Johannesburg and Turffontein subgroups by the Booysens Formation (Mellor, 1917). The Central Rand Group also hosts meta-diamictite units associated with the Main and Kimberley formations. The meta-diamictite units are interpreted to have formed through a debris-flow mechanism associated with erosional channels in the immediate footwall (Martin et al., 1989; Beukes and Nelson, 1995).

The Johannesburg Subgroup is dominated by aggradational quartzite units deposited in a fluvial braid-plain environment, and several degradational cycles that were followed by either transgression or progradation of braid plains (McCarthy, 2006). Buck and Minter (1985) and Els (1991) suggested that the degradational cycles deposited placer gold as a result of episodic erosion resulting from a lowered base level. The erosion caused braid-plain peneplanation, resulting in widespread disconformities (McCarthy, 2006). Subsequent marine transgressions onto peneplaned surfaces resulted in deposition of heavy mineral concentrates, and placer reworking and upgrading of the heavy minerals (Brink, 1986; Verrezen, 1987). A basin-wide transgression occurred towards the end of deposition of the Krugersdorp Formation, during which slate units of the Booysens Formation were deposited as the top of the Johannesburg Subgroup (McCarthy, 2006).

The final phase of deposition in the Witwatersrand Basin occurred in alluvial braid-plains (Els, 1998) and less dominant alluvial fan environments (Kingsley, 1984, 1987). This is represented in the Turffontein Subgroup by generally upward-coarsening units that indicate prograding braid-plains that Loen (1992) and McCarthy (2006) attributed to uplift by tectonic processes. Erosional reworking of alluvial gravels by marine and fluvial processes deposited economic placer gold-bearing horizons (e.g., Kimberley Reefs and the Upper Elsburg Reefs; McCarthy, 2006). The sedimentation culminated in prograding alluvial fan deposits of the Mondeor Formation (Kingsley, 1987).

2.1.3. Ventersdorp Supergroup

The Witwatersrand Supergroup is both conformably and unconformably overlain by the Ventersdorp Contact Reef (VCR), which is in turn overlain by the 2.72-2.63 Ga (Kositcin and Krapež, 2004) Ventersdorp Supergroup (Figure 2.1; Jolley et al., 2004; McCarthy, 2006; Dankert and Hein, 2010; Manzi et al., 2013). The VCR belongs to the Venterspost Formation, which has a maximum detrital age of 2.729 ± 0.019 Ga (Kositcin and Krapež, 2004). The Venterspost Formation consists of <1.5 m thick immature, auriferous meta-conglomerate units (Manzi et al., 2013) that host subangular to rounded pebbles (McCarthy, 2006). The conglomeratic VCR represents an erosional surface (Tankard et al., 1982; Frimmel et al., 1999; Gibson, 2004, 2005; McCarthy, 2006). The contrast in acoustic impedance of the VCR and overlying basalts creates a laterally definable reflector horizon for 2-D and 3-D seismic methods in the goldfields and around the Kaapvaal Craton (c.f., Tinker et al., 2002 and Manzi et al., 2013).

The Ventersdorp Supergroup is composed primarily of ultramafic and mafic volcanic rocks of the Klipriviersberg Group, as well as metasedimentary and bimodal volcanic rocks of the Platberg Group and the younger Pniel Group that are collectively ~8 km thick (Van der Westhuizen et al., 1991; Crow and Condie, 1998). The Klipriviersberg Group consists of tholeiitic flood basalts up to 5.1 km thick, as well as feeder dykes and sills (Winter 1976; Van der Westhuizen et al., 1991). The group is divided into the Westonaria, Alberton, Orkney, Jeannette, and Edenville formations (McCarthy, 2006). Armstrong et al. (1991) dated amygdaloidal basalt of the Alberton Formation to 2.714 ± 0.008 Ma (single zircon U-Pb).

Nelson et al. (1999), Eriksson et al. (2002), and Condie (2004) suggested that the emplacement of the Klipriviersberg Group may have occurred at the same time as mantle overturning related to development of a superplume beneath the Kaapvaal Craton, which also formed massive granite-greenstone belts in the Yilgarn and Superior cratons. This is evidenced in the presence of ultra-high-temperature xenoliths of the lower crust in the Lace-Voorspoed kimberlites and the Kimberley Block (Schmitz and Bowring, 2003). The Klipriviersberg Group is overlain by the Platberg and Pniel groups (up to 2.9 km thick; Winter, 1976) that do not crop out in the study area; therefore they are not discussed further.

2.1.4. Transvaal Supergroup

The 2.588-2.100 Ga Transvaal Supergroup (Figure 2.1: Armstrong et al., 1991; Eriksson et al., 2005) is preserved in three structural basins on the Kaapvaal Craton; the Transvaal, Griqualand West, and Kanye basins (Eriksson et al., 2006). The unconformity between the Ventersdorp Supergroup and the Transvaal Supergroup is marked by the Black Reef Formation. The age of the Transvaal Supergroup unconformity is taken as 2.588 Ga, which is the published age of tuff units in the Oaktree Formation (single zircon U-Pb SHRIMP; Martin et al., 1998) considered as the proxy for the base of the Transvaal Supergroup. The Black Reef Formation is composed of an older quartzite unit, with

subordinate meta-conglomerate and slate (Henry et al., 1990; Eriksson et al., 2006; Manzi et al., 2013). It represents a significant reflective interface in 3D seismic surveys owing to the extensive acoustic impedance contrast between the overlying high-velocity, high-density Chuniespoort Group (Transvaal Supergroup) dolomites and the low-velocity and less dense Ventersdorp Supergroup meta-basalts (Manzi et al., 2013). This unconformity is also an important reflector horizon in seismic sections produced by Tinker et al. (2002) and Manzi et al. (2013), especially in relation to Neoarchaeal structures interpreted from these sections.

The Black Reef Formation is overlain, successively, by the Chuniespoort and Pretoria groups. The older and approximately 2 km thick Chuniespoort Group (2.58 – 2.47 Ga; Walraven et al., 1990; Martin et al., 1998) comprises dolomite, banded iron formation, and lacustrine deposits (Eriksson et al., 1995, 2006; Frimmel, 2005). The 6-7 km thick Pretoria Group (2.35 – 2.25 Ga; c.f., Hannah et al., 2004) is separated from the Chuniespoort Group by a non-conformity representing a hiatus of about 80 million years (Eriksson and Reczko, 1995; Eriksson et al., 1995). The Pretoria Group is dominated by intercalated mudstone-sandstone units, andesite, and subordinate meta-conglomerate, meta-diamictite, and carbonate units (Eriksson et al., 2006).

2.1.5. Bushveld Complex

The 2.059 – 2.061 Ga Bushveld Complex (U-Pb titanite age from a vesuvianite-, wollastonite-, and garnet-rich xenolith; Buick et al., 2001) is the world's largest layered intrusion (Cawthorn et al., 2006). It crops out along the main Western, Northern, and Eastern limbs, as well as in smaller outcrops of the Far Western Limb and as sub-outcrop of the Southeastern Limb (Figure 2.4). The geology is discussed by, amongst others, Button (1976), Buchanan (1977), White (1994), Hatton and Schweitzer (1995), Eales and Cawthorn (1996), and Cawthorn et al. (2006).

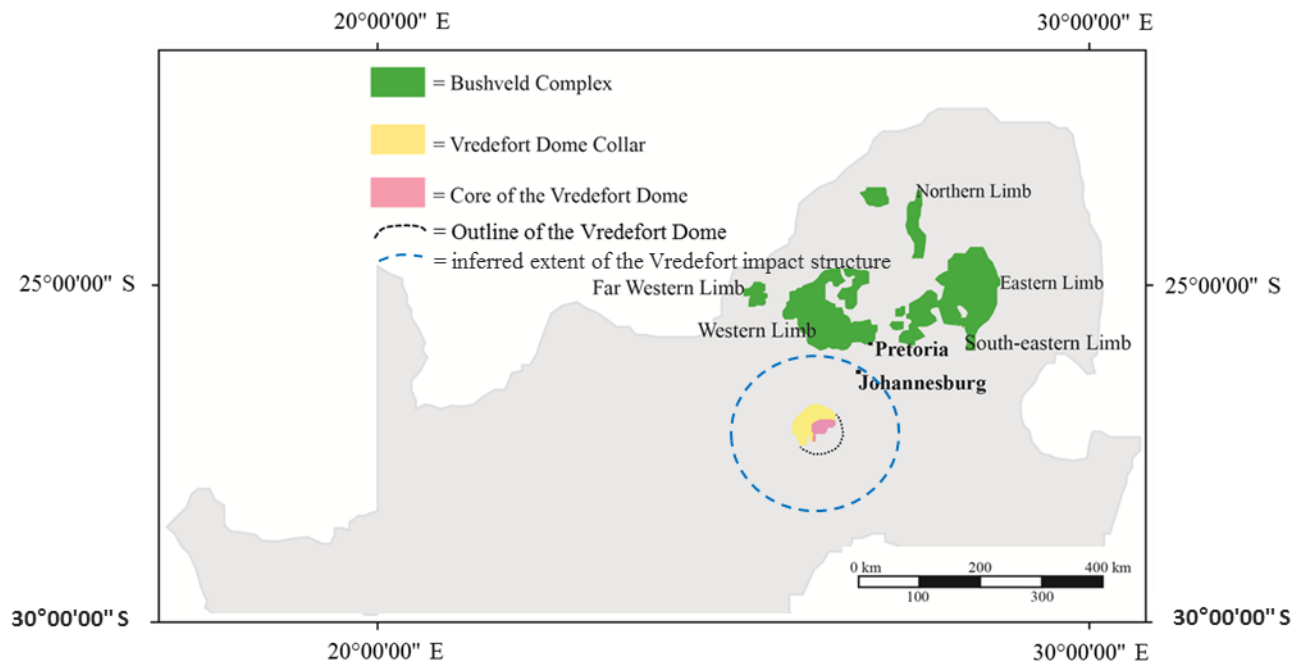


Figure. 2.4. Map of the northern part of South Africa showing the spatial relationship between outcrops of the Bushveld Complex and the Vredefort Dome. The inferred extent of the now eroded impact structure is also included (Therriault et al., 1997).

The heat released into the country rocks during cooling formed a contact metamorphic aureole, particularly in the Pretoria Group (Walraven, 1985; Cawthorn et al., 2006). Metamorphic mineral assemblages are hosted along a lateral distance of up to 50 km. These include porphyroblasts of andalusite, garnet, staurolite, cordierite, biotite, and chloritoid in metapelitic lithologies that also show evidence of syn-metamorphic deformation recorded in some parts of the aureole (Sharpe and Chadwick, 1982; Uken and Watkeys, 1997). Uken and Watkeys (1997) and Cawthorn et al. (2006) proposed that deformation associated with the metamorphism was driven by downsagging of the Bushveld Complex, as a result of its large size and density. Regional cleavages, folds, and mylonitic shears described by McCarthy et al. (1986) and Gibson et al. (1999) in the Transvaal Supergroup (north of the Vredefort and Johannesburg domes) are attributed to this deformation; evidenced by the observation that the fabrics and folds postdate contact metamorphism associated with the Bushveld Complex but predate the Vredefort impact (Gibson et al., 1999).

Magmatism associated with emplacement of the Bushveld Complex is also suggested to have caused regional greenschist facies metamorphism of the Witwatersrand Supergroup along the margins of the Witwatersrand Basin, and amphibolite to greenschist facies metamorphism of the equivalent units in the Vredefort Dome (Phillips and Law, 1994; Gibson and Stevens, 1995; Gibson and Wallmach, 1995).

2.2. Structural geology of the Witwatersrand Basin

Studies of the structural geology of the Witwatersrand Basin have been conducted in the goldfields, mostly focused on the relationship between structures and gold mineralisation, or fault displacement of gold-bearing reefs (c.f., Coward et al., 1995; Jolley et al., 2004; Beach and Smith, 2007; Dankert and Hein, 2010; Manzi et al., 2013). These studies have used regional 2D seismic lines and 3D seismic surveys, underground mapping, and core logging (c.f., Jolley et al., 2004, 2007; Beach and Smith, 2007; Dankert and Hein, 2010; Mambane et al., 2011; Mashabela, 2011; Manzi et al., 2013). There are generally five deformation events recognised from the beginning of deposition of the Central Rand Group and the Vredefort impact at 2.02 Ga (Figure 2.5; c.f., Bickle and Eriksson, 1982; Burke et al., 1985, 1986; Killick, 1992; Jolley et al., 2004; Frimmel, 2005; Beach and Smith, 2007; Dankert and Hein, 2010; Manzi et al., 2013).

2.2.1. Regional D1 deformation at 2.90 Ga

Manzi et al. (2013) indicated that an unconformity exists between the West Rand and Central Rand groups, which they attributed to a ca. 2.90 Ga deformation, termed the Asazi Event. The deformation was associated with uplift and tilting of the West Rand Group, which resulted in its erosion. The unconformity is evidenced in 3-D seismic profiles from the Carletonville and West Rand goldfields (Manzi et al., 2013). The cause of uplift and tilting is not known, however the timing is consistent with descriptions by McCarthy et al. (1982), Roering et al. (1991), Barnicoat et al. (1997), and Brink et al. (2000) of early compressional tectonics in the Witwatersrand Supergroup. They associated the event with eastward vergence (Brink et al., 2000) and open fold-thrust formation at the end of deposition of the West Rand Group, and possibly during early deposition of the Central Rand Group at 2.90 Ga.

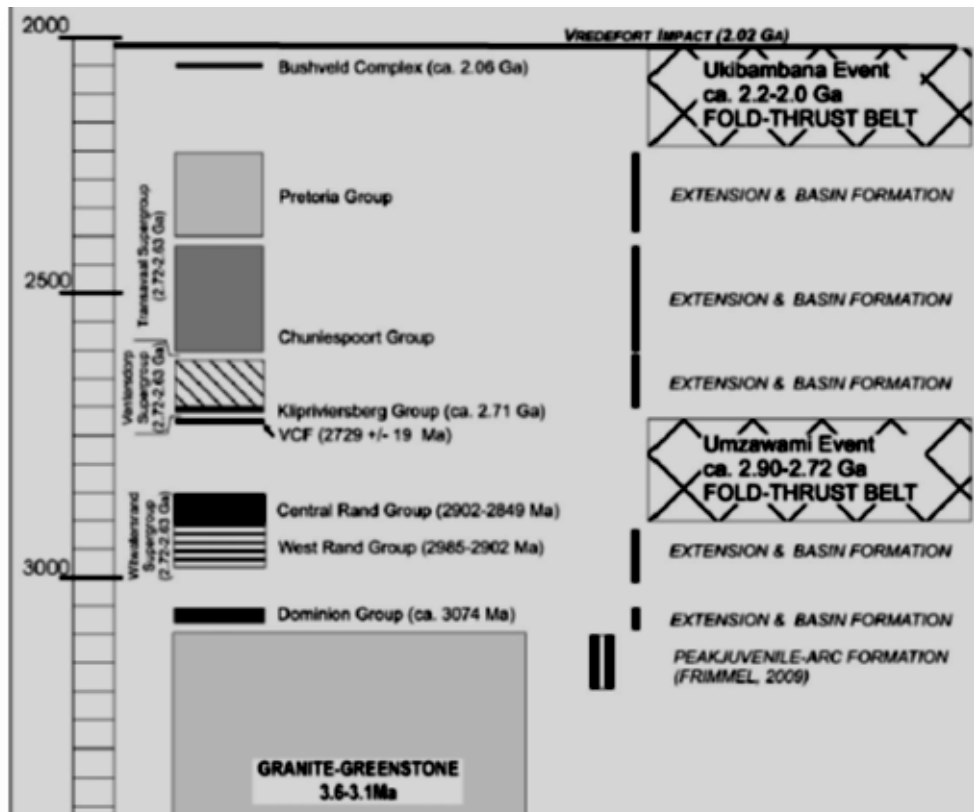


Figure 2.5. Sketch showing deformations reported in the Witwatersrand goldfields, as reviewed by Dankert and Hein (2010).

2.2.2. Regional D2 deformation at 2.73 Ga

The D2 regional deformation at approximately 2.73 Ga formed thrusts and folds in the Witwatersrand Supergroup (Beach and Smith, 2007; Jolley et al., 2007; Manzi et al., 2013). Thrust duplexes are interpreted to have formed in the West Rand Group and broad-scale folds in Witwatersrand Supergroup strata located in the West Rand, Carletonville, Klerksdorp, and Welkom goldfields (c.f., Stanistreet et al., 1986; De Wit and Tinker, 2004; Schmitz et al., 2004; Frimmel, 2005; Beach and Smith, 2007). Dankert and Hein (2010) termed this the Umzawami Event.

Amongst the folds attributed to this compressional event, Manzi et al. (2013) described the northerly-trending, upright, and tight-to-open Libanon Anticline in the West Rand goldfield (Figure 2.7). From 3-D seismic reflection data, they indicated that the anticline has a wavelength of approximately 8 km and amplitude of 2 km, and is associated with second-order parasitic folds (e.g., the Panvlakte anticline-syncline pair) as well as reverse faults such as the Tandeka and Jabulani thrusts (Figure 2.6). These structures, along with those described by Barnicoat et al. (1997), Beach and Smith (2007), and Jolley et al. (2007), were formed during a single compressive tectonic event that led to the development of an easterly-verging fold-thrust belt (Figure 2.7).

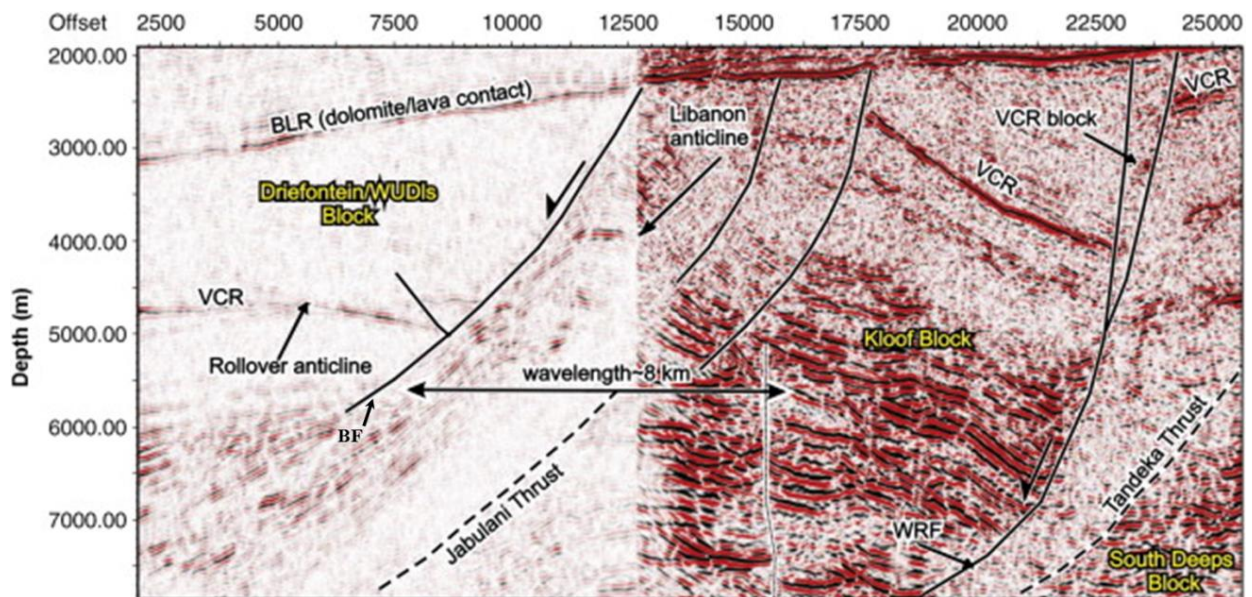


Figure 2.6. Seismic section of the Witwatersrand, Ventersdorp, and Transvaal supergroups in the West Rand and Carletonville goldfields (Manzi et al., 2013). VCR: Ventersdorp Contact Reef, BLR: Black Reef Formation, BF: Bank Fault, WRF: West Rand Fault.

2.2.3. Craton-wide D3 extensional tectonics at 2.70-2.64 Ga

Bickle and Eriksson (1982), Burke et al. (1985), Stanistreet et al., (1986), Brink et al. (2000), Frimmel and Minter (2002), Tinker et al. (2002) and other authors described listric faults that displace the Witwatersrand and Ventersdorp supergroups, but that are truncated by the base of the Transvaal Supergroup (the Black Reef Formation). Brink et al. (2000), Jolley et al. (2004, 2007), and Beach and Smith (2007) proposed that the normal faults were formed by extensional collapse initiated during deposition of the Ventersdorp Supergroup (Figure 2.7). Manzi et al. (2013) termed the extensional deformation the Hlukana-Platberg Event, and proposed that it formed tear/scissor faults (e.g., West Rand and Bank faults, Figure 2.6). Clendenin et al. (1988), Crow and Condie (1988), Eriksson et al. (2001, 2009) associated extension to lithospheric thinning, attributed to mantle plume heating. Extension formed horsts and half-grabens across the Kaapvaal Craton (c.f., Stanistreet et al., 1986; Tinker et al., 2002; Beach and Smith, 2007; De Kock et al., 2012). This event correlates with the extrusion of basaltic sequences on the Pilbara Craton (Arndt et al., 1991; Nelson et al., 1999; Blake, 2001; Thorne and Trendall, 2001; Blake et al., 2004), indicating a wider event.

2.2.4. Regional D4 deformation at 2.20-2.02 Ga

A fold-thrust belt was developed in the Central Kaapvaal Craton subsequent to deposition of the upper Pretoria Group (Andreoli, 1988; Bumby et al., 1998; Killick and Scheepers, 2005; Alexandre et al., 2006). Andreoli (1988) and Dankert and Hein (2010) termed this compressive deformation the Transvaalide and Ukubambana events respectively. The deformation formed regional cleavages, folds, and mylonitic shears described by McCarthy et al. (1986), Gibson et al. (1999), and Alexandre et al (2006) in the Transvaal Supergroup. Gibson et al. (1999) showed that the cleavage postdated contact metamorphic assemblages of the Bushveld Complex but predated the Vredefort impact at 2.02 Ga. The deformation was characterised mainly by north- to NNE vergence (Gibson et al., 1999; Alexandre et al., 2006), however Alexandre et al. (2006) reported other different vergence directions attributed to the polyphase nature of the deformation.(e.g., SW-directed vergence of the D2 phase).

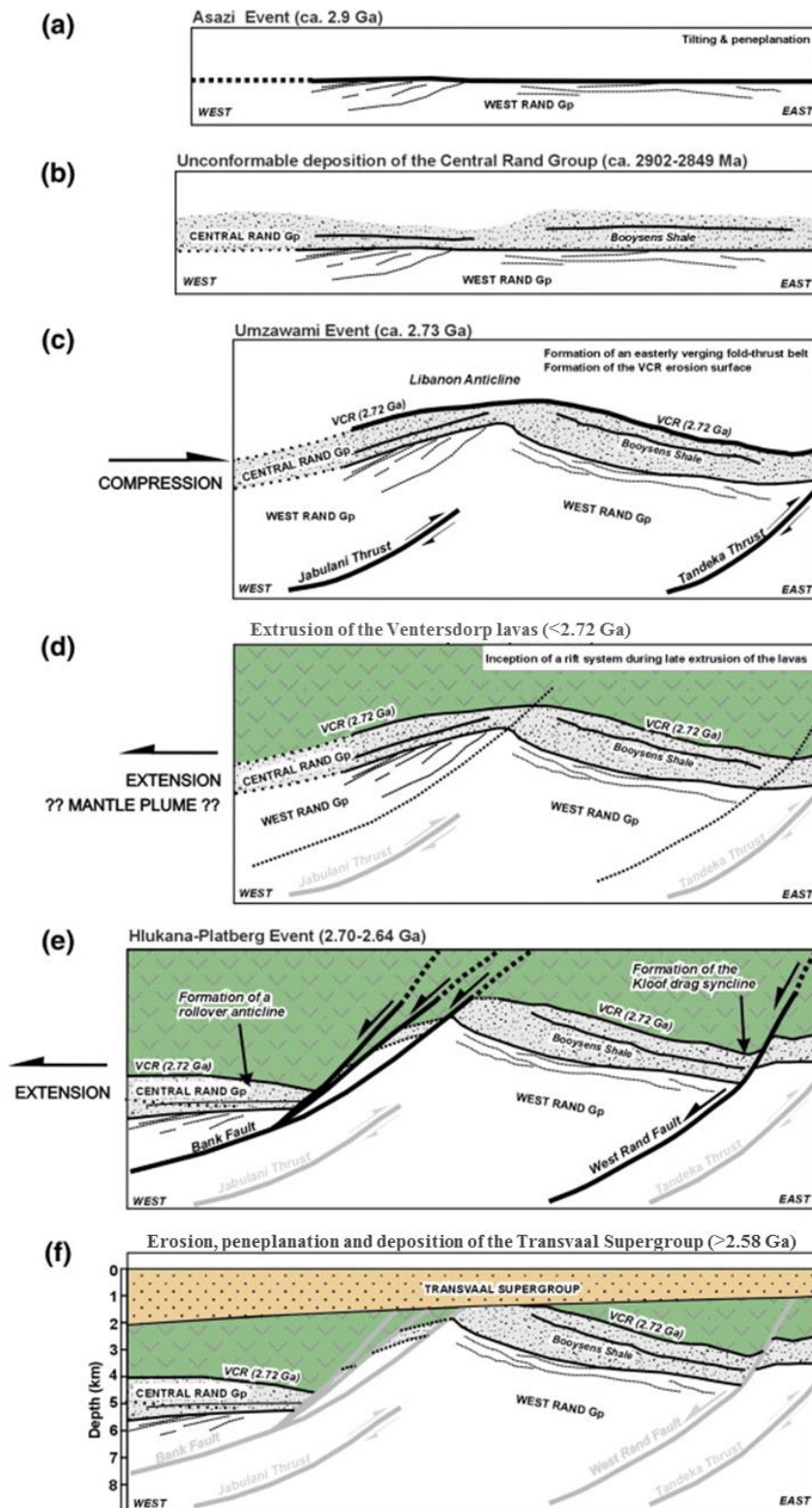


Figure 2.7. A schematic diagram of the Neoarchaean history of the West Wits Line and West Rand goldfields, as proposed by Manzi et al. (2013). (a) Exposure of the West Rand Group during uplift, with tilting syn- to post-erosion; (b) unconformable deposition of the Central Rand Group; (c) compressive deformation during the Umzawami Event; (d) Initiation of extensional tectonics during extrusion of Ventersdorp lavas; (e) Extensional tectonics during the Hlukana-Platberg Event; and (f) Deposition of the Transvaal Supergroup.

2.2.5. D5 Vredefort impact at ca. 2.02 Ma

The meteorite impact occurred at 2.02 ± 0.004 Ga (U-Pb zircon; Kamo et al., 1996; Gibson and Reimold, 2008), affecting the entire Mesoproterozoic to Palaeoproterozoic sequence on a regional scale, with the 90 km-wide central uplift surrounded by a rim synclinorium that occupies the bulk of the present Witwatersrand Basin. The erosion of the resultant impact crater exposed the Vredefort Dome and the Potchefstroom Synclinorium as deep structural remnants of the crater, but one of the enduring unknowns is the radial extent of impact-induced structures (Simpson, 1977; Spray et al., 1995; Kamo et al., 1996; Gibson et al., 1997a; Moser, 1997). The structural detail of the Vredefort Dome is described in Section 2.4.

2.3. Metamorphism in the Witwatersrand Basin

Metamorphic assemblages hosted in the Witwatersrand Supergroup have been established from studies of: (a) extensive pelitic horizons (e.g., Parktown and Coronation slate formations; Borrego and Phillips, 1987; Palmer et al., 1987, 1989; Phillips and Law, 1994); (b) narrower pelitic horizons (e.g., the Black Reef below the Main Leader Reef; Phillips, 1987); (c) quartzite units (Fuller, 1958; Law, 1991; Sutton et al., 1990); (d) conglomeratic units (Young, 1917; Hallbauer, 1986); (e) intrusive and extrusive mafic rocks (e.g., the Crown and Bird meta-basalt units; Pegg, 1950; Palmer, 1986); (f) kerogen (Law et al., 1991); and (g) lithological horizons proximal to fault planes (e.g., the Master Bedding Fault; Palmer, 1986; Phillips, 1988). These analyses indicated that the Witwatersrand Supergroup has been subjected to widespread greenschist facies metamorphism, which is evident throughout the goldfields (Bisschoff, 1982, 1988; Phillips and Law, 1994 and references therein). Phillips and Law (1994) proposed that most of the goldfields host metamorphic mineral assemblages composed of pyrophyllite, chloritoid, muscovite, chlorite, quartz, tourmaline, rutile, and pyrite, with local assemblages of kaolinite, paragonite, biotite, pyrrhotite, and arsenopyrite (Figure 2.8). However, distinct areas along the northwestern margin of the basin, northeast of the Evander goldfield, and the collar of the Vredefort dome host higher grade mineral assemblages of andalusite, kyanite, garnet and/or cordierite (Bisschoff, 1982; Phillips and Law, 1994; Gibson and Wallmach, 1995; Foya, 2002). These mineral assemblages are crosscut by impact shock features in the Vredefort Dome, indicating a pre-2.02 Ga origin (c.f., Bisschoff, 1982, 1988; Gibson and Wallmach, 1995).

The Ventersdorp Supergroup hosts chlorite, epidote, and muscovite in Klipriviersberg metabasalts, interpreted as alteration products of feldspar, quartz, rutile and pyrite (Phillips and Law, 1994). A metamorphic assemblage composed of muscovite, quartz, amphibole, rutile, and pyrite is associated with foliation along the VCR unconformity (Palmer, 1986).

A metamorphosed shale unit of the Black Reef Formation hosts muscovite, chlorite, quartz, carbonate, tourmaline, and albite. Metapelitic units of the Black Reef Formation host higher grade metamorphic assemblages of andalusite, anthophyllite, biotite and/or cordierite in the northeastern

part of the Evander Goldfield (Figure 2.8; Phillips and Law, 1994). Gibson et al. (1999) also described retrogressed andalusite in metapelite units of the Chuniespoort Group north of Johannesburg.

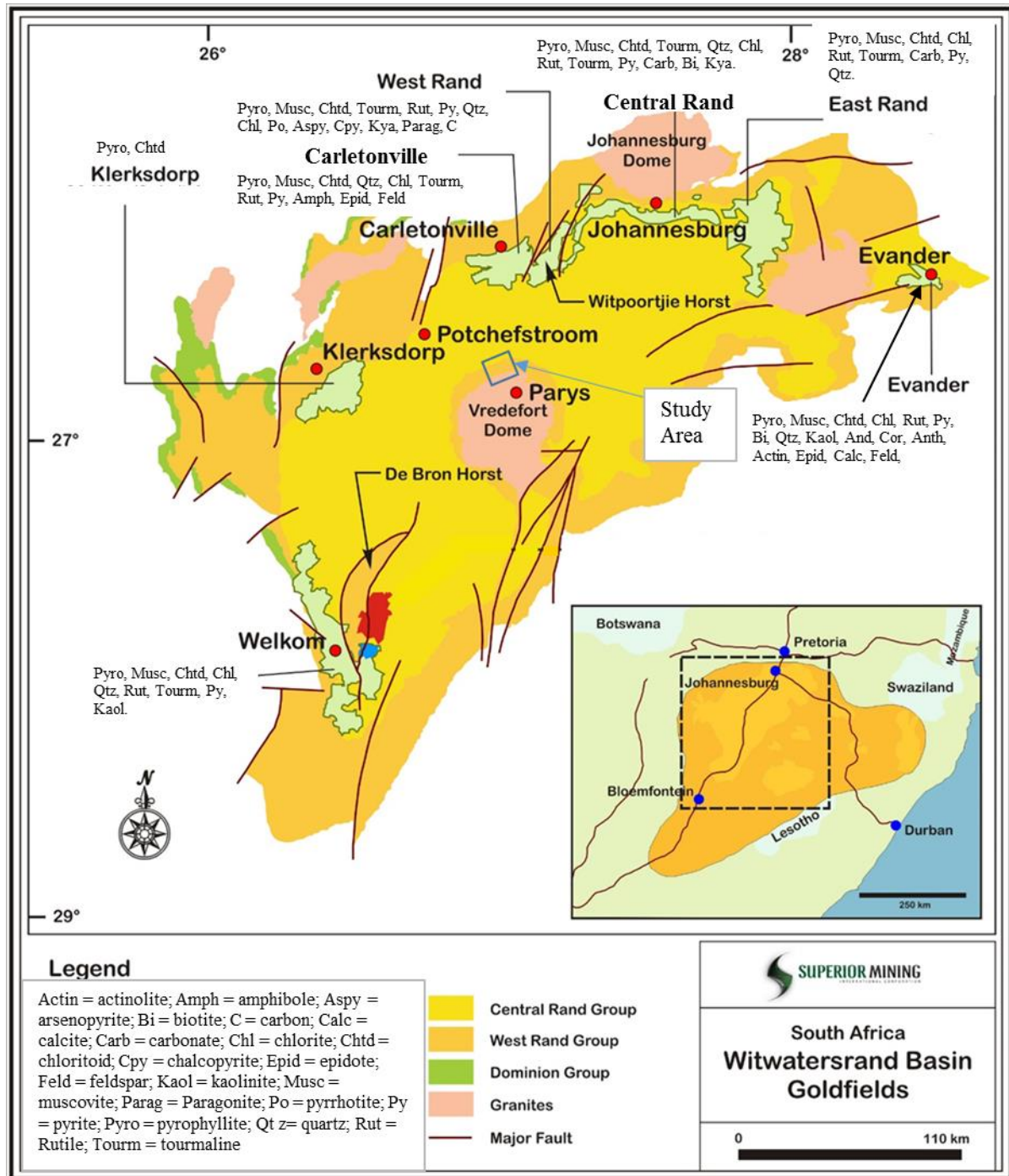


Figure 2.8. Distribution of metamorphic mineral assemblages across the Witwatersrand goldfields (Phillips and Law, 1994) (Map published with permission from Superior Mining International Corporation).

Pressure and temperature conditions of metamorphism across the Witwatersrand Basin were estimated by Wallmach and Meyer (1990) and Phillips and Law (1994) using the pyrophyllite-chloritoid-quartz assemblage. Interpretations of metamorphic textures indicate that peak metamorphic conditions of 300-360°C and 2-3 kbar were reached during metamorphism in most parts of the basin (c.f., Phillips and Law, 1994; Foya, 2002), signifying a geothermal gradient of 40-50°C/km (Bisschoff, 1982; Phillips and Law, 1994; Gibson and Wallmach, 1995). Bisschoff (1982) and Gibson and Wallmach (1995) described mid-amphibolite facies metamorphism in rocks of the lower Witwatersrand Supergroup in the Vredefort Dome, which they attributed to deeper levels of burial of the lower Witwatersrand Supergroup rocks relative to the goldfields during the same metamorphic event.

Phillips and Law (1994), Gibson and Stevens (1995), and Gibson and Wallmach (1995) attributed this metamorphism to underplating and interplating of mafic magma during Bushveld magmatism (2058.9 ± 0.8 Ma U-Pb age of titanite; Buick et al., 2001). This regional metamorphism was accompanied by deformation (e.g., Gibson, 1993); large aqueous fluid fluxes, element mobility, and in some parts carbon and gold remobilization (Phillips and Law, 1994; Stevens et al., 1997).

2.4. The Vredefort Dome

The Vredefort Dome is an 80-90 km wide feature centred 120 km southwest of Johannesburg, South Africa (Figure 2.9; Lana et al., 2003a; Wieland et al., 2005; Grieve et al., 2008). It is composed of a 40 km wide core bounded by a 20-25 km wide collar, which is in turn surrounded by a 50-70 km wide rim synclinorium (c.f., Gibson and Reimold, 2008). These are structural remnants of an eroded ancient crater estimated to have been 250-300 km wide (Therriault et al., 1997; Henkel and Reimold, 1998). The estimation of the size of the Vredefort impact crater was partly based on the occurrence of pseudotachylitic breccias in the Witwatersrand goldfields (Fletcher and Reimold, 1989; Killick and Reimold, 1990), which have been reported by Trieloff et al. (1994) as having similar ages to those hosted in the Vredefort Dome. This makes it one of the three largest, multi-ring, impact craters known on Earth, along with the approximately 250 km wide Sudbury crater (1.85 Ga) and the 180-200 km Chicxulub (0.66 Ga) crater (Spudis, 1993; Grieve and Therriault, 2000).

A comprehensive review of the Vredefort Dome was provided by Gibson and Reimold (2008). Studies of the dome have included: (1) melt-rock research (c.f., Shand, 1916; Hall and Molengraaff, 1925; Nel, 1927; Wilshire 1971; Dressler and Reimold, 2004; Reimold and Gibson, 2006); (2) research focused on impact-related features such as shatter cones, planar deformation features (PDF), coesite and stishovite, shocked zircons and other minerals (c.f., Dietz, 1960, 1961; Hargraves, 1961; Carter, 1968; Martini, 1978; Leroux et al., 1994; Kamo et al., 1996; Gibson et al., 1997b, 1998; Reimold, 1998; Reimold and Gibson, 2005, 2006); (3) metamorphic analyses (c.f., Bisschoff, 1982;

Gibson and Wallmach, 1995; Gibson and Stevens, 1998; Foya, 2002); and (4) structural analyses (c.f., Simpson, 1977; Lilly, 1978; Colliston and Reimold, 1990; Corner et al., 1990; Martini, 1992; Lana et al., 2003a, 2003b; Wieland et al., 2005). Metamorphic studies by Gibson and Wallmach (1995), Foya (2002), Lana et al. (2003a, 2003b), and Ogilvie (2010) showed that the Vredefort Dome is a polymetamorphic terrain, exhibiting a maximum granulite facies in the core and mid-amphibolite to lower greenschist facies in the collar. These assemblages were attributed to metamorphic events at 3.10 Ga, 2.06 Ga, and 2.02 Ga, discussed in Section 2.4.4. Structural analyses show that the Vredefort Dome is structurally complex (c.f., Simpson, 1977; Lilly, 1978; Wieland et al., 2005) and most of its large-scale complexity has been attributed to the impact event and, more specifically, to the formation of the central uplift beneath the crater, now exposed as the dome.

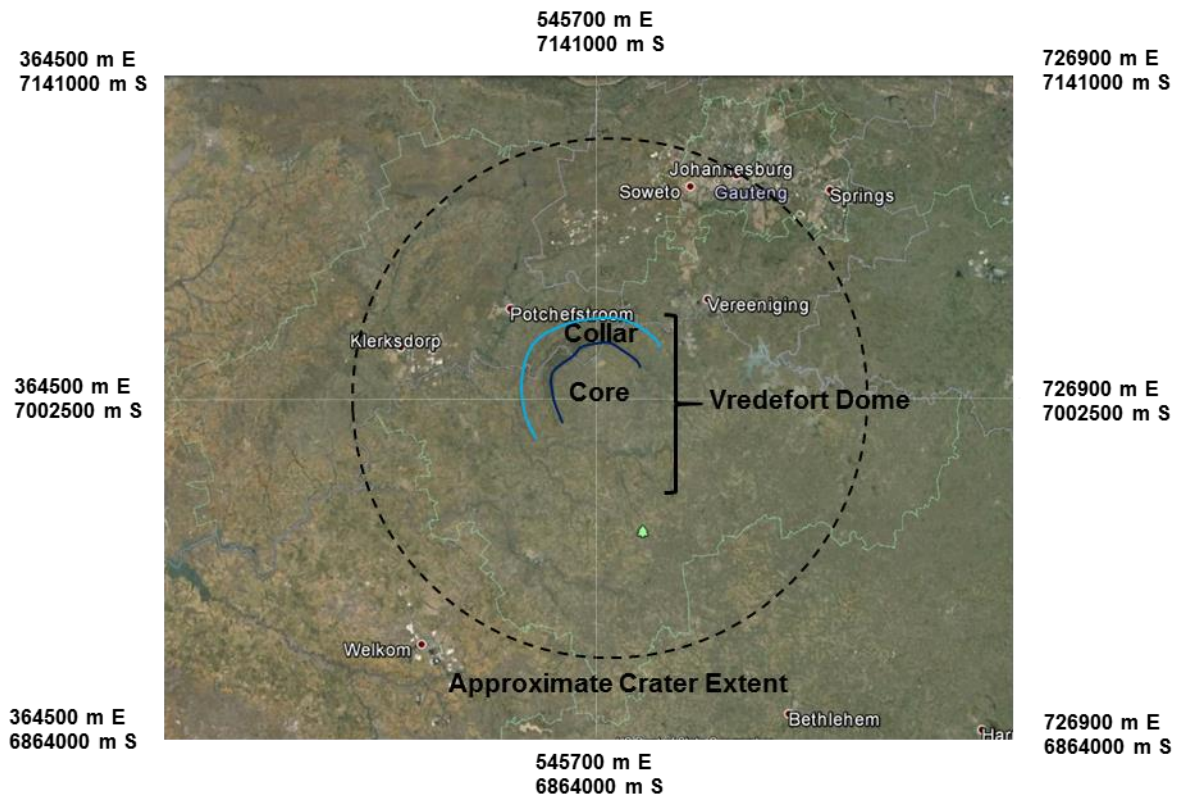


Figure 2.9. Google EarthTM image showing the estimated extent of the Vredefort impact crater by Therriault et al. (1997) and Henkel and Reimold (1998), showing the uplifted region (Vredefort Dome) in the centre.

2.4.1. Impact craters formation – A brief outline of principles

Experimental production of small craters (e.g., Gault, 1973), geological studies of larger impact structures on Earth (e.g., Shoemaker, 1963; Dence et al., 1977), and a combination of theoretical and experimental studies of shock waves (e.g., Melosh, 1989) have contributed to the knowledge of impact processes. French (1998) showed that meteorite impact craters form through complex processes that can be divided into three overlapping stages (Figure 2.10).

The initial, contact-compression stage begins when the projectile contacts the target rocks, resulting in the transfer of kinetic energy as a shock wave that is generated at the projectile-target interface (Kieffer and Simonds, 1980; O’Keefe and Ahrens, 1982, 1993; Melosh, 1989). The reflected wave reaches the back of the projectile and is reflected forward as a release wave (Melosh, 1989), which unloads the projectile from high shock pressures and temperatures, resulting in complete melting and vaporisation of the projectile. The duration of this stage is controlled by the behaviour of the shock wave that is reflected back into the projectile from the interface. It generally lasts for only a few seconds, even in the largest impacts (French, 1998).

The excavation stage results in exhumation of the target rocks by interactions between the shock wave and the original ground surface, causing an outward acceleration of target rocks (Figure 2.10; Melosh, 1989; Grieve, 1991). This stage produces a symmetric excavation flow around the centre of the developing crater, forming a bowl-shaped depression referred to as the transient crater (French, 1998). This stage lasts for a few minutes for large impacts (about 90 seconds for a 200 km-diameter crater, Melosh, 1989). The shock wave eventually decreases in intensity away from the point of impact, giving way to the modification stage (French, 1998).

The modification stage alters the transient crater through gravity and rock mechanical processes, causing collapse of the transient crater walls as well as uplift and subsequent outward collapse of the central part (French 1998; Wieland et al., 2005). Rocks beneath the centre of the transient crater rise to form a central uplift (Figure 2.10), changing the shape of the transient crater floor (Grieve et al., 1981). The immediate part of the modification stage lasts for a few minutes in large impacts (Melosh, 1989). The extent of transient crater modification depends on the size of the crater and the properties and structure of target rocks. Depending on the extent of modification, 3 impact structures can be formed, namely: (a) simple bowl-shaped craters that are a few kilometres across, and have no centrally uplifted region; (b) complex craters (diameter >4 km) characterised by a centrally uplifted region that could be a central peak or peak ring; and (c) multi-ring basins that are typically a few hundred kilometres or more in diameter, characterised by multiple concentric uplifted rings (French, 1998). These estimated crater diameters are for terrestrial impacts only; they differ from planet to planet owing to different planetary gravities (Melosh, 1989).

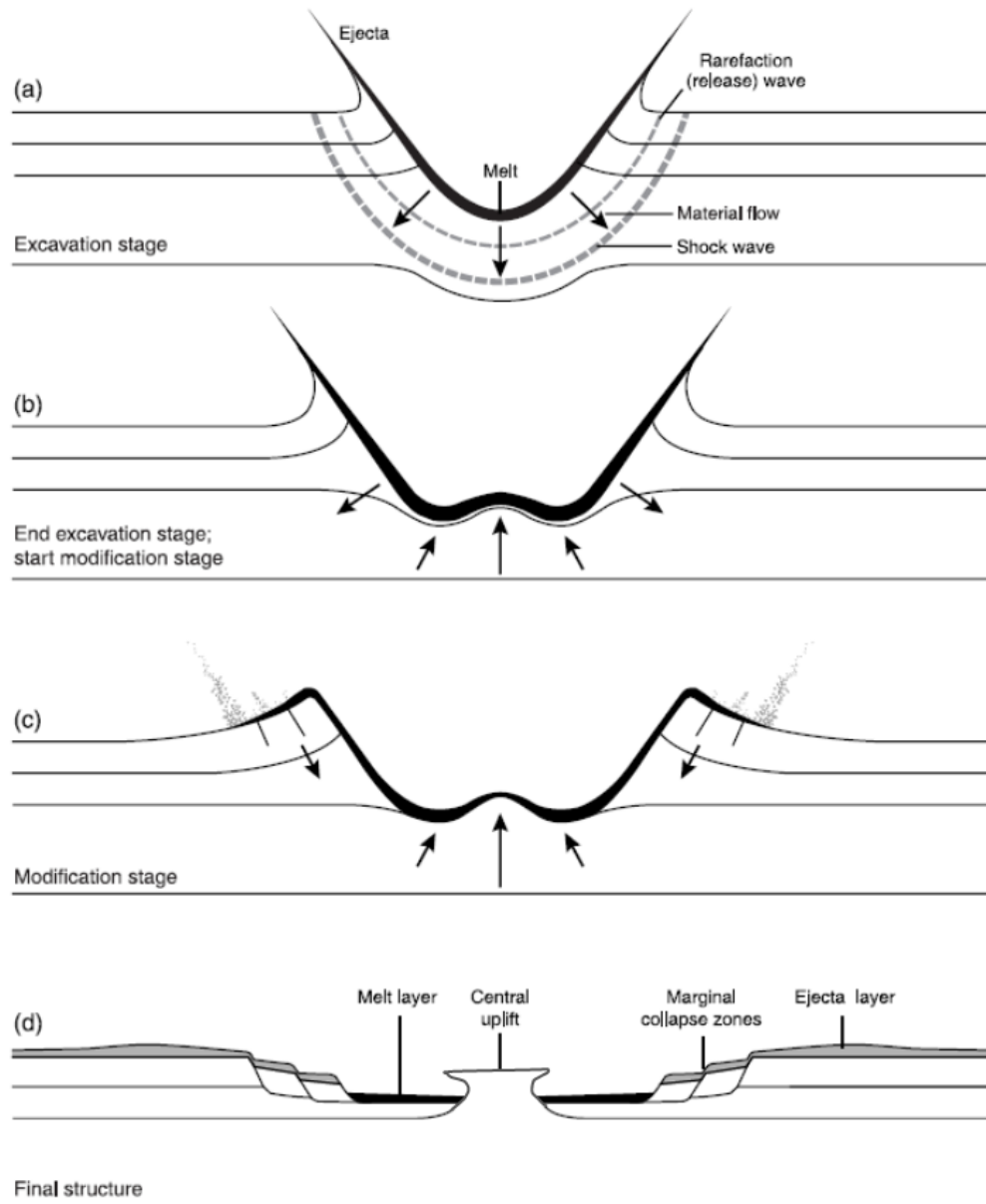


Figure 2.10. A sketch showing the development stages for a complex impact crater: (a) Transient crater formation by the excavation stage, similar to that of simple craters; (b) initial development of central uplift during the modification stage; (c) initiation of peripheral collapse, accompanied by thinning of the original melt layer with continued uplift in the centre; (d) the final structure (central peak complex crater) with exposed deeper rocks in the central uplift, surrounded by a flat plain and a rim formed by inward displacement along normal faults (French, 1998).

2.4.1.1. The Vredefort impact structure

The Vredefort crater was formed by an approximately 10 km wide projectile (French and Nielsen, 1990). Through the modification stage, a central uplifted region 80-90 km in diameter was formed, the eroded remnant of which is now referred to as the Vredefort Dome. The main control on whether a crater exhibits a central peak vs. peak ring vs. multi-ring is the diameter of the transient crater, which reflects how the transient crater is modified (Grieve et al., 1981; Melosh, 1989; Spudis, 1993). The transitional diameters of simple to central peak craters and central peak to peak ring craters are 2-4 km and 20-25 km, respectively; whereas multi-ring craters are the largest planetary impact structures with diameters of a few hundred kilometres to >1000 km. Despite the lack of preservation of the original Vredefort crater owing to erosion, it is most likely that the Vredefort crater was a multi-ring basin.

The formation of a central uplift is induced by inward and upward directed convergent flow of the crater floor, after the target rock has been vertically and centrifugally compressed during formation of the transient crater (Kenkmann et al., 2005). The rocks undergo temporary strength degradation, allowing the flow to occur (Melosh, 1979). Melosh (1979, 1989) described acoustic fluidization as a mechanism that allows for the temporarily weakened state of the rocks during uplift. Acoustic fluidization decreases towards the end of the process of outward collapse, resulting in the fluid-like material regaining strength to behave as a solid (Melosh and Ivanov, 1999).

In the Vredefort crater, the central uplift resulted in widespread overturning of supracrustal strata that now make up the collar of the Vredefort Dome, with increasing amounts of rotation from the core to the collar (Lana et al., 2003b, 2003a; Wieland et al., 2005). The Potchefstroom Rim Syncline was also formed through upward drag of the centrally uplifted region (Simpson, 1977), which McCarthy et al. (1990) stated as the probable cause of preservation of the gold reefs over much of the Witwatersrand Basin. The entire process of forming the central uplift is characterised by: (a) initial outward and downward displacement of the transient crater accompanying shock compression; (b) upwards and inwards displacement associated with post-shock relaxation; and (c) outward collapse of the uplifted rocks. Constriction occurs briefly during (b). Wieland et al. (2005) suggested that tangential shortening during (b) formed radial folds and radial and oblique faults hosted in the collar of the Vredefort Dome. The West Rand and Central Rand groups are estimated to have been shortened by approximately 7 - 17% (Manton, 1965; Lilly, 1978; Wieland et al., 2005) and 5 - 7% (Wieland et al., 2005), respectively. Simpson (1977) reported negligible tangential shortening in the Black Reef Formation.

If the amount of structural uplift is ~ 10% of the diameter of the final crater (French, 1998), an uplift of 20-30 km is implied for the Vredefort crater (Therriault et al., 1997; Henkel and Reimold, 1998). Henkel and Reimold (2002) suggested a minimum of 10-12 km uplift of the core-collar contact in the Vredefort Dome. When added to the amount of erosion, it suggests at least 20 km of uplift (Gibson et al., 1998; Henkel and Reimold, 1998; Gibson and Reimold, 2008). Henkel and Reimold

(1998) and Melosh and Ivanov (1999) estimated that central uplift formation lasted approximately 2-3 minutes. The sequence (a) to (c) for central uplift formation gives rise to highly complex and rapidly changing stresses; consequently structures are expected to be complex, with multiple overprinting relationships spanning as little as a few minutes. Erosion has since removed the crater rim and impact breccia fill, and exposed the deeper parts of the central uplift (Vredefort Dome) and the Potchefstroom Rim Synclinorium (McCarthy et al., 1990; Gibson et al., 1998; Henkel and Reimold, 1998).

2.4.2. The geology of target rocks

The core is underlain by Mesoarchaeon basement gneisses, whereas the collar comprises generally vertical to overturned rocks of the Dominion Group and Witwatersrand, Ventersdorp and Transvaal supergroups, with minor mafic to granitic intrusions of varying ages (Figure 2.2; Wieland et al., 2005; Gibson and Reimold, 2008). The outermost units of the Transvaal Supergroup exhibit a relatively gentle dip that forms the Potchefstroom Rim Synclinorium.

2.4.2.1. The Archaean Granite Complex core

The core of the Vredefort Dome is composed of 3.2-3.1 Ga granitoid basement gneisses (Archaean Basement Complex; Moser et al., 2001; Lana et al., 2003a; Gibson and Reimold, 2008). The Archaean Basement Complex is migmatitic and composed of trondhjemite-, granodiorite- and granite-gneisses and granite, with minor mafic and metasedimentary xenoliths (Lana et al., 2003a; Reimold and Gibson, 2006). These rocks were metamorphosed to upper amphibolite facies grade in the outer core and granulite facies in the inner core of the dome before the impact event (Section 2.4.4; M₁; de Wit, 1992; Moser et al., 2001; Ogilvie, 2010).

2.4.2.2. The collar

The collar consists of vertical to overturned, locally right-way-up, supracrustal strata of the Dominion Group and the Witwatersrand, Ventersdorp, and Transvaal supergroups (Figure 2.1; Lana et al., 2003a; Wieland et al., 2005; Reimold and Gibson, 2006). The major difference between supracrustal rocks exposed in the dome and those exposed in the wider Witwatersrand Basin is the metamorphic grade (Phillips and Law, 1994). Additionally the Dominion Group (3.074 ± 0.006 Ga; Armstrong et al., 1991) consists of only tholeiitic-andesitic lava rocks in the collar (Jackson 1992, 1994; Bisschoff, 2000), whereas the Dominion Group also hosts quartzite and meta-conglomerate units in other parts of the Witwatersrand Basin (Bisschoff, 2000; Marsh, 2006).

Meta-dolerite sills of suspected Ventersdorp age are intruded into the Witwatersrand Supergroup, Dominion Group, and part of the Archaean Basement Complex (Pybus, 1995; Bisschoff, 2000). Also, several alkali granitic igneous bodies (Figure 2.2) and mafic sills intruded the strata

during Bushveld magmatism at approximately 2.06 Ga (Bisschoff, 1982; Graham et al., 2005; De Waal, 2008). The igneous bodies, along with the supracrustal rocks, were deformed by the 2.02 Ga impact (Gibson and Reimold, 2008).

The deformed target rocks were subsequently intruded by a series of post-impact Umkhondo-age tholeiitic intrusions (^{40}Ar - ^{39}Ar age of approximately 1.1 Ga; Reimold et al., 2000), including the Anna's Rust Sheet (Gibson and Reimold, 2008). The southeastern half of the dome is covered by Phanerozoic (0.30-0.18 Ga; Graham et al., 2005) Karoo Supergroup sediments and dolerite sill intrusions (Figure 2.2).

2.4.2.3. The Potchefstroom Rim Synclinorium

The Potchefstroom Rim Synclinorium is the outermost large-scale structural remnant produced by the Vredefort Impact. The synclinorium was exposed by the 5 – 10 km erosion along with the core and collar, exposing deeper levels of the crater (McCarthy et al., 1990; Gibson et al., 1998; Henkel and Reimold, 1998). The synclinorium is defined by folded strata of the Transvaal Supergroup, which host pseudotachylitic breccia associated with fold hinges and faults (Simpson, 1977). The age of pseudotachylitic breccia in the synclinorium and wider goldfields area is consistent with the Vredefort impact, however multiple generations of pseudotachylite veins and breccias are also associated with pre-Transvaal Supergroup faults (Section 2.4.3.1; Mambane et al., 2011; Manzi et al., 2012a, 2012b).

2.4.3. Melt-rocks and impact-related features

The geological evidence for impact-related shock includes shatter cones (Hargraves, 1961), planar deformation features (PDF) in quartz (Fricke et al., 1990; Leroux et al., 1994), shock deformation of zircons (Reimold et al., 2002), and the occurrence of coesite and stishovite (Martini, 1978, 1992). Coesite and stishovite are high-pressure silica polymorphs that are hosted adjacent to and in thin pseudotachylitic breccia veins, with stishovite as a rarer occurrence (Martini, 1992). PDF are formed in quartz along crystallographic planes as closely spaced, parallel planar lamellae, under shock pressures ranging from 8-25 GPa (French, 1998 and references therein). The outer limit of PDF in the dome is the Central Rand Group, whereas the outer limit of shatter cones, which can form at pressures as low as 4 GPa, is the Pretoria Group (Figure 2.1; Grieve et al., 1990; Lilly, 1978; Wieland et al., 2006).

Shatter cones are characterised by conical or part-conical fracture surfaces with striations that diverge from an apex (French, 1998). They are believed to form during reflection and scattering of a shock front against inhomogeneities and discontinuities (e.g., joints, bedding, and grains) in the rock, at minimum shock pressures of ~ 4 GPa (Milton, 1977; French, 1998). The stress intensity of the reflected and scattered wave varies locally (Baratoux and Melosh, 2003); this is reflected by the wide range of striation angles of shatter cones about the apex (c.f., Wieland et al., 2006). Furthermore, the

direction of wave scattering is highly variable owing to different shapes of heterogeneities, resulting in shatter cones having inconsistent orientations (Wieland et al., 2006). Studies by Sagy et al. (2002, 2004) proposed that shatter cones form as tensile fractures under decompression, immediately after passage of the shock pulse. This occurs between the end of the contact/compression phase and the inception of the excavation phase (Milton, 1977; Milton et al., 1996).

The Vredefort Granophyre and pseudotachylitic breccia are impact-generated melt-rocks recognised in the Vredefort Dome (c.f., Shand, 1916; Reimold and Gibson, 2005, 2006). The Vredefort Dome is particularly famous for hosting some of the largest volumes of pseudotachylitic breccias on Earth (Reimold and Gibson, 2006).

2.4.3.1. *Pseudotachylitic breccia*

Pseudotachylitic breccia is observed in almost every lithology in the Vredefort Dome (Reimold and Gibson, 2006), and four mechanisms of formation have been suggested: (a) localized shock melting by extreme shock pressure fluctuations, (b) slip surface friction melting, (c) a combination of shock and friction melting; and (d) decompressional melting (Reimold and Gibson, 2005, 2006). The different mechanisms of formation create ambiguity in the terminology of the melt-rock produced, which is why a more generalised, non-generic term “pseudotachylitic breccia” is used in this thesis (Lana et al., 2003b; 2003a; Reimold and Gibson, 2006; Hein, 2012).

Pseudotachylitic breccias occur in a wide variety of geometries in the dome, from straight and irregular veins, to <1 mm wide vein networks and massive dykes up to several metres wide (Reimold and Gibson, 2006). The core of the dome hosts 10 m wide and 100 m long pseudotachylitic breccia dykes (Dressler and Reimold, 2004; Lieger et al., 2009). In contrast, the most voluminous occurrences of pseudotachylitic breccias in the collar rocks are hosted in fault zones, and within highly fractured hinge zones of km-scale folds. The width of individual pseudotachylitic breccia veins and their overall abundance in the collar decrease radially outward from the West Rand Group (Wieland et al., 2005) to the Transvaal Supergroup (Simpson, 1977). Other outcrops of pseudotachylitic breccia in the collar include centimetre-wide veins that normally follow bedding contacts, and en echelon vein sets (Reimold and Gibson, 2006). Gibson et al. (1997b) showed abundant micro-veins associated with spaced fracture cleavage hosted in units of the West Rand Group. The fracture cleavage involves mm- to cm-scale displacements of bedding.

Geochronological studies suggest that pseudotachylitic breccia in the dome and in the Witwatersrand goldfields formed during the Vredefort impact (2.01-2.02 Ga ^{40}Ar - ^{39}Ar white mica, Trieloff et al., 1994; 2.02 ± 0.004 Ga U-Pb SHRIMP from authigenic zircons in pseudotachylitic breccia, Kamo et al., 1996). Although the age of pseudotachylitic breccia in the goldfields is consistent with the Vredefort impact, multiple generations of pseudotachylite veins and breccias formed by frictional melting are also associated with pre-Transvaal Supergroup faults in the West

Rand goldfield (Mambane et al., 2011; Manzi et al., 2012a, 2012b). 3D seismic data by Manzi (2010) showed that these pseudotachylite-bearing faults do not penetrate the base of the Transvaal Supergroup. Mambane et al. (2011) concluded that the majority of pseudotachylite veins in the West Rand goldfield are tectonically induced, rather than being formed by the Vredefort impact. This conclusion is consistent with that of Killick and Reimold (1990) and Berlenbach and Roering (1993).

2.4.3.2. Vredefort Granophyre

In addition to pseudotachylitic breccias, the dome hosts nine vertical, clast-laden, 10-60 m wide impact-melt dykes, called the Vredefort Granophyre (Bisschoff, 1972; Reimold et al., 1990; Koeberl et al., 1996; Reimold and Gibson, 2006). These dykes are hosted in single or en echelon fractures that are broadly radial or tangential to the dome. The dykes are up to several kilometres long, and intrude both the core and inner collar (Reimold and Gibson, 2006). Koeberl et al. (1996, 2002) used Re and Os isotope data to show that the Granophyre contains a <0.2% extra-terrestrial component.

Gibson et al. (1998), Gibson and Reimold (2000), Reimold and Gibson (2006) suggested that the dykes represent intrusive offshoots of the homogenised impact melt that once lay within the crater. This is supported by evidence that the dykes have a uniform bulk composition (Therriault et al., 1996, 1997; Reimold and Gibson, 2006). The dykes are not affected by large-scale folds or oblique and radial faults that have been linked to central uplift formation, suggesting that they intruded and crystallised after overturning and tangential shortening of the collar; during or after the final collapse of the central uplift (Wieler et al., 2005).

2.4.4. Metamorphism in the Vredefort Dome

Rocks of the Dominion Group and Witwatersrand Supergroup exposed in the Vredefort Dome host metamorphic assemblages formed under higher temperature and pressure conditions than those found in some units exposed around the edges of the Witwatersrand Basin (Figure 2.8; Phillips and Law, 1994; Foya, 2002; Gibson and Reimold, 2008). Metamorphic grades in the Vredefort Dome increase with stratigraphic age from lower greenschist facies (approximately 350 °C and 2-3 kbar; Foya, 2002) in the Central Rand Group to mid-amphibolite facies (570-600 °C and 3.5-4.5 kbar; Gibson and Wallmach, 1995 and references therein) in the lower West Rand Group. Upper amphibolite facies assemblages occur in the outer core and granulite facies in the inner core of the dome, reaching a maximum of ~875 °C and ~7.6 kbar (Ogilvie, 2010). However, the granulite-grade facies metamorphism in the Archaean Basement Complex is attributed to a 3.1 Ga event (M_1 ; de Wit, 1992; Moser et al., 2001; Ogilvie, 2010), which formed gneissic textures and predates deposition of the supracrustal collar rocks. Consequently, this metamorphism is not considered further here.

M_2 metamorphism is described by Schmitz and Bowring (2003) as a 2.72-2.69 Ga (Ventersdorp magmatism age) ultra-high-T metamorphic event in the lower crustal granulite xenoliths from,

amongst others, the Lace and Voorspoed mines located ~45 km southwest of the centre of the dome. However, no metamorphic textures associated with this metamorphic event have been identified in rocks located in the central parts of the Vredefort Dome. It thus seems likely that at least some metamorphism affected the Witwatersrand Supergroup during eruption of Ventersdorp Supergroup basalts (Dawson and Smith, 1987; Dawson et al, 1997); however, it is unclear whether significant metamorphic assemblages were formed.

