

Geology of selected sites along the northern margin of the Barberton
Greenstone Belt

Masters' Dissertation

Submitted by

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

Plagiarism Declaration

I, Nonkululeko Phumelele Mashele (458861), 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.



Nonkululeko Phumelele Mashele

28th day of September 2022 in Windischeschenbach

Abstract

An increasing number of UNESCO World Heritage Sites are recognised for their geological significance. One such site forms part of the Barberton Greenstone Belt (BGB), which has been widely acknowledged as an outstanding site for Archean geology research. Though members of the geoscience community are familiar with the outstanding universal value of the Barberton Greenstone Belt, the people who live on it often do not. With the World Heritage Status (2018) comes an obligation to unpack the geological heritage of the Barberton region to local people and visitors, and to promote further research so the true value of the geological importance of the region may be better understood. For this reason, this project set out to interpret and contribute to concepts formulated from studies of BGB rocks to produce a thesis usable for generating various science communication media and topics for further research. Themes discussed in this thesis include granite-greenstone relationships, alteration of ultramafic rocks and the associated economic products, as well as deformation of various lithologies. Ease of access guided the selection of eight mostly roadside outcrops along the R38 secondary road linking Barberton to the southern gate of the Kruger National Park. These sites link up to form a geological trail along the northern margin of the Barberton Greenstone Belt, allowing visitors to easily interact with outcrops on their way to other popular attractions.

By driving along the R38 with science communication media produced from this project, locals and visitors will learn how geology shaped their world. They will gain insight on how mineable concentrations of verdite, talc and magnesite occur in certain ultramafic bodies. They will also learn of different ways in which rocks respond to stress as a common perception of rocks among laypersons is that they are solid, stable and unchanging.

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†: Passed on

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0.1 Introduction

0.1.1 Barberton Granite-Greenstone Terrane

Archean landmasses remain our primary source of information about early Earth processes. The volcano-sedimentary sequences and adjacent plutonic rocks contained in Archean landmasses such as the Barberton Granite-Greenstone Terrane (BGGT) are not only the best-preserved in quantity, continuity of sequences and in quality (Byerly *et al.*, 2018 and references therein), they are also easily accessible. For this reason, the Barberton region is of outstanding universal value and worthy of world heritage status (World Heritage Committee, 2008). Through the years, researchers have studied various components of the BGGT, collectively producing plausible reconstructions of early Earth conditions and processes (Byerly *et al.*, 2018 and references therein).

Extensive work has been done covering such diverse topics as the evolution of the lithosphere and construction of continents (Anhaeusser, 1969, 1975; 1984; De Wit, 1992; Heubeck and Lowe, 1994; Van Kranendonk, 2010; Bedard, 2017), the origins of life (Barghoorn and Schopf, 1966; Walsh and Lowe, 1985; Ward, 1999; Schopf, 2006; Tice and Lowe, 2006; Kohler *et al.*, 2013) and its emergence from marine to terrestrial environments (Noffke *et al.*, 2006; Homann *et al.*, 2015; Eriksson, 1977), the evolution of the atmosphere (Pflug, 1967; Lowe and Knauth, 1978; Reimer, 1980; Hessler *et al.*, 2004; Lowe and Tice, 2004; Sleep and Hessler, 2006; Hessler and Lowe, 2006), the response of the early crust to regional stresses (Anhaeusser, 1969, 1972; Gay, 1969; Heubeck and Lowe, 1994b; Dirks *et al.*, 2013) and the implications for ore-forming processes (Anhaeusser, 1969, 1975, 2012; Dirks *et al.*, 2011; Gloyn-Jones and Kisters, 2018). Although many of these topics - or aspects thereof - remain debatable, the global significance of the BGGT is universally acknowledged. As a result, and for geoscientific reasons, a large part of the BGGT was recently (2018) declared a World Heritage Site.

This study addresses the scientific outreach objective of both CIMERA and the management authority of the Barberton Makhonjwa Mountains World Heritage Site (BMM WHS) by collating, updating and disseminating geological knowledge about selected outcrops within the BGGT. To broaden the appeal of the Barberton Makhonjwa Mountains World Heritage Site as a tourist destination, the focus is on eight geologically significant and instructive sites along or near the R38 road, skirting the northern margin of the Barberton Greenstone Belt (BGB). This road links Barberton to the southern gates of the Kruger National Park. The sites represent contact zones of the BGB supracrustal sequence and surrounding granitoids, mineralisation

associated with ultramafic intrusives, highly strained conglomerates, and low-grade metamorphism. Objectives vary between original scientific work (including high-resolution geological mapping, petrographic and bulk chemical analysis, elemental mapping using μ -XRF) and synthesising pre-existing scientific data.

By placing the selected sites in the context of modern BGB stratigraphy and tectonics as well as preparing learning objectives for laypersons, this study will produce both original geological literature as well as laying the base for educational products aimed at the general public, including adequately phrased pamphlets and guidebooks.

0.1.2. Barberton-Makhonjwa Mountains World Heritage Site

The town of Barberton was developed in 1884 during a gold rush that predates the Witwatersrand goldrush by two years (see guidebook in appendices). The discovery of world class gold deposits during the 1880s brought the world's attention to the town of Barberton. Once again, the town gains international attention as nearby mountains have become a UNESCO World Heritage Site (2018). This status applies to a site inscribed as the Barberton-Makhonjwa Mountains World Heritage Site (BMM WHS). It incorporates c. 40% of the Barberton Greenstone Belt (BGB) to the exclusion of areas associated with mining rights or those occurring in Eswatini.

With the aim to boost geotourism activity around the then tentatively inscribed BMM WHS, the Barberton-Makhonjwa Geotrail was developed (2014). Located along the R40 road between Barberton and the Bulembu in the western part of Eswatini, the R40 Geotrail intercepts stratigraphic sequences of the BGB. Along this road, visitors can view rocks and information panels that describe the early Earth, local flora and fauna, as well as mining and cultural heritage of the region. It is currently the only geotourism physical product within the Barberton region and is the only road through the BMM WHS that intercepts the widest variety of lithologies. The R40 Geotrail has potential to gain more visitors if more attractions are created in the vicinity thereof.

To explore that potential, the present study lays the foundation for extending the existing R40 Geotrail to the N4 national road connecting Pretoria to Mozambique. This extension will be done by developing a Geotrail on the tarred R38 road between the existing R40 Geotrail and the N4 road passing through Kaapmuiden. Along the R38 road are several geosites, most of

which are exposed as roadcut surfaces. These geosites display structures and minerals formed through multiple deformation events and metasomatism.

Though R38 geosites present interesting features, laypersons may need more than just the geoheritage content to enjoy a geotourism tour. During the experience of the author in guiding laypersons to BGB geosites or giving talks about Barberton geoheritage, it became apparent that geotourism is better received by laypersons when coupled with other interests. This approach has the potential to work well in the Barberton region because of the following reasons. The region has a long history of human occupation spanning from the stone age to the present day (Liebenberg, 2016). It also has an active heritage society that conducts excursions to many geosites. There are documented sites rich in biodiversity (Emery *et al.*, 2016) which are visited by birding clubs and ecology researchers who continue to make discoveries (Oosthuizen, 2021). The mining history extends for more than a century, and even longer when pre-colonial mines exploited during the iron-age are included (Delius, 2007; Liebenberg, 2016). Various natural or cultural heritage sites - published or those known only by locals - are often linked to rocks. For instance, a number of granitic bodies of the region preserve rock art (local knowledge; Liebenberg, 2016) or contain warm-water springs associated with mafic dykes (Visser *et al.*, 1956). Other geosites double as cultural sites. This is the case for the painted boulders in Low's Creek and ironstone exposures mined for red ochre since pre-colonial times (Myburgh, 1949; Liebenburg, 2016). For this reason, geology chapters of this thesis, which are numbered according to the sequence of the proposed R38 Geotrail sites from Barberton to Kaapmuiden, include cultural and/or historical attractions that will help tourism service providers attract groups of various interest to the R38 Geotrail and the associated BMM WHS.

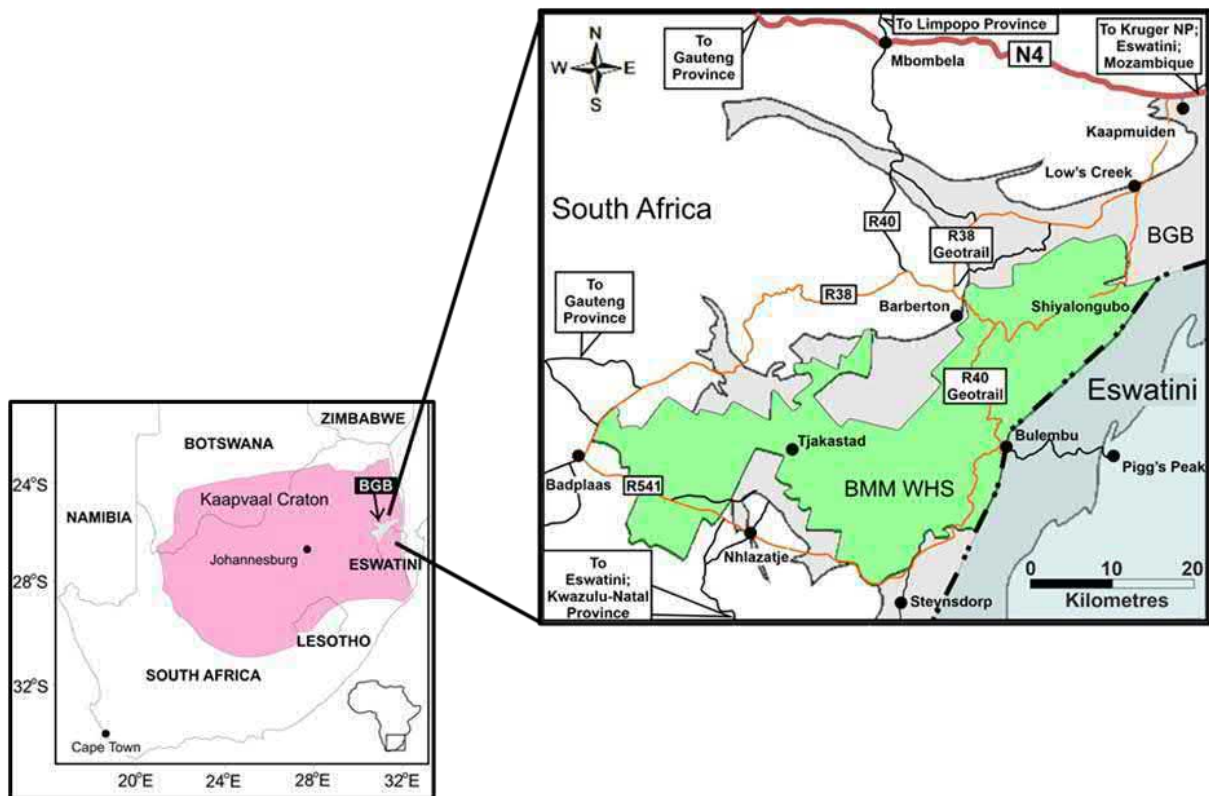


Figure 1. Geological context of the Barberton-Makhonjwa Mountains World Heritage Site (BMM WHS: green filled polygon). The site includes 40% of the Barberton Greenstone Belt (BGB: grey filled polygon) and is best explored along a road network called the ‘Genesis Route’ (orange). This route intercepts key lithologies and other attractions.

0.1.3. Proposed R38 Geotrail Sites

The R38 road winds along the northern edge of the Barberton Greenstone Belt, cutting across two granite bodies, various greenstones and good outcrops in two river systems. From about 7 km outside Barberton, the Kaap River can be seen winding in the valley adjacent to the Geotrail, providing wonderful scenery all year round. This river has been included in close-up maps of different sites. By spotting features of the river e.g., a sharp bend to the east or where the river flows closest to the road, visitors can ensure that they are at the right site in the absence of navigation tools.

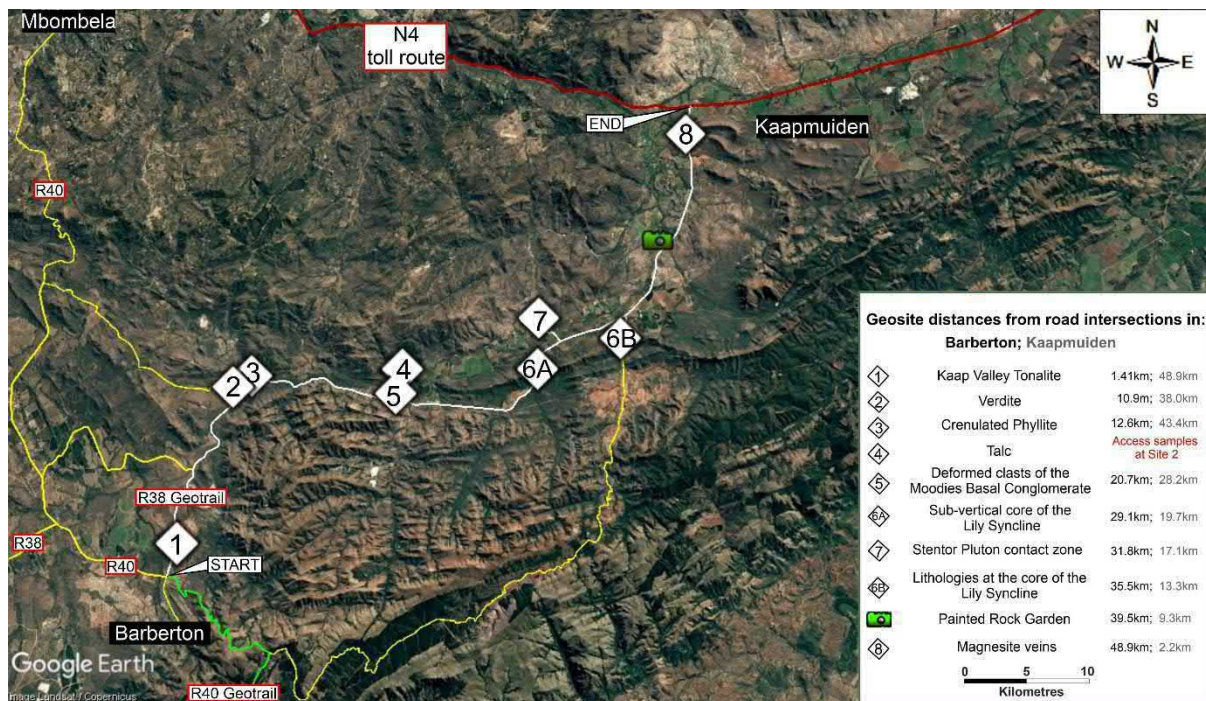


Figure 2. Selected roadside exposures along the R38 Geotrail.

As a proposed geotrail, this section of the R38 between Barberton and Kaapmuiden currently contains no information panels, bins, shops or toilet facilities. However, one of the proposed R38 Geotrail sites (Site 2) is located outside Diggers' Retreat, which is a hotel and restaurant facility where toilets can be used after purchasing refreshments.

0.1.4. Regional Geology

The Barberton Granite Greenstone Terrain (Figure 1) records more than 900 Ma (ca. 3644 to ca. 2740 Ma) of Earth history from the Paleo- to Neoproterozoic (Kröner and Compston, 1988; Tegtmeier and Kröner, 1987; Kamo and Davis, 1994; Robb *et al.*, 2006; Byerly *et al.*, 2018). Preserved components of the BGGT indicate at least three cycles of crustal evolution from ultramafic to felsic magmatism (Kohler and Anhaeusser, 2002; Robb *et al.*, 2006; Kröner *et al.*, 2013; 2018). Most rocks that constitute the BGGT are magmatic and occur either as volcanic components of the Barberton Greenstone Belt (BGB) supracrustal sequence or as intrusive layered ultramafic sequences or plutons of various compositions (Viljoen and Viljoen, 1969; Robb *et al.*, 2006; Kohler, 2003). Ovoid and circular granitoid plutons of the region surround the BGB (Anhaeusser, 1976; Anhaeusser and Robb, 1981; Robb *et al.*, 1986).

The supracrustal rock sequence has been divided into three lithostratigraphic units. These units are, from oldest to youngest, the Onverwacht, Fig Tree and the Moodies Groups. They have

corresponding thicknesses of >10 000 m, ca. 1 800 m and ca. 3 000 m, respectively (Byerly *et al.*, 2018).

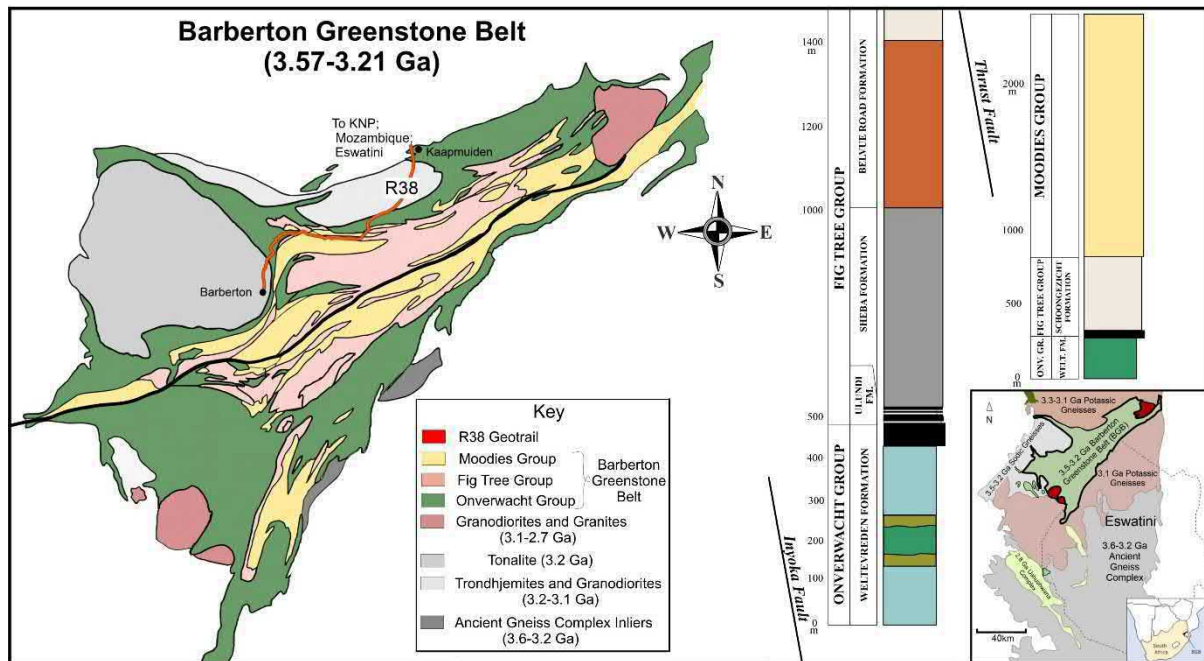


Figure 3. Simplified geological map of the R38 Geotrail in the context of the Barberton Greenstone Belt. Modified after Visser *et al.*, (1956) and Anhaeusser *et al.* (1981). Stratigraphic column modified after Byerly *et al.*, 2018.

0.1.4.1. Stratigraphic units

The Barberton greenstone belt (BGB) has been categorised into three tectonostratigraphic blocks which, from south to north, are the Songimvelo, Umuduha and Kaap Valley blocks (Byerly *et al.*, 2018). Though consisting of similar lithostratigraphic units, these terrains have distinct stratigraphy and tectonic histories, generally becoming younger towards the north (Lowe and Byerly, 2007). This study area lies on the Kaap Valley block of the BGB.

The 3550-3270 Ma Onverwacht Group of the Kaap Valley Block consists of ultramafic to mafic volcanics and their serpentinised equivalents (Anhaeusser, 1976; Lowe and Byerly, 2007; Byerly *et al.*, 2018). Minor occurrences of intermediate to felsic volcanics also occur, with the former often found to have ages overlapping with those of nearby intermediate granitoids (Agangi *et al.*, 2018). Cherts and silicified clastic units form a volumetrically minor but conspicuous and stratigraphically important component (Lowe and Byerly, 2007; Hofmann *et al.*, 2013).

Conformably overlying the Onverwacht Group is the 3260-3225 Ma Fig Tree Group, mostly composed of argillite, greywacke, various chemical sediments and intermediate volcanics. It

contains several meteorite-impact spherule beds (Lowe *et al.*, 2014; Jones and Kisters, 2018). The Bien Venue Formation that forms part of the upper Fig Tree Group defined in the northeastern Barberton Greenstone Belt is dominated by quartz-muscovite schists (Kohler and Anhaeusser, 2003). Stratigraphically, the Bien Venue Formation is overlain by felsic schists and agglomerates of the Schoongezicht Formation that are in a transitional contact with the Moodies Group (Byerly *et al.*, 2018). The 3224-3214 Ma Moodies Group is comprised of terrigenous sediments interlayered with iron-rich shales, jaspilites, rhyodacitic ash-fall tuffs and a regional basaltic unit (Visser *et al.*, 1956; Anhaeusser, 1976; Eriksson, 1978; Heubeck and Lowe, 1994).

0.1.4.2. Deformation

The present fabric of the Barberton Greenstone Belt (BGB) is a result of multistage deformation during the deposition of BGB supracrustal deposits (Visser *et al.*, 1956; Anhaeusser, 1966; 2019; De Ronde and De Wit., 1994; Heubeck and Lowe, 1994; Byerly, 1999; Kisters *et al.*, 2003; Van Kranendonk, 2011; Byerly *et al.*, 2018; Dziggel and Kisters, 2019). In the northern BGB, the earliest deformation event (D₁) produced a series of cleavage-less, doubly-plunging folds of a northeast-southwest trend (Ramsay, 1963; Visser *et al.*, 1956) understood to have resulted from gravity-induced sagging of BGB supracrustals (Anhaeusser, 1966). The folds verge to the north and northwest, with the southern limb overturned and invariably thicker than the northern limb due to tectonic truncation by high-angle faults (Anhaeusser, 1976; Byerly *et al.*, 2018). The high-angle faults coincide with poorly preserved anticlinal structures between regional synclines (Anhaeusser, 1976). The former were considered to have initially been low-angle thrust faults separating synclines in a fold-and-thrust belt during D₁ shortening (De Wit and De Ronde, 1994). These faults were reactivated as strike-slip faults during events subsequent to D₁ (Anhaeusser, 1976). The second deformation produced slaty cleavage of a northeast-southwest trend at an oblique angle to bedding and fold axes of D₁ (Ramsay, 1963). However, the similar orientation of D₁ and D₂ principal strain axes may indicate a single event of progressive deformation (Ramsay, 1963). The third event in Ramsay (1963) is considered to be northeast-southwest shortening which Anhaeusser (1963, 1966, 1972, 1975) associated with the emplacement of the Kaap Valley Tonalite (Anhaeusser, 1966; Gay, 1969). This third event of shortening was coincident with the refolding of the Eureka and Ulundi Synclines and their detachment folding about a northwest-southeast axis (Anhaeusser, 1968; 1972; 1975; 2019). A final deformation event is indicated by sub-vertical shortening of D₂ slaty cleavage to produce sub-horizontal crenulation

lineations, kink bands as well as conjugate and chevron folds (Anhaeusser, 1963, 1968, 1975; Viljoen, 1963). However, after noting the consistent juxtaposition of Moodies and Fig Tree sequences over Fig Tree and Onverwacht sequences in the northern BGB, Byerly *et al.* (2018) proposed a northwards thrusting event as the final deformation event.

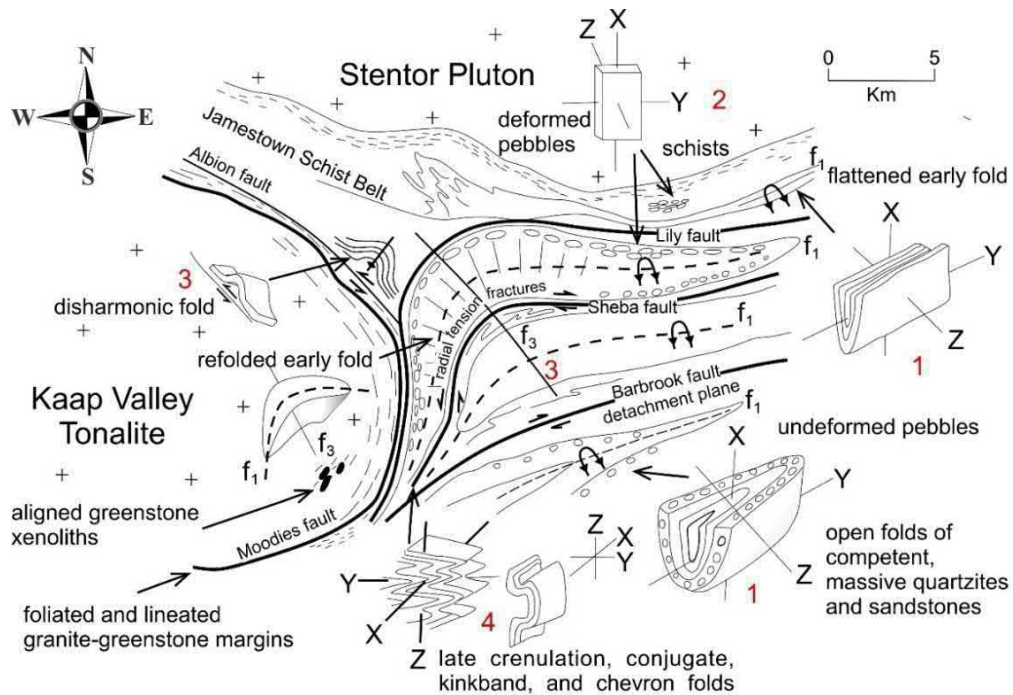


Figure 4. Chronological order (red numbers) of characteristic structures of main deformation events that affected the northern Barberton Greenstone Belt. Redrawn from Anhaeusser (1975).

0.1.5 Present Study

0.1.5.1 Aims

To identify roadside outcrops in the northern Barberton Greenstone Belt about which existing and new geological information with an appeal to laypersons would be generated and further research would be inspired. To produce written and visual content for science outreach, geoheritage awareness and geotourism development in the Barberton region.

0.1.5.2 Objectives

To undertake a literature review, collate data and to generate new information for sites with unresolved key features. Detailed accounts of objectives are listed in individual chapters.

0.1.5.3. Methods: Science outreach and thesis components

To fulfil science objectives of the funding agency, the geoheritage of the Barberton region was translated to laypersons through T.V., radio, a media article and a talk at a local university where c. 300 tourism and heritage students were in attendance (Appendix A).

General to all sites is the synthesis of existing information, the digitisation of map sections relevant to each site and redrawing of figures. New figures were drawn where new information was produced.

Due to travel restrictions and limited lab access resulting from the COVID-19 pandemic, original lab-intensive studies meant for some sites were substituted by comparative field studies of other sites. This was to confirm apparent characteristic features and lithological associations observed in R38 outcrops.

Unless otherwise stated in individual chapters, thin sections of characteristic samples of each site were produced from oriented samples photographed within the sampled outcrop. Sample composition and fabrics were described using a petrographic microscope. Photomicrographs were created in the Wits Imaging Lab using the Olympus BX41 and BX53M photographic microscopes connected to the Olympus DP74 imaging system (Fig. 5 A).

All bulk chemical analyses were carried out on samples trimmed of weathered surfaces and subsequently processed using a jaw crusher and disc mill at the Wits Geosciences Earthlab. Major element analysis of samples was conducted using the Panalytical Axios x-ray spectrometer to measure x-ray fluorescence. Elements were determined using the Norrish Fusion technique (Norrish and Hutton, 1969) together with in-house correction procedures.

Major elements were fused using Johnson Matthey Spectroflux 105 at 1000 °C and raw data corrected using in-house software. Standard calibrations were made using synthetic oxide mixtures and international standard rocks as well as in-house controls. Sample weight used was 0.35 gm and flux weight 2.0 gm. Calibration standards were primary International Reference Materials USGS series (USA) and NIM series (South Africa). Precision was taken as 1% for elements in abundance of greater than 5% by weight, and 5% for elements in abundance less than 5%. Standard calibrations used BCR2, NIM-P, NIM-D, W-2, NIM-S, GSP-2, NIM-N, NIM-G, PCC1, BHVVO2, AGV2, G2 DTS1. Bulk chemical analyses of trace elements of samples were conducted using the Phillips PW2404 x-ray spectrometer. Samples were pressed into pellets using a Moviol solution binder. Standardisation was carried out using International Reference Materials USGS series (USA) and NIM series (South Africa). Precision was determined on the basis of counting time and was taken as 5% for elements in abundance greater than 100ppm, and 10% for elements in abundance between 10 and 100 ppm.

Elemental maps were generated using a Bruker Tornado M4 micro-X-ray fluorescence analyser at the Friedrich-Schiller University of Jena, Germany. Selected parts of thin sections were carbon coated using graphite and irradiated with an electron beam. The ratio of counts per second to applied electrical potential from the electron beam was plotted against the applied electrical potential to generate characteristic peaks of phases within the analysed sample. Peaks on a spectrum of reference materials were matched with peaks on a spectrum of the analysed sample to identify phases.

Confirmation of phases and the characterisation of advancedly altered components was achieved using the liberation analysis option of the VEGA3 X64 TESCAN TIMA 1.7.0 (Fig. 5 B). scanning electron microscope. In particular, the energy-dispersive X-ray spectroscopy component was used together with the VEGA3 X64 TESCAN TIMA 1.7.0. software within the Wits Automated Mineralogy Lab of the School of Geosciences at the University of the Witwatersrand, Johannesburg. Beam conditions during liberation analyses on the VEGA3 X64 TESCAN SEM were 25 kV at accelerating voltage and an emission current of 43 µA. Working distance of 15 mm, a specimen beam current of 17.68 nA, spot size of 590.00 nm and an absorption current of < 1 pA were selected. The scanning mode was set at resolution mode, the heating at 47 %, the filament live time was at 431 hours and the gun pressure was at 1.6×10^{-3} Pa while the column pressure was at 5.5×10^{-3} Pa. Acquisition of high-resolution maps was conducted at a pixel spacing of 2 microns, x-ray counts of 1000 and an imaging speed of 4 microseconds.

Raman spectra were produced from four spots on a freshly cut, unpolished surface of outcrop sample MJS were examined on a HORIBA LabRam Raman instrument (Fig. 6) at the Department of Geosciences, Friedrich-Schiller University of Jena, Germany. The microscope magnification was 20x and a laser wavelength of 532 nm was used for 9 seconds and two runs and at 5 % laser intensity. Acquired spectra were found to lie on curves which were flattened by simple curve fitting. Spectra also showed (poorly defined) periodicities of ca. 90 cm^{-1} between 300 and 1300 cm^{-1} . These deficiencies may be due to weak electric current changes, effects of room lighting or fluctuations in laser intensity. Corrected spectra were compared to those proposed by the HORIBA KnowItAll Spectral library and the public RRUFF Raman spectra database.