Metapelite units of the West Rand Group in the northern and northwestern parts of the collar exhibit andalusite, garnet, cordierite, staurolite, and biotite ($\pm$ fibrolite) peak metamorphic assemblages (Bisschoff, 1982; Gibson, 1993; Phillips and Law, 1994; Gibson and Wallmach, 1995; Gibson and Reimold, 2008); whereas the Central Rand Group hosts pyrophyllite-chloritoid-chlorite-muscovite and kyanite-andalusite-chlorite-muscovite peak metamorphic assemblages, with the latter occurring in quartzite horizons (M_3 ; Bisschoff, 1982; Phillips and Law, 1994; Foya 2002). These “knotted” quartzite horizons host spherical aggregates up to 3 cm in diameter formed by intergrown kyanite and andalusite porphyroblasts (Bisschoff, 1982). Metamorphic assemblages hosted in pelitic units in the West Rand Group indicate peak conditions of 570-600°C and 3.5-4.5 kbar (M_3 ; Gibson and Wallmach, 1995), consistent with an estimated burial depth of 14-16 km at the time of metamorphism. Gibson and Wallmach (1995) used metamorphic reactions to infer that the metamorphism was characterised by an anticlockwise P-T path (Figure 2.11).

Phillips and Law (1994), Gibson and Stevens (1995), and Gibson and Wallmach (1995) concurred that the metamorphism of the supracrustal rocks in the collar must have been aided by advective heat transfer by magma owing to an unusually high thermal gradient of 35-40°C/km, in contrast with the expected background gradient of 15-20°C/km for the craton (Gibson and Jones, 2002).

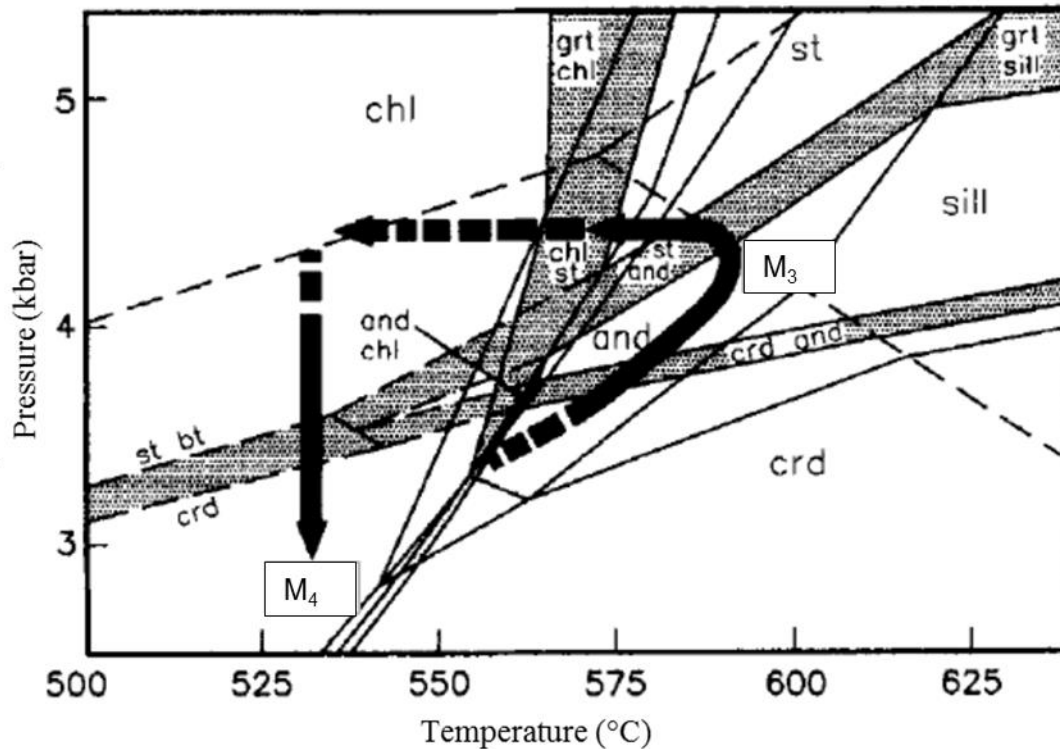


Figure 2.11. An anticlockwise *P-T* path indicated by *M*₃ metamorphic mineral textures, with *M*₄ metamorphic path included (modified from Gibson and Wallmach, 1995). *Chl* = chlorite, *grt* = garnet, *st* = staurolite, *sill* = sillimanite, *and* = andalusite, *crd* = cordierite.

Bisschoff (1982) suggested that the metamorphic isograds in the collar are elliptical, with an approximately 50 km long northwest-southeast axis and a shorter 40 km northeast-southwest axis. This was linked to the interpretation that alkali granite intrusions hosted in the collar were the source of heat. However Gibson and Wallmach (1995) disputed the interpretation owing to the regional extent of metamorphism and lack of evidence of increasing metamorphic grade towards the intrusions. Nonetheless, West Rand Group rocks in the northeastern collar show a lower grade (upper greenschist facies; andalusite-chloritoid±biotite) than their equivalents in the northwestern collar (Bush, 2014; Nemutandani, 2014). Bush (2014) and Nemutandani (2014) attributed this variation to displacement of the West Rand Group strata to different structural depths along pre-metamorphic faults.

Attempts to date the pre-impact metamorphism in the West Rand Group have, however, produced ambiguous results; yielding ages ranging from 2.02 to 1.85 Ga (e.g., 2.00 ± 0.016 Ga biotite ^{40}Ar - ^{39}Ar ; Allsopp et al., 1991; 1.94 ± 0.092 Ga, Rb-Sr whole rock; Hart et al., 1981; 2.027 ± 0.014 Ga, biotite ^{40}Ar - ^{39}Ar ; Gibson et al., 2000).

All rocks in the Vredefort Dome were subsequently subjected to *M*₄ metamorphism through shock-heating and impact-induced uplift at 2.02 Ga (Wieland et al., 2005). Foya (2002) interpreted relatively uniform post-impact thermal conditions of approximately 300 °C in the Kimberley Reefs

(Central Rand Group) across the basin, but Gibson et al. (1998), Gibson (2002) and Ogilvie (2010) noted an increase towards the centre of the dome, to about 525 °C ($P < 3.5$ kbar; Figure 2.11) in the lower West Rand Group (inner collar of the dome), and >1000 °C in the core of the dome. The M_4 metamorphism is represented by: (1) chlorite-sericite alteration that overprints the peak, pre-impact, lower greenschist facies chloritoid-pyrophyllite assemblage in the Kimberley Reefs (Foya, 2002; Reimold and Gibson, 2006); and (2) post-shock recrystallization adjacent to and in the matrix of pseudotachylitic breccia veins (Gibson and Wallmach, 1995; Reimold and Gibson, 2006). Gibson et al. (1998) concluded an approximately 40 – 50 °C/km lateral thermal gradient across the dome related to increasing levels of shock heating and differential uplift to the centre of the dome.

2.4.5. Structural framework

Limited investigations of the large-scale structural character of the collar of the Vredefort Dome have been conducted (c.f., Wieland et al., 2005; Gibson and Reimold, 2008). One of these studies, by Antoine et al. (1990), suggested that the dome has a polygonal shape, rather than being circular. Their study resulted in the division of the exposed section of the dome into six segments, at 40° to 45° to each other, with two more segments possibly covered by the Karoo Supergroup rocks. As a result, recent structural studies on the collar of the dome initially treated the collar as being divided into these six segments (c.f., Lana et al., 2003a & b; Wieland et al., 2005). However, Wieland et al. (2005) showed that the collar does not host large radial faults between the polygonal segments as suggested by Antoine et al. (1990); instead bedding strike varies in a more continuous way around the dome (Figure 2.12; Table 1).

The collar strata were overturned during the modification stage (Wieland et al., 2005). These overturned to steeply-dipping right-way-up strata are separated from gently outward-dipping, right-way-up strata within the Potchefstroom Synclinorium by concentric normal faults (Simpson, 1977). Furthermore, Wieland et al. (2005) described local areas of right-way-up strata in the northern and northeastern segments of the dome, and in the southeast (Figure 2.12d, e, g). Lana et al. (2003a & b) and Wieland et al. (2005) suggested that supracrustal rocks that are not overturned in the collar were possibly affected by pre-impact structures and/or impact-induced fault-block rotation.

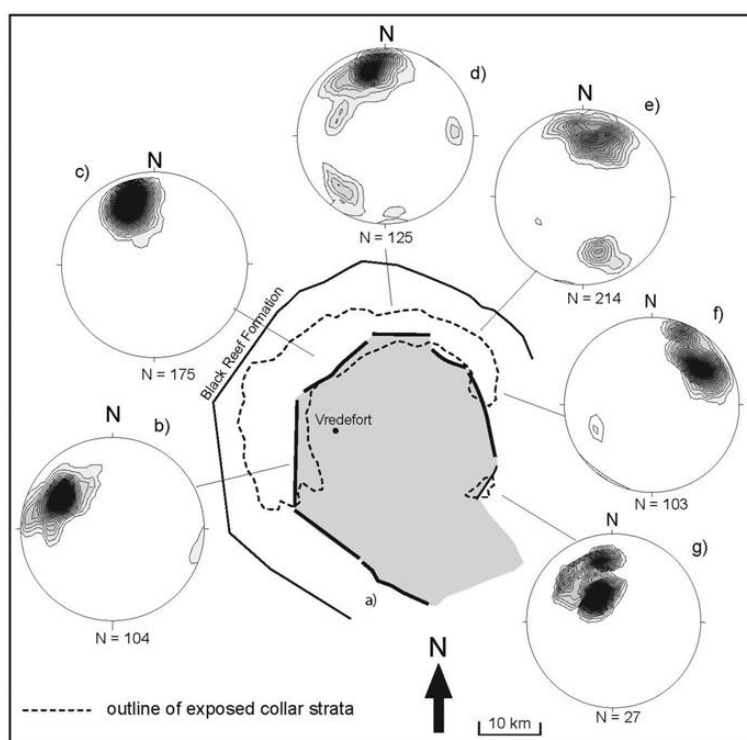


Figure 2.12. Equal area stereographic projections and contour plots of poles to bedding of the Witwatersrand Supergroup collar rocks in the Vredefort Dome (after Wieland et al., 2005). Plots 'b' to 'g' represent six segments of the collar as suggested by Antoine et al. (1990).

Table 1. Mean and Maximum density values for Schmidt net projections shown in Figure 2.12 (Wieland et al., 2005).

	b	c	d	e	f	g
Mean	36 °/300 °	30 °/336 °	24 °/349 °	36 °/007 °	36 °/053 °	60 °/144 °
Maximum density	27.8	25.9	16.9	11.8	17.3	30.7

2.4.5.1. *Folds*

In the collar rocks, several prominent impact-induced folds exist in the West Rand Group (Wieland et al., 2005); however, none are found in the Central Rand Group (Bisschoff, 2000). This may partly be because of a relatively large competency contrast between quartzite and slate units in the West Rand Group (whereas the Central Rand Group is more homogeneous); but also because tangential shortening decreases radially outwards (Wieland et al., 2005).

Wieland et al. (2005) described kilometre-scale open to tight symmetric folds whose axial planes and fold hinges are oriented radial to the core of the dome, and open asymmetric homoclinal folds with axial planes oblique (an orientation that is between tangential and radial) to the dome in the West Rand Group. Intense folds are developed in the Hospital Hill Subgroup (northwestern sector of the dome), whereas the rest of the collar generally exhibits local bedding variations related to gentle folding and fault-related block rotation (Wieland et al., 2005). In the Transvaal Supergroup on the inner limb of the Potchefstroom Synclinorium, axial planes of major, km-scale folds are mostly arranged concentric to the dome, parallel to the fold axis of the synclinorium (Simpson, 1977). Simpson (1977) interpreted that these folds verge radially outwards and are characterised by faulted hinges, consistent with rapid formation during the impact event. This is similar to observations by Wieland et al (2005) who also showed that fold hinges in the West Rand Group are commonly brecciated.

From the observation that axial planes of major and minor folds are radial or tangential to the core of the Vredefort Dome, Simpson (1977) suggested that folding, warping, and overturning were simultaneous with updoming, which is consistent with later interpretations of folds developed in the West Rand Group by Wieland et al. (2005). Based on numerical modelling of the central uplift formation (Collins et al., 2004), the folds would have formed in the intermediate, constrictional, stage of central uplift development.

2.4.5.2.. *Faults*

Faults crosscutting the collar are described by Simpson (1977), Lilly (1978) and Wieland et al. (2005) as being radial and oblique to the dome. The dominant displacement along the faults in the collar and rim synclinorium is left-lateral and, less commonly, right-lateral strike-slip (Lilly, 1978; Wieland et al., 2005), with subordinate normal faults (Simpson, 1977; Wieland et al., 2005). Normal faults in the rim synclinorium are both concentric and radial to the dome (Simpson, 1977). These faults cut across impact-induced folds, indicating they postdate ductile impact-induced deformation (Simpson, 1977; Lilly, 1978; Wieland et al., 2005).

Intensely fractured fault zones in quartzite units in the West Rand Group and the Pretoria Group host pseudotachylitic breccia vein networks (Simpson, 1977; Wieland et al., 2005). Lilly (1978) and Wieland et al. (2005) attributed the formation of faults in the collar to tangential shortening around the dome during the final stages of rise of the central uplift, but before the outward

collapse phase. This was based on the slip geometries and relation of the faults to smaller-scale microfaults. The amount of shortening postulated is 7 - 17% (Manton, 1965; Lilly, 1978; Wieland et al., 2005) in the West Rand Group and 5 – 7 % in the Central Rand Group (Wieland et al., 2005).

In an earlier study, Lilly (1978) highlighted that primary/major faults in the western and northwestern collars have intermediate oblique angles to the dome's circumference, whereas those in the northern collar make high angles with the dome's circumference (Figure 2.13). Lilly (1978) showed that the maximum compressive stress during fault formation was tangential to the dome in the western and northwestern collars, but radial to the dome in the northern collar. In general, tangential shortening seems to have formed the majority of faults in the collar (Wieland et al., 2005), with only the northern collar being inconsistent (Lilly, 1978). One of the suggested reasons for this inconsistency is the presence of pre-impact faults within the northern collar.

The northern collar thus provides an excellent natural laboratory to distinguish possible pre-impact structures from those formed by impact-related deformation and is, thus, the subject of this study.

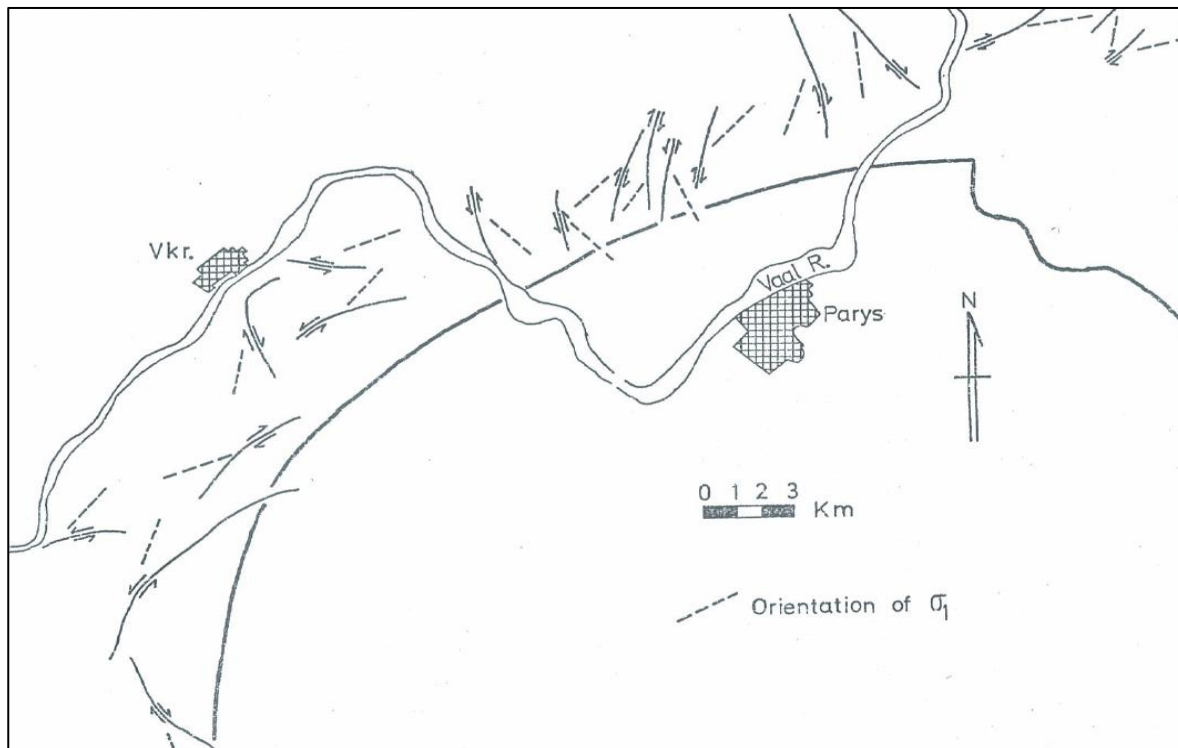


Figure 2.13. Pattern of primary-scale (major) faults in the collar of the Vredefort Dome with interpreted orientations of maximum principle stress that are oblique to the dome in the western and north-eastern sectors, but radial in the northern sector (after Lilly, 1978).

3.0 METHODOLOGY

The study involved a combination of field and desktop mapping, petrography, and chemical analyses of samples. Desktop mapping was implemented early in the study to plan field work.

3.1. Lineament analysis

Orthophotographs and satellite images of the northern collar of the Vredefort Dome were used to delineate major structures cutting across the collar rocks. Orthophotographs at 1:10 000 scale within geographic blocks 2627 C & D, were obtained, and georeferenced to UTM WGS84. The orthophotographs were supplied by Surveys and Mapping (Mowbray, Cape Town). Satellite images were acquired from Google EarthTM. These images were overlapped using CorelDRAW X3[®] to form a high-resolution mosaic as base maps (Figure 3.1).

Although compilation of Google EarthTM images produces a mosaic with better resolution than the orthophotographs, it is associated with distortions. The images change in scale and the angle of topographic features, resulting in distortions of tens of metres. This occurs even when the zoom size is kept constant. The distortions mostly occur where there are large changes in elevation across the area being imaged, i.e., ridges versus valleys. To reduce the distortion effect, large overlaps were used (up to 50% overlap). The distortion was also reduced by starting the mosaic from the centre outwards, so that most of the distortion lay at the edges of the mosaic (Figure 3.1).

Compiled layers of satellite images, orthophotographs, and a geological map (Bisschoff, 2000) were used to select target areas of favourable outcrops, which made field mapping more efficient. These layers were overlain and georeferenced to a single coordinate system (Latitude/Longitude in degrees, minutes, and seconds) using ArcGIS 10.1TM. This programme was also used to draw geological maps of the study area. These include: (1) a geological map (Figure 3.2); (2) structural map (Figure 3.4); (3) metamorphic mineral distribution map (Figure 3.5); (4) pseudotachylitic breccia distribution map (Figure 3.6); and (5) pseudotachylitic breccia volume map (Figure 3.7). A better quality version of these maps is available on the .pdf format of the thesis; also, A1 size copies of these maps are available from the library of the University of Witwatersrand.

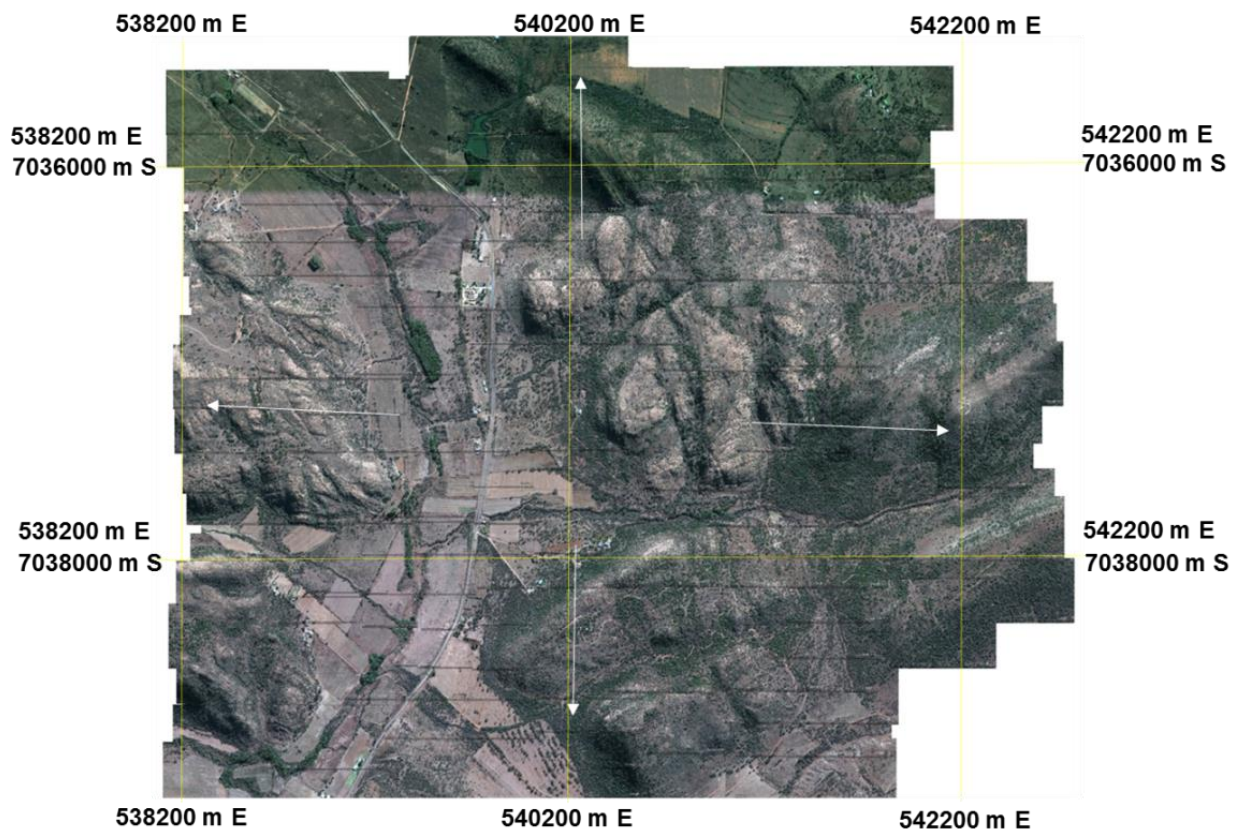


Figure 3.1. Google EarthTM image mosaic of the north-eastern portion of the study area. The arrows indicate the directions in which the mosaic was compiled.

3.2. Field mapping

The high number of private farms in the study area (Figure 1.2), some of which have wild animals on them, presented a degree of limitation to field mapping. Personal communication and trust had to be established to negotiate permission to work on the farms; several farms with absentee owners remained inaccessible.

Field mapping was carried out using a Silva compass clinometer, an Etrex Garmin high sensitivity GPS, measuring tape, and Samsung ES30 and Canon EOS 1100D cameras. Field photographs discussed in the thesis include coordinates where the outcrop was photographed. Most of the mapping (lithological and structural) was undertaken within the Witwatersrand Supergroup because it has well exposed outcrops with better preserved structures. Structural mapping involved recording geometric and kinematic data from bedding and fractures (over 1600 measurements; Appendix II). Other measured structural elements included cleavages and lineations (striations and intersection lineations). Lithological mapping delineated marker horizons, which were used to establish and estimate the sense and magnitude of displacement along faults. Metamorphic mineral

assemblages were noted and samples collected because they have significant implications for the geological history of the dome, particularly the timing of deformation relative to metamorphism.

Outcrop photographs from the mapping sessions, used in the thesis, are labelled in a standard format, which includes: (1) the name of the lithological unit located on the top-left of the figure; and (2) the locality coordinates included as part of the figure title (e.g., Figure 3.8).

A database for the study area was created. The data are presented in a number of database files. Numerical data (orientation data of bedding and structural features) were saved as Excel files (Appendix II). Google EarthTM was used to save spatial data (locations of station points and samples), polygonal (e.g., lithology outcrops), and linear (e.g., fault locations) data. The data were saved as .kml files, which were later exported into ArcGISTM 10.1 for further processing. The location of samples on Google EarthTM was important for structural and comparative studies of metamorphic mineral assemblages across faults.

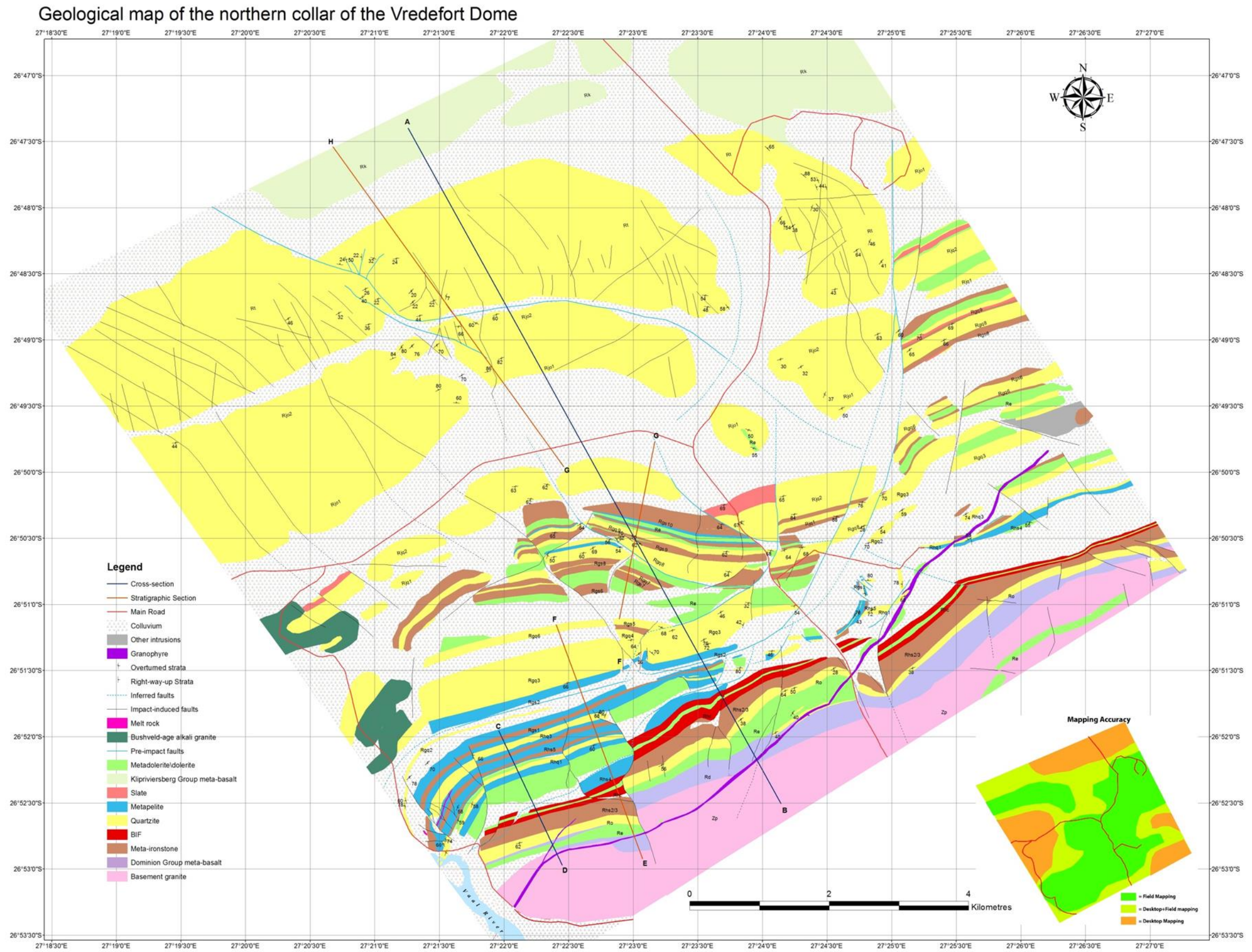


Figure 3.2. Geological Map of the study area. It includes section lines A-B and C-D for the cross-sections (Figure 3.3), and sections E-H along which lithology thicknesses of the stratigraphic column were calculated.

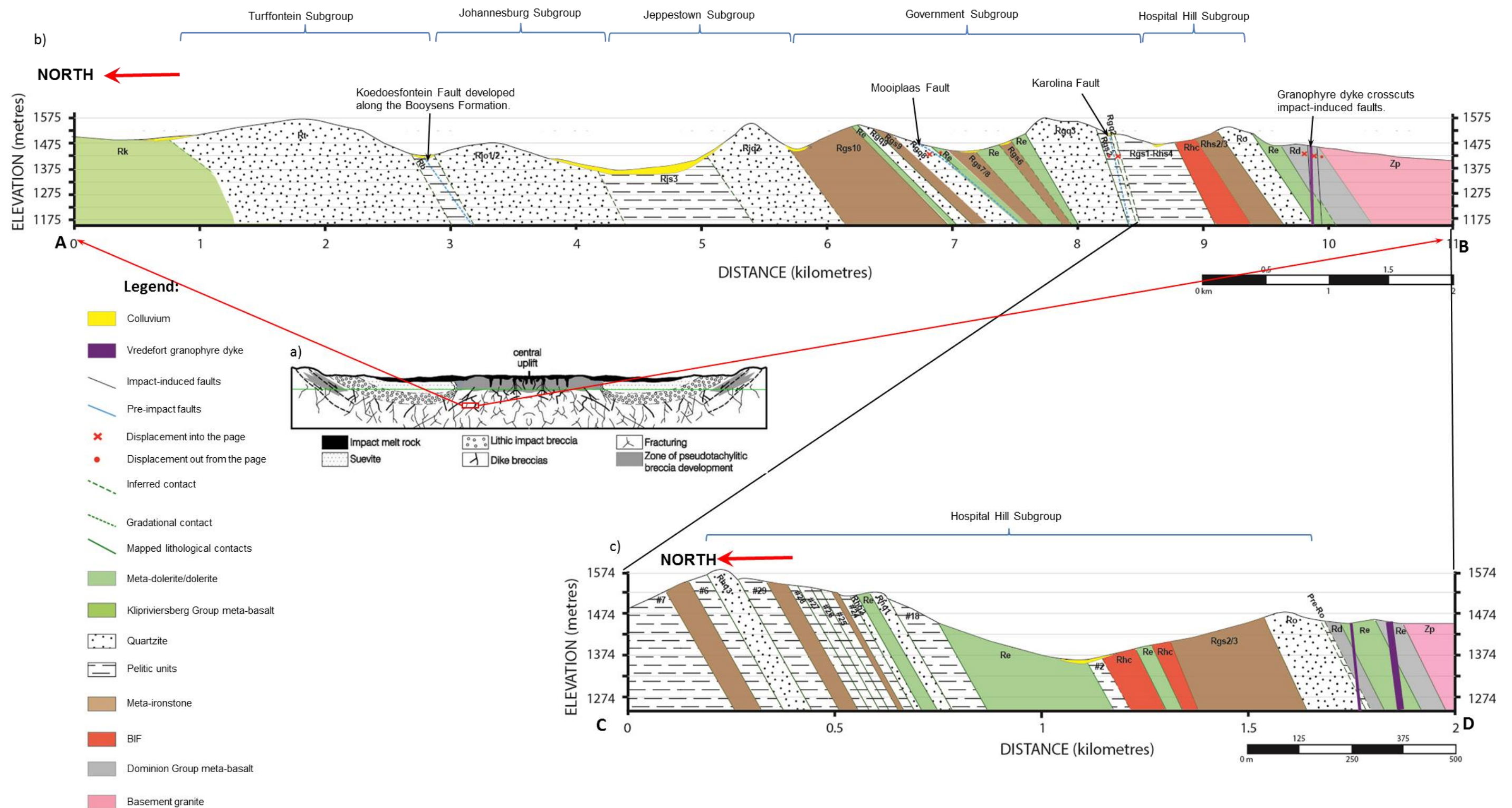


Figure 3.3. (a) A cross-section of the impact structure, showing the approximate location of the study area cross-sections. (b) Cross-section A-B from Figure 3.2, with 2× vertical exaggeration (note the general increase in the degree of overturning towards the north). The section also shows the approximated orientation and apparent displacements of faults. (c) Cross-section C-D, without vertical exaggeration (includes some of the Hospital Hill Subgroup units missing in cross-section A-B due to the effect of faulting). The overturned strata young towards the north.

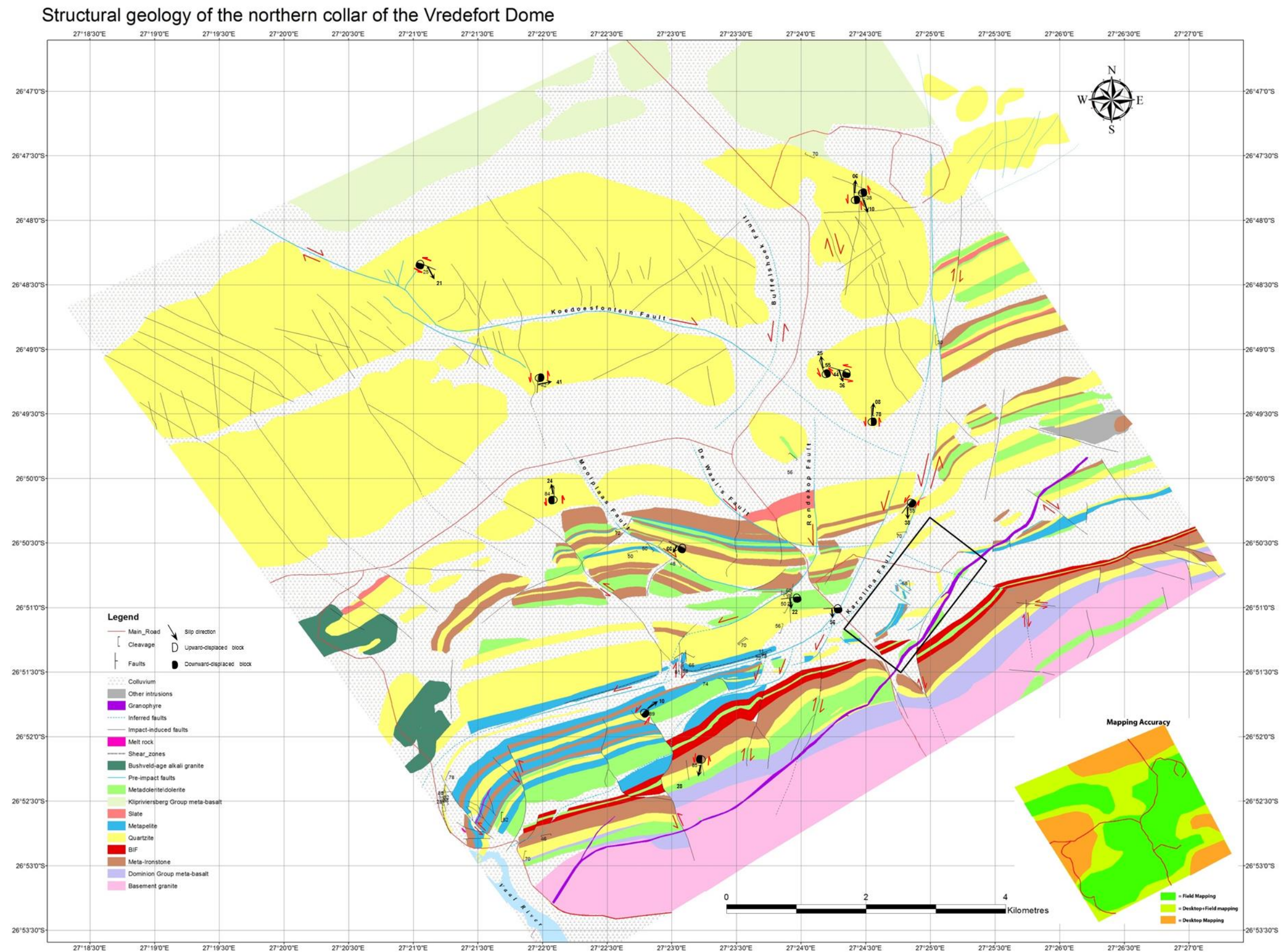


Figure 3.4. Map showing the structural features of the study area. The map includes: (a) orientations of the S_2 cleavage; (b) absolute displacements along impact-induced faults; and (c) The Listric Shift Area (Section 8.3.2) indicated by the black rectangle.

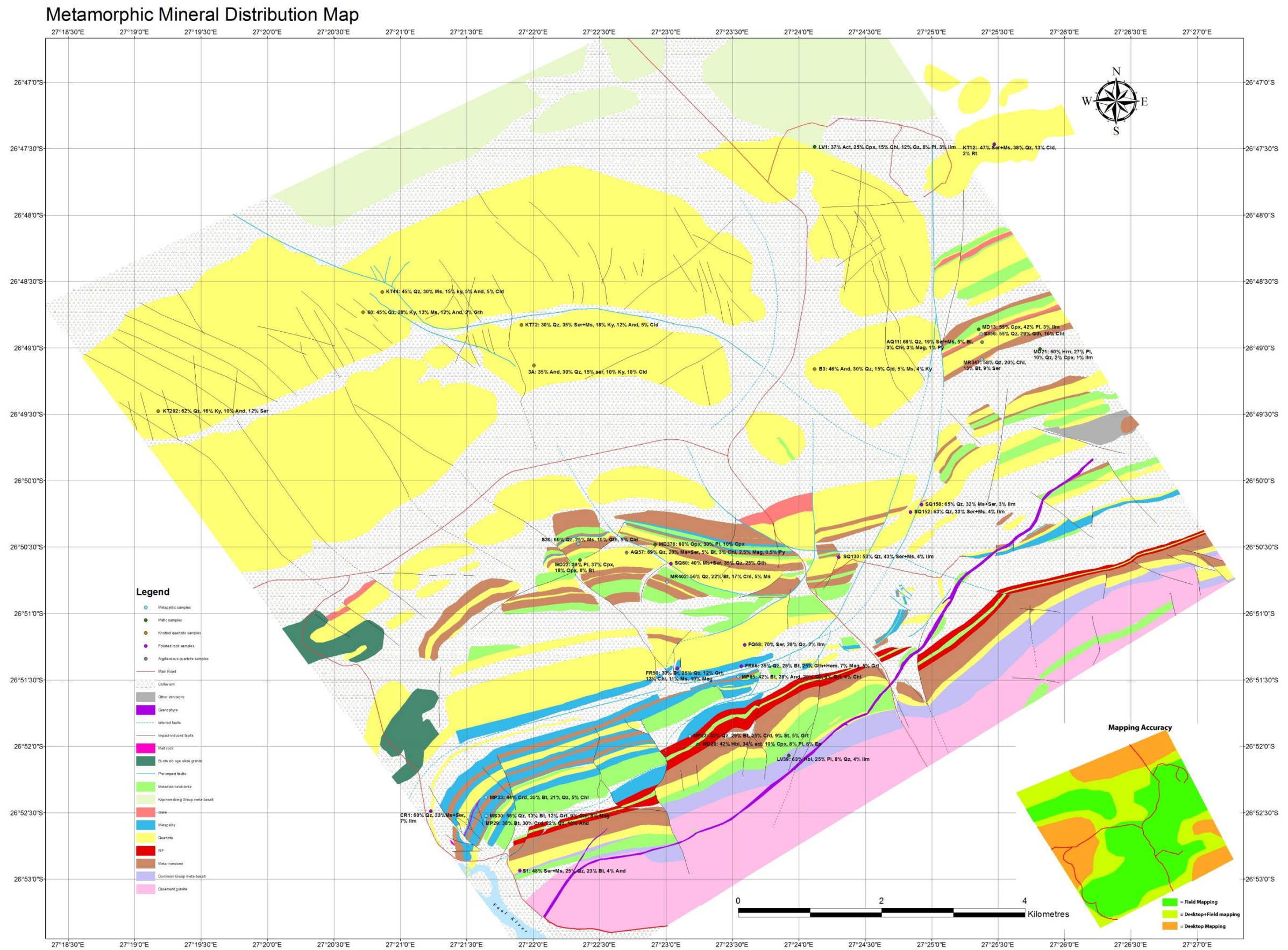


Figure 3.5. Map showing the distribution of metamorphic mineral assemblages across the study area, which were identified from petrography samples included in Appendix III.

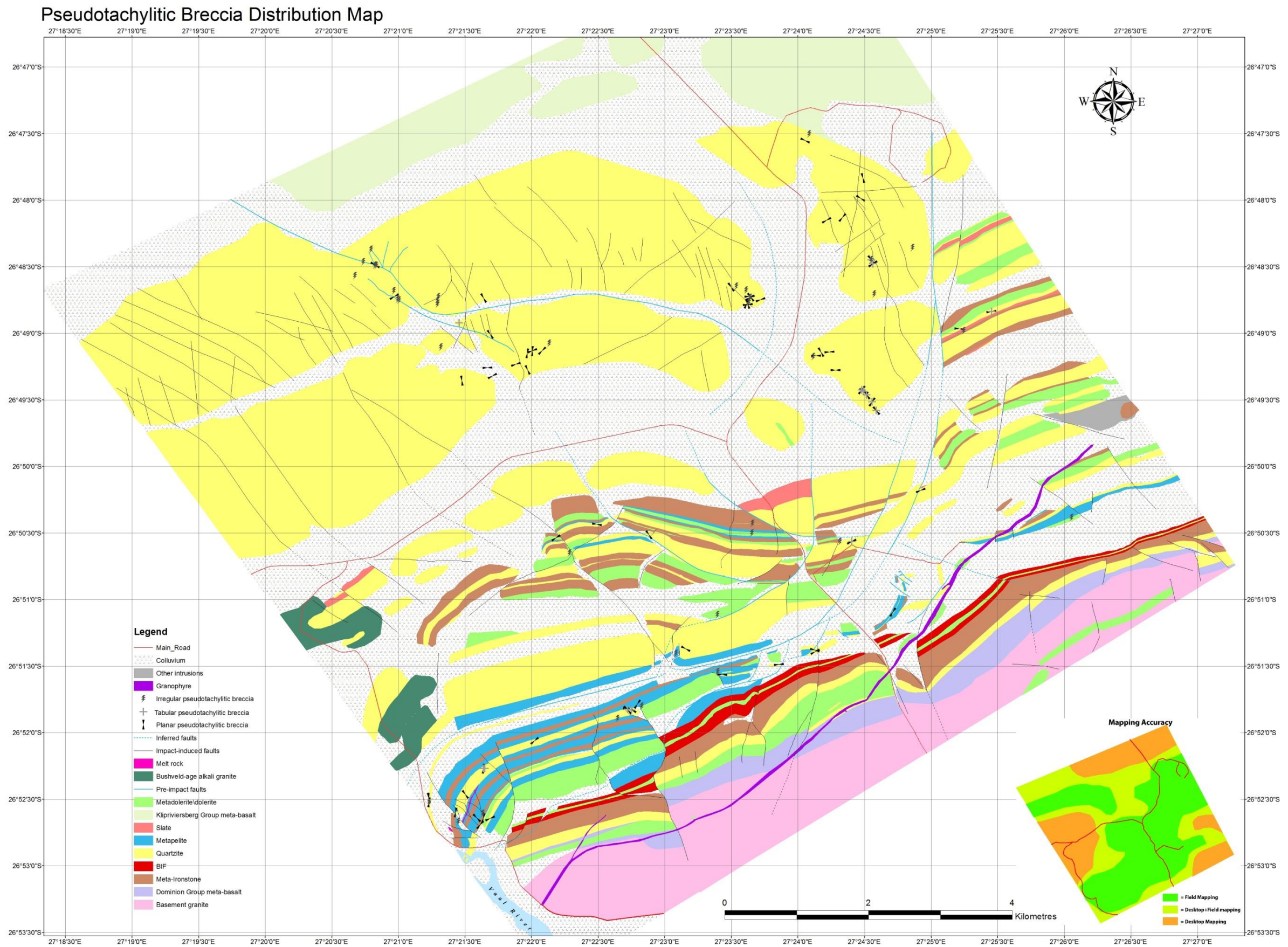


Figure 3.6. Map showing the distribution and geometries of pseudotachylitic breccias mapped in the study area.

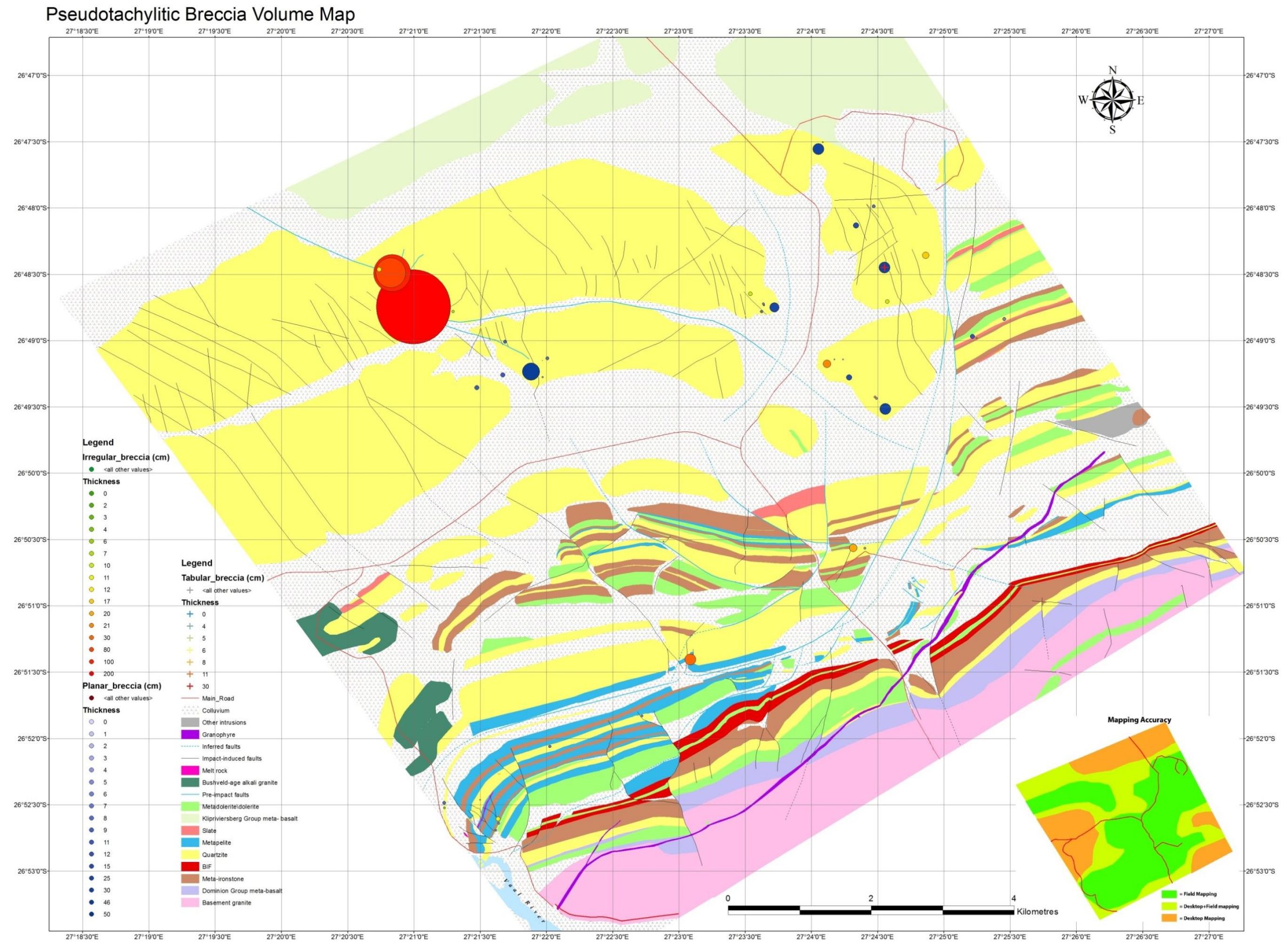


Figure 3.7. Map showing volumes of pseudotachylitic breccia mapped in the study area.

3.3. Geological map, stratigraphy, and graphs

3.3.1. Geological map design

Google Earth™ does not have functions that can represent orientation data. As a result, the .kml files were exported to ArcGIS™ to create complete geological maps (Figures 3.2-3.7). ArcGIS™ files are saved as shapefiles (.shp), and were edited and saved in the programme. To simplify editing, the .kml files were divided into categories on Google Earth™ (e.g., lithological categories: quartzite, meta-ironstone, metapelite). These categories were imported into ArcGIS™ as layers that could be edited individually. To plot bedding orientations, bedding locations were plotted with the appropriate strike-dip symbol, then rotated accordingly to represent the strike and dip angle of that unit.

3.3.2. Construction of the stratigraphic column

The stratigraphy of the study area (Chapter 4) was constructed using Google Earth™ and the field location of lithological contacts. Most rocks in the study area are sedimentary; their dip information and outcrop width were used to calculate true thickness (Figure 3.9). The outcrop width of lithologies was measured perpendicular to the two contacts on Google Earth™. The distance between the two contacts was represented as ‘map length’ and ‘ground length’ on the programme. Ground length accounts for elevation difference, whereas map length is a horizontal measurement. Map length was used for the purpose of this calculation. After true thicknesses were calculated, Adobe Illustrator CS6 was used to construct the stratigraphic column (Figure 4.2).

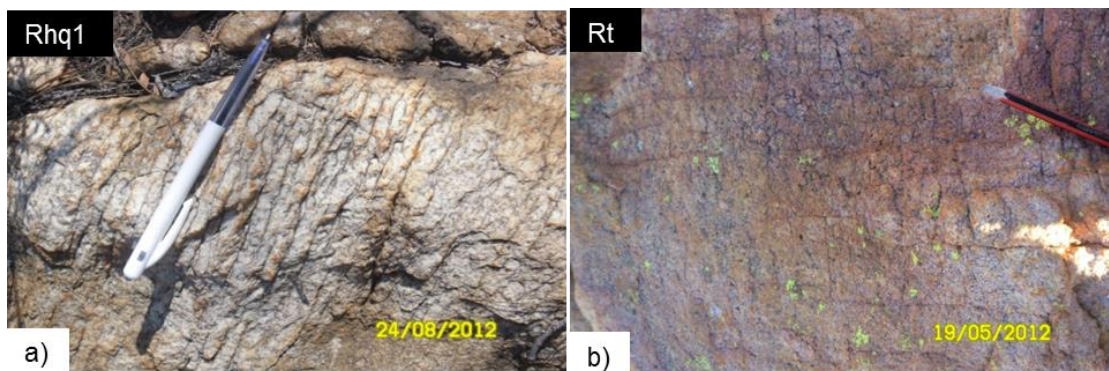


Figure 3.8. A format showing how outcrop photographs are presented in the thesis, with the name of the lithology unit listed on the top-left and coordinates (35J 535653 m E, 7026637 m S and 35J 53510 m E, 7034879 m S, respectively) included in the title.

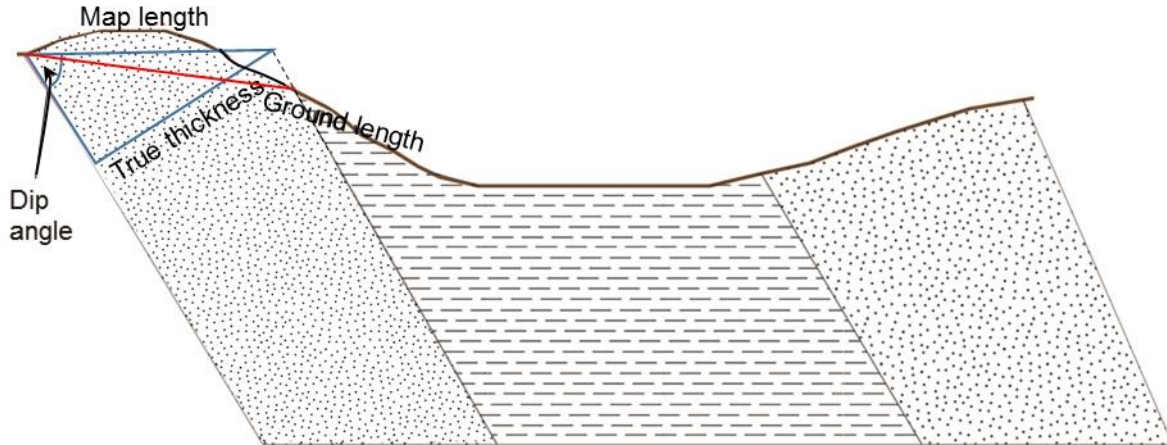


Figure 3.9. Sketch illustrating the calculation used to estimate the true thicknesses of lithological units. Map Length assumes horizontal distance between contacts whereas ground distance accounts for difference in elevation.

3.3.3. Stereonets, roseplots, and graphs

The numerical data collected during field work were analysed using stereographic projections, rose plots, and graphs. All plotted projections are equal area stereographic projections (Schmidt plots), and were plotted using GEOrient® ver. 9.5.0. The numerical data were imported into GEOrient® as text files, categorised according to strike, dip angle, and dip direction. Rose plots were also plotted using GEOrient®. These plots represent the distribution of orientation data with calculated means, but this study also required a spatial representation of orientations of bedding and structural data. Excel scatter plots were used to represent orientation data spatially, through plots of northing or easting coordinates versus strike or dip angles.

3.4. Sample selection

A total of 172 samples were collected from the study area for petrographic and microstructural analyses (Figure 3.10). In general, three types of samples were collected: (1) oriented samples of foliated rocks; (2) lithologies; and (3) melt-rocks produced by the impact. Foliated rocks were studied for microstructural characters, i.e., nature of the foliation and deformation-metamorphism relationships. Melt-rock samples were analysed for their textural and compositional characteristics. All sample locations and descriptions are presented in Table A1.4 – A1.8 (Appendix I). All figures and tables referenced from appendices are labelled with an ‘A’ prefix (e.g., Table A3.2 is Table 3.2 in Appendix III).

3.4.1. Thin section samples

Of the 172 samples, 54 were cut for thin section analysis (location and descriptions in Table A1.4 – A1.8; Appendix I). These samples were oriented in outcrop, so that the true orientations of microstructural features could be ascertained. Samples were cut in multiple sections to show the full 3D character of microstructures. Thin sections were analysed using an Olympus BX51 microscope at the School of Geosciences, mounted with an Olympus DP26 camera. The microscope is fitted with an Olympus TH4-200 light box for reflected light microscopy. The whole system is run by an Olympus Stream EssentialsTM (ver. 1.7) software. This system was used to identify and estimate mineral proportions of the samples. The abundance of minerals was determined through visual estimation of differentiated minerals (modal abundance), on the largest possible scale of the thin section depending on grain size.

An Olympus BX63 microscope at the Microscopy and Microanalysis Unit (MMU, University of the Witwatersrand, Johannesburg) was used to capture a mosaic of several thin sections. It is mounted with an Olympus DP80 camera and run by Olympus CellSensTM Dimension 1.9 software (Build 11514). This system is run by a Windows 7 Mecer Graphite computer (Intel Core i7, 16 gigabyte RAM, and 3.8 GHz CPU).

Photomicrographs in this thesis are represented in a format shown in Figure 3.11. The figure includes the code of the lithology sampled, sample name, and the scale of the sample. Also, minerals are abbreviated consistent with the abbreviation system by Whitney and Evans (2010). The abbreviations are recognised by the International Mineralogical Association (IMA), and supplement those drafted by Kretz (1983). The list of mineral abbreviations used in the thesis is attached in Table A3.8, Appendix III.

4.0 LITHOLOGIES AND STRATIGRAPHY

4.1. Introduction

Lithological mapping and petrographic analyses show that the study area is almost wholly composed of metamorphic rocks (Figure 4.1.). The rocks include the basement granite and supracrustal rocks that belong to the Dominion Group and the Witwatersrand and Ventersdorp supergroups, as well as multiple igneous bodies, mafic sills, and dykes (Figure 4.2.). Previous studies have linked the mafic sills to the 2.72 Ga Klipriviersberg Group and the 2.06 Ga Bushveld Complex (c.f., Pybus, 1995; Bisschoff, 1982; Coetzee et al., 2006; De Waal, 2008); however the sill generations are not distinguished in this study as a full geochemical analysis was not performed. The stratigraphic unit codes used in this thesis for lithologies accords with the code system used by Bisschoff (2000) (Figure 4.2). Outcrop descriptions of the units are coupled with brief petrographic descriptions that include types of mineral assemblages hosted and their proportions. Detailed petrographic descriptions of the lithologies are discussed in the metamorphic grade distribution chapter (Chapter 6).

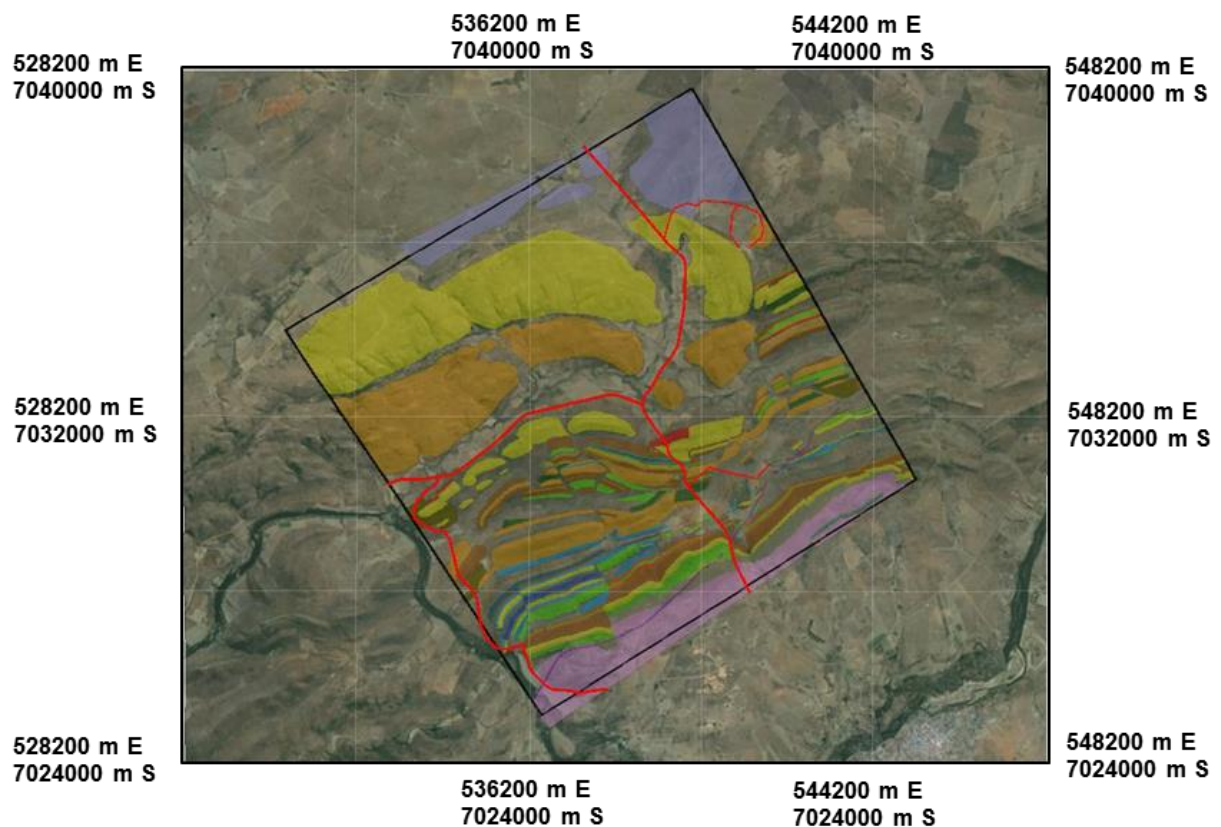


Figure 4.1. Google EarthTM image showing the 10×11 km² study area in the northern collar of the Vredefort Dome, with the main lithologies included.

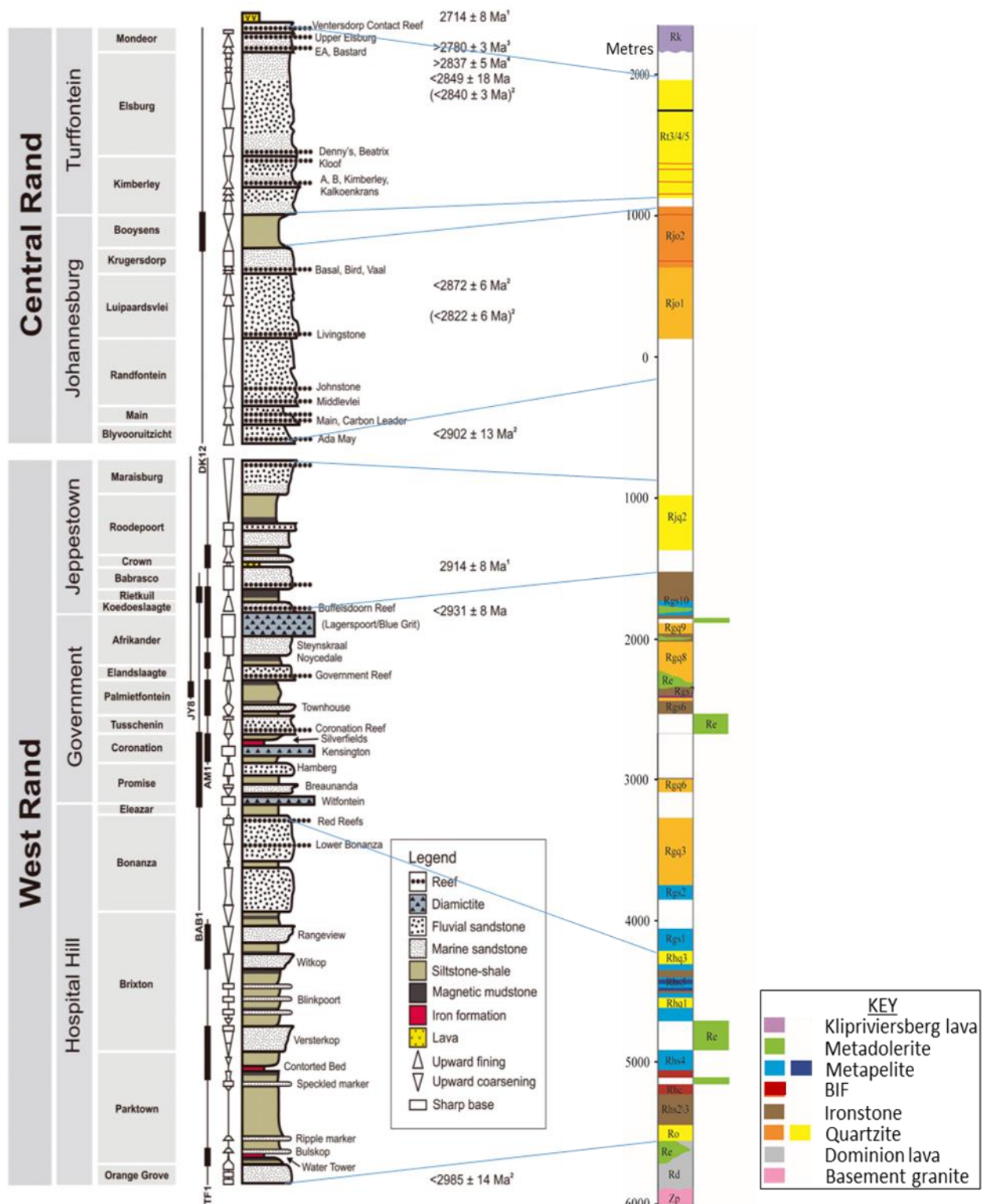


Figure 4.2. General stratigraphic column of the northern collar of the Vredefort Dome (right), compared with the stratigraphy of the Witwatersrand Supergroup by Guy et al. (2010) (left). Symbols used in the right-hand column are explained in the text.

4.2. Stratigraphy

4.2.1. *The Basement Granite*

The basement granite is the oldest rock type in the area, and forms tors located in the relatively flat core of the Dome (Figure 4.3.). The basement is reported to be migmatitic and gneissic further south of the study area (e.g., Lana et al., 2003a), but is made up of massive, locally weakly foliated, granite in the mapping area. The granite is brown on weathered surfaces (e.g., Figure 4.4.) and peach to light pink on fresh surfaces. It is composed of 31% quartz, 36% K-feldspar, 25% plagioclase and 8% biotite (outcrop-scale proportions). The granite is coarse-grained with interlocking crystals up to 6 mm in size; however megacrystic feldspar up to 2 cm occurs locally. It is crosscut by irregular pseudotachylitic breccia, quartz veins (Figure 4.4), and the Vredefort Granophyre (Figure 3.2).

4.2.2. *Dominion Group meta-basalt*

The extrusive rocks of the Dominion Group crop out as the southern pediment of the Orange Grove Formation (Ro) quartzite ridge (Figure 3.2), separating the quartzite ridge from the granite basement. Macroscopically, the Dominion Group is composed of locally amygdaloidal meta-basalt, which overlies an erosional contact with the older granite to the south. The meta-basalt is up to 200 m thick in the study area when assumed to be subparallel to the Hospital Hill Subgroup, and has quartz amygdales typically 2-3 mm in length, but up to 1 cm (Figure 4.5.). The meta-basalt is brownish-black on weathered surfaces and greenish-grey on fresh surfaces. It looks superficially similar to the matrix of the Vredefort Granophyre; however, it is more mafic and amygdales and lithic clasts hosted in the meta-basalt and granophyre, respectively, distinguish the two.

Petrographically, the meta-basalt (sample LV39; Table A3.4, Appendix III) is composed of 63% hornblende, 25% plagioclase, 8% quartz, and 4% ilmenite. Quartz crystals in amygdales are coarser-grained than those in the fine-grained matrix (Figure 4.6). Hornblende is generally hosted as radiating skeletal crystals and coarser irregular porphyroblasts up to 700 μm in length, which are coarser around quartz amygdales (Figures 4.6a, b respectively). The metamorphic minerals form a texture that is coarser than primary textures of basalt, which also indicates the metamorphic nature of the meta-basalts (Chapter 6).

4.2.3. *Witwatersrand Supergroup*

Most of rocks in the study area can be assigned to the Witwatersrand Supergroup. Outcrops are dominated by quartzite units that are intercalated with metamorphosed ironstone, banded iron-formation (BIF), pelite, and meta-diamictite units in the West Rand Group, and slate and meta-conglomerate units in the Central Rand Group (Figures 3.2, 4.2).

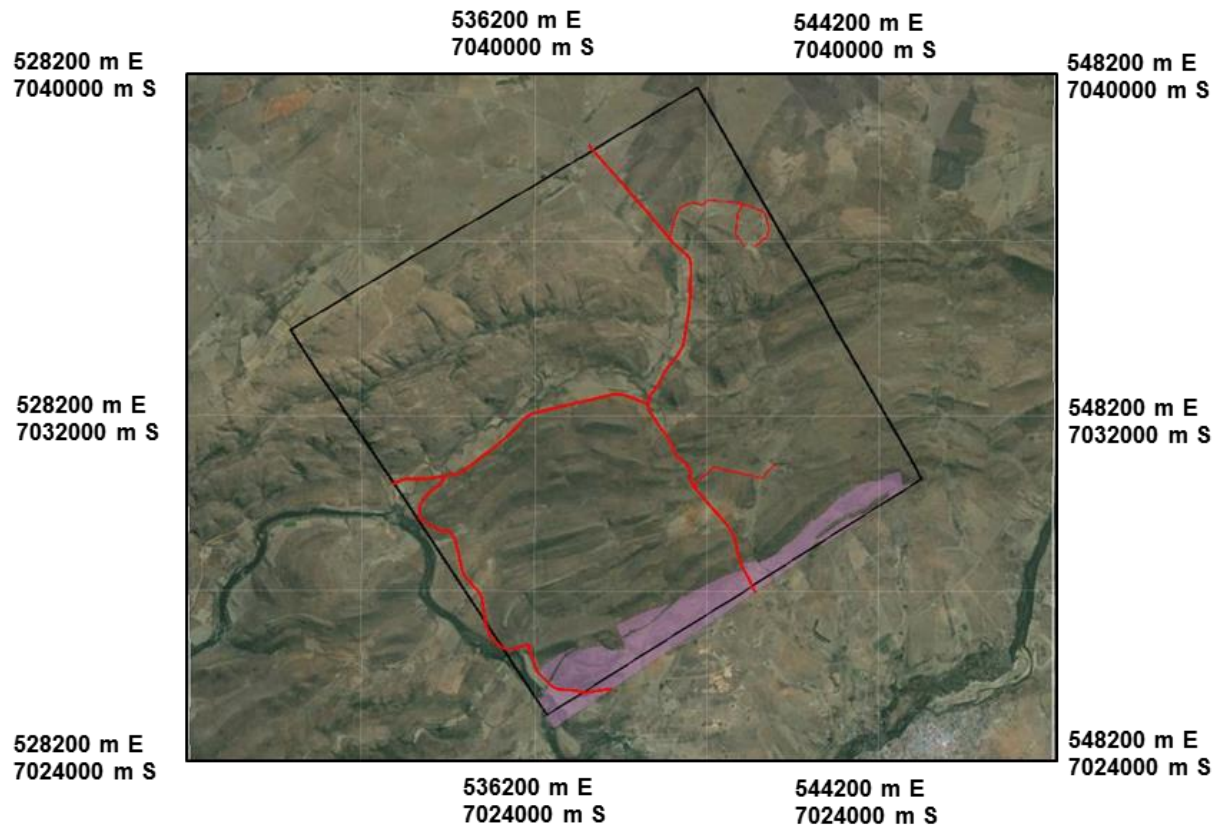


Figure 4.3. Google Earth™ image showing outcrop distribution of the basement granite.



Figure 4.4. Archaean basement granite crosscut by a ~1.5 cm wide quartz vein (35 J 536623 m E, 7026395 m S). The granite does not host foliation like other basement rocks described in the core of the dome.



Figure 4.5. Amygdaloidal meta-basalt belonging to the Dominion Group (35 J 539605 m E, 7028142 m S).

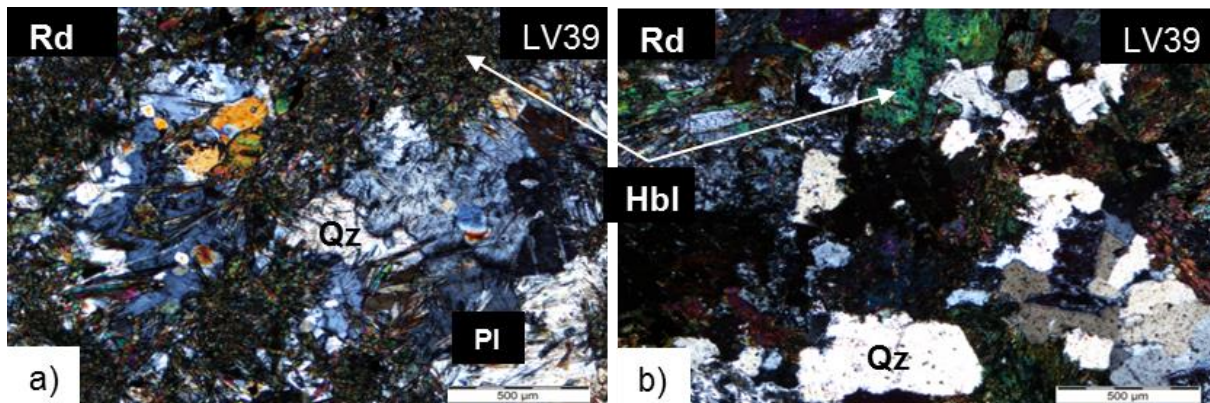


Figure 4.6. Photomicrographs of Dominion Group meta-basalt with a hornblende-plagioclase-quartz-ilmenite assemblage. (a) Skeletal hornblende hosted around quartz crystals. (b) Coarser-grained and interstitial hornblende (both XPL).

4.2.3.1. Quartzite

Quartzite is the most competent rock in the study area, and underlies most of the ridges of the collar (Figure 4.7). It is partly this character that makes quartzite ideal for recording structures, especially in rocks as old as the Witwatersrand Supergroup. In general, the thickness of the quartzite units, grain size, and compositional immaturity increase with younging, towards the north. The quartzite units show different colours on weathered surfaces (cream-white, grey, and brown), but are generally cream-white to light-brown on fresh surfaces of the West Rand and Central Rand quartzite units, respectively (e.g., Figure 4.8).

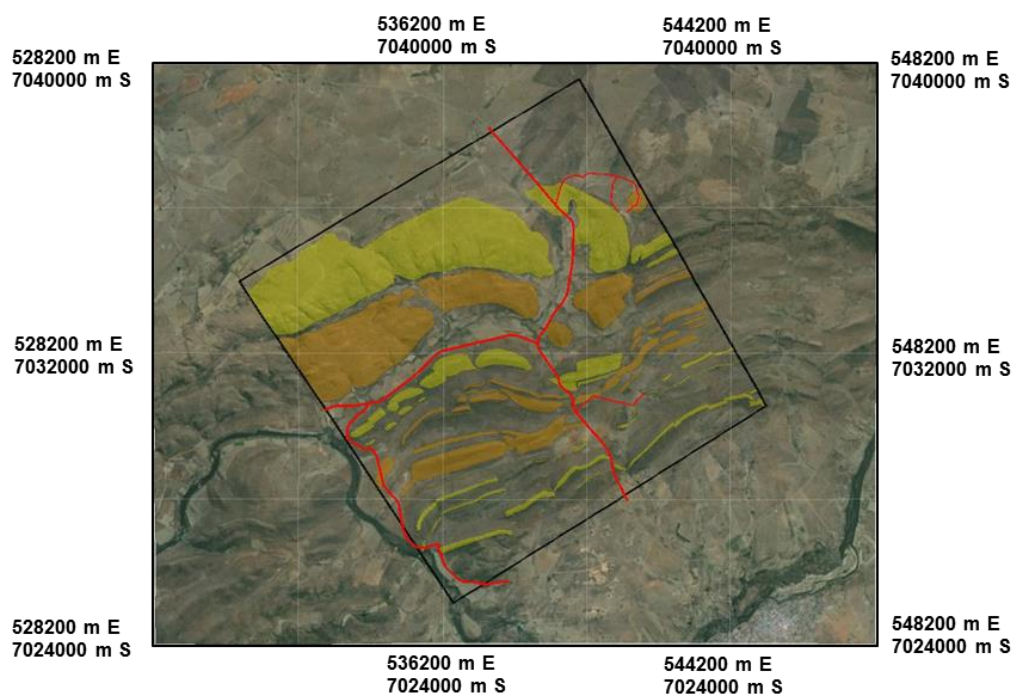


Figure 4.7. Quartzite units of the Witwatersrand Supergroup, showing an increase in thickness towards the north. Alternating yellow and orange colours represent the different subgroups of the Witwatersrand Supergroup.

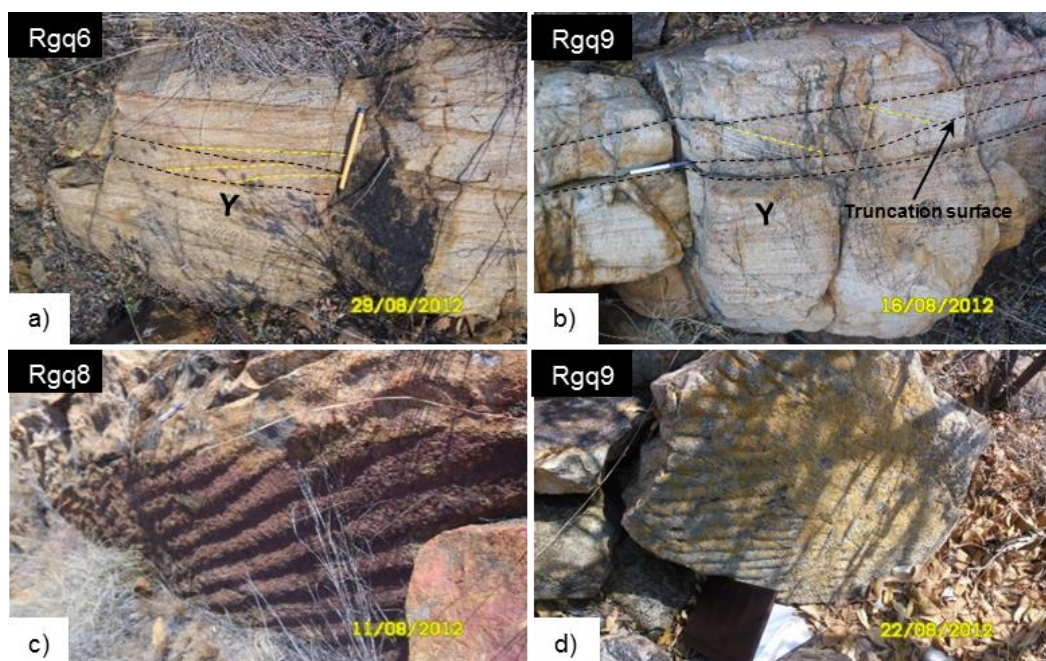


Figure 4.8. (a) & (b) Strike-subparallel sections showing trough and planar cross-bedding, respectively. Dip is towards the south (top of photographs) and younging to the north, indicative of overturned strata (35 J 538297 m E, 7029725 m S; and 35 J 537671 m E, 7031248 m S, respectively). (c) & (d) Symmetric straight to sinuous ripple marks developed in Rgq8 and Rgq9 quartzite units (35 J 541203 m E, 7031045 m S; and 35 J 537241 m E, 7031179 m S, respectively).

The Central Rand Group quartzite units are thicker, coarser and less pure (i.e., immature) than quartzite units in the West Rand Group. Quartzite units of the Central Rand Group are divided by the poorly exposed Booyens Formation slate into the Johannesburg and Turffontein groups. The units show a general upward-coarsening sequence (Figure 4.9), which is consistent with the upward change in grain size of facies in the Turffontein Subgroup (slate and meta-siltstone to quartzite and pebbly quartzite up-stratigraphy) as described by Holland (1990) and Holland et al. (1990). The quartzite units host poorly-developed planar and trough cross-beds relative to the West Rand Group (Figure 4.9). Other sedimentary structures described by Holland (1990) and Holland et al. (1990) include planar lamination and trough cross-lamination, which are less common. The immature nature of quartzite units of the Central Rand Group is shown by the presence of “knotted quartzite” horizons (Bisschoff, 1982), which host centimetre-scale aluminosilicate porphyroblasts. Porphyroblasts include kyanite, andalusite, and chloritoid developed along bedding surfaces and cleaved horizons (Figure 4.12). These horizons are generally brown on both weathered and fresh surfaces.

Quartzite units of the West Rand Group proximal to the core of the dome exhibit well-developed sedimentary structures. Most of these quartzite units are trough cross-bedded, with subordinate planar and herringbone cross-bedding, and ripple marks developed in a number of units (Figures 4.8, 4.9). Primary sedimentary features such as trough cross-bedding and truncated surfaces show that the strata are largely overturned; they stratigraphically young to the north (Figure 4.9). Quartzite units hosted in the Government Subgroup are commonly intercalated with thin pelitic horizons. The scale of intercalation ranges from metre- to millimetre-scale, and cross-laminations are associated with oxidised heavy minerals (Figure 4.10). The immature nature of the quartzite units is well shown by the Rgg8 argillaceous quartzite, which has a quartz content <75% (Table A4.2). Petrographically, the matrix of samples AQ11 and AQ57 of the quartzite is fine-grained and hosts ~25% sericite, biotite, and chlorite. The laminations host magnetite associated with maghemite and pyrite, which form ~4% of the sample (Figure 4.11). This unit is a distinctive marker horizon, which was used in field mapping to estimate the magnitude of displacement of crosscutting faults (Chapter 5).

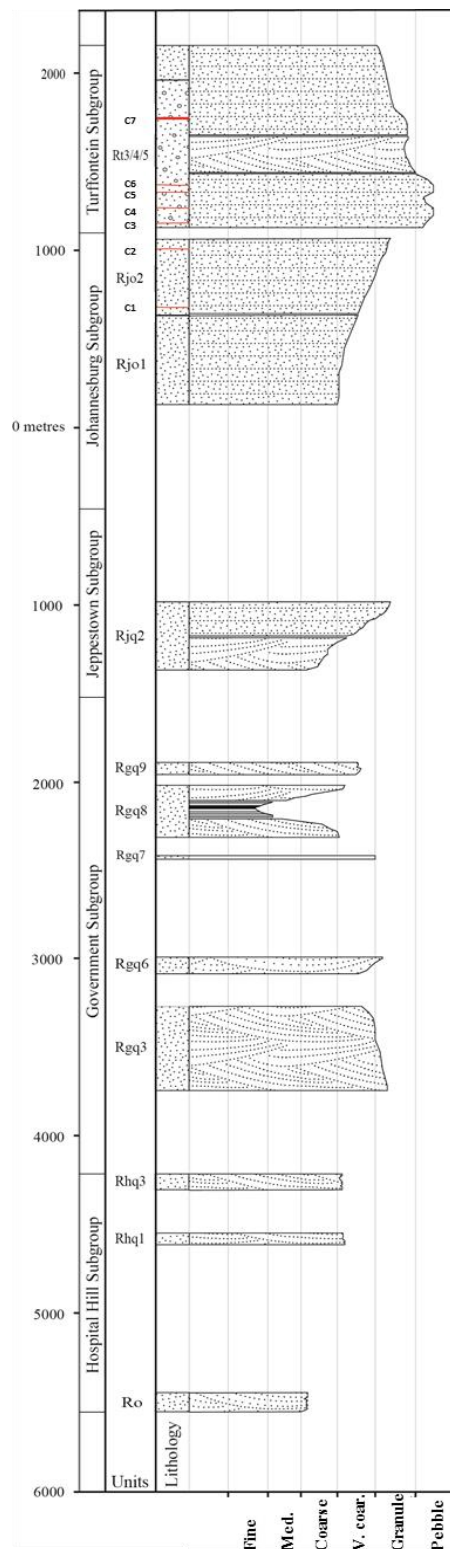


Figure 4.9. Stratigraphic log showing the sedimentology of quartzite units of the Witwatersrand Supergroup (fine < 0.25 mm, medium = 0.25–5 mm, coarse = 0.5–1 mm, v. coarse = 1–2 mm, granule = 2–4 mm, and pebble = 4–64 mm; classification by Wentworth, 1922).

The **Rt** quartzite unit is generally granular, with pebbly quartzite horizons that are interbedded with meta-conglomerate. The Rt units fine upwards into very coarse-grained quartzite. Rt units host the most number conglomeritic units (C3–C7 in Table 3). The meta-conglomerate units are generally matrix-supported and host clasts up to 7 cm in size (Subsection 4.2.3 VII)

Rjo1 and **Rjo2** quartzite units are very coarse-grained to granular, showing an upward-coarsening pattern. Rjo2 hosts meta-conglomerate units C1 and C2. Both units are massive and have poorly developed sedimentary structures.

Rjq2 quartzite unit is coarse-grained with well-developed laminations and trough cross-beds. It coarsens upwards into a granular and massive unit with poorly-developed sedimentary structures and a 50 cm-thick meta-conglomerate unit.

The **Rgq9** quartzite is very coarse-grained and hosts symmetric ripple marks (Figure 4.8d). The **Rgq8** quartzite unit is very coarse-grained and host symmetric ripple marks, and trough and herringbone crossbeds (Figure 4.8c). Rgq8 quartzite hosts a well-laminated argillaceous quartzite (Figure 4.10). It also hosts well-developed trough and planar crossbeds.

The **Rgq8** quartzite unit is very coarse-grained and host symmetric ripple marks, and trough and herringbone crossbeds (Figure 4.8c). Rgq8 quartzite hosts a well-laminated argillaceous quartzite (Figure 4.10).

Rgq7 and **Rgq6** quartzite units are very coarse-grained to granular, and host well-developed trough cross-beds.

Rgq3 quartzite unit is granular to very coarse-grained. It fines upwards, and hosts well-preserved trough and planar cross-beds. This unit is cleaved and hosts phyllitic horizons in Rondekop Farm (Figure 6.30).

Rhq3 and **Rhq1** quartzite units are very coarse-grained and host ripple marks. Rhq1 hosts planar cross-beds, whereas Rhq3 hosts herringbone and trough cross-beds with fuchsite along bedding planes.

Ro quartzite unit is coarse-grained, massive, and has poorly developed trough cross-beds. Thickness of beds is typically 45 cm.

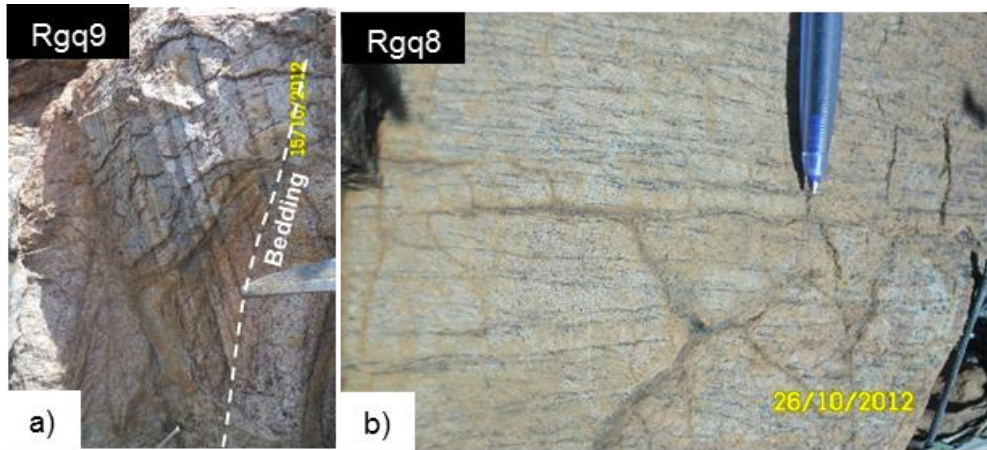


Figure 4.10. (a) Oxidised heavy minerals define planar crossbeds in Rhq9 (35 J 540239 m E, 7030900 m S). (b) Pelitic laminations in arkosic/argillaceous quartzite unit Rgq8 (35 J 541819 m E, 7033763 m S); see Figure 4.11 for photomicrographs.

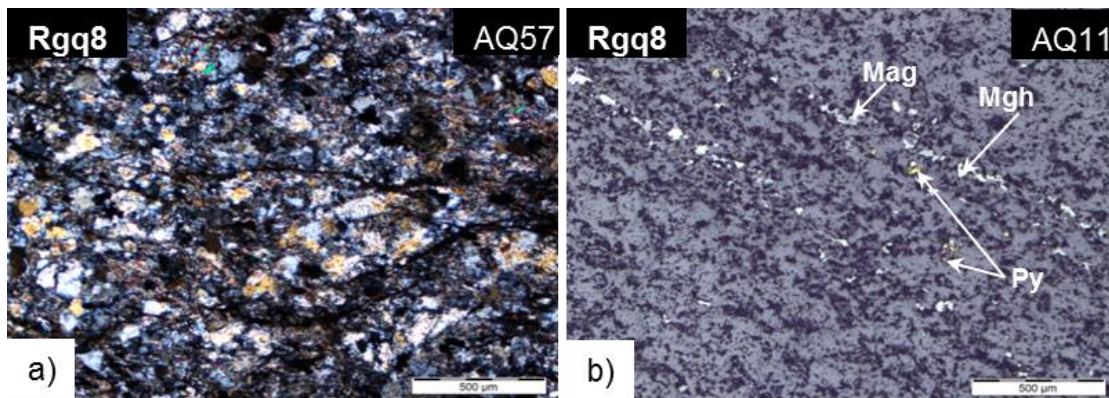


Figure 4.11. (a) Argillaceous quartzite Rgq8 composed of 69% quartz, 19% muscovite+sericite, 5% biotite, 3% chlorite, and 4% pyrite, magnetite and maghemite (XPL). (b) Laminations in argillaceous quartzite composed of magnetite (grey), maghemite (light grey), and pyrite (yellow; reflected light = RL).

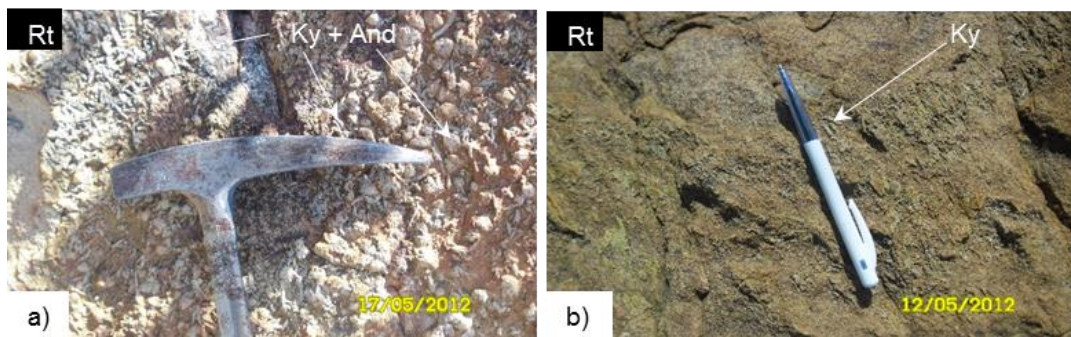


Figure 4.12. Knotted quartzite texture (a) & (b) Andalusite (idioblastic) and kyanite (subidioblastic) porphyroblasts developed on bedding planes of quartzite units of the Turffontein Subgroup (35 J 534666 m E, 7034055 m S; and 35 J 535798 m E, 7033959 m S, respectively).

4.2.3.2. Meta-Ironstone

Meta-ironstone units in the study area are highly weathered, resulting in formation of pediments and valleys between the quartzite ridges (Figure 4.13). There are 9 main meta-ironstone units in the area, all located in the West Rand Group (Figure 4.2). The meta-ironstone units commonly grade into Fe-rich slate; the slates are generally less magnetic than the meta-ironstones. The meta-diamictite units have a blocky to slabby appearance, and are dark-brown on weathered surfaces and blue-black/grey on fresh surfaces. Meta-ironstone units younger than the Rhs5 meta-ironstone unit (Figure 4.2) are less competent and show poorly developed bedding. The oldest meta-ironstone units (Rhs1 and Rhs2/3) host subidioblastic garnet porphyroblasts up to 4 mm in diameter (Figure 4.14). The Rhs2/3 meta-ironstone is well-bedded relative to the younger units, and grades up-stratigraphy into the Rhc BIF (Contorted Bed). The strength of magnetism of the Rhs2/3 meta-ironstone also increases towards the Contorted Bed.

The more massive Rhs2/3 meta-ironstone is composed of 50% grunerite, 25% quartz, 20% magnetite, and 5% garnet; thus, it is an amphibolite facies mineral assemblage (Mel'nik, 1982). Radiating aggregates of grunerite surround a nucleus of subidioblastic grunerite (Figure 4.15). Grunerite is distinguished from other amphiboles by twinning shown in sample MI18 (Table A4.3).

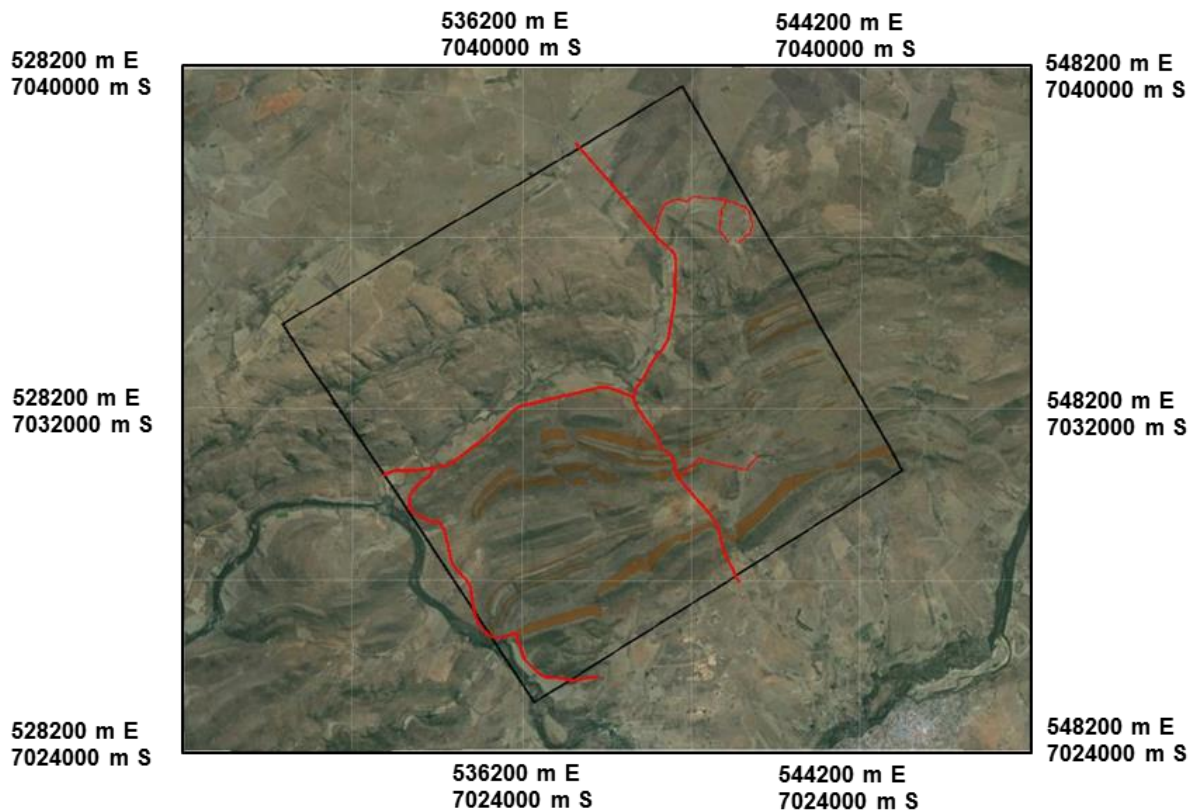


Figure 4.13. Google EarthTM image showing outcrop distribution of meta-ironstone units hosted in the West Rand Group.



Figure 4.14. Garnet-bearing Rhs2/3 meta-ironstone (35 J 536563 m E, 7026801 m S).

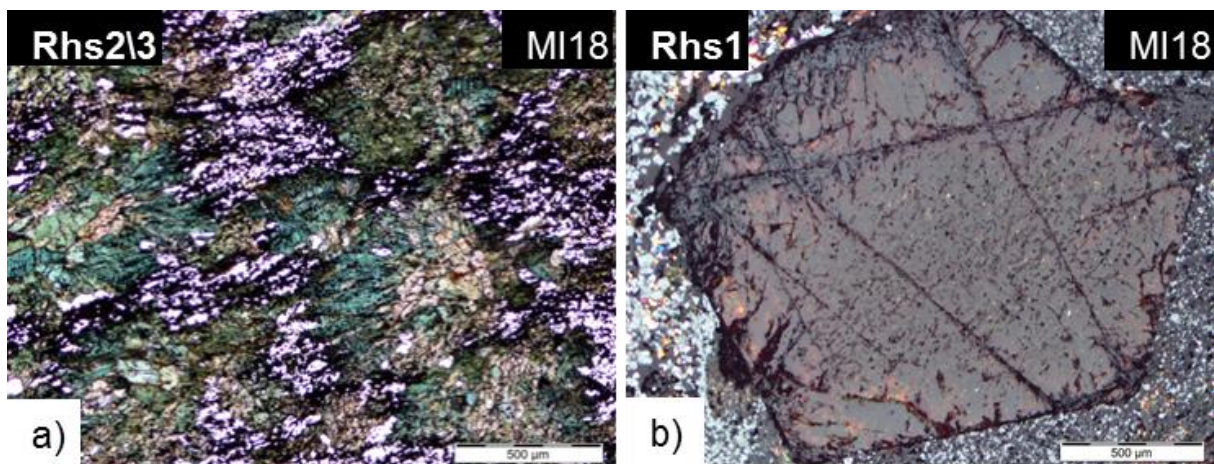


Figure 4.15. (a) Elongate grunerite in radiating or sheaf-like aggregates in the Rhs2/3 meta-ironstone (PPL). (b) Idioblastic garnet porphyroblast developed in the ~7m thick Rhs1 meta-ironstone located between the Ro quartzite units (reflected light=RL+XP).

4.2.3.3. Banded Iron Formation (BIF)

There is a single BIF unit located in the West Rand Group (Rhc, Contorted Bed). It forms a gradational contact with the Rhs2/3 meta-ironstone (Figure 4.16). It is magnetic, with typically 2 cm thick chert bands, which have been locally boudinaged and folded. The folds are predominantly tight to isoclinal and are Z-shaped, with a consistent right-lateral sense of shear and subvertical fold axial planes and fold hinge (Figures 4.17). The Rhc BIF is more resistant than the Rhs2/3 meta-ironstone, forming a small ridge relative to the meta-ironstone. This unit is generally about 90 m thick. It crops out well across the area and is a good marker horizon for detecting fault displacements.

Petrographically, the Rhc BIF is composed of 45% quartz, 35% magnetite, 10% grunerite, 6% biotite, and 4% garnet. Grunerite is developed interstitial to magnetite and quartz crystals (Figure 4.18). The magnetite-rich bands are intercalated with the quartz-rich chert bands, and are crosscut by perpendicular quartz veins.

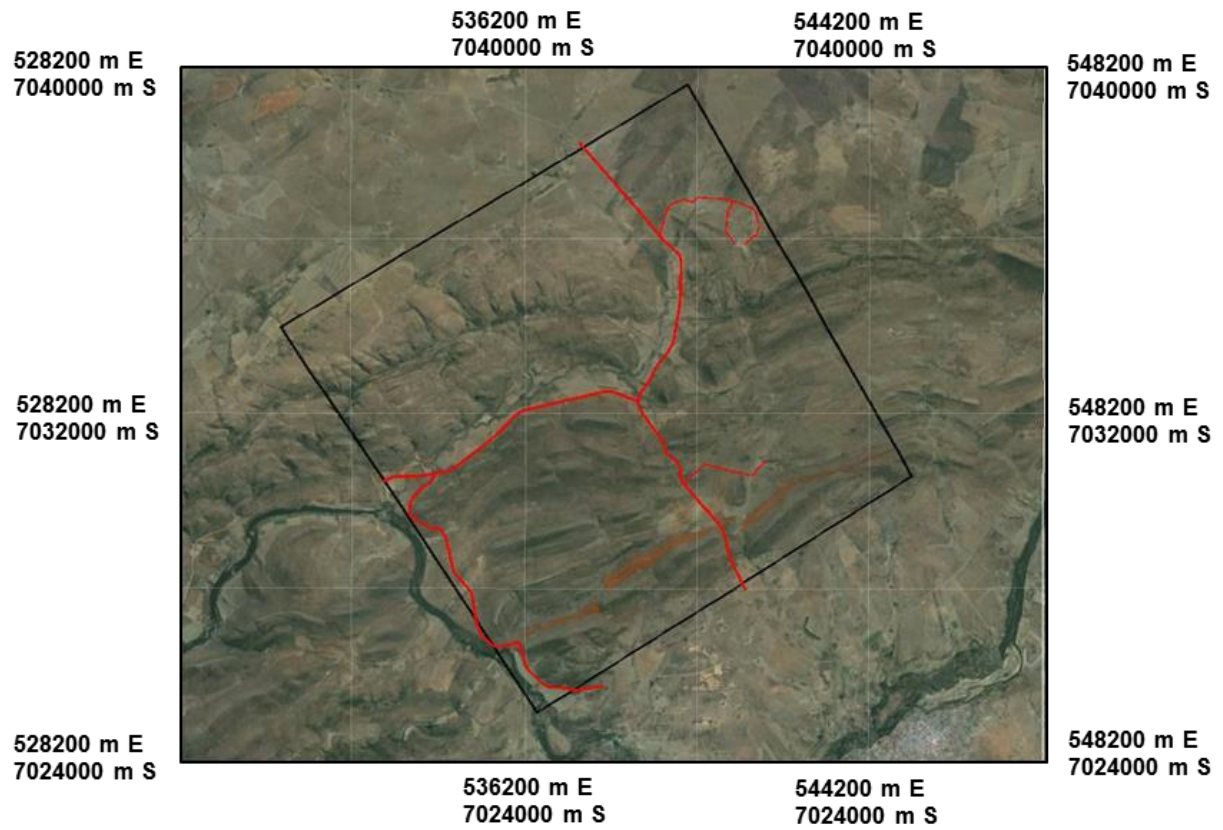


Figure 4.16. Google Earth TM image showing outcrop distribution of the Rhc Banded Iron Formation in the study area.

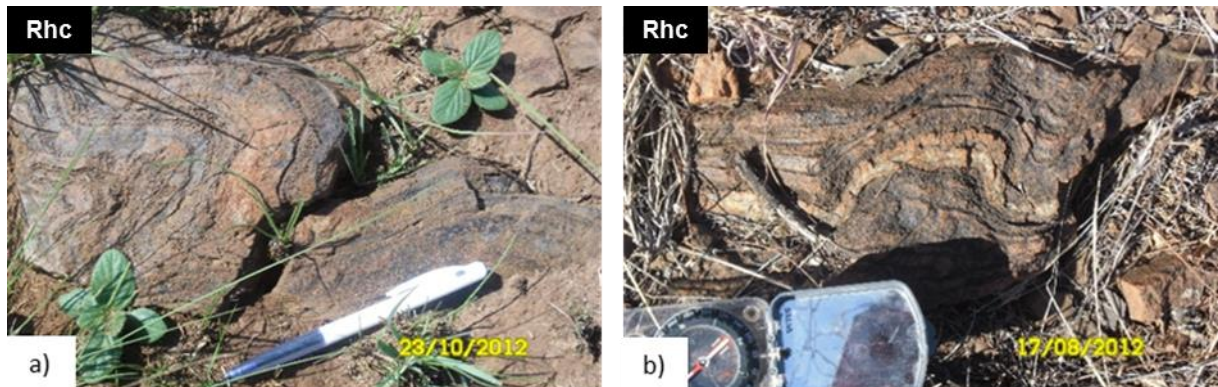


Figure 4.17. (a) Tight to isoclinal folds and (b) box folds developed in the Rhc BIF (35 J 538478 m E, 7028312 m S; and 35 J 538879 m E, 7028633 m S).

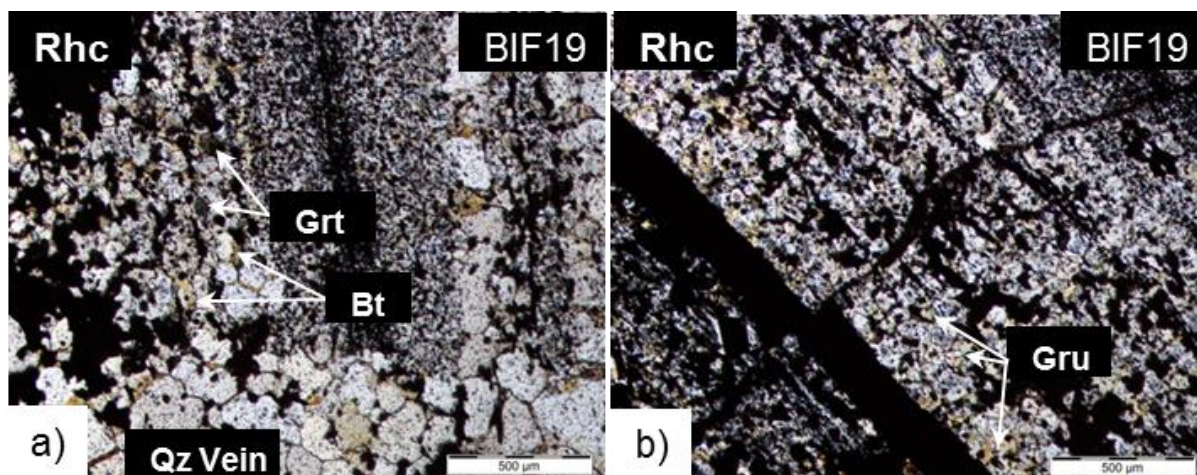


Figure 4.18. Photomicrographs of the Rhc Banded Iron Formation composed of 45% quartz, 35% magnetite, 10% grunerite, 6% biotite, and 4% garnet (PPL). Vertical bedding in (a) is cut by a quartz vein that is more coarsely-crystallised than the host rock.

4.2.3.4. Metapelite

This is the single most diverse rock type in the study area. There are 8 metapelite units interbedded with meta-ironstone and quartzite in the West Rand Group (Figure 4.19). Metapelitic units of the West Rand Group are brown to dark-grey on weathered surfaces and commonly form gradational contacts with meta-ironstone; some of them are, thus, magnetic (e.g., #27 Rhs 5 metapelite in Table 4.1). This gradational contact is also shown by transition from an Fe-rich assemblage (magnetite, quartz, grunerite, garnet in Rhc) to a relatively Al-rich assemblage (quartz, biotite, cordierite, staurolite, garnet in Rhs4), owing to the higher clay content in the primary composition of the pelites.

Metapelite units hosted in the Government and Hospital Hill subgroups can be distinguished according to sedimentary characteristics and metamorphic assemblages, as summarised in Table 4.1. The units are generally well-bedded, with the exclusion of Rhs4 (#2) and Rgs1. The metapelites show intercalation of pelitic beds with quartzite beds that are a few millimetres (e.g., #2 Rhs4) to several centimetres (#18 Rhs4) thick. Upward-fining sequences are exhibited in units that have well-developed quartzite intercalations, indicating sedimentary textures of a Bouma Sequence consistent with a turbidite environment (Figure 4.20). Depositional bases of Bouma Sequences, as well as trough cross-bedding developed in the quartzite intercalations, indicate the overturned nature of the metapelite units (Figure 4.20a). The units also show soft-sediment deformation features indicated by quartzite clasts (e.g., #27 Rhs5 in Table 4.1).

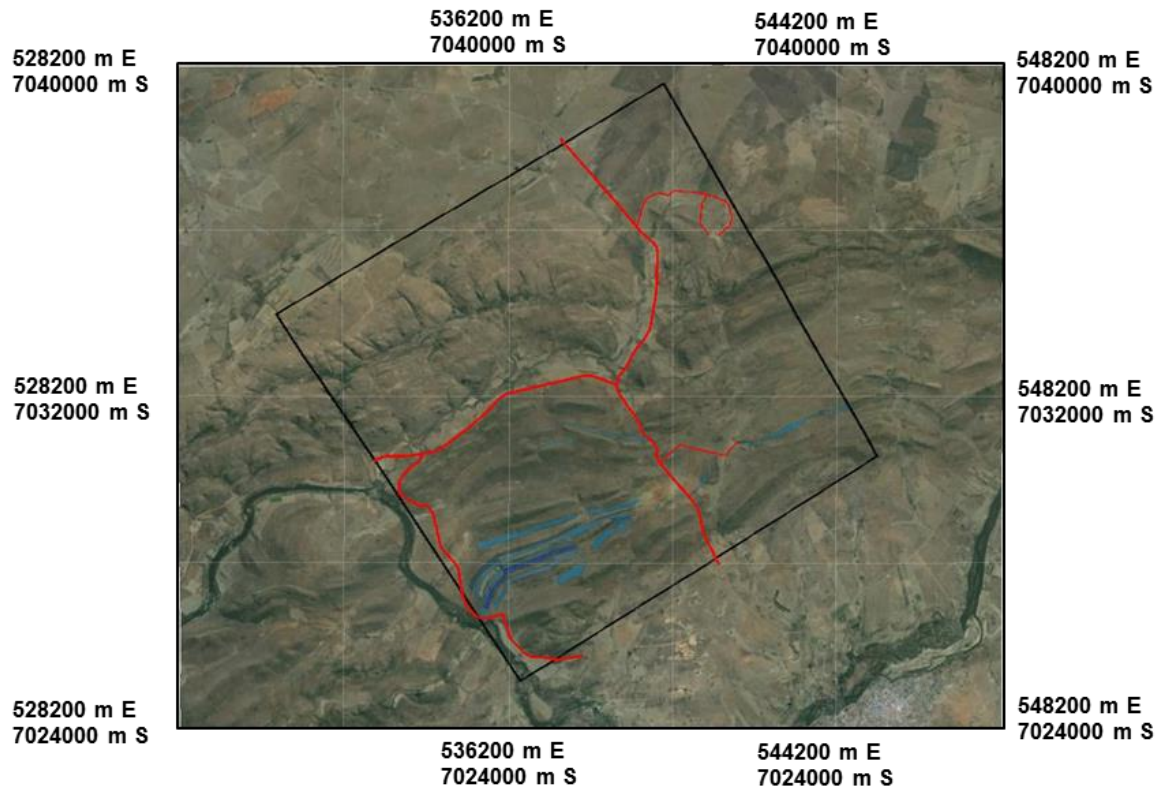


Figure 4.19. Google Earth™ image showing outcrop distribution of metapelite units in the study area.

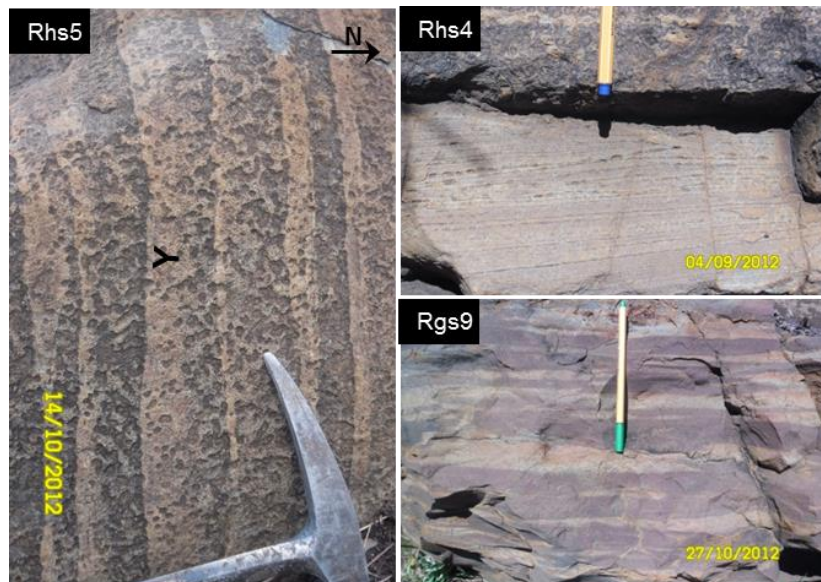


Figure 4.20. (a) Outcrop photograph of upward-fining layers indicative of parts of Bouma Sequences; which show northward younging (35 J 535834 m E, 7027367 m S). (b) Trough cross-bedding developed in quartzite intercalations of the Rhs4 metapelite (35 J 537588 m E, 7028137 m S). (c) Highly siliceous Rsg9 metapelite that does not exhibit mesoscale metamorphic assemblages (35 J 541951 m E, 7033947 m S).

Table 4.1. Metapelite units of the West Rand Group (oldest at the bottom).

Lithology Unit no.	Bisschoff's (2000) key	Sedimentary Features	Metamorphic features	Outcrop appearance
5	Rgs2	Well-bedded metapelite with quartzite intercalations typically 1.5 cm or less. The quartzite intercalations are boudinaged along bedding. It is mostly dominated by 1.2 cm-spaced fracture fabric and minor dextral micro-faults (35 J 539183 m E, 7029462 m S).	There are garnet porphyroblasts <1 mm. The metapelite is magnetic, which is due to the presence of secondary magnetite <0.5 mm. This unit is composed of 42% quartz, 34% biotite, 10% garnet, 9% magnetite, 5% hematite (Sample FR23).	
7	Rgs1	Poorly-bedded andalusite-biotite metapelite. It contains the most pelitic material relative to all other metapelites. Its arenite content increases towards the north (35 J 538988 m E, 7029241 m S).	Andalusite porphyroblasts are up to 5 cm long, and form ~25% of the rock at certain station points. Biotite porphyroblasts up to 3 mm long are hosted in the matrix and within andalusite porphyroblasts. The unit is composed of 42% biotite, 28% andalusite, 20% quartz, 06% garnet, 04% chlorite (Sample MP65).	
6	Rgs1	A well-bedded cordierite-biotite-andalusite metapelite. It has quartzite lenses that are typically 2 cm thick, closely intercalated with ~3 cm thick pelitic beds. It has similar sedimentary features to the #26 (Rhs5) metapelite unit (35 J 538679 m E, 7028805 m S).	Dominated by cordierite porphyroblasts typically 1 cm in diameter, and andalusite ranging from 0.6-3 cm, with an increase in size towards the Rgs2 andalusite metapelite to the north. There's also biotite typically 2 mm large. Cordierite has darker rims of biotite.	
29	Rhs5	Well-bedded cordierite metapelite. It has quartzite intercalations up to 5 cm thick, separated by pelitic beds down to 2 cm thick (35 J 535756 m E, 7027467 m S).	It hosts cordierite porphyroblasts typically 5 mm in diameter, and it is poor in andalusite. This unit is composed of 44% cordierite, 30% biotite, 21% quartz, 05% chlorite (Sample MP33).	

Table 4.1. Continued...

Lithology Unit no.	Bisschoff's (2000) key	Sedimentary Features	Metamorphic features	Outcrop appearance
28	Rhs5	Well-bedded cordierite metapelite. Quartzite interlaminationations are less than 4 mm in thickness, which increases towards the #29 metapelite to the north (35 J 535804 m E, 7027415 m S).	Cordierite porphyroblasts are typically 3 mm in diameter. The size increases towards the north.	
27	Rhs5	Poorly-bedded metapelite with quartzite intercalations that are boudinaged and isoclinally folded. The deformed quartzite intercalations are typically up to 4 cm thick. There is no consistency in the sense of displacement. This, together with the intense folded (rolled) quartzite blocks indicates soft-sediment deformation (35 J 535831 m E, 7027316 m S).	It has no visible porphyroblasts, and is magnetic. This unit is composed of 58% quartz, 13% biotite, 12% garnet, 9% chlorite, 8% magnetite (Sample MS30).	
26	Rhs5	Well-bedded cordierite-biotite-andalusite metapelite. It is distinguished by the closely spaced sandy and muddy intercalations, which form upward-fining sequences (Bouma sequences) (35 J 535820 m E, 7027266 m S).	Dominated by cordierite and biotite porphyroblasts typically 9-23 mm in diameter and 1.2 mm long respectively. There is minor andalusite up to 2.4 cm long. The pelitic beds are composed of 38% biotite, 30% cordierite, 22% quartz, 10% andalusite (Sample MP29).	
25	Rhs5	Poorly-bedded Fe-rich metapelite, similar to #27. It has ~3 cm thick discontinuous quartzite intercalations that are thinned and folded in a chaotic manner, reflecting soft sediment deformation. It is located on the northern gradational contact of a meta-ironstone (See Figure 3.3) (35 J 535818 m E, 7027174 m E).	It has no visible porphyroblasts, and is magnetic	

Table 4.1. Continued...

Lithology Unit no.	Bisschoff's (2000) key	Sedimentary Features	Metamorphic features	Outcrop appearance
24	Rhs5	Well-bedded cordierite metapelite. It has quartzite beds typically 8 cm thick intercalated with pelitic beds (35 J 535820 m E, 7027100 m S).	Dominated by cordierite porphyroblasts ~3 mm in diameter. There's minor andalusite porphyroblasts up to 1 cm long.	 12/10/2012
18	Rhs4	Well-bedded cordierite-andalusite metapelite. This unit has the thickest quartzite interbeds, up to 15 cm thick. The quartzite beds show trough cross-bedding indicative of overturned strata. It has thinner quartzite intercalations (≤ 1 cm) that are folded towards its southern contact, indicative of soft -sediment deformation (35 J 537814 m E, 7027887 m S).	It is dominated by cordierite and andalusite porphyroblasts typically 8 mm and 2.7 cm large respectively	 04/08/2012
2	Rhs4	Poorly-bedded Fe-rich cordierite-garnet metapelite. Evidence for sedimentary layering is represented by poorly-developed laminations. It grades northwards into #18 metapelite, with increasing andalusite content (35 J 538374 m E, 7028420 m S).	It hosts cordierite and garnet porphyroblasts typically 1.2 cm and 2 mm large respectively. This unit is composed of 32% quartz, 29% biotite, 25% cordierite, 9% staurolite, and 5% garnet (Sample MP23).	 13/10/2012

Different metamorphic mineral assemblages hosted in metapelite units reflect the different primary compositions of the units. Minerals visible in outcrop include garnet, cordierite, staurolite, biotite and andalusite porphyroblasts, which range from a few millimetres (garnet, biotite, and staurolite) up to 5 cm in size (andalusite). Generally, garnet and cordierite are more abundant in the southernmost, older units (e.g., #2); whereas andalusite is abundant in younger metapelite units (e.g., #7). However, a metapelite unit located along the southern contact of the Ro quartzite unit hosts andalusite (Samples S1 and S2 in Table A4.6). This makes the pre-Ro unit the oldest metapelite unit in the study area, rather than metapelite unit Rhs4 as suggested by Bisschoff (2000).

Petrographic analysis of the metapelite units was performed on only a limited number of six representative samples (Table A4.2), because some units are similar (e.g., #25 and #27 in Table 4.1). All the samples host biotite, which ranges from 13-42%, in association with varying combinations of cordierite, andalusite, garnet, staurolite, muscovite, chlorite, ilmenite, magnetite, and quartz (Figure 4.21). Cordierite (25-44%) is restricted to the Rgs1 metapelite and older units towards the south, whereas garnet (5-10%) is hosted across the entire range of metapelite units (Rhs4-Rgs2; Table 4.1). Magnetite, hematite, and ilmenite are hosted in Fe-rich and magnetic metapelite units (Table 4.1, Figure 3.5). Garnet, cordierite, and chloritoid porphyroblasts are associated with a syn-metamorphic foliation, a relationship which has important implications for the deformational history of the collar (Section 6.2).

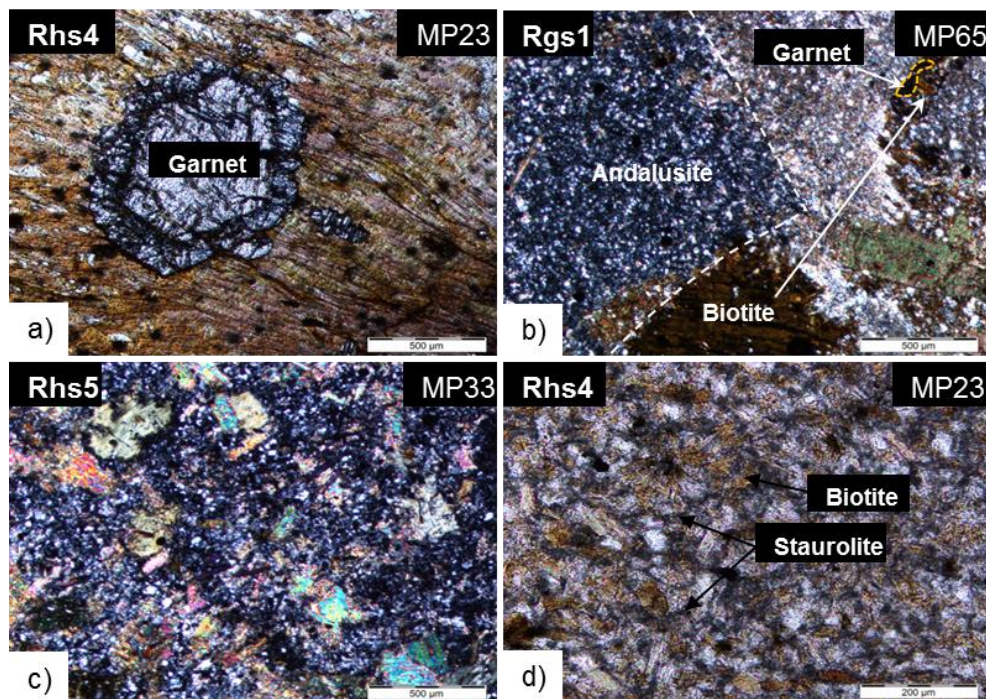


Figure 4.21. Photomicrographs of metapelite units hosted in the lower West Rand Group: (a) A zoned garnet porphyroblast associated with a cleavage (PPL); (b) andalusite-rich Rgs1 metapelite unit (XPL); (c) Rhs5 matrix dominated by biotite, chlorite, and quartz (XPL); (d) biotite and staurolite replacements in cordierite (PPL; Section 6.2).

4.2.3.5. *Meta-diamictite*

The Government Subgroup hosts two meta-diamictites in the Rgs7 and Rgs3 units (Figure 3.2), with the former as the more extensive unit in the study area. The units are similar to meta-diamictite units described by McCarthy (2006) in the Government Subgroup. The meta-diamictite units are up to 20 m thick and generally form poor outcrops. Outcrops of the Rgs3 and Rgs7 meta-diamictite units are both dark grey and host subrounded quartzite clasts that have a generally uniform size ranging from a few millimetres to 3 cm in diameter (Figure 4.22). These units are also important fault displacement markers because they are distinctive.

Although no outcrops were found, boulders of meta-diamictite were also observed proximal to the Kimberley Reefs of the Central Rand Group. The boulders are possibly part of meta-diamictite units intercalated with the Kimberley Reefs as described by Martin et al. (1989) and Beukes and Nelson (1995).

Petrographically, the Rgs7 meta-diamictite unit is composed of approximately 55% quartz, 20% chlorite, 17% biotite, and 8% sericite and muscovite (Figure 4.23a; Samples MR347 and MR402 in Table A2.2). The matrix of the samples is dominated by biotite and skeletal chlorite typically <150 μm (Figure 4.23).



Figure 4.22. *Meta-diamictite unit with a generally uniform size of quartzite clasts, developed in the Government Subgroup (35 J 538090 m E, 7030560 m S).*

4.2.3.6. Slate

Slate units are pelitic outcrops that have a shale-like appearance, but thin sections show evidence of cleavage. Slate units are hosted in both the West Rand and Central Rand groups. These units are weathered, except where they are interbedded with meta-ironstone in the West Rand Group (e.g. Rgs9 unit). The Turffontein Subgroup of the Central Rand Group hosts a thin slate unit up to 10 m thick within the Kimberley Reefs. This unit is weathered and presents varying colours (yellowish-brown to greenish-grey). The slate unit was used to determine offset along faults cross-cutting the Central Rand Group.

Samples of the pelitic units were collected for petrography along the Rgs9, Rjs2, and pelitic units associated with the Kimberley Reefs (Samples S08-S356 in Table A4.2). Most of these units are weathered; as a result metamorphic mineral textures are not well preserved. Relatively well-preserved samples from the Rgs9 unit show that the slate is composed of about 55% fine-grained quartz (<100 μm), 10% oxidised skeletal lath-shaped porphyroblasts, and 35% fine-grained matrix predominantly composed of sericite $\pm$ chlorite (Figure 4.24). The typically 200 μm skeletal and oxidised porphyroblasts resemble the chloritoid seen in the knotted quartzite horizons (Section 6.2).

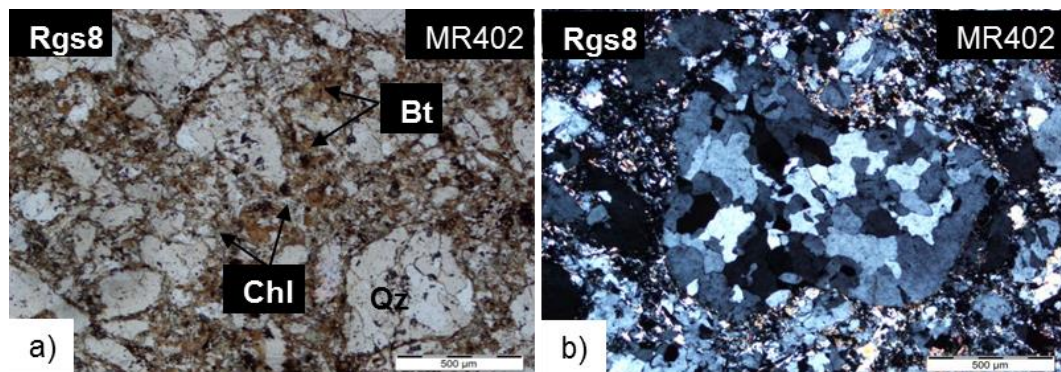


Figure 4.23. Photomicrographs of the Rgs8 meta-diamictite unit composed of: (a) biotite-chlorite-muscovite matrix, and subangular polycrystalline quartz clast (PPL). (b) Close-up of polycrystalline quartz clast (XPL).

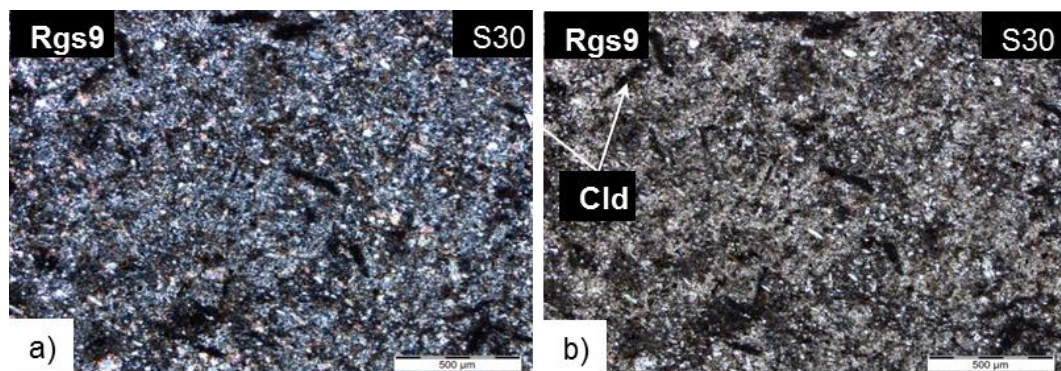


Figure 4.24. Skeletal oxidised porphyroblasts, interpreted as chloritoid hosted in a fine-grained matrix of quartz, sericite, and chlorite matrix of slate. (XPL and PPL respectively).

4.2.3.7. Meta-conglomerate

All mapped meta-conglomerate units belong to the Central Rand Group. The meta-conglomerate units show an increase in clast proportion and size up-stratigraphy, and are generally oligomictic with vein quartz as the main clast component; the exceptions are the C4 and C7 meta-conglomerates that have slate and chert clast components, respectively (Table 4.2).

The oldest meta-conglomerates in the study area are the relatively thin oligomictic units of the Johannesburg Supergroup (C1 and C2, Table 4.2), which are typically 0.6-1.2 m thick. They host quartz clasts up to 3 cm in diameter. The clast size increases from 3 cm in Johannesburg Subgroup, to 7 cm in the Kimberley reefs (Turffontein Subgroup). The C3-C7 conglomeritic units hosted in the Turffontein Subgroup are up to 10 m thick, and are matrix or clast-supported (Table 4.2). These units are useful marker horizons for estimating offset along faults that crosscut the thick massive quartzite units of the Central Rand Group.

The C4 meta-conglomerate was used as a marker horizon to determine the offset along the Koedoesfontein Fault (Figure 3.4) because it is the only unit that hosts slate clasts (Table 4.2). Poorly developed trough cross-beds in the Kimberley Reefs and bedding surfaces of the C5 meta-conglomerate indicate that the rocks are overturned (Figure 4.25a, b). Bedding surfaces of the C5 unit exhibit depressions formed by clasts during their deposition, which indicate that the sequence is inverted. The lack of evidence of deformation along the meta-conglomerate unit indicates that the bedding depression is most likely a primary sedimentary feature.