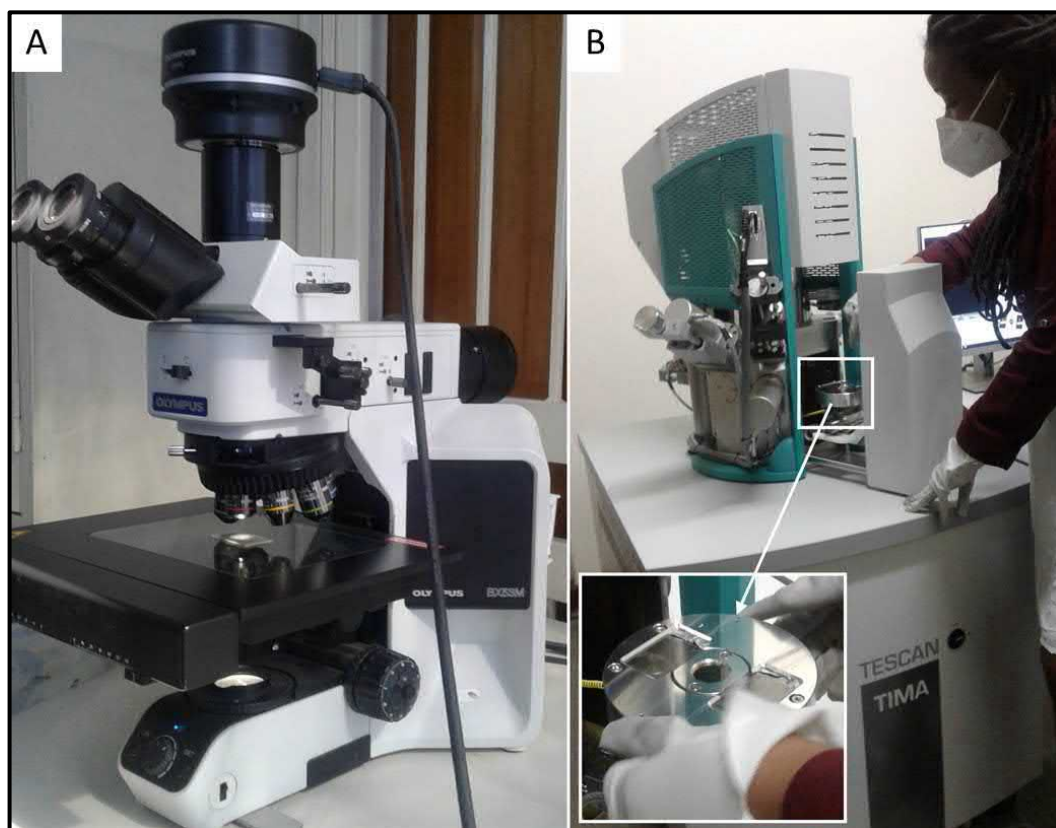


Figure 5. Instruments used for analyses. (A): Olympus BX53M petrographic microscope and DP740 digital camera (top component) used to generate photomicrographs. (B): TESCAN integrated mineral analysis (TIMA) instrument. Insert shows carbon-coated thin sections mounted on a stage.



Figure 6. Analytical and data-processing instruments. (A) Sample MJS mounted on a stage prior to analysis. (B): Hardware coupled with spectrometer for data acquisition, processing, and export.

Chapter 1: Kaap Valley Tonalite (Crumbly Granite) in a Quarry

Way in at 25°45'0.28"S; 31° 3'12.07"E

1.1 Site description

Weathered outcrops of the Kaap Valley Tonalite are exposed in a municipal sandpit where tonalite fragments and sand are extracted for various uses in construction and road maintenance (Fig. 1.1). Where intact, outcrops of the tonalite occur as medium- to coarse-textured domes within masses of friable weathered material. Northwest-trending veinlets of epidote are common within the quarry. These are of a similar trend to dolerite dykes that invade the tonalite in the area between Barberton and Mbombela (Fig. 1. 3). The various mafic enclaves in the tonalite will be described in later sections of this chapter. As a public location, this site may be accessed by other people at any given time. Care should be taken to discourage harm or theft by accessing the site in groups and locking or hiding away valuable items in vehicles.

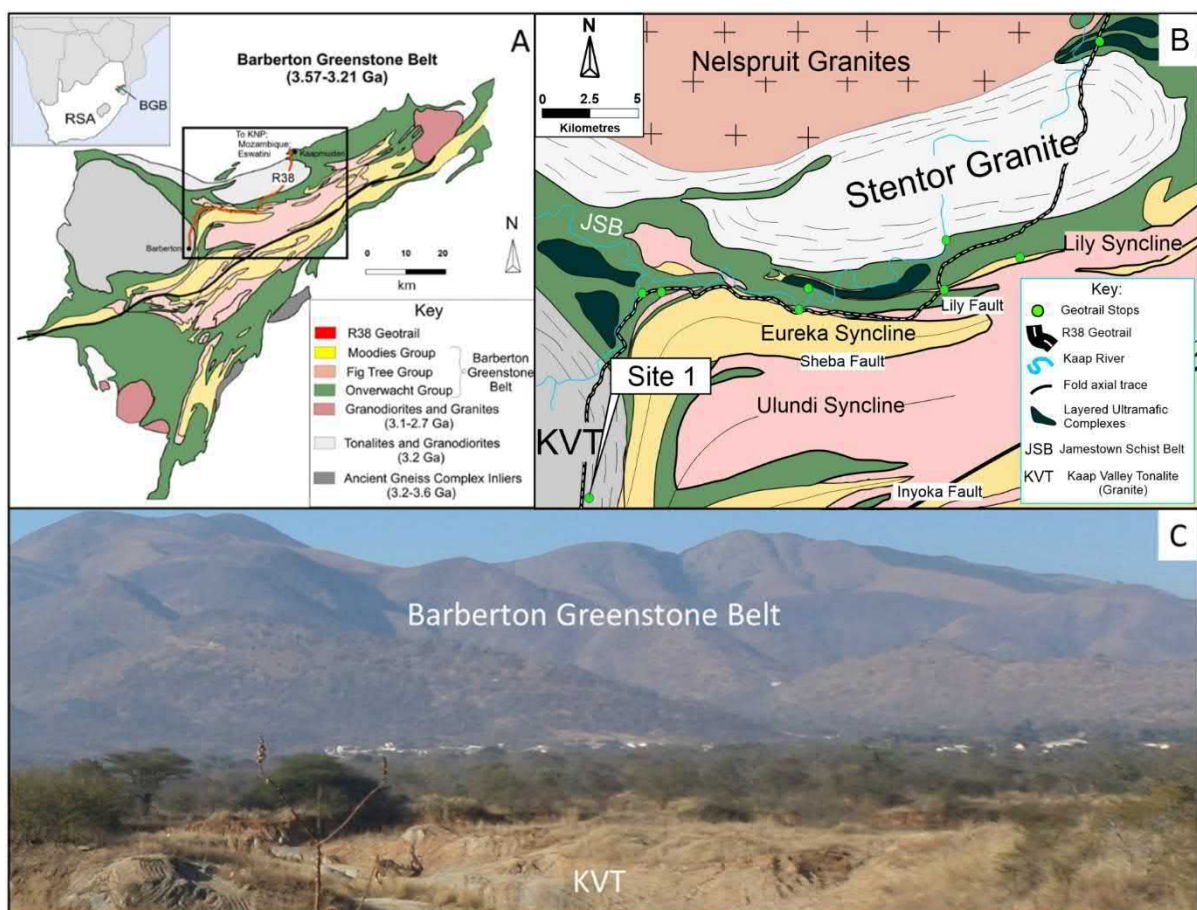


Figure 1.1 Main features of Site 1. (A): Simplified geological map of the R38 Geotrail showing the context of Site 1 in the Barberton Greenstone Belt (BGB). (B): A detailed view of the black rectangle in (A) shows the location of Site 1 in the Kaap Valley Tonalite at the western margin of the BGB. (C) View of Site 1 looking southeast. Maps modified after Visser *et al.* (1956) and Anhaeusser *et al.* (1981).

1.2 Aspects of the site that may appeal to laypersons during site visits

Solid outcrops of the Kaap Valley Tonalite occur in and around Barberton. At the sandpit (Fig. 1.1), tectonic fabrics and fractures in the marginal zones of the tonalite can be observed. These contribute to an understanding of the emplacement history of the pluton. They may be highlighted for their significance concerning deductions made about the emplacement history of the pluton. Visitors will also see well-preserved outcrops of the tonalite found within cores of friable plutonic rock on the quarry floor. Such exposures provide examples of weathering processes that form part of the rock cycle.

Fundamentally, concepts of weathering and variable resistance to erosion are easily seen here because the drastic change of slope at margins of the valley is very apparent. The weathering of the quartz-poor tonalite into a near-circular valley also contrasts with the resistance of the high hills formed by quartz-rich phases of the Nelspruit Batholith at the end of the R38 Geotrail.

On a more advanced level, coarse crystals making up the tonalite in the sand pit will be a discussion point on the control of cooling rate on crystal size and shape.

Various mafic enclaves and dykes (Fig. 1.2) can be used to discuss the usefulness of inclusions and cross-cutting relations in establishing relative ages of rocks.

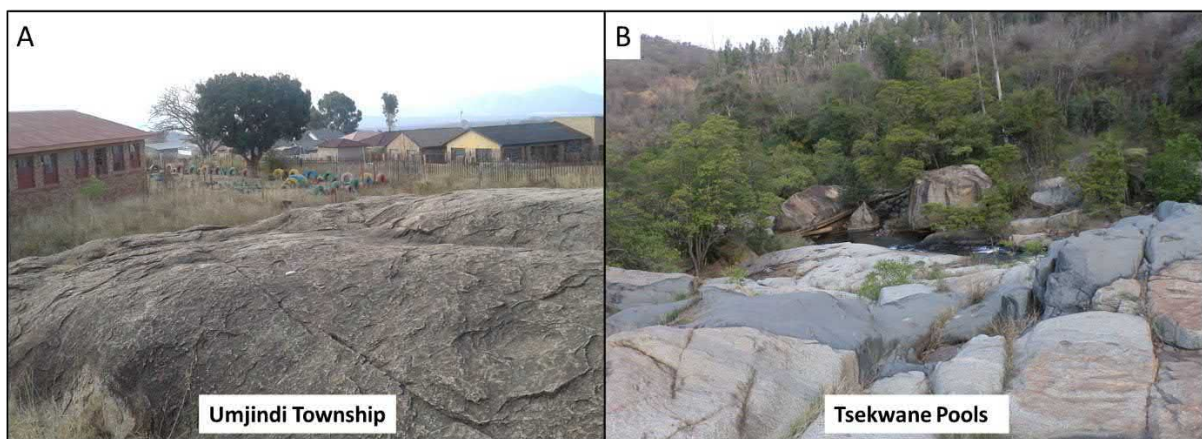


Figure 1.2 Kaap Valley Tonalite exposures at selected public and recreational localities in and around Barberton. (A): Extensive tonalite outcrops are a common occurrence in school yards and playgrounds. (B): A picnic location c. 16km WSW of Barberton (Tsekwane/Tiguan Pools) allows relative dating from the use of cross-cutting relationships between a basaltic dyke that crosscuts mineral layering and hosts angular tonalite fragments.

1.3 Geology

1.3.1 Background

1.3.1.1 Kaap Valley Tonalite

Archean cratons contain granitoids of varying compositions (Moyen *et al.*, 2010; Halla, 2018). These granitoids broadly fall into two main categories, the Trondhjemite-Tonalite-Granodiorite (TTG) series and the Granodiorite-Monzogranite-Syenogranite (GMS) series. During the Archean, granitoid magmatism often occurred episodically, and often produced composite TTGs (Halla, 2018). Further studies of TTG geochemistry led to the identification of two subcategories of TTGs, those with low Y and HREE, and those with high Y and HREE. These subgroups respectively correspond to deep and shallow crustal melting of oceanic crust with and without garnet in the residual phase (Halla, 2018).

In the Barberton-Granite-Greenstone-Terrane (BGGT), TTGs constitute the oldest remaining granitic bodies of the region and were formed during three main magmatic episodes (Condie and Hunter, 1975; Anhaeusser and Robb, 1981; Kamo and Davis, 1994; Agangi *et al.*, 2018; Kröner *et al.*, 2018). The youngest of these magmatic episodes produced TTGs in the western and northern margins of the BGB during a period of rapid crustal thickening between 3250-3225 Ma (Kamo and Davis, 1994; Byerly *et al.*, 2018). It is during this period that the Kaap Valley Tonalite was emplaced.

The tonalite is a medium- to coarse-grained (0.5-6mm) plagioclase-rich granite (albite-to-andesine) mainly with hornblende and lesser biotite as mafic phases (Robb *et al.*, 1986). It is more mafic than granitoids of the region (Anhaeusser, 1972; Robb and Anhaeusser, 1980; Robb *et al.*, 1986). Fine- to medium-grained tonalitic dykes with no apparent chill margin occur in central parts of the pluton (Robb *et al.*, 1986), while feldspar porphyries, quartz and carbonate veins commonly occur at and near pluton margins (De Ronde *et al.*, 1988; Faure, 1989). The higher hornblende content in the tonalite compared to other granites, was initially ascribed to magma contamination by hornblende xenocrysts from ultramafic country rocks (Condie and Hunter, 1975). However, complete assimilation of mafic-ultramafic country rocks, or mixing of rhyolitic and mafic-ultramafic magma were considered by Robb *et al.* (1986) to be appropriate based on petrographic and geochemical data.

The most precise age determined for the tonalite is a U-Pb age of 3227 ± 1 (Kamo *et al.*, 1990). Subsequently, Ar-Ar ages between 3250 and 3035 Ma were determined from hornblende and

biotite crystals from central and marginal zones of the tonalite (Layer *et al.*, 1992). The upper limit of the 3250-3035 Ma age spectrum corresponds to crystallisation of the tonalite after semi-solid diapiric emplacement to shallower crustal levels (Robb *et al.*, 1986). Crystallisation was soon followed by a pre-3214 Ma magnetic reversal proposed by Layer *et al.* (1996) to possibly be the oldest preserved magnetic reversal on Earth. The younger ages of the 3250-3035 Ma spectrum represent hydrothermal events presumably linked to the c.3.1 Ga intrusion of the Nelspruit Batholith and associated gold mineralisation of the northern part of the Greenstone Belt (Layer *et al.*, 1992).

Once solid, the tonalite was affected by northwest-trending shear zones (Robb *et al.*, 1986; Faure, 1989). These shear zones became sites for the emplacement of sub-economic gold-bearing quartz-chlorite-limonite veins with gold grades ranging from 0.03 to 0.9 ppm (Faure, 1989). Two later generations of pre-Transvaal basaltic and gabbroic dykes with U-Pb baddeleyite ages of c.2965-2967 Ma and c. 2662-2686 Ma intrude the Kaap Valley Tonalite (Olsson *et al.*, 2010). The older of these is the northwest trending Badplaas Dyke Swarm which occurs in a c.80 by 100 km² area between Eswatini and the Nelspruit region. It is correlated with lavas of the Nsuzi Group (Olsson *et al.*, 2010). Some warm-water springs of the Badplaas region occur where dykes of this swarm intercept sheared Archean plutons (Visser *et al.*, 1956).

The younger swarm is known as the Rykoppie Dyke Swarm (Olsson *et al.*, 2010). It extends over a c.50 by 100 km² area within the Archean basement part of the Barberton region and is correlated with the upper Ventersdorp Supergroup. The Rykoppie Dyke Swarm formed in a rift basin that created accommodation space for the deposition of Transvaal Group sediments (Olsson *et al.*, 2010).

1.3.1.2 Enclaves

According to previous workers cited in Barbarin and Didier (1991) and Barnes *et al.* (2021), mafic magmatic enclaves may be genetically related or unrelated to the enclosing granitic host. Enclaves may form through:

- chemical alteration of xenoliths (Bower, 1922);
- the disaggregation of synplutonic mafic dykes (Roddick and Armstrong, 1959);
- partial melting of restite material to produce the enclosing magma (White and Chappell, 1977);

- magma mixing where granitic magma is recharged by incoming mafic magma transporting undercooled, (semi) consolidated fragments to be later enclosed in the granitic magma (Eichelberger, 1978; Gamble, 1979; Barnes *et al.*, 1986) or
- the incorporation (into granitic magma) of fragments of a more mafic, earlier phase crystallised from an evolving magma chamber (Pabst, 1928; Bateman *et al.*, 1963).

It has been widely accepted that TTG magma is generated through 15-30 % partial melting of hydrous ultramafic-mafic rocks (Holloway and Burnham, 1972; Moyen *et al.*, 2010). This, according to Beard *et al.* (2004) occurs through incongruent melting of anhydrous mafic minerals in an assemblage of, for example, orthopyroxene-clinopyroxene-calcic-plagioclase-FeTi oxides to produce trondhjemites or tonalites comprised of hornblende-biotite-sodic plagioclase-quartz. In this way, preserved pyroxenitic restites may be associated with the resulting TTGs bodies.

Less enigmatic associations of mafic fragments and TTGs are seen in angular mm- to km-scale enclaves which are understood to be rafts stopped into TTG magma (Anhaeusser and Robb, 1981; Robb *et al.*, 1986). These enclaves are abundant near TTG margins due to the emplacement of most Barberton regional TTGs directly against ultramafic and mafic sequences of the Onverwacht Group (Anhaeusser, 1969; 1976; Robb and Anhaeusser, 1981). The tonalite margins may therefore contain enclaves of both obvious and contentious origin. According to Robb *et al.* (1986), the enclave population of the tonalite is made up of large and small fragments of serpentinites and amphibolites. The larger fragments were regarded as xenoliths and smaller fragments which were partially digested and found to have reacted with the enclosing tonalite were regarded as comagmatic phases. Comparative geochemical analyses by Faure (1989) found that hornblende from enclaves in the tonalite and that from the host tonalite overlapped in chemical composition.

The challenges in distinguishing the origins of enclaves were highlighted in Barbarin and Didier (1991). These authors described how the isotopic equilibration of enclaves with the enveloping melt may lead to a loss of the original isotopic fingerprint of enclaves. However, studies that include detailed comparisons of texture and chemical evolution of zoned plagioclase crystals from enclaves and the enclosing host allow for better assessments of enclave origins (e.g., Barnes *et al.*, 2021).

Distinct phases of the tonalite (Fig. 1.3) were identified in different spatial and topographic zones of the pluton (Robb *et al.*, 1986). The dominant of these phases contains hornblende and

lesser biotite as primary mafic minerals. Less common was a phase that contained hornblende as the only mafic phase in low topographic zones of the tonalite whereas the latter was in high topographic zones (Robb *et al.*, 1986).

In contrast, field observations and petrographic analyses by Faure (1989) contended that the tonalite was originally homogeneous. The heterogeneous occurrence of biotite in the tonalite was attributed to complete chloritisation of biotite within the original biotite-hornblende assemblage due to higher susceptibility of biotite to chloritisation (Faure, 1989). Studying 900 points on 78 tonalite samples, Faure (1989) provided a modal average of 58.1% primary plagioclase, 22% quartz, 5.8-20% hornblende and 1-10% biotite with 10.1% secondary chlorite, 2.0% epidote and 0.9-40% of secondary calcite in the samples studied.

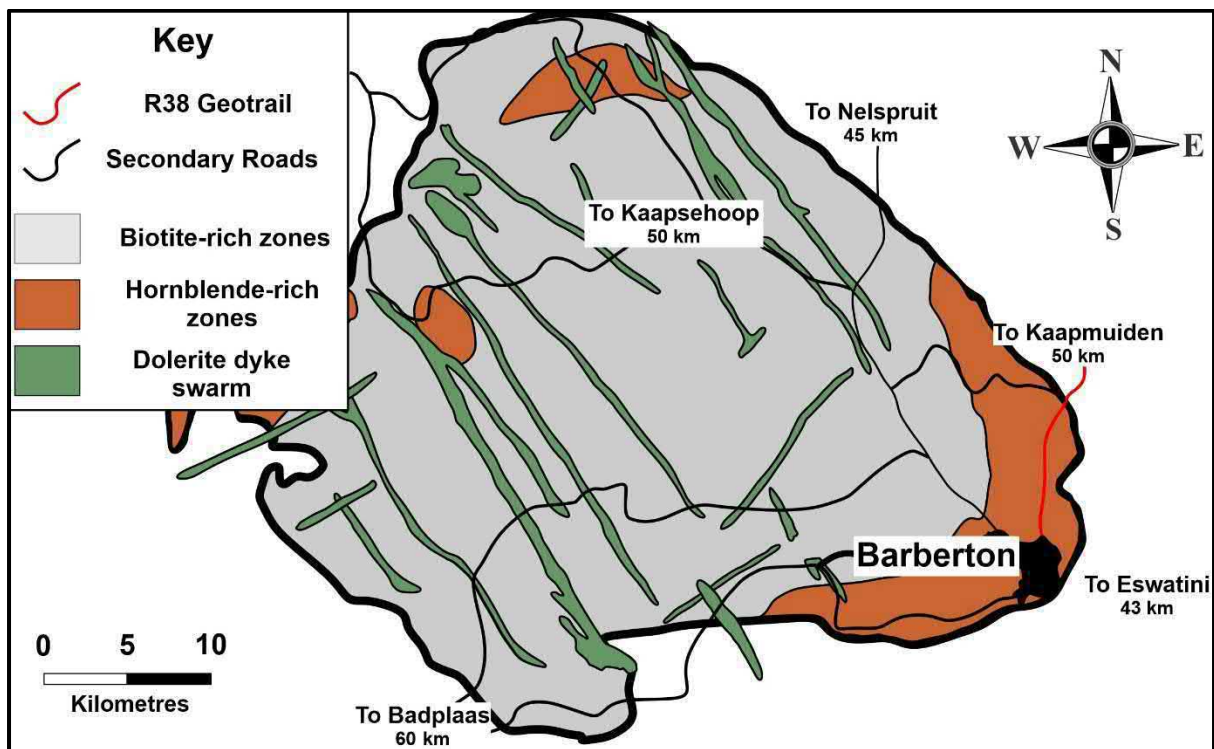


Figure 1.3 Distribution map of different tonalite phases of the Kaap Valley Tonalite. Also shown are NW-trending and roughly E-W-trending dyke swarms. Map redrawn from Robb *et al.* (1986).

1.3.1.3 Margins

Except for the western margin which is overlain by Proterozoic Transvaal Group metasedimentary rocks, the margins of the Kaap Valley Tonalite are adjacent to sheared metavolcanics of the Onverwacht Group (Anhaeusser, 1966, 1969). These sheared metavolcanics are locally associated with gold deposits that have been mined for more than a

century (Anhaeusser, 1966, 1969, 1976, 2013). Unlike the 3.1 Ga potassic granites of the Barberton region, the tonalite margins are not discordant intrusive bodies (Anhaeusser, 1966, 1969, 1972). Instead, interleaving contact zones comprised of concordant, alternating layers of sheared tonalite and metavolcanic country rocks are observed within an amphibolite-to-greenschist metamorphic aureole of less than 5 km (Anhaeusser, 1966, 1969; Robb *et al.*, 1986; Faure, 1989; Dziggel *et al.*, 2006). According to Anhaeusser (1966, 1969) and Faure (1989), lithologies on either side of the tonalite margin present a strong contact-parallel mineral foliation and lineation that dips or plunges southwest towards the core of the pluton at certain localities (Fig. 1.4). These fabrics decrease in intensity away from the contact zone, shallowing towards the core of the pluton suggesting that the margins of the tonalite were semi-solid during emplacement into the metavolcanics of the Onverwacht Group resulting in metasomatised schists (Anhaeusser, 1966, 1969). In this way, margins of the tonalite were sheared and recrystallised as they steepened and caused shearing of metavolcanic country rocks (Anhaeusser, 1969; Robb *et al.*, 1986) and - according to Faure (1989) - produced a cataclasite zone in the southeastern contact of the tonalite. Both the sub-vertical and sub-horizontal strain within the Barberton Greenstone Belt was attributed to the shallow diapiric emplacement of the tonalite (Viljoen, 1963; Anhaeusser, 1968, 2019).

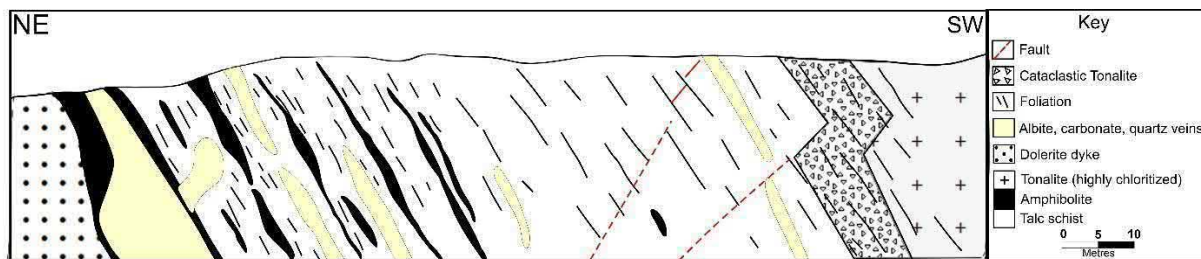


Figure 1.4 Schematic diagram of the interleaved contact between the Kaap Valley Tonalite and metasomatised volcanics of the Onverwacht Group. This outcrop can be accessed at Site 1 of the R40 Barberton-Makhonjwa Geotrail. Note the southwest-dipping foliation in the country rocks that also occurs in the tonalite. Modified after Faure (1989).

Within the contact aureole of the tonalite are pluton-parallel extensional shear zones that were responsible for the downward displacement of Onverwacht metavolcanics (Anhaeusser, 1976). Beyond the contact aureole, the effects of the emplacement of the tonalite caused sub-horizontal northeast-southwest shortening and detachment folding of the inner Eureka and Ulundi folds about a northwest-southeast fold axis. Progressive shortening led to northwestwards thrusting of the refolded Eureka and Ulundi folds during D₃ northeast-southwest shortening (Anhaeusser, 1976, 2019). This deformation was bound by the Barbrook fault in the southeast and the Lily Fault in the northwest (Anhaeusser, 1976, 2019).

1.3.2 Present study

1.3.2.1 Aim and Objectives

To aim of this research was to make field descriptions of enclaves and their contacts with the tonalite and to describe textural and compositional characteristics of sampled enclaves and their margins.

1.3.2.2 Methods

Where possible, characteristic enclaves were viewed in three dimensions and were described in the field in terms of composition and texture, size, nature of contacts and orientation with respect to tonalite foliation.

Representative samples of enclave variants within the host tonalite were cut into thin sections for petrographic study. The focus of descriptions was composition, identity, and texture of minerals on either side of the enclave-tonalite contacts. To verify enclave descriptions made at Site 1 of the R38 Geotrail, enclaves from three other sites that are easily accessible were also described. These include a river pavement from the Caledonia Siding, outcrops on a school playground in the Barberton Township and a river pavement at Tsekwane/Tiguan Pools that are respectively c.10 km north, c. 2 km NNW and c. 16 km WSW of the Barberton Central Business District. All sites studied occur within a hornblende-tonalite phase of the Kaap Valley Tonalite identified by Robb *et al.* (1986) in the southeastern margin of the pluton (Fig. 1.3).

Elemental maps of representative samples of the tonalite and of a sub-angular mafic enclave with sharp margins were produced from complementary offcuts of blocks from which thin sections were cut.

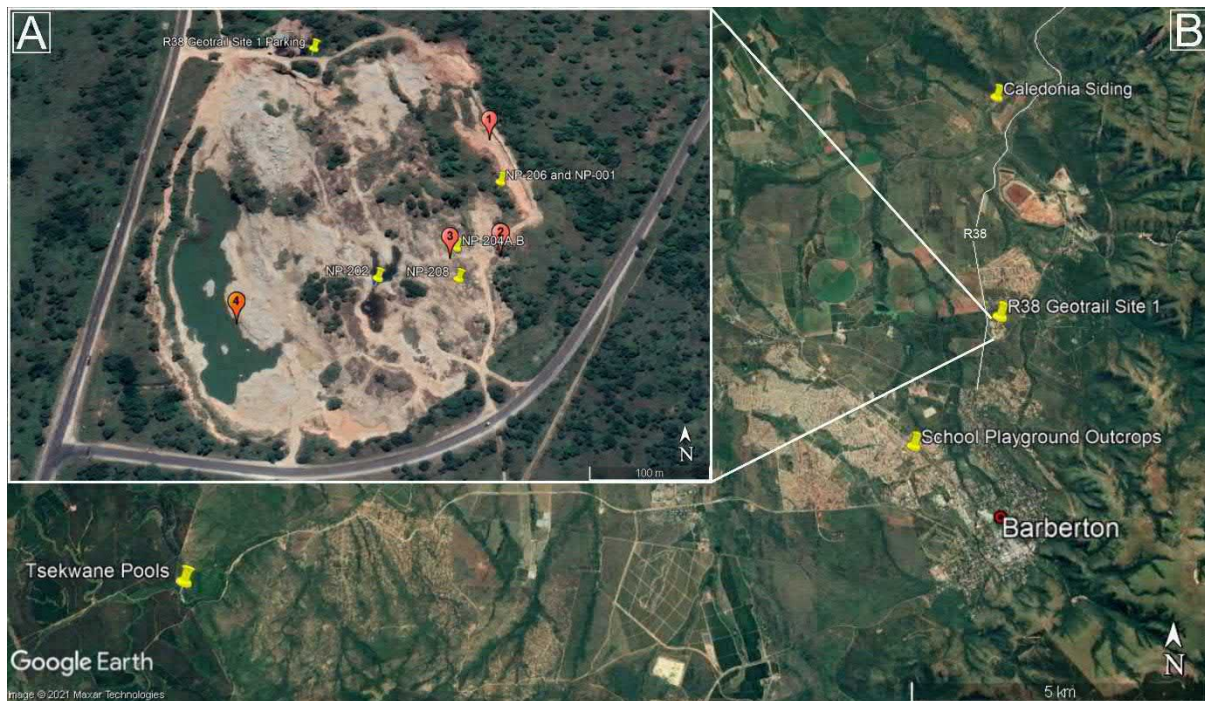


Figure 1.5 Exposures of the Kaap Valley Tonalite at localities that are easily accessed by locals. (A): Site 1 quarry showing sampling locations (yellow pins) and other general features to see (red pins). (B): Localities accessed for enclave descriptions.

1.4 Results

1.4.1 Field observations

Enclaves were more rare (<5 in a $10 \text{ by } 10 \text{ m}^2$ area) at the school playground locality than in the three other localities situated closer to tonalite-BGB contact zones so enclaves are only described from Site 1 and two river pavements: Caledonia Siding and Tsekwane/Tiguan Pools. The enclave population observed mainly contains 5 to 50 cm long, isolated enclaves that are either well-rounded or angular. Rare enclave clusters (Fig. 1.6 I) are comprised of 5 to 40 cm enclaves that do not have mutually complementary margins suggestive of an origin through brecciation of a single mafic dyke or larger enclave (e.g., disaggregated mafic synplutonic dykes of Barnes *et al.* 2021). Instead, enclave clusters are comprised of individual angular to sub-angular enclaves ranging in hornblende abundance from 40 to 70 %. No consistent link was observed between enclave size or shape and hornblende content within the enclave clusters, or the enclave population studied.

All enclaves observed were comprised of at least 20% hornblende and had no apparent rims and or internal enclaves e.g. “double enclaves” of Barnes *et al.* (2021). Internal structures include quartzo-feldspathic dykelets that occur in both circular and angular enclaves (Fig. 1.6 A-C); as well as schistosity in angular variants. The long axis of angular enclaves and their

internal mineral foliation are parallel or sub-parallel to tonalite foliation. In contrast, unshaped circular enclaves do not show any internal alignment or rotation of crystals into alignment with the foliation of the host rock (Fig. 1.6 D, H).

Contacts of the enclaves range from sharp to gradational. Unless partially resorbed, rounded variants mostly have sharp contacts, and the angular varieties exhibit sharp and gradational contacts, sometimes even within different ends of the same enclave. While some enclaves are conspicuous due to a high content of hornblende, some are barely noticeable due to having similar mineral proportions to the host tonalite as well as diffuse margins (Fig. 1.6 C).

Coarse (2-5 mm), isolated feldspar crystals occur heterogeneously within angular and circular enclaves commonly without a systematic increase in modal abundance from margins to cores. In other enclaves, feldspar content is uniform throughout the enclave. The latter are mainly composed of chloritised hornblende, 30-50 % plagioclase and <20% quartz.