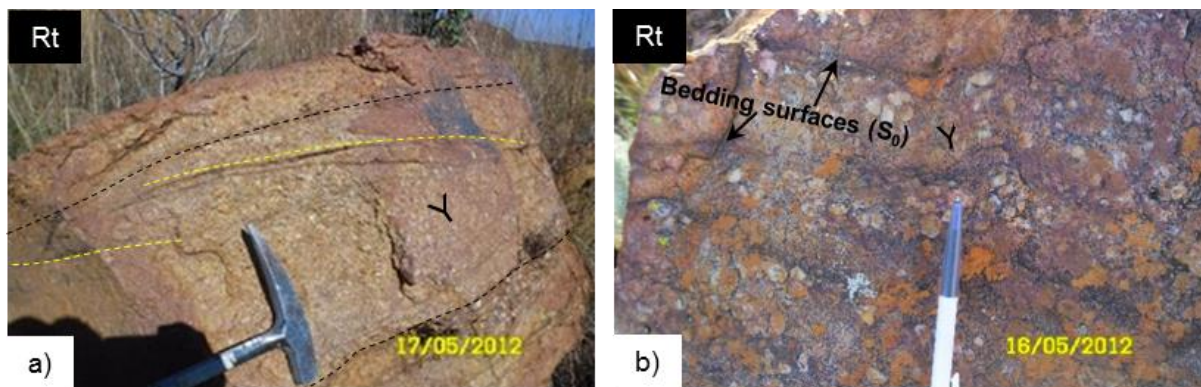


Figure 4.25. (a) & (b) Cross-sections of one of the Kimberley Reefs and the C5 meta-conglomerate, respectively showing trough cross-bedding and sedimentary depressions indicative of overturned strata (35 J 534304 m E, 7034176 m S, and 35 J 535286 m E, 7034394 m S, respectively).

Table 4.2. Meta-conglomerate units of the Central Rand Group (oldest at the bottom).

Lithology Unit no.	Sedimentary Features	Outcrop appearance
C7	A 10 m thick matrix-supported unit, with clast size ranging from 0.8-12 cm. It is immature with angular clasts in a coarse-grained (~4 mm) matrix. Clast populations are quartz, quartzite, and chert (35 J 534742 m E, 7034958 m S).	
C6	Clast-supported oligomictic conglomerate unit is divided into 3 units of 0.8 m total thickness. These units are separated by coarse-grained pebbly quartzite. The clasts are subrounded with a typical clast size of 1.3 cm, but up to 2.8 cm (35 J 535242 m E, 7034479 m S).	
C5	0.72 m thick matrix-supported oligomictic conglomerate. Unlike the other units, it is well-bedded and has clasts that are generally 1.2 cm in diameter, but up to 2 cm. Bedding indicates overturned strata (35 J 535286 m E, 7034394 m S).	
C4	Clast-supported polymictic conglomerate (1.2 m thick). It hosts poorly-sorted subrounded clasts that are 1.5-3.7 cm large. It is composed of quartz, siltstone, and shale clast populations (35 J 534322 m E, 7034404 m S).	
C3	Clast-supported, 1 m thick oligomictic conglomerate unit. Clasts are typically 1.4 cm large, and up to 4 cm. It has a coarse-grained quartzite matrix typically 4 mm in grain size (35 J 535405 m E, 7034071 m S).	

Table 4.2. Continued...

Lithology Unit no.	Sedimentary Features	Outcrop appearance
Kimberley Reefs	These are mostly clast-supported oligomictic conglomerate units. There are 11 units in total, with generally subrounded clasts 1.2-7 cm in size. One of the units has poorly developed trough cross-beds indicative of overturned strata (35 J 534336 m E, 7034170 m S).	 A photograph showing a clast-supported oligomictic conglomerate outcrop. The clasts are subrounded and vary in size. A blue tool is placed against the rock for scale. The date 17/05/2012 is visible in the bottom right corner.
C2	Matrix-supported, 1.2 m thick oligomictic conglomerate. It hosts subrounded clasts typically 0.5-3 cm in diameter. It is associated with up to 4 thinner units, with thickness ranging from 2-7 cm. The matrix has a general grain size of 3 mm (35 J 535941 m E, 7034071 m S).	 A photograph showing a matrix-supported oligomictic conglomerate outcrop. The matrix is fine-grained, and the clasts are subrounded. A red string is placed against the rock for scale. The date 11/05/2012 is visible in the bottom right corner.
C1	C1 is composed of two clast-supported oligomictic conglomerate units up to 0.6 m thick. It hosts relatively well-sorted quartz clasts 2-3 cm large, in a coarse-grained (~3.5 mm) quartzite matrix (35 J 536459 m E, 7033590 m S).	 A photograph showing a clast-supported oligomictic conglomerate outcrop. The clasts are well-sorted and subrounded. A blue tool is placed against the rock for scale. The date 18/05/2012 is visible in the bottom right corner.

4.2.4. Ventersdorp Supergroup

Meta-dolerite sills and amygdaloidal meta-basalt belonging to the Ventersdorp Supergroup (c.f., Bisschoff, 2000) form about 8% of the lithologies in the study area. The sills intruded broadly parallel to bedding planes of the Witwatersrand Supergroup, but are locally discordant or stepped, consistent with an en-echelon pattern (Figure 3.2). The meta-basalt crops out north of the Central Rand Group, whereas the sills mostly intruded the West Rand Group (Figure 4.26).

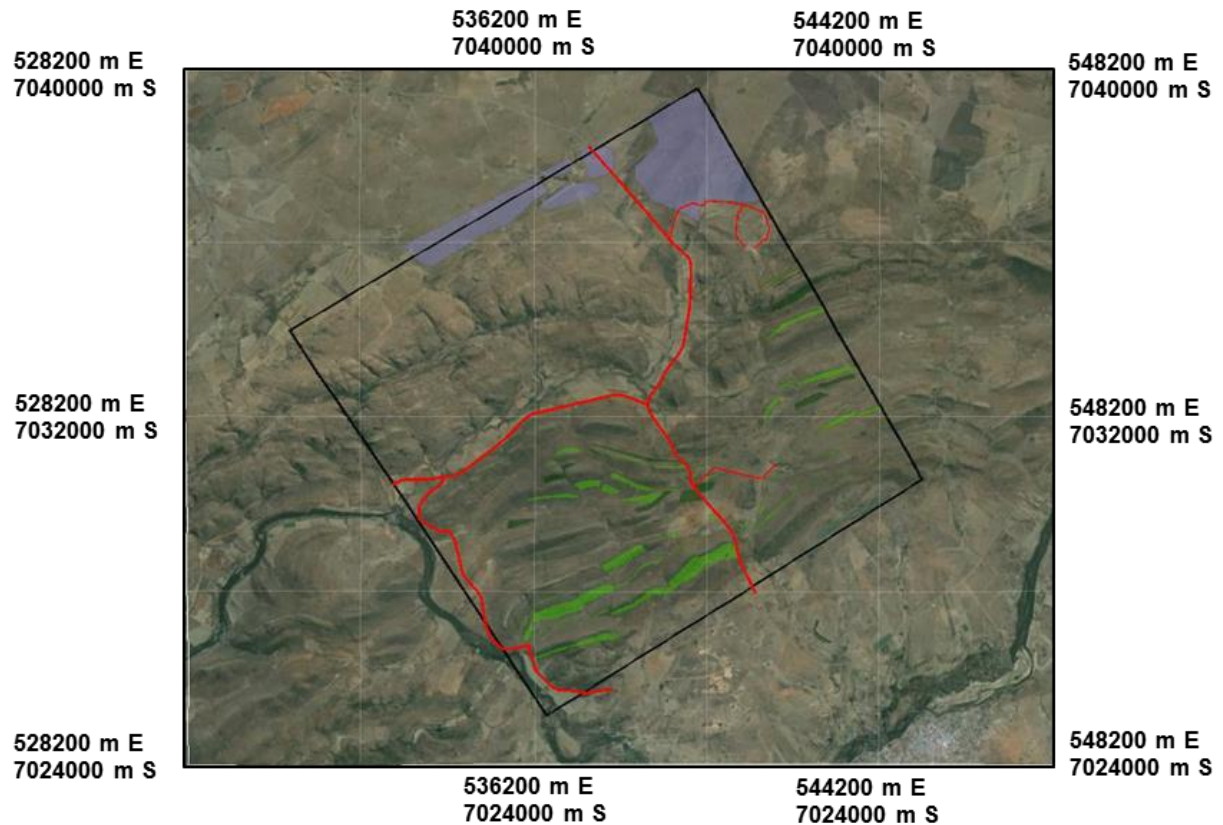


Figure 4.26. Google Earth™ image showing outcrop distribution of metabasalt (purple) and metadolerite/dolerite (green) sills of the Ventersdorp Supergroup.

4.2.4.1. Meta-dolerite sills

All meta-dolerite sills sampled in the study area have intruded the West Rand Group, except for one meta-dolerite sill hosted in the Johannesburg Subgroup (Central Rand Group). The meta-dolerite sills are 30-300 m in outcrop width, with a medium- to coarse-grained texture, and are dark grey (with millimetre-scale white plagioclase and black/green pyroxene/amphibole crystals) in colour. Meta-dolerite units that are located in the southern part of the study area host amphibole needles up to 15 cm long. In places the sills exhibit a coarse-grained texture in their interiors that is reminiscent of gabbro (Figures 4.27). Some sills that are intruded within the Jeppestown Subgroup do not host amphibole and retain an ophitic to subophitic igneous texture (Figure 4.28d). This may reflect preservation of primary textures owing to lack of water ingress during metamorphism and the lower metamorphic grade at greater radial distances (c.f., Gibson and Wallmach, 1995) (Table A4.4). However, Coetzee et al. (2006) also noted several Bushveld-age sills southwest of the study area in the upper West Rand Group, which indicates potentially two possible generations of sills including the dominant Ventersdorp-age sills. Geochemical analysis to distinguish the Bushveld-age sills was beyond the scope of this study.

Petrographically, the sills are composed of amphibole and/or pyroxene as the mafic minerals; they are thus meta-dolerite (Figure 4.28a-c) to dolerite (Figure 4.28d). These intrusions are composed of 0-85% hornblende and actinolite, 10-55% clinopyroxene, 7-42% plagioclase, and minor concentrations of biotite, epidote, chlorite, quartz, and ilmenite. Sample FMD 231 (Figure 4.28a) is the most metamorphosed sill sample with ~78% amphibole, whereas sample MD13 in the northern part of the study area shows primary igneous textures with virtually no amphibole (Table A3.4, Appendix III). The metamorphic assemblages show a generally northward increase in actinolite concentration (decreasing hornblende) and a northward decrease in the size of porphyroblasts (Compare sample locations in Figure A3.1 and Table A3.4, Appendix III).

4.2.4.2. Klipriviersberg Group meta-basalt

The Klipriviersberg Group amygdaloidal meta-basalt underlies the relatively flat topography in the far northern part of the study area (Figure 4.26). The meta-basalt is brownish-grey on a weathered surface and greenish-grey on a fresh surface, with quartz amygdales that are typically ≤ 1.5 cm in diameter (Figure 4.29). The meta-basalt is approximately 800 m thick if assumed to be subparallel to the orientation of the Central Rand Group, but is thicker in other parts of the Witwatersrand Basin (Section 2.1.3).

Photomicrographs show that the meta-basalt is highly altered; as a result mineral proportions are difficult to estimate (Figure 4.30). However there is evidence of clinopyroxene, plagioclase, ilmenite, actinolite, chlorite, and quartz in meta-basalt sample LV1 (Table A4.4). Quartz occurs both in amygdales and in the matrix.

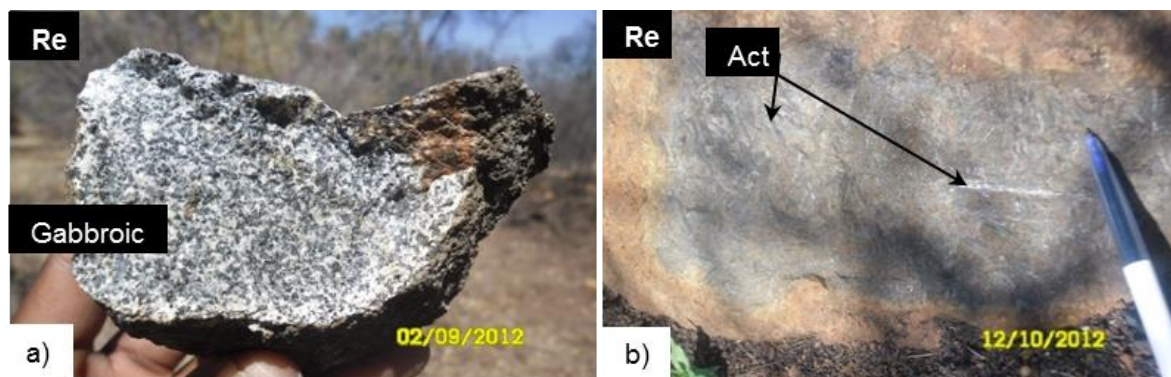


Figure 4.27. (a) Coarse-grained and (b) actinolite-bearing metadolerite sills of the Ventersdorp Supergroup (35 J 538295 m E, 7027610 m S; and 35 J 536084 m E, 7026942 m S, respectively).

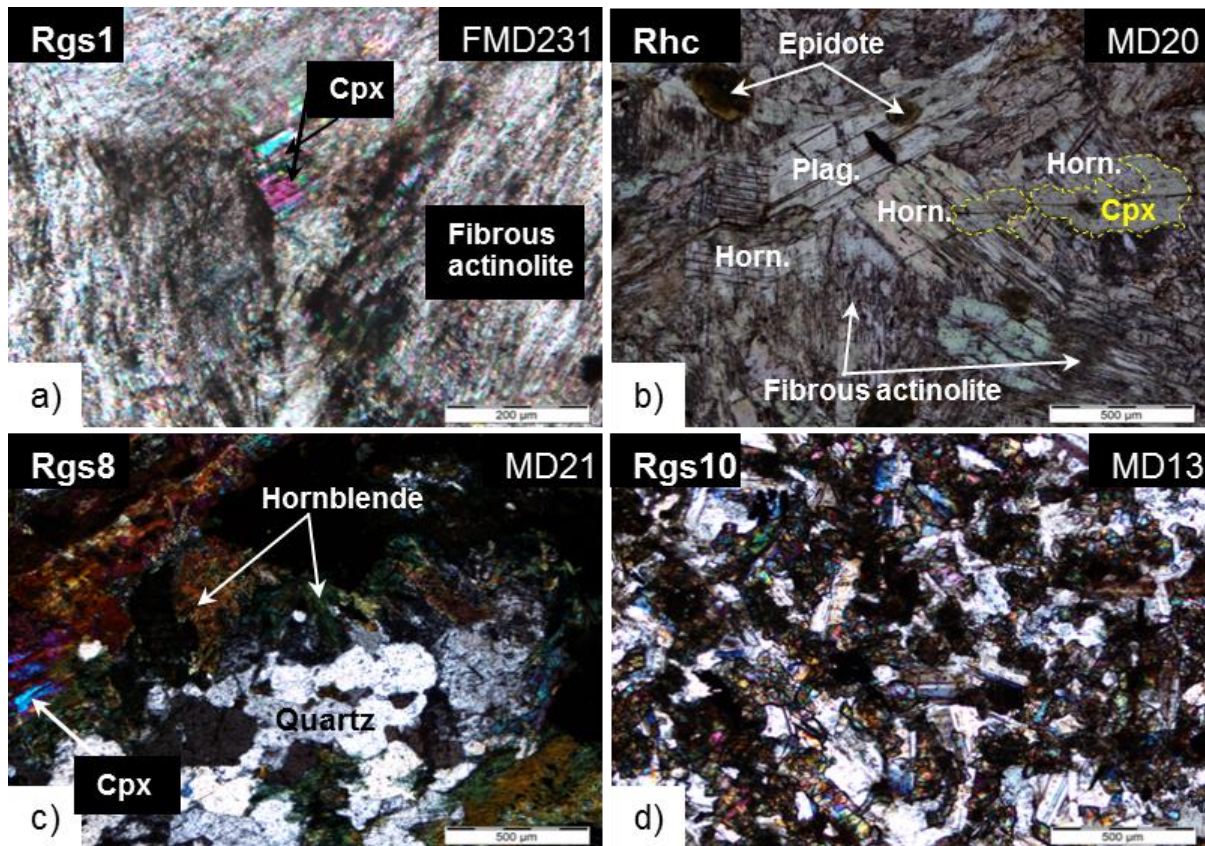


Figure 4.28. (a) Foliated metadolerite dominated by fibrous actinolite (XPL). (b) Clinopyroxene replaced by subidioblastic hornblende, which is in turn replaced by fibrous actinolite in sample MD20 (PPL). (c) Hornblende associated with secondary quartz (XPL). (d) A dolerite sill composed of primary clinopyroxene and plagioclase (XPL; Table 3.4, Appendix III).



Figure 4.29. Amygdaloidal meta-basalt belonging to the Klipriviersberg Group (35 J 539110 m E, 7030197 m S).

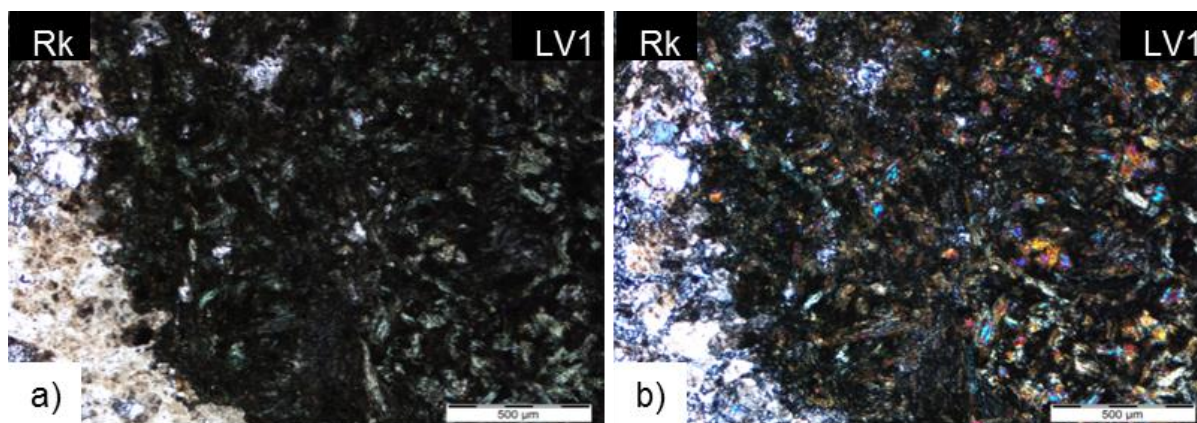


Figure 4.30. Photomicrographs of the amphibole-bearing Klipriviersberg Group meta-basalt (PPL and XPL respectively).

4.2.5. Alkali granite bodies

Two small alkali granite plutons crop out close to the Vaal River in the western part of the area (Figure 3.2). The northernmost intrusion covers an area of about 1.1 km² whereas the southern one covers ~0.78 km² within the limits of the study area. Field mapping for this study did not include these plutons, however Graham et al. (2005) showed that the fine- to medium-grained granite is composed of 43% albite, 25% quartz, 22% k-feldspar, 9% aegirine, and 1% biotite. These intrusions have been dated at 2.052 ± 0.014 Ga (U-Pb zircon from the Schurwedraai alkali granite) by Graham et al. (2005), who linked the intrusions to the Bushveld magmatic event.

4.2.6. Melt rocks

Melt-rocks that are attributed to the Vredefort impact in the study area include granophyre dykes and pseudotachylitic breccias. The largest pseudotachylitic breccias, up to several metres across and tens of metres long, are hosted along impact-induced faults, but small veins and pods as thin as <1 mm are abundant in metapelite units in the south of the study area. The 10 – 35 m wide and ~10 km long granophyre dyke in the southern part of the study area crosscuts the granite basement, Dominion and West Rand groups, and an impact-induced fault (Figure 3.2). Textures of these rocks are discussed in Chapter 7.

4.2.7. Colluvium

Colluvium in the study area is generally developed towards the bottom of valleys, along highly faulted zones, and along less resistant rock types such as slate and meta-ironstone (Figures 3.2, 3.4). It is also developed in the less metamorphosed rocks of the Central Rand Group (e.g., along the Booyens Formation) and part of the Klipriviersberg Group.

Colluvium is generally composed of soil with lithic fragments derived from local rocks, however, in some areas the fragments are cemented, forming consolidated colluvium up to 1.5 m thick. The consolidated fragments comprise ferricrete, which is composed of a medium- to coarse-grained Fe-rich matrix (weathered meta-ironstone and Fe-rich slates) that has cemented quartzite, meta-ironstone, and metapelite boulders depending on the composition of the adjacent rock types (Figure 4.31). The boulders vary largely in size, with the largest one mapped having a 30 cm diameter. Some of the clasts/boulders have been eroded away, leaving a pitted texture within the matrix (Figure 4.31b). Ferricrete is widely distributed within the study area, being located at Buffelshoek, De Waal's, Reikies, Donkervliet, and Eldorado farms (Figure 1.2). In places, stream incision indicates that soil and colluvium may be >3 m thick.

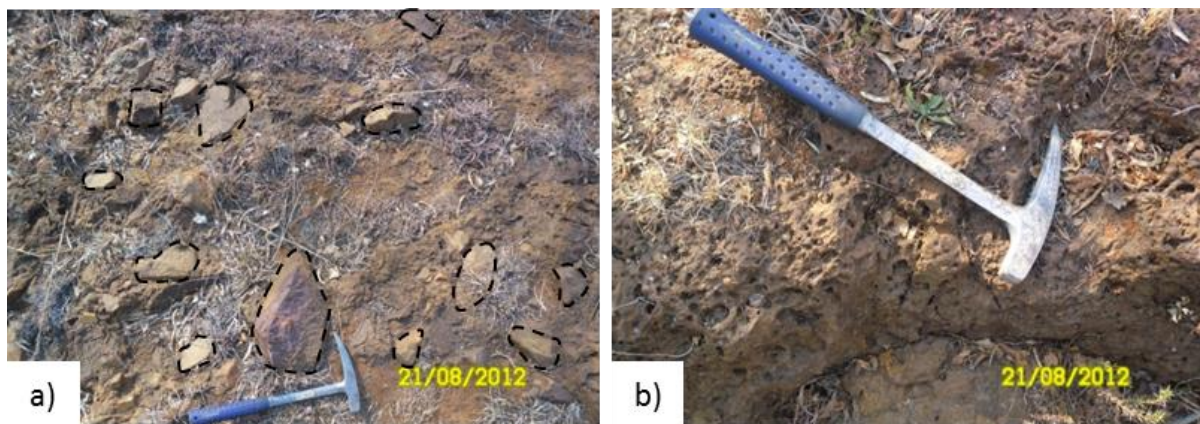


Figure 4.31. Field photographs of ferricrete located in De Waal's Farm (35 J 539869 m E, 7031026 m S).

4.3. Summary

An unconformity divides the Archaean Basement Complex from supracrustal rocks that form the collar of the Vredefort Dome (Figure 3.2). The unconformity is overlain by supracrustal rocks that young towards the north. The strata are south-dipping and overturned as evidenced from inverted cross-bedding developed in quartzite and metapelite units. The oldest supracrustal lithology is the meta-basalt unit belonging to the Dominion Group, which was deposited onto an erosional surface on the Basement Complex. Older units of the Witwatersrand Supergroup (the West Rand Group) were deposited onto the Dominion Group, under deep and shallow marine settings, which allowed for the formation of: (1) ironstone units; and (2) characteristic sedimentary structures such as herringbone cross-bedding developed in the Rhq3 quartzite unit and trough cross-bedding developed in the Ro-Rgq9 quartzite units (Figure 4.9). The younger Central Rand Group was deposited mostly through fluvial processes, which formed the thick aggradational and less mature quartzite units. The fluvial setting was associated with high-energy streams that deposited meta-conglomerate reefs (Table 4.2).

The evolution of depositional environments of the Witwatersrand Supergroup thus shows an overall marine regression, as evident from the thicker clastic units with coarser grain size up-stratigraphy (Pretorius, 1976; Beukes and Nelson, 1995; McCarthy, 2006; Guy et al., 2010). Manzi et al. (2013) suggested that an angular unconformity exists between the West Rand Group and the Central Rand Group, however this could not be confirmed in the study area because impact-induced faults have differentially rotated the rocks into different domains, making it hard to recognise a primary angular relationship between the two groups (Figure 3.2).

No evidence of an unconformity could be found at the base of the Klipriviersberg Group amygdaloidal meta-basalts, although Kositsin and Krapež (2004) maintain that a 129 Ma age gap exists between the Turffontein Subgroup and the Klipriviersberg Group; an unconformity should exist between the two. Ventersdorp-age meta-dolerite/dolerite sills and Bushveld-age granite bodies and mafic sills subsequently intruded the strata, prior to the impact and formation of the dome.

5.0 STRUCTURAL GEOLOGY

5.1. Introduction

Structural studies in the collar of the Vredefort Dome have been completed by Simpson (1977), Lilly (1978), Colliston and Reimold (1990), Brink et al. (2000), Wieland et al. (2005, 2006), Bush (2014) and others (see Chapter 2). Wieland et al. (2005) studied structures mainly in quartzite units of the West Rand Group; Simpson (1977) studied those in the Potchefstroom Synclinorium north of the study area (Figure 1.1), whereas Lilly (1978) conducted a regional photo-lineament analysis of structures in the collar of the dome. In summary, these studies recorded that the collar hosts a variety of structures: (1) kilometre-scale open to tight symmetric folds and open asymmetric homoclinal folds with axial planes that are radial and oblique to the dome, respectively (Wieland et al., 2005); (2) wrench faults that are radial and oblique to the dome (Simpson, 1977; Lilly, 1978; Wieland et al., 2005); (3) joints and shear fractures that are radial, oblique, and tangential to the dome (Lilly, 1978; Wieland et al., 2005; Reimold and Gibson, 2006); and (4) possible pre-impact, large-scale fold (Colliston and Reimold, 1990) and fault structures (Lilly, 1978; Bisschoff, 1982; Tinker et al., 2002; Wieland et al., 2005; Lana et al., 2006), and syn-metamorphic foliations (Gibson, 1993). This study focuses on mapping of geological structures hosted across the Witwatersrand Supergroup in the northern collar of the dome, in an attempt to link structural studies conducted in the inner collar by Wieland et al. (2005) to those conducted in the outer collar by Simpson (1977). Labelling of deformations interpreted from the structures, and associated metamorphic events, is based on crosscutting relationships observed in the study area rather than deformation and metamorphic events described in the literature review (i.e., M_2 in the literature review does not necessarily correspond to M_2 described in the study area). To avoid confusion, regional metamorphism and metamorphism described in the study area will be referred to as $M_{(REG)}$ and $M_{(NC)}$ hereafter, respectively.

The rocks in the study area record four deformation events labelled D1, D2, D3, and D4. The D4 Vredefort impact succeeded deformations D3, D2, and D1 (Table 5.1). The D1 deformation is brittle, and characterised by listric fault formation. The ductile deformation D2 is characterised by regional S_2 composite cleavage. The S_2 cleavage is syn- to post-genetic to $M_{2(NC)}$ metamorphic assemblages described in Chapter 6. D3 deformation is defined by an S_3 crenulation foliation that overprints the S_2 foliation. D4 is defined by brittle and brittle-ductile structural features produced by the Vredefort impact, in addition to shock microdeformation (PDF), shatter cones, shock cleavage (Gibson and Reimold, 2008) and rotation of the collar sequence. The D4 structures include faults, symmetric and homoclinal folds, pseudotachylitic breccia veins, a spaced fracture cleavage (S_{4b}), and kinks. D4 features constrain pre-impact deformations through crosscutting relationships with D1-D3 structural features (Table 5.1).

Table 5.1. Interpretation of crosscutting relationships exhibited in the study area.

Deformations		Deformation features			Structural Elements
		1 st -order features	Macroscopic features	Microscopic features	
D4 Deformation	Impact-induced deformation features	Granophyre dyke			F ₄ S _{4a} , S _{4b} & S _{4c} M ₄
			Pseudotachylitic breccia		
		Drag folds			
		Generally north-south trending, pseudotachylitic breccia-bearing faults.	micro-faults		
		Thrust-folded Rgq3 quartzite unit	Kinks		
			Joints		
			Spaced fracture cleavage (SFC)		
D3 Deformation				Crenulations of the S ₂ cleavage (S ₃)	L _{3a} & L _{3b} F ₃ S _{3a} & S _{3b}
D2 Deformation	Ductile deformation	Regional development of an S ₂ foliation, with local S ₂ -C shear	Knotted quartzite horizons developed in the Johannesburg and Turffontein subgroups.	Quartz-muscovite-sericite cleavage in knotted quartzite.	S ₂ cleavage M ₂
			Foliation developed parallel and oblique to bedding	Quartz-biotite ± muscovite ± chlorite cleavage in metapelite.	S ₂ -C shears
			Localised east-verging shear zones	Quartz-sericite-muscovite cleavage in quartzite.	F ₂ intrafolial quartz vein folds.
			Quartz Veins		V ₂
D1 Deformation	Brittle, extensional deformation features	Rollover and drag folds developed along the Karolina Shear.			
		Arcuate Faults and the Karolina and Mooiplaas shear zones.			
				A relict planar fabric defined by quartz inclusions	S _x

During D4 deformation, beds of the West Rand Group (average dip of 56° S) in the study area were rotated by up to 124° in a clockwise direction when looking WSW. Thus, a correction is applied where necessary for the pre-impact data (i.e., data are rotated 124° anti-clockwise looking WSW). A map view of the study area is, thus, effectively a crustal section for pre-impact structures, making descriptions of the precise original orientations of the structural elements before overturning difficult, particularly as evidence exists that differential block rotation has occurred (e.g., Wieland et al., 2005). However, in this chapter, orientations of pre-impact structural elements are described in their overturned positions. The strata that host the elements are subsequently restored back to horizontal, to (approximately) analyse the geometry of the elements before the Vredefort Impact. As a note of caution, the rotational angle of 124° is based on a mean dip of 56° S of overturned bedding (West Rand Group), which is inferred to have been horizontal before impact and rotation took place. Owing to the fact that the actual pre-impact bedding orientation of the target rocks in the Vredefort Dome has not been established, the orientation of the target rocks before impact-induced deformation is assumed to have been horizontal; this is an assumption considering that 2 deformations have affected rocks of the Witwatersrand Supergroup (c.f., Manzi et al., 2013). The least affected region of the supergroup is the lower West Rand Group, which is located in the footwall of Umzawami-age listric faults associated with the largest displacement of the Witwatersrand Supergroup (Section 2.2.3; relatively consistent bedding in southeast part of Figure 3.2). Considering this, only lower West Rand Group bedding was rotated back to the horizontal when estimating pre-impact orientation of structures (e.g., foliation in Section 5.3.1).

5.2. D1 deformation

The D1 deformation is represented by a set of arcuate faults. The faults range from 1st-order (macroscopic) to 3rd-order (mesoscopic) in scale, and exhibit listric geometries in their restored orientation. The faults were interpreted from Google Earth images and established during field work from discontinuous geological units (Figures 3.4, 5.1). The 1st order-scale Karolina Fault (Figure 3.4) extends for a minimum of 15 km from the northeast of the study area, where it offsets the Central Rand and Klipriviersberg groups, to the southwest, where it displaces the Government Subgroup of the West Rand Group. The study area exhibits poorly exposed and preserved outcrops of the fault zone rocks; as a result the geometry and kinematic information of the Karolina Fault are not described. Marker horizons on either side of the Karolina Fault shows an apparent 3 km left-lateral separation of the Witwatersrand and Ventersdorp supergroups. Furthermore, the northwestern block (restored hangingwall) has been rotated clockwise to its present position, consistent with reverse drag/hangingwall collapse in the restored orientation. The fault terminates against the base of the Transvaal Supergroup (Figure 5.1) suggesting it was only active prior to deposition of the Transvaal

Supergroup. The age of the Transvaal Supergroup unconformity is taken as 2.588 Ga, which is the published age of tuff units in the Oaktree Formation (single zircon U–Pb SHRIMP; Martin et al., 1998) considered as the proxy for the base of the Transvaal Supergroup (Henry et al., 1990; Eriksson et al., 2006; Manzi et al., 2013). The Karolina Fault is thus older than 2.588 Ga.

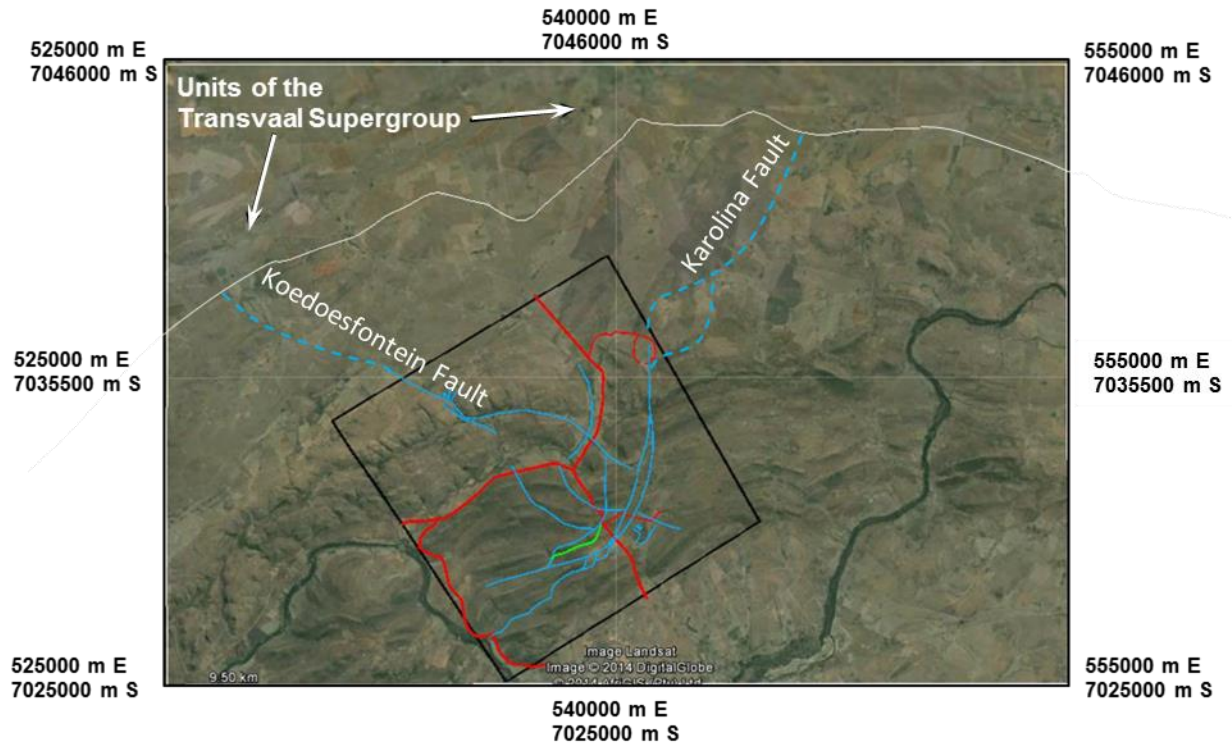


Figure 5.1. Google Earth™ image showing D1 arcuate faults (up to 3 km slip) terminated against the base of the Transvaal Supergroup represented by the white line (c.f., Bisschoff, 2000). All the faults are labelled in Figure 3.4.

The northwestern block (hangingwall) of the Karolina Fault hosts arcuate 2nd order faults, namely the Koedoesfontein, Mooiplaas, De Waal's, Buffelshoek, and Rondekop faults (Figure 3.4). The 12.5 km long Koedoesfontein Fault trends northwest-southeast and exhibits an apparent right-lateral offset of the Central Rand Group by ~700 m. The Koedoesfontein Fault is also terminated against the base of the Transvaal Supergroup, indicating that activity on the Koedoesfontein Fault may be older than 2.588 Ga. The Mooiplaas Fault is 4 km in length and exhibits an apparent right-lateral offset of the Government and Jeppestown subgroups by tens of metres. The De Waal's Fault has a minimum length of 2.8 km, and exhibits an apparent right-lateral offset of the Jeppestown Subgroup by ~800 m. The Rondekop Fault lies subparallel to the Karolina Fault and is approximately 4 km in length; it offsets the Government and Jeppestown subgroups left-laterally (apparent) by ~300 m. The Buffelshoek Fault is inferred from an apparent left-lateral offset of <100 m. It lies sub-parallel

to the Karolina Fault and is about 5 km in length (Figure 3.2). Like the Karolina Fault, fault zone outcrops of the 2nd-order faults do not crop out, but are covered by colluvium.

Second-order D1 faults bound rollover/reverse drag-folds and drag folds in the northwestern block of the Karolina Fault. A 1st-order, synformal anticline, rollover fold in the northwestern block has a wavelength of ~5.3 km and amplitude of ~0.8 km. It is bounded by the Koedoesfontein Fault to the west and the Karolina Fault to the east. The largest drag fold in the study area is an antiformal syncline, bound by the southern-most segments of the Karolina and Mooiplaas faults (Figure 3.4; area 'b' in Figures 5.2a). The drag fold has a wavelength of 6.2 km and amplitude of 1.5 km. The rollover and drag folds are overturned and have fold closures that generally face to the north. Stereographic projections (areas 'a' and 'b' in Figure 5.2) show that the folds are characterised by moderately curved limbs, an interlimb angle ~70°-120° and form broad fold hinges that plunge at moderate to steep angles towards the south (52°-60°).

The southeastern block (restored footwall) of the Karolina Fault hosts a series of 3rd order-scale, NNE- to ENE-trending faults that offset the Rhq1-Rhq3 quartzite units left-laterally (apparent) by magnitudes up to 200 m (Rondekop) (Figure 3.4). These fault displacements have pulled apart the Rhq1-Rhq3 quartzite units (Figure 3.2) on Rondekop and Doringbos farms (Figure 1.2).

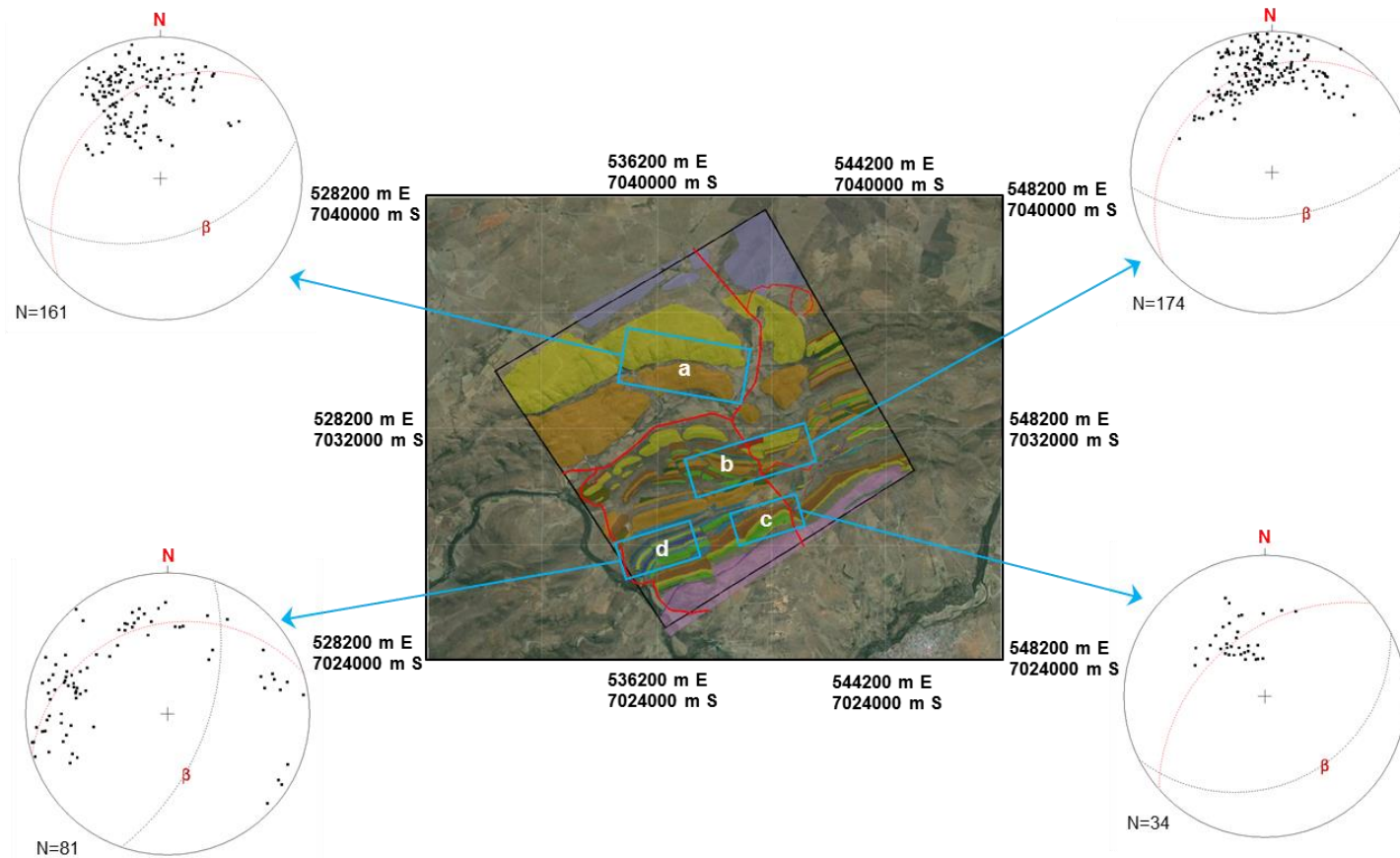


Figure 5.2. Equal area stereographic projections of poles to bedding planes of drag and rollover folds hosted in the Witwatersrand Supergroup. (a) represents a synformal anticline (rollover anticline; $\beta = 52^\circ$ towards 136° , mean orientation = 55° towards 165° , and interlimb angle = 110°); (b) a gently-folded antiformal syncline (drag fold, $\beta = 59^\circ$ towards 140° , mean orientation = 64° towards 174° , and interlimb angle = 069°); and (c) & (d) synformal anticlines with β equals 53° towards 163° (interlimb angle = 113° and mean orientation = 37° towards 153°), and 36° towards 139° (interlimb angle = 77° and mean orientation = 66° towards 109°) respectively (drag folds).

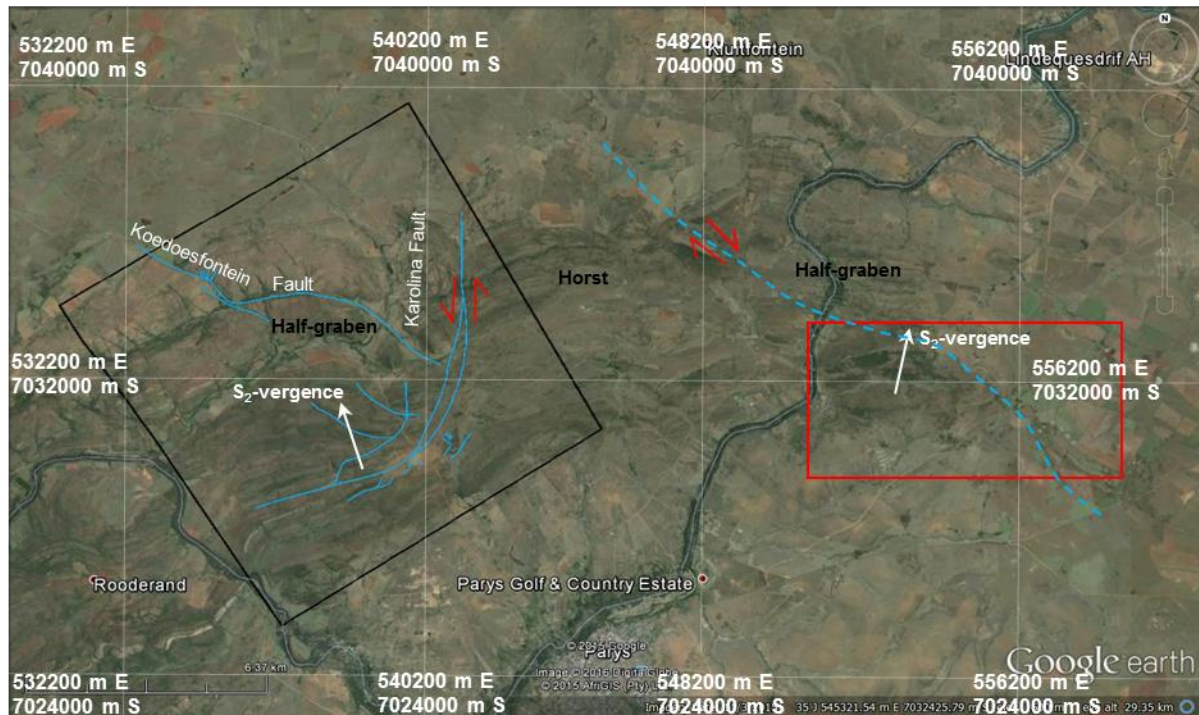


Figure 5.3. Google EarthTM image showing the approximate location of the eastern arcuate fault (Bush, 2014), which, when interpreted to be pre-impact (Ventersdorp-age), forms horst and half-graben structures with the Karolina Fault. Bush's (2014) study area is represented by the red box. The white arrows indicate the direction of S_2 vergence discussed in section 5.3.1.

5.3. D2 deformation

5.3.1. S_2 cleavage

A heterogeneously developed, but regional, foliation (S_2) is developed as discrete/narrow (typically <1 m wide) cleavage horizons that are found throughout the study area (spatial distribution of samples shown in Figure A3.1, Appendix III). S_2 cleavage is generally parallel-oblique (<35°) to bedding (Figure 3.4), and is differentially rotated by impact-induced faults and folds (e.g., Figure 5.4). S_2 occurs in the knotted quartzite horizons of the Central Rand Group, as well as quartzite units, metapelite, meta-ironstone, and meta-dolerite hosted in the West Rand Group (Table A3.6, Appendix III). Gibson (1993) noted that pelitic 'hornfels' in the lower West Rand Group contain two foliations but that these are generally obscured in hand specimen by the post-impact thermal annealing and metamorphic recrystallization, therefore, S_2 is likely more common.

Petrographically, the S_2 cleavage is generally defined by phyllosilicates (biotite, muscovite, sericite and chlorite), amphibole (grunerite, actinolite, hornblende), and deformed quartz grains depending on which rock formation it crosscuts. It is characterised by the following metamorphic

assemblages: (a) quartz-biotite-muscovite $\pm$ chlorite in metapelite units of the West Rand Group; (b) quartz-muscovite-sericite $\pm$ chlorite in phyllitic horizons of quartzite units of the West Rand and Central Rand groups; (c) quartz-magnetite $\pm$ grunerite in meta-ironstone units of the West Rand Group; and (d) fibrous actinolite-chlorite-plagioclase in meta-dolerite (Table A3.6, Appendix III). The S_2 cleavage wraps chloritoid, cordierite and garnet porphyroblasts in pelitic units and garnet in meta-ironstone (e.g. Figures 5.5). It can thus be interpreted that S_2 formed syn- to post-kinematic to the porphyroblasts ($M_{2(NC)}$ metamorphism; Chapter 6).

In the Central Rand Group of the study area, five knotted quartzite horizons were mapped, with the thickest and most extensive unit being located along the northern contact of the Kimberley Reefs (Figures 5.6, 5.7; and KT samples, and samples B3 and 60 in Figure A1.9, Appendix I). The R_t knotted quartzite horizon at the stratigraphic level of the Kimberley Reefs is important for determining the relative chronology of structural features because it is thick enough (approximately 35 m wide zone) to be traced across the greater mapping area and thus forms a marker unit (Figure 3.2). Google EarthTM images combined with field mapping show that this horizon is displaced by a set of north-south and northwest-southeast trending pseudotachylitic breccia-bearing faults (Figures 5.7). This crosscutting relationship, along with the observation that S_2 is crosscut by pseudotachylitic breccia and kinks (Figure 5.8), confirms that S_2 hosted in the Central Rand Group knotted quartzite formed prior to the Vredefort Impact. Additionally, Bush (2014) observed that the kink bands crosscutting S_2 host pseudotachylitic breccia in a locality ~10 km east of the study area, confirming the impact origin of the kink bands. Knotted quartzite horizons host oxidised chloritoid, kyanite and/or andalusite porphyroblasts that attain 3 cm in length (Figure 5.6). Seven samples of knotted quartzite were cut for thin section analysis (Table A3.5, Appendix III). The S_2 cleavage in knotted quartzite is defined by the assemblage sericite-muscovite-quartz (Figure 5.8). The cleavage is syn to post-kinematic to the formation of subidioblastic kyanite porphyroblasts as evidenced by the deflection of S_2 cleavage around kyanite porphyroblasts (Figure 5.9a).

In an effort to determine the pre-impact vergence of S_2 , lower West Rand Group bedding was back-rotated to the horizontal to undo the overturning effect of the impact (Figure 5.10); therefore rotating S_2 foliations hosted in the lower West Rand Group to their pre-impact orientation when the strata (S_0) were horizontal. Considering calculated mean bedding orientation ($063^\circ/62^\circ$ S), the strata were rotated by 118° anticlockwise, looking WSW along an axis oriented 243° (strike-parallel). The rotation results in a pre-impact S_2 orientation of $044^\circ / 12^\circ$ S, indicating a NW-directed vergence relative to horizontal bedding (Figure 5.10). This is broadly consistent with interpretations by Bush (2014) that S_2 (S_1 in Bush, 2014) verges N to NNE, considering that bedding strikes at $090^\circ - 150^\circ$ in the latter study compared to 063° in this study area (Figure 5.3). This indicates a generally northwards vergence across the two study areas.

In an effort to establish crosscutting relationships between D_2 and D_1 in the Central Rand Group, an attempt to compare S_2 along strata folded by D_1 faults was made. Owing to the narrow and

non-extensive nature of S_2 horizons along folded regions, this relationship could not be established. However, extensive and wider S_2 horizons are hosted in the West Rand Group, where there is insignificant folding by the D1 deformation (compare cleavage plots in Figure 3.5). Also, Bush (2014) and Nemutandani (2014) described metamorphic mineral assemblages similar to $M_{2(NC)}$, which exhibit different metamorphic grades along the same units that were displaced by D1 faults (Section 8.2).

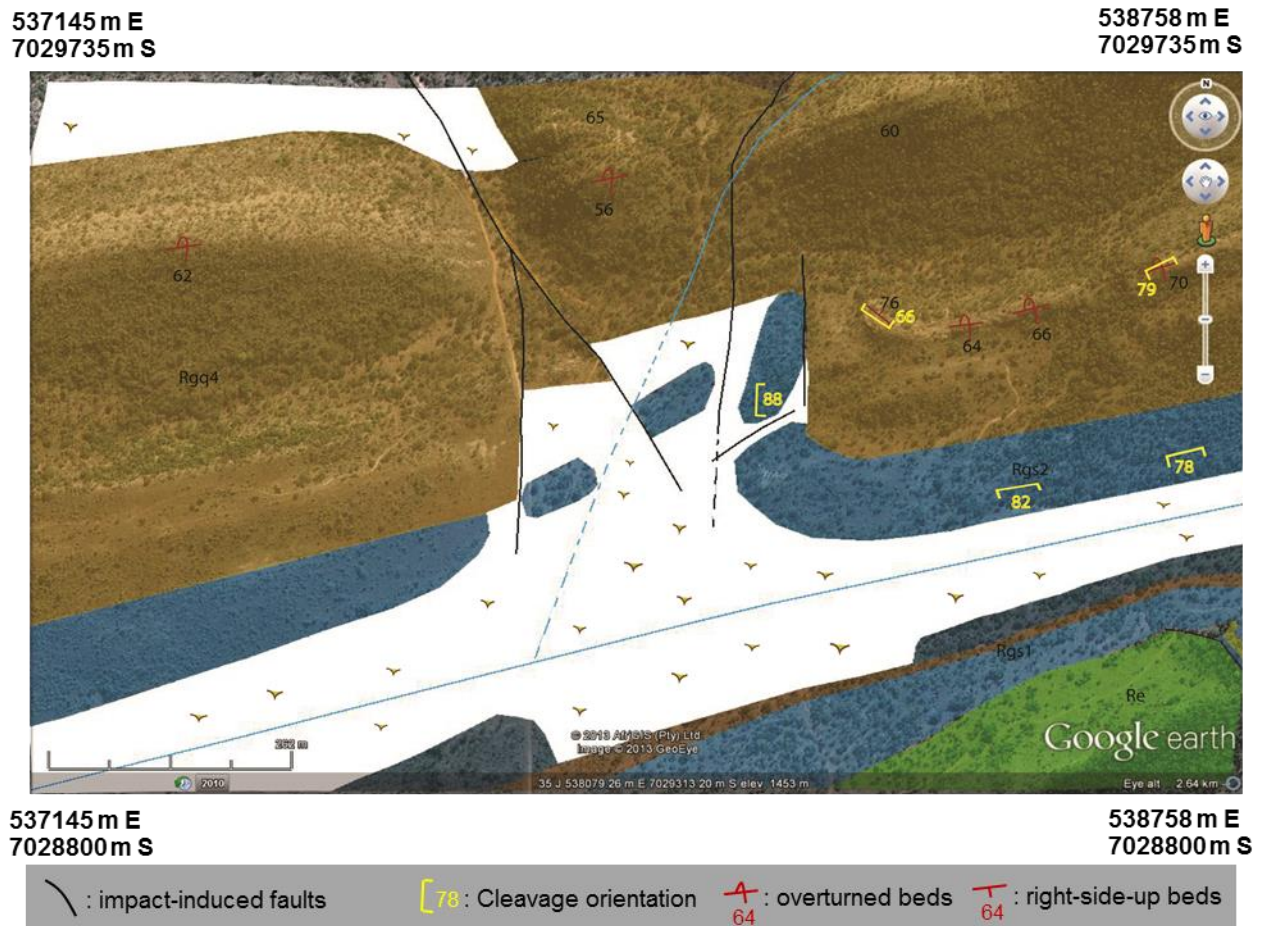


Figure 5.4. S_2 and S_2 -C cleavages rotated by pseudotachylitic breccia-bearing D4 syncline (also see Figure 5.17).

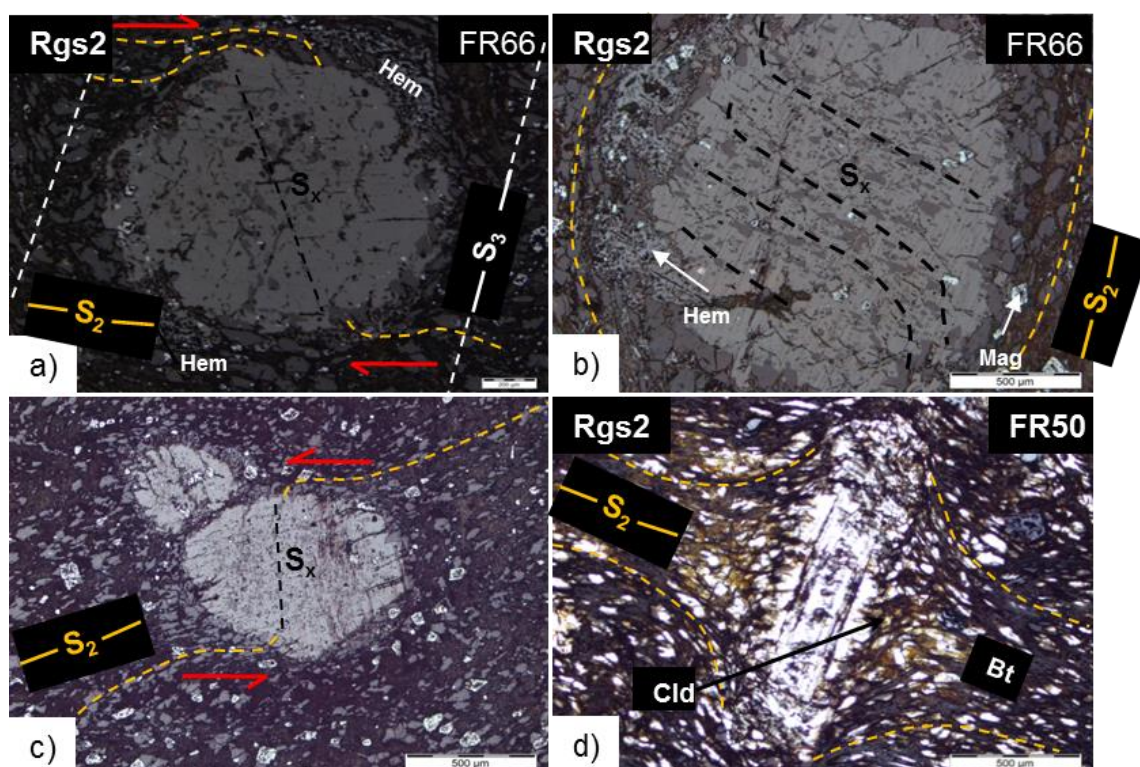


Figure 5.5. Syn to post-kinematic S_2 cleavage developed on a subhorizontal section of sample FR66, showing S_x inclusion trails in garnet (a-c). (a) S_2 cleavage indicative of a clockwise rotation; (b) S_2 cleavage developed perpendicular to the S_x inclusion trails; (c) S_2 cleavage indicative of a counter-clockwise rotation (all three in RL). (d) Chloritoid (Cld) pseudomorph (replaced by muscovite) porphyroblast wrapped by a syn-kinematic S_2 cleavage (PPL). Figures a-c show an outward increase in the size of magnetite porphyroblasts hosted in garnet.



Figure 5.6. Knotted quartzite horizons hosted in the Central Rand Group (35 J 531767 m E, 7032996 m S; and 35 J 534576 m E, 7034398 m S respectively), with kyanite and/or andalusite wrapped by the syn-metamorphic S_2 .



Figure 5.7. Knotted quartzite horizon (red line) shown in Figure 5.6 is displaced by generally northwest-trending D4 fault sets (blue circle) in the Turffontein Subgroup. Samples collected from the horizon are also plotted.

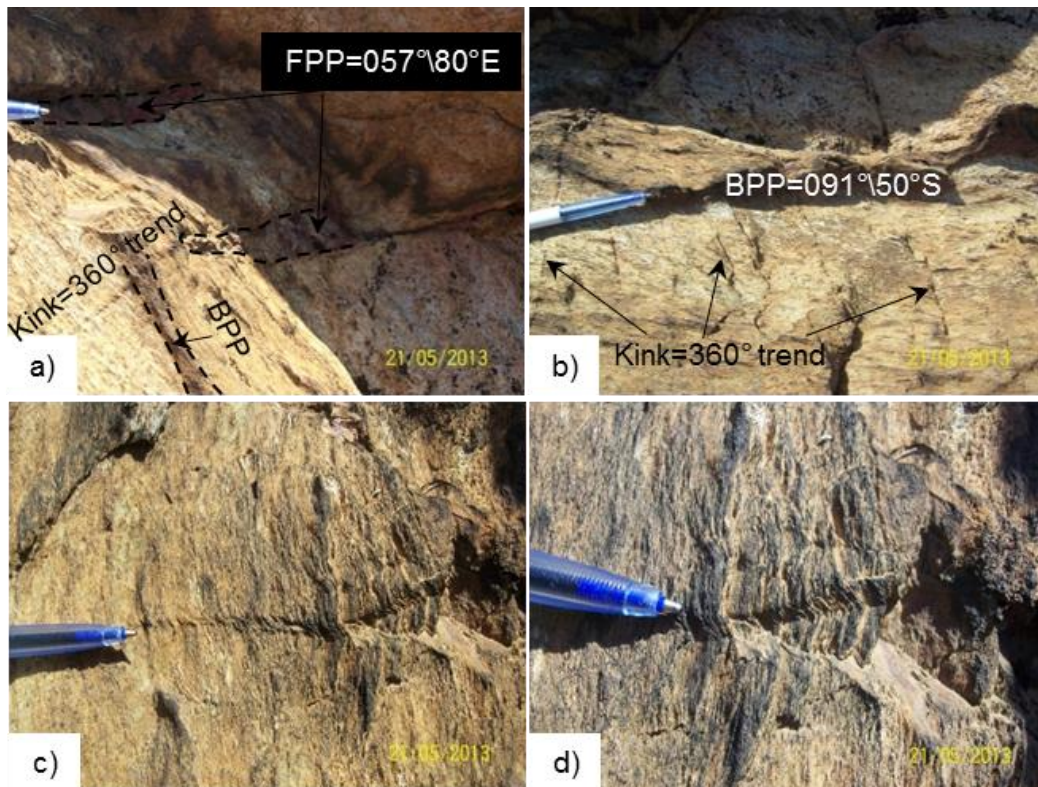


Figure 5.8. (a) & (b) S_2 cleavage developed along knotted quartzite horizons is crosscut by: (a) & (b) bedding-parallel (BPP) and fracture-hosted pseudotachylitic breccia (FPP); and (c) & (d) kink bands trending 360° (35 J 538918 m E, 7034450 m S).

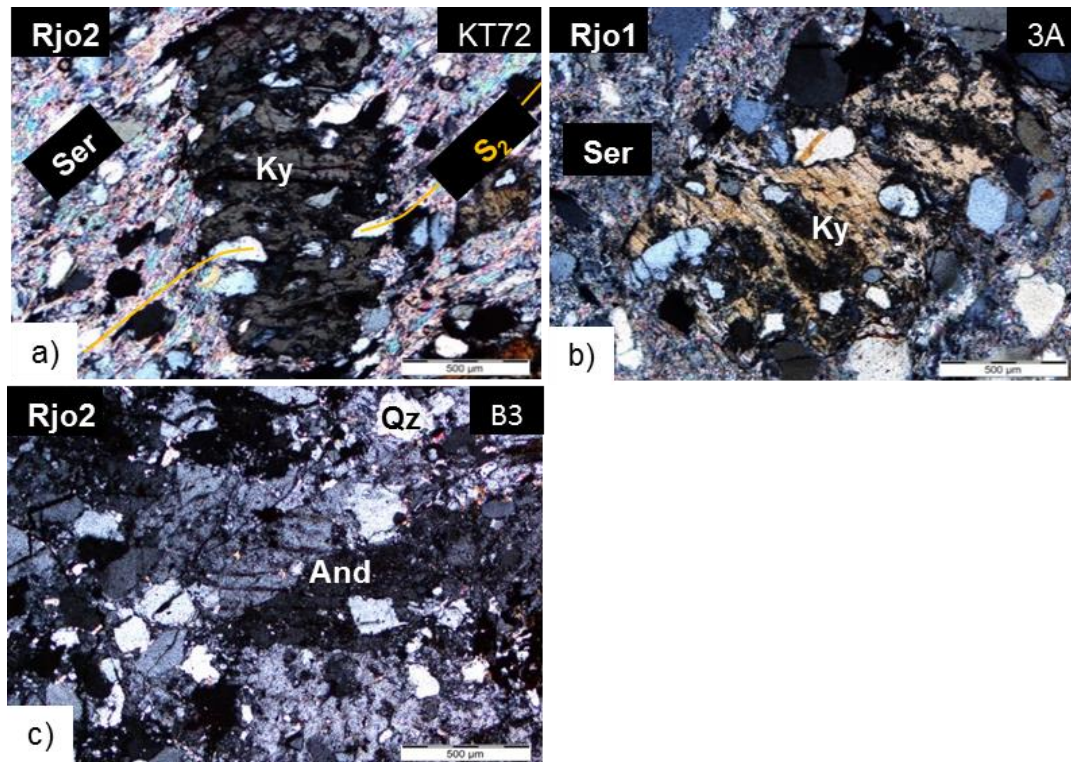


Figure 5.9. (a) A syn to post-kinematic S_2 cleavage developed in knotted quartzite. (b) & (c) Subidioblastic kyanite and xenoblastic andalusite porphyroblasts respectively. Samples are composed of: (a) 35% sericite+muscovite, 30% quartz, 18% kyanite, 12% andalusite, and 5% chloritoid; (b) 35% andalusite, 30% quartz, 10% oxidised chloritoid, 15% sericite, and 10% kyanite; and (c) 46% andalusite, 30% quartz, 15% chloritoid, 5% muscovite, and 4% kyanite (XPL).

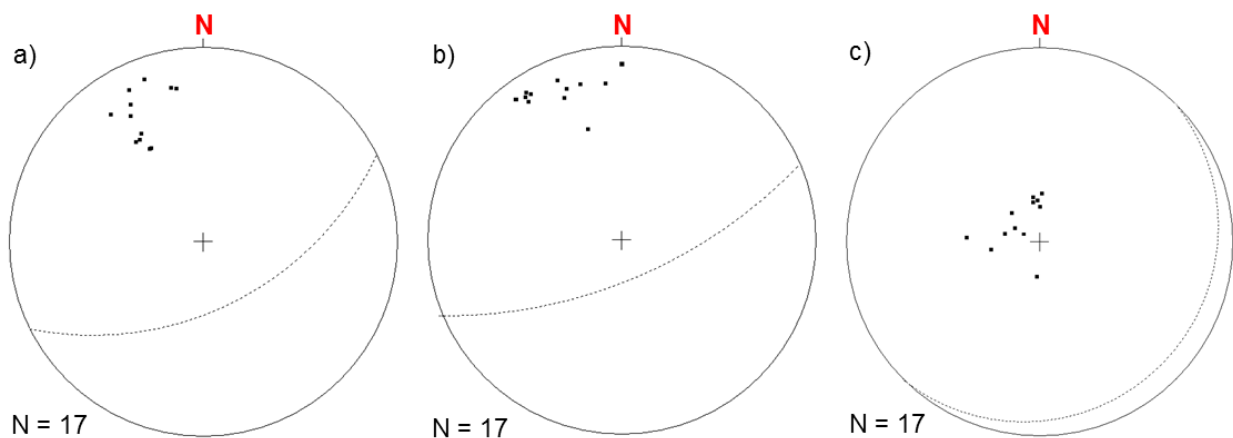


Figure 5.10. Stereographic projections of: (a) poles to S_0 planes, indicating an average S_0 orientation of $063^\circ / 62^\circ S$; (b) poles to S_2 planes, indicating an average S_2 orientation of $067^\circ / 73^\circ S$; (c) resulting S_2 pole positions when the D4 impact overturning is undone (118° rotation to the south, along a 063° axis), showing a pre-impact S_2 orientation of $044^\circ / 12^\circ S$ and, thus, a NW-directed (314°) vergence.

5.3.2. S_2 -C cleavage

The regional S_2 cleavage is associated with zones of localised sigmoidal shear on Rondekop Farm. S_2 -C is developed along the Rgs2 metapelite and Rgq3 quartzite units. In its unrestored form, S_2 -C developed in the Rgs2 metapelite generally trends ENE-WSW and dips $\sim 72^\circ$ S, and that developed in the Rgq3 unit trends NE-SW and dips $\sim 70^\circ$ S (Figure 3.4). The S_2 component of this composite cleavage indicates a NW-directed, pre-impact vergence (Figure 5.10).

At the outcrop-scale, the relationship of S to C in the Rgq2 quartzite unit indicates a right-lateral sense of slip on shear bands (Figure 5.11) which is consistent with that indicated from asymmetric intrafolial folds located within the same cleaved horizon (V_1 and F_2 ; Figure 5.12). The S_2 -C cleavage is itself crenulated by an S_3 cleavage (Figure 5.13; Section 5.4).

Garnet porphyroblasts in samples of the Rgs2 metapelite generally host a relict fabric (S_x) defined by inclusion trails of quartz (Figures 5.5a-c). The S_x fabric could be bedding (S_0), a pre-existing foliation (pre- S_2), or early S_2 : as a result it was not classified with respect to the other fabrics. The inclusion trails are generally straight, with curvature of the foliation generally developed adjacent to edges of the porphyroblasts (Figures 5.5a-c). There are some exceptions where a garnet porphyroblast contains a sigmoidal and curved inclusion trails (e.g., Figure 5.14 and samples S1 and S2 in Table A3.6, Appendix III). These textures indicate that the S_2 -C cleavage is syn- to post-metamorphic to the porphyroblasts. This is consistent with observations by Gibson (1993) that cordierite, andalusite, staurolite, and garnet porphyroblasts show syn-tectonic textures to S_2 foliation in metapelitic units of the Hospital Hill Subgroup; and the post-metamorphic nature is consistent with the observation that S_2 is also developed in Bushveld-age granite intrusions in the northeastern collar (Bush, 2014).

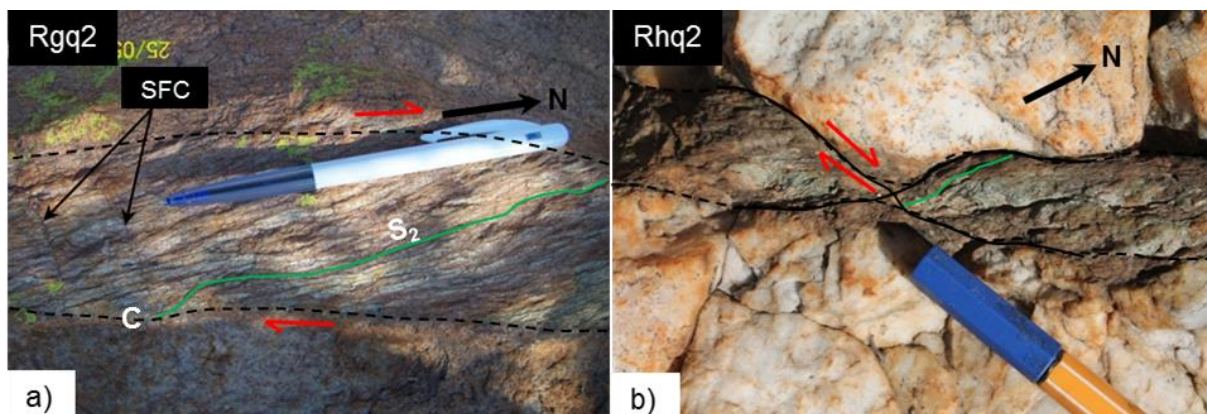


Figure 5.11. (a) Sigmoidal shear cleavage (S_2 -C) developed in argillaceous layer in the Rgq2 quartzite is crosscut by a spaced fracture cleavage ($SFC=S_{4b}$). (b) An extensional shear band that has boudinaged the shear cleavage developed in the Rhq2 quartzite (35 J 535140 m E, 7027318 m S; and 35 J 535900 m E, and 7027094 m S).

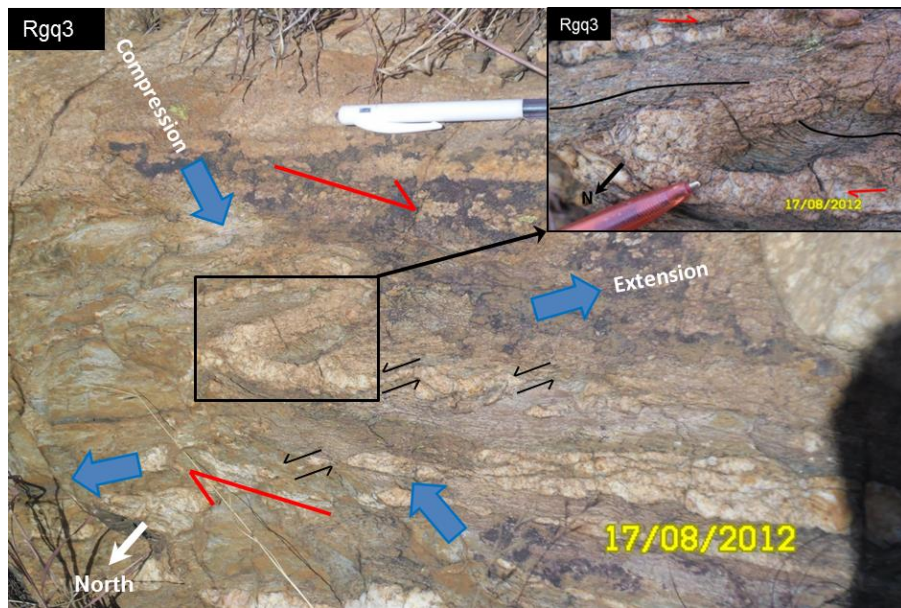


Figure 5.12. A tight, Z-shaped, intrafolial fold (F2) of quartz veins (F1) hosted within the S_2 -C cleavage (35 J 539060 m E, 7029681 m S). The maximum compressive stress is oriented east-west.

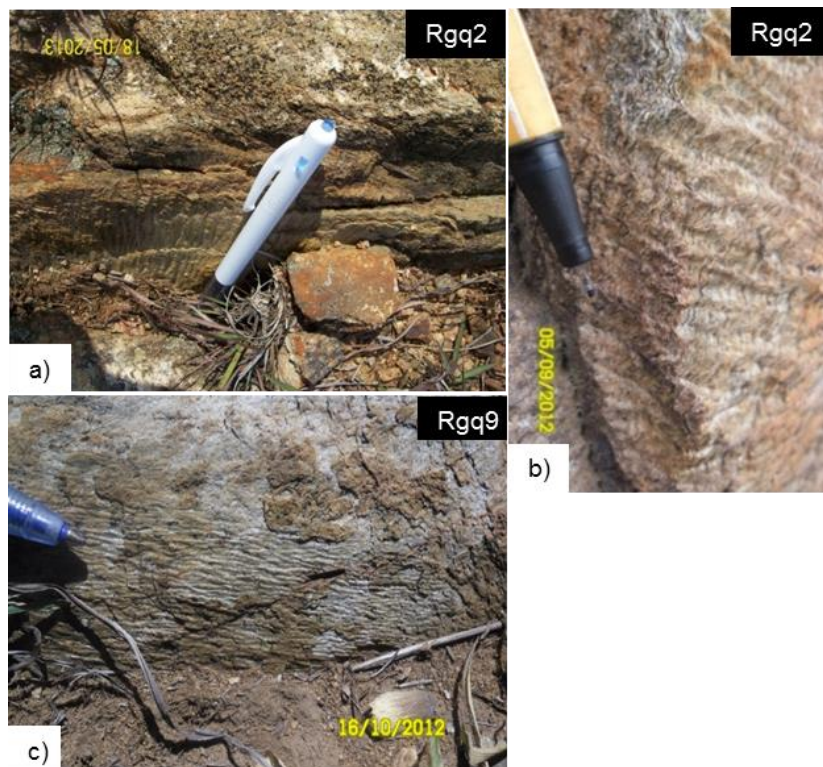


Figure 5.13. S_2 cleavage crenulated along two different planes in Rgq2, Rgq3, and Rgq9 quartzite units (West Rand Group). (a) Subvertical planes (S_{3a}) of crenulation that formed subvertical crenulation lineation (35 J 535162 m E, 7027477 m S). (b) and (c) subhorizontal planes (S_{3b}) of crenulation that formed subhorizontal crenulation lineations (35 J 353142 m E, 7027383 m S and 35 J 541106 m E, 7031541 m S, respectively).

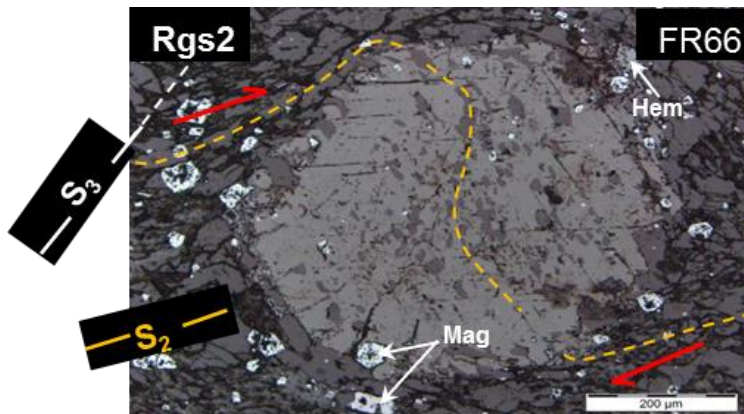


Figure 5.14. Photomicrograph of a weakly-developed sigmoidal cleavage in a garnet porphyroblast hosted in the Rgs2 metapelite.

5.4. D3 deformation

S_2 -C and S_2 cleavages developed in the West Rand and Central Rand groups are crenulated to form a generally weakly-developed S_3 cleavage (e.g., Figures 5.13, 5.15). The F_3 crenulations are exhibited on outcrop scale and microscopically. The main cleavage, S_{3a} , is generally oriented WNW-ESE and is steeply south-dipping ($\sim 110^\circ/55^\circ$ SW); however there are local variations in the orientation owing to impact-induced rotation. There is a subordinate, subhorizontal, S_{3b} cleavage, which is likely a conjugate to the former (Figure 5.13 and 5.16). The conjugate nature was not confirmed because of the rare occurrence of S_{3b} . The S_3 crenulations form intersection lineations (L_{3a} and L_{3b}), oriented either subparallel to the dip direction or strike of the S_2 cleavage, depending on which S_3 conjugate intersects S_2 (Figure 5.13). S_3 is rotated with bedding adjacent to D4 faults (Figure 3.4) and is crosscut by impact-induced spaced fracture cleavage (S_{4b} ; Figure 5.11a); indicating its pre-impact nature.

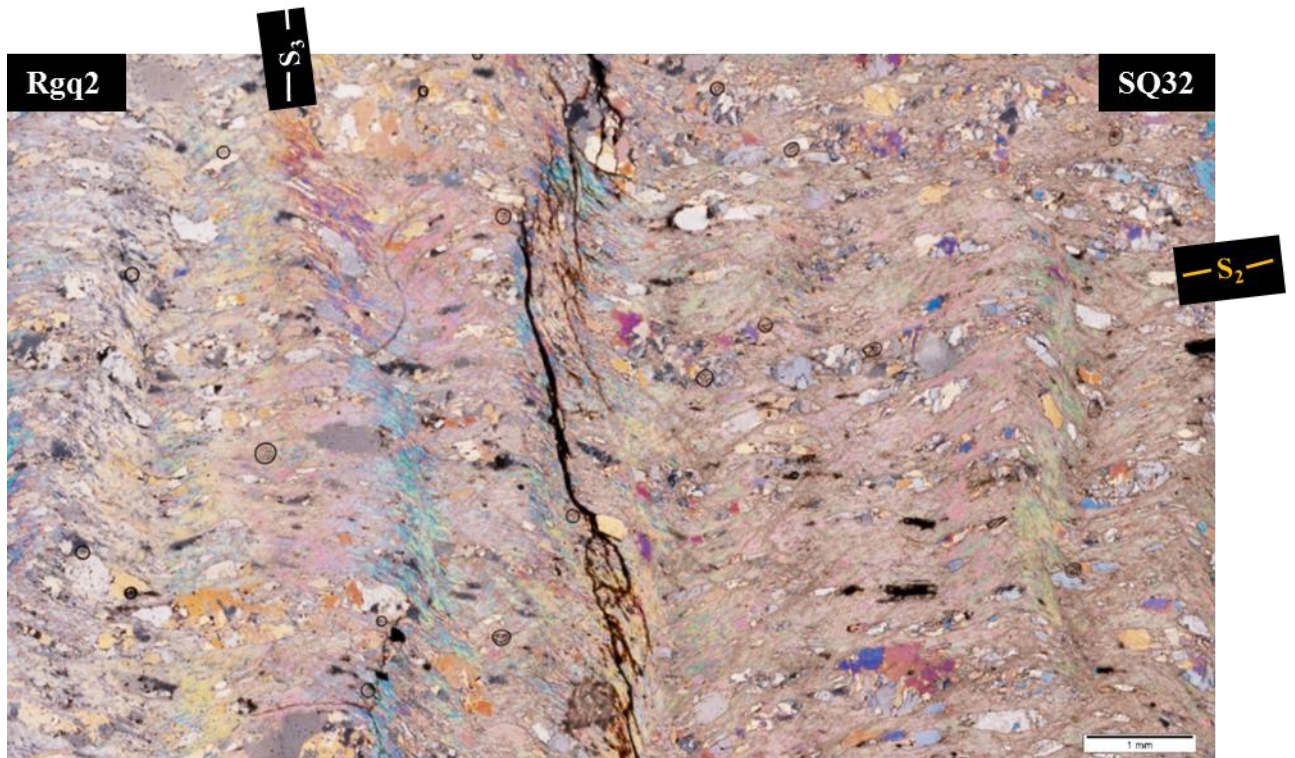


Figure 5.15. Photomicrograph mosaic of the crenulated S_2 cleavage developed along the Rgq2 quartzite unit. The crenulations are crosscut by a goethite-filled fracture (DIC=Differential Interference Contrast).

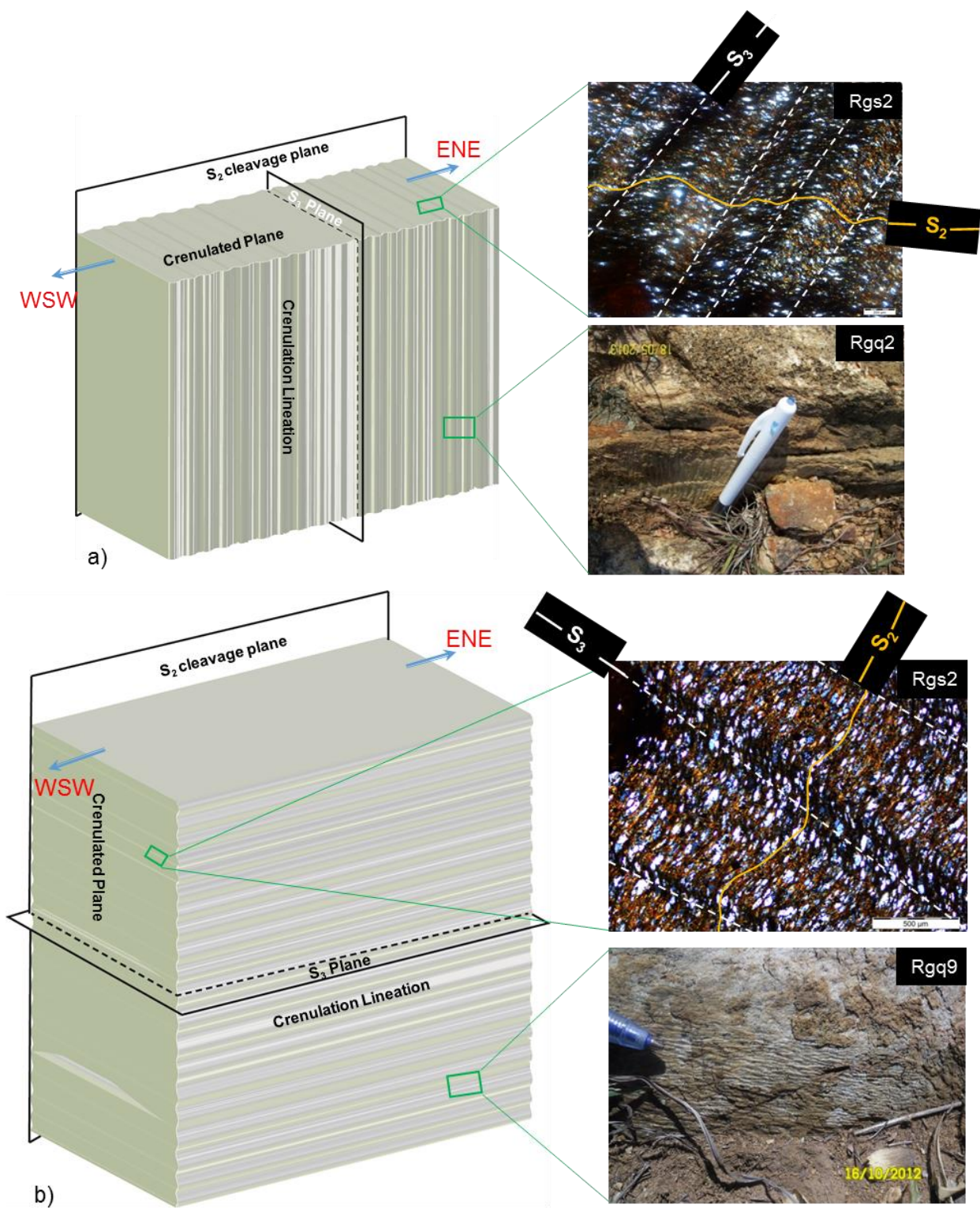


Figure 5.16. An illustration of orientations of S_3 crenulation planes developed in quartzite and metapelite units. (a) Subvertical planes of crenulation, with the resultant crenulated S_2 cleavage shown along a subhorizontal plane (35 J 535162 m E, 7027477 m S). (b) Subhorizontal planes of crenulation, with the resultant crenulated S_2 cleavage shown along a subvertical plane (35 J 541106 m E, 7031541 m S).

5.5. D4 deformation (impact-induced structures)

D4 structural features include all structures formed during the crater-forming process as described in Section 2.4.1. The largest structures associated with D4 are pseudotachylitic breccia-laden folds and faults.

5.5.1. *Folds*

Impact-induced folds (F4) are hosted in quartzite units of the West Rand Group (Figure 5.2c-d). These are synclines (Figure 5.4) and synformal anticlines (Figure 5.2c-d) developed adjacent to D4 faults. The folds are characterised by narrow/chevron-type fold hinges, and wavelengths and amplitudes of up to 1000 m and 180 m, respectively; with generally north-facing fold closures (Figures 5.2, 5.4, 5.17). Fold-hinge zones host pseudotachylitic breccia, and there is evidence for rotation of the S_2 cleavage by the folds (Figure 5.17). Some D4 folds are associated with a weakly-developed, localised, axial planar, S_{4a} cleavage (e.g., Figure 5.17a). Axial planes of the F4 folds exhibit a consistent NNE to NE strike with different dip angles (Figures 5.2c-d, 5.17), which is likely a result of heterogeneous deformation, an effect of variably oriented planes (e.g., D1 rotations), or disruption by later rotations.

5.5.2. *Faults*

In the field, the few D4 fault surfaces mapped are generally sub-vertical (Figure 5.18). Equal area stereographic projections of the faults show they are radial with respect to the inferred centre of the dome (Figure 5.19). Striations on fault planes generally plunge shallowly and indicate that displacement along D4 faults was mostly strike-slip along the north-south direction (Figure 5.19b). Stereographic plots of the faults generally show normal-left-lateral displacement with eastern and southern blocks down (Figures 3.4, 5.20). Stereographic projections of the few fault planes mapped indicate that these sets are not conjugate, but rather show southward displacement of the strata (Figure 5.20b). These displacements, combined with apparent offsets derived from Google Earth™ images, suggest a general east-west-oriented compressive stress (σ_1) that bisects the obtuse angle of the faults (Figure 3.4). Some fault plane striations are parallel to bedding with a plunge that is parallel to the dip of the layers; this is indicative of dip-slip along bedding planes with a normal sense of slip (Figure 5.21). The different sense of slip indicates that the strike-slip and dip-slip faults were formed by separate stress patterns during the impact. En-echelon and conjugate sets of D4 micro-faults displace quartz veins and $M_{2(NC)}$ metamorphic assemblages (Figure 5.22), confirming their pre-impact nature.

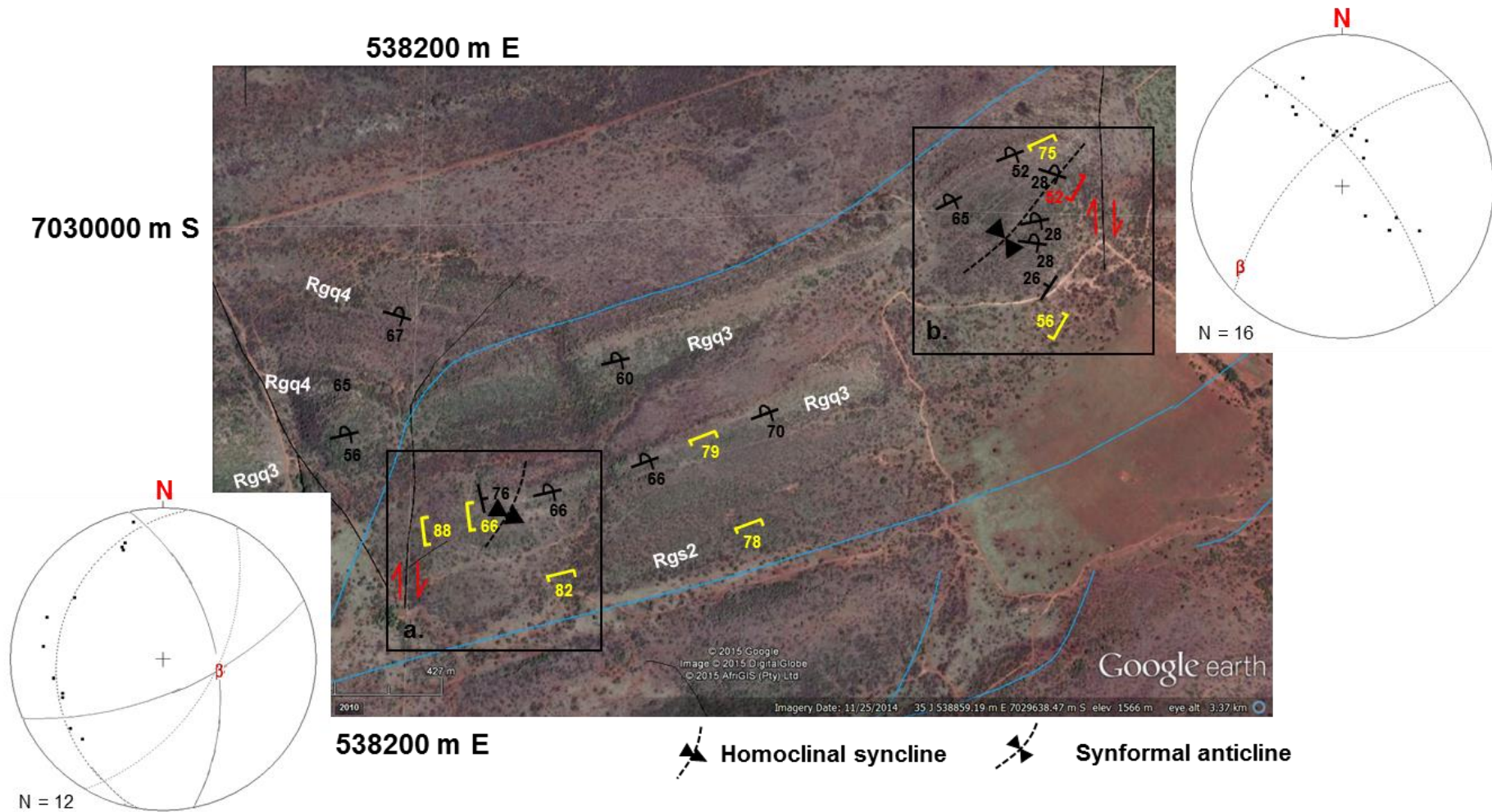


Figure 5.17. Google EarthTM photograph showing D4 folds that are bounded by impact-induced faults (black lines). The pseudotachylitic breccia-laden fold hinge shows rotation of S_2 cleavage along with the strata. Area 'a' homoclinal syncline has a fold hinge and axial plane oriented 59° on 100° and $033^\circ/61^\circ$ E respectively. Area 'b' synformal anticline has a fold hinge and axial plane oriented 16° on 232° and $045^\circ/66^\circ$ N respectively.

On a regional scale, Google Earth™ images show that impact-induced faults in the West Rand Group generally trend north-south, whereas those in the Central Rand Group generally trend northwest-southeast. Interpretation of Google Earth™ images suggests the faults are characterised by dominant left-lateral (apparent) offsets of up to 800 m in the West Rand Group (Figure 3.4).

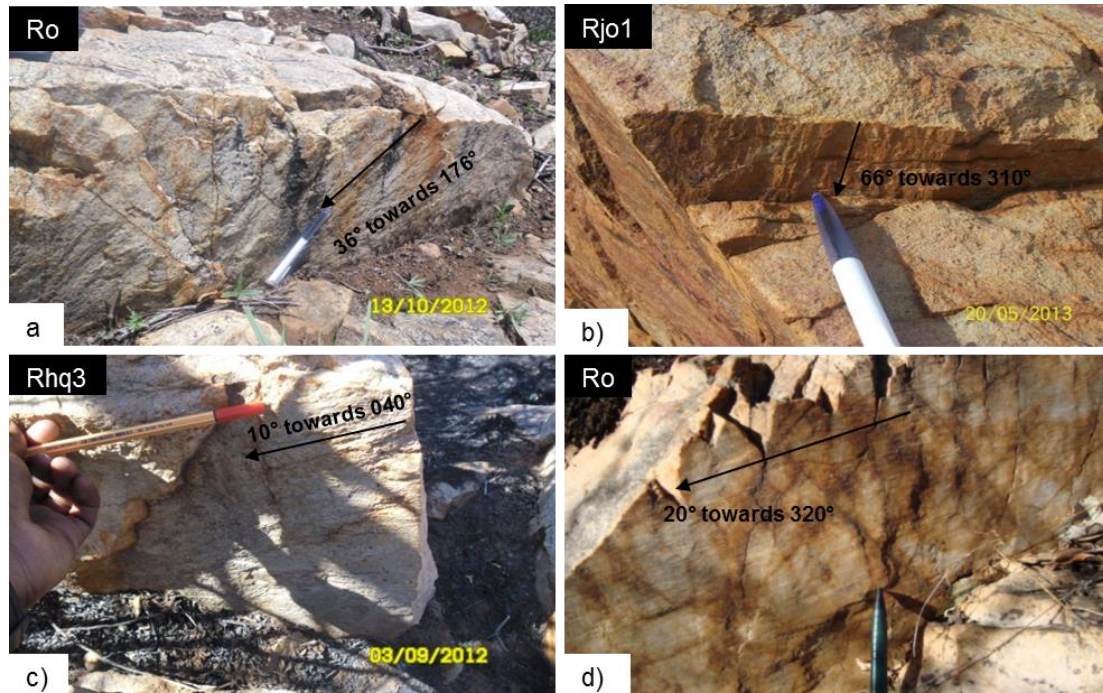


Figure 5.18. Images of striated D4 fault planes hosted in the study area. The striations indicate: (a) oblique-slip (35 J 538336 m E, 7027816 m S), (b) dip-slip (35 J 541884 m E, 7037457 m S), and (c) and (d) strike-slip (35 J 537776 m E, 7028658 m S; and 35 J 541324 m E, 7028892 m S respectively).

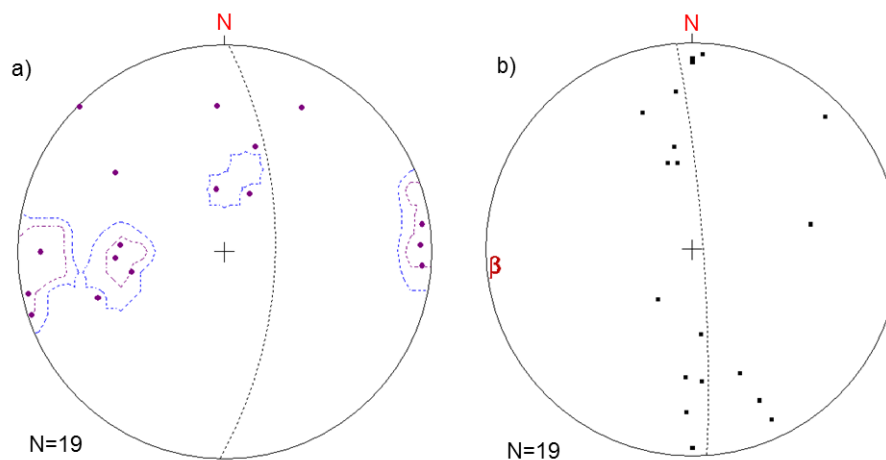


Figure 5.19. Equal area stereographic projections of: (a) poles to fault planes (mean orientation = 001°/70° E); and (b) striations developed along the fault planes.

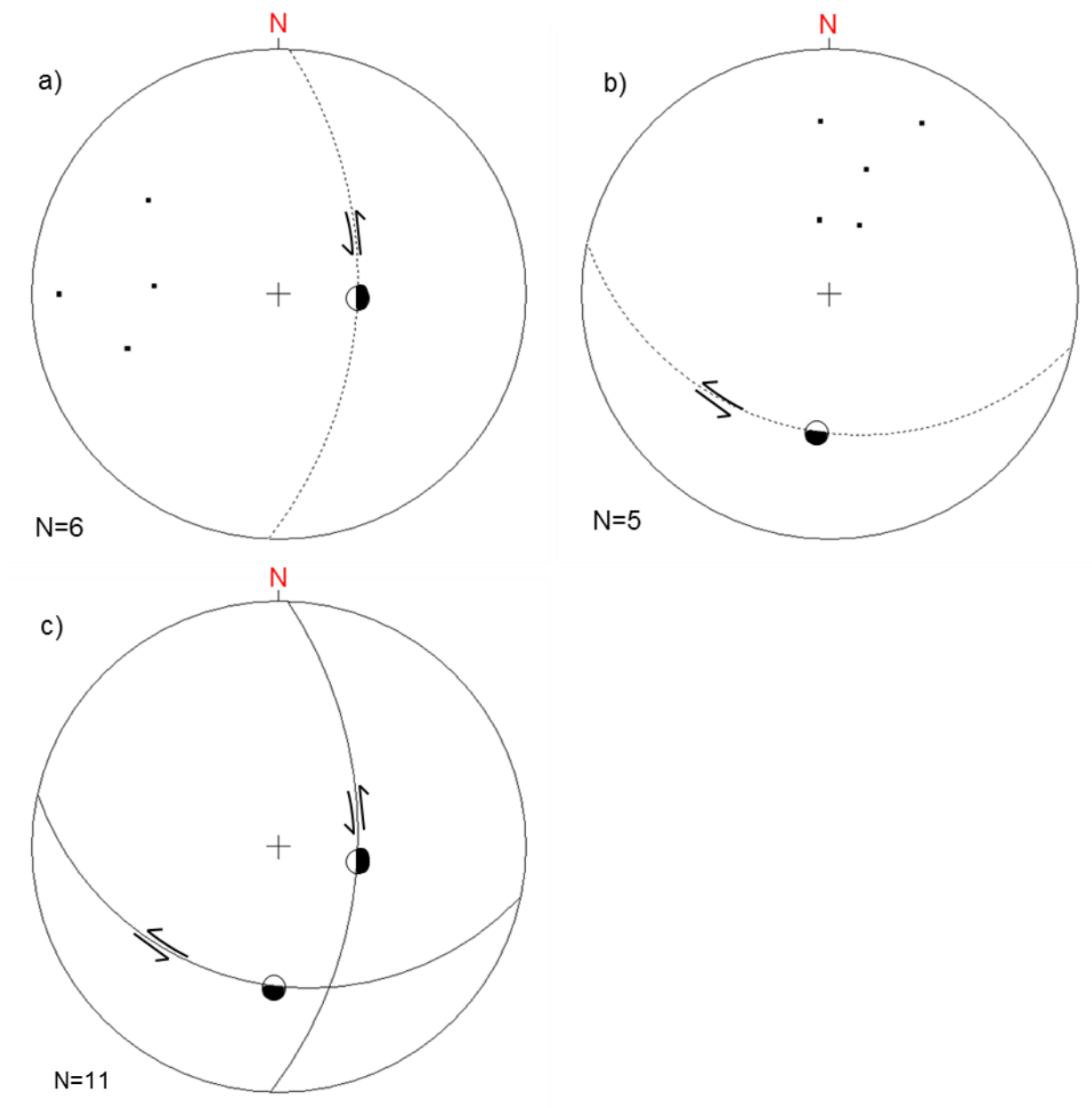


Figure 5.20. (a) & (b) Equal area stereographic projections of poles to faults showing the dominant displacements in the study area (West Rand Group). The displacements are mostly normal-sinistral, with the eastern and southern blocks collapsed. (c) Mean orientations of the displacements show consistent collapse of the south-eastern block.



Figure 5.21. Striated bedding planes indicative of dip-slip (35 J 541884 m E, 7037457 m S; and 35 J 541789 m E, 7036718 m S).

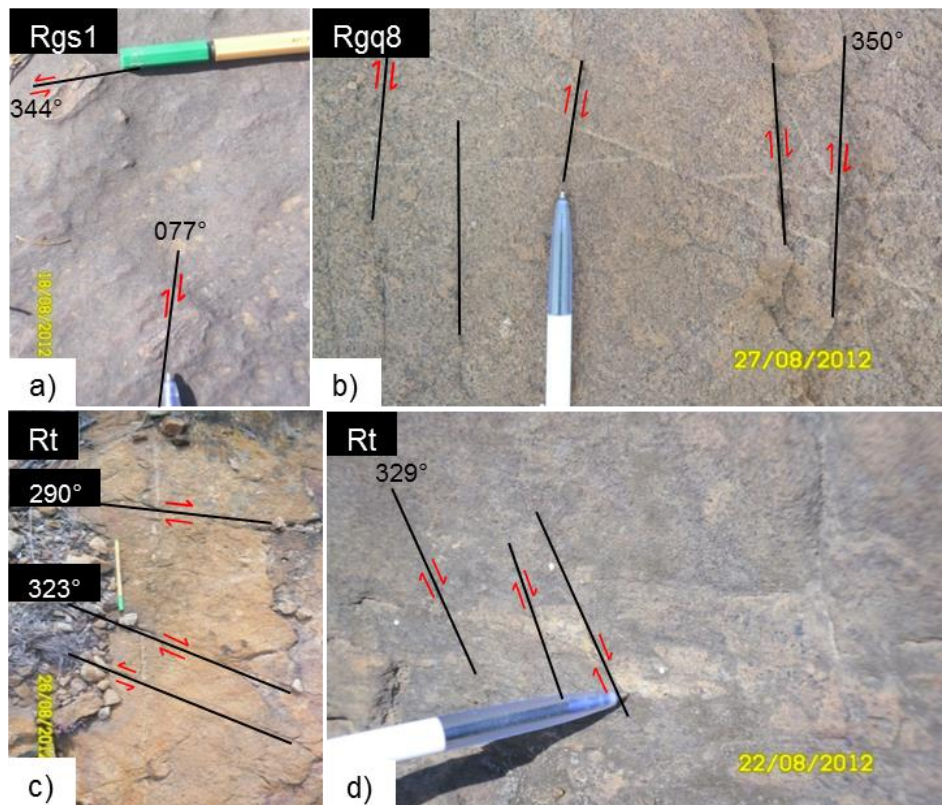


Figure 5.22. (a) Conjugate micro-faults that offset $M_{2(NC)}$ andalusite porphyroblast (35 J 538918 m E, 7034450 m S). (b), (c), and (d) En-echelon micro-faults that offset quartz veins (35 J 540395 m E, 7035628 m S; 35 J 539843 m E, 7036473 m S; and 35 J 536670 m E, 7030898 m S respectively). The micro-faults exhibit a predominantly dextral displacement.

5.5.3. *Fractures/Joints*

Based on a survey of 293 joints measured in the study area, the strike of joints is averaged to 358° (Figure 5.23). Generally, joints in the collar occur as: (a) subvertical joint sets; (b) subvertical spaced fracture cleavage; and (c) gently-dipping, bedding-perpendicular joints that generally dip to the north.

The subvertical joint sets are oriented in two main directions ranging from 350°-011° and 314°-330° (Figure 5.23). The acute angle between average orientations of the two sets is 35°, consistent with that shown by a shear fracture and tension joint combination (Ramsay and Huber, 1987); however, the sets do not exhibit distinguishing features. The subvertical joints are: (a) pervasive in the study area; (b) well-preserved; (c) generally orientated radial to the dome; and (d) locally host pseudotachylitic breccia (Chapter 7).

Spaced fracture cleavage (S_{4b}) is the most easily recognisable fracture set. It is classified as closely-spaced fractures with a typical fracture spacing <3 cm. The main fracture cleavage is very similar to the subvertical fractures in terms of orientation (Figure 5.24a), and forms a rectangular fracture pattern when it intersects with the less common, perpendicular fracture cleavage (Figure 5.24b). The S_{4b} fracture cleavage crosscuts S_2 and S_3 cleavages developed in phyllitic zones of the Rgq2 quartzite unit (Figure 5.11a). Also, $M_{2(NC)}$ assemblages are crosscut by conjugate shock cleavage in the West Rand Group (Figure A4.2, Appendix IV).

The subsidiary fracture set generally has a shallow dip towards the north (Figure 5.25), perpendicular to the dip direction of bedding. The orientation of these fractures is variable because they follow orientations of rotated bedding planes, but most of them form the east-west-trending fracture set in Figure 5.23. This fracture set hosts pseudotachylitic breccias, which occur as either injections or offshoots to/from bedding-parallel pseudotachylitic breccia. Similar to the subvertical fracture set, this association indicates that these fractures were formed during the Vredefort impact.

5.5.4. *Kink bands*

The S_2 cleavage developed along knotted quartzite horizons is cut by kink bands and pseudotachylitic breccia veins that are hosted along bedding and fractures (Figure 5.8). The kinks and pseudotachylitic breccia are oriented 057°/80° SE and 360°/78° E, respectively, both of them subparallel to the orientation of subvertical fracture sets trending 001° and 067° proximal to the outcrop. Although not mapped in the study area, Bush (2014) showed that the kink bands in the West Rand Group do host pseudotachylitic breccia. This crosscutting relationship between the kink bands, pseudotachylitic breccia and S_2 confirms that S_2 in the knotted quartzite horizons was formed before impact (Table 5.1).

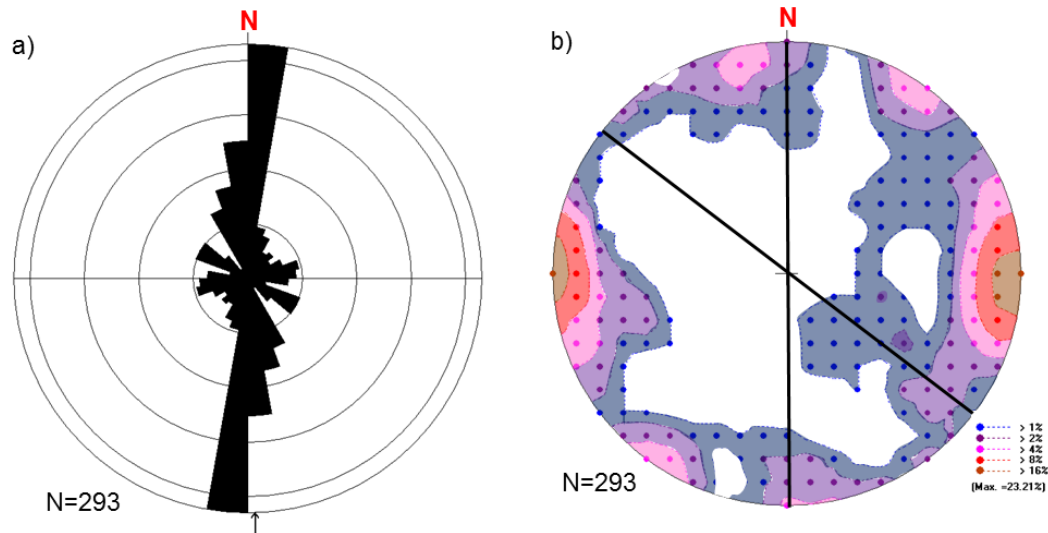


Figure 5.23. (a) A rose plot of strike orientations of D4 fractures that show an average orientation of 358° , and (b) stereographic projection of D4 fracture planes. A conjugate set of subvertical fractures orientated approximately 360° and 325° are dominant.

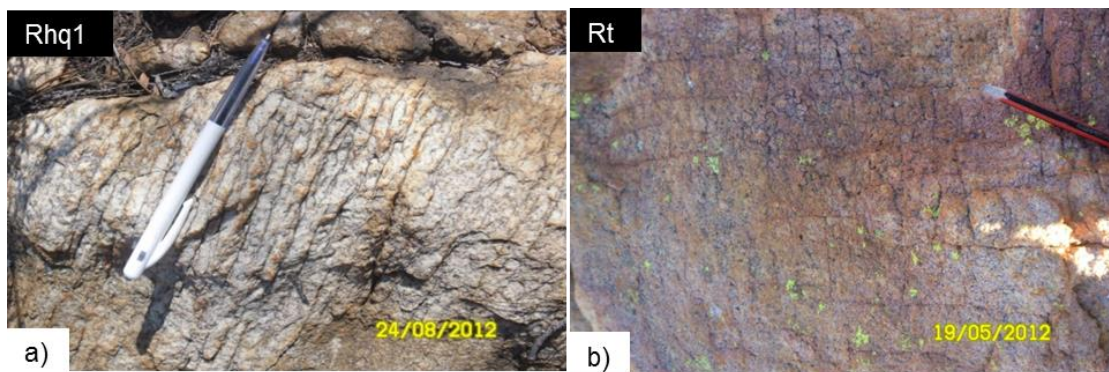


Figure 5.24. (a) Show subvertical spaced fracture cleavage with fracture spacing of 1.8 cm and 0.3 cm respectively ($35J\ 535653\ m\ E$, $7026637\ m\ S$). (b) A rectangular fracture pattern formed by perpendicular spaced fracture cleavage ($35J\ 53510\ m\ E$, $7034879\ m\ S$).

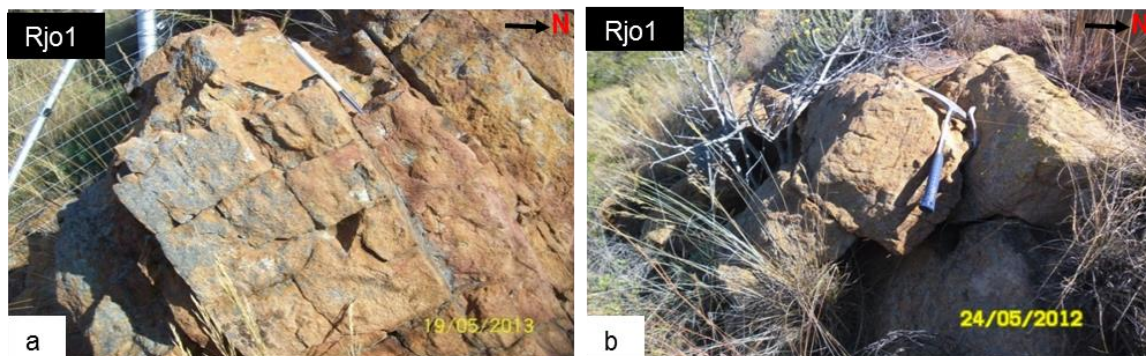


Figure 5.25. Bedding-perpendicular fractures hosted in the Rjo1 quartzite unit.

5.5.5. Shatter cones

Shatter cones are well developed in quartzite units of the Witwatersrand Supergroup, and are a prominent structural feature in the study area. Shatter cones are characterised by lineations developed along a curved surface that is usually conical in shape (Milton, 1977; French, 1998; Reimold, 1998; Wieland et al., 2006); however complete cones have not been found in the study area. The lineations converge at the apices of the cones (Figure 5.26). Most of these apices measured in the study area generally point towards the north at shallow to steep plunge angles. The length of cones mapped is up to 45 cm, and their surfaces are commonly stained with iron oxides. The curvature of half-cones is gentle in some locations, such that they could easily be mistaken for striated fault surfaces. The difference is that lineations developed along shatter cone surfaces vary more in orientation than fault striations, and the lineations are frequently curved.

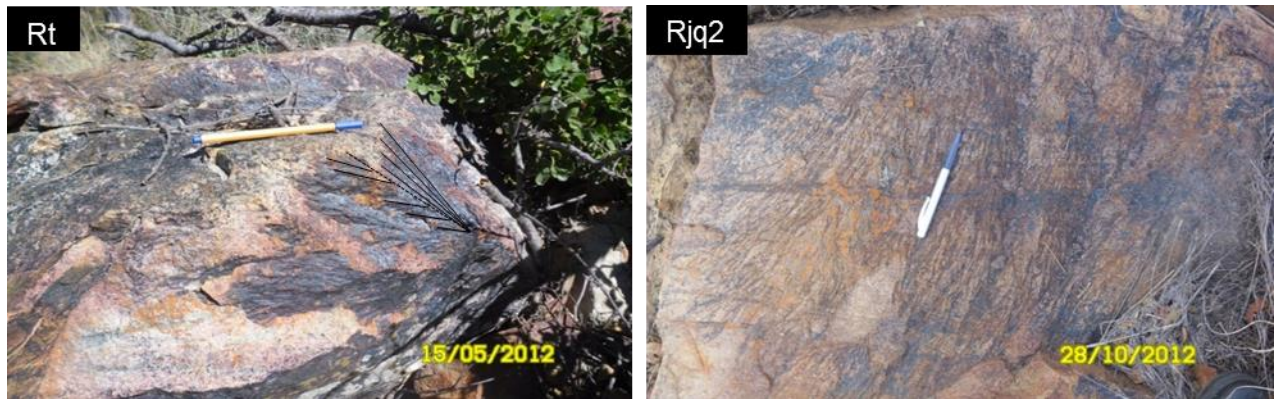


Figure 5.26. Shatter cones hosted along the Rt and Rjq2 quartzite units (35 J 534584 m E, 7034234 m S; and 35 J 536555 m E, 7031759 m S respectively).

5.6. Summary

Cross-cutting relationships of structures hosted in the study area show that a range of ductile and brittle pre-impact structures exist in the study area. These structures are represented by D1 extensional listric faults, a heterogeneously-developed D2 foliation, and D3 crenulation cleavage. The pre-impact nature of these structures is evident because they are crosscut in all cases by D4 impact-induced structural features such as faults, folds and pseudotachylitic breccia, kink bands, joints and shear fractures, and are rotated in some cases by impact-related folds.

In D1, kilometre-scale listric faults formed during ~SE-NW extension of the strata. Displacement along these faults resulted in a first-order-scale rollover fold in the northwestern block (pre-impact hangingwall), as well as drag folds in the southeastern block (pre-impact footwall). D1 structural features do not affect the <2.588 Ga Transvaal Supergroup, placing their timing as Ventersdorp-age.

The D2 deformation is characterised by a regional S_2 foliation, associated with local S_2 -C shear cleavage. The bedding-parallel shearing is represented by intrafolial quartz vein folds (V2) and S_2 -C, which are exhibited both on outcrop-scale (Figures 5.12, 5.11) and microscopically (Appendix). These features suggest that D2 deformation was characterised by NW vergence in the study area. Microscopic textures show that the D2 foliation was developed syn- to post-formation of garnet, kyanite and chloritoid porphyroblasts (e.g., Figure 5.5). The porphyroblast-cleavage microtextures and cross-cutting relationships indicate that S_2 -C and S_2 formed by progressive deformation during the peak of pre-impact metamorphism.

During D4, the meteorite impact formed structural features that crosscut D1, D2, and D3 structural features. The structures include 2nd-order scale faults, folds, pseudotachylitic breccia, joints, spaced fracture cleavage, kinks, and shatter cones. D4 also involved ~124° rotation of the strata, leading to their overturned and steep southward dips.

6.0 METAMORPHISM

6.1. Introduction

Petrographic analyses of horizons displaying the composite S_2 cleavage shows that S_2 formed at the same time as garnet, chloritoid, kyanite, cordierite, and andalusite porphyroblasts. These mineral assemblages are part of the $M_{2(NC)}$ (derived from association with the D2 deformation) metamorphic event, which reached amphibolite grade in older units of the West Rand Group and greenschist grade in the upper West Rand Group and Central Rand Group (Figure 6.1). Bisschoff (1982), Gibson (1993), Phillips and Law (1994), Gibson and Wallmach (1995), Foya (2002), and Gibson and Reimold (2008) described such grades, with andalusite, garnet, cordierite, staurolite, and biotite ($\pm$ fibrolite) peak metamorphic assemblages in the West Rand Group; whereas the Central Rand Group hosts pyrophyllite-chloritoid-chlorite-muscovite and kyanite-andalusite-chlorite-muscovite peak metamorphic assemblages.

In the study area, the Hospital Hill and Government subgroups host well-exposed $M_{2(NC)}$ porphyroblasts up to a few centimetres in size (e.g. andalusite and cordierite). The assemblages are hosted in metamorphosed pelite (Table 4.1), as well as meta-ironstone units and meta-dolerite sills. Metamorphic assemblages hosted in these units and sills are crosscut by impact-induced micro-faults and pseudotachylitic breccia (Section 7.3), indicative of their pre-impact age. Besides the West Rand Group units, knotted quartzite horizons in the Central Rand Group also host centimetre-scale porphyroblasts (e.g. andalusite- kyanite intergrowths). The knotted quartzite horizons are also displaced by impact-induced faults and kinks containing pseudotachylitic breccia (e.g., Figure 5.8; Section 5.3).

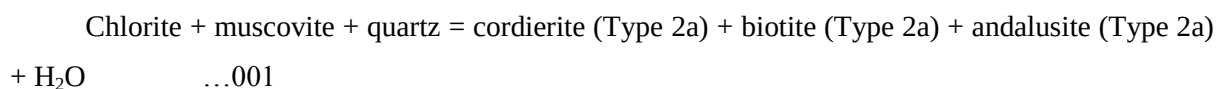
Field mapping, crosscutting relationships and petrography show that $M_{2(NC)}$ metamorphic assemblages occur regionally in the study area, decreasing in grade up-stratigraphy and that they are postdated by $M_{4(NC)}$ (impact induced) metamorphic assemblages.

6.2. Metamorphic mineral assemblages

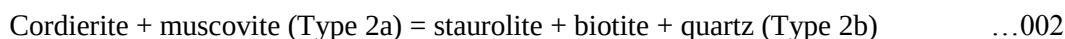
The most diverse metamorphic mineral assemblages in the collar are hosted in pelitic units and slate horizons. The major metamorphic mineral assemblages developed in these units are: (a) biotite-quartz $\pm$ andalusite $\pm$ cordierite $\pm$ garnet $\pm$ chlorite $\pm$ chloritoid $\pm$ staurolite $\pm$ magnetite $\pm$ ilmenite assemblage developed in metapelite; (b) muscovite-sericite-quartz $\pm$ kyanite $\pm$ andalusite $\pm$ chloritoid $\pm$ chlorite $\pm$ ilmenite $\pm$ rutile assemblage developed in phyllitic horizons within quartzite units; (c) grunerite-garnet-magnetite-quartz $\pm$ chlorite assemblage developed in meta-ironstone; and (d) [clinopyroxene-plagioclase]-quartz-hornblende $\pm$ actinolite $\pm$ epidote $\pm$ chlorite $\pm$ muscovite $\pm$ ilmenite $\pm$ biotite assemblage hosted in metamorphosed dolerite and basalt (Figure 6.1). The clinopyroxene and some of the plagioclase in the meta-dolerite are relict igneous minerals. The distribution of the metamorphic assemblages shows that amphibolite-grade was attained in units older than the Rgq6 quartzite unit, and greenschist-grade attained in younger units (Figures 3.5, 6.1). This transition is evidenced by the absence of hornblende, grunerite, garnet, and cordierite in metamorphic assemblages hosted north of the Rgq6 quartzite unit (Figure 6.1). The metamorphic minerals are amphibolite- to greenschist-grade facies Type 2 (attributed to $M_{2(NC)}$ metamorphism) and/or Type 4 attributed to impact heat (Type 4b) with precursor shock features (Type 4a).

6.2.1. Biotite-quartz $\pm$ andalusite $\pm$ cordierite $\pm$ garnet $\pm$ chlorite $\pm$ muscovite $\pm$ chloritoid $\pm$ staurolite $\pm$ magnetite $\pm$ ilmenite (amphibolite-grade facies)

Metapelite units have the most diverse metamorphic mineralogy in the study area, owing to the variety of primary compositions. Most of the units show a biotite-andalusite-cordierite-quartz assemblage, which is complemented by a number of subordinate minerals that may include garnet, chlorite, muscovite, chloritoid, staurolite, magnetite and/or ilmenite (Table A3.2, Appendix III). Judging from the observation that there is minimal chlorite in these metapelite samples, the dominant biotite-andalusite-cordierite assemblage was formed through reaction 001. These mineral assemblages show dominant prograde textures (Type 2a minerals), with retrograde reactions (Type 2b minerals) evident in some samples. Prograde textures are evidenced by the replacement of lower temperature minerals by peak metamorphic assemblages (e.g., biotite $\rightarrow$ garnet), whereas retrograde reactions are evidenced by the breakdown of the peak assemblages to lower temperature minerals (e.g., hornblende $\rightarrow$ actinolite).



Cordierite mostly occurs as subidioblastic Type 2a porphyroblasts that show prograde textures, with subordinate Type 4b cordierite coronae that rim the Type 2a cordierite and overprint the S₂ cleavage (Figure 6.2). In contrast to Type 4b cordierite, Type 2a cordierite porphyroblasts are wrapped by the S₂ biotite-quartz cleavage. They also enclose Type 2a garnet in some samples (Figure 6.3). Type 2a garnet was, thus, included by Type 2a cordierite during prograde metamorphism. The cordierite was subsequently replaced by Type 2b biotite + staurolite, a reaction described by Gibson and Wallmach (1995) as a high-temperature, fine-grained retrograde reaction 002.



The staurolite-biotite replacement assemblage is itself replaced by Type 4b cordierite and biotite, which have overgrown S₂ along the porphyroblast margins (Figures 6.2, 6.3). This is a later cooling reaction that is the inverse of reaction 002.

The dominant Type 2a garnet (up to 4 mm in diameter) is generally subidioblastic to idioblastic and hosts a relict S_X fabric defined by quartz inclusion trails, which is generally oblique to the matrix foliation (Figure 6.4). This relationship is similar in timing to the S₁ and S₂ relationship described by Gibson (1993) in West Rand group metapelites. The garnet porphyroblasts also host fine-grained idioblastic magnetite, and are wrapped by the S₂ cleavage (Figure 6.4a; Section 5.3.1). Type 2a garnet commonly replaces Type 2a biotite, indicating a prograde reaction (Figures 6.4c and d). Disintegrated garnet fragments are hosted in the groundmass of irregular pseudotachylitic breccia in the Hospital Hill Subgroup, indicative of its pre-impact nature (e.g., Sample PS7 in Table A3.7, Appendix III).

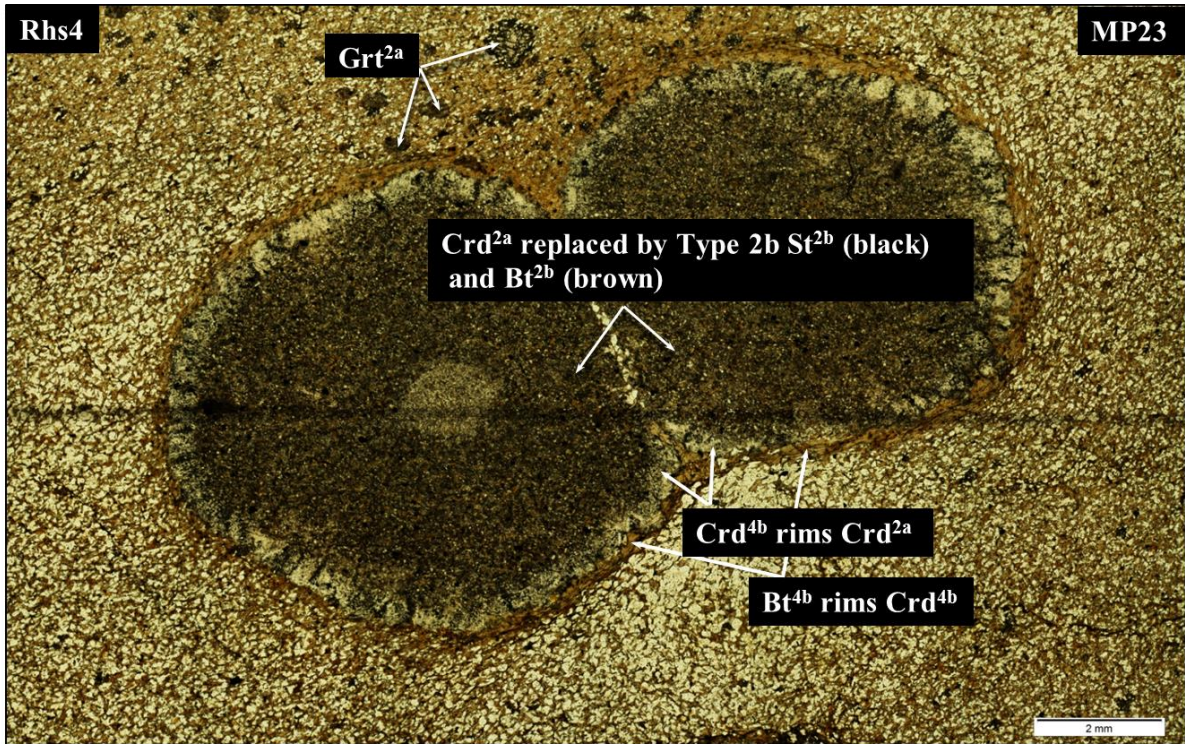


Figure 6.2. Idioblastic Type 2a cordierite replaced by Type 2b biotite and staurolite. Type 4b cordierite rims Type 2a cordierite, and is in turn rimmed by Type 4b biotite (PPL). Type 4b cordierite and biotite overgrow the S_2 cleavage.

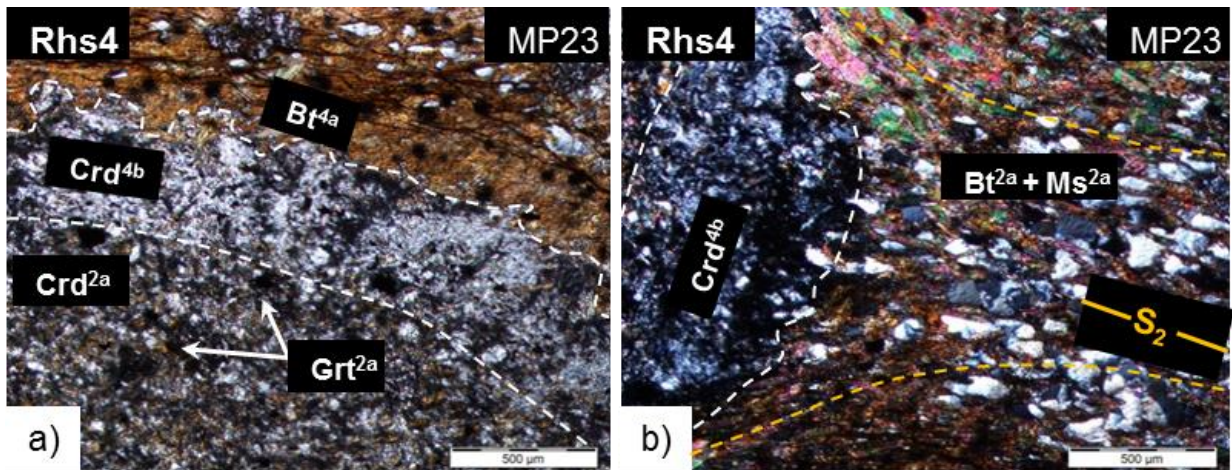


Figure 6.3. (a) Type 2a garnet included in Type 2a cordierite (PPL). (b) Replaced edges of the Type 2a cordierite porphyroblast wrapped by a syn-kinematic (Type 2b biotite) S_2 cleavage (XPL). Type 4b cordierite and biotite have overgrown S_2 in (a) the mid-bottom part of (b).