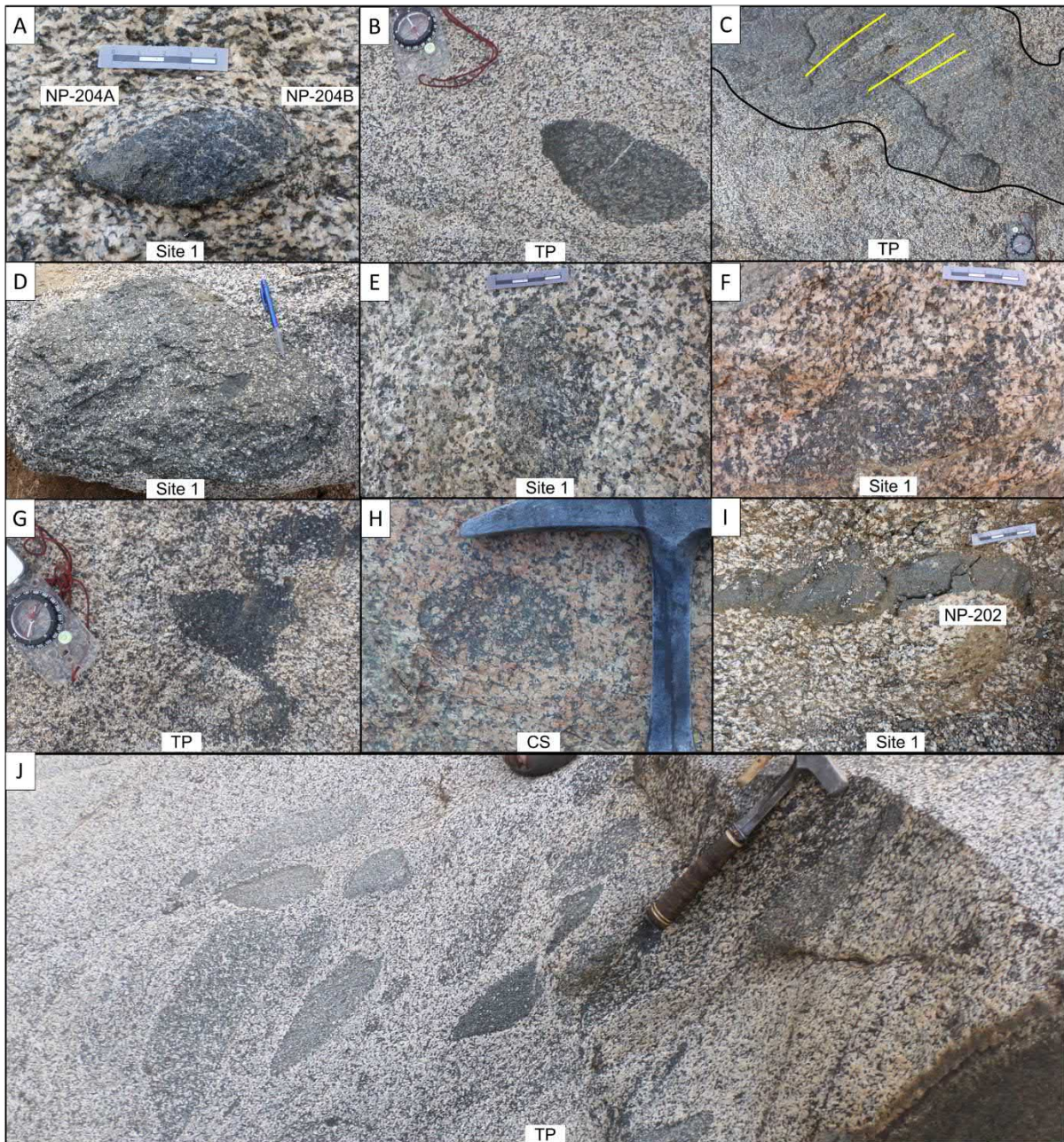


Figure 1.6 Representative mafic enclaves of the Kaap Valley Tonalite from Site 1 of the R38 Geotrail, Tsekwane Pools (TP) and Caledonia Siding (CS). Sample names are included near enclaves, see appendix for other samples. Scale bar has 1 cm segments. (A-C): Enclaves invaded by quartzo-feldspathic dykelets include angular enclaves (A), sub-rounded enclaves (B) and inconspicuous, tabular enclaves (C). Other enclaves are circular (D-E) or angular (F) mafic enclaves with uniformly distributed, coarse feldspar crystals. (G-I): Non-uniform distribution of feldspar in mafic enclaves is observed in enclaves with all or some diffuse margins and more mafic cores (G-H), or discontinuous internal zones of higher feldspar content (I). (J): A rare cluster of enclaves of varying hornblende proportions.

1.4.2 Petrography

Samples studied were predominantly of the hornblende-rich tonalite phase. Texturally, the tonalite is coarse-grained and mainly comprised of euhedral to subhedral crystals of albite and hornblende 2-8 mm in size. The groundmass contains quartz, sericite, subhedral apatite and anhedral zoisite. Where present, euhedral to subhedral biotite is entrained in a weakly developed foliation that anastomoses around angular enclaves. Euhedral accessory apatite, magnetite and sphene are common inclusions in both hornblende and biotite. Hornblende in both the tonalite and enclaves is chloritised. Except for fractured enclaves, the main difference in angular enclaves and the tonalite host was the lower quartz content in enclaves (Fig. 1.6).

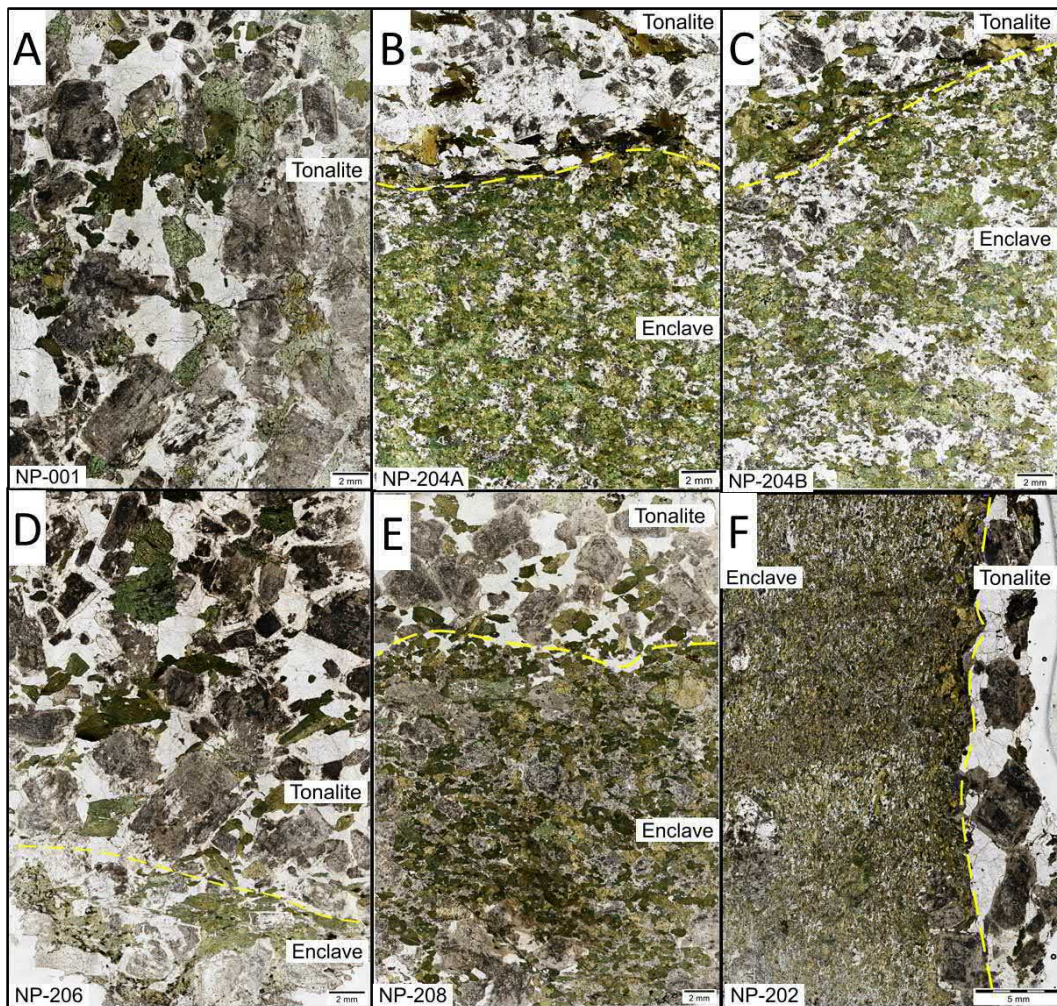


Figure 1.7 PPL photomicrographs of representative samples of the tonalite and enclaves from Site 1, sample numbers are shown in the bottom left corners. (A): Kaap Valley Tonalite. (B-C): Sharp (B) and diffuse contacts of a single hornblende-albite-quartz enclave in a biotite-bearing phase of the tonalite. Note how the quartz-filled diagonal fractures in the lower half of the thin section have transformed part of the enclave into a less mafic phase. (D): Gradational boundary between the tonalite and enclave where the only difference apparent is the higher abundance of chloritised hornblende in the enclave. (E-F): Sharp tonalite-enclave contact where the enclave is a finer grained and quartz-poor version of the host tonalite.

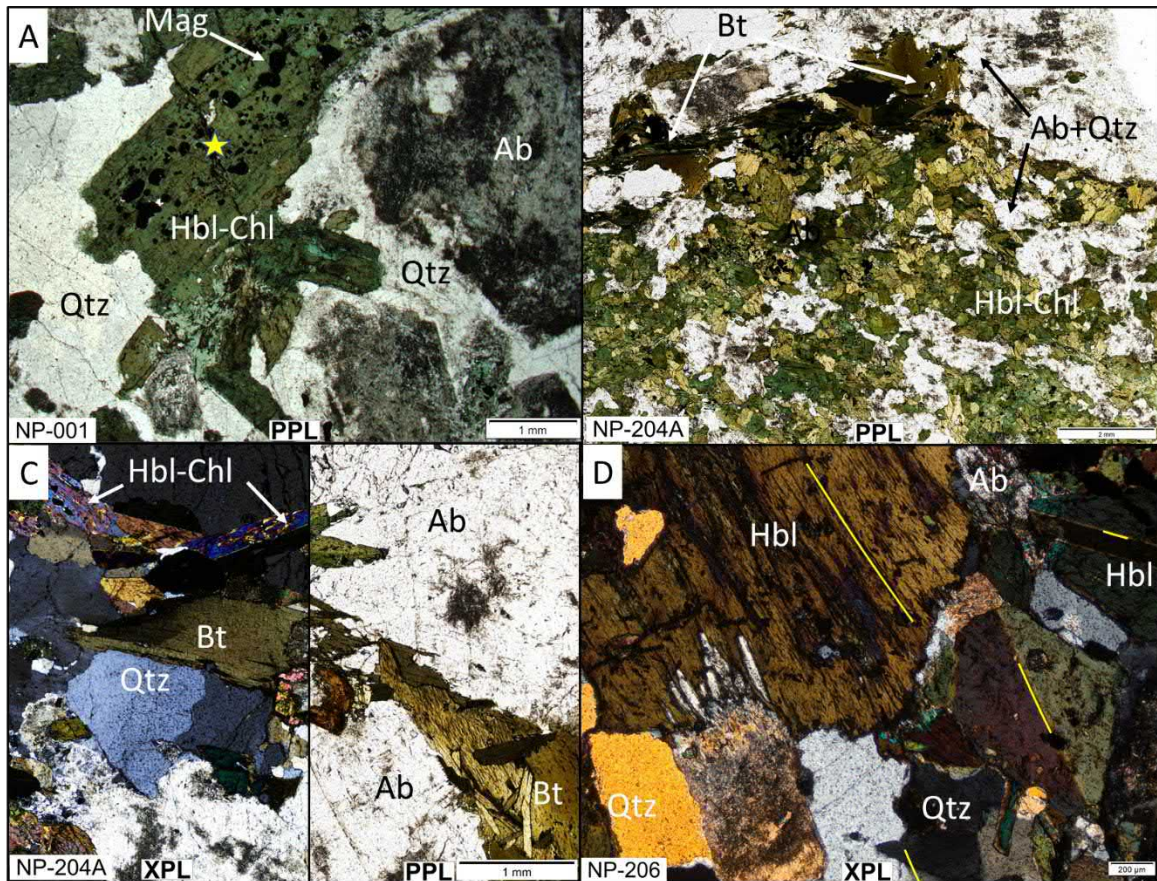


Figure 1.8 Photomicrographs of hornblende textures in representative samples of the Kaap Valley Tonalite from Site 1. PPL, XPL labels are in photomicrographs produced using plane-polarised or cross-polarised light, respectively. (A): Magnetite exsolution in chloritised hornblende (Hbl-Chl). Note the microfault (yellow star) (B): Mainly unchloritised biotite (Bt) along margins of a hornblende (Hbl) enclave that lacks biotite. (C): Mainly unchloritised biotite adjacent to chloritised hornblende. (D): Quartz (Qtz) growth along hornblende cleavage planes. Note the alignment of hornblende cleavage with fold axes of kinked albite (Ab) and quartz (Qtz) in the tonalite sector of enclave NP-206.

1.4.3 Elemental Maps

Element distribution maps of the representative samples of the Kaap Valley Tonalite (NP-001) and a sharp margin of a representative mafic enclave show the following. Magnesium, calcium, iron, titanium, sulphur, chromium and rare copper occur in hornblende. Pyrite is highlighted where iron and sulphur co-exists while ilmenite is highlighted where titanium and iron clusters occur in association with hornblende. Albite crystals coincide with discrete zones of sodium, aluminium and lesser calcium-which is less abundant in the feldspar than the adjacent hornblende crystals. Potassium is concentrated at albite margins and in cores sericitised crystals.

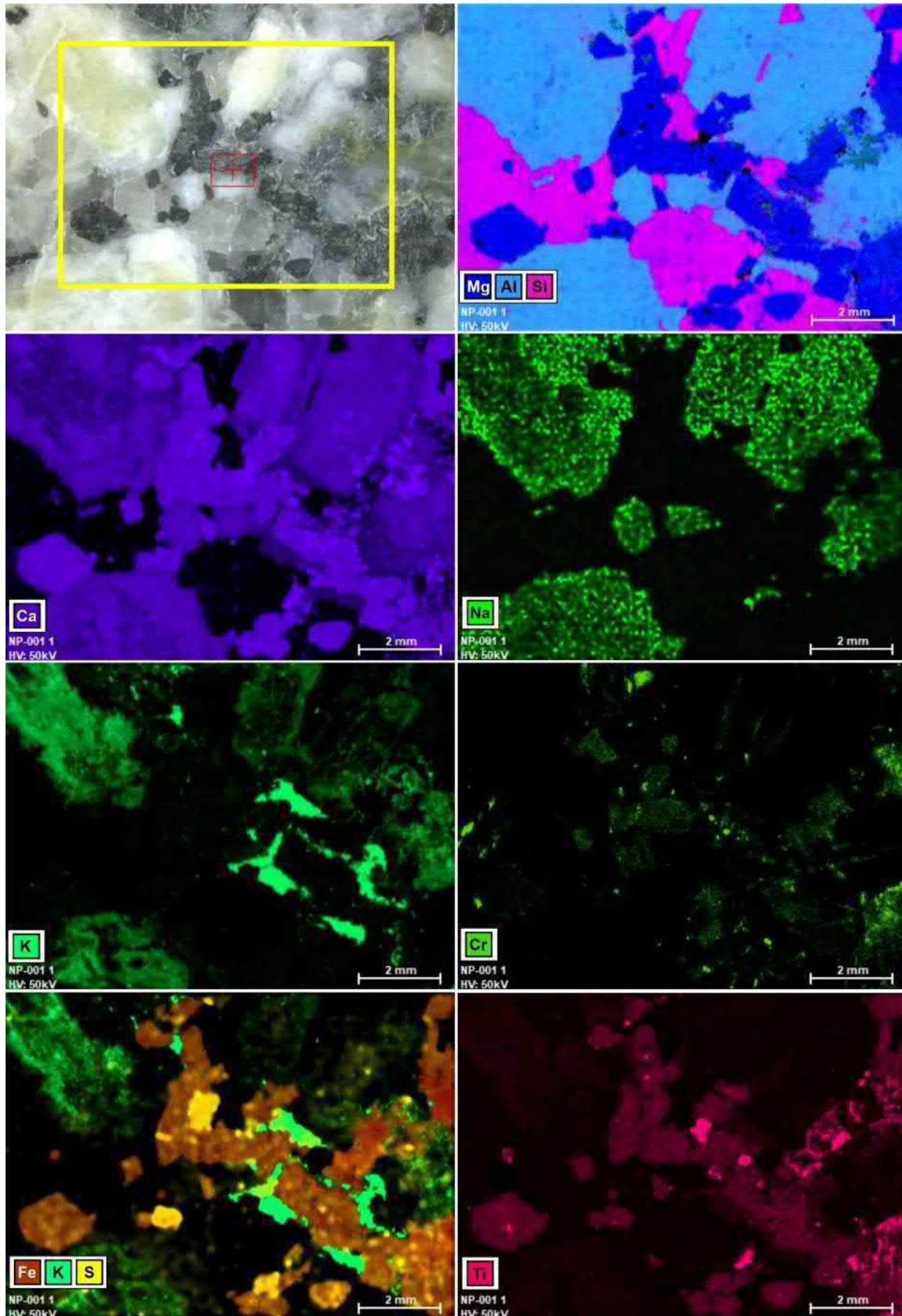


Figure 1.9 Element distribution maps of a representative sample of the tonalite (NP-001) shows an association of albite-forming elements (Na, Al, Ca and lesser K from potassic alteration). Fe, S, Ti and K occur in quartz within interstices of hornblende mineral clusters and Cu is associated with some sulphur clusters. Mafic minerals coincide with Mg, Fe and Cr-rich zones and are associated with discrete sulphur-rich zones.

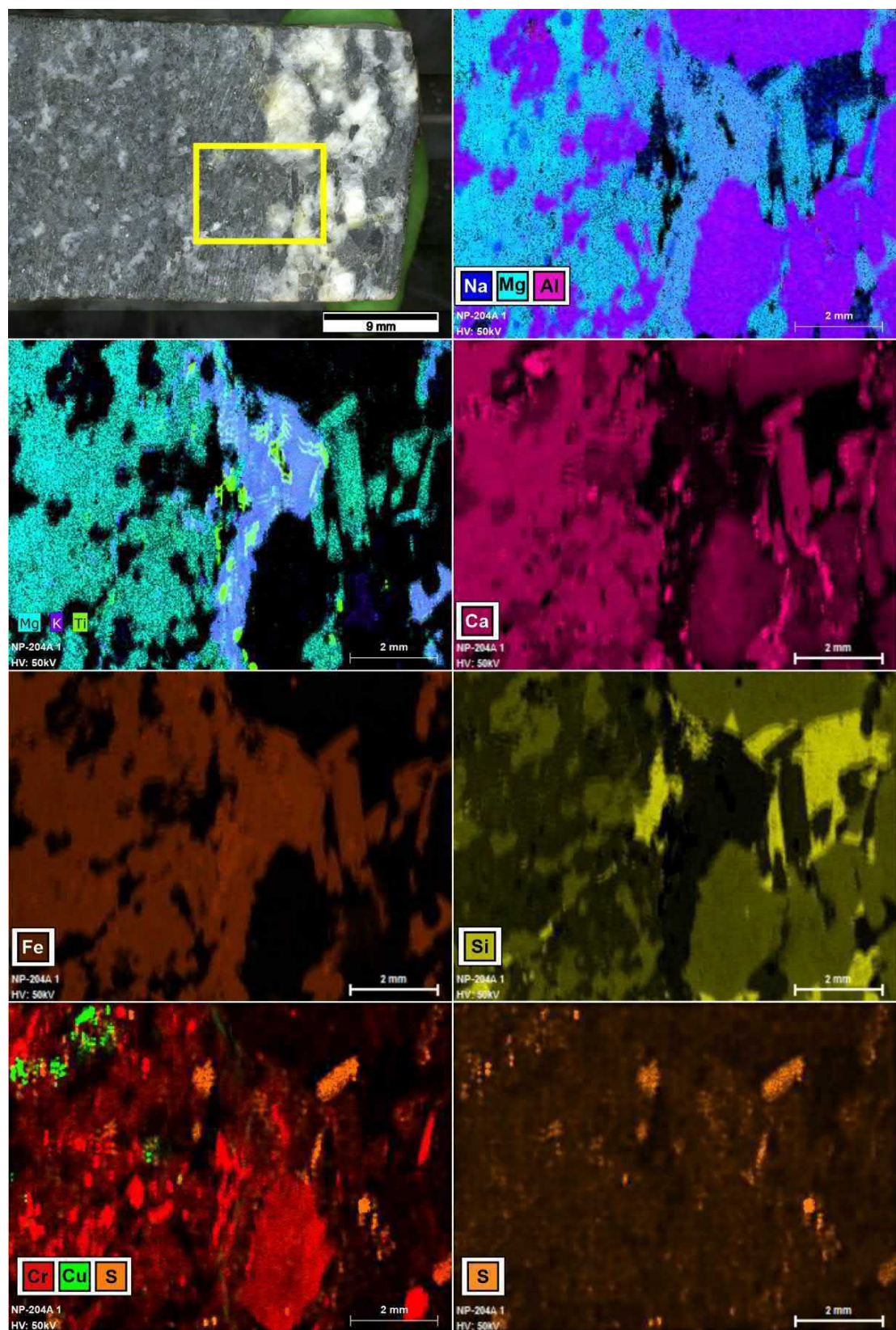


Figure 1.10 Element distribution maps of a representative sample of a fine-grained mafic enclave (NP-204A) shows an association of albite-forming elements (Na, Ca and lesser K). Fe, S, Ti and K occur in quartz that occupies interstices of hornblende mineral clusters while Cu is associated with some sulphur clusters suggesting traces of chalcopyrite. Mafic minerals coincide with Mg, Fe and Cr-rich zones.

1.5 Discussion

Different enclaves of the Kaap Valley Tonalite reveal the following. Some enclaves viewed in the field were inconspicuous due to a similar texture and composition as the host tonalite. They were only identified because they have a slightly higher proportion of hornblende. Such enclaves appear to be a more mafic, possibly an earlier phase of the tonalite, or they may have undergone an expulsion of felsic components due to compression prior to solidification (e.g., Barnes *et al.*, 2021). These inconspicuous enclaves may be angular or circular. The angular variants may be fragments of earlier-crystallised layers of the tonalite, whereas the circular variants may represent unmixed globules or undercooled fractions of basaltic magma introduced to tonalite magma during magma recharge (e.g., Barnes *et al.* 2021 and references therein). However, the presence of brittle fractures even in rounded enclaves (e.g., Fig. 1.6 B) might indicate thermochemical or mechanical rounding of once-angular enclaves that were fractured prior to being incorporated into tonalite magma.

Fracturing of the enclaves facilitated mineralogical equilibration with the tonalite magma. This is seen in the higher feldspar content in enclave margins compared to cores and in highly fractured sectors of deformed enclaves (Fig. 1.6 G-I).

The occurrence of sub-angular enclaves of lower hornblende content within a few centimetres to similar-sized, subangular enclaves of a higher hornblende content (Fig. 1.6 I) suggests an origin of enclaves from different sources. According to Flinders and Clemens (1996), methods of enclave transportation may include magmatic flow that may facilitate clustering of enclaves from unrelated sources.

Petrographic studies of sampled margins of enclaves and the host tonalite revealed the following. The only biotite observed occurred in samples NP-204A and B which are respective thin sections of sharp and gradational margins found at different boundaries of a non-uniform enclave (Fig. 1.6 A). These samples contained clusters of euhedral biotite that were aligned with hornblende cumulates at the tonalite-enclave margin in the tonalite sector of the sample. No biotite was found in the enclave sector of the sample (Fig. 1.8 B, C). According to Faure, (1989), this lack of biotite in the Kaap Valley Tonalite may indicate more advanced chloritisation of biotite in the fractured enclave. However, in the biotite-bearing thin sections studied (NP-204A and B), chloritised hornblende occurred adjacent to biotite that is generally not chloritised (Fig. 1.8 B, C), possibly indicating a composition that does not support chloritisation. Alternatively, the enclave may be a fragment of a less evolved, biotite-free,

earlier fraction of the tonalite enclosed in a younger, biotite-bearing, more evolved, shallow-level fraction of the tonalite as suggested by Robb *et al.* (1986).

The chloritisation and formation of epidote from mafic components occurs throughout the tonalite-enclave samples. This alteration was found to be more advanced in (poorly) mineralised gold-bearing shear zones of the tonalite (Faure, 1989).

Elemental maps show the association of sphene, pyrite ($\pm$ chalcopyrite) and magnetite with hornblende, and presence of sodium and lesser calcium in albite, indicating a composition towards oligoclase. This albite-oligoclase content of the feldspars in tonalite is in agreement with previous descriptions by Robb *et al.* (1986) and Faure (1989).

1.6 Conclusion

Mineralogical similarities between the enclave and host tonalite suggest mineralogical equilibration of angular mafic xenoliths with the tonalite and the partial digestion of early-crystallised phases of the tonalite. Microfaults and kink-folding of mafic and felsic minerals in the tonalite-side of the tonalite-enclave samples of this study (Fig. 1.8) indicates post-crystallisation brittle deformation.

Chapter 2: Verdite at Diggers' Retreat

2.1 Site description

Site 2 is a roadside stop outside a hotel and restaurant called Diggers' Retreat (Fig. 2.1). It is situated near the intersection of the R38 Geotrail and the road to Consort Mine at 25°40'43.60"S; 31° 4'47.69"E

Rocks to be seen here are in a cocopan container filled with green verdite in a grey-green host rock. Also in the container are samples of talc and talc-carbonate schist in serpentinite (see talc description in the Chapter on site 4). Both verdite and talc were transported with permission- from the Handsup Verdite Mine and the Scotia Talc Mine after the mines closed in 1999 and 2019 respectively. A lot of parking space is available around the cocopan on either side of the road.

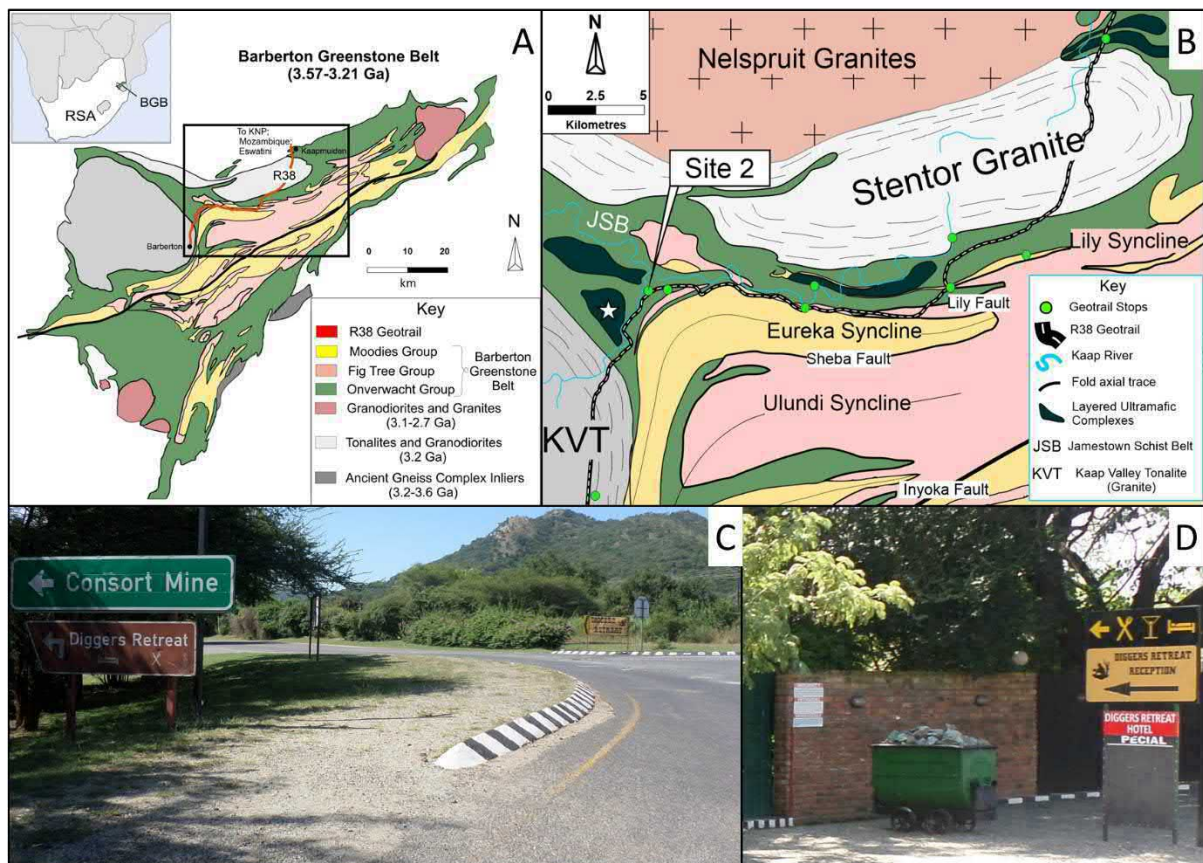


Figure 2.1. Main features of Site 2. (A): Simplified geology of the R38 road in the context of the Barberton Greenstone Belt. Maps modified after Visser *et al.* (1956) and Anhaeusser *et al.* (1981). (B): a detailed view of the area outlined in (A) shows the location of site 2 northeast of the Handsup Layered Ultramafic Complex (white star) where verdite was once mined. (C): View of the turn off to site 2. (D): Cocopan at site 2 filled with verdite samples from the Handsup Mine and talc samples from Scotia Talc Mine.

2.2 Aspects of verdite and Site 2 that may interest laypersons

Verdite (a variety of fuchsite), buddstone (a mottled or layered green stone related to verdite that is sometimes called “South African Jade”) and other colourful semi-precious ornamental stones were mined in the Barberton area from alteration zones of ultramafic lithologies. On the way to site 2, waste rock of the Handsup Mine will be an obvious feature on the western side of the R38 road about 9 km north of Barberton. This old mine is one of many historical mines of the region where colourful minerals of chromium were extracted. The Handsup quarry is visible from the turn-off to Site 2 and from the backyard of Digger’s Retreat.

Visitors could be told about barbertonite ($\text{Mg}_6\text{Cr}_2(\text{OH})_{16}[\text{CO}_3]\cdot 4\text{H}_2\text{O}$), a colourful chromium hydrotalcite aggregate named after the town of Barberton (Fron del, 1941). It was discovered in 1941 within the Barberton region, in a layered ultramafic host of the Kaapsehoop Asbestos Mine c. 30 km northwest of Barberton (Fron del, 1941). Barbertonite is also known as stichtite-2H (Mills *et al.*, 2011) as it is a polytype of, and commonly occurs with stichtite (Ashwal and Cairncross, 1997; Mills *et al.*, 2011). It occurs as pink, lilac or purple masses within altered chrome-rich serpentinites (Ashwal and Cairncross, 1997; Mills *et al.*, 2011).



Figure 2.2 Hand specimen of barbertonite (stichtite-2H), a lilac-to-purple chromium hydrotalcite mineral. It is a polytype of, and usually co-occurs with Stichtite-3R (Mills *et al.*, 2011).