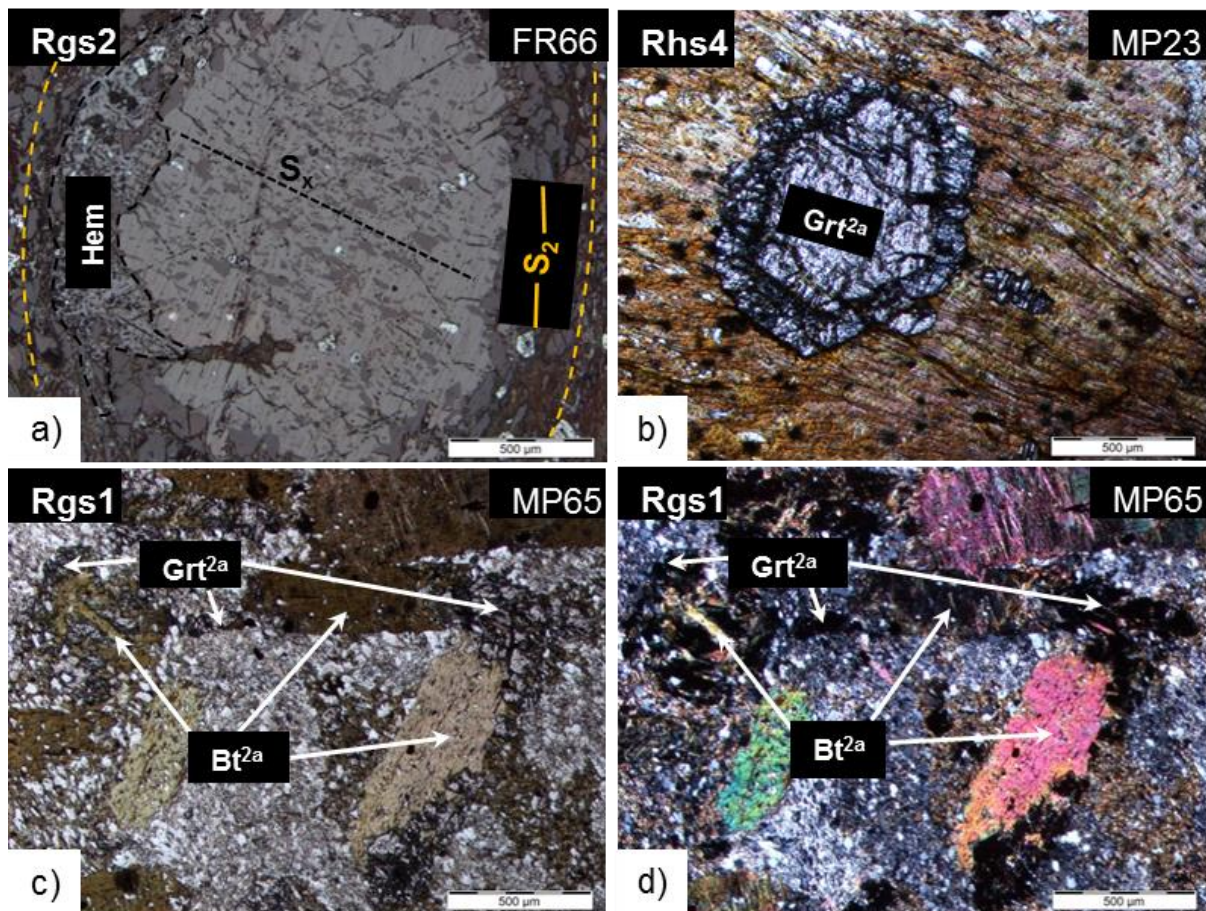


Figure 6.4. (a) Type 2a garnet porphyroblast with planar S_x quartz inclusion trails. Oblique to external S_2 cleavage (RL). (b) Idioblastic Type 2a garnet exhibiting overgrowth (PPL). (c) & (d) Type 2a biotite replaced by Type 2a garnet (PPL and XPL, respectively).

Biotite is classified into Type 2a porphyroblasts (up to 3 mm long), smaller aggregates defining the S_2 foliation, Type 2b biotite that replaces cordierite, and Type 4b biotite that rims Type 4b cordierite (Figure 6.2). Type 2a porphyroblasts are skeletal and subidioblastic, commonly included in Type 2a andalusite (e.g., Rgs1 metapelite; Figure 6.5a). The porphyroblasts are replaced by Type 2a garnet (prograde) and chlorite (retrograde; Table A3.2, Appendix III). Type 2a biotite partly defines the S_2 cleavage, which wraps garnet, chloritoid, and cordierite porphyroblasts (e.g. Figure 6.5b. Type 2b biotite and staurolite replace Type 2a cordierite porphyroblasts, representative of a fine-grained retrograde reaction (Gibson and Wallmach, 1995; Figure 6.3). This texture is hosted in samples of the lower Rhs4 metapelite unit and the Government Subgroup. The staurolite is skeletal and very fine-grained ($<100\ \mu\text{m}$), as a result it exhibits a dark grey, almost opaque, pleochroism instead of the normal yellow pleochroism (Figure 6.6).

Andalusite is developed in the Rhs4, Rhs5, Rgs1, and Rgs2 metapelite units. It commonly occurs as Type 2a porphyroblasts (up to 5 cm in size) that host Type 2a biotite (Figure 6.7a). These porphyroblasts contain inclusion trails that display relative rotation to the syn-to post-metamorphic S_2

cleavage (Samples S1 and S2 in Table A3.6, Appendix III). The syn-kinematic texture is also shown by distortion of the S_2 foliation in and around Type 2a chloritoid porphyroblasts (Figure 6.7c). In the Rgs2 metapelite, chloritoid is replaced by fine-grained Type 2b muscovite (Figure 6.7d).

Muscovite is developed as Type 2a Muscovite matrix (locally fibrous) that partly defines the S_2 cleavage (Figure 6.3a) and Type 2b muscovite polymorph. Type 2b muscovite is hosted as partial replacements of skeletal chloritoid porphyroblasts in the Rgs2 metapelite unit (Figure 6.7d). Type 2a sericite-muscovite defines the S_2 cleavage in foliated horizons of the West Rand and Central Rand groups. Type 2a sericite-muscovite cleavage shows relative rotation along edges of andalusite in the pre-Ro metapelite (samples S1 and S2 in Table A3.6, Appendix III).

Magnetite is developed in Fe-rich metapelite units of the West Rand Group. It is hosted as inclusions in Type 2a garnet porphyroblasts, and as subidioblastic porphyroblasts hosted in the matrix of the Rgs2 metapelite (Figure 6.8a). The size of magnetite increases into the matrix, consistent with increasing temperatures associated with prograde metamorphism. Hematite forms part of Fe-oxide alteration of Type 2a garnet + chloritoid porphyroblasts (Figure 6.8b).

Chlorite is developed in the matrix of pelitic units and slate horizons hosted in the West Rand Group. Chlorite is classified into Type 2 assemblages that form part of $M_{2(NC)}$, and Type 4 chlorite that overprints impact-induced pseudotachylitic breccia. Type 2a chlorite is developed in the Central Rand Group as oxidised skeletal porphyroblasts, and retrograde Type 2b skeletal chlorite replaces Type 2a biotite in the Hospital Hill and Government subgroups (Figures 6.9a). Type 4 chlorite occurs as radiating skeletal crystals, forming rosettes that overprint the aphanitic pseudotachylitic breccia groundmass (Figure 6.9b). The chlorite rosettes overprint the groundmass of pseudotachylitic breccia hosted in the Hospital Hill Subgroup, and, thus post-date impact-induced pseudotachylitic breccia. This is indicative of thermal metamorphism ($M_{4b(NC)}$) after the impact, as a result of residual heat generated by the impact ($M_{4(REG)}$; Gibson et al., 1997a, 1998).

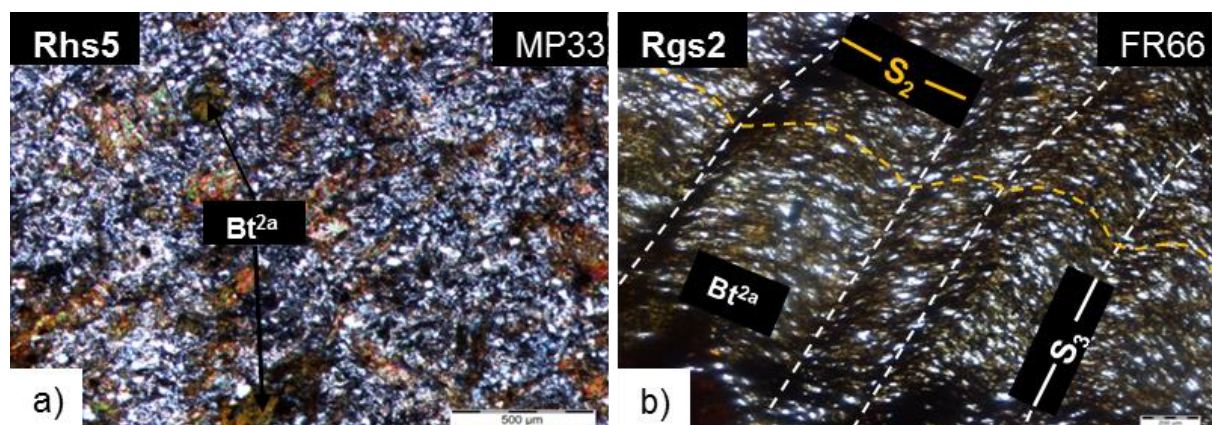


Figure 6.5. (a) Elongate Type 2a biotite porphyroblasts hosted in metapelite. (b) A crenulated S_2 cleavage composed of Type 2a biotite and quartz assemblage (Both under XPL).

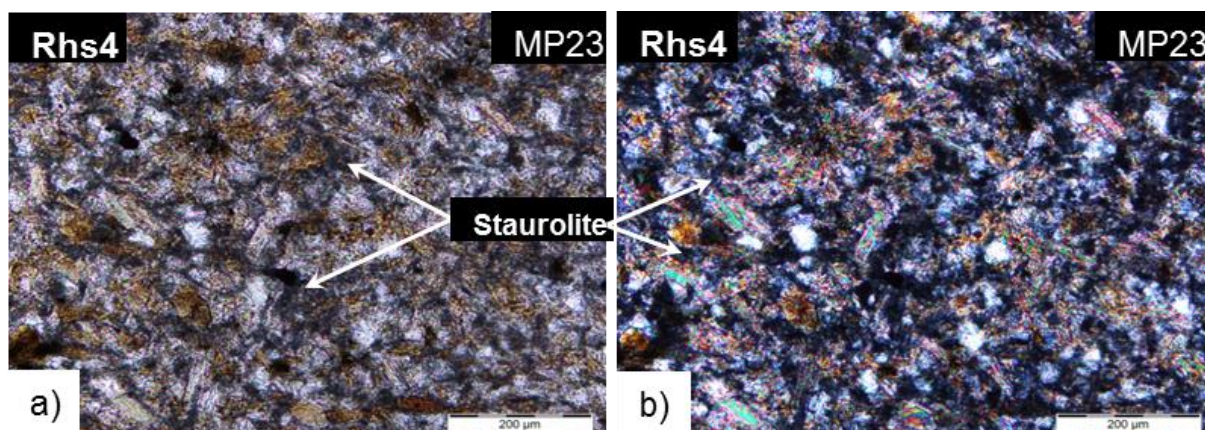


Figure 6.6. (a) & (b) Type 2a cordierite porphyroblast replaced by fine-grained Type 2b staurolite and biotite (PPL and XPL, respectively).

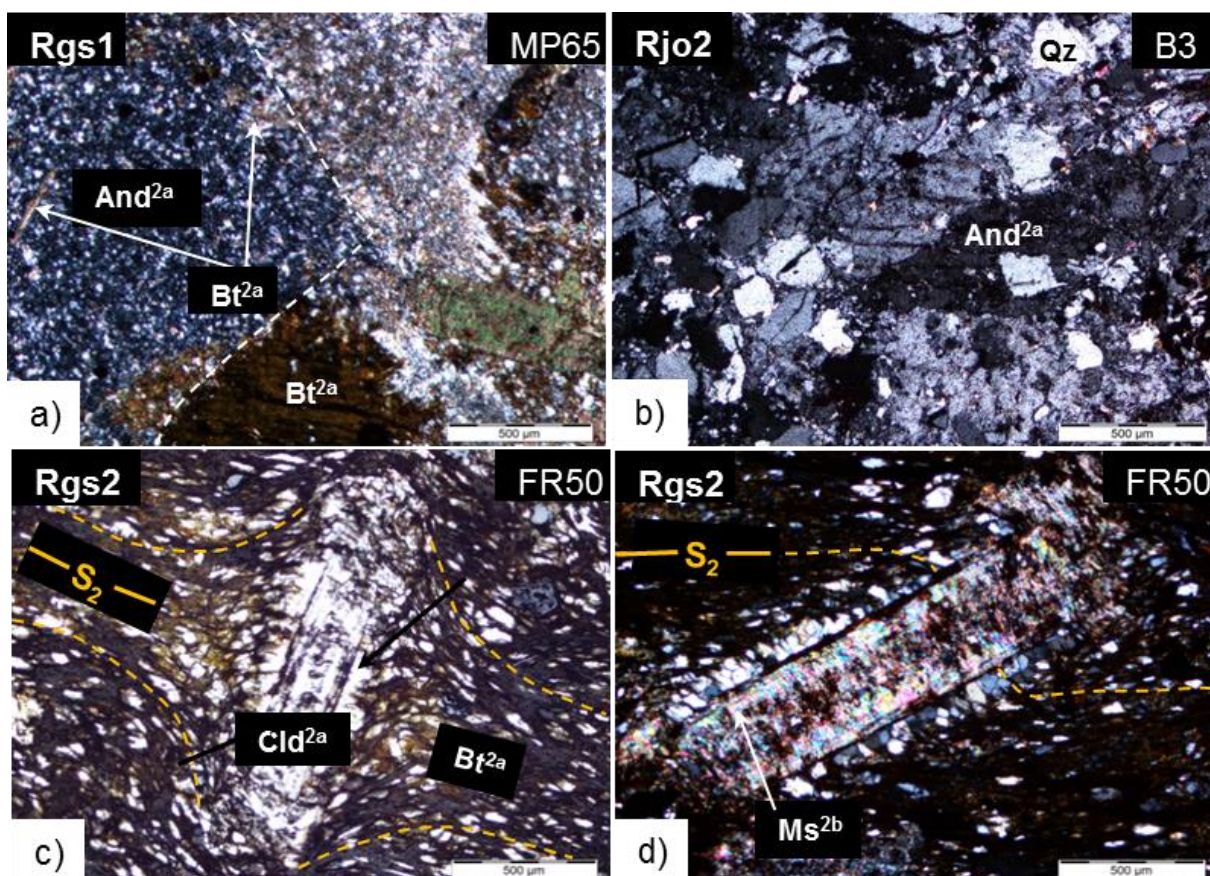


Figure 6.7. (a) Type 2a biotite inclusions in idioblastic Type 2a andalusite porphyroblasts hosted by metapelite units of the West Rand Group (XPL). (b) Xenoblastic Type 2a andalusite hosted in knotted quartzite horizons (XPL). (c) Type 2a chloritoid porphyroblast wrapped by a syn to post-kinematic S_2 cleavage (RL+XP). (d) Type 2a chloritoid porphyroblast replaced by Type 2b muscovite (XPL).

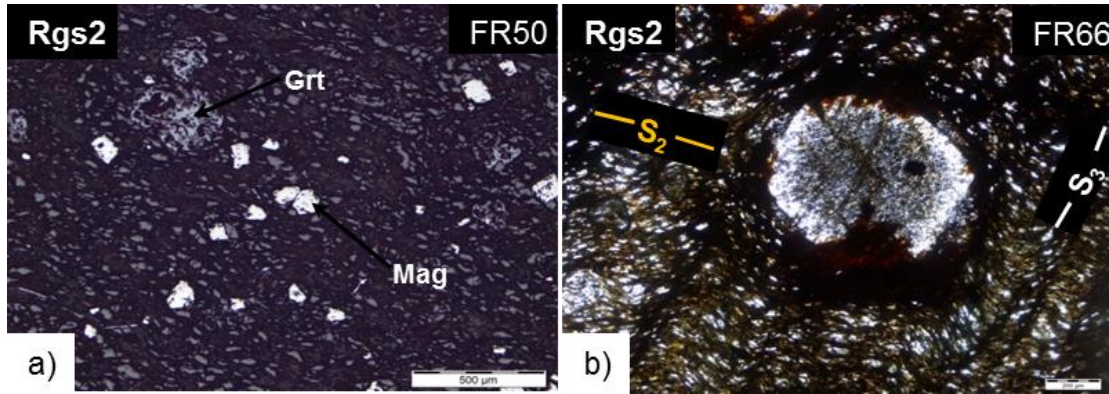


Figure 6.8. (a) Magnetite porphyroblasts hosted as Type 2a garnet inclusions and in the matrix of Rgs2 metapelite (RL). (b) Type 2a garnet rimmed by hematite (PPL).

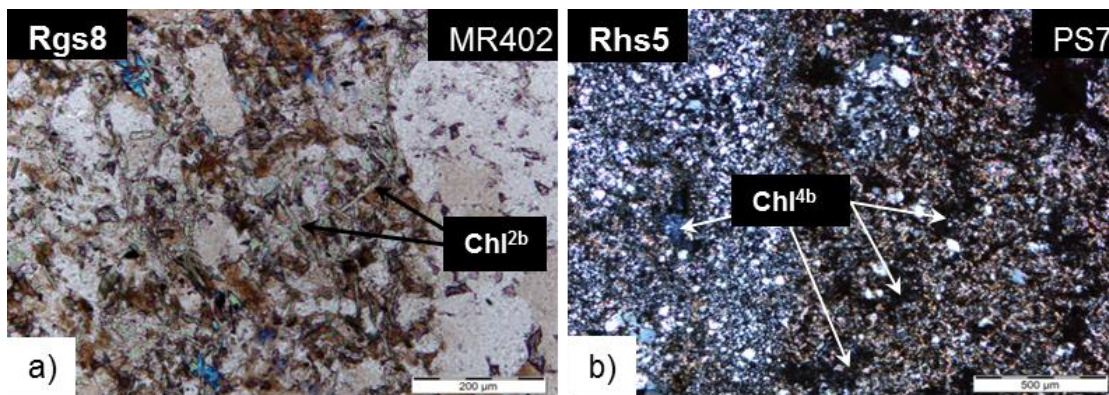


Figure 6.9. (a) Skeletal Type 2b chlorite hosted as skeletal replacements of Type 2a biotite porphyroblasts (PPL). (b) Type 4b chlorite rosettes overprint compositionally layered pseudotachylitic breccia (XPL).

6.2.2. *Muscovite-sericite-quartz ± kyanite ± andalusite ± chloritoid ± chlorite ± ilmenite (generally greenschist-grade facies)*

In the Central Rand Group very few bulk rock compositions conducive to metamorphic mineral growth are found. Type 2a sericite and muscovite define the S₂ cleavage developed in phyllitic horizons in quartzite units and in knotted quartzite horizons (e.g., Figure 5.9a). In knotted quartzite horizons, the cleavage shows relative rotation against Type 2a kyanite porphyroblasts, which contain a poorly-defined internal foliation (Figure 5.9a). The subidioblastic kyanite porphyroblast grew during D2. This is a similar texture that Type 2b biotite-bearing S₂ forms around Type 2a garnet in the Government Subgroup (e.g., Figure 5.14). These similarities show that S₂ cleavage and M_{2(NC)} assemblages formed in both the West Rand and Central Rand groups as the syn-tectonic M_{2(NC)} event evolved. Sample B3 hosts Type 2a kyanite that is hosted as irregular inclusions in Type 2a andalusite (Figure 6.10). The lack of replacement reactions suggests that kyanite and andalusite most likely formed coeval, through reaction 003; judging from the lack of pyrophyllite in these samples.

Chloritoid porphyroblasts developed in the Rgs9 and knotted quartzite horizons are highly oxidised and consequently, do not show any clear textural relationships with S₂ cleavage (Figure 6.11).

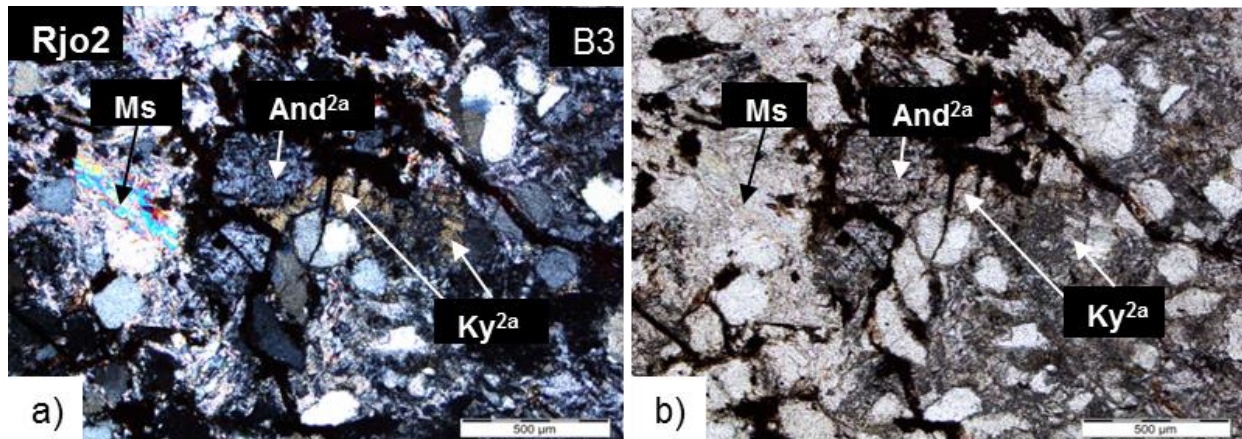


Figure 6.10. Type 2a kyanite hosted in sample B3 as irregular inclusions in Type 2a andalusite (XPL and PPL respectively).

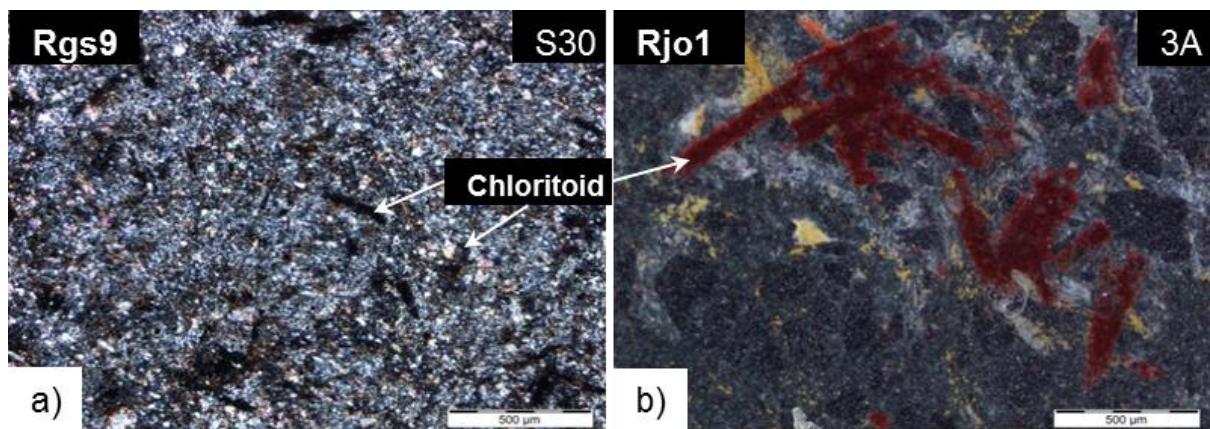


Figure 6.11. Skeletal chloritoid developed in (a) slate (XPL) and (b) knotted quartzite (Natural/Ambient light=NL).

6.2.3. Grunerite-garnet-magnetite-quartz (amphibolite-grade facies)

Type 2a grunerite, garnet, and magnetite are developed as subidioblastic and skeletal porphyroblasts in samples of the Rhs2/3 meta-ironstone (Table A3.3, Appendix III). Grunerite is developed as radiating to sheaf-like aggregates (Figure 6.12c). Subidioblastic Type 2a garnets are wrapped by the S₂ cleavage, showing a similar texture to other Type 2a garnets developed in Fe-rich metapelites. The porphyroblasts host poorly-developed inclusion trails and are displaced by microfaults (see Sample 13D4 in Appendix III). At least two growth zones in garnet were described

by Bisschoff (1982) and Gibson (1993), however only one zone was seen in these samples (e.g., Figure 6.4b).

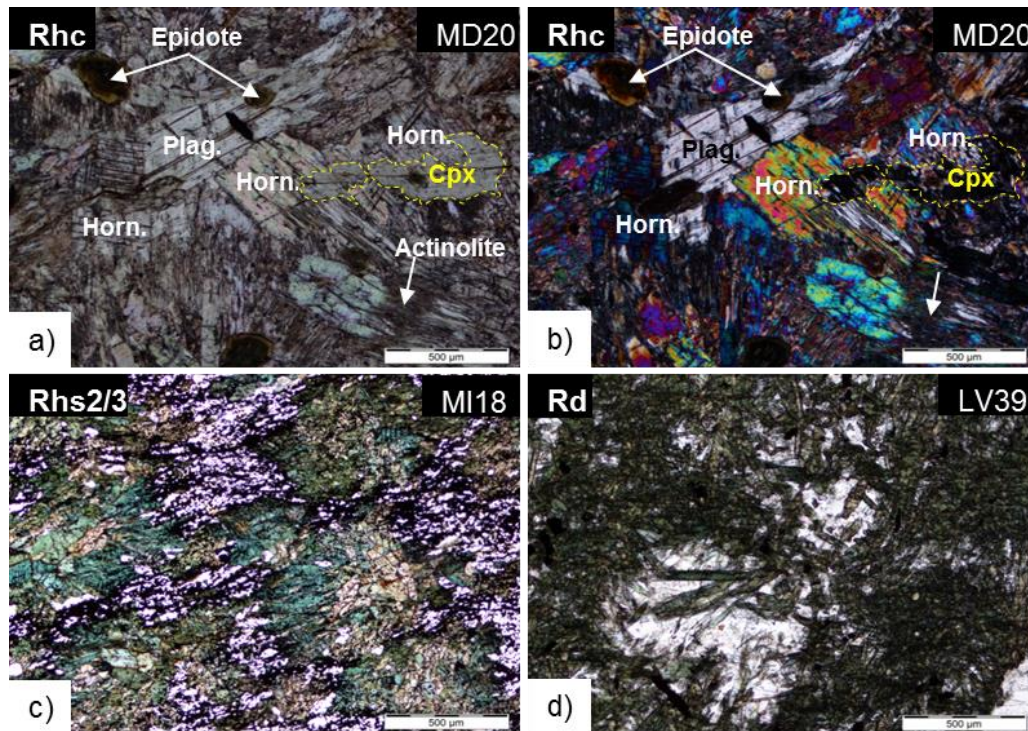


Figure 6.12. (a) & (b) Clinopyroxene and plagioclase replaced by epidote and hornblende in metadolerite sample MD20. The hornblende is in turn replaced by fibrous actinolite (PPL and XPL respectively). (c) Subidioblastic and radiating skeletal grunerite developed in meta-ironstone (PPL). (d) Skeletal hornblende developed in the Dominion Group meta-basalt (PPL).

6.2.4. Hornblende-plagioclase-quartz $\pm$ actinolite $\pm$ epidote $\pm$ chlorite $\pm$ ilmenite $\pm$ biotite (amphibolite-grade facies)

Hornblende, epidote, and actinolite are mainly developed in meta-dolerite sills hosted in the West Rand Group. Hornblende and epidote are developed as subidioblastic porphyroblasts in a meta-dolerite sill intruded along the Rhc BIF (Figure 6.12). Skeletal and subidioblastic hornblende porphyroblasts are developed in a meta-dolerite sill intruded into the Rhc, Rgs1, and Rgs7 unit. Actinolite is hosted as skeletal and radiating crystals up to 5 cm in length in the meta-dolerite sill intruded along the Rhs metapelite (35 J 536105 m E, 7027085 m S). Fibres of actinolite replace hornblende in sample MD20 (Figure 6.12a), showing retrograde reaction textures. Hornblende is also developed in meta-basalts of the Dominion Group (amphibolite-grade) and actinolite in the Klipriviersberg Group (greenschist-grade). Plagioclase and chlorite are generally hosted as subidioblastic to interstitial minerals (e.g., Figure 6.12a-b).

6.2.5. *Paragenesis*

The interpretation of metamorphic mineral textures generally shows that the supracrustal strata were affected by $M_{2(NC)}$ peak metamorphism (pre-impact) and $M_{4(REG)}$, syn- and post-impact, metamorphism. Metamorphic mineral textures show that the $M_{2(NC)}$ metamorphism is characterised by prograde and retrograde assemblages, and that $M_{4(NC)}$ peak conditions were slightly lower than $M_{2(NC)}$ (equivalent to $M_{4(REG)}$ and $M_{3(REG)}$, respectively; Gibson and Wallmach, 1995).

6.2.5.1. $M_{2a(NC)}$ prograde metamorphism

Metamorphic minerals formed during the prograde phase are generally idioblastic to subidioblastic (Type 2a); they may be partially replaced by a Type 2b paragenesis. Some of these assemblages (e.g., Type 2a garnet, cordierite, andalusite, and kyanite porphyroblasts) are wrapped by the regional S_2 cleavage (Figures 6.2, 6.4a, 6.7c, 6.8b). The fact that all these Type 2a minerals are crosscut by impact-induced features (Section 7.3) constrains them to the pre-impact, $M_{2a(NC)}$, metamorphism. These assemblages are the earliest metamorphic minerals hosted in the collar rocks of the study area, together with Type 2a biotite, sericite + muscovite, quartz, hornblende, epidote, and magnetite. The prograde nature of $M_{2a(NC)}$ porphyroblasts is shown by: (a) the outward increase in size of magnetite inclusions in Type 2a garnet in samples FR50 and FR66 (e.g., Figure 6.8a); (b) replacement of lower grade hydrous minerals by higher grade minerals (e.g. Type 2a biotite $\rightarrow$ Type 2a garnet in Figure 6.4c). The syn-kinematic nature of $M_{2a(NC)}$ to D2 is shown by the relative rotation of S_2 along edges of Type 2a garnet and kyanite porphyroblasts (Figures 5.5 and 5.9a, respectively).

6.2.5.2. $M_{2b(NC)}$ retrograde metamorphism

Mineral assemblages formed by the retrograde phase partially replace $M_{2a(NC)}$ porphyroblasts. These are $M_{2b(NC)}$ assemblages, which include Type 2b biotite and staurolite that replace Type 2a cordierite porphyroblasts (Figure 6.3), representative of a fine-grained retrograde reaction described by Gibson and Wallmach (1995). Retrograde reactions are also evidenced by the replacement of hornblende by actinolite in amphibolite-grade mafic rocks (e.g., Figure 6.12a). These reactions represent decreasing metamorphic conditions after $M_{2a(NC)}$ prograde metamorphism had reached its peak pressure and temperature.

6.2.5.3. $M_{4(NC)}$ metamorphism

Impact-related metamorphism is termed $M_{4(NC)}$ owing to its association with the D4 impact deformation, and it can be divided into two: (a) $M_{4a(NC)}$ shock metamorphism that occurred together with the impact; and (b) $M_{4b(NC)}$ thermal metamorphism that is a result of residual heat after the impact, characterised by temperatures up to 525 °C in the West Rand Group (Gibson et al., 1998; Gibson, 2002) and 300 °C in the Central Rand Group (Gibson et al., 1998). $M_{4b(NC)}$ features are shatter cones (e.g., Figure 5.26), PDF, and coesite and stishovite at pressures of 20 – 120 kbar (Reimold, 1993; Gibson and Wallmach, 1995). These features have not been analysed further because they are beyond the scope of this study. $M_{4b(NC)}$ metamorphic assemblages are represented by: (a) recrystallized pseudotachylitic breccia matrix; and (b) annealed impact-induced micro-faults (Figure 6.13). $M_{4b(NC)}$ is also evidenced by Type 4b chlorite rosettes that overprint pseudotachylitic breccia hosted in the Hospital Hill and Government subgroups (Figure 6.9b); as well as Type 4b cordierite and biotite coronae that rim Type 2a cordierite and overprint the S_2 cleavage (Figures 6.2, 6.3).

6.3. Metamorphic grade distribution across the Karolina Fault

Part of the aims of this study was to test whether there is a significant change in the grade of metamorphism in units located across the Karolina Fault (Figure 3.4). Nemutandani (2014) and Bush (2014) showed that West Rand Group units ~10 km east of the study area do show significantly lower grade metamorphic assemblages. Metapelitic units of the Hospital Hill and Lower Government subgroups in the study area host coarse-grained metamorphic assemblages (amphibolite-grade facies); however the Karolina Fault does not significantly displace these units owing to its listric nature (Figure 3.4). To compensate for this inconvenience, samples of the stratigraphically younger units were collected for metamorphic petrology. These units include meta-diamictite, cleaved horizons, meta-dolerite, argillaceous quartzite (all located in the Government and Jeppestown subgroups), knotted quartzite horizons (Central Rand Group), and slate (Government and Turffontein subgroups); whose sample locations are shown in Figure A3.1. In general, samples of these units/horizons show little or no change in metamorphic grade across the Karolina Fault; with samples from the knotted quartzite showing some contrast in metamorphic grade.

Samples KT12, 60, 3A, B3, KT1, KT44, KT72 and KT92 from knotted quartzite horizons were petrographically analysed. These samples are composed of 30-62% primary quartz, 5-35% sericite $\pm$ muscovite (cleavage), 0-50% kyanite and andalusite, 2-15% chloritoid, and 0-15% goethite (Table A3.5, Appendix III). Samples 60, 3A, B3, KT44, and KT72, which were collected from the pre-impact hangingwall of the Karolina Fault, host kyanite and/or andalusite proportions ranging from 24-50% (western block; Table A3.5). In contrast, sample KT12, the only sample collected from the pre-

impact footwall of the Karolina Fault (~2.7 km offset from the former samples), has a finer matrix and no andalusite/kyanite porphyroblasts (Figure 6.14). Sample KT12 is located at a similar stratigraphic location as samples B3 and KT72; however it does not host andalusite/kyanite (Figure 6.14b), indicating the units were possibly displaced before $M_{2(NC)}$ metamorphism. What the study did not cover is the analysis of bulk compositions of these samples, to prove/disprove whether the primary compositions of these samples are similar; which will be conclusive evidence for/against the suggestion that $M_{2(NC)}$ assemblages are discordant across the Karolina Fault. Perhaps this is a worthwhile investigation that needs to be conducted in the study area going forward.

The rest of the sampled units/horizons do not show any clear difference in metamorphic grade (Table A3.2-3.6, Appendix III). Meta-diamictite samples MR402 and MR347 are composed of the assemblage quartz-chlorite-biotite-sericite, showing little textural difference. Samples of a phyllitic quartzite (Rhq8) horizon from either side of the Karolina Fault host a similar muscovite-sericite-quartz assemblage of the S_2 cleavage. Samples S08, S15, and S17 sampled from a slate unit within the Kimberley Reefs (Rt) show poorly-developed mineral textures. This is because they are masked by goethite, making it hard to compare the samples. Pelitic laminations within the Rgq8 argillaceous quartzite are <0.5 mm, and do not show mineral assemblages large enough to be compared/contrasted.

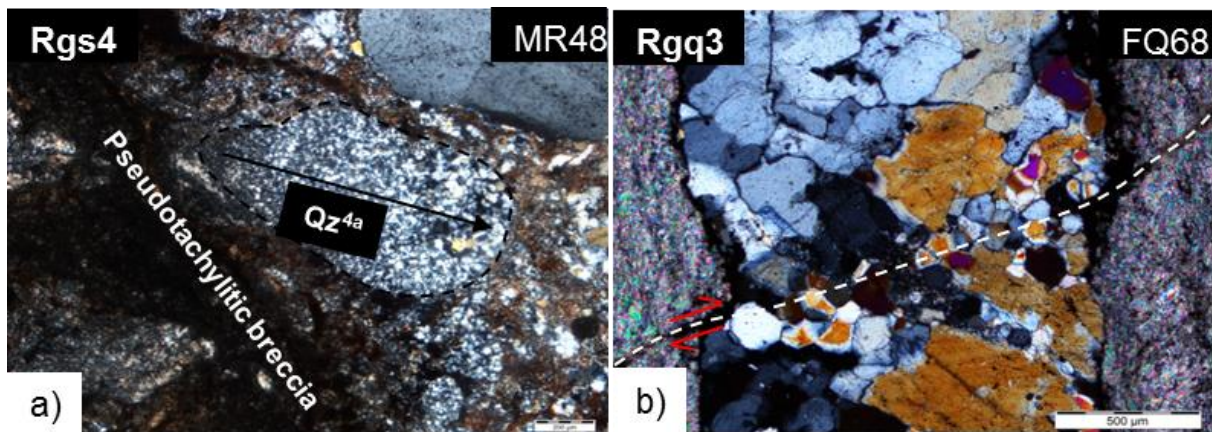


Figure 6.13. (a) Type 4a recrystallised quartz associated with pseudotachylitic breccia. (b) Impact-induced microfault annealed by Type 3a quartz (all under XPL).

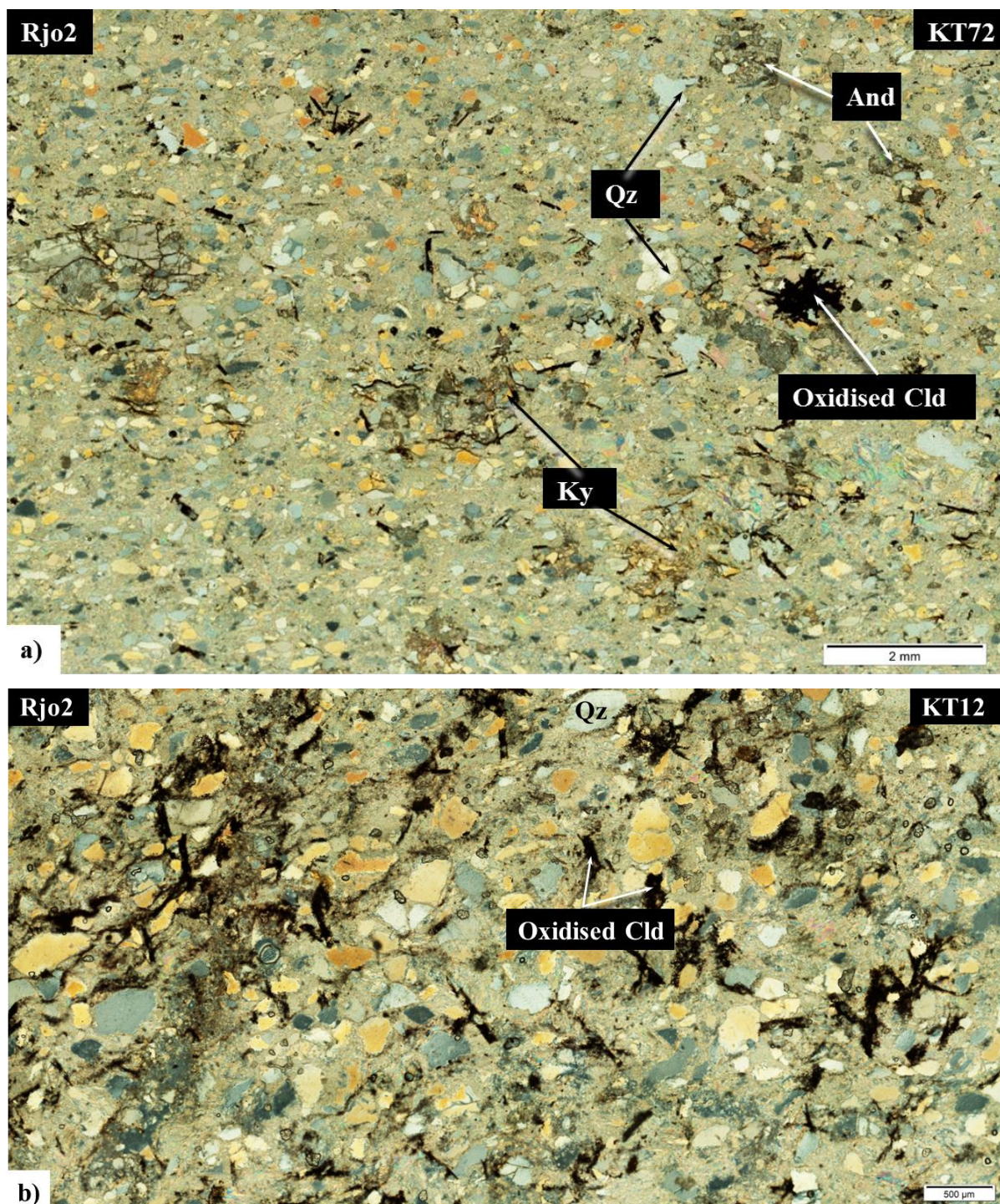


Figure 6.14. Photomicrograph mosaics of kyanite and andalusite-bearing Sample KT72 and lower-grade Sample KT12 (from the same stratigraphic unit Rjo2), composed of: (a) 35% sericite+muscovite, 30% quartz, 18% kyanite, 12% andalusite, and 5% chloritoid, and (b) 38% quartz, 47% sericite+muscovite, 13% chloritoid, 2% rutile.

6.4. Summary

Metamorphic mineral textures show that rocks in the study area were subjected to two metamorphic events. An earlier, $M_{2(NC)}$, metamorphism formed textures that are crosscut by impact-induced features (e.g., pseudotachylitic breccia, kink bands and spaced fracture cleavage). The $M_{2(NC)}$ metamorphism is divided into prograde $M_{2a(NC)}$ metamorphism and subsequent $M_{2b(NC)}$ retrograde metamorphism. Prograde metamorphism formed metamorphic assemblages containing mm- to cm-scale garnet, cordierite, andalusite, biotite, hornblende, epidote, grunerite, chloritoid, chlorite, and kyanite porphyroblasts. Retrograde reactions formed metamorphic assemblages that include muscovite, actinolite, staurolite, biotite, and chlorite as replacements of the earlier $M_{2a(NC)}$ assemblages. The $M_{2(NC)}$ metamorphism developed syn- to post-kinematic to the regional S_2 foliation hosted in both the West Rand and Central Rand groups (Section 5.3). The later, $M_{4(NC)}$, metamorphism occurred during and immediately after the Vredefort impact (D4). The $M_{4b(NC)}$ metamorphism annealed impact-induced micro-faults and formed chlorite-, cordierite-, biotite-, and possibly muscovite-bearing assemblages that overprinted the groundmass of pseudotachylitic breccias, Type 2a cordierite, and chloritoid in the lower West Rand Group.

7.0 IMPACT-RELATED MELT-ROCKS

7.1. Introduction

The impact-related melt-rocks mapped in the study area include a granophyre dyke (with at least one offshoot) and pseudotachylitic breccias. Most pseudotachylitic breccias hosted in the Vredefort Dome and the Witwatersrand Basin are reported to have been formed by the Vredefort impact (Dressler and Reimold, 2004; Reimold and Gibson, 2006), although some evidence of pre-impact pseudotachylitic breccia has been described in the goldfields (e.g., Berlenbach and Roering, 1993; Mambane et al., 2011). It is part of the aims of this study to determine whether the activation of pre-impact structures was associated with formation of pseudotachylitic breccia in the study area. Evidence from field mapping has demonstrated that pseudotachylitic breccias crosscut pre-impact structures, and show syngenetic relationships and textures to impact-induced structures; thus, they are regarded as D4 deformation features. The only exception is the largest volume (~2 m wide) of pseudotachylitic breccia mapped adjacent to the NW-SE trending segment of the D1 Koedoesfontein Fault (Figures 3.6, 3.7). These breccias are interpreted as impact-induced pseudotachylitic breccia injected into the pre-existing NW-SE-trending segment during the formation of NW-SE trending D4 faults in the Central Rand Group. It is unlikely that they formed during the activation of the Koedoesfontein fault because: (a) they are not planar; (b) they don't show any evidence for shearing; (c) their volume is too large for friction-generated pseudotachylitic breccia (e.g., Mambane et al., 2011).

7.2. Granophyre Dyke

In the study area a single 10-35 m wide, northeast-trending granophyre dyke crosscuts the core-collar boundary of the Vredefort Dome (Figures 3.2, 7.1). The dyke is characterised by a fine-grained matrix that predominantly hosts angular quartzite and granite clasts up to 22 cm in size (Figure 7.2). It is also characterised by an offshoot and en-echelon geometry exhibited on Donkervliet and Rondekop farms respectively (Figure 3.2). The granophyre dyke crosscuts a D4 fault hosted in the West Rand Group (Figure 3.4). As the D4 faults are related to tangential shortening (Section 2.4.1), this relationship indicates that dyke emplacement occurred after tangential shortening during the rise of the central uplift. The tangential and radial orientation of the granophyre dykes across the Vredefort Dome, noted by Theriault et al. (1996), led these authors to propose that downward intrusion of the impact melt occurred along vertical fractures that opened as a result of the collapse of the central uplift.

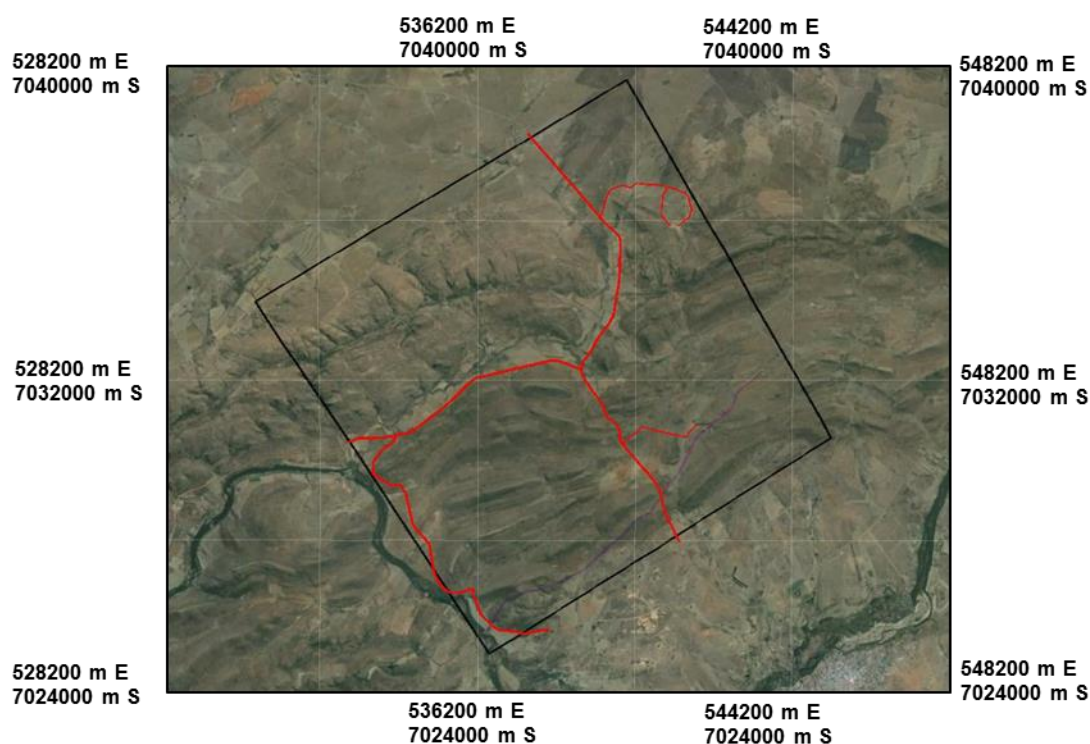


Figure 7.1. Google Earth™ image showing outcrop distribution of the granophyre dyke (purple).



Figure 7.2. Outcrop photographs of the clast-laden granophyre dyke (35 J 541806 m E, 7030608 m S; and 35 J 541448 m E, 7029891 m S respectively).

7.3. Pseudotachylitic breccias

The mechanism of pseudotachylitic breccia formation and the timing of formation (c.f., Dressler and Reimold, 2004; Reimold and Gibson, 2005, 2006) have been subjects of interest in the dome. Some evidence of the timing and mechanism of formation are indicated through relative chronology of crosscutting events, and geometry and textural characteristics of the pseudotachylitic breccias.

Pseudotachylitic breccias hosted in the collar are heterogeneous in appearance. The appearance and composition of the breccias differ depending on the rock type in which they are hosted. The pseudotachylitic breccias occur as aphanitic veins that are generally grey on weathered surfaces, but colours range from black to brown depending on the host rock.

Pseudotachylitic breccias are well-preserved in quartzite units of the study area. They generally occur as: (a) irregular vein networks hosted proximal to D4 faults; (b) planar breccia veins hosted along fractures and bedding; and (c) tabular breccia that consists of intersecting networks of bedding-parallel and fracture-hosted pseudotachylitic breccia (Figure 3.6). The pseudotachylitic breccias occur throughout the study area, with a slight concentration of breccias in the thicker quartzite units of the Central Rand Group. The volume of pseudotachylitic breccias mapped, especially bedding-parallel breccia, is larger in the Central Rand Group than in the West Rand Group (Figure 3.7); however there is a general decrease in the volume of pseudotachylitic breccia in the collar (e.g., Reimold and Gibson, 2006).

7.3.1. Irregular pseudotachylitic breccia (non-planar)

Most irregular pseudotachylitic breccias are hosted proximal to D4 impact-induced fault zones (Figure 3.6). Irregular vein networks are up to 2 m in width (including brecciated country rock clasts) and host the largest and most diverse clast populations (Figure 7.3). Irregular veins that host clasts of multiple lithologies are developed in the Hospital Hill and Government subgroups. The polymictic clast composition of irregular pseudotachylitic breccias includes meta-dolerite, quartzite, meta-ironstone, and metapelite (Figure 7.4). The pseudotachylitic breccias are characterised by: (a) aphanitic to microcrystalline groundmass that forms >60% of the sample petrographically, making the pseudotachylitic breccia matrix-supported; (b) subangular-subrounded clast populations; and (c) poorly-developed flow textures. These characters are atypical of cataclasite (French, 1998).

7.3.2. Fracture-hosted pseudotachylitic breccia

The planar veins are mostly sub-vertical, with a thickness typically less than 10 cm. These veins are locally hosted along D4 joints, and show consistent trends to local joints (Figure 7.5). The most consistent characteristic of fracture-hosted pseudotachylitic breccia is that the veins tend to have a lighter grey central zone that is more clast-rich, bounded by darker and finer-grained margins with fewer clasts (Figure 7.6a-c). This texture also occurs in bedding-parallel pseudotachylitic breccia. Also, fracture-hosted veins are monomictic and the clasts are significantly smaller than in irregular veins; typically a few millimetres in size. The small clast size is coupled with a common ellipsoidal flow texture developed in relatively thick veins (typically a few centimetres thick; Figure 7.7). The ellipsoidal texture is a compositionally and grain size variable texture seen on outcrop (Figure 7.7)

and microscopic scale (e.g., Sample B7 in Table 3.7, Appendix III). Fracture-parallel pseudotachylitic breccia crosscuts S_2 cleavage in the Central Rand Group (Figure 5.8).

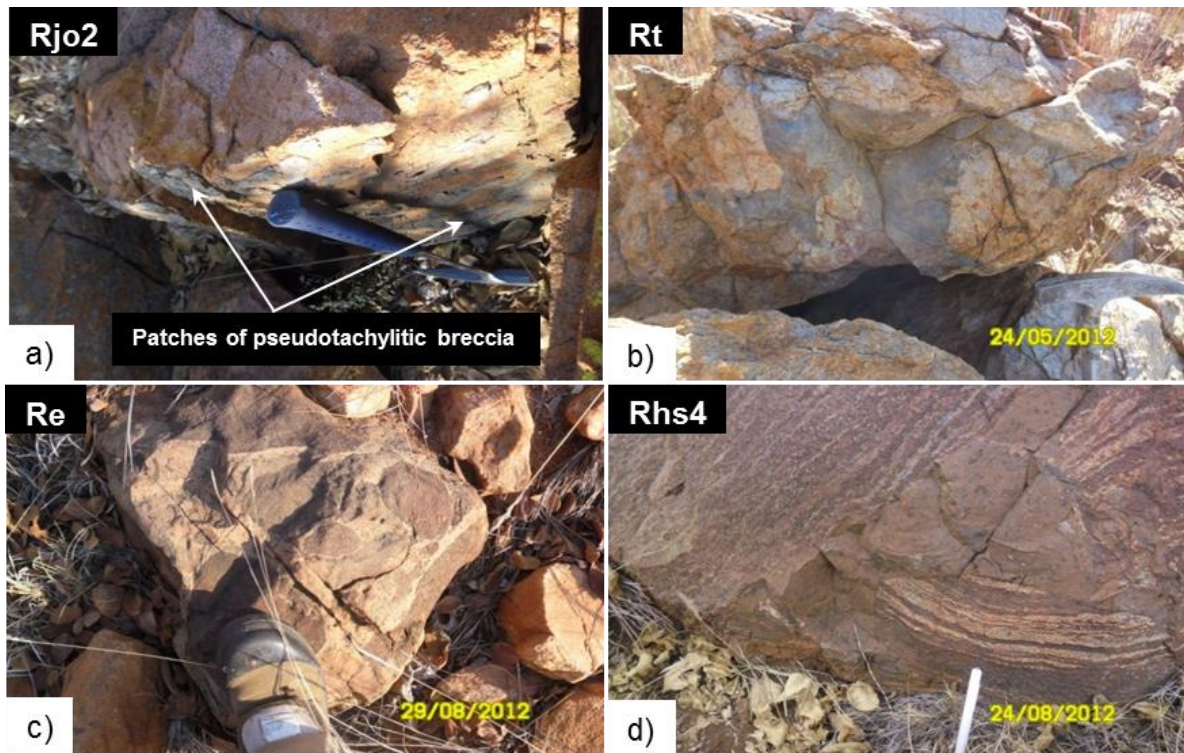


Figure 7.3. Irregular pseudotachylitic breccia hosted in: (a) Rjo2 quartzite (35 J 539976 m E, 7033497 m S); (b) Rt quartzite (35 J 540720 m E, 7032736 m S); (c) metadolerite (35 J 538738 m E, 7029907 m S); and (d) metapelite (35 J 543145 m E, 7031244 m S).

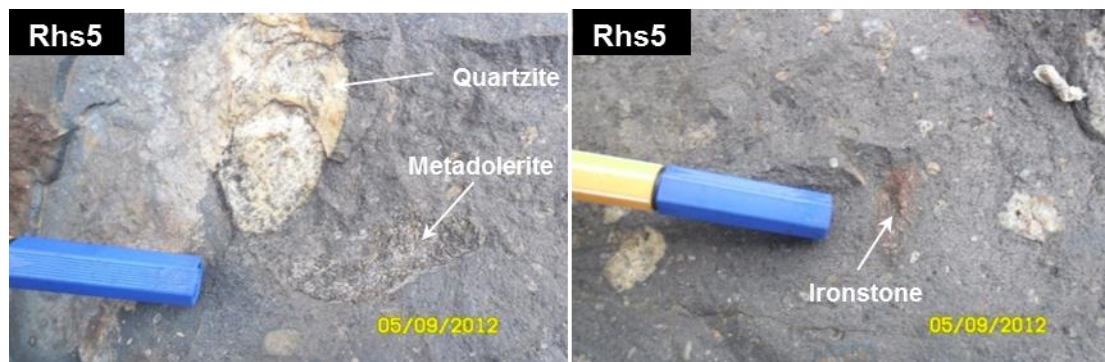


Figure 7.4. Polymictic clast population comprising quartzite, metadolerite, and meta-ironstone xenoliths hosted in irregular pseudotachylitic breccia that crosscuts the Hospital Hill Subgroup (35 J 535523 m E, 7026949 m S).

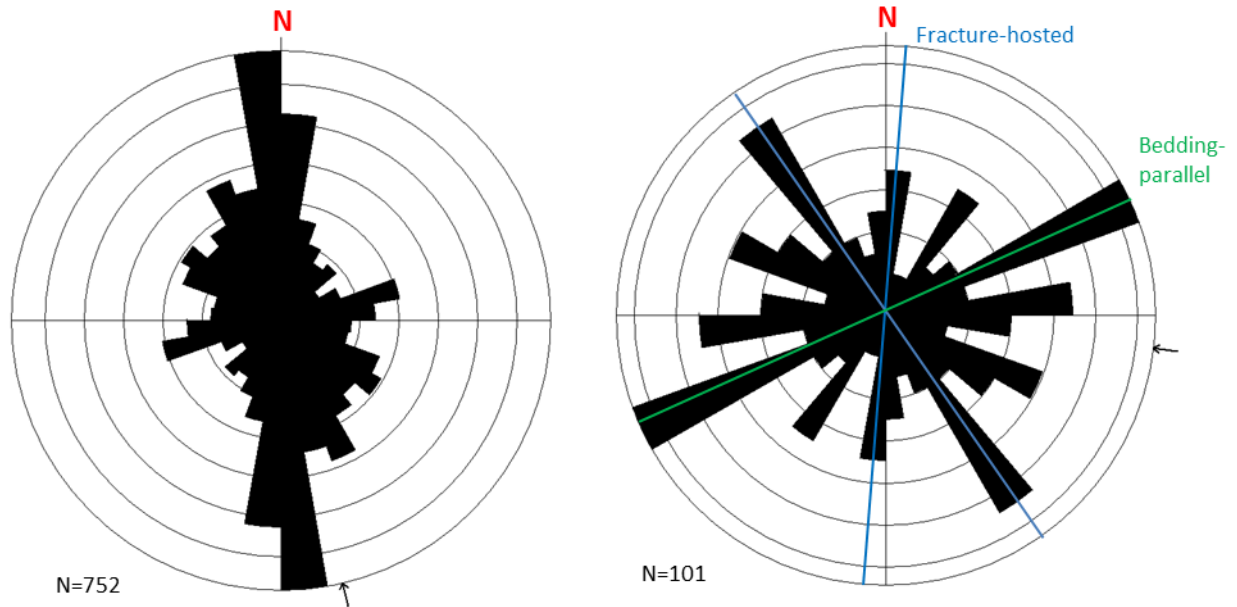


Figure 7.5. (a) A rose plot of trends of subvertical joints, showing a consistent NW-SE to N-S trend (N=752). (b) A rose plot of trends of planar to sub-planar pseudotachylitic breccia (N=101); showing consistent trends to NW-SE and N-S joints and general bedding trend.

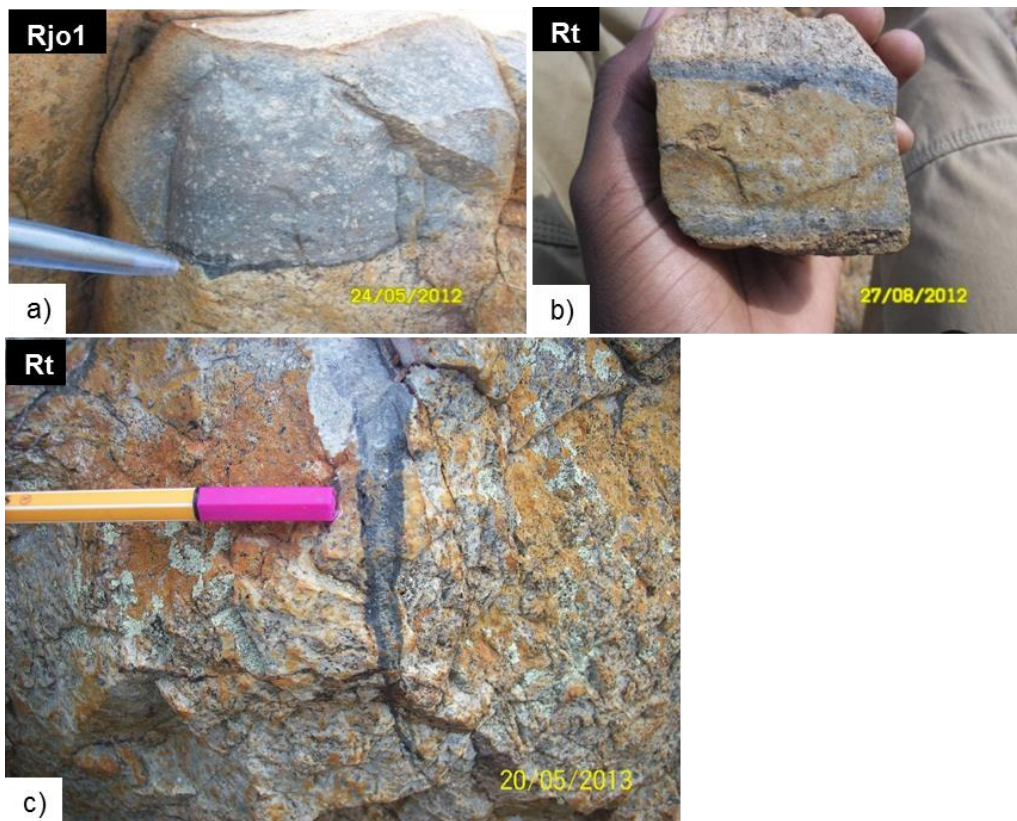


Figure 7.6. (a) – (c) Pseudotachylitic breccias exhibiting light-grey, clast-rich cores and darker finer-grained margins (35 J 540720 m E, 7032736 m S; 35 J 540147 m E, 7035440 m S; and 35 J 542291 m E, 7036574 m S).

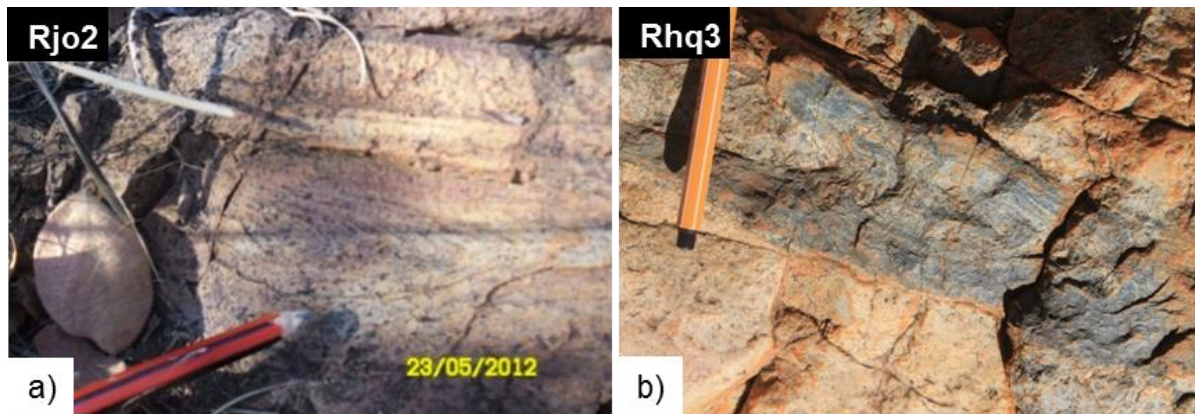


Figure 7.7. Ellipsoidal flow texture developed in: (a) fracture-hosted and (b) bedding-parallel pseudotachylitic breccia, respectively (35 J 540035 m E, 7033546 m S; and 35 J 535821 m E; 7027785 m S).

7.3.3. Bedding-parallel pseudotachylitic breccia

Pseudotachylitic breccias that are developed along bedding are the most abundant breccias mapped in the study area, and could be traced for up to 2 km along the Rhq3 unit, where they locally grade into irregular networks (Figure 3.6). Bedding-parallel pseudotachylitic breccias are generally thicker than those hosted in fractures (Figure 7.8), measuring up to 45 cm in width. The veins have variable orientations that conform to different local bedding orientations.

Bedding-parallel pseudotachylitic breccias have relatively small clasts, and their margins are generally finer-grained and darker than the central zone (Figure 7.6). The only exception is in the Rhq3 quartzite unit (Hospital Hill Subgroup), where bedding-parallel pseudotachylitic breccias display a darker and finer-grained central zone than the margins (Figure 7.9). The lighter grey margins are interpreted as cataclastic zones, with a finer and darker pseudotachylitic breccia zone hosted in the middle. The bedding-parallel breccia is extensive along the length of the Rhq3 quartzite in Donkervliet Farm, associated with wedge-shaped footwall and hangingwall offshoots (Figures 7.11b, 7.12a). The main breccia is up to 8 cm wide, and is located at approximately the same stratigraphic location across D4 faults (Figure 7.10). This crosscutting relationship indicates that bedding-parallel breccia was formed before activation of the pseudotachylitic breccia-laden D4 faults. Bedding-parallel pseudotachylitic breccias crosscut S_2 cleavage hosted in the Central Rand Group (Figure 5.8).

7.3.4. Tabular pseudotachylitic breccia

Fracture-hosted and bedding-parallel pseudotachylitic breccias are described as separate breccia formations above; however, there are a number of locations where the overlap between the two is exhibited. For convenience, the geometry of breccias involving intersecting bedding-parallel and fracture-hosted pseudotachylitic breccias is referred to as tabular pseudotachylitic breccia (Figure 3.6). Tabular pseudotachylitic breccias display a 3D geometry in outcrop (Figure 7.11). The geometry and intersection of the bedding-parallel and fracture-hosted pseudotachylitic breccia suggests that the fractures most likely host injection veins into bedding, rather than offshoots from bedding (Figure 7.11).

The offshoots/injections generally intersect with bedding-parallel pseudotachylitic breccia. One exception is in the Rhq3 quartzite, where the bedding-parallel breccia is separated from the offshoots/injections by a cataclastic zone (Figure 7.12).

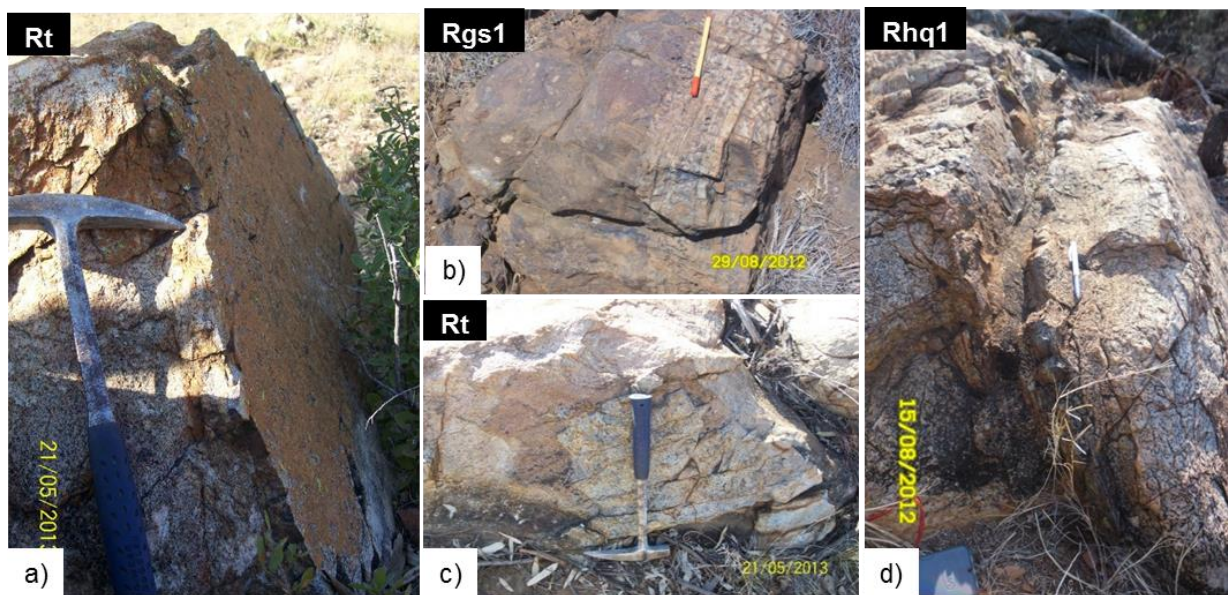


Figure 7.8. Bedding-parallel pseudotachylitic breccia in metapelite and quartzite units. (a) & (c) 35 J 539286 m E, 7034274 m S; (b) 35 J 538720 m E, 7029134 m S; (d) 35 J 541117 m E, 7030014 m S.



Figure 7.9. Bedding-parallel pseudotachylitic breccia composed of a darker clast-poorer grey core, and lighter and clast-rich margins (35 J 537665 m E, 7028589 m S).

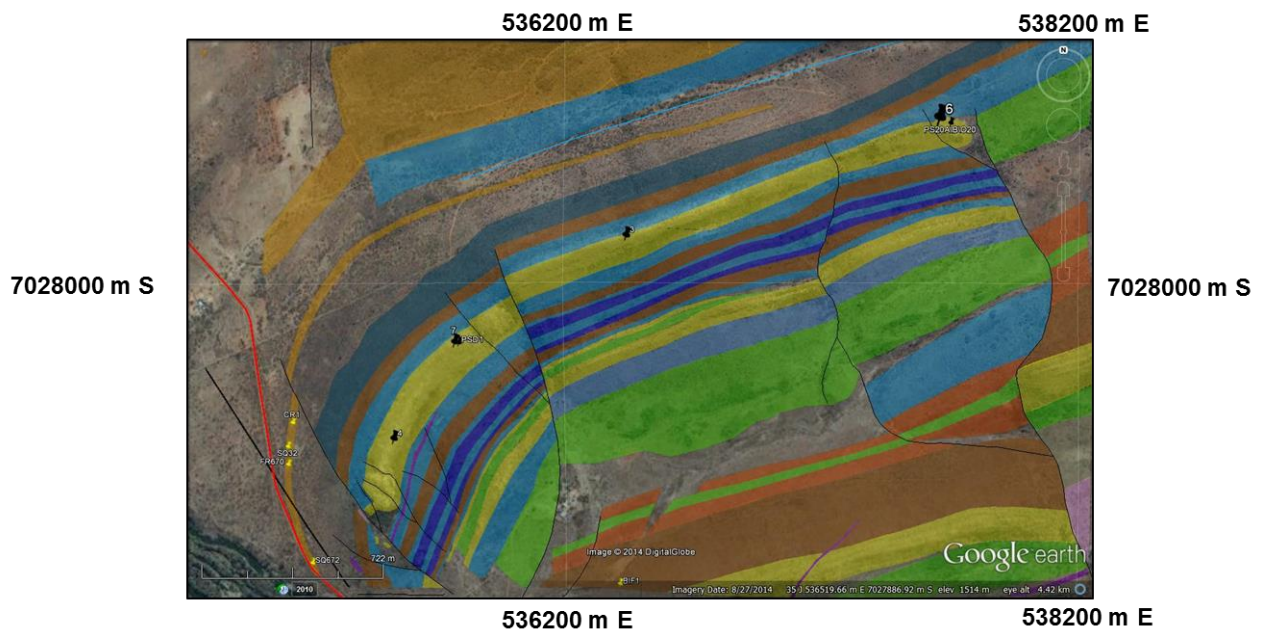


Figure 7.10. Locations of pseudotachylitic breccia hosted along the same stratigraphic horizon in the Rhq3 quartzite unit (black pins). The location of the breccia is not affected by impact-induced displacements of the quartzite unit.

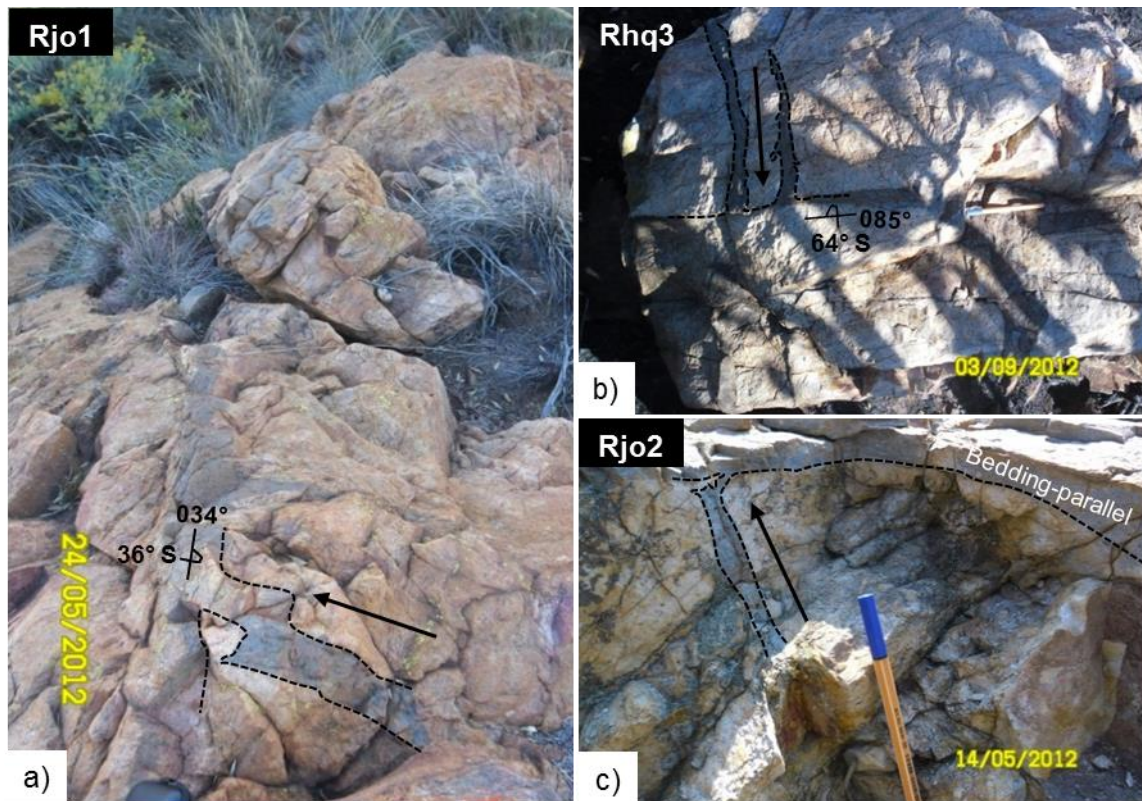


Figure 7.11. Tabular pseudotachylitic breccia associated with what are most likely footwall injections into bedding planes. The injections are interpreted from apparent flow directions, estimated from the geometry of the veins. (a) 35 J 540540 m E, 7033019 m S; (b) 35 J 537665 m E, 7028589 m S; (c) 35 J 535536 m E, 7033963 m S.



Figure 7.12. Injection/offshoot separated from the main pseudotachylitic breccia vein by a cataclastic zone (35 J 537665 m E, 7028589 m S).

7.4. Pseudotachylitic breccia petrography

Microscopically, the pseudotachylitic breccias are generally composed of a fine-grained groundmass that is characterised by flow bands and flow folds, with quartz clasts and/or lithic clasts. The colour of the groundmass ranges from light-brown to black depending on the composition of the host rock(s) (Figure 7.13).

Finer and darker margins of pseudotachylitic breccia described in Sections 7.3.2 and 7.3.3, which are characterised by flow bands, have a microcrystalline texture that is finer than that of the central zone (e.g., sample PS32B in Figures 7.14c, d). Flow bands are more common along the pseudotachylitic breccia margins, but may also be developed within the central zone of some breccias (Figures 7.14a, b). In contrast to finer margins, sample MR48 hosts pseudotachylitic breccia with coarser microcrystalline margins (Figure 7.15). The crystal size decreases towards the flow bands in the middle. The inward fining character of pseudotachylitic breccia is also exhibited along bedding of the Rhq3 unit (Sample PS20A, Figure 7.9). The breccia is composed of a central zone with clasts that are 0.20-0.50 mm in size, forming <15% of the central zone (~85% aphanitic groundmass). The marginal zones host clasts that are 0.2-1.1 mm in size, forming ~50% of the marginal zones (~50% aphanitic groundmass, Figure 7.16); which is most likely a cataclastic zone. Ellipsoidal/eye-shaped flow textures (Figure 7.7) are characterised by alternating layers with different compositions and grain size (Figure 7.17).

The pseudotachylitic breccias predominantly host mono- and/or polycrystalline quartz clasts (Table A3.7, Appendix III). The quartz clasts are recrystallized in some samples, with decreasing recrystallization into the clast (coarsening inwards; Figure 7.18a). Dolerite lithic clasts and quartz-chlorite, and garnet clasts also form part of the clast population. Andalusite, meta-dolerite and meta-ironstone fragments were also observed as part of the clast populations in outcrops (Figure 7.4). Petrographically, the fragments are composed of plagioclase, clinopyroxene, elongate and radiating chlorite crystals ('rosettes'), and quartz (Figure 7.18b-d). The chlorite 'rosettes' also overprint the groundmass of pseudotachylitic breccia in samples PS7, MR9, and PS192 (Table A3.7, Appendix III).

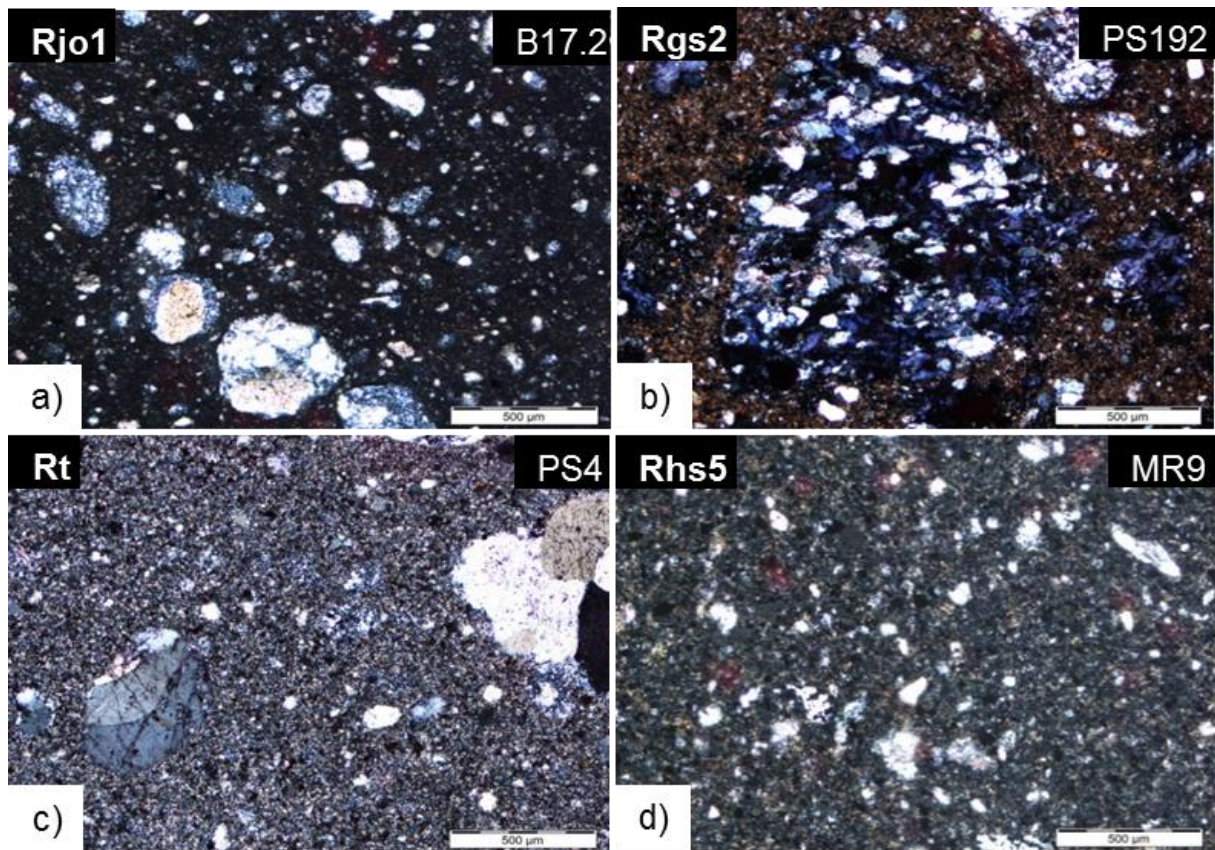


Figure 7.13. Photomicrographs of structureless pseudotachylitic breccia showing varying compositions of clasts and the groundmass. (a) Microcrystalline groundmass with quartz inclusions (XPL); (b) aphanitic groundmass overprinted by Type 4b chlorite rosettes (XPL); (c) & (d) aphanitic groundmass with quartz inclusions (XPL).

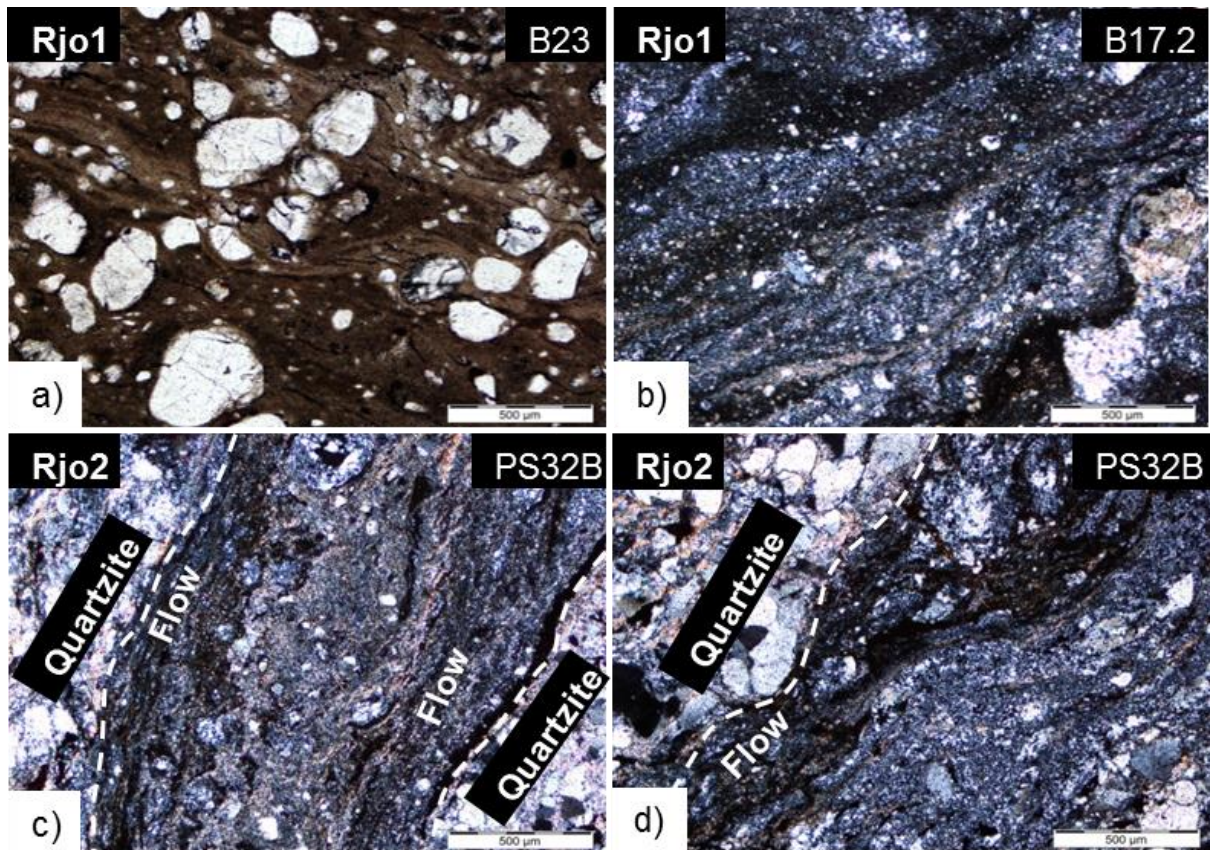


Figure 7.14. Photomicrographs showing flow banding in pseudotachylitic breccia. (a) & (b) Flow banding developed within the core of pseudotachylitic breccia (PPL and XPL respectively). (c) & (d) Flow banding along pseudotachylitic breccia margins (both under XPL).

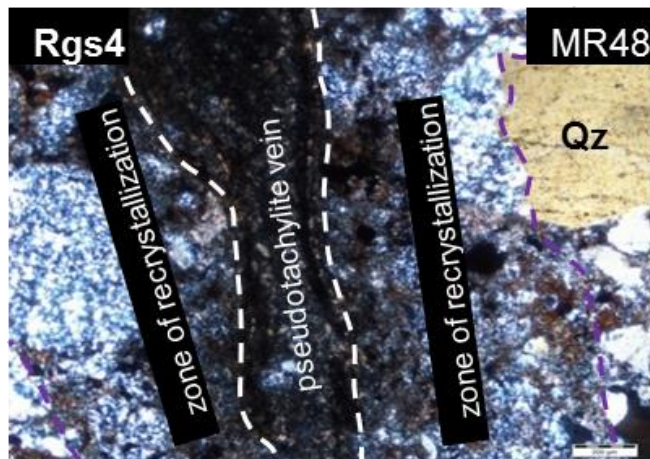


Figure 7.15. Pseudotachylitic breccia associated with marginal recrystallization (XPL).

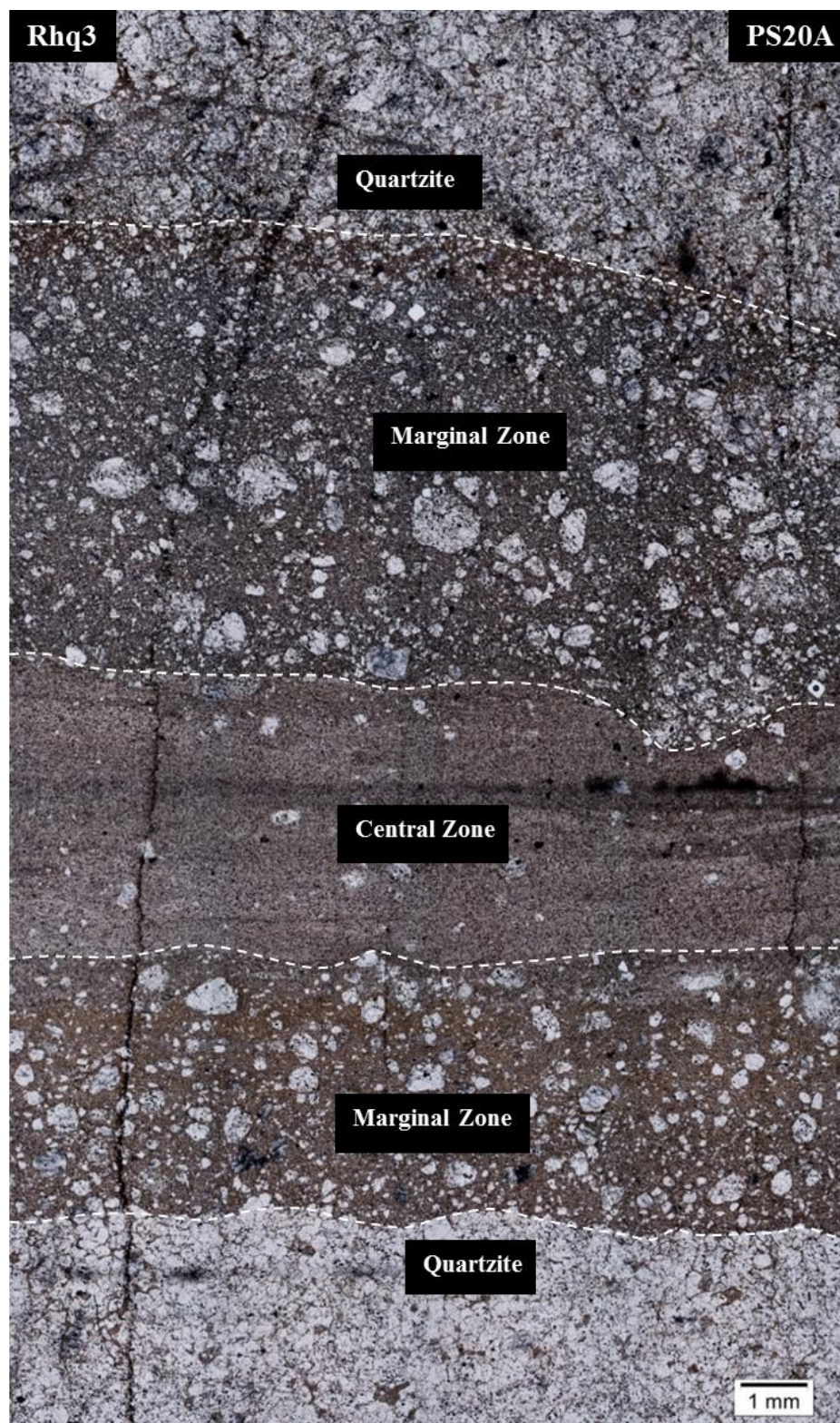


Figure 7.16. Photomicrograph mosaic of the layered, bedding-parallel PS20A pseudotachylitic breccia. The breccia fines inwards. The central zone of the breccia has fewer clasts (~15%) than the marginal zones (~50%; PPL).

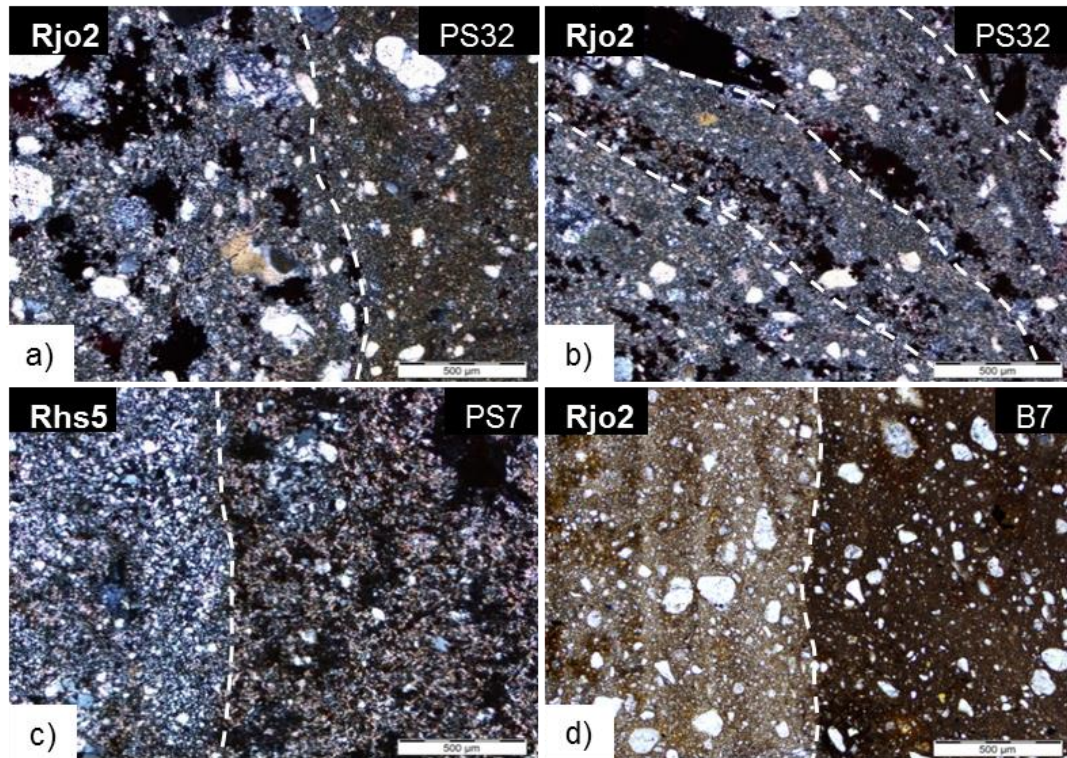


Figure 7.17. (a) – (d) Photomicrographs of ellipsoidal/eye-shaped flow texture characterised by compositional layering (a-c=XPL, d=PPL).

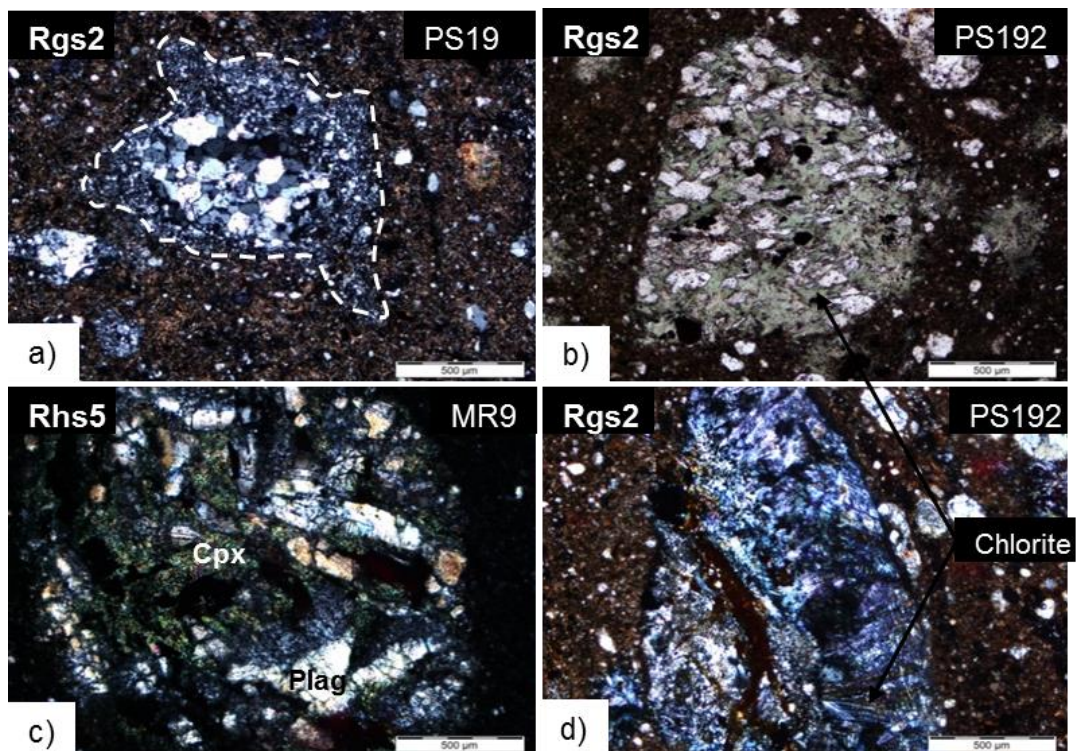


Figure 7.18. Photomicrographs of inclusions hosted in pseudotachylitic breccia. (a) Recrystallized quartz inclusions hosted in an aphanitic groundmass, with an inward increase in grain size. (b) - (d) Metadolerite and dolerite inclusions hosted in microcrystalline to aphanitic groundmass.

7.5. Summary

The study area generally hosts irregular (fault-hosted), planar fracture-hosted and planar layer-hosted pseudotachylitic breccias that crosscut pre-impact features. The irregular pseudotachylitic breccias are generally structureless and some show polymictic textures, whereas planar pseudotachylitic breccias are characterised by flow banding and compositional layering. The association of pseudotachylitic breccias with D4 faults and joints (Figure 7.5), as well as the observation that pseudotachylitic breccias cut across D2 features, show that these pseudotachylitic breccias were generated by the impact.

8.0 DISCUSSION

8.1. Introduction

Evidence acquired from desktop mapping, field mapping, and analytical methods show that the northern collar of the Vredefort Dome is characterised by: (a) multiple deformations that generated brittle and ductile structures (D1-D4); and (b) polyphase metamorphism ($M_{2(NC)}$ and $M_{4(NC)}$) producing a range of mineral assemblages that show key temporal relationships with the brittle and ductile features. Cross-cutting relationships of these characteristic features show that a significant pre-impact deformational history exists in the northern collar of the Vredefort Dome.

8.2. Deformations recorded in the northern collar of the Vredefort Dome

Most faults in the collar of the Vredefort Dome have been attributed to the Vredefort impact at 2.02 Ga (e.g., Simpson, 1977; Lilly, 1978; Colliston and Reimold, 1990; Lana et al., 2003b; Wieland et al., 2005). Although the sequence of deformation during impact processes is not fully understood, Wieland et al. (2005) suggested that most of the faults were formed under tangential shortening during a constrictional interval in central uplift formation. This excluded two faults described by Lilly (1978) in the northern collar (Figure 2.13). The faults were attributed to an earlier deformation on the basis that they are terminated by the Black Reef Formation, making them older than 2.588 Ga (Figure 8.1).

Fieldwork for this study was conducted across the southern extensions of the faults described by Lilly (1978), where the relationship to the Transvaal Supergroup was not encountered. This study confirmed the existence of several generations of pre-impact structures within the study area; the data confirms that multiple deformations exist. Pre-impact deformations are evidenced by: (a) characteristics of pre-impact faults compared to impact-induced faults, (b) possible metamorphic grade variation in stratigraphic units displaced across pre-impact faults, and (c) cross-cutting relationships of D1-D4 features.

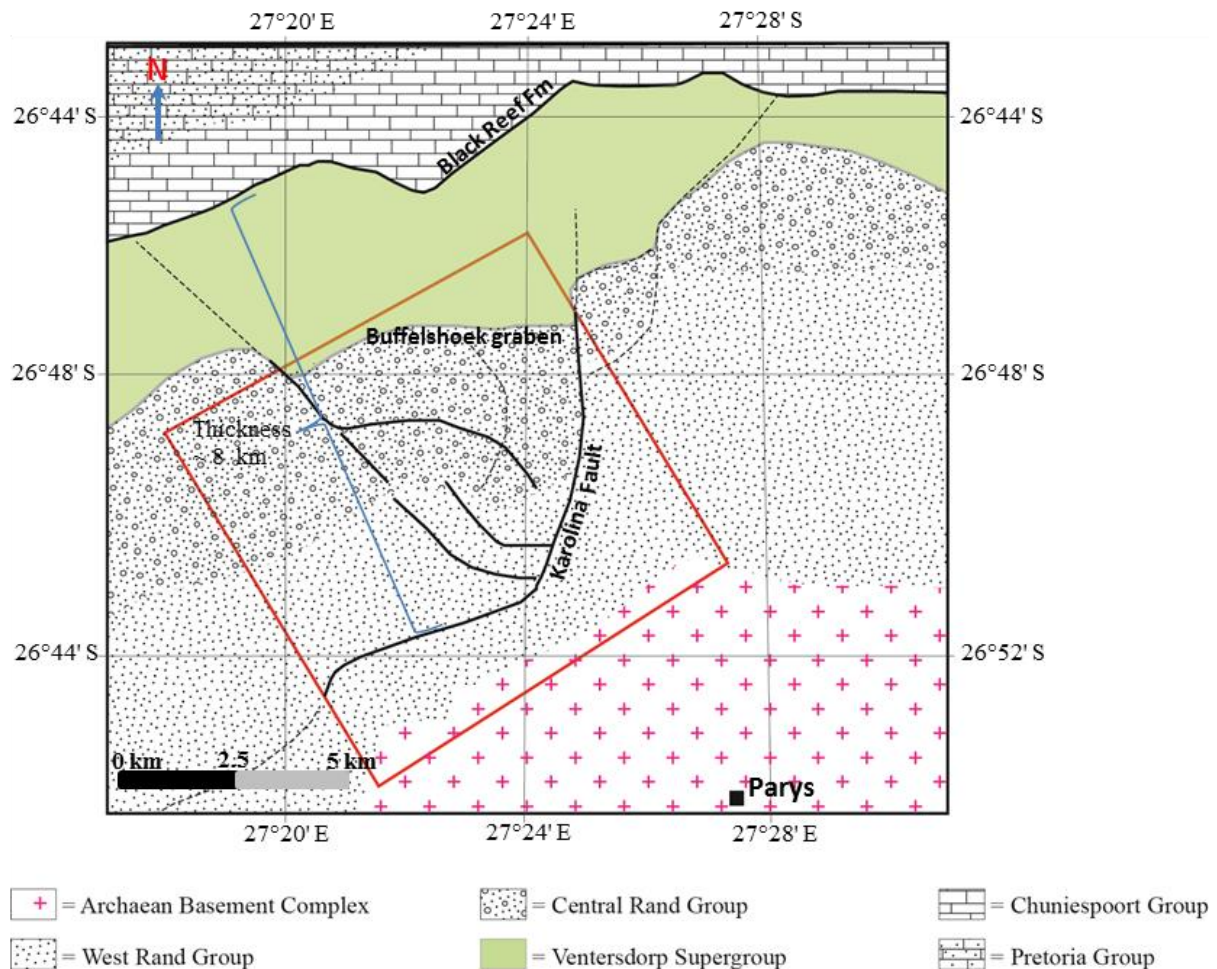


Figure 8.1. A sketch of the D1 listric faults extended up to the Black Reef Formation contact (Study area indicated with the red rectangle).

8.2.1. Characteristics of pre-impact faults

Pre-impact faults (D1) have a kilometer-scale listric geometry that distinguishes them from impact-induced faults (D4; Figure 3.4). The magnitude of stratigraphic offset along the listric faults increases towards the north (outer collar), whereas that of impact-induced faults is greatest in the southern part of the study area (inner collar; Sections 5.2, 5.5); the latter indicating that rocks closer to the centre of the dome were subjected to more deformation during impact. This is consistent with the observation by Wieland et al. (2005) that strata shortening decreases outwards from the core-collar boundary. In contrast to impact-induced faults, which host pseudotachylitic breccia in the study area; there is no evidence of pseudotachylitic breccia in any pre-impact fault.

8.2.2. Distribution of metamorphic grade across the Karolina Fault

The lower West Rand Group hosts mid-amphibolite facies metamorphic assemblages, whereas the Central Rand Group contains greenschist facies metamorphic assemblages (generally attributed to $M_{2(NC)}$). Considering the proposed 2.06 Ga age of metamorphism (Gibson and Stevens, 1995; Gibson and Wallmach, 1995; Phillips and Law, 1994), metamorphic assemblages of the same units hosted on either side of D4 faults should be similar because the displacement would have occurred after metamorphism (Figure 8.2 a). Gibson and Wallmach (1995) proposed that the peak metamorphic geotherm during this event was 35 – 40 °C/km. Thus, metamorphic assemblages developed in the same stratigraphic units should show some difference in grade across D1 displacements, because offset would have occurred before metamorphism (Figure 8.2 b). However, the listric nature of the Karolina Fault means that little vertical displacement of the pelite-rich West Rand Group appears to have occurred in the study area. However, Bush (2014) and Nemutandani (2014) described metapelite units of the Hospital Hill Subgroup (Wild Waters Ridge, ~10 km east of the study area) in the footwall of the Karolina Fault and another suspected pre-impact listric fault (Section 5.2), which should have been displaced upwards relative to the same units in the study area. The metapelites at Wild Waters Ridge are finer-grained (porphyroblasts <4 mm) relative to those in Donkervliet Farm (porphyroblasts up to 50 mm; Figure 8.3). Also, the former hosts the assemblage chlorite + chloritoid + muscovite + andalusite + quartz + ilmenite + rutile ± magnetite (estimated temperatures ~440 °C; Nemutandani, 2014). These are upper greenschist facies grade compared to the mid-amphibolite facies hosted in the West Rand Group of the study area (Section 6.2).

8.2.3. Crosscutting structural relationships

Cleaved horizons (S_2) were mapped in both the West Rand and Central Rand groups, which allowed for comparison of cross-cutting relationships across the study area. There are a number of cross-cutting relationships that indicate that cleavage was formed before the impact: (a) a knotted quartzite horizon in the Turffontein Subgroup is displaced several times by north-south oriented D4 impact-induced faults (Figure 5.7); (b) a cleaved quartzite horizon in the Turffontein Subgroup is crosscut by D4 pseudotachylitic breccia and kink bands (Figure 5.8); (c) The S_2 -C foliation in the Lower Government Subgroup is crenulated, and in turn displaced and rotated, by D4 impact-induced folds and faults that host pseudotachylitic breccia (Figure 5.4); (d) The S_2 -C fabric hosts isoclinally-folded intrafolial quartz veins, representative of a progressive ductile deformation rather than a rapid deformation associated with impact processes (Figures 5.12); and (e) $M_{2(NC)}$ metamorphic minerals are included in pseudotachylitic breccia (meta-dolerite in Figure 7.4) and are crosscut by D4 microfaults (Figure 5.22a).

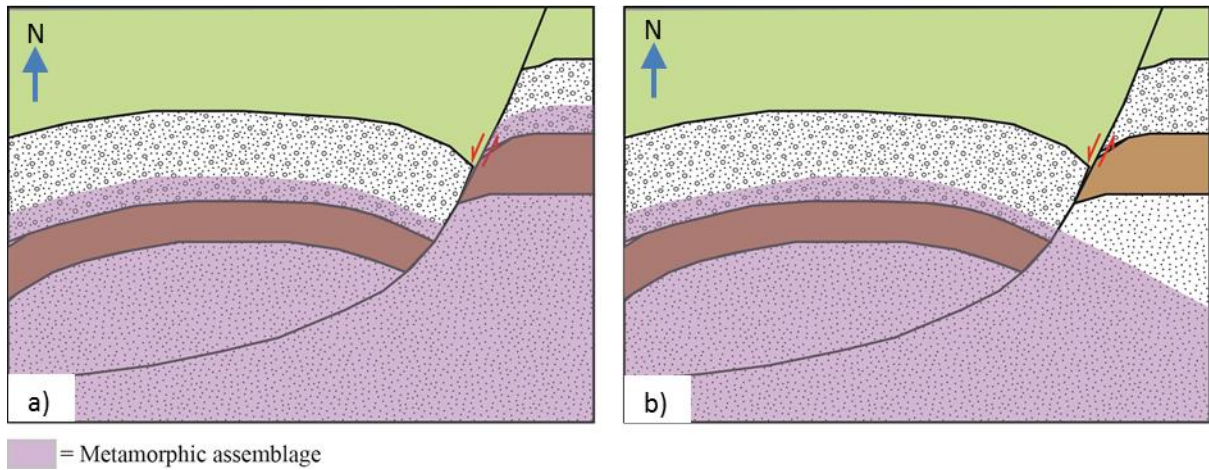


Figure 8.2. A sketch showing the expected distribution of metamorphic assemblage, provided: (a) $M_{2(NC)}$ metamorphic assemblages (similar across the fault) formed before activation of D1 listric faults; or (b) $M_{2(NC)}$ mineral assemblages (different across the fault) that postdate D1 listric faults.

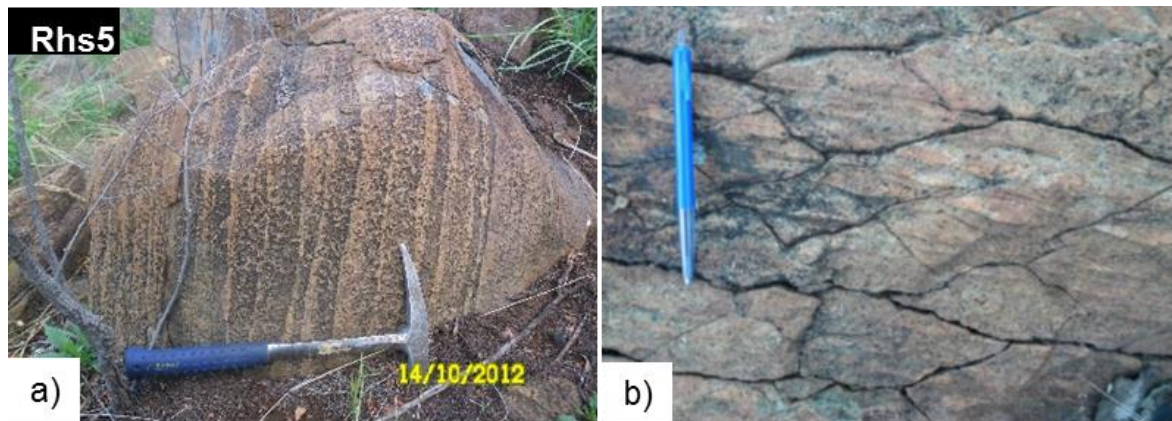


Figure 8.3. Outcrop images of Al-rich metapelite units of the Hospital Hill subgroup, located in: (a) Donkervliet farm, with andalusite porphyroblasts up to 50 mm long; and (b) Wild Waters Ridge, with andalusite porphyroblasts up to 4 mm (Nemutandani, 2014).

8.3. D1 listric faults

The D1 faults in their present geometry display apparent strike-slip offsets that are inconsistent with those shown by pseudotachylitic breccia-bearing impact-related faults. Whilst the latter indicate tangential (east-west) shortening, the D1 faults indicate east-west extension and north-south shortening. Bearing in mind that both the strata and D1 faults experienced $\sim 124^\circ$ of rotation related to the impact event, restoration to their pre-impact orientation indicates a vertical shortening and tangential east-west extension, i.e., conditions consistent with extensional tectonics (Ramsay and Huber, 1987). The overall effect of these geometries is an approximately east-west extension of the strata. This extension is consistent with the Ventersdorp-age extension/Hlukana-Platberg Event described by Tinker et al. (2002) and Manzi et al. (2013). Both these studies suggested a series of kilometre-scale half-grabens with a northwest-block-down displacement. Furthermore, Bush (2014) described a southeast-block-down (~ 3.5 km), displacement of the West Rand Group in the northeastern collar (Figure 5.3); the geometry suggests that it is potentially a pre-impact listric fault. This fault, together with the Karolina Fault, bound a potential horst block when the impact-induced overturning is restored.

8.3.1. Geometry of listric faults

Listric faults are initiated as steep ($\sim 60^\circ$ dip) normal faults (Ramsay and Huber, 1987). They develop when the extension is regional, causing strata to be under tensile forces, resulting in normal dip-slip with the progressive formation of low-angle decollement fault zones (Davis and Reynolds, 1996). The initial development of a listric fault system is associated with a space problem. As the fault system develops, antithetic accommodation faults and reverse drag must form in the hangingwall to eliminate gaps (Figure 8.4; Davis and Reynolds, 1996; Ramsay and Huber, 1987). This results in the formation of a hangingwall ‘rollover’ antiform, horsts and grabens/half-grabens, and drag folds along fault zones. All these features can be seen in the study area once structures are restored to their pre-impact orientation (Figure 3.4). The antithetic accommodation faults form a conjugate set with the main listric faults; both bound horsts and grabens (Figure 3.4). Hangingwall deformation described by Ramsay and Huber (1987) is represented by a hangingwall roll-over synformal anticline (present-day geometry) developed in the Central Rand Group (Figure 3.4), with a wavelength of ~ 5.3 km and amplitude of ~ 0.8 km. The listric faults have architecture and cross-cutting relationships similar to Neoarchaean listric faults developed in the goldfields interpreted from seismic profiles and mine plans (Tinker et al., 2002; Jolley et al., 2004, 2007; Beach and Smith, 2007; Dankert and Hein, 2010; and Manzi et al., 2013).

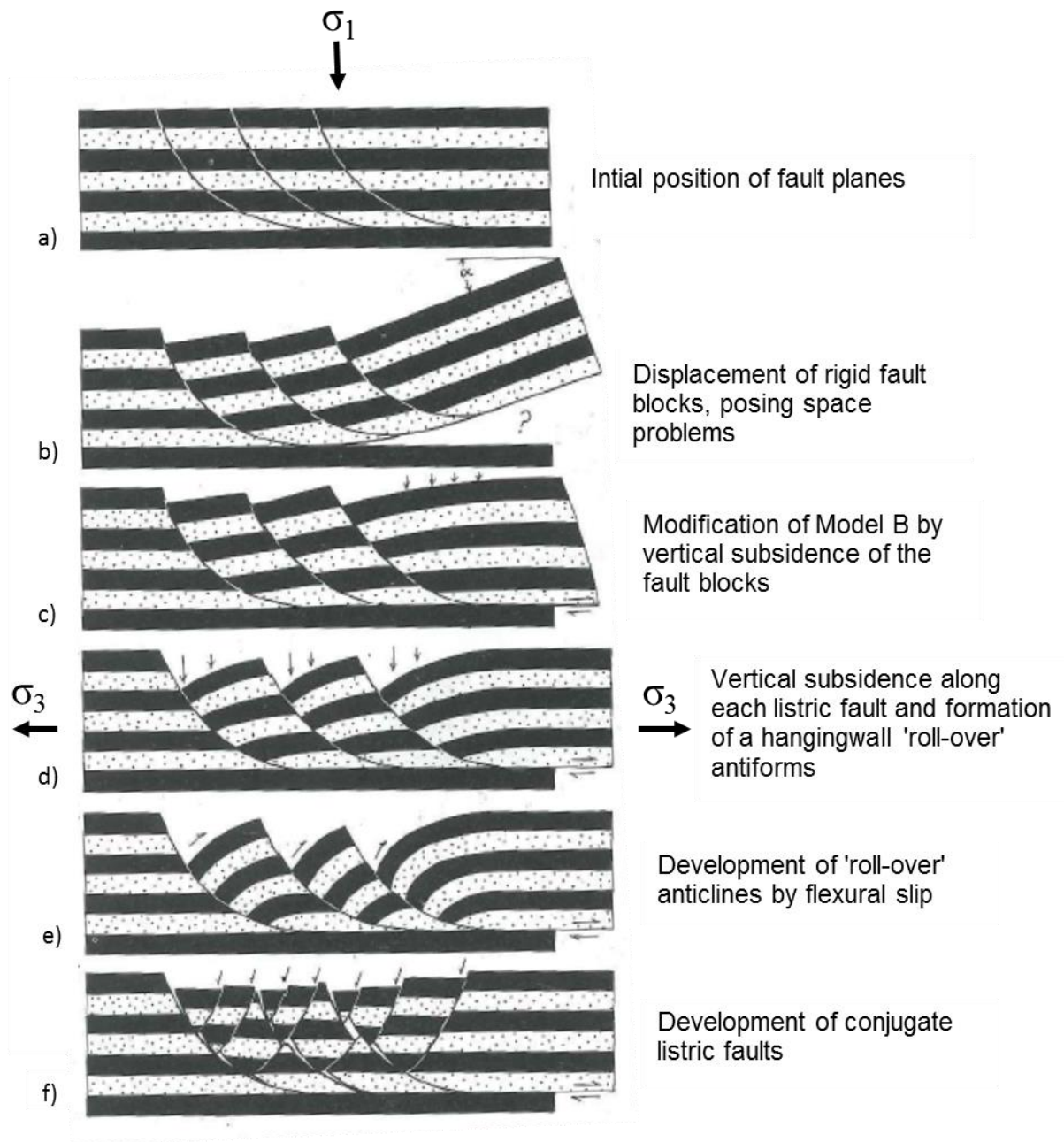


Figure 8.4. A diagram illustrating development stages of listric faults. (modified from Ramsay and Huber, 1987).

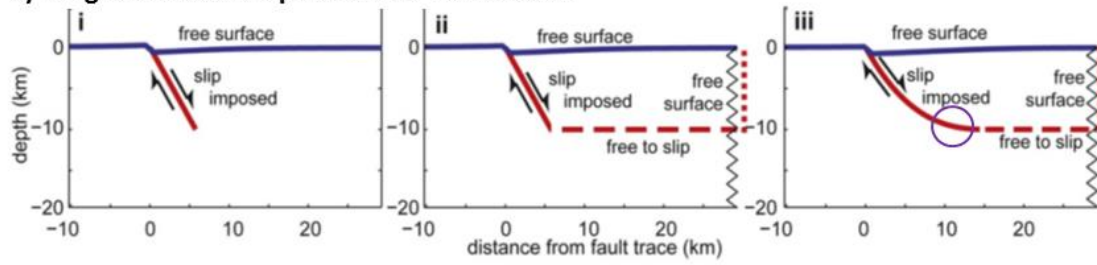
8.3.2. Stress distribution associated with activation of listric faults

The presence of a ‘rollover’ antiform in the hangingwall of a normal fault is regarded as evidence for a listric fault geometry (Laubscher, 1956; Hamblin, 1965; Shelton, 1984). However, Barnett et al. (1987) and Resor and Pollard (2012) indicated that a rollover antiform can also be developed in the hangingwall of sub-planar faults. In an attempt to find alternative structural characters for identifying listric faults, Resor and Pollard (2012) modelled deformations of planar, planar detached, and listric detached faults (Figure 8.5).

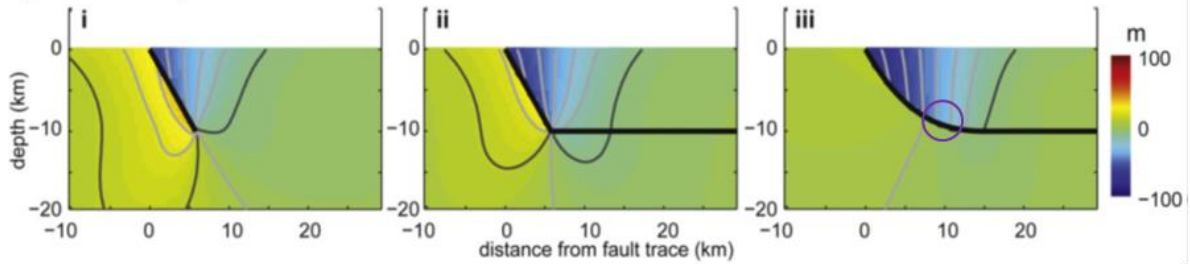
The models show that the degree of collapse in the hangingwall, and the width of the region being down-warped increases from planar faulting to listric detached faulting. Another essential observation made was that the magnitude of footwall uplift (vertical displacement), and the width of the region of footwall uplift decreased with development of a listric decollement zone (Figure 8.5 b). This means that progressive deformation that results in the formation of listric faults is characterised by a decrease in the magnitude of footwall uplift close to the decollement zone. This concept is consistent with the southward decrease in the magnitude of offset along the Karolina Fault as it flattens into the West Rand Group (Figure 3.4).

In order to predict areas that will be jointed and those that will be faulted, the distribution of tensile minimum compressive stress versus positive coulomb stress were investigated (Figure 8.5 c). Resor and Pollard (2012) showed that the area that experiences the most compressional stress is the footwall of the listric shift area (LSA), characterised by horizontal compression (Figure 8.5 e). A region of increased tension is developed mainly in the hangingwall, with the extensional axis subhorizontal whereas it is subvertical in the immediate footwall of the LSA (Figure 8.5 d). These observations are consistent with structural features observed across the Karolina Fault (Section 5.2). The most chaotic zone along the Karolina Fault is the footwall of the LSA located in Doringbos Farm, which is drag folded and brecciated along en-echelon set of pinnate faults (black rectangle area in Figure 3.4). Hancock (1985) and Hancock and Barka (1987) describe pinnate joints/faults as structures initiated as tensile features whose orientation defines the maximum compressive stress (σ_1). This implies a northwest-southeast oriented σ_1 (Figure 3.4), which is consistent with horizontal extension attributed to the listric faults when impact-induced overturning is undone. This region correlates with an interpretation that the LSA is subjected to high compressive stresses during the formation of listric faults (Resor and Pollard, 2012). The footwall is characterised by an east-side up, ramp-like structure that thinned quartzite, metapelite, and meta-ironstone units that have been rotated and reoriented north-south (Figures 3.2, 3.4). This area also has the least displacement (down to 800 m) associated with the Karolina Fault because it is adjacent to the decollement zone.

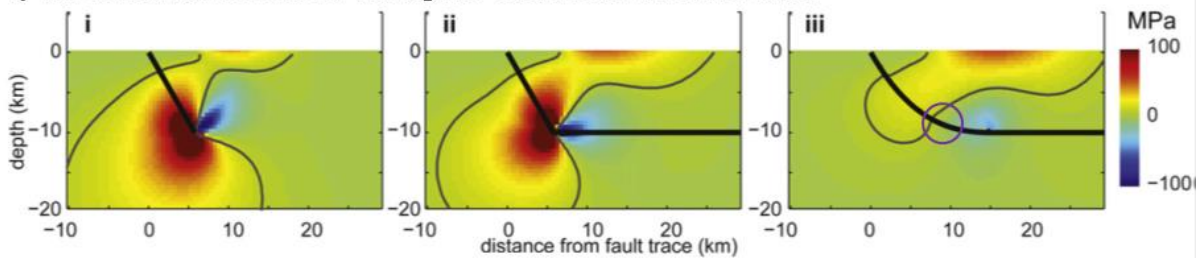
a) Progressive development of a listric fault.



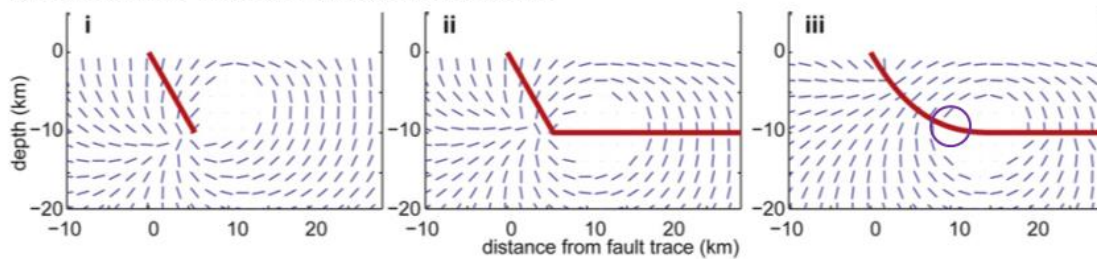
b) Vertical displacement field.



c) The distribution of stress during the formation of a listric fault.



d) Direction of the least compressive stress.



e) Direction of the maximum coulomb stress.

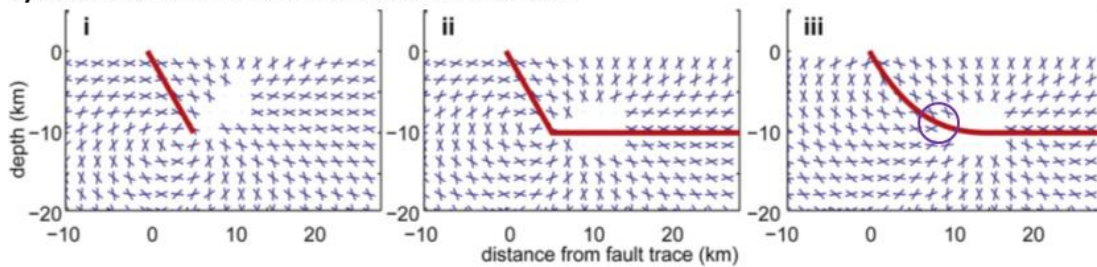


Figure 8.5. A diagram illustrating different stress fields associated with progressive activation of listric faults (modified from Resor and Pollard, 2012). The listric shift area (LSA) is marked with a purple circle on pictures located in the third column.

8.3.3. Comparison between listric faults hosted in the collar and the goldfields

Aside from listric faults developed in the Vredefort Dome, 3-D seismic survey profiles show large listric faults that cut across the Witwatersrand and Ventersdorp supergroups in the West Rand and Carletonville goldfields (Figures 8.6 and 8.7; Tinker et al., 2002; Gibson, 2005, 2010; Manzi et al., 2012a, 2012b, 2013). These faults are reported to have formed through extensional tectonics, related to eruption of the Klipriviersberg Group at ~2.7 Ga (Beach and Smith, 2007; Dankert and Hein, 2010; Jolley et al., 2004, 2007; Manzi et al., 2013; and Tinker et al., 2002). Tinker et al. (2002) suggested that the listric faults hosted in the Vredefort Dome are the southern extensions of listric faults in the goldfields, which are located ~40 km northeast of the dome. This association raised questions on whether the geometry of listric faults in the goldfields and the dome are similar, and whether their sense of displacement is comparable.

If consideration is given to impact rotations, the geometry and scale of the Bank, West Rand, and Karolina listric faults are very similar. Similar to the Bank (BF) and West Rand (WRF) faults, the Karolina Fault (KF) is west-facing and terminates against the base of the Transvaal Supergroup. The displacement is approximately 2 km normal throw along the WRF, up to 2.5 km normal throw along the BF, and up to 3 km left-lateral slip along the overturned Karolina Fault (Figure 8.6; Figure 3.4). The listric shift areas of the WRF and BF are located at 7.5 km and 6 km depths below the BLR respectively, whereas that of the KF is at a maximum depth of 8.5 km below the BLR when the impact-induced rotation is undone. This shows a general increase in depth of location of the LSA with increasing magnitude of displacement (KF>BF>WRF), also noting that shallower parts of these faults were eroded before deposition of the Transvaal Supergroup. All these faults flatten out to form low-angle decollement zones located in the West Rand Group (Figures 8.1, 8.6).

The Bank and West Rand faults have a similar imbricate geometry and are associated with hangingwall strata that define a km-scale rollover anticline (Figures 8.6, 8.7), which is similar to the synformal anticline developed along the Karolina Fault (Figure 3.4). The hangingwall of the WRF hosts synthetic normal faults, and the footwall is crosscut by antithetic faults (Figure 8.7) that have a similar geometry to the Mooiplaas, Koedoesfontein, and De Waal's faults developed in the pre-impact hangingwall of the Karolina Fault (Figure 3.4).

The geometry of listric faults developed in the study area and those in the West Rand and Carletonville goldfields are thus, comparable. It is therefore concluded that the study area hosts Neoarchaean listric faults that were rotated by ~124° during the Vredefort impact at 2.02 Ga. This fault system is interpreted to have been formed by an extensional event during deposition of the Klipriviersberg Group, similar to the WRF and BF systems. In the Vredefort Dome, overturning of the strata did not distort the original structure of the target rocks to a significant degree. The collar of the Vredefort Dome is, thus, an excellent locality to study the nature of the large-scale Neoarchaean listric faults affecting the Witwatersrand Basin.

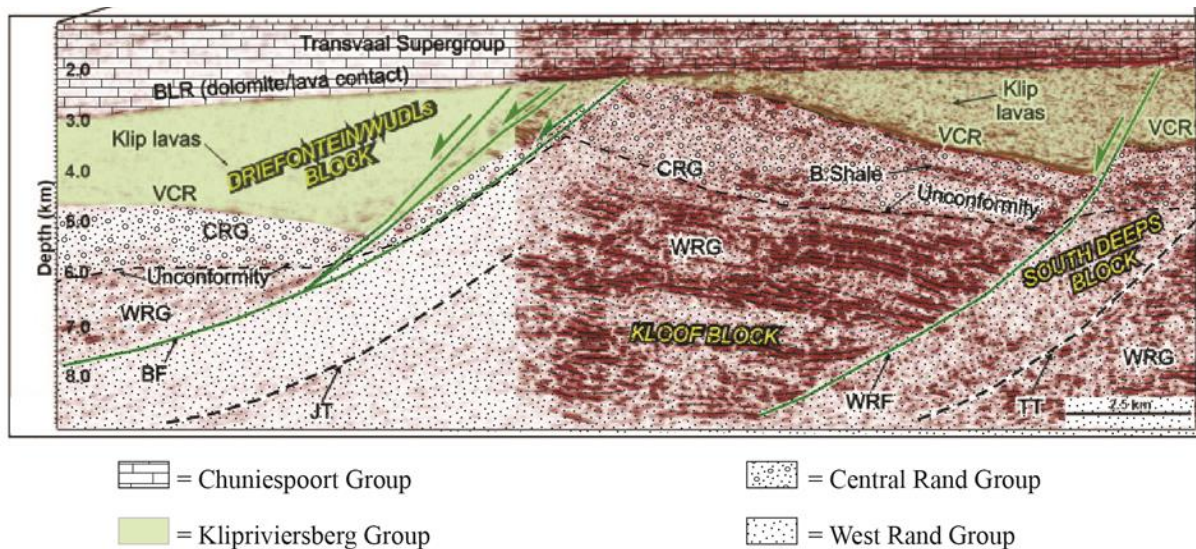


Figure 8.6. Seismic profile of the West Rand and Bank faults terminated against the base of the Transvaal Supergroup (modified from Manzi et al., 2013).