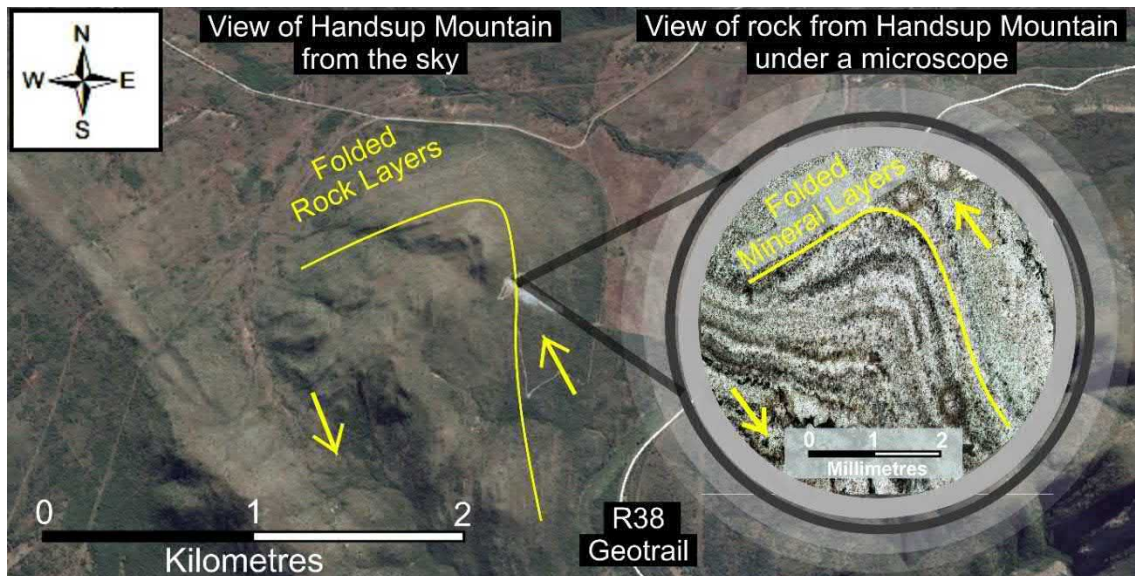


Fig. 2.3 An example of related geological features that occur at different scales. The structural fabric near the Handsup verdite quarry can be used to show visitors how features present clues used by geologists to make plausible reconstructions.

Within Digger's Retreat there is a display of described rock samples of selected local lithologies. They include banded ironstones of the Fig Tree Group that provided sub-economic iron oxide deposits which have been mined in the region since pre-colonial times (Fig 2.4, Myburgh, 1949).



Figure 2.4 Local Swati women carrying red ochre from Dumaneni, a pre-colonial iron mine at the northeastern margin of the Barberton Greenstone Belt. Image sourced from Myburgh (1949).

2.3 Geology

2.3.1 Background

2.3.1.1 Handsup Layered Ultramafic Complex

Situated at the southeastern margin of the Jamestown Schist Belt (JSB), the Handsup Layered Ultramafic Complex occupies a tectonically complex zone affected by multiple deformation episodes. These include the regional shortening events of the BGB, possibly associated with the forceful shallow emplacement of the Stentor Pluton (c. 3250 Ma) in the northeast, the emplacement of the Kaap Valley Tonalite (c. 3227 Ma) in the south, and the intrusion of Nelspruit granites in the north (c. 3109 Ma) (Anhaeusser, 2019).

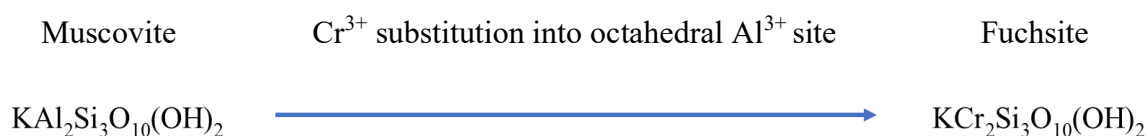
The Handsup Layered Ultramafic Complex occurs as a sub-vertical anticline detached in a top-to-the-northwest sense along the Albion Fault (Anhaeusser, 1969, 2019). It is made up of cyclic compositional layering that may include a basal dunitic layer, overlain respectively by a harzburgitic, websteritic, gabbroic and marginally anorthositic gabbro layers. Unlike similar layered complexes at the northeastern and southwestern margins of the BGB, lithologies of the Handsup Layered Ultramafic Complex are generally advancedly altered. Common alteration includes serpentinisation, carbonatisation, steatitisation, uralitisation, chloritisation, saussuritisation, sericitisation, propylitisation and silicification (Anhaeusser, 1969, 2019).

Layered ultramafic complexes of the BGB intruded only into sequences of the Onverwacht Group (Anhaeusser, 2006; Hofmann *et al.*, 2021). In the northern margin of the BGB, they intrude volcanosedimentary sequences of the Weltevreden Formation which is diachronous with uppermost Onverwacht Group strata of the southern BGB (Hofmann *et al.*, 2021). A Sm-Nd age of c. 3286 and Re-Os age of c. 3266 were determined by Lahaye *et al.* (1995) and Connolly *et al.* (2011) respectively.

No direct age has been determined from the Handsup Layered Ultramafic Complex. However, Pb-Pb ratios of ultramafic and mafic lithologies from the adjacent Mundt's Concession Layered Ultramafic Complex defined an age of c. 3244 Ma (Dupré and Arndt, 1990). More recently, a direct age of $3\,244 \pm 11$ Ma was determined from a gabbroic unit of the same Layered Ultramafic Complex (Hofmann *et al.*, 2021). If the Handsup and Mundt's Concession Layered Ultramafic Complexes constitute a continuous layered sequence as proposed by Anhaeusser (1969), the abovementioned ages may be regarded as age estimates for lithologies of the Handsup Layered Ultramafic Complex.

2.3.1.2 Verdite

Verdite is a variety of fuchsite schist found in altered serpentinites (Anhaeusser, 1963). According to Burns and Burns (1975) and Treloar (1987), the green colour in verdite results from a substitution of Cr^{3+} for Al^{3+} in octahedral sites of muscovite.



True verdite is described as a green semi-precious ornamental stone with a hardness range between 3 and 4, and a specific gravity range between 2.70 and 2.87 (Harding and Jobbins, 1984; Groeneveld, 1975). Compositionally, it is described as a schist composed mainly of fuchsite (chrome-muscovite) and varying proportions of rutile, quartz, albite, chlorite, talc, and corundum (Anhaeusser, 1969, 1975; Groeneveld, 1975; Antenen, 1991).

The name ‘verdite’ in Wybergh (1932) and Anhaeusser (1963) was used to describe a soft, green, fine-grained, fuchsite-rich alteration product of mafic-to-ultramafic serpentinite. According to Antenen (1991), the name was derived from ‘verde’, the Spanish, Portuguese or Italian word for green. ‘Verde’ (Kerrich *et al.*, 1987), verd antique or ‘African Jade’ (Hobbs, 1982) as well as serpentinite (Anhaeusser, 1984) are other terms which have been used in literature to describe verdite. Except for serpentinite/verd-antique, these various names for verdite all described a soft, green, fuchsite-rich schist with abundant rutile and varying proportions of quartz, feldspars and margarite, an Al-rich muscovite (Anhaeusser, 1963, 1975; Groeneveld, 1975; Harding and Jobbins, 1984).

Verdite from the Barberton region has been known since 1907 before it was discovered in Zimbabwe where it is mined as a by-product of corundum production. Zimbabwean reserves have outlasted South African counterparts, the majority of which were confined to the northern margin of the Barberton Greenstone Belt (Groeneveld, 1975; Antenen, 1991). Best occurrences in the Barberton region, such as the one mined at the prominent Handsup quarry visible from the R38 road 12 km north of Barberton, were found within some multiply deformed and metasomatised Archean ultramafic sequences.

According to Groeneveld (1975), three varieties of verdite were identified in quarries of the Barberton Greenstone Belt. The first type was pale green in colour, of a granular texture, compact and hard. This variety was found in a gradational contact with a white rock with the same texture. The second type was a dark green, foliated, much softer variety. The third, a

green and white brecciated variety with white fragments often drawn out into thin bands folded together with the verdite. The latter variety is a semi-precious decorative stone locally known as buddstone. The three variants were found to grade into each other (Groeneveld, 1975).

Later work by Antenen (1991) identified sub-economic verdite deposits reported to be altered mafic xenoliths sheared together with the encompassing 3.1 Ga potassic Nelspruit Batholith. No paragenetic sequence or metamorphic grade was described due to the ultramylonitic fabric of samples studied by the same author.

A number of workers have investigated the composition of verdite within the Barberton Greenstone Belt, the Mashishimala deposit in the Murchison Greenstone Belt to the north, as well as the 'O'Briens Claims' and the 'Concessions' deposits found in the 2.73 Ga Salisbury Greenstone Belt of Zimbabwe (Partridge, 1936; Schreyer *et al.*, 1981; Kerrich *et al.*, 1987; Antenen, 1991). In all of these deposits, all samples were dominated by chrome-muscovite with a whole-rock concentration of 0.2-4 wt % Cr (Table 1), >25wt. % alumina and >7wt. % potash with abundant rutile and lesser chromite and magnetite as main opaque mineral phases.

Various lithological associations of verdite occurrences were reported in attempts to characterize verdite deposits. It was initially proposed that verdite occurrences were linked to Cr-metasomatism of felsic schists (Hutton, 1942). Partridge (1936) proposed a genetic-and sometimes-spatial link of verdite to granitic bodies (Partridge, 1936). In contrast, an association of verdite occurrences with sheared contacts between altered serpentinitised pyroxenites and metagabbros were noted by Anhaeusser (1963, 1972). In support of the latter, Groeneveld (1975) reported a close association of talc-chlorite lenses and verdite deposits in shear zones near contacts of a serpentinitised metagabbro. This linked verdite formation to intrusive ultramafic lithologies which have been multiply metasomatised, initially through serpentinitisation, then steatitisation to produce talc, and later K-, Al-, Cr and Si- metasomatic processes of "verditisation" under conditions that are only partially understood.

At a later stage, Pearton (1981) proposed a sequence of alteration events whereby K-, Al-, Cr- and Si- metasomatism associated with verdite petrogenesis, was subsequent to the carbonitisation of mafic/ultramafic wallrocks. In this way, carbonates were understood to form part of the composition of verdite (Anhaeusser, 1969; Groeneveld, 1975; Antenen, 1991; Ward, 1999, Cairncross, 2004).

In later years, more verdite occurrences were discovered in the area north of the BGB. These deposits were described as northwest-trending, sheared ultramafic xenoliths within the

Nelspruit Batholith (Antenen, 1991; Cairncross and Dixon, 2007). A total of three deposits occurring from ca. 5 km (Nelspruit quarry) up to 30 km north of the northern margin of the BGB were identified (Antenen, 1991).



Figure 2.5 Background of the old Handsup Mine. (A) Polished verdite ornament sourced from the quarry. (B) View of the unrehabilitated verdite quarry visible from the R38 road situated within the Jamestown Schist Belt. (C) Modified geological map of the Handsup Layered Ultramafic Complex (Anhaeusser, 1975) superimposed on a Google Earth image. White rectangle mark shows that the Handsup verdite quarry is situated at a lithological boundary between a pyroxenitic serpentinite and a metagabbro associated with a pyroxenite.

2.3.2 Present study

2.3.2.1 Research question

Verdite from the Barberton Greenstone Belt has been described by several workers, however, aspects of the mineralogy, occurrence and paragenetic model remain unclear.

2.3.2.2 Aims

This work aims to characterise, and propose a paragenetic model of, the best quality verdite from the Barberton Greenstone Belt and from a widely unknown, possible extension of the Jamestown Schist Belt located in Nelspruit (Fig. 2.10).

2.3.2.3 Methods

2.3.2.3.1 Sampling

The Handsup Layered Ultramafic Complex (LUC) was selected for sampling due to the occurrence of high quality verdite. Due to the rarity of in-situ verdite within the quarry, samples thereof were collected from the quarry dump. Care was taken to collect representative variants in order to link composition of samples to physical features observed.

Verdite host rocks were sampled together with intrusive bodies proximal to the only remaining verdite-bearing alteration zone still in-situ within the quarry. Less altered equivalents of the host rocks were also sampled from outside the quarry along strike from those bearing verdite within the quarry. This was to identify assemblages only present in altered verdite-bearing host rocks.

Additional samples of verdite and the host rocks were obtained from a different geological context which lies in Nelspruit at a location ca. 19 km north of the Handsup LUC verdite (Fig. 2.10). This was done to assess the validity of a petrogenetic and paragenetic model proposed for the Barberton verdite sampled from the Handsup LUC.

2.3.2.3.2 Petrography

Labelled samples were cut into thin sections. Phase maps of thin sections and elemental maps of thin section off-cuts were produced to confirm petrographic observations made on fine grained, and highly altered samples of this study.

2.3.2.3.3 Bulk Chemical Analysis

X-ray fluorescence spectrometry was used to determine major and trace element compositions of selected samples.

2.4 Results

2.4.1 Field Occurrence: Handsup quarry

At the Handsup quarry, a NW-trending shear zone at the centre of the quarry intercepts north-northeast trending strata of a westerly dip. From the quarry entrance in the southeast to the only remaining in-situ verdite in the northwestern part of the quarry are asbestos-bearing serpentinites (Sample HUVH) that are interlayered with thin (<25cm) grey chert layers. Some of these serpentinites contain disseminated grey-white millimetre-scale masses, are multiply deformed and in some areas within the quarry, occur as thin shards known as “pencils” formed by the intersection of two cleavage planes.

Verdite occurs in southwards-plunging lenses within a grey-brown, sub-vertically oriented chlorite-epidote schist (Fig. 2.6 B C, D). Disseminated within the schist are white, millimetre-scale ovoid masses of albite that either occur as individual masses or lenticular aggregates in northwest-plunging tension gashes. Sheared and altered equivalents of these are altered to sericite $\pm$ kaolinite possibly from recent weathering. Sub-vertical quartz-feldspar porphyritic dykes occur at both margins of the verdite-bearing chlorite schist and are the likely source of the sericite $\pm$ kaolinite in the tensional fractures of the chlorite-epidote schist. A few tens of centimetres northeast of the verdite lenses, the verdite-bearing sericite-chlorite-epidote schist and an altered, southwest-dipping quartz-feldspar porphyritic dyke are truncated by a harzburgite dyke. A thicker quartz-feldspar porphyry dyke occurs ca. 1m south of the verdite zone (Fig. 2.6 A) and is offset in a top-to-the southeast sense at the top of the verdite-bearing zone.

In situ verdite observed invariably occurs as lenses between the chlorite-epidote schist and films of a powdery kaolinite-sericite schist (Fig 2.6 B, C). A few tens of centimetres towards the northeast of the verdite lenses, the verdite-bearing sericite-chlorite-epidote schist and an altered, southwest-dipping quartz-feldspar porphyritic dyke are truncated by a harzburgite dyke. A thicker quartz-feldspar porphyry dyke occurs ca. 1m south of the verdite zone (Fig 2.6 A, B, sample CDWV) and appears to be offset in a top-to-the southeast sense at the top of the verdite-bearing zone. At the northeastern edge of the verdite-bearing zone is a layer-parallel zone of carbonated serpentinite (sample HUT). However, no obvious zones of pervasive carbonate development were observed in the quarry or vicinity.

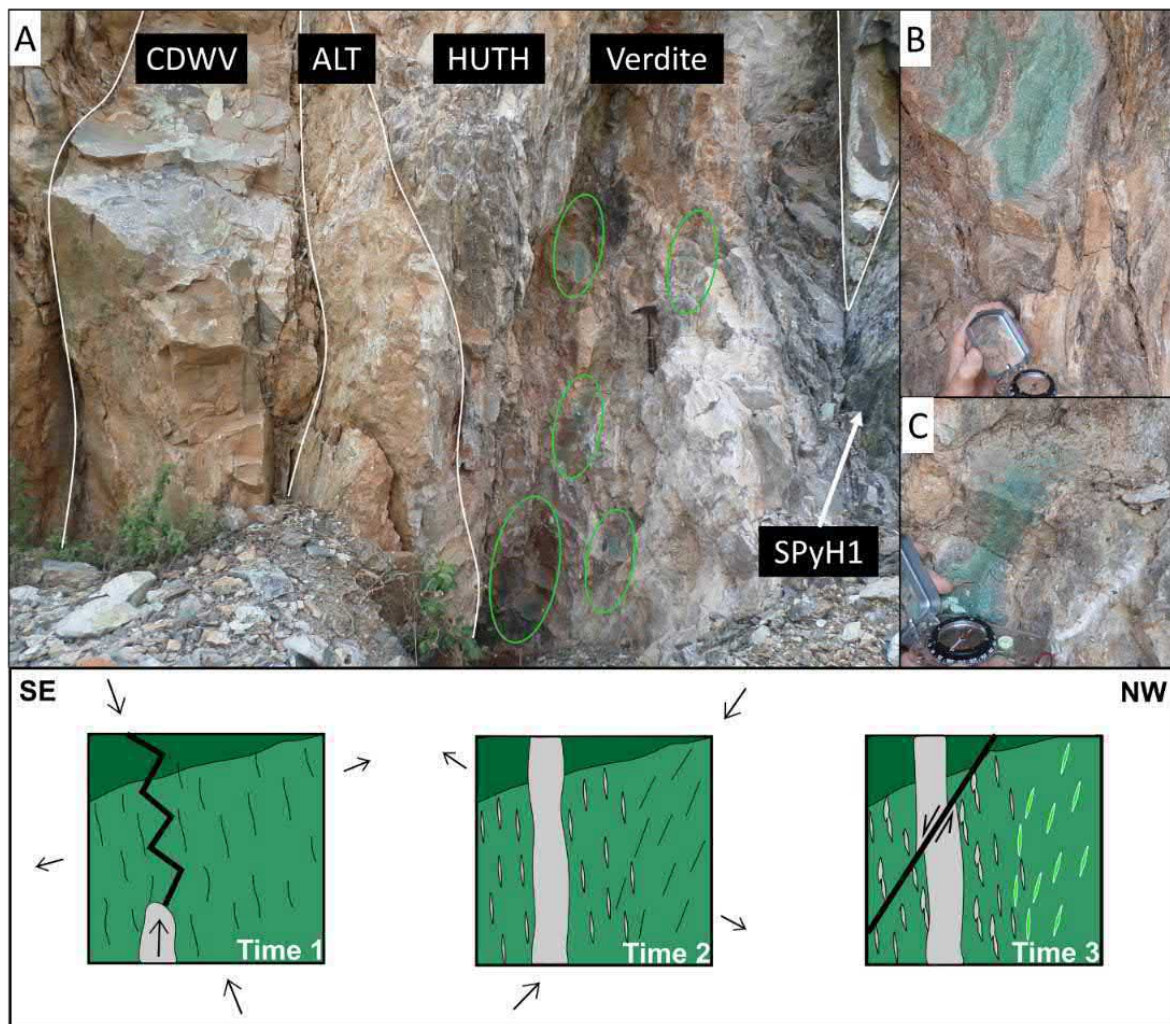


Figure 2.6 Lithological context of the remaining in-situ verdite pods in the Handsup Quarry. (A): Verdite lenses in a sub-vertical array of southwest-plunging tension gashes. Labels refer to sample names except “ALT” which represents sericitised and kaolinitised feldspar porphyry. (B-C): The verdite lenses occur within earlier sub-vertical tension gashes that plunge towards the northwest. (D): Illustrated diagrams show the development of tension gashes mentioned above. The deformation sequence is based on features of the intrusive quartz-feldspar porphyritic dyke (CDWV) and the orientation of tension gashes. See Appendix B, Figure 1.

2.4.2 Petrography: Handsup quarry

The chlorite schist host rock of verdite (HUTH/CHWVH) is comprised of chlorite groundmass/oikocrysts that envelop kinked, ovoid albite masses that sometimes exhibit a Maltese Cross extinction (bottom right part of Fig. 2.7 A). The chlorite crystals are partially replaced by muscovite. Fine aggregates of chromite are disseminated throughout the sample, occurring as inclusions in albite crystals as well as in the muscovite-rich groundmass (Fig. 2.7 B and E). Tensional fractures of different generations cut through the schist. The earliest is a feldspar-rich filled phase now occurring as muscovite-rich lineaments at a slight angle to

schistosity (centre of Fig. 2.7 A). The second generation is characterised by spinel-bearing fractures that offset the layer-parallel felsic veins. The latter are kinked sub-horizontally in a top-to-the northwest shear sense (Fig. 2.7 B-C).

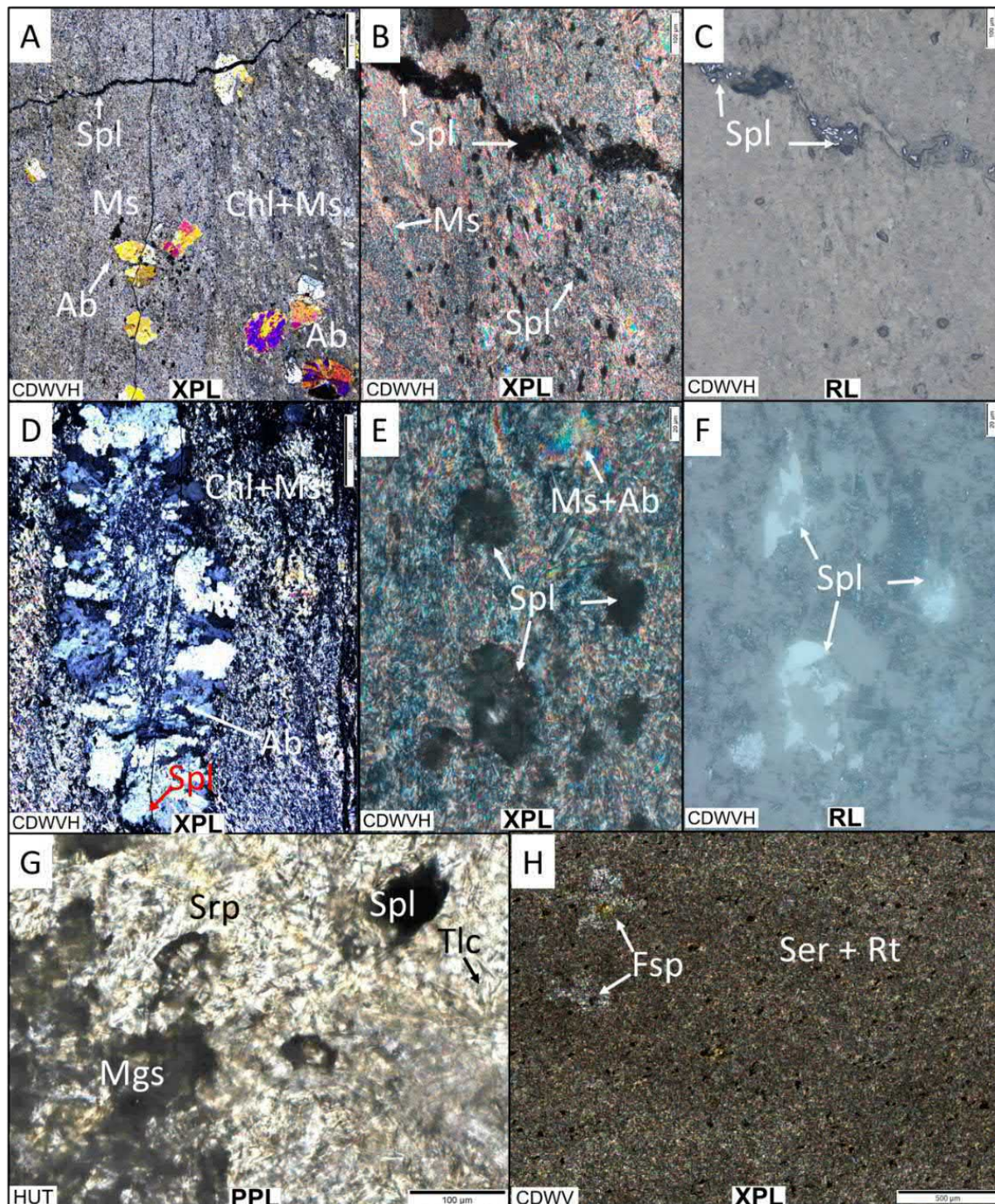


Figure 2.7 Characteristics of the chlorite-muscovite-albite schist host of verdite in the Handsup quarry. Sample names are in bottom-left corners and PPL, XPL and RL labels are in photomicrographs produced using plane-, cross-polarised or reflected light, respectively. (A): Subvertical muscovite-rich lenses with albite crystals at kinks of an opaque vein. (B-C): Photomicrographs of disseminated clusters of chromite in a micaceous groundmass cut by a vein kinked along the orientation of mica fish. Magnetite crystals occur in dilational jogs of the kinked vein. (D): Syntaxial feldspar vein kinked normal to vein opening is cut obliquely by a thin opaque vein. (E-F): Bladed chromite crystals growing at margins of advancedly altered ovoid crystals of albite. (G): Steatized and carbonated serpentine in the flanks of the verdite-bearing zone. (H): Advancedly sericitized feldspar porphyry. Cr-Ms: chrome-muscovite, Ms: muscovite, Chl: chlorite, Ab: albite, Spl: spinel (chromite/magnetite), Srp: serpentine, Tlc: talc, Mgs: magnesite, Ser: sericite, Fsp: feldspar, Rt: rutile.

Though verdite from Zimbabwe and the Limpopo Province of South Africa may contain red spots due to the presence of red rubies (Schreyer *et al.*, 1981), the colours of Handsup verdite are mainly different shades of green, white, and yellow. These colourful layers are respectively made up of chrome muscovite, stringers of albite ovoid masses or lenses, and clusters of euhedral rutile crystals associated with sericitized albite layers. The chrome-muscovite which dominates the samples studied occurs as radial or cross-shaped aggregates with no preferred orientation, but where deformed, strong mineralogical layering characterizes the samples. Three main textures that were observed are described below and shown in Fig. 2.8.

Type 1

A sample of this variant is the closest to true verdite (NP-004). It is the softest sample and is vividly green. It consists of inter laminations of distinct, fine-textured mineral layers. The thickest and dominant layers are composed of chrome-muscovite intergrown with chlorite. Thinner layers include very fine sericite-feldspar-quartz layers and layers of rutile aggregates responsible for the characteristic yellow streaks in verdite hand specimens. Very rarely are albite masses and lenses preserved in this variant. The few observed have been heavily sericitized or resorbed into the micaceous groundmass and are better preserved within microfold hinges and along boundaries of mineral layers (Fig 2.8, H-I). Late-stage veinlets of albite cut across mineral layers. Opaque minerals were too fine to identify.

Type 2

Samples of this variant are dominated by lenticular aggregates of albite with chrome-muscovite rims separated by micro-shear zones filled with coarse, interlocking muscovite crystals. These

lenticular aggregates of albite are set within chlorite partially replaced by randomly-oriented muscovite (Fig. 2.8.C-E). In sheared samples, the felsic lenses are subdivided into shear bands with margins lined by sillimanite (Fig. 2.8 I). No sillimanite was found in the chrome-muscovite zone which lies in a transitional contact with the sillimanite-bearing albite lenses.

Aggregates of euhedral rutile occupy fractures and cleavage planes of albite. Clusters of chromite and magnetite are disseminated in the fine-grained, vividly green chrome-muscovite groundmass and show no preferred orientation except when sheared into alignment with the schistosity of the sample (Fig. 2.8 C). Unsheared clusters of these oxides are associated with pale-coloured muscovite (Fig. 2.8 C).

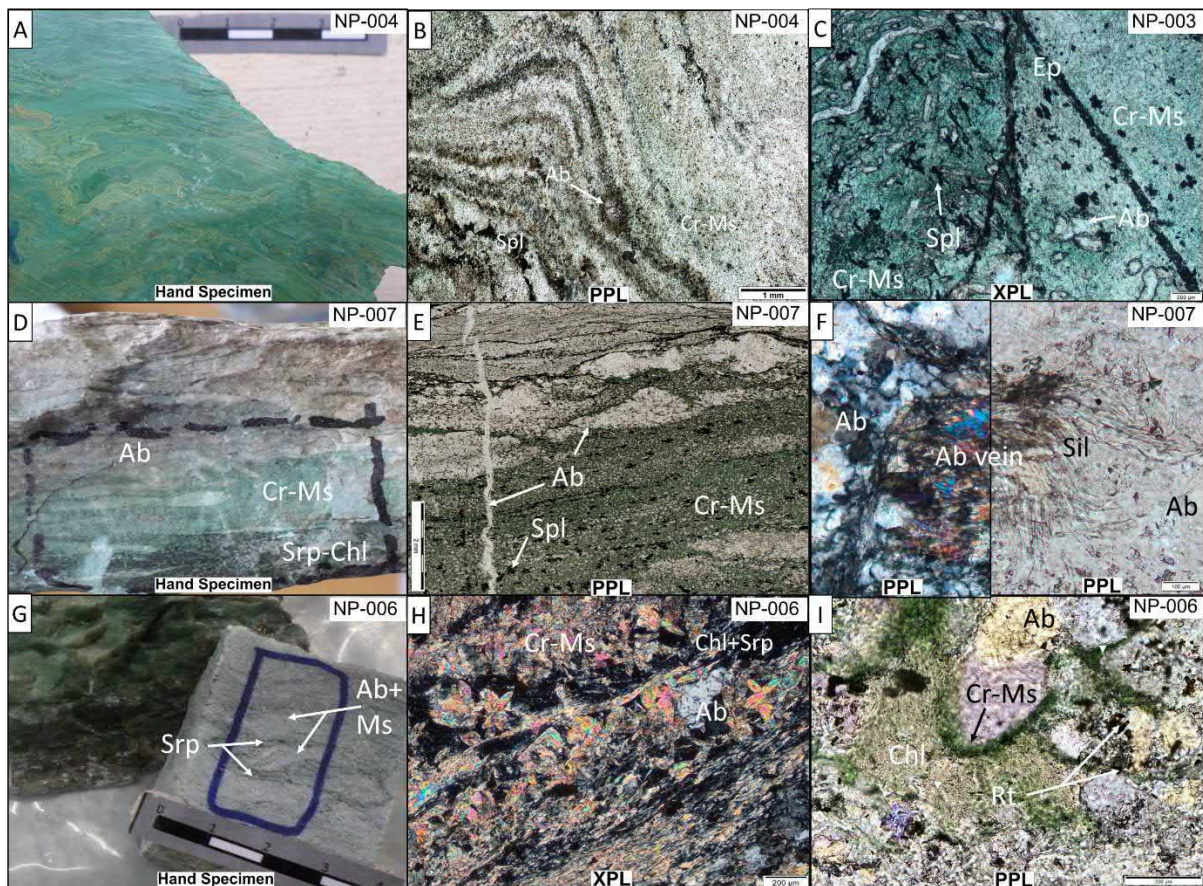


Figure 2.8 Mineral composition of the verdite-bearing samples and host rocks. Sample names are in the top right corners and PPL and XPL labels are in photomicrographs produced using plane-polarised or cross-polarised light, respectively. (A-B): Folded mineral layers in true verdite (Type 1). (C): Layered sample of incomplete verdite-formation shows more chrome-muscovite in a sheared sector of the same fuchsitic schist. (D-F): Lens-bearing verdite (Type 2) at the interface of a chloritised serpentinite and an intrusive, porphyritic albite ± quartz dyke with sillimanite. (F). (G-I): Chloritised serpentinite in-filled by muscovite along tension fractures. This sample contains chrome-muscovite at margins of albite crystals that occur within lenses filled by porphyritic albite ± quartz fluid. (I). Ab: albite, Spl: spinel (chromite/magnetite), Cr-Ms: chrome-muscovite, Ep: Epidote, Srp-Chl: chloritised serpentine, Sil: sillimanite, Chl: chlorite, Rt: rutile.

2.4.3 Field Occurrence: Nelspruit mafic ridge

Additional samples of verdite and host rocks (Fig. 2.9) were collected in Nelspruit (Fig. 2.10) for comparison. At this locality, verdite lenses within quarry A in Nelspruit occur within a mafic ridge, in a schistose foliation that is strike-parallel to a northwest-trending shear zone. The shear zone cuts through a sub-vertical chlorite schist with interfolial veins of white quartz either encrusted with goethite or bearing prismatic voids lined with hematite.

The chlorite schist in the southwestern quarry wall is folded sub-horizontally and the fold amplitudes become larger towards the southeast (f1 in Fig. 2.11 A-D). The larger structure of the folded quarry wall is here proposed to be a southeast-plunging antiformal syncline with z-folds forming the southeastern wall of the quarry. A proposed interpretative model (Fig. 2.11 D) depicts the proposed structural context of verdite in quarries A and B where verdite was once mined.

Quartz $\pm$ feldspar veins intrude verdite in both quarry A and B. In quarry A, verdite occurs between an altered leucogranite and the chlorite schist. The contact is part of a sub-vertical, top-to-the-southeast, transpressional shear zone that runs parallel to the schistose foliation of the chlorite schist. The altered leucogranite is slightly sheared (Fig. 2.9 A) and marks the northwestern margin of the quarry.

At the southeastern edge, the same chlorite schist is bound to the south by a northwest-trending fault zone occupied by a cohesive fault rock (Fig. 2.9 G) which is in turn bound to the west by boulders of massive cryptocrystalline white quartz with voids (quartz-feldspar porphyry?) that occupies at least 100 m² of grassland. No contact was observed between the white quartz and the adjacent verdite zone or with the closest potassic granite which lies c. 100 m southwest of the quarry.

Texturally, the fault rock at the southern extent of quarry A contains sub-rounded to sub-angular, partially digested, boulders and cobbles comprised of alternating layers of white quartz $\pm$ feldspar and pale green verdite (Fig. 2.9 F). These clasts represent a semi-precious carving stone known locally as buddstone (Anhaeusser, 1969, 2013). Buddstone clasts in the fault rock are aligned with the northwest-trend of the verdite-bearing shear zone in the quarry. A similar lithology was sampled at quarry B where advanced silicification and compression had not occurred (Fig 2.9. G). The latter could be separated into the verdite matrix (NBN4A) and powdery cobbles of sericite-rich nodules in a verdite groundmass (sample NBN4B/C). These

cohesive faultrocks at Quarry B only occur as boulders close to sparse outcrops of the mafic ridge which, at that locality, occurs as sparse, metre-scale massive bodies of pale to dark green chrome muscovite. The latter are invaded by veins of pinkish, euhedral quartz veins with prismatic voids (Fig. 2.9 E).

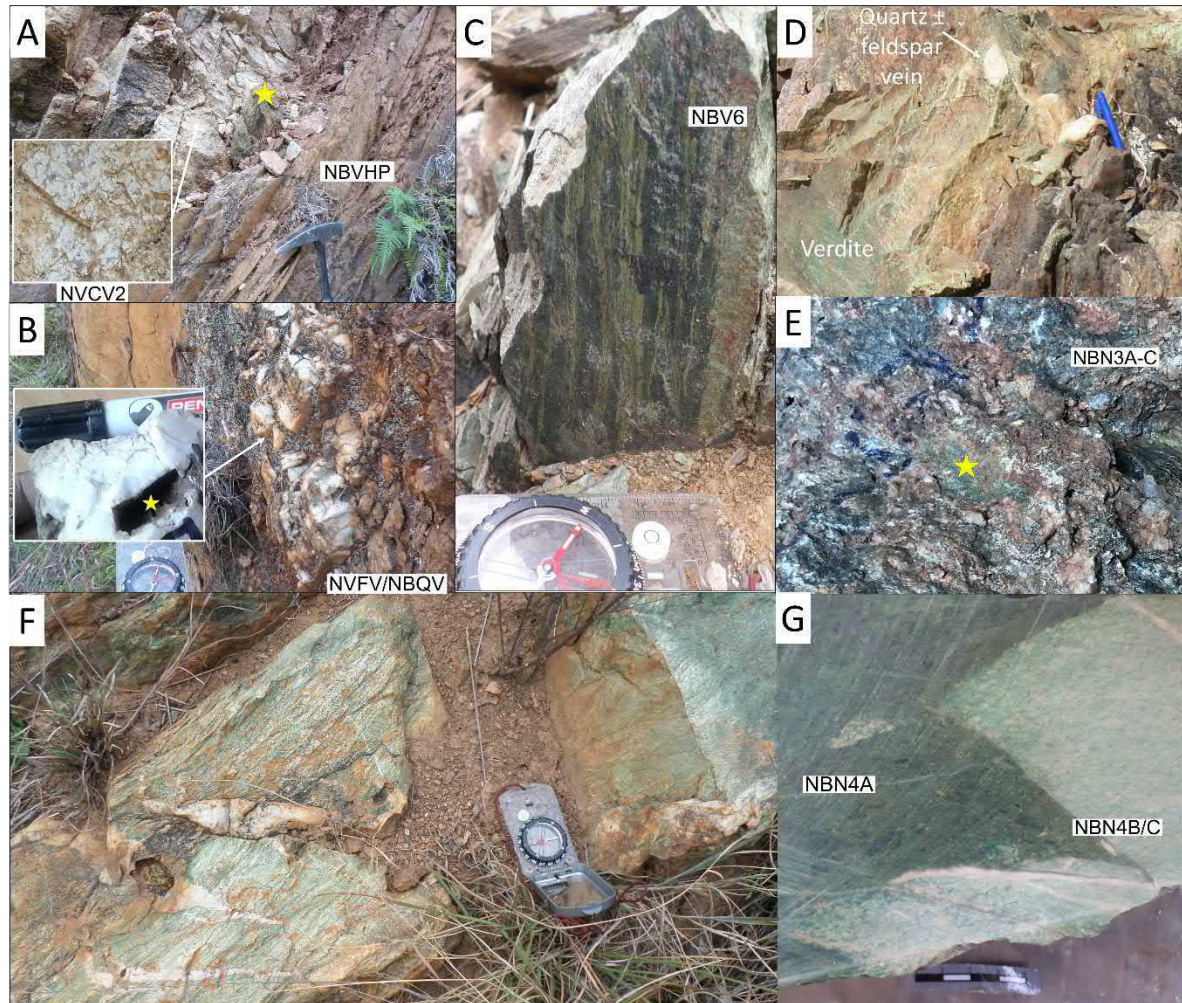


Figure 2.9 Lithologies in Nelspruit verdite quarry A and B (images E and G) labelled with sample names. Fuchsite is used here to describe siliceous verdite. (A): Fuchsite at the interface of the chlorite schist verdite host and a massive, foliation-parallel quartz-feldspar granitoid. (B): Layer-parallel quartz vein with prismatic voids (e.g. yellow star) lined with iron oxides. (C): Sub-vertical verdite sample. (D): One of the common layer-parallel felsic veins interlayered with verdite in the layer-parallel shear zone. (E): Sample of chlorite schist with incipient verdite formation (green in yellow circle) near veins of euhedral K-feldspar and quartz. (F): An apparent clast of banded fuchsite and quartz $\pm$ feldspar (buddstone) in a matrix of dark fuchsite schist. (G): Unsilicified angular clasts of massive fuchsite and sericite aggregates in a dark green fuchsite schist.

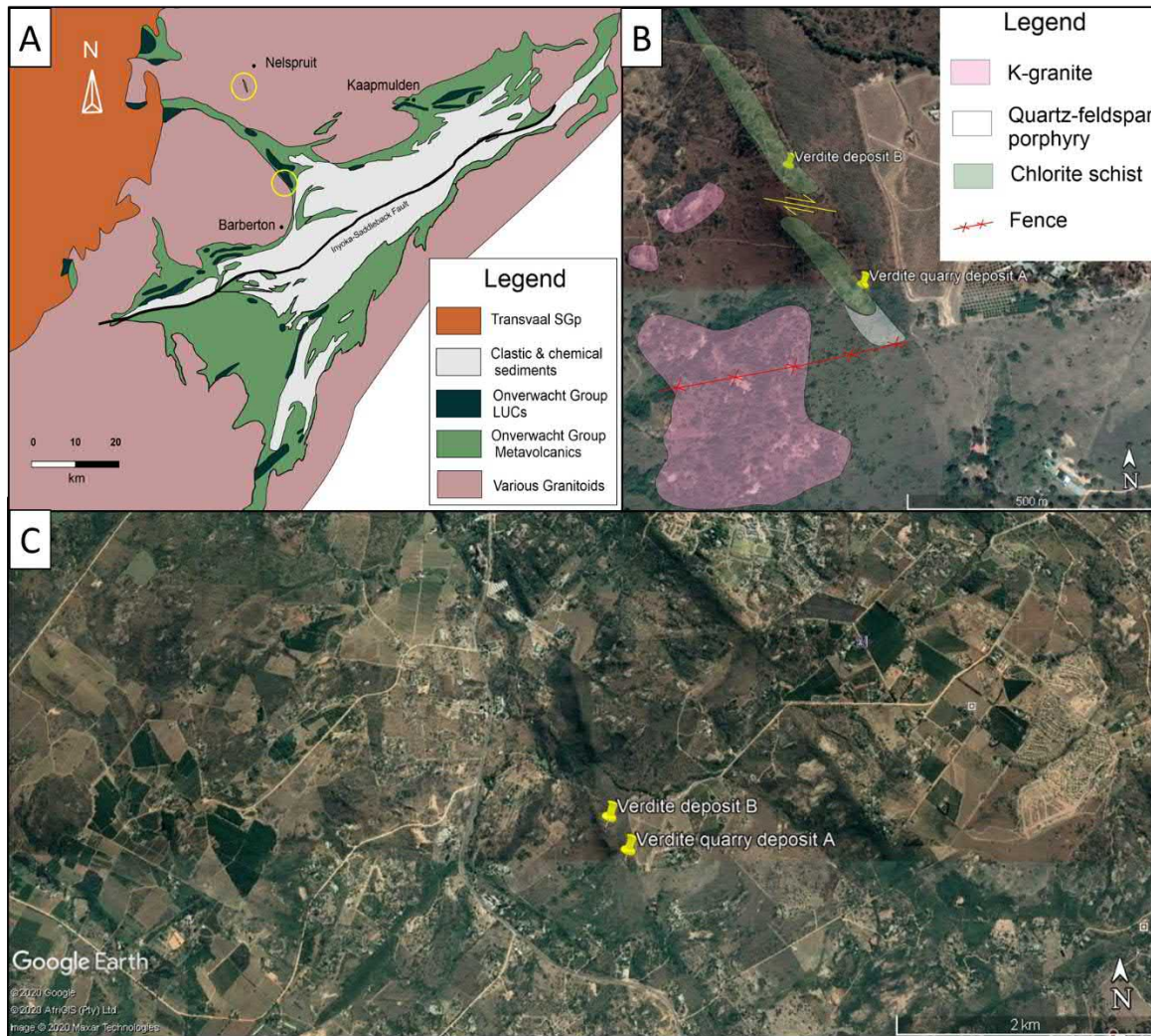


Figure. 2.10 Widely unknown verdite deposits 6 km south of Nelspruit (Quarry A: 26 32 30.36 S; 30 57 52.50 E. Quarry B: 25 32 21.30 S; 30 57 46.26 E). (A): Simplified map of the Barberton Granite Greenstone Belt with two yellow circles marking the context of verdite deposits studied. The southern circle surrounds the Handsup Layered Ultramafic Complex of the BGB, and the northern circle marks the location of a verdite-bearing mafic ridge (quarries A and B) of Onverwacht Group affinity (B): Two verdite deposits sampled during the present study (quarries A and B) are separated by apparent dextral displacement. (C) Wider context of the verdite-bearing ridge marked with locations of verdite quarries reveals a similarly-trending ridge to the east.

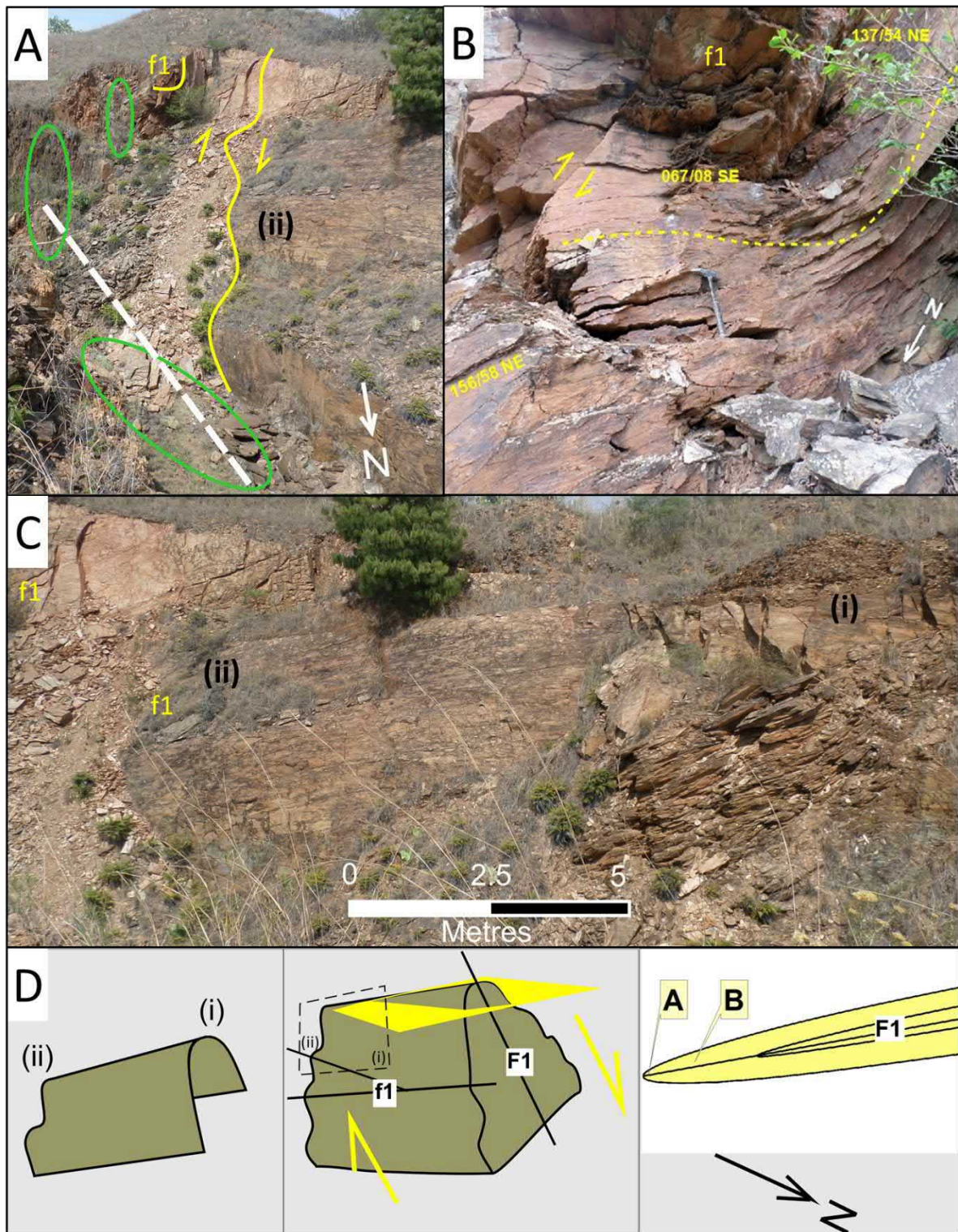


Figure 2.11 Nelspruit verdite quarry A. (A): Verdite lenses (green ovals) occur within a foliation- and strike-parallel northwest-trending shear zone cutting through a chlorite schist. (B): Close up view of metre-scale fold in the quarry wall (f1) in image (A). (C): Close-up image of left section of image (A) shows the merging of f1 folds (ii) into a larger fold (i) towards the right edge of the image. (D): Z-folds on the eastern margin of quarry may (Images A-C) suggest that the verdite-bearing shear zone occurs in the eastern limb of a northwest-trending antiform.

2.4.4 Petrography: Nelspruit mafic ridge

The main difference between the host rocks to the verdite in Handsup Quarry and Nelspruit quarries is the higher iron content in the Nelspruit chlorite schist which also contains a higher proportion of iron oxides and hydroxides at contacts with invading quartz veins. The Nelspruit verdite samples are also dominated by chrome-muscovite and an increasing proportion of preserved subhedral albite from quarry A to quarry B. Albite in this context is mostly sheared into layers. Where unsheared and preserved as distinct crystals, advanced alteration to muscovite is predominant (Fig. 2.12, C-E). The alteration of feldspar and the resorption of andalusite porphyroblasts into a groundmass of muscovite and partially sericitised feldspar appear to be the source of muscovite in the samples. Unlike samples from Handsup, the alteration of chlorite by muscovite-precipitating fluids was not observed in the Nelspruit verdite samples as all samples were completely composed of muscovite-after-feldspar interlayered with or surrounded by chrome-muscovite. As observed in Handsup verdite, euhedral rutile also occurs in association with rarely preserved albite.

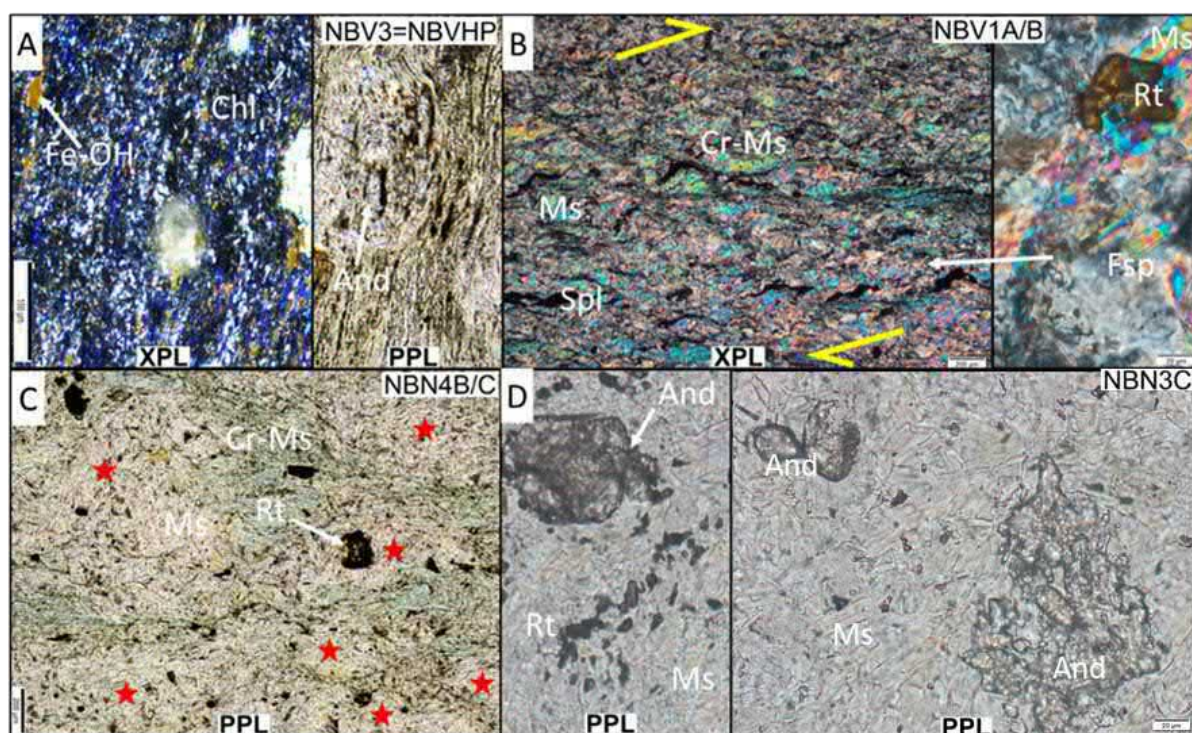


Figure 2.12 Main minerals related to the formation of Nelspruit verdite from quarry A and B. Sample numbers at top right corner PPL and XPL labels are in photomicrographs produced using plane-polarised or cross-polarised light, respectively. (A): Photomicrographs of the Fe-rich chlorite schist verdite host. (B): Dextrally sheared verdite with rarely-preserved feldspar crystals (insert). Quarry B samples show that the angular clasts in a fuchsite matrix (Fig. 2.9 G) are comprised of nodular zones of muscovite aggregates (red stars) in a chrome-muscovite groundmass (C); and that resorbed andalusite occurs in some fuchsite schists of that quarry (D). Fe-OH: iron hydroxides, Chl: chlorite, And: andalusite, Ms: muscovite, Spl: spinel (chromite/magnetite), Cr-Ms: chrome-muscovite, Rt: rutile.

2.4.5 Elemental Maps

In both the Handsup and Nelspruit verdite, the green Cr-muscovite-rich and yellowish rutile-rich mineral layers common in high-quality verdite are detected as bands rich in Cr or Ti respectively. Cr, Fe, Mg, Nb, Y, V and As are homogeneously dispersed throughout samples in fine crystals, and except for Cr, coarser mineral crystals of these elements are concentrated in Ti-rich layers dominated by rutile (Fig 2.13 and 2.14). This suggests that muscovite replaced a phase that was rich in these elements-such as mafic minerals. Only V, Nb and Zr occur directly in rutile crystals, indicating the absence of ilmenite or titanomagnetite in the samples.

In the Handsup sample of true verdite, quartz occurs as frameworks of resorbed crystals mottled with secondary chrome-muscovite (Fig. 2.13). Mg, Ca, Fe, Ni, Al and K are equally distributed in the sample, suggesting potassic alteration of a mafic precursor.

Foliation-parallel quartz lenses in the chlorite schist verdite host of quarry A in Nelspruit (Fig. 2.15) contain irregular spots of Fe intergrown with K and finer spots of Al while the chlorite is rich in Fe, and contains Al while lacking K. This suggests that the quartz vein contained some iron-bearing potassic feldspar, and that the chlorite host may be the iron-rich, aluminous variety i.e. chamosite: $(\text{Fe}_5\text{Al})(\text{AlSi}_3)\text{O}_{10}(\text{OH})_8$. Within verdite from quarry A, no free quartz was observed. Instead, well-distributed Si is associated with Al and K, confirming the presence of muscovite-which dominates the sample.

The partially verditised sample NBN3C of quarry B in Nelspruit shows K-rich nodules intergrown with Si-rich zones corresponding to muscovite after feldspar (Fig. 2.15). The altered feldspar is surrounded by a chrome-muscovite groundmass rich in Al, Cr, Ti, Nb, indicating a mafic precursor-presumably the chlorite schist.

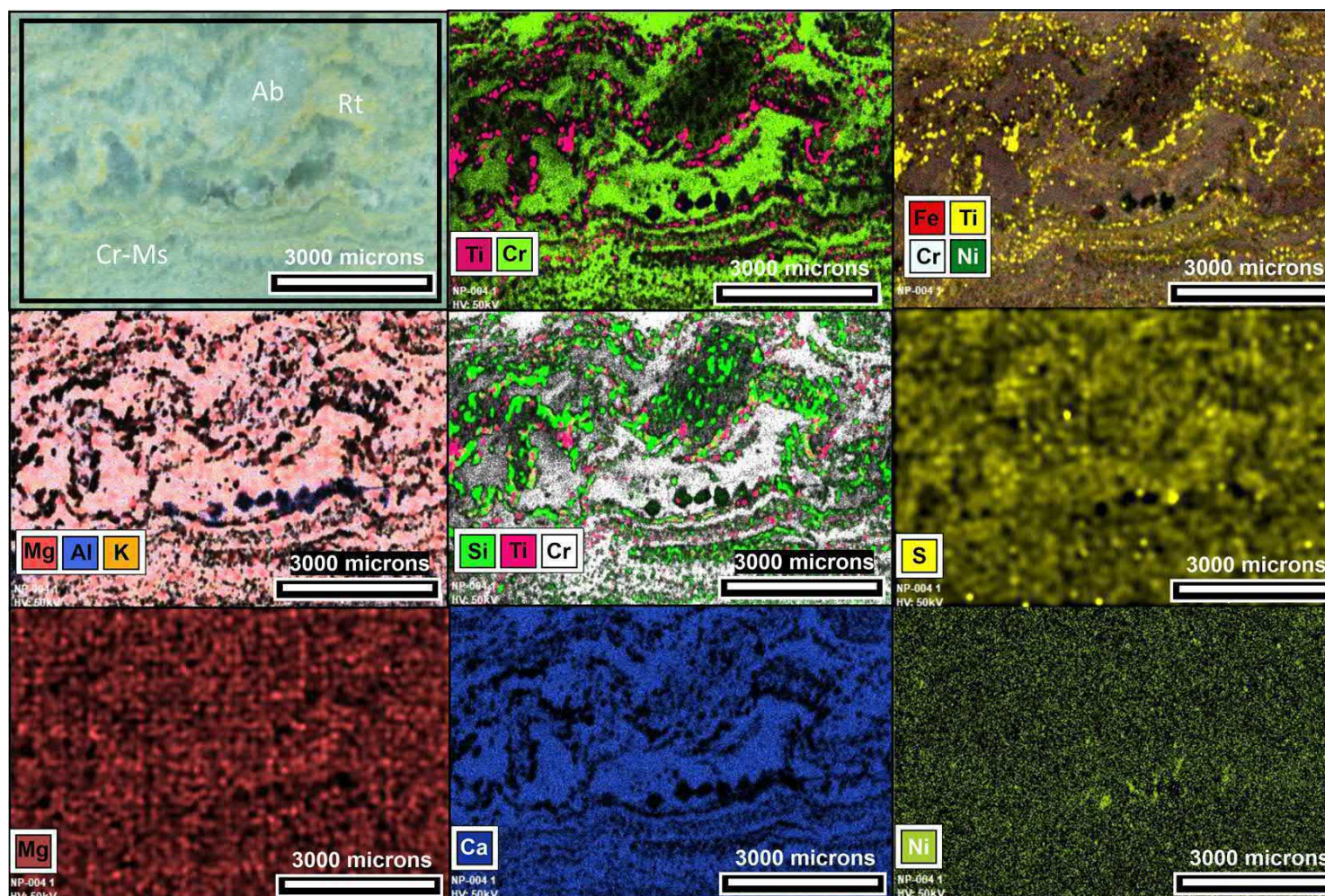


Figure 2.13 Distribution maps of selected elements within a compositionally layered verdite sample from the Barberton Handsup Layered Ultramafic Complex (sample NP-004). Mapped area of thin section offcut in the top left corner is marked by a black rectangle. Cr is less abundant in Fe-rich layers which are associated with immobile elements such as Al, V, Nb, Ti. In addition to occurring as distinct clusters, Cr and Ni, are homogeneously distributed throughout the sample in fine opaque minerals and in altered serpentine. No layering can be seen in the distribution of Mg, Ca and S which occupy the Cr-muscovite $\pm$ chlorite $\pm$ serpentine groundmass.

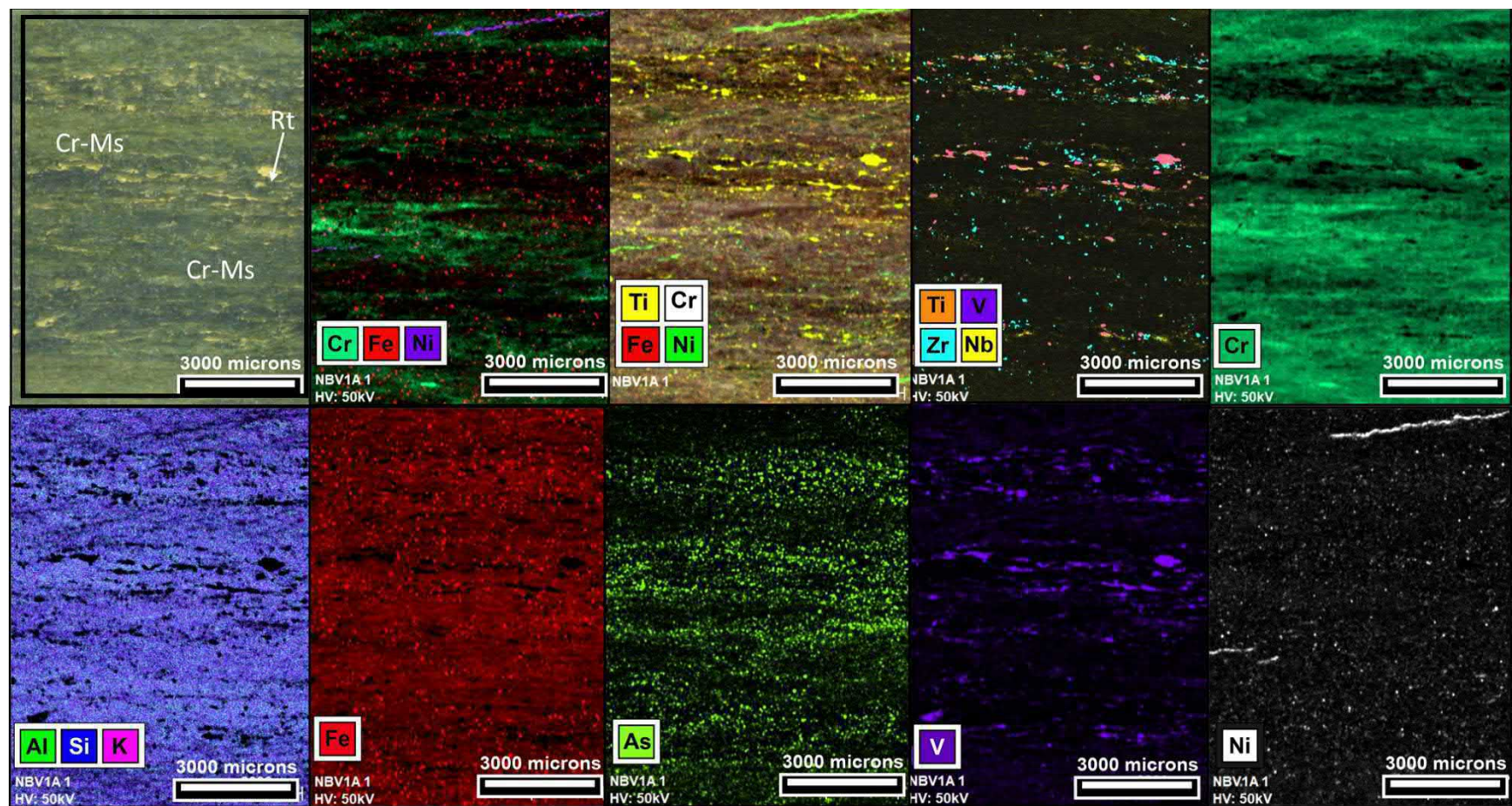


Figure 2.14 Distribution maps of selected elements within a compositionally layered verdite sample from quarry A of the Nelspruit mafic ridge (sample NBV1/N1V1). Mapped area of thin section offcut in the top left corner is marked by a black rectangle. Compositional layering of chrome-muscovite and rutile correspond to layers rich in Cr and those rich in Fe and Ti. Rutile-rich layers are associated with immobile elements such as Al, V, Zr and Nb. Ni is concentrated in late, foliation-discordant iron-bearing veins and in clusters of fine iron-bearing opaque minerals.