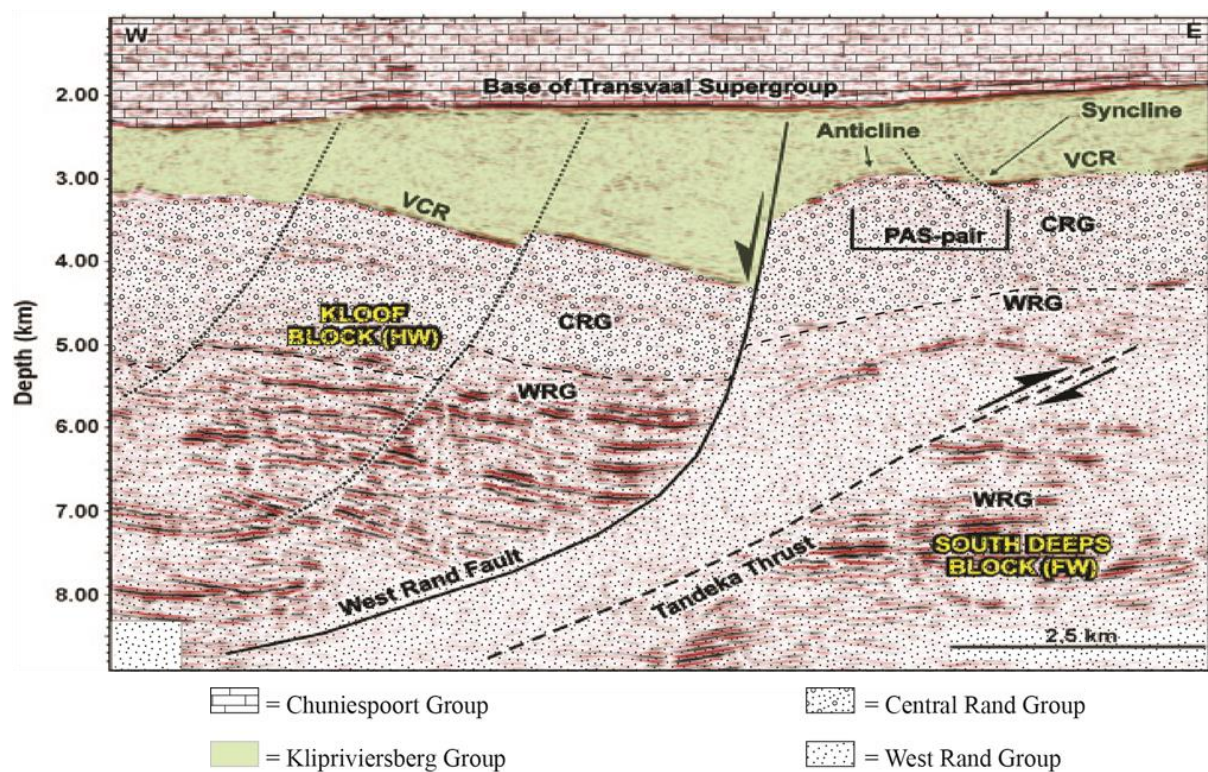


Figure 8.7. Seismic profile showing synthetic faults and antithetic faults (Panvlakte Anticline-Syncline pair = PAS-pair) developed in the hangingwall and footwall of the West Rand Fault respectively (modified from Manzi et al., 2013).

8.4. Metamorphic history of the northern collar of the Vredefort Dome

8.4.1. Introduction

In the Vredefort Dome, rocks of the Witwatersrand Supergroup host higher-grade metamorphic assemblages than the general lower greenschist grade reported from most of the goldfields (Bisschoff, 1982; Phillips and Law, 1994; Gibson and Wallmach, 1995; Foya, 2002). These collar assemblages show an increase in metamorphic grade towards the core of the Vredefort Dome (Gibson and Stevens, 1995; Gibson and Wallmach, 1995; Gibson et al., 2000). Some of the critical metamorphic assemblages described by Nel (1927) and Bisschoff (1982) include kyanite, andalusite, muscovite and chloritoid, developed in knotted quartzite units of the Central Rand Group.

In the Central Rand Group, borehole analysis of the Kimberley Reefs hosted in the collar of the dome shows that the reefs are characterized by chloritoid, with minor pyrophyllite (Foya, 2002). Like Bisschoff (1982), Foya (2002) reported that the Turffontein Subgroup is rich in chloritoid, but unlike the wider Witwatersrand Basin, with minor or no pyrophyllite. Foya (2002) also described impact-induced micro-deformation features that cross-cut the metamorphic assemblages; which have also been altered to a chlorite-sericite assemblage by heat associated with the Vredefort impact. This relationship indicates that the Turffontein Subgroup hosts a chloritoid-bearing peak metamorphic paragenesis in the western and northwestern parts of the collar (M_1 ; Foya, 2002), which is crosscut by a second, impact-related, metamorphic assemblage (Gibson et al., 1998; Gibson, 2002; Ogilvie, 2010). The lack of kyanite and andalusite in the Kimberley Reefs regionally in the Witwatersrand goldfields and surrounding lithologies indicates that kyanite and andalusite are specifically hosted in knotted quartzite horizons that have only been mapped in the northern collar thus far (Section 5.3.1; Bisschoff, 1982).

In the southern part of the study area, petrographic analysis of metapelite units also showed that the lower part of the Witwatersrand Supergroup was subjected to 2 metamorphic events (Gibson and Wallmach, 1995; Gibson et al., 1998). Gibson and Wallmach (1995) proposed that the M_1 metamorphism formed porphyroblasts of biotite, andalusite, cordierite, staurolite, and garnet. The assemblages are consistent with metamorphic minerals described in the West Rand Group of the study area as $M_{2(NC)}$ assemblages (Section 6.2). These $M_{2(NC)}$ peak metamorphic assemblages are proposed to have formed under pressures of 3.5-4.5 kbar and temperatures of 570°C-600°C (Gibson and Wallmach, 1995). In the West Rand Group, $M_{2(NC)}$ metamorphic assemblages are reported to have evolved along an anticlockwise P-T path (Gibson and Wallmach, 1995; Foya, 2002), under peak a geotherm of 35-40 °C/km (Gibson and Wallmach, 1995; Gibson and Jones, 2002). These assemblages are cross-cut by impact-induced brittle fracture cleavage, which is in turn overgrown by $M_{4(NC)}$ metamorphic assemblages described by Gibson and Wallmach (1995) as fine-grained cordierite and biotite.

In the study area, microstructural and petrographic analyses show a polymetamorphic history distinguished into two main events: (1) A dynamo-thermal $M_{2(NC)}$ metamorphic event that is characterised by prograde and retrograde reactions, syn-metamorphic S_2 cleavage, and crosscut by impact-induced structures; and (2) $M_{4(NC)}$ impact-induced metamorphic event that is characterised by shock metamorphism ($M_{4a(NC)}$) and thermal metamorphism ($M_{4b(NC)}$). P-T conditions and reasonable age constraints for these events are discussed in sections below.

8.4.2. *Dynamo-thermal $M_{2(NC)}$ metamorphism in the study area*

8.4.2.1. *P-T conditions of the Central Rand Group*

The lack of pyrophyllite (based on microscope petrology) and the presence of a kyanite-andalusite assemblage in knotted quartzites have significant implications for the metamorphic conditions to which the collar rocks were subjected. The intergrowth of andalusite and kyanite (Section 6.2.2) indicates that kyanite and andalusite were formed at the same time during $M_{2(NC)}$ metamorphism, associated with minimal pyrophyllite. The metamorphic textures indicate that the assemblage was formed under conditions represented along the andalusite-kyanite phase boundary, with minimum pressure conditions of 2.7 kbar and temperatures of approximately 390°C (Spear, 1993 isograd; Figure 8.8). The pressure is equivalent to a minimum of 9.45 km overburden (Johnson and Harley, 2012). An approximately 9.45 km overburden requires a geothermal gradient of about 39 °C/km (370 °C/9.45 km) to reach the 390 °C temperature, assuming that the surface temperature was 20°C. If different experimental data are used for the Kyanite-Andalusite-Sillimanite system (i.e., Dymoke and Sandiford, 1992), the Vredefort kyanite-andalusite assemblage plots at $P=3.4$ kbar and $T = 395$ °C. These conditions require a geothermal gradient of ~31.5 °C/km (375 °C/11.9 km) to produce the assemblage.

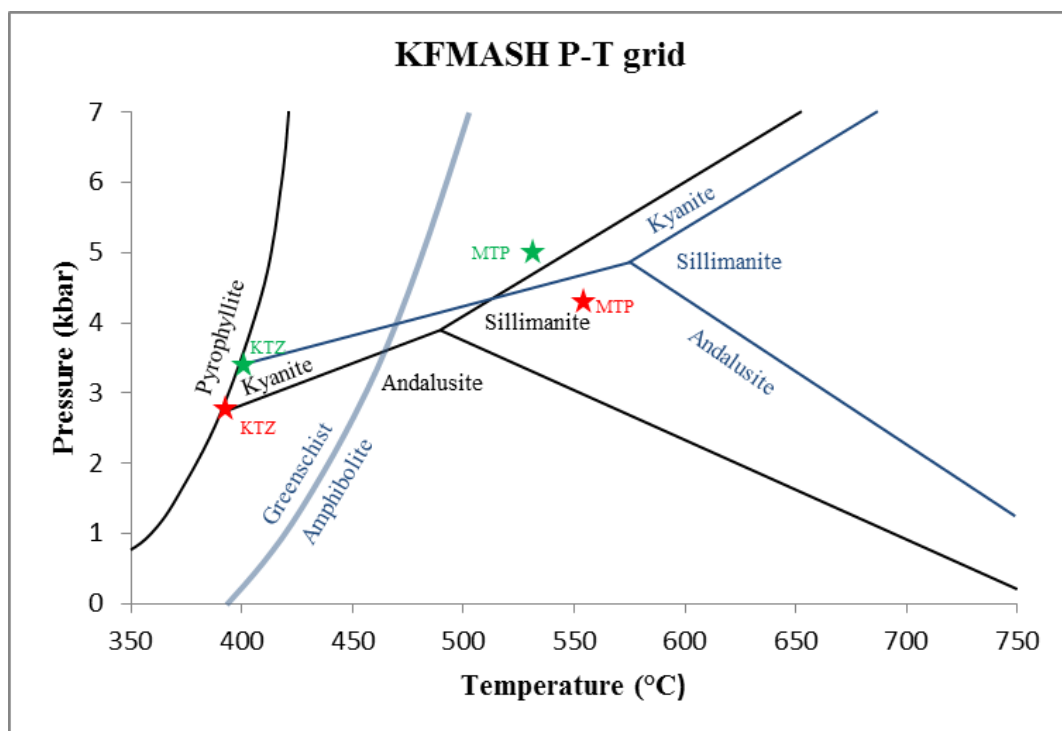
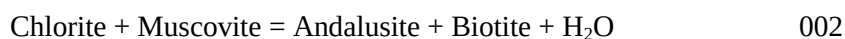
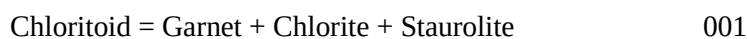


Figure 8.8. KFMASH P-T plots of metamorphic mineral assemblages of knotted quartzite (KTZ) and metapelite (MTP). Green plots are with respect to the P-T graph proposed by Dymoke and Sandiford (1992; blue phase boundaries), whereas the red plots are with respect to the P-T graph proposed by Spear (1993; black phase boundaries).

8.4.2.2. P-T conditions of the lower West Rand Group

Metapelites hosted in the lower West Rand Group are dominated by the assemblage biotite-quartz-andalusite-cordierite-garnet, which reflect an andalusite, cordierite, garnet, and biotite stability field (Figure 8.9). The minerals are stable under P-T conditions of 3.5-4.2 kbar and 500-600 °C (Figure 8.9). In order to relate the metamorphic mineral assemblages to the P-T-t path of $M_{2(NC)}$, reaction textures were related to common mid-amphibolite facies metamorphic reactions (Figure 8.9). The reaction textures show that, after Type 2a biotite began to form, Type2a garnet was formed by reaction 001, which marks an upper termination of the stability field of chloritoid during prograde metamorphism. The continued formation of biotite and andalusite, as well as subsequent formation of cordierite is consistent with reactions 002 and 003 respectively, which terminate the upper stability field of chlorite (Figure 8.8; Spear, 1993). These reactions indicate the prograde path of $M_{2(NC)}$ metamorphic assemblages, reaching peak assemblages dominated by cordierite, garnet, andalusite, and biotite (Section 6.2). The metapelite assemblages show similar mineral reactions to those formed during low pressure anticlockwise metamorphism of high-Al pelites (Spear, 1993).



Chlorite + muscovite + quartz = cordierite (Type 2a) + biotite (Type 2a) + andalusite (Type 2a)
+ H₂O ...003

The prograde metamorphic assemblages show textural evidence of partial replacement by lower temperature assemblages. Type 2b chlorite and muscovite replace earlier (Type 2a) biotite and chloritoid, respectively (Figures 6.9 a, 6.7d), as well as the replacement of hornblende by actinolite in amphibolite-grade mafic rocks (e.g., Figure 6.12a). The pre-impact nature of retrograde actinolite (Type 2b) is evidenced by it defining the S₂ cleavage in meta-dolerite (Sample FMD 231 in Table A3.6, Appendix III); however, there was no timing evidence to distinguish Type 2b muscovite and chlorite from M_{4b(NC)} assemblages, so they may have formed during M_{4b(NC)}.

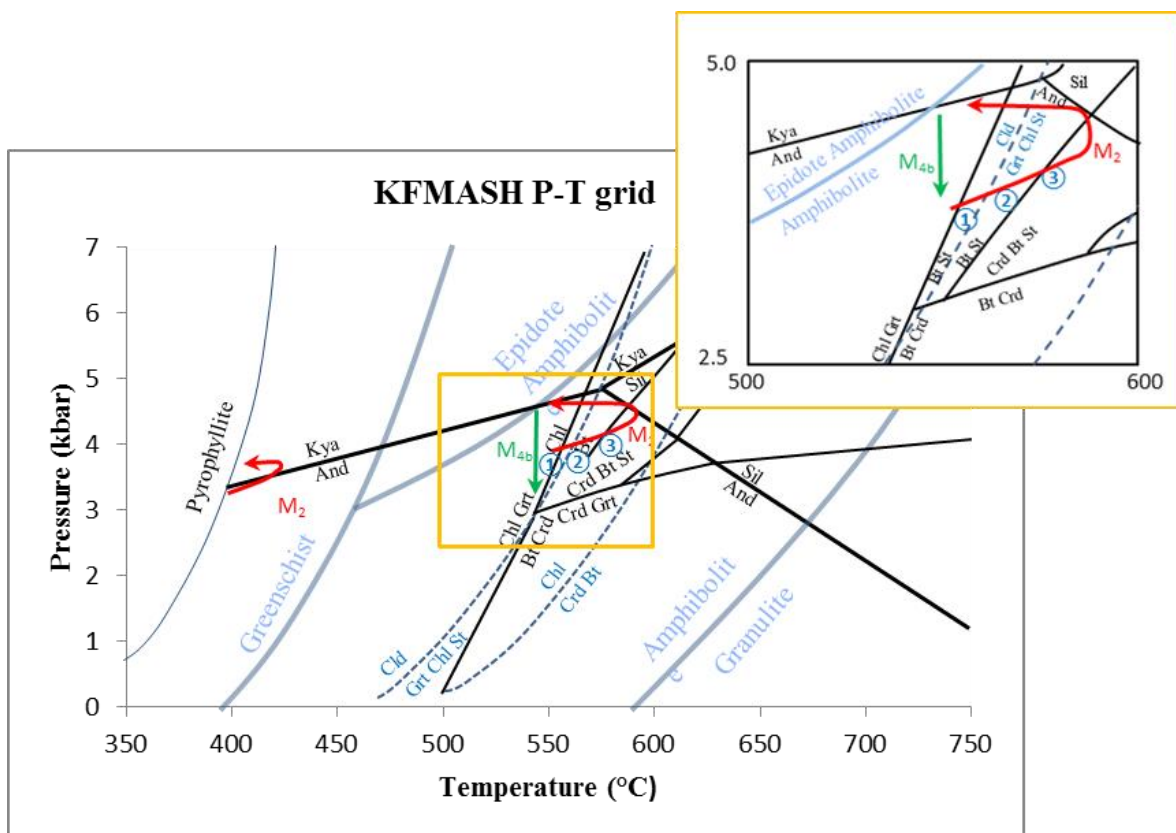


Figure 8.9. Estimated P-T paths for mineral assemblages of knotted quartzite and metapelitic units in the Central Rand and West Rand Groups respectively. The paths are plotted on PT- grids suggested by Spear (1993; blue lines) and Dymoke and Sandiford (1992; black isograds). Some of the reactions associated with the grids are labelled 1-3.

8.4.2.3. Relationship between peak conditions of the lower West Rand and Central Rand groups

A large proportion of $M_{2(NC)}$ metamorphic assemblages are hosted in metapelitic units located at a stratigraphic depth approximately 5 km below the knotted quartzite horizons. Considering the pressure and temperature conditions suggested for the knotted quartzite horizons and the geothermal gradient then, the West Rand metapelite units would have been under pressures of about 3.7 kbar $((9.45+5 \text{ km})/3.5 \text{ km/kbar})$, which corresponds to $T = 563^{\circ}\text{C}$ $(14.45 \text{ km} \times 39^{\circ}\text{C/km})$ under a geothermal gradient of 39°C/km (Spear, 1993). These conditions lie well within the sillimanite stability field (Figure 8.8). However, metamorphic assemblages hosted in the West Rand Group are poor in sillimanite and kyanite (Figure 3.5). The only sillimanite reported in the northern collar is an accessory assemblage developed adjacent to the Schurwedraai alkali granite (Bisschoff, 1982). The lack of sillimanite could be a result of: (a) the low-pressure and high-temperature conditions associated with the $M_{2(NC)}$ metamorphism (Figure 8.8); (b) different plots of the andalusite-kyanite-sillimanite stability fields that show varying metamorphic conditions (e.g., KFMASH by Spear, 1993 vs. Dymoke and Sandiford, 1992); and (c) an effect of the primary composition of the pelites, which varies the temperatures at which some assemblages form (Spear, 1993). The effect of variations in the aluminium silicate boundaries is illustrated by comparison between the Dymoke and Sandiford (1992) and Spear (1993) grids (Figure 8.8), with $M_{2(NC)}$ metapelite assemblages plotting in the kyanite stability field of the latter plot $((11.9+5\text{km})/3.5 = 4.8 \text{ kbar}, 16.9 \text{ km} \times 31.5 = 534^{\circ}\text{C})$. XRF and microprobe analyses are recommended to better understand the metamorphic conditions in the West Rand Group; this was beyond the scope of this study.

The study area was thus subjected to mid-amphibolite to greenschist facies metamorphism increasing in grade with stratigraphic age (Figures 3.5, 6.1), and this is consistent with findings by Phillips and Law (1994), Gibson and Stevens (1995), Gibson and Wallmach (1995), Gibson et al. (2000). These assemblages are post-dated by impact-induced $M_{4(NC)}$ metamorphic assemblages (Section 6.2), which also overprint pseudotachylitic breccia. Type 4b cordierite coronae (Figure 6.2), and possibly chlorite described as Type 2b (Figure 6.9a), formed by thermal metamorphism resulting from the impact are consistent with cordierite, biotite, and chlorite described by Gibson and Wallmach (1995) as impact-induced assemblages. The $M_{2(NC)}$ metamorphic assemblages hosted in both the West Rand and Central Rand Group are associated with ductile deformation (Section 5.3).

8.4.2.4. Relationship between $M_{2(NC)}$ metamorphism and deformation

Williams (1994) proposed that there is a significant link between metamorphism and deformation in different (contact and regional) metamorphic terranes of varying scales. This proposition was based on the presence of similar sigmoidal inclusions and porphyroblast relationships, which are hosted in distinct metamorphic assemblages formed by two plutons in central Arizona and by regional metamorphism in northern New Mexico (Figure 8.10). The relative rotation of inclusions vs. porphyroblasts and flattening/alignment of minerals are indicative of the sequence of metamorphism and deformation, with sigmoidal inclusions indicative of syn-metamorphic deformation (Figure 8.11).

In the study area, inclusion trails are hosted in garnet and andalusite porphyroblasts (Section 5.3). The inclusions are both planar (Figure 5.5a) and sigmoidal (Figures 5.14, 8.11), the latter consistent with syn-metamorphic foliation described by Williams (1994). The magnitude of apparent rotation of planar S_x is different in the porphyroblasts (compare Figures 5.5 a & c), consistent with a heterogeneous strain distribution (Bell, 1981). In the West Rand Group, syn-metamorphic deformation is represented by $M_{2(NC)}$ cordierite, andalusite, and garnet porphyroblasts that are wrapped by an S_2 cleavage, with the latter showing relative cleavage rotation (e.g., Figure 5.5). In the knotted quartzite, kyanite contains planar to slightly curved inclusion trails and is wrapped by S_2 (e.g., Figure 5.9a); chloritoid shows a similar texture in the West Rand Group (Figure 5.5d). The inclusion patterns and external foliation are consistent with deformation that is syn- to post- metamorphism (Figures 8.11, 8.12), as illustrated by Shelley (1993). S_x is thus interpreted to be an earlier S_2 cleavage that was included in $M_{2(NC)}$ porphyroblast as $D2$ - $M_{2(NC)}$ evolved.

The metamorphic assemblages are similar to syn-tectonic cordierite, andalusite, staurolite, and garnet porphyroblasts described by Gibson (1993) in the Hospital Hill Subgroup, which contain helicytic S_1 inclusion trails (Gibson, 1993). However, Gibson (1993) was not able to determine a consistent orientation of S_1 and S_2 , most likely because of annealing and recrystallization reactions related to the post-impact thermal event.

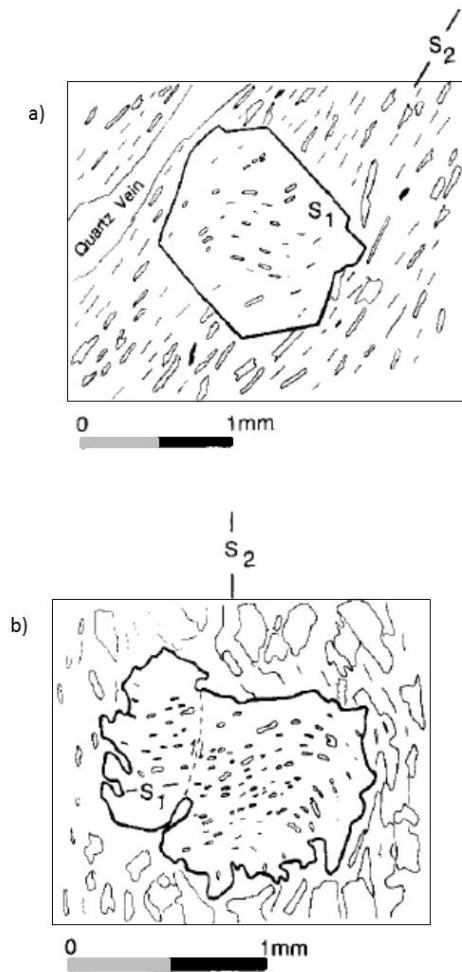


Figure 8.10. Sigmoidal inclusion trails hosted in garnet porphyroblasts from the Crazy Basin (Central Arizona) and Cerro Colorado area (Tusas Mountains) respectively (Modified from Williams, 1994).

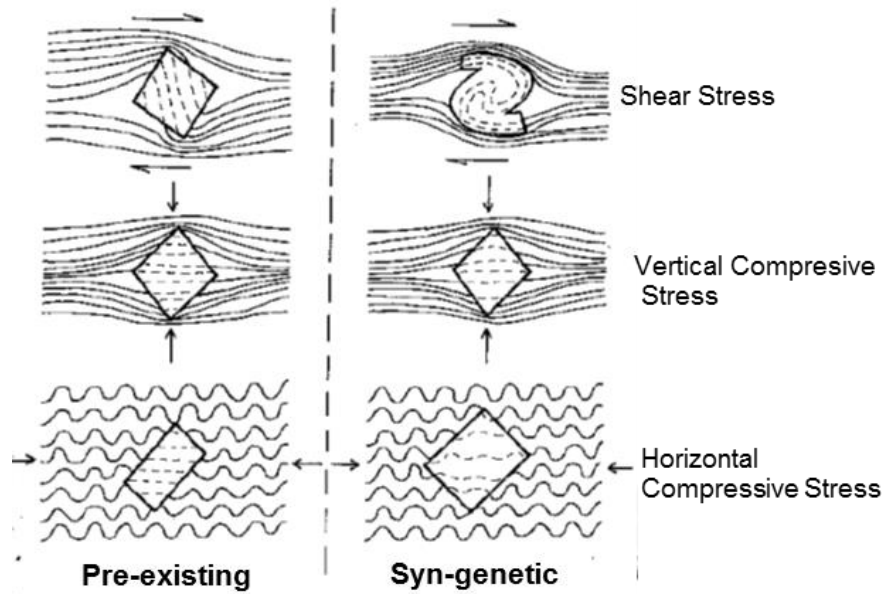


Figure 8.11. A sketch illustrating textures of a metamorphic mineral that formed pre- and syn-genetic to foliation, respectively (modified from Shelley, 1993).

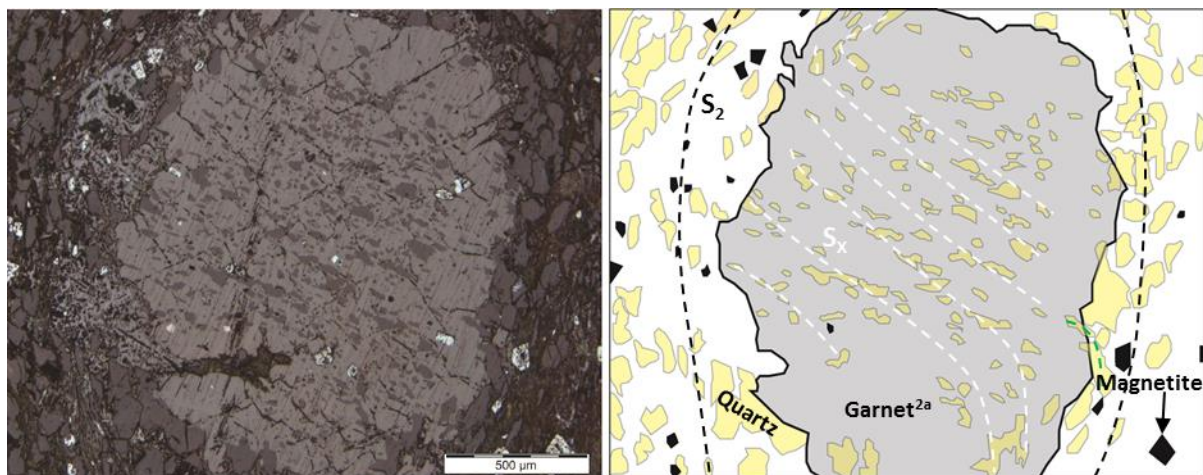


Figure 8.12. A representative sketch of a photomicrograph of an $M_{2a(NC)}$ garnet, which hosts an earlier sigmoidal cleavage S_x .

8.4.3. Age constraints to $M_{2(NC)}$ metamorphism

Major deformation and thermal events that occurred after the deposition of the Witwatersrand Supergroup and before the Vredefort impact are: (a) the Meso- to Neoarchaeon Umzawami event; (b) the Neoarchaeon Ventersdorp-age event; (c) the Paleoproterozoic Bushveld-age event; and (d) the Paleoproterozoic orogenic event (Transvaalide/Ukubambana event).

8.4.3.1. $M_{2(NC)}$ metamorphism as a Meso- to Neoarchaean event

Brink et al. (2000), Jolley et al. (2004, 2007), Beach and Smith (2007) and Manzi et al. (2013) proposed that the Umzawami Event was initiated pre- to syn-deposition of the Ventersdorp Supergroup (Figure 2.7). At this time, knotted quartzite horizons were located at an approximate depth of ~0.5 km (Turffontein Subgroup; Figures 4.2, 8.13). The depth is equivalent to pressure conditions approximately equal to 0.14 kbar. Assuming that the surface temperature was 20°C, a geothermal gradient of 750°C/km (375°C/0.5 km; Table 8.1) is required to reach the minimum metamorphic temperature estimated for the kyanite-andalusite assemblage ($P=3.4$ kbar and $T=395^{\circ}\text{C}$; Figures 8.8, 8.9). These thermal conditions are extremely unrealistic compared to the expected cratonic geothermal gradient of 15-20°C/km (Gibson and Jones, 2002); the pressure was too low to form kyanite regardless of which andalusite-kyanite-sillimanite stability field is used (Figure 8.8).

Considering the regional stratigraphy (Figure 4.2), pelitic units in the West Rand Group would have been located at a depth of about 6 km; an equivalence of 1.7 kbar (Figure 8.13; Table 8.1). At this depth, it would have required a geothermal gradient of about 92 °C/km (550 °C / 6 km) to form the mid-amphibolite facies hosted in the West Rand Group ($P = 3.5 - 4.5$ kbar and $T = 570 - 600$ °C; Figure 8.9). This is still an unrealistically high geothermal gradient for an Archaean craton. It is, thus, unlikely that thermal metamorphism formed the $M_{2(NC)}$ metamorphic assemblages during the Umzawami Event.

	Meso- to Neoarchaean		Ventersdorp age		Bushveld age		Paleoproterozoic orogeny	
	Knotted qtzite	Metapelite	Knotted qtzite	Metapelite	Knotted qtzite	Metapelite	Knotted qtzite	Metapelite
Overburden thickness (km)	0.5	6	5.7	11.2	12	17	12	17
Pressure (kbar)	0.14	1.7	1.6	3.2	3.4	4.8	3.4	4.8
Temperature (°C)	395	570	395	570	395	570	395	570
Geothermal gradient (°C/km)	750	92	66	49	31	32	31	32

Table 8.1. Estimated metamorphic conditions (overburden thickness, pressure, temperature, and geothermal gradients) during: (1) the Meso- to Neoarchaean Umzawami event; (2) Ventersdorp-age magmatism; (3) Bushveld-age magmatism; and (4) Paleoproterozoic orogenic event.

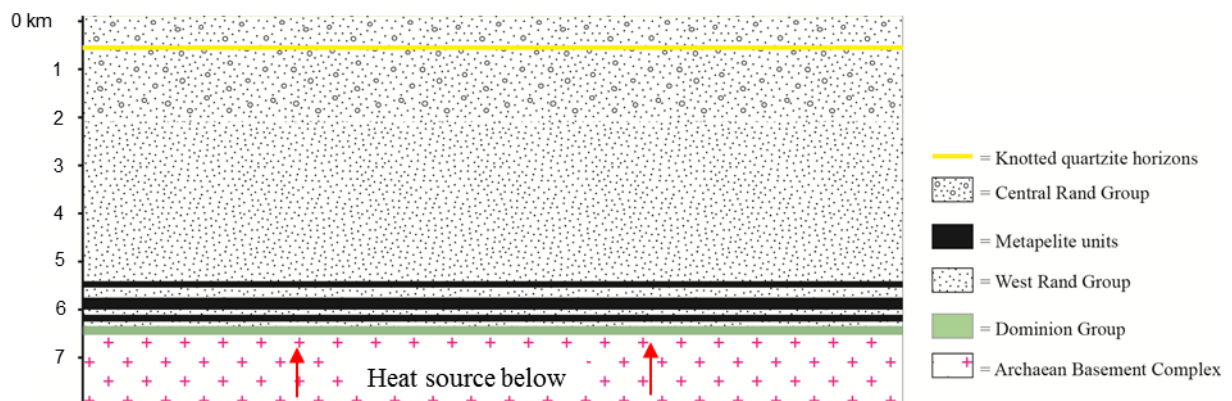


Figure 8.13. A schematic cross-section of the study area at the beginning of the Umzawami Event, showing the stratigraphic location of the knotted quartzite horizons and metapelite units.

8.4.3.2. $M_{2(NC)}$ metamorphism as a Neoarchaeon, Ventersdorp-age event

The Ventersdorp event produced lavas up to 5.1 km in thickness, as well as an extensive sill and dyke system in the underlying Witwatersrand Supergroup. Knotted quartzite horizons would have been at a depth of ~5.7 km (.05+5.1 km) at the end of deposition of the Klipriviersberg Group. Kimberlite xenoliths from the southwestern pipes (Lace and Voorspoed mines) show that the lower crust experienced ultra-high-temperature metamorphic conditions (1000 °C, 11 kbar; Dawson and Smith, 1987; Dawson et al, 1997; Schmitz and Bowring, 2003) during eruption of the lavas, the conditions indicate an average crustal geotherm of 26 °C/km. Assuming that the surface temperature was 20 °C and the 5.7 km thick overburden for the knotted quartzites, the quartzite horizon would have been subjected to temperatures ~168 °C and pressures ~1.6 kbar (Figure 8.14; Table 8.1). A geothermal gradient of 66 °C/km (375°C/5.7 km) is required to reach the minimum metamorphic temperature estimated for the kyanite-andalusite assemblage (P=3.4 kbar and T=395°C; Figure 8.9). This geothermal gradient is more than twice the gradient estimated for Ventersdorp-age metamorphism (Dawson and Smith, 1987; Dawson et al, 1997; Schmitz and Bowring, 2003), and the pressure is too low to form kyanite.

Pelitic units of the West Rand Group would have been at a depth of about 11.2 km after eruption of the Klipriviersberg lavas. The pelitic units would have been subjected to temperatures of ~311 °C and pressures of ~3.2 kbar. It would have required a geothermal gradient of about 49 °C/km (550°C/11.2 km; Table 8.1) to cause the mid-amphibolite facies metamorphism at this depth (P = 3.5 – 4.5 kbar and T = 570 – 600 °C; Gibson and Wallmach, 1995; Figure 8.9). The pressure conditions are close to the minimum required to form the mid-amphibolite-grade facies, however, the geothermal gradient required to reach temperatures estimated for the mid-amphibolite facies is almost double the gradient estimated for Ventersdorp-age metamorphism. This means that with a Klipriviersberg overburden of ~5.1 km, metamorphosed units of the Central Rand and West Rand groups would have required approximately double the geothermal gradient estimated for the Ventersdorp-age metamorphism, to attain greenschist-grade and mid-amphibolite-grade facies in the groups, respectively. Also, Ventersdorp-age sills in the West Rand Group are metamorphosed to mid-amphibolite-grade facies ($M_{2(NC)}$; Section 6.2), indicating that they were not the heat source for $M_{2(NC)}$ metamorphism.

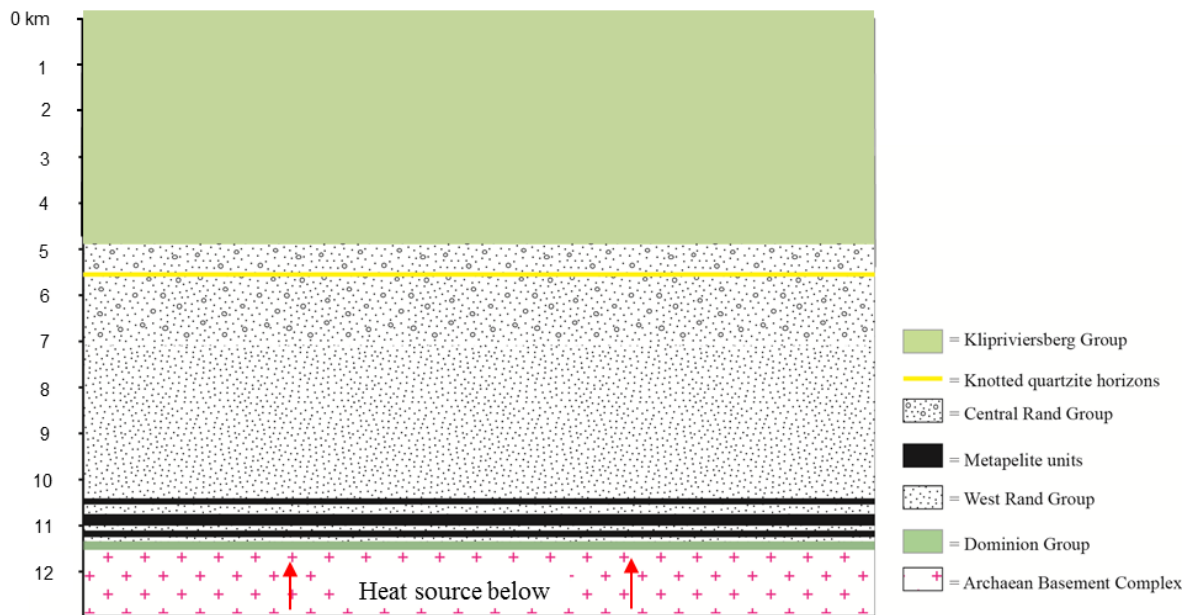


Figure 8.14. A schematic cross-section of the study area during eruption of the Klipriviersberg lavas, showing the stratigraphic location of the knotted quartzite horizons and metapelite units.

8.4.3.3. $M_{2(NC)}$ metamorphism as a Paleoproterozoic, Bushveld-age event

During intrusion of the Bushveld Complex at 2.06 Ga (Walraven, 1997), the Central Rand Group was already overlain by the Ventersdorp and Transvaal supergroups (Figure 8.15), adding up to overburden thickness of ~12 km (about 4 km of the eroded Ventersdorp Supergroup plus 8 km of the Transvaal Supergroup; Eriksson and Reczko, 1995; Eriksson et al., 1995; Reimold and Gibson, 2006). This is a depth equivalent to up to 3.4 kbar, which would require a geothermal gradient of about 31°C/km (375°C/12km) to form a kyanite-andalusite assemblage (Figure 8.9; Table 8.1). This pressure is high enough to be associated with more realistic, lower geothermal gradient than the estimated geothermal gradient of 35-40°C/km during $M_{2(NC)}$ metamorphism (Phillips and Law, 1994; Gibson and Stevens, 1995; Gibson and Wallmach, 1995).

The West Rand Group pelites would have been at depths of up to 17 km, which is equivalent to up to 4.8 kbar (Figure 8.15). At this depth a geothermal gradient of 32 °C/km depending on the peak temperature of the mid-amphibolite assemblages hosted in the West Rand Group (550 °C/ 17 km; Table 8.1). Estimated $M_{2(NC)}$ temperatures (570 – 600 °C; $M_{3(REG)}$ in Gibson and Wallmach, 1995) could be attained in these conditions. However, such a pressure would restrict the aluminosilicate polymorph to kyanite and sillimanite. As andalusite is the stable polymorph in the West Rand Group metapelites, a slightly lower pressure and, therefore, slightly higher geotherm is more acceptable. As a result, the Bushveld magmatism is proposed as an event that was associated with the most suitable metamorphic conditions for the formation of $M_{2(NC)}$ greenschist to mid-amphibolite facies hosted in the Central Rand and West Rand groups respectively. Also, the actinolite-plagioclase-epidote±hornblende foliation developed in meta-dolerite indicates that the rocks were still being

metamorphosed after the Ventersdorp-age sills were intruded (e.g., Sample FMD231 in Table A3.6, Appendix III). The event is consistent with interpretations by Bisschoff (1982), Phillips and Law (1994), and Gibson and Wallmach (1995); who also attributed the peak metamorphic assemblages hosted in the wider Witwatersrand Basin to Bushveld magmatism.

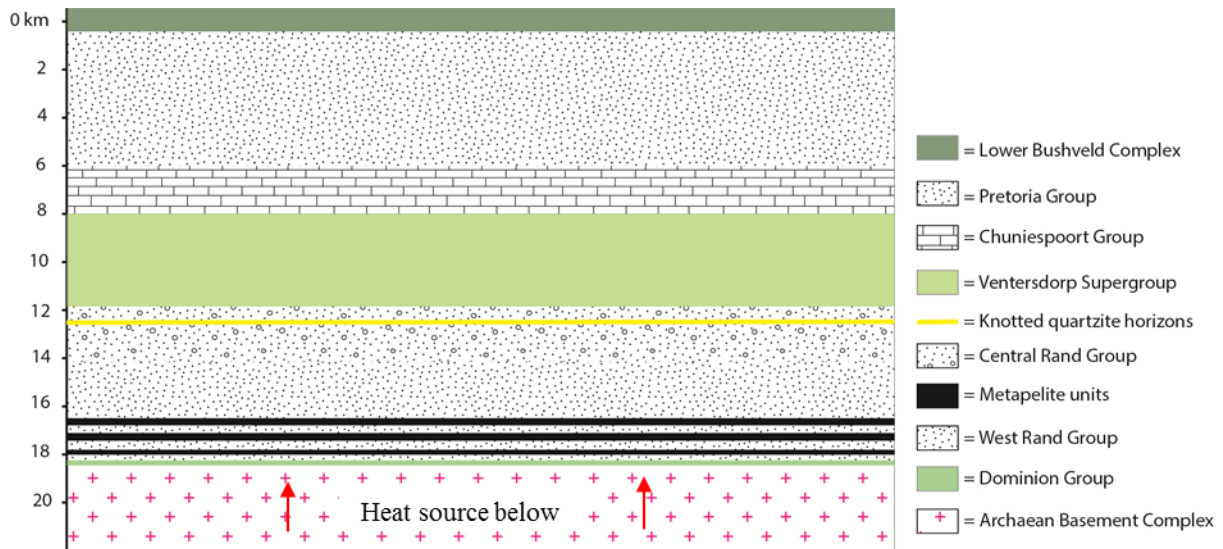


Figure 8.15. A schematic cross-section of the study area during magmatism associated with the Bushveld Complex, showing the stratigraphic location of the knotted quartzite horizons and metapelite units.

Uken and Watkeys (1997) and Cawthorn et al. (2006) proposed that downsagging resulting from the large size and density of the Bushveld Complex was the cause of syn-metamorphic deformation. The timing of deformation is consistent with textures and crosscutting relationships in the study area, however the sense of shear is not. For downsagging, shear sense will be different depending on location relative to the Bushveld Complex. In contrast, the S_2 foliation in the collar of the Vredefort Dome and Johannesburg Dome shows consistent vergence generally directed north to NNE (Roering, 1986; Hilliard, 1994; Hilliard and McCourt, 1995; Gibson et al., 1999; Bush, 2014), with NW-directed vergence in the study area due to local variation in strike rotation caused by the impact (Sections 5.3).

8.4.3.4. $M_{2(NC)}$ metamorphism as a Paleoproterozoic orogenic event

Generally consistent, Northwards S_2 vergence is indicated in the northern collar (Bush, 2014) and the Johannesburg Dome (Roering, 1986; Hilliard, 1994; Hilliard and McCourt, 1995; Gibson et al., 1999; Alexandre et al., 2006) that is ~100 km northeast of the study area. This vergence and timing is most consistent with a compressive deformation that Andreoli (1988) and Dankert and Hein (2010) termed the Transvaalide and Ukubambana events, respectively (2.2 – 2.0 Ga). During this

event, minimum metamorphic conditions would have been similar to those discussed in subsection 8.4.3.4 (Figure 8.15; Table 8.1), during Bushveld magmatism. Unlike previous interpretations by Dankert and Hein (2010) that the regional S_2 shows minimal effect on Bushveld-age intrusions, evidence indicates that S_2 affects granite intrusions (Bush, 2014). Additionally, it is evident in the study area that S_2 predates the Vredefort impact, thus, restricting deformation to syn- $M_{2(NC)}$ and not later than 2.02 Ga (pre-dates the Vredefort impact).

8.4.4. $M_{4(NC)}$ metamorphism

The initial $M_{4a(NC)}$ metamorphism recrystallized pseudotachylitic breccia matrix and annealed impact-induced micro-faults (Gibson and Wallmach, 1995; Figure 6.13). $M_{4b(NC)}$ assemblages have only been identified in pelitic units of the West Rand Group in this study. Like observations by Gibson and Wallmach (1995), the most obvious $M_{4b(NC)}$ reaction is shown by Type 4b cordierite (Figure 6.2, 6.3). Type 4b cordierite forms coronae around replaced Type 2a cordierite porphyroblasts, coronae that also overprint Type 2a biotite. Type 2a biotite is part of an assemblage that defines the S_2 cleavage, indicating that the Type 4b cordierite coronae are indeed post $M_{2(NC)}$ - S_2 . Other suggested reactions for $M_{4b(NC)}$, which do not exhibit solid evidence that they are $M_{2b(NC)}$ or $M_{4b(NC)}$, are Type 2b muscovite and chlorite (Figure 6.7d, 6.9a). Furthermore, Nemutandani (2014) reported a cordierite-biotite assemblage that overprints pseudotachylitic breccia 10 km east of the study area, consistent with $M_{4b(NC)}$ metamorphism.

Metamorphic conditions for $M_{4b(NC)}$ are estimated to be temperatures up to 525 °C in the West Rand Group (Gibson et al., 1998; Gibson, 2002; Ogilvie, 2010) and 300 °C in the Central Rand Group (Foya, 2002).

8.4.5. Summary

The study area hosts mid-amphibolite to greenschist facies metamorphic assemblages that were formed by an $M_{2(NC)}$ metamorphic event, which is post-dated by lower-grade assemblages formed by the impact-induced $M_{4(NC)}$ metamorphism. $M_{2(NC)}$ metamorphism in the lower West Rand Group is characterised by an earlier prograde path that formed peak assemblages dominated by Type 2a cordierite, andalusite, biotite, and garnet ($M_{2a(NC)}$) followed by a retrograde path that formed replacements of the prograde assemblages ($M_{2b(NC)}$, staurolite and biotite). The $M_{2(NC)}$ metamorphism is also associated with the S_2 -C and S_2 foliations. This association has been identified in $M_{3(REG)}$ assemblages described in separate studies by McCarthy et al. (1986), Gibson (1993) and Bush (2014). Considering both burial depth and the syn-metamorphic geotherm, metamorphic conditions of the mid-amphibolite to greenschist assemblages are most consistent with development during 2.06 Ga Bushveld magmatism. It is during the emplacement of the Bushveld Complex that overburden and high heat flow were sufficient to create ideal metamorphic conditions for the assemblages. The $M_{2(NC)}$

metamorphic assemblages were subsequently crosscut by $M_{4b(NC)}$ assemblages which also decrease in grade outwards as a result of differential uplift and shock heating levels.

8.5. D4 Impact-induced structures

Meteorite impact processes, as described by French (1998), form shatter cones, PDF, shock metamorphism, shock/shatter cleavage, pseudotachylitic breccia, impact melt-rock, suevite, lithic impact breccia, dyke breccias (e.g., Figure 3.3a); as well as folds and faults during the modification stage. The magnitude and extent of these impact features is dependent on the size of the impact. Most of these features have been described in the Vredefort Dome, except for relatively shallow features like suevite, lithic impact breccia, and dyke breccias because of the eroded nature of the Vredefort Dome.

8.5.1. Shock features

Shock features were developed during the contact-compression phase of the impact (French, 1998). Shock features developed in the collar include shatter cones, shock metamorphism ($M_{4a(NC)}$), shock/shatter cleavage, coesite and stishovite (Martini, 1978, 1992), PDFs, pseudotachylitic breccia and impact melt. Shatter cones are hosted throughout the study area (Section 5.5.5), consistent with the suggestions by Grieve et al. (1990) and French (1998) that shatter cones can form at pressures as low as 4 GPa; relative to PDF that form between 8 and 25 GPa (French, 1998). PDFs are developed in the study area; however they were not given much attention because of the wide scope of this study. Shock also produced impact melt, which later intruded tangential joints and lithified to form the granophyre dykes (Gibson et al., 1998; Gibson and Reimold, 2000; Reimold and Gibson, 2006).

8.5.2. Faults and folds

Structural analysis of the study area indicates that the northern collar hosts generally NNW-trending impact-induced faults (Figure 3.4). Displacements along impact-induced faults show a tangential orientation of the maximum compressional axis hosted in older quartzite units of the Hospital Hill Subgroup; a geometry consistent with the tangential shortening model suggested by Lilly (1978) and Wieland et al. (2005). The angle bisected by the compressional axis varies from 60° to 100°, which makes it harder to predict the orientation of the compressional axis when only one set of faults is exhibited. Younger rocks towards the north (Central Rand Group, Buffelshoek area) host faults that indicate a more variable orientation of the compressional axis, with a number of faults showing a north-south oriented compressional direction (Figure 3.4). This is most likely a result of interference with the adjacent D1 faults during formation.

Collins et al. (2004) and Wieland et al. (2005) reported that tangential shortening occurs towards the end of central uplift formation. The faults formed during upwards and inwards displacement (constriction) of the strata, before outward collapse of the central uplift. It is during this

phase that some of the joints are interpreted to have formed; however joints were not classified, which makes it challenging to distinguish them from those developed during outward collapse of the crater. The brittle phase of central uplift formation most likely formed kink bands and spaced fracture cleavage. It is during this stage that the radial and oblique displacements shorten the overturned strata (Collins et al., 2004; Wieland et al., 2005). The few D4 faults surfaces mapped host a single set of striations, which represent tangential shortening because D4 faults collectively show generally strike-slip (Section 5.5.2), consistent with increasing shortening of strata towards the south described by Wieland et al. (2005).

D4 folds are also interpreted to have formed during tangential shortening, as deformation was evolving from brittle-ductile to mostly brittle (French, 1998). This is evidenced by the similar sense of shear the folds have to D4 folds (Section 5.5.1). This is consistent with interpretations by Wieland et al. (2005).

8.5.3. Pseudotachylitic breccia

The largest pseudotachylitic breccias in the collar of the Vredefort Dome are reported to be hosted along impact-induced faults, hinges of impact-induced folds, and bedding planes (Wieland et al., 2005; Reimold and Gibson, 2006). This is consistent with pseudotachylitic breccia hosted in the study area; which are: (a) hosted along D4 faults and folds; and (b) associated with displacements of the S_2 -C and S_2 cleavages and host $M_{2(NC)}$ metamorphic minerals. Although all pseudotachylitic breccia mapped shows evidence for impact-origin, there seems to be multiple generation stages of the breccias. This is shown by pseudotachylitic breccia (Rhq3 unit) that has been displaced by impact-induced faults, some of which host irregular pseudotachylitic breccia (Section 7.3). This might indicate that the pseudotachylitic breccias were generated at different stages of the impact, however this would need further investigation as it was beyond the scope of this study.

Suggested mechanisms for the generation of pseudotachylitic breccias include shock brecciation/melting, shock decompression, and/or frictional melting (Reimold and Gibson, 2006). Proposed models of pseudotachylitic breccia generation in the Vredefort Dome have evolved from fluid/gas fluxing (Shand, 1916), to shock brecciation and melting (Schwarzman et al., 1983; Martini, 1991), to frictional melting (Manton, 1962). The study of these models involved field, mineralogical, and chemical evidence; an analytical level that was out of the scope of this study. Therefore, the proposed model (s) of pseudotachylitic breccia generation is based on cross-cutting relationships, petrographic analysis, and literature; which are limited data to confidently conclude on the generation model.

Textures and composition of pseudotachylitic breccias generally show that they did not form in situ, particularly those hosted in the West Rand Group. This is shown by some heterogeneous clast compositions and flow textures that indicate migration from the point of generation (Section 7.3). The

veins are generally characterised by finer-grained, banded margins that are indicative of heterogeneous flow in planar pseudotachylitic breccia (Figure 7.6). The aforementioned characteristics indicate that melt was mobile, without evidence of in-situ melt generation. This could indicate frictional melting, whereby the generated melt is channelled through faults and joints from the source. Additional techniques are needed to conclude on the model for pseudotachylitic breccia generation in the study area.

8.5.4. Outward collapse features

Outward collapse is the last stage that occurs during the formation of the centrally uplifted region (Grieve et al., 1981; French, 1998). Outward collapse of the central uplift is evidenced by sedimentary structures that indicate that the strata have been overturned, inverting the younging direction of the rocks (Section 4.2.3). The collapse formed joints (Wieland et al., 2005) and overturned earlier impact-induced features. Also, the collapse overturned pre-existing Neoarchaean listric faults (D1) and the S₂ foliation to their sides. Judging from the present nature of the pre-impact faults, it seems the impact event did not catastrophically deform these structures.

8.5.5. The granophyre dyke

Homogeneous impact melt, generated during the compression phase of target rocks, intruded radial and tangential joints (central uplift formation features) to form the granophyre dykes. The homogeneous nature is proven by evidence that the dykes have a uniform bulk composition (Therriault et al., 1996, 1997; Reimold and Gibson, 2006). The granophyre dyke mapped in the study area cuts across all structures in the study area (Figure 3.2) and is not crosscut by any pseudotachylitic breccia (Reimold and Gibson, 2006), which makes it the youngest feature related to the impact event, formed after outward collapse of the centrally uplifted region (French, 1998).

9.0 CONCLUSION

9.1. Introduction

Lithological, structural and metamorphic studies of the northern collar of the Vredefort Dome show that the collar rocks have been subjected to multiple deformation and metamorphic events. Despite meteorite impacts being highly complex structural events, careful mapping of features has established crosscutting relationships that indicate that rocks in the study area were subjected to pre-impact deformations and metamorphism. The structural and metamorphic complexity of the northern collar is, thus, not only attributed to the Vredefort impact but also to the structural architecture and metamorphic state of the target rocks before the impact event. These crosscutting relationships, combined with comparative structural studies of the Witwatersrand Basin, show that the deformational history of supracrustal rocks in the northern collar of Vredefort Dome spans from 2.70 to 2.02 Ga.

9.2. The geological history of the northern collar of the Vredefort Dome

9.2.1. *Neoarchaeon craton-wide extension (2.70-2.64 Ga)*

The earliest deformation event recorded in the study area is a Neoarchaeon extensional event that occurred after the deposition of the Dominion Group and Witwatersrand Supergroup, and syn- to post-deposition of the Klipriviersberg Group. The event occurred on a craton-wide scale, and is recorded in both the Kaapvaal and Pilbara cratons (Hlukana-Platberg and Fortescue events respectively; c.f., Tinker et al., 2002; De Vries et al., 2008). In both cratons, the extensional event was initiated by the eruption of basaltic lavas (e.g., Klipriviersberg and Fortescue groups, respectively). The crustal extension formed listric normal faults of variable orientation across the Kaapvaal and Pilbara cratons. In the Kaapvaal craton, the listric faults displaced the Witwatersrand Supergroup and basaltic units of the Klipriviersberg Groups (e.g., Bank, West Rand, and Karolina Fault systems; Figures 8.6, 8.7, and 8.1).

Arcuate strike-slip faults mapped in the northern collar are interpreted as overturned conjugate listric faults with an original ~NNE strike and predominantly west-side-down slip. These faults are similar to those described by Tinker et al. (2002), Jolley et al. (2004, 2007), Beach and Smith (2007), Dankert and Hein (2010), and Manzi et al. (2013) in the Carletonville and West Rand goldfields, characterised by normal displacements typically 2-3 km in magnitude (Section 8.3.3). Displacements led to the formation of hangingwall rollover anticlines and drag folds, with a decollement zone developed in the pelitic units of the West Rand Group (Section 5.2).

The extensional event was preceded by the Umzawami Compressional Event in the Witwatersrand Basin (Dankert and Hein, 2010); however, structural features formed by this event are not exhibited in the study area.

9.2.2. $M_{2(NC)}$ Metamorphism and Paleoproterozoic orogenic event (2.20-2.02 Ga)

Elevated thermal conditions during the intrusion of the Bushveld Complex formed regional greenschist- to amphibolite-grade facies metamorphism in the study area ($M_{2(NC)}$) and the entire Witwatersrand Basin ($M_{3(REG)}$; c.f., Phillips and Law, 1994). The metamorphism formed prograde and retrograde $M_{2(NC)}$ metamorphic assemblages hosted in the collar of the Vredefort Dome. The assemblages are mid-amphibolite grade in the lower West Rand Group (cordierite, andalusite, garnet, staurolite, biotite, actinolite, epidote, muscovite, sericite, chlorite assemblages) and greenschist grade in the Central Rand Group (kyanite, andalusite, chloritoid, sericite, muscovite assemblages) of the Vredefort Dome. The $M_{2(NC)}$ metamorphic assemblages were formed simultaneous to an S_2 and S_2 -C cleavage (Sections 5.3, 8.4), forming a regional shear foliation that verges towards the NW. When strike rotation by the impact is accounted for, this vergence is consistent with generally northward-directed vergence of post-Transvaal Supergroup foliation described in the northeastern collar of the Vredefort Dome (Bush, 2014) and in the Johannesburg Dome (Roering, 1986; Andreoli, 1988; Hilliard, 1994; Hilliard and McCourt, 1995; Gibson et al., 1999; Alexandre et al., 2006). The generally north-verging foliation was formed during a Paleoproterozoic orogeny that Andreoli (1998) and Dankert and Hein (2010) termed the Transvaalide and Ukubambana events (2.2 – 2.0 Ga; Dankert and Hein, 2010), respectively. Unlike previous interpretations that the foliation shows minimal effect on Bushveld-age intrusions (Dankert and Hein, 2010), there is evidence that the foliation post-dates Bushveld contact metamorphism (Gibson et al., 1999; Alexandre et al., 2006) and Bushveld intrusions (Bush, 2014); as a result the Ukubambana Event might have lasted longer than it is currently reported. Crosscutting relationships in the study area reflect a deformational period of 2.06 Ga to no less than 2.02 Ga.

The S_2 foliation is crenulated by a later, unidentified S_3 cleavage. The S_3 crenulations can be attributed to either late stages of the Ukubambana Event or early stages of the Vredefort impact, before the spaced fracture cleavage was formed.

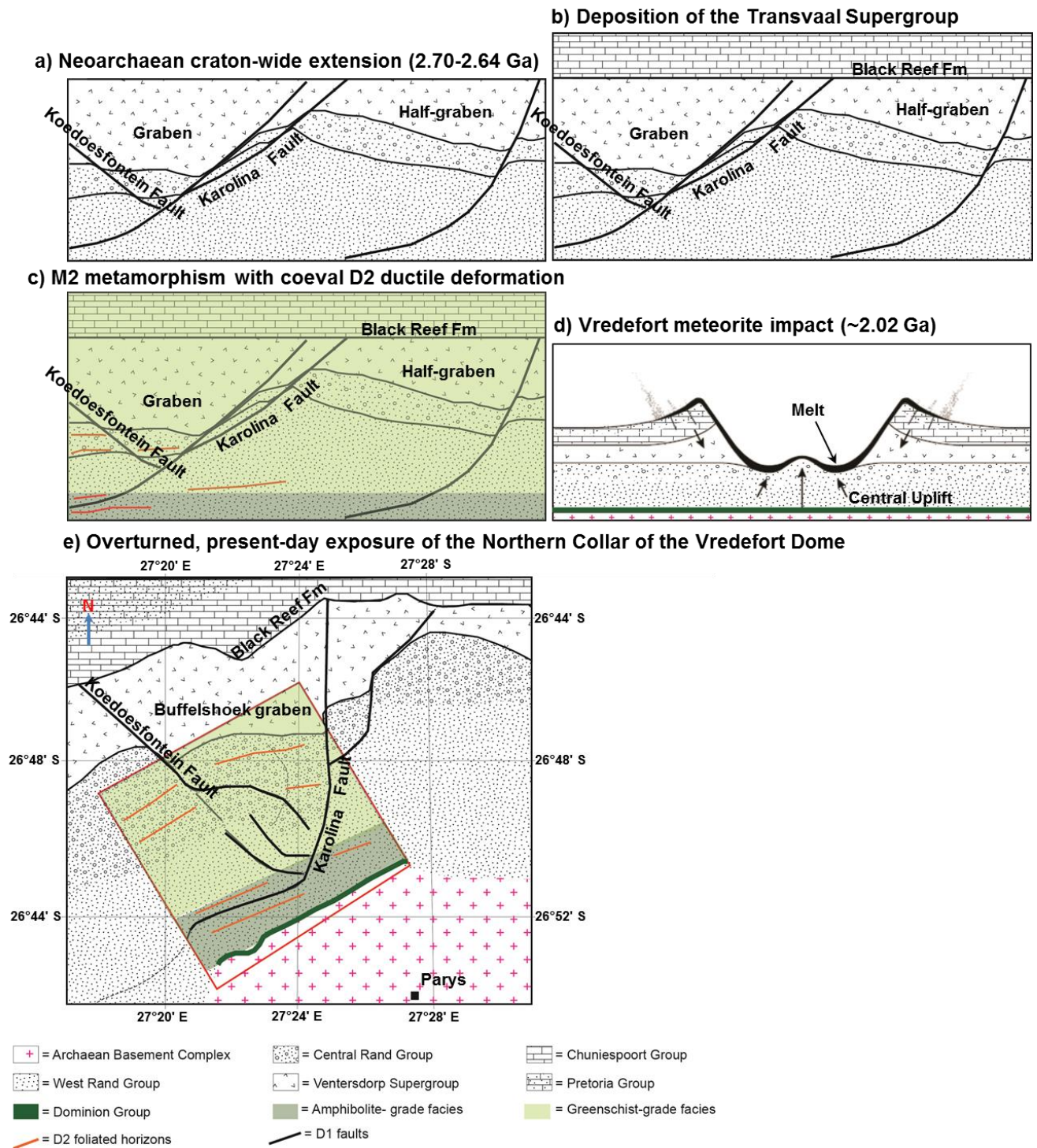


Figure 9.1. A schematic diagram of the Neoarchaean to Paleoproterozoic history of the northern collar of the Vredefort Dome. (a) Neoarchaean craton-wide extension; (b) deposition of the Transvaal Supergroup; (c) M2 metamorphism with coeval D2 ductile deformation; (d) Vredefort meteorite impact; and (e) present-day exposure.

9.2.3. Vredefort meteorite impact (2.02 Ga)

The already deformed strata (Archaean Basement Complex, Dominion Group, Witwatersrand, Ventersdorp, and Transvaal supergroups) were subjected to high shock pressures and significant uplift and overturning during the 2.02 Ga Vredefort impact. Shock pressures associated with the contact phase of the impact formed shatter cones, PDF, shock metamorphism ($M_{4a(NC)}$), shock/shatter cleavage and shock-induced pseudotachylitic breccia and impact melt-rock hosted in the collar. Following the contact and excavation phases, the target rocks were folded during early stages of the central uplift formation (Wieland et al., 2005; Reimold and Gibson, 2006).

The D4 strike-slip faults are attributed to tangential shortening during the inward and upward (constriction) displacement of the strata, before the central uplift collapsed outwards (Section 2.4.1). The faults formed together with kink bands, joints and spaced fracture cleavage. The subsequent outward collapse formed more joints and overturned earlier impact-induced features. The outward collapse also rotated the pre-existing Neoarchaean listric faults, now hosted as the D1 Karolina listric fault system in the collar of the dome. Also overturned was the S_2 foliation and associated metamorphic assemblages. Heat associated with the impact formed $M_{4b(NC)}$ metamorphic assemblages, which overprinted pseudotachylitic breccia and partly replaced the pre-existing $M_{2(NC)}$ metamorphic assemblages (Section 6.2.5).

Impact-induced deformation was culminated by the intrusion of the Granophyre Dyke into tangential en-echelon fracture system related to central uplift collapse. The granophyre cuts across all faults hosted in the northern collar of the Vredefort Dome (Section 7.2).

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