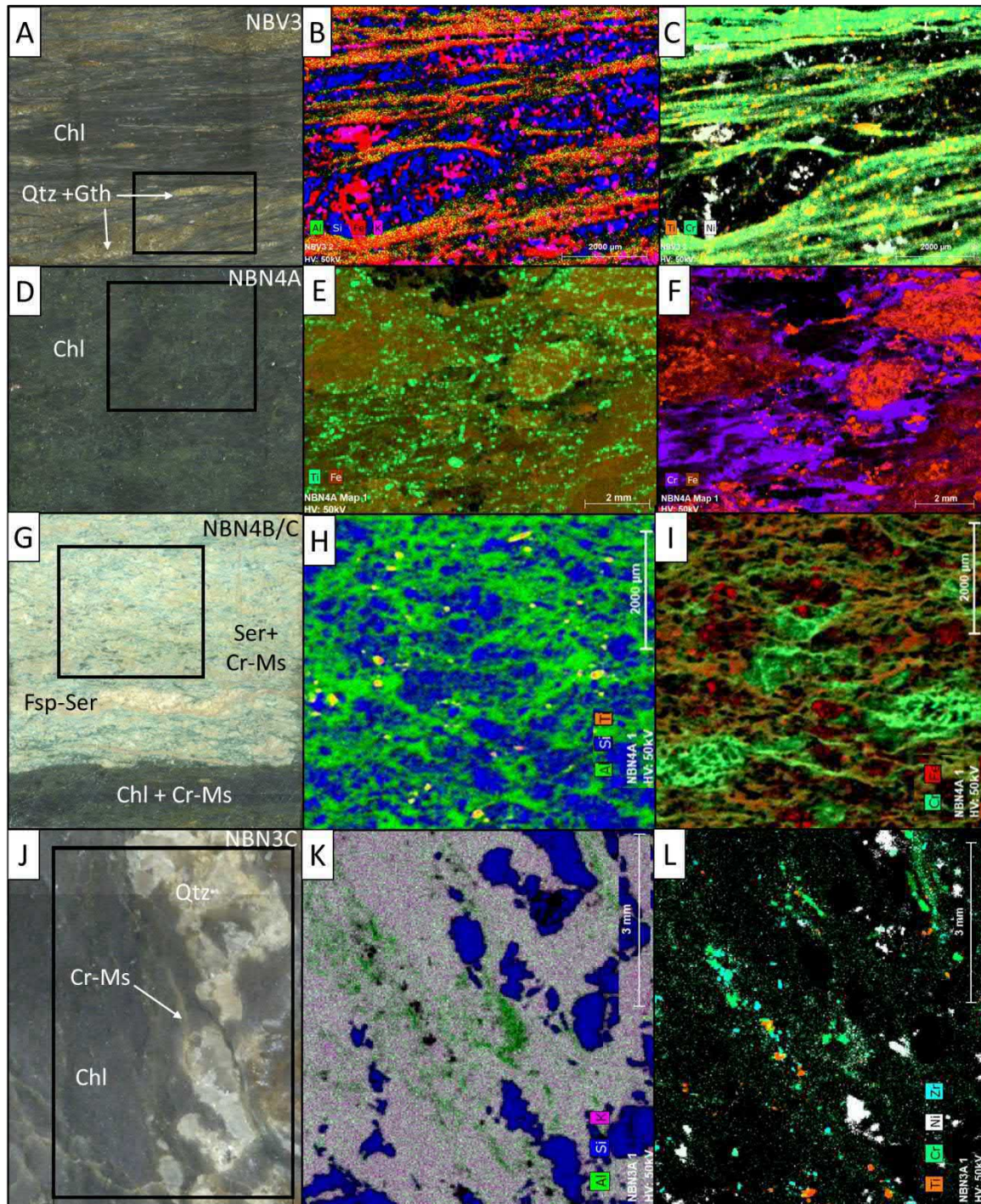


Figure 2.15 Distribution maps of selected elements within key lithologies of the Nelspruit mafic ridge, sample names at top right corners. (A-C): Fe-rich chlorite host of quarry A verdite. (D-F): Dark green verdite from quarry B. This forms the groundmass component of a cohesive fault rock containing angular blocks of sericite-chrome muscovite schist (G-I). (J-L): Chrome muscovite containing high abundances of immobile Al, Cr, Zr and Ti was intruded by a vein comprised of quartz and crystals replaced by chrome-poor muscovite aggregates rich in K, Al and Si. Mapped areas of thin section offcuts are marked by a black rectangle in (A), (D), (G) and (J).

2.4.6 Phase maps

TIMA phase maps (Fig. 2.16) of a representative sample of nodular verdite confirm the association of albite crystals with rutile aggregates. They reveal the presence of sodium-titanium silicates associated with the albite-rutile assemblage that is interlayered with- or forms nodules in a groundmass of Cr-muscovite. Blebs of kaolinite occur within Cr-muscovite-rich zone where highly sericitised albite occurs.

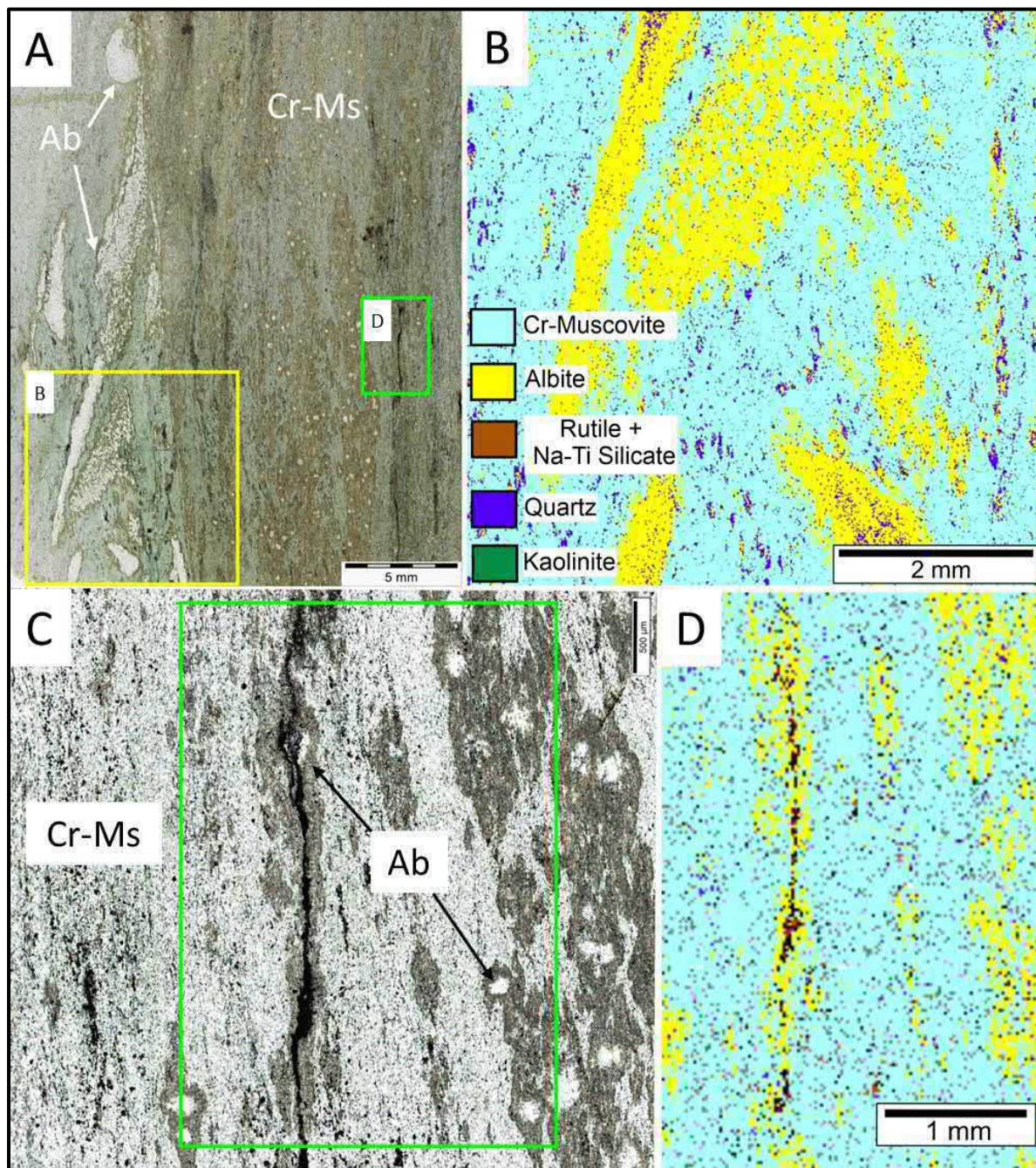


Figure 2.16 Phase maps of sample NP-008. (A): PPL photomicrograph marked with borders that mark mapped areas shown in (B) and (D). (B): Albite blebs resorbed into a groundmass dominated by Cr-muscovite. (C): Close-up view of the left-central sector of (A) shows a kinked vein of rutile and Na-Ti silicates that overgrew what may have been a kinked albite vein or a train of albite crystals similar to that described in the petrography section.

2.4.7 Chemical Data

Barberton and Nelspruit verdite contain higher K and Al than chlorite host rocks whereas chromium, magnesium and iron show the opposite trend.

Table 2.1 Compositions of verdite samples and selected host rocks.

XRF Anhydrous Bulk Composition (this study)				
wt%	BGB Verdite (NP-004)	BGB Verdite Host (HUTH)	Nelspruit Verdite (NBV6)	Nelspruit Verdite Host (NBVHP)
SiO₂	61,72	38,24	50,24	59,49
Al₂O₃	27,47	20,95	33,72	11,99
Fe₂O₃	0,05	4,57	1,06	12,08
MnO	0,01	0,08	0,01	0,07
MgO	0,36	31,18	2,11	14,92
CaO	0,01	0,78	-0,02*	0,15
Na₂O	0,55	1,19	0,16	0,00
K₂O	7,74	0,01	11,64	0,02
TiO₂	0,86	0,83	0,77	0,68
P₂O₅	0,01	0,01	0,03	0,10
Cr₂O₃	1,20	1,96	0,28	0,38
NiO	0,00	0,20	0,01	0,11
EPMA Analyses of Cr-Muscovites (Harding and Jobbins, 1984)				
wt. %	Verdite from eastern RSA mine n=4	Barberton verdite n=7	Zim verdite n=10	Zim ruby-verdite n=3
SiO₂	45.08	47.42	45.52	45.14
TiO₂	0.39	0.42	0.23	0.34
Al₂O₃	30.89	31.78	35.16	35.10
Fe₂O₃	0.33	0.20	<0.2	<0.2
MgO	1.72	1.61	0.37	<0.2
Na₂O	0.43	0.59	0.91	0.68
K₂O	10.36	10.88	10.23	10.19
Cr₂O₃	4.20	1.86	1.23	2.60
Total:	93.40	94.76	93.65	94.05

*Concentration regarded by lab technician as 0 wt% oxide.

In the Barberton Handsup Layered Ultramafic Complex context, the chlorite schist host rock has lower Si, Al, K and Ti abundances than verdite samples. Except for Si, the same distribution of elements was observed between the Nelspruit verdite and host rock. Though Ti is more abundant in verdite, the enrichment is relatively minor (<1 wt %). The higher Al content of verdite may be a result of residual Al in Al-poor variolitic basaltic komatiites together with magmatic or hydrothermal inputs indicated by various generations of albite veins and muscovite veins described in the petrography section above.

The high Al content of verdite samples of verdite differentiates samples from common green alteration products of mafic and ultramafic lithologies (Fig. 2. 19). These include products of calcium metasomatism to produce rodingite/hibschite assemblages dominated by hydrogrossular and diopside (e.g., Anhaeusser, 1979); silica-carbonate alteration to produce lisvenite/listwanite assemblages of Mg-carbonate and Cr-Ni micas (Kashkai and Allakhverdiev, 1965); and K-Si metasomatism to produce fuchsitic schists and cherts (e.g. Duchac *et al.*, 1987). In a ternary plot of Al-Fe+Ti-Mg oxides, verdite is more closely associated with K-Si metasomatised BGB komatiites analysed by Duchac *et al.* (1987).

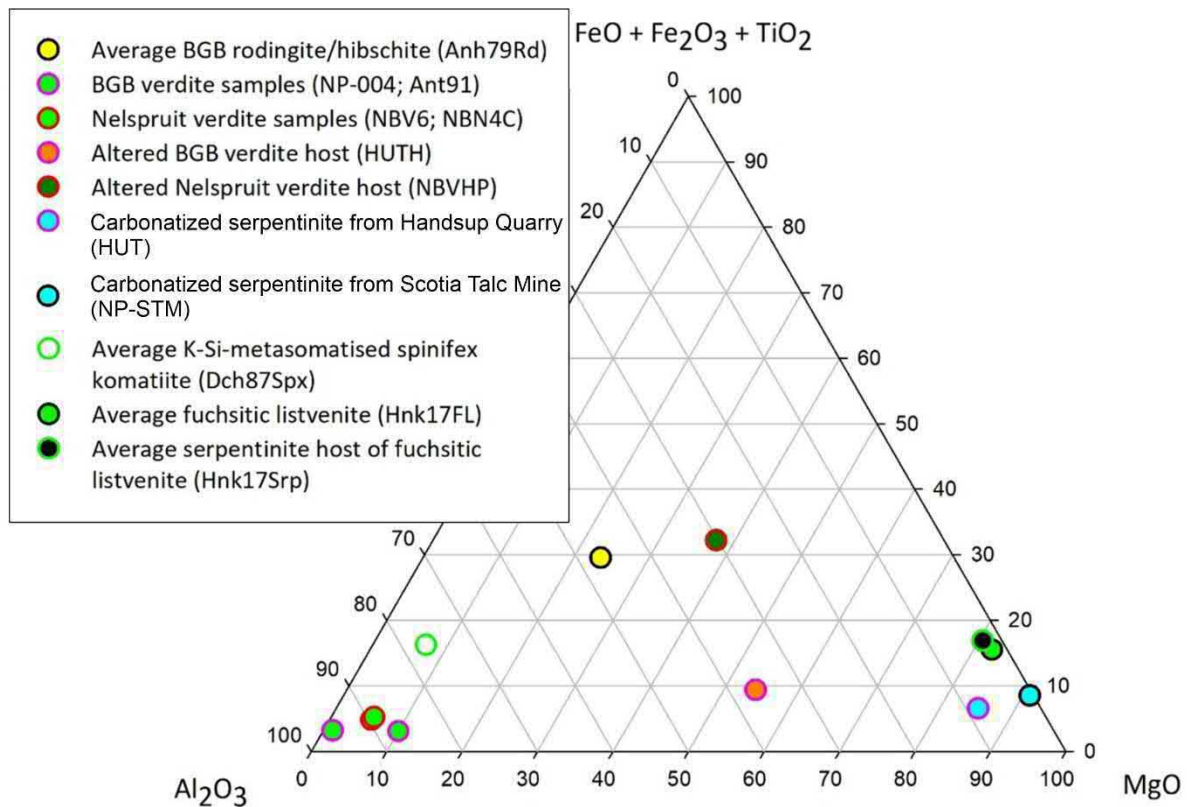


Figure 2.17 Modified Jensen (1976) Al-Fe+Ti-Mg oxide plot of verdite samples plotted for comparison with direct host lithologies and a variety of alteration products of ultramafic rocks mainly from the Barberton Greenstone Belt. Sources and chemical analysis data listed in Table 1 of Appendix B, Table 1.

According to the classification diagram of Winchester and Floyd (1977) for altered lithologies, the direct verdite host rocks studied are both magnesium-rich basaltic lithologies (Fig. 2.18). A ternary plot of Al-Fe+Ti-Mg oxides populated with compositions of verdite host rocks and other BGB lithologies (Fig. 2.19) indicates that the two verdite bearing chlorite schists had compositions that were close to that of komatiitic basalts.

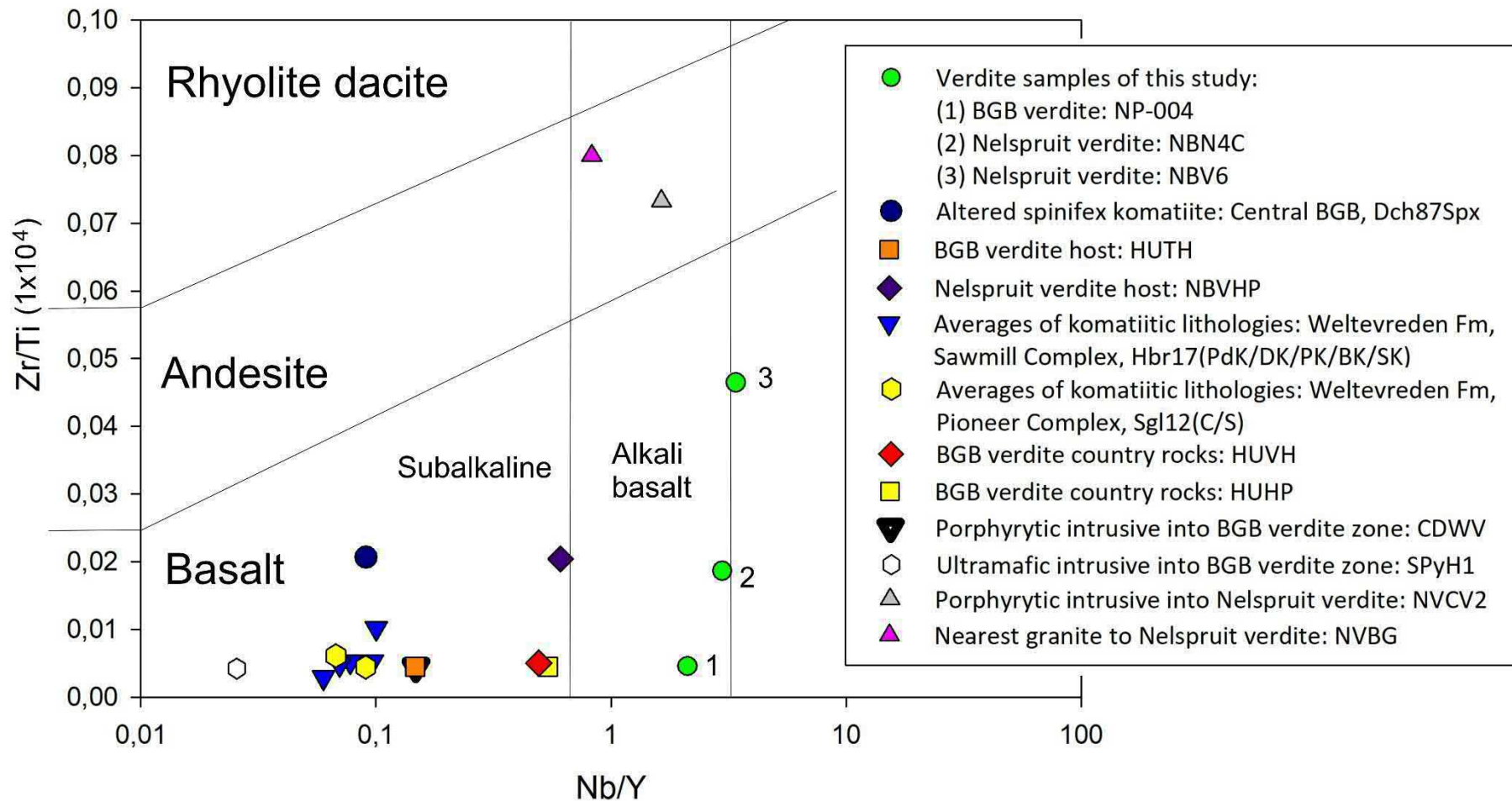


Figure 2.18 Classification diagram for altered lithologies from Winchester and Floyd (1977), modified after Agangi *et al.* (2018). Relative abundances of immobile elements within verdite samples suggest a protolith of a basaltic composition. Various ultramafic lithologies of the Weltevreden Fm exposed in less deformed, less altered layered ultramafic verdite samples the basalt subcategory as verdite host rocks. Sources and chemical analysis data listed in Appendix B, Table 3. Sample codes and sources are described in Appendix B, Table 1.

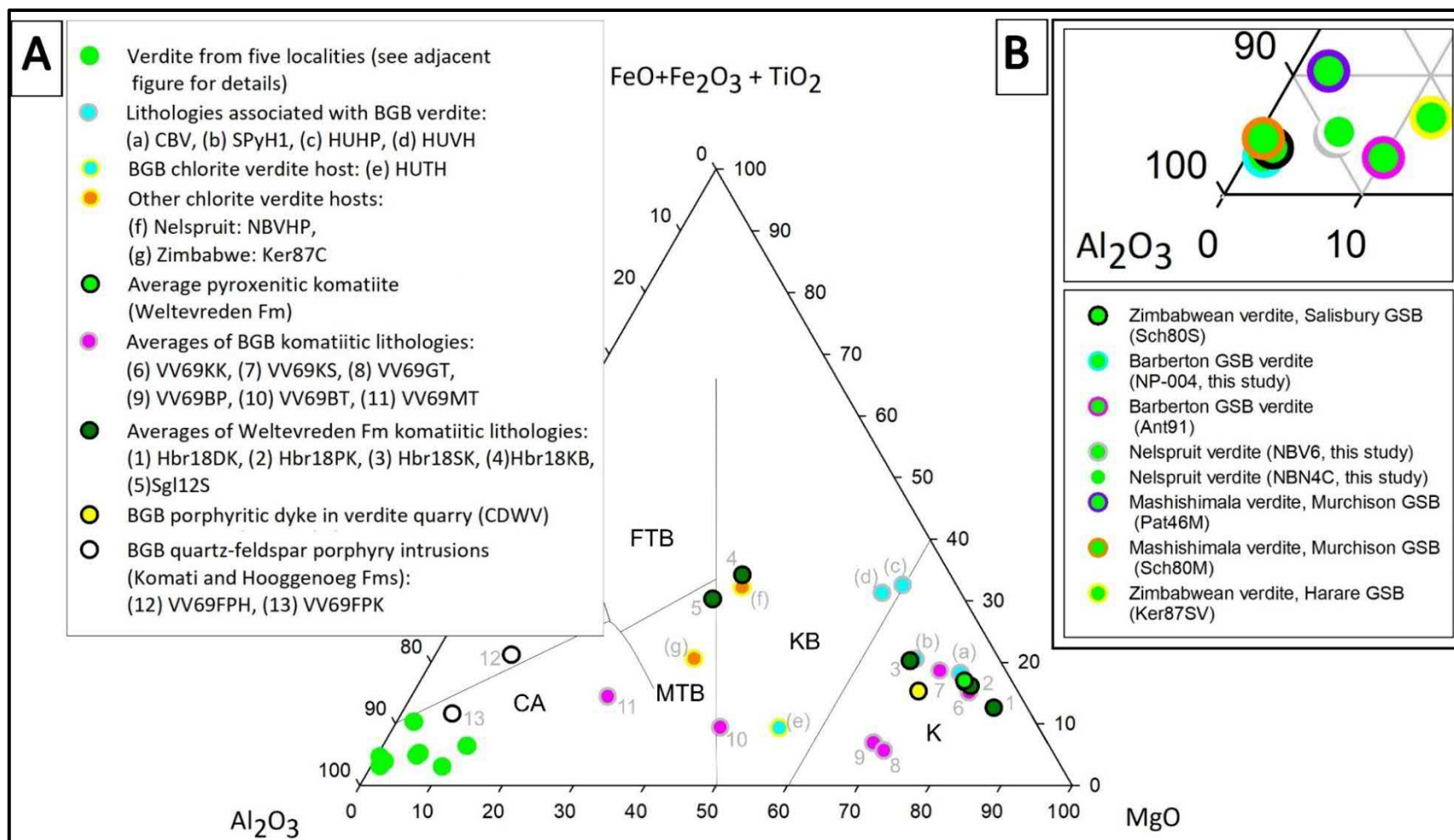


Figure 2.19 Modified Jensen (1976) Al-Fe+Ti-Mg oxide plots with lithological zones K: Komatiite, KB: Komatiitic basalt, MTB: Magnesium-rich tholeiitic basalt, FTB: Iron-rich tholeiitic basalt, CA: Calc-alkaline lithologies generalised from Arndt (1975), Naldrett and Turner (1977), Arndt *et al.* (1977) and Nesbitt *et al.* (1979). (A): Lithologies associated with BGB and Nelspruit verdite occurrences. Ultramafic Onverwacht lithologies from the northern (1-5) and southern (6-11) BGB are plotted for comparison with lithological suites of BGB and Nelspruit verdite. (B) Detailed view of verdite cluster from (A). Sources and chemical analysis data listed in Appendix B, Table 1.

2.5 Discussion

Serpentinites at the margins of the only remaining verdite-bearing zone in the Handsup Quarry were subjected to silicification along fractures. This produced discordant veins of chrysotile asbestos. Pervasive silicification allowed ubiquitous serpentine replacement by talc through steatitisation of a serpentinite situated in the eastern flanks of the verdite-bearing shear zone. The same steatitised serpentinite was later pervasively carbonated. However, carbonates were not observed in verdite samples studied.

The occurrence of sub-vertical verdite lenses in a feldspar-bearing serpentinite altered to a chlorite-epidote-sericite schist was initially understood to indicate a gabbroic verdite protolith (Anhaeusser, 1969; Groeneveld, 1975). However, the petrographic features of the feldspar in verdite may indicate an original komatiitic basalt that contained feldspathic varioles which, according to Arndt and Fowler (2004), may occur as spherical, sheath or comb feldspar structures. The variolitic komatiitic basalt would have been albitised, chloritized, later fractured and infilled by fluids that precipitated feldspar and later muscovite. Subsequent saussuritisation and sericitisation of albite would have produced epidote and mica that in the presence of chromium from serpentine, would become a chrome-muscovite schist, viz. verdite.

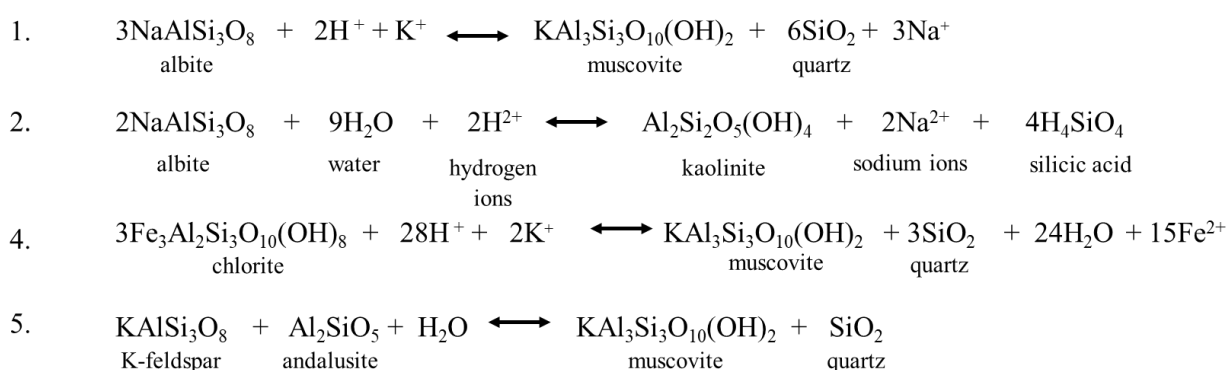
Additionally, field observations show that the remaining verdite lenses in the Handsup quarry occur at the interface of albite-filled tension gashes and a chloritised serpentinite that contains saussuritised and sericitised ovoid albite aggregates and mm- to dm-scale albite $\pm$ quartz veins. This may indicate that the verdite precursor was a serpentinised variolitic komatiitic basalt that was chloritised, extended and in-filled by albite-bearing magmatic or hydrothermal fluids (e.g. dyke CDWV in Fig. 2.6). Later extension and infilling by potassic fluids produced muscovite (Fig. 2.8 H) and may have caused the sericitisation of albite in the chlorite schist. Muscovite formation was followed by later shearing, that appears to have promoted the incorporation of chromium into muscovite (Fig. 2.8 C). This suggestion is supported by samples with incomplete “verditisation” (NP-003, NP-006). They show that chrome-muscovite developed in sheared sectors of the same spinel-bearing, muscovite-after-chlorite schist (Fig. 2.8 C), and at margins of kinked and sheared albite masses (Fig. 2.8 I).

To produce verdite, the albite-infilled, serpentinised, variolitic komatiitic basalt was sheared and metasomatized (e.g. Reactions 1-5 below) during separate deformation events. The earlier deformation event produced sub-vertical tension gashes that plunge to the northeast. A later deformation event produced sub-vertical tension gashes that plunge to the southwest. The

selective occurrence of verdite in southwest-plunging tension gashes filled with feldspar porphyry appears to be a function of proximity to mafic-ultramafic lithologies (Fig. 2.6). This may have been combined with being favourably oriented to remain open during a subsequent deformation event that may have kept the northeast-plunging fractures closed.

In the Nelspruit context, verdite also occurs in chlorite-muscovite schists that contain feldspar but only as highly sericitised aggregates. In addition to feldspar, isolated crystals and clusters of euhedral andalusite are resorbed into a groundmass of sericitised feldspar and oikocrystic chrome-muscovite, presumably after chlorite. Within the verdite-bearing shear zone, the chlorite and feldspar were respectively metasomatised to chrome-muscovite and chrome-poor muscovite. This suggests that Nelspruit verdite may have been a gabbro or variolitic basalt that was mylonitised into the chrome-rich and chrome-poor layers observed in hand specimens and in elemental maps (Fig. 2.9 C). Later brittle deformation at margins of the verdite-bearing shear zone produced cohesive fault rocks comprised of angular blocks of mylonitised and metasomatised gabbro (buddstone?) or variolitic komatiitic basalt enveloped by fuchsitic fault gouge (Fig. 2.9 F-G).

Contact metamorphism associated with the intrusion of the leucogranite/coarse quartz-feldspar porphyry in quarry A and the potassic granite in quarry B caused the growth of andalusite porphyroblasts in the Cr-muscovite-sericitised feldspar schist. Retrograde metamorphism caused the replacement of andalusite and potassic feldspar by muscovite $\pm$ quartz (Fig. 2.12 D) according to reaction 5 below (Winter, 2010).



The verdite-bearing mafic ridge in Nelspruit is extensive and lies parallel to other local lineaments that occupy an area tens of kilometres north of Nelspruit (see Olsson *et al.*, 2010). This contrasts with the xenolithic identity of the Nelspruit verdite host that was suggested by Antenen (1991) and may indicate that the ridge is a dyke through the batholith. However, the observed sub-horizontal folding (Fig. 2.11) and strong schistosity (Fig. 2.9 A) of the chlorite

schist adjacent to an essentially massive leucogranite (coarse quartz-feldspar porphyry?) suggests that the chlorite schist is either older (and is a host to the leucogranite) or that the schist is younger, and strain was partitioned into it due to lower competency relative to the leucogranite. The advanced alteration of the leucogranite to soil prevents an assessment of field relationships beyond the quarry. Therefore, the current data cannot confirm if the mafic ridge forms part of the locally exposed 2.9 Ga Badplaas dyke swarm. However, the andalusite and quartz ($\pm$ feldspar) veins in the altered chlorite schist suggests that the gabbroic or variolitic basalt protolith of the chlorite schist may have been invaded by felsic fluids and later thermally metamorphosed by the intrusion of the 3.1.-2.7 Ga phases of the Nelspruit Batholith. In this way, the mafic ridge may form part of the 2.9 Ga Badplaas dyke swarm, or it may be an extension of the NW-trending Jamestown Schist Belt as the BGB is understood to extend northwards beneath the surface cover of the Nelspruit and Pigg's Peak batholiths (De Beer, 1988).

2.6 Conclusion

In both the Handsup and Nelspruit verdite quarries, the incomplete replacement of chlorite schists by muscovite $\pm$ feldspar sericitisation is associated with foliation-parallel dilational structures within foliation-parallel ductile shear zones later deformed in a brittle manner. In both quarries, verdite occurs in chlorite schists at the interface with felsic lithologies: a quartz-feldspar porphyry dyke in the Handsup quarry and a massive leucogranite (coarse quartz-feldspar porphyry?) in Nelspruit. The felsic lithologies adjacent to the verdite-bearing chlorite schists in both quarries are advancedly sericitised with rare occurrences of feldspar phenocrysts. Through potassic metasomatism, both phases were altered to muscovite, and later to chrome-muscovite. The incorporation of chromium depended on the proximity to a chromium-bearing source like sheared chromite crystal aggregates (Fig. 2.8. C: top left segment of NP-003) or the chlorite schist as it is presumably an alteration product of a chrome-bearing mafic mineral (Fig. 2.8, I: NP-006 with Cr-muscovite at albite-chlorite interfaces). This is supported by the presence of chrome-poor muscovite adjacent to sheared, vivid green chrome muscovite in some samples (samples NP-007, NP-003; NBN4A).

The scarcity of in-situ verdite-bearing zones limits these interpretations of lithological and alteration associations to outcrops observed. Therefore, these interpretations may only be one aspect of the verdite-formation process.

Chapter 3: Crenulated Phyllite

Parking: 25°40'24.75"S; 31° 5'21.26"E; Representative outcrop: 25°40'24.12"S; 31° 5'23.82"E

3.1 Site description

This is a roadside exposure of a sheared mafic phyllite just south of the confluence of the Noordkaap and Suidkaap Rivers to form the Kaap River (Fig. 3.1 B). Just beneath the northern edge of the road is a railway line that used to transport goods between Pretoria and Mozambique from the late 1890s (Bornman, 2002). This railway runs almost parallel to a section of the R38 road built between 1884 and 1927 to avoid crossing the Kaap River multiple times between Site 2 and Site 5 (Bornman, 2002).

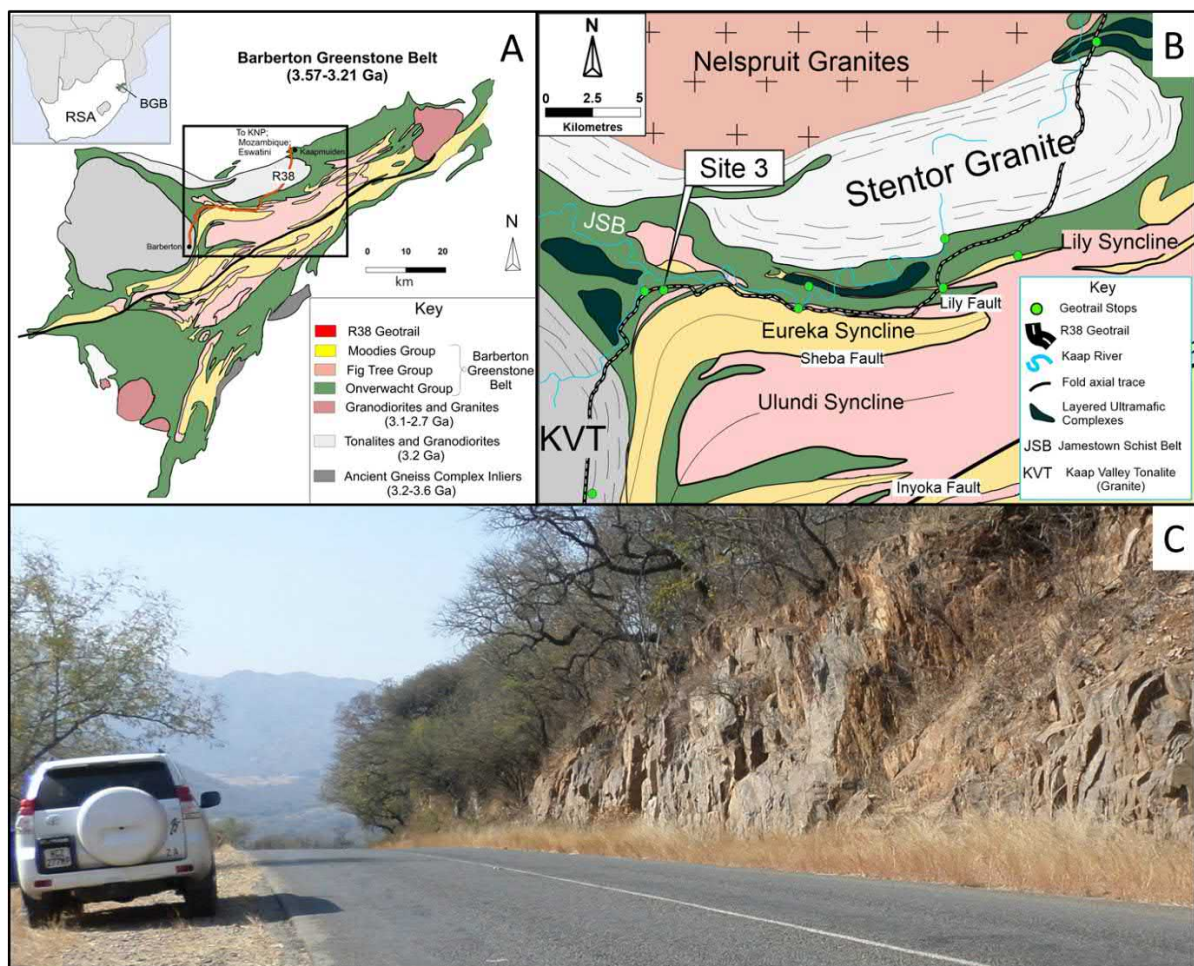


Figure 3.1 Main features of Site 3. (A): Simplified geological map of the R38 Geotrail in the context of the BGB. (B): a detailed view of the black rectangle in (A) shows the location of Site 3 in the northern edge of Eureka Syncline. (C) View of site 3 taken from the west. Maps modified after Visser *et al.* (1956) and Anhaeusser *et al.* (1981).

3.2 Aspects of Site 3 that may interest laypersons

To history enthusiasts, the appeal of this site would be the view of the Kaap River and Noordkaap. According to Bornman (2002), a confluence of tributaries of the Kaap River was the site of an alluvial gold discovery (1882) that started the Barberton gold rush. Consequent to the gold rush, a number of mining towns such as Noordkaap were soon established near mining sites (e.g. Fig. 3.2).



Figure 3.2 Kaap River gold mining during the 19th century. Image sourced from the Barberton Museum archives.

Part of the Kaap River was a precolonial divide between Swati tribes in the south and Pedi tribes in the north (Bornman, 2002). Having settled in Barberton from Eswatini during the late 1860s (Myburgh, 1949), Swati tribes fought fierce battles with Pedi tribes of the region, eventually causing their northwards retreat to areas north of the Kaap and Crocodile Rivers. In the present day, only stonewalls (Liebenburg, 2016) and beacon names make reference to Pedi tribes who once occupied the region (Delius, 2007). One such beacon is the Three Sisters Mountain east of the settlement near Low's Creek. This mountain is locally known as iMbayana (Myburgh, 1949) or Mbayiyane (Delius, 2007), referencing the Pai Pedi tribe displaced by Swati regiments following a fierce battle near the mountain.

3.3 Geology

3.3.1 Background

3.3.1.1 Lithology

The crenulated surfaces of Site 3 outcrops (Fig. 3.1, 3.3) were described as talc-chlorite or quartz-sericite schists folded sub-vertically to produce shallow-plunging microfold axes of an east-west trend (Ramsay, 1963; Anhaeusser and Viljoen, 1965; Anhaeusser, 1969). The sub-vertical shortening deformation that produced crenulations at Site 3 was ascribed to a late deformational event ($D_{3/4}$) that postdates major structures and sub-horizontal shortening of the BGB (Ramsay, 1963; Viljoen, 1963; Anhaeusser, 1969, 1976). Other structures formed during this late-stage sub-vertical shortening include sub-horizontal chevron and conjugate folds found in Moodies and Fig Tree lithologies (Anhaeusser, 1969, 1976).

Crenulations such as those observed at Site 3 (Fig. 3.1) form from the folding of an existing planar fabric (Ramsay, 1963; Passchier and Trouw, 2005; Winter, 2010). This fabric may be schistosity or cleavage shortened by subsequent compressional deformation. In the case of Site 3, a northeast-trending schistose fabric was sheared sub-vertically to generate D_4 sub-horizontal crenulation lineations (Fig. 3.4, 3.5) (Ramsay, 1963; Viljoen, 1963). This, according to Anhaeusser (1976) occurred after the refolding of the northeast-trending Eureka Syncline about a sub-vertical northwest-trending fold axial plane (D_3).

3.3.1.2 Structural context

A high-angle thrust fault was identified along the length of the road at this locality (Viljoen, 1963; Ramsay, 1963). According to cross-sections A-B and C-D of Figure 2 in Viljoen (1963), this thrust fault separates a hanging wall comprised of upper Onverwacht, Fig Tree and Moodies lithologies, from a footwall of the lower Onverwacht Group. The site of interest along the R38 road at Site 3 exposes the lower part of the hanging wall where some phyllites of the Onverwacht Group are preserved.

This fault was understood by previous workers to be part of a thrust zone transporting the Barberton Greenstone Belt towards the northwest in response to compressional stress mainly from the southern and southeastern margins of the greenstone belt (Roering, 1965; De Ronde and De Wit, 1994). Later work by Anhaeusser (1975, 1984) highlighted the key role of the forceful, shallow emplacement of the Kaap Valley Tonalite in facilitating sub-horizontal north-westward transport of the Eureka Syncline.

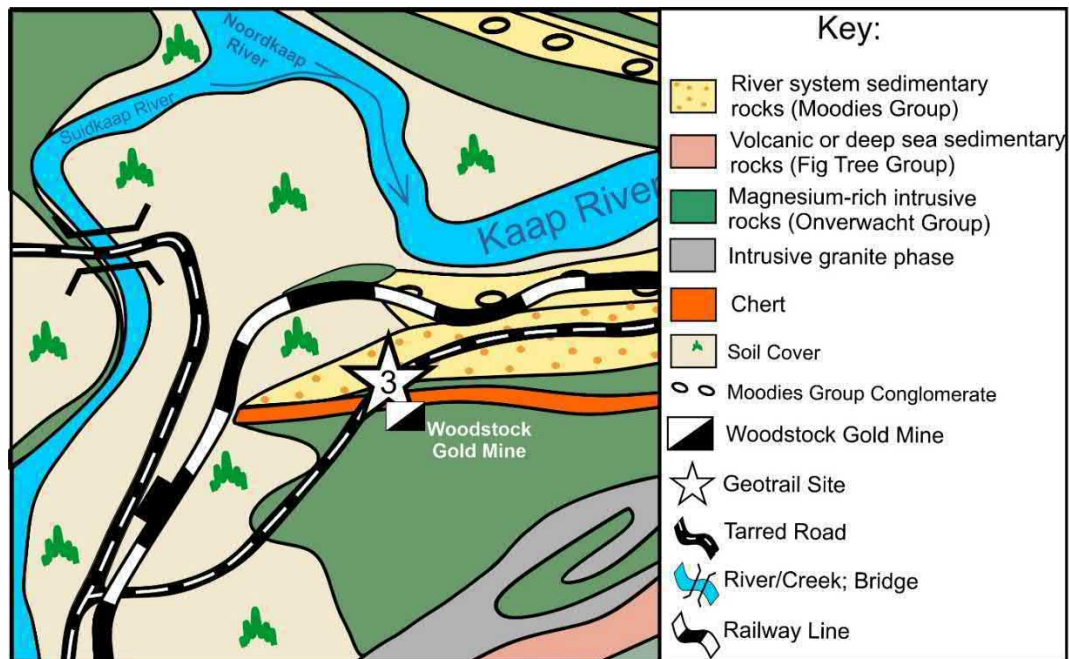


Figure 3.3 Lithological context of the mafic schist at Site 3.

3.3.2 Present study

3.3.2.1 Research question

The focus of previous work at this locality was the structural fabrics linked to crenulation cleavage. This was accompanied by a brief statement on the composition of the phyllite/schists of this locality. Little is known about the veins and mineral composition of the phyllite exposed at Site 3.

3.3.2.2 Aim

To add to existing descriptions of the phyllite exposed at site 3.

3.3.2.3 Objective

To produce thin section descriptions and supplement them with element distribution maps of the crenulated phyllite.

3.3.2.4 Methods

The variety of vein and microfold orientations was investigated and structural measurements taken over a 20 m outcrop area where crenulation cleavage is best developed. Measurements taken were those of microfold axes and schistosity. A representative sample was taken to study the composition and distribution of minerals and elements within the phyllite (NP-212). The sample was cut into a thin section that exposes the ZX plane of the applicable strain ellipse and the mineral composition was described using a petrographic microscope. Element distribution maps were generated using non-destructive XRF spectrometry to analyse millimetre-scale areas of the sample.

3.4 Results

3.4.1 Field observations

Crenulated outcrops of this locality display mm-scale microfolds on a foliation that trends northeast-southwest. They have fold axes plunging sub-horizontally towards the east and northeast. These sub-horizontal fold axes are best developed at contacts between the crenulated phyllite and sub-vertical quartz veins that were crenulated with the phyllite.

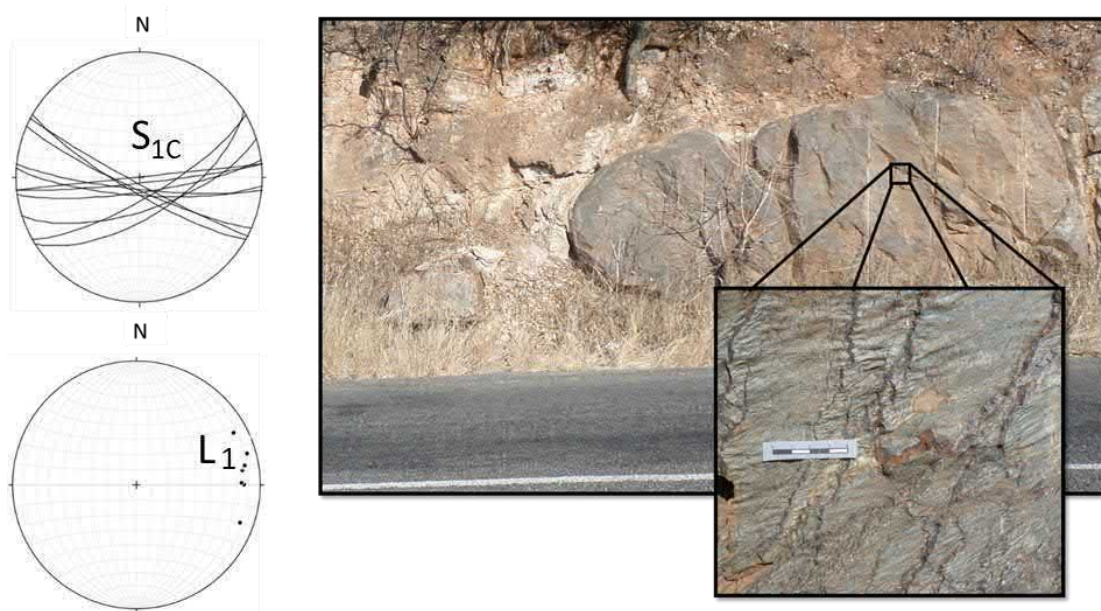


Figure 3.4 Equal-area stereonet plots of schistosity (S_{1c}) and crenulation lineation measurements (L_1) from a sheared phyllite invaded by black quartz veins that accentuate the crenulation fabric.

A detailed look at Site 3 outcrops in the XZ plane of the sub-horizontal microfolds mentioned above (i.e. one of the west-facing planes that cut through the fold profiles of the microfolds which have sub-horizontal fold axes) reveals a later set of folds. These are seen more easily in carbonate and black quartz veins that cut across the earlier crenulation at a high angle (e.g. Dark vein in Fig. 3.5). These later veins dip steeply towards the west and bear no evidence of prior sub-vertical shortening associated with the S_{1c} fabric preserved in the phyllite. Folds of this later generation (i.e., folded S_2 foliation) have an apparent vergence towards the northwest but the rarity of such folds and the lack of three-dimensional exposures prevents a conclusive statement.

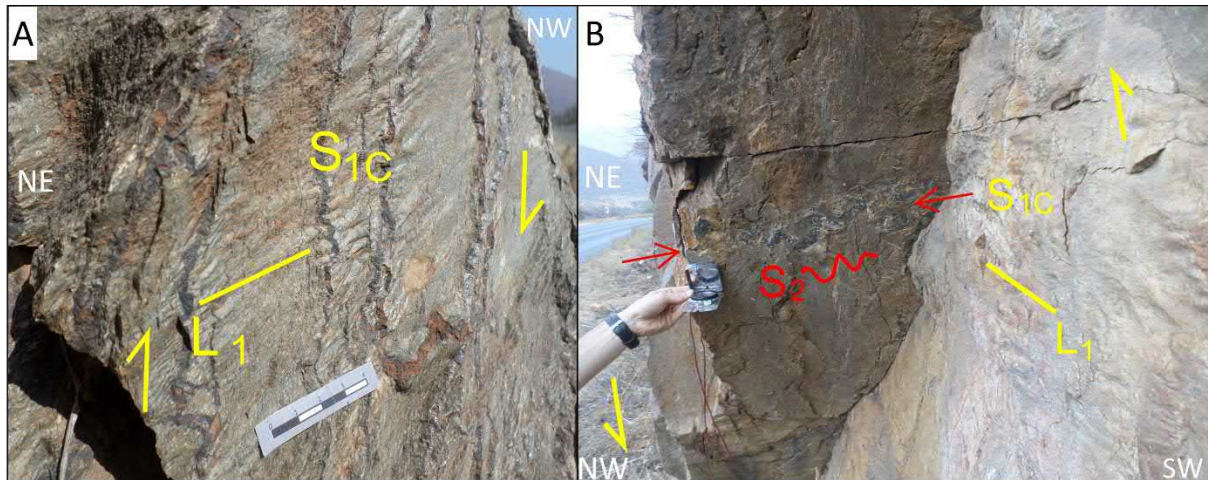


Figure 3.5 Two main directions of compressional forces recorded at Site 3. (A): Sub-vertical folds of a shallow southeast plunge (yellow labels). (B): Later veins were shortened to produce sub-vertical folds (red labels) exposed on an oblique fracture plane.

3.4.2 Petrographic analysis

The phyllite is made up of muscovite, chlorite and minor talc interlayered with carbonates produced through the breakdown of rarely preserved clinopyroxene (Fig. 3.6 C). This assemblage forms a groundmass to sericitised albite crystals sparsely distributed within the sample. Oxides observed include titanite which forms coarse, isolated euhedral crystals in hinge zones of microfolds. Ilmenite occurs as trains of fine anhedral crystals aligned with the crenulated schistose fabric. Accentuating the rock fabrics are quartz and dolomitic veins that are folded into sub-horizontal and sub-vertical microfolds. In sample NP-212, a folded dolomitic vein is associated with clusters of coarse (0.5-1.0 mm) euhedral pyrite preferentially occupying hinges of microfolds.

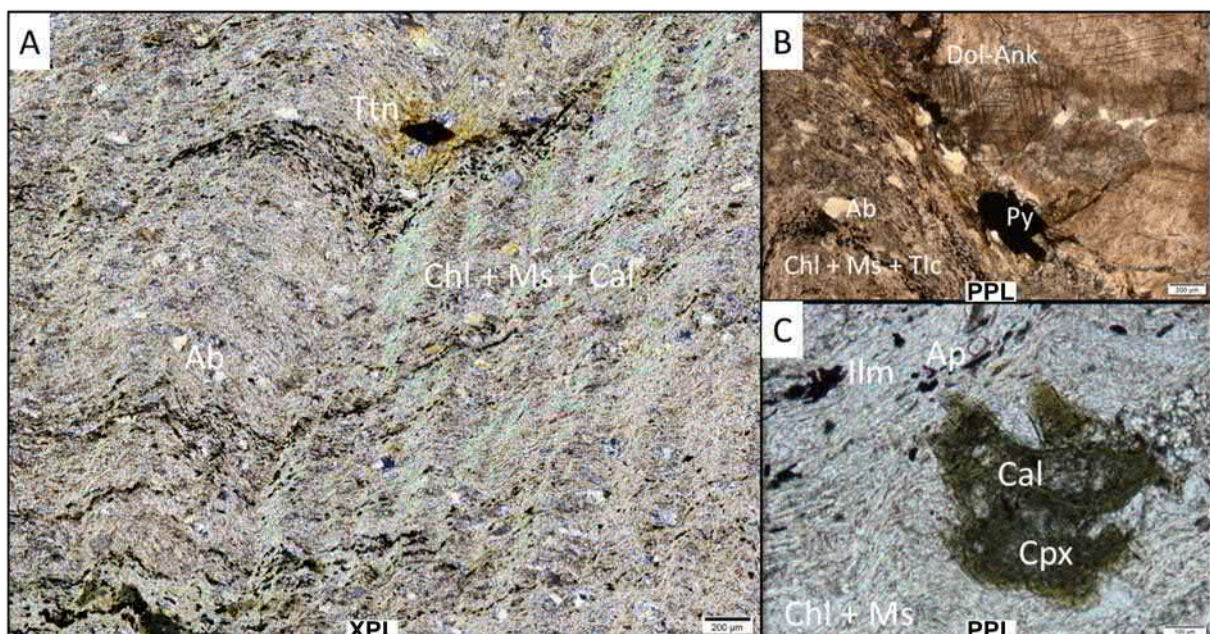


Figure 3.6 Main mineralogy of the phyllite section affected by later sub-horizontal shortening (sample NP-212). PPL and XPL labels are in photomicrographs produced using plane-polarised, or cross-polarised light respectively. (A): Chlorite-muscovite-albite-quartz phyllite assemblage is dotted by trains of ilmenite and isolated titanite crystals. (B): Photomicrographs of the phyllite reveal pyrite-bearing dolomitic veins that predate the crenulation folding of the phyllite. (C): Photomicrograph of the phyllite reveals the alteration of clinopyroxene to calcite, forming blocky calcite aggregates entrained in the foliation. Ab: albite, Ttn: titanite, Chl: chlorite, Ms: muscovite, Cal: calcite, Dol: dolomite, Ank: ankerite, Tlc: talc, Py: pyrite, Cpx: clinopyroxene.

3.4.3 Elemental maps

The phyllitic groundmass is rich in potassium, aluminium, silica and chromium due to the predominance of muscovite and lesser magnesium-poor chlorite. Other elements occur in sparsely distributed crystals in the groundmass. Discrete occurrences of iron and titanium correspond to opaque phases of titanite or ilmenite and pyrite-in the case of iron. Trains of quartz crystals accentuate the crenulated schistosity as does a prominent carbonate vein in the centre of the sample.

The microfolded carbonate vein contains higher calcium, magnesium and iron than the groundmass. Thin silica zones at a high angle to vein margins are more common at fold hinges. These represent microfractures within the dolomite vein. Hinge zones of the folded vein contain clusters of sub- to anhedral pyrite crystals ranging in length from 0.2 to 1 mm.

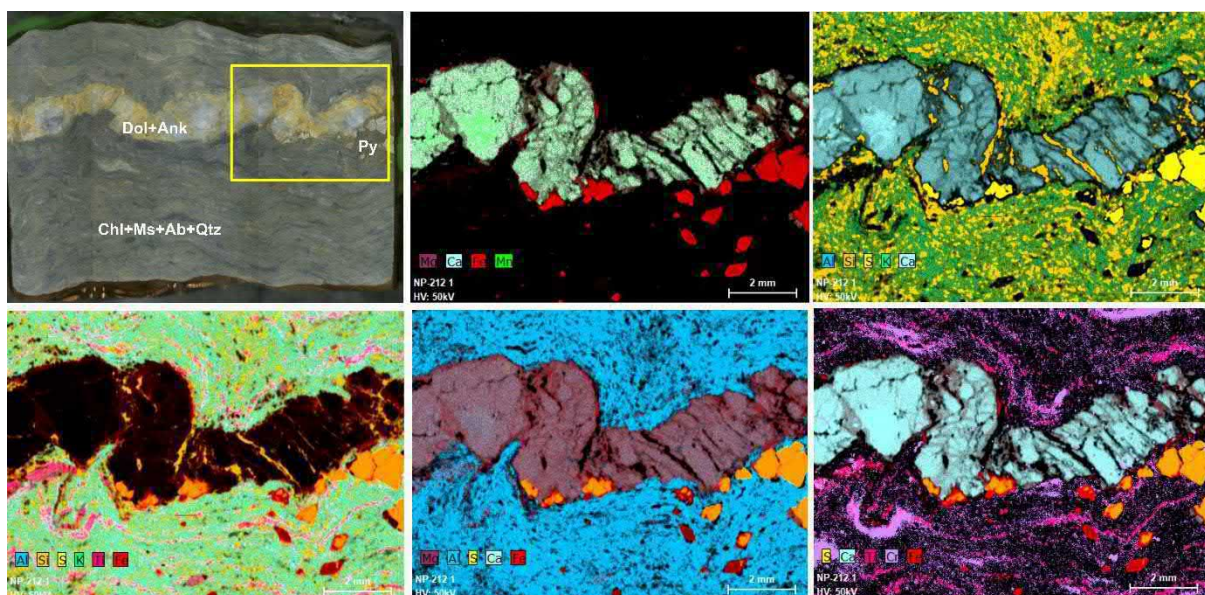


Figure 3.7 Element distribution map of the crenulated phyllite sample NP-212. This shows euhedral pyrite and titanite crystals associated with hinges and kinks of an ankeritic dolomitic vein.

3.5 Discussion

If the sub-horizontal crenulation folds at Site 3 formed after tilting of strata to a sub-vertical orientation (Viljoen, 1963), the following events are understood to have formed the crenulation fabrics. An initial phase of northwest-southeast flattening would have been applied to the unit. This, according to Passchier and Trouw (2005), would destabilise original blocky grains or crystals of the magnesium-rich protolith of Onverwacht Group affinity (Visser *et al.*, 1956; Ramsay, 1963). Consequently, the blocky grains or crystals would be replaced by sheet silicates like chlorite and muscovite to form a phyllitic fabric here denoted as S_1 . Thereafter, subvertical shortening would cause folding of the phyllitic cleavage to produce crenulation cleavage (S_{1C}) and crenulation lineations formed by microfold hingelines (L_1). Later southeast-verging sub-vertical folds prominent in dolomite and quartz veins that crosscut the S_{1C} crenulations would have formed in response to a top-to-the-southeast event of shearing. Contrary to previous deformation sequences proposed (Viljoen, 1963; Anhaeusser, 1969, 1976), cross-cutting relationship suggest that the sub-vertical shortening of planar fabrics at Site 3, and possibly similar sites elsewhere, precedes at least one event of sub-horizontal northwest-southeast shortening. Hinge zones and cleavage planes created by the latter deformation served as loci for pyrite precipitation from hydrothermal fluids (Fig 3.6, 3.7).

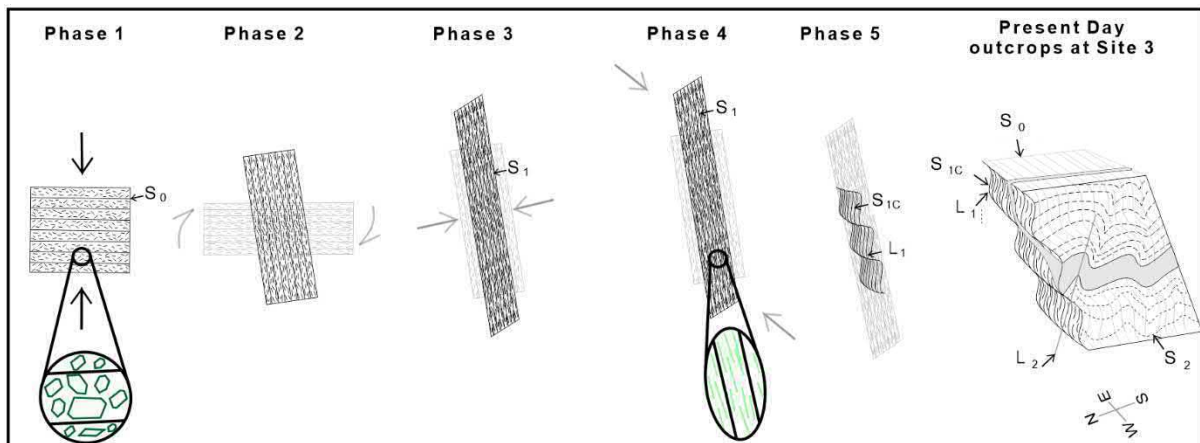
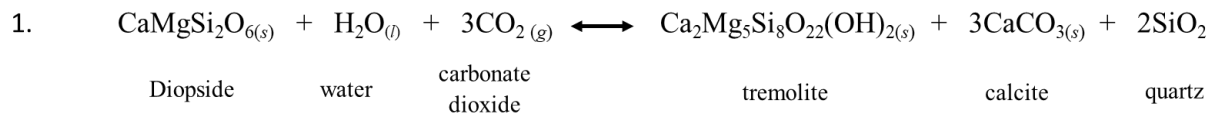


Figure 3.8 Sequence diagram showing associated changes that result in the development of crenulated schists or phyllites. Such changes include flattening associated with the growth of secondary planar minerals that form a schistose fabric which is later rotated and then sheared sub-vertically to produce sub-horizontal crenulation lineations. Diagram inspired by a model proposed by Ramsay (1963).

Due to advanced alteration, the origin of the abundant muscovite in the phyllite could not be confirmed. Literature suggests, however, that mafic components of the Onverwacht Group have undergone propylitic and potassic alteration. It is therefore conceivable that such a

chlorite-epidote-albite assemblage would have been altered to muscovite to produce the phyllite.

The rare occurrence of clinopyroxene in the sample provided a clue to the origin of some of the carbonates in the interfolial spaces of the phyllite. This decarbonation reaction may have proceeded according to reaction 1 below which corresponds to a minimum temperature of 300°C at zero CO₂ partial pressure and possibly followed by further carbonation of tremolite (Winter, 2010).



3.6 Conclusion

The chromium-rich groundmass surrounded by isolated pyroxene crystals with decarbonation textures suggests that the protolith may have been a mafic or ultramafic volcanic unit with isolated clinopyroxene phenocrysts. Decarbonation may have resulted from prograde metamorphism during a regional metamorphic event that predates F₁ and F₂ fabrics of outcrops at Site 3.

Chapter 4: Talc Mineralisation

Parking: (25°40'21.28"S; 31°10'11.22"E); No guaranteed safe access as the mine has been closed and vandalised. Visitors might intercept illegal occupants of the mine property, as was the case during one of two post-closure visits to the mine.

4.1 Site description

Scotia Talc Mine (Fig. 4.1) has been closed since early 2019. For this reason, material from the mine was transported to Site 2 (Chapter 2) of the R38 Geotrail. The following descriptions are based on observations and photographs produced prior to the present study.

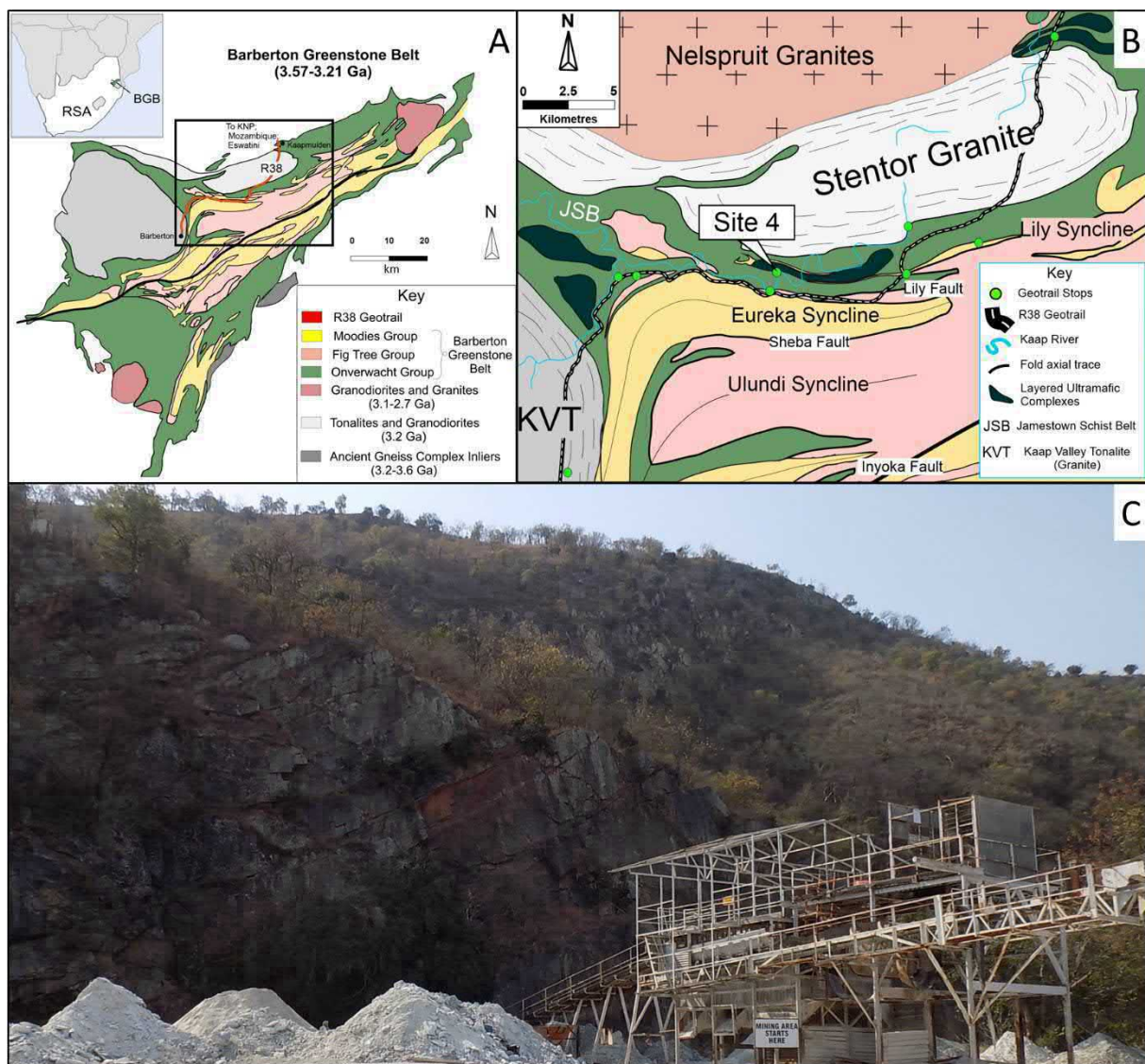


Figure 4.1 Main features of Site 4. (A): Simplified geology of the Barberton Greenstone Belt, rectangular border outlines part of the research area. (B): Insert from (A) showing selected sites of the R38 Geotrail and the location of Site 4 in a Layered Ultramafic Complex. (C): Heaps of low-grade talc, serpentinite and talc-carbonate schists lie exposed near the mine entrance. Maps modified after Visser *et al.* (1956) and Anhaeusser *et al.* (1981).

Talc is a hydrous magnesium silicate ($\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_2$) of a soapy/slippery texture. Because of associated minerals (magnesite, chlorite, mica or serpentine), talc can occur in a spectrum of compositions with a difference in colour and hardness or presence of grit (Tahir *et al.*, 2018). This allows grade classification according to Table 4.1 below.

Table 4.1 Grade scheme of talc samples ranging from most to least valuable (Grade I-IV respectively). Information sourced from Halder (2013).

Talc Grade	Description
I	Pure white colour, smooth texture, grit-free.
II	Tinted, smooth feel, grit-free.
III	Off-colour, smooth feel, grit-free.
IV	White or coloured, contains grit.

4.2 Aspects of talc and Site 4 that may interest laypersons

Displayed in a cocopan at site 2 are various samples of talc and similar-coloured rocks that include talc-carbonate schist, carbonatised serpentinite and various forms of verdite and green cherts. The visitor will identify talc among other rocks through touch and the ability to scratch it with fingernails (Whitney *et al.*, 2007).

Visitors will also be told about the international significance of the Bon Accord Layered Ultramafic Complex (LUC) which hosts the Scotia Talc Mine deposit. Other mineral deposits of the Bon Accord LUC are covered in the current chapter.

4.3 Geology

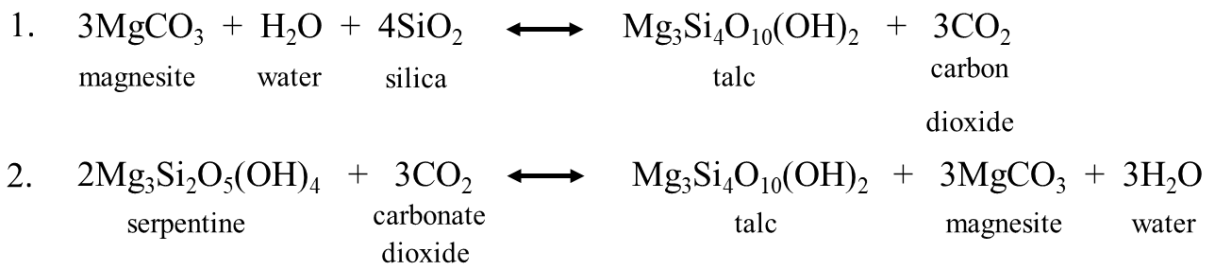
4.3.1 Background

4.3.1.1 Talc in the northern Barberton Greenstone Belt

Being a soft and inert substance, talc is used in cosmetics and in industrial products that include ceramics, insulators and paint fillers (Ward, 1999; Anhaeusser, 2012; Tahir *et al.*, 2018). It was first mined in the Barberton region from the ‘Verdite Mine’ in the Mundt’s Concession Layered Ultramafic Complex of the Jamestown Schist Belt (Anhaeusser, 2012). Other significant occurrences and deposits are concentrated in the north and northwestern margins where they have been exploited since 1918 (Groeneveld, 1975). The largest deposit occurs at the Scotia Talc Mine where reserves remain unexploited due to the closure of the mine in early 2019. The Scotia Talc Mine deposit is hosted in the Bon Accord LUC. This complex intrudes or is folded into younger quartzites of the subvertical Lily Syncline (Anhaeusser, 2006; 2019).

Prominent talc deposits in the BGB are typically hosted in shear zones within talc-carbonate schists, and to a lesser extent, in talc schists and serpentinites (Groeneveld, 1975). According to Van Zyl (1940) in Groeneveld (1975), talc from Site 4 commonly occurs in talc-carbonate- and talc schists. In these schists, high-quality talc was selectively developed at lithological contacts with a marked competency contrast. These include sites where the above-mentioned schists were in contact with more competent quartz-sericite schists, quartz bands and amphibole schists. This, according to Groeneveld (1975), was an indication of a more complete shearing at such contacts, allowing better permeation of hydrothermal fluids and a more complete alteration of the host rock.

It was proposed by N.R. Schindler quoted in Groeneveld (1975) that the talc-carbonate and talc schists which commonly host layer-parallel talc lenses were derived from the carbonatisation of serpentinites. In this way, talc-formation preceded the carbonatisation of serpentinites altered according to the following reactions:



The CO₂-rich hydrothermal fluids were proposed to have been sourced from magmatism related either to the Kaap Valley Tonalite or the Nelspruit Batholith, or from meteoric water (Groeneveld, 1975). Ni-rich content of the serpentinites associated with the talc-carbonate and talc schists was suggestive of an ultramafic or mafic igneous protolith. Notably, a number of historical talc deposits yielding economic grades were typically located near porphyritic intrusions related to the Kaap Valley Tonalite (Anhaeusser, 1976b). The presence of tourmaline in some talc samples is understood to indicate the role of granitic fluids in talc formation (Anhaeusser, 1976a)

4.3.1.2. Other deposits associated with talc-bearing ultramafics at Site 4

Millerite (nickel sulphide) occurs as an accessory mineral in the talc-carbonate and talc schists of the Scotia Talc Mine (Groeneveld, 1975). Minable concentrations occurred in sub-economic deposits at the western and eastern segments of the Bon Accord LUC. From the western extent a rare nickel-rich spinel called Trevorite (NiFe₂O₄) was recovered (Trevor, 1920; Tredoux *et*

al., 1989). This deposit became the type locality of trevorite and was suggested to have been sourced from a meteorite (De Waal, 1978, 1979), or an ophiolite bearing a siderophile-rich fragment of the Earth's core (Tredoux *et al.*, 1989) or a komatiite-hosted massive sulphide deposit (O'Driscoll *et al.*, 2014) later oxidised and desulphurised to form trevorite (Tredoux *et al.*, 1989, O'Driscoll *et al.*, 2014). The eastern segment of the same layered complex contained iron, nickel and copper sulphide mineralisation mentioned in decreasing order of abundance.

The latter mineralisation extends from the northern margin of the Scotia Talc Mine operations, extending eastwards along a 1400 m strikelength (Keenan, 1986). The mineralised zone occurs at the contact of a banded chert unit of the Lily Syncline and an ultramafic serpentinite altered to talc schist that was mined for talc. The zone is about 25 m wide and ranges-within that width-between disseminated and semi-massive ore (Keenan, 1986). In order of abundance, the mineralised zone contains pyrite, pyrrhotite and pentlandite with minor to trace amounts of gersdorffite, chalcopyrite, millerite, magnetite and chromite (Keenan, 1986). Notably, trace amounts of magnetite and rutile were found in some sulphide-bearing serpentinites while chromite occurred in all samples.

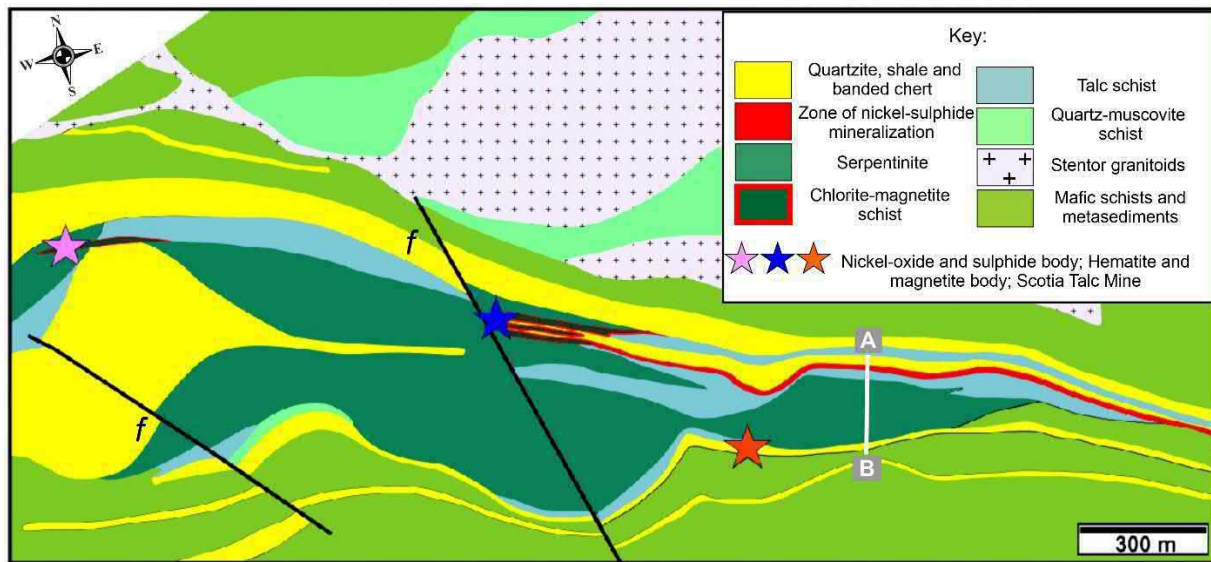


Figure 4.2 Simplified geological map of the northern BGB between Joe's Luck Siding and Eureka Siding. Depicted in the map are the locations of the Trevorite (Ni-rich Magnetite) type locality (purple star); an iron oxide body associated with nickel sulphide occurrences (blue star) and the Scotia Talc Mine. Modified after O'Driscoll *et al.* (2014) who adapted the original Keenan (1986) map. A schematic section of line A-B is included below.

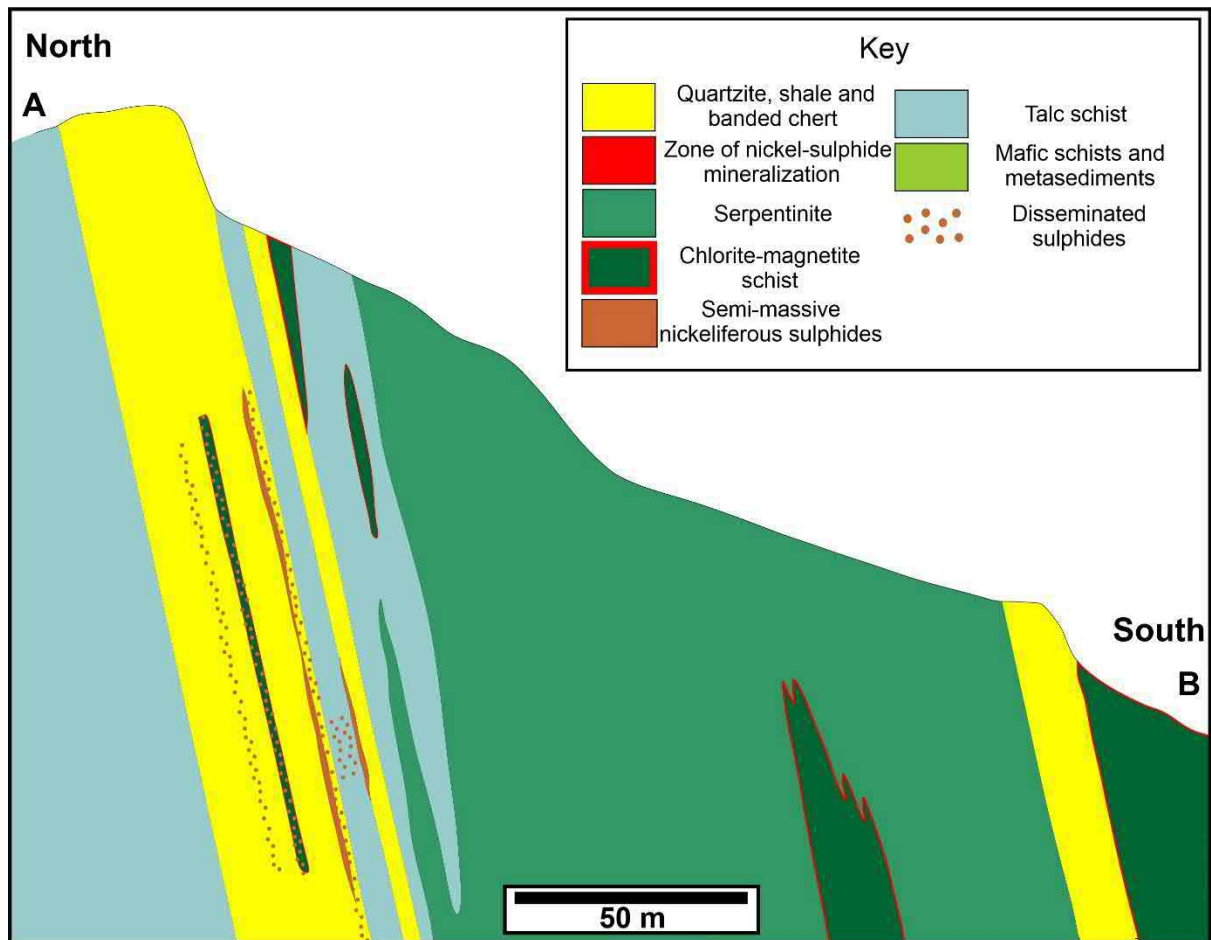


Figure 4.3 Schematic diagram of lithologies along line A-B in the previous map show the relationship between steep ridges of the Lily Syncline and nickel-sulphide mineralised zones. Modified after Keenan (1986).

4.3.2 Present study

4.3.2.1 Research question

The serpentinised layered ultramafic intrusion that was mined at Scotia Talc Mine was already known to be the host of talc mineralisation. What was unconfirmed was the textural relationship between talc and associated minerals such as magnesite and millerite (NiS).

4.3.2.2 Aim

To collate existing information about talc mineralisation at Scotia Talc Mine and to describe phases that occur within the talc host rock. Samples were obtained before the mine was closed.

4.3.2.3 Objectives

To summarise existing literature and describe phases in a talc-bearing host rock by studying thin sections and describing phase and elemental maps.

4.3.2.4 Methods

To identify phases that predate talc formation, a talc-poor sheared and a metasomatised serpentinite (NP-STM) that are abundant in the mine dump were collected after the mine closed down in 2019. Weathered surfaces were cut off and fresh surfaces were cut for thin sectioning and bulk chemical analysis, the results of which are shown in comparative plots of Chapter 2 (Fig. 2.17).

4.4 Results

4.4.1 Outcrop observations

High-grade talc deposits of the Scotia Talc Mine occur in metasomatised peridotites. The earliest and most pervasive alteration observed in the mine was serpentinisation. The serpentinites were carbonatised to magnesite along sub-vertical, interfolial dilational zones as well as in sub-horizontal veins. Later talc development favoured kinks in deformed veins as well as interfolial spaces between serpentinised schists.

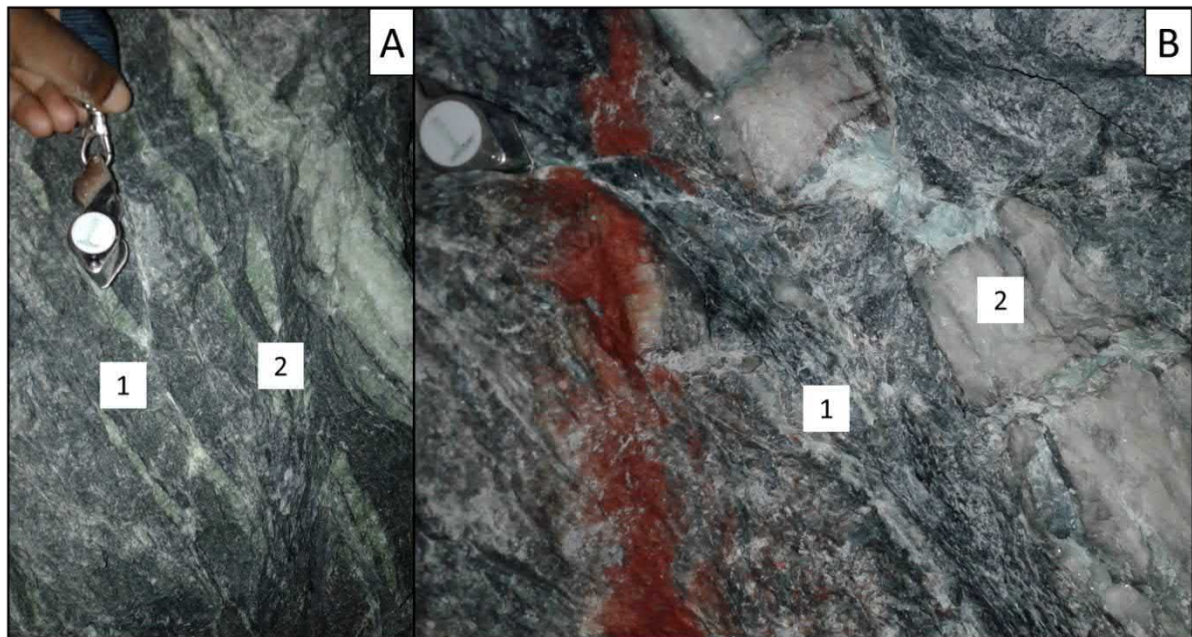


Figure 4.4 Altered serpentinites of the Bon Accord Layered Ultramafic Complex at Scotia Talc Mine. Numbers tell of relative ages (1=oldest). (A): Evidence of carbonatization (2) of serpentinite (1) in sub-vertical dilational structures. (B): Talc (1) in a steatized serpentinite close to a sub-vertical quartz vein (2).

4.4.2 Petrographic analysis

Sample NP-STM, the host rock of talc was studied to describe associated alteration processes according to a sample obtained after Scotia Talc Mine had been closed. From the representative sample studied, the dominant mineral components are clusters of serpentine resorbed into a groundmass of magnesite and minor talc (Fig.4.5). Magnesite also occurs as foliation-discordant veins through the sample. These veins are kinked along the same direction as the preferred orientation of serpentine aggregates and the enclosing groundmass.

Magnesite also forms apophyses that extend into aggregates comprised of millerite-after-chromite (Fig. 4.6 C).

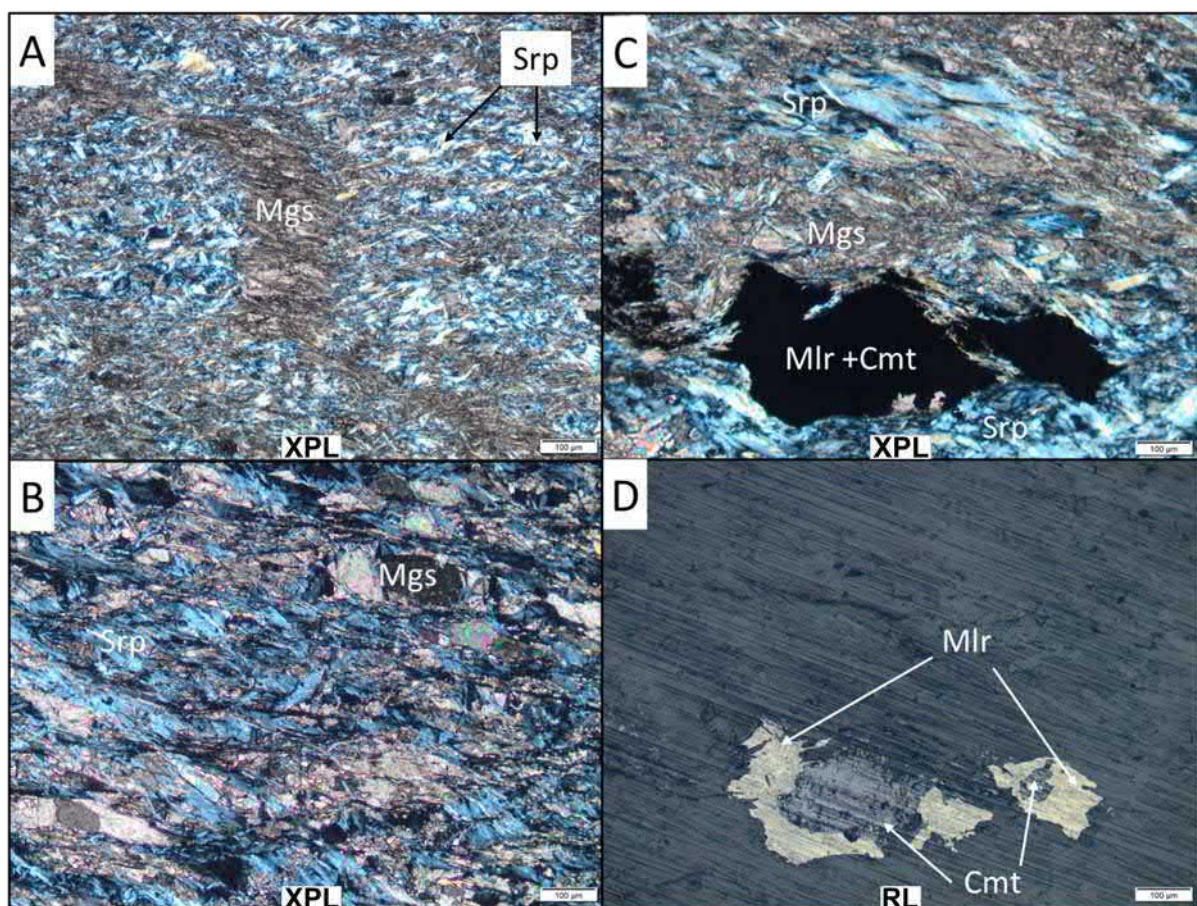


Figure 4.5 Photomicrographs of a talc-bearing carbonatised serpentinite from Scotia Talc Mine (Sample NP-STM). XPL and RL labels are in photomicrographs produced using cross-polarised or reflected light, respectively. (A): Foliation parallel magnesite layers in serpentinite intersected by a discordant magnesite vein later kinked by shear planes that coincide with talc-bearing sites (B): Kinked foliation parallel magnesite lenses. (C): Serpentinite lenses are associated with opaque phases invaded by magnesite along cracks. (D): RL photomicrograph shows the paragenesis of millerite (Mlr) to be the alteration of chromite (Cmt).

4.4.3 Phase maps

Incipient talc development in sample NP-STM (Fig. 4.6) occurs along foliation-parallel fractures through a groundmass of serpentine and magnesite, and in fractures along which magnesite veins were kinked. This relationship is only perceivable in phase maps of the sample included below. Millerite envelopes chromite grains and magnesite-filled fractures occur in both opaque minerals.

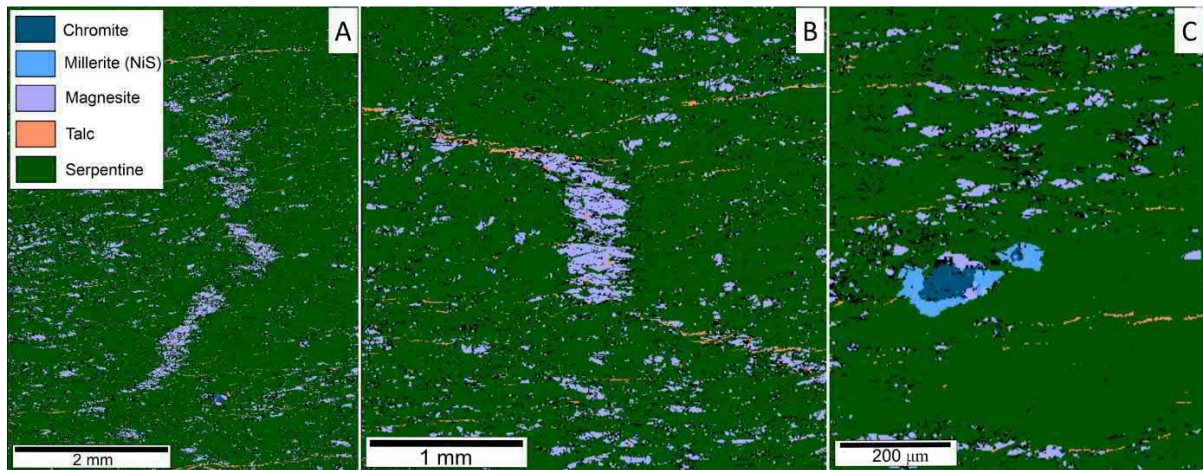


Figure 4.6 TIMA phase maps of talc-bearing, carbonatised serpentinite from Scotia Talc Mine (Sample NP-STM). B and C correspond to XPL photomicrographs shown above (A,C). In phase maps (A) to (C), talc occupies late-stage brittle fractures and tension gashes parallel to the offsetting of carbonate veins. It can also be seen that talc occupies fractures and tension gashes within the magnesite veins and the serpentinite and occupies fractured parts of opaque minerals.

An element distribution map of a representative sample (NP-STM, Fig. 4.7) shows the following associations. Except for iron, clusters of transitional metals reflect rutile, millerite, and chromite crystals. Iron occupies diffuse magnesitic and dolomitic zones associated with magnesium, calcium and a lack of silica. Calcium also occupies discrete microcrystals of plagioclase coincident with discrete aluminium-rich zones that contain silica.

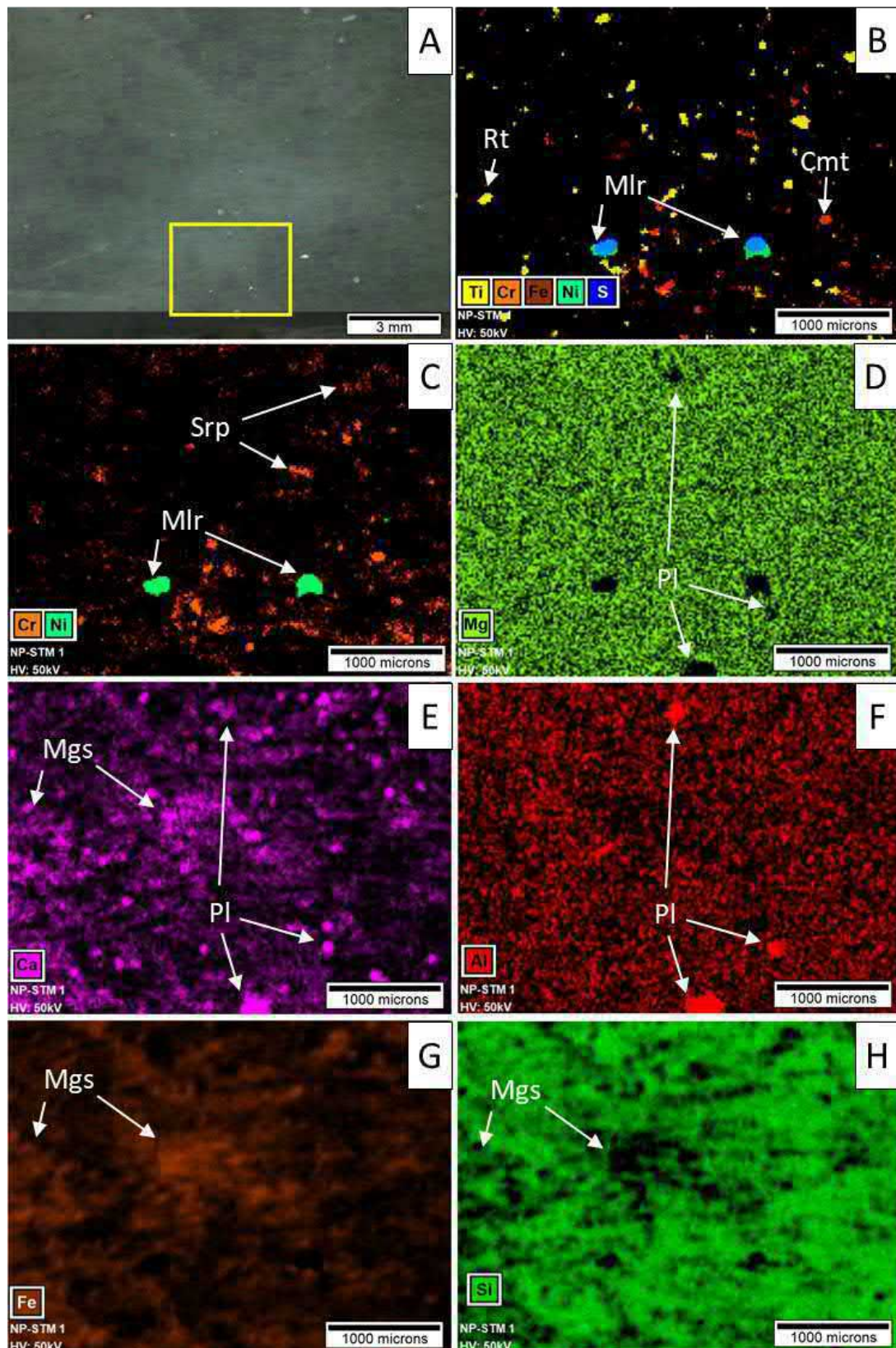


Figure 4.7 Elemental maps of selected elements within a carbonatised serpentinite (Sample NP-STM). (A): Mapped area on serpentinite with incipient talc mineralisation shows discrete occurrences of transitional metals Ti, Ni and Cr whereas Fe, Mg and Ca are predominantly diffuse in occurrence. Some Ca-rich discrete zones coincide with discrete aluminium-rich zones that contain silica.

4.5 Discussion

According to a classification scheme cited in a review by Abu-Jaber and Kimberly (1992), talc mineralisation can be grouped into four main categories depending on the origin. These categories differentiate between talc of an ultramafic, mafic, metasedimentary and metamorphic origin (McCarthy, 2006). Respective to the categories mentioned above, talc-forming processes may involve the conversion of serpentinitised and carbonatised peridotite or gabbro to talc; the hydrothermal silicification of dolomite and magnesite, or the steatitisation of tremolite during retrograde metamorphism. A talc-formation model applicable to the Scotia Talc Mine deposit, would include the serpentinitisation and carbonatisation of peridotite. Subsequent silicification of secondary magnesite would produce talc (Groeneveld, 1975). Though talc was not seen to invade any fractures within millerite or chromite, the presence of magnesite-filled fractures on both opaque minerals suggests that the replacement of chromite by millerite predates talc mineralisation (because carbonatisation predates steatitisation in the samples studied). This relationship supports the formation of millerite in a komatiite seafloor prior to later deformation and metasomatism as proposed by O'Driscoll *et al.* (2014).

4.6 Conclusion

According to previous authors who assessed the talc mineralisation at Site 4 and the data presented from a representative sample during the present study, it appears that talc mineralisation is structurally controlled. The observed replacement of magnesite by talc along brittle structures is in agreement with previous descriptions of talc growth through the silicification of magnesite according to equation 1 from Groeneveld (1975) above.

4.7 Recommendation

Once Scotia Talc Mine is reopened and safe to access, a study of the spatial relation of tourmaline-bearing talc-schists and zones of high-quality talc would provide a better understanding of the nature of fluids involved in paragenesis.

Chapter 5: Moodies Basal Conglomerate

Parking: 25°41'1.45"S; 31°10'3.09"E; Representative outcrop: 25°41'4.68"S; 31° 9'56.25"E.

5.1 Site description

Site 5 is a scenic roadside stop where the Kaap River winds closest to the road (Fig. 5.1). This stretch of road receives high traffic volume and occurs near two bends. Care should be taken when leaving and returning to the parking spot.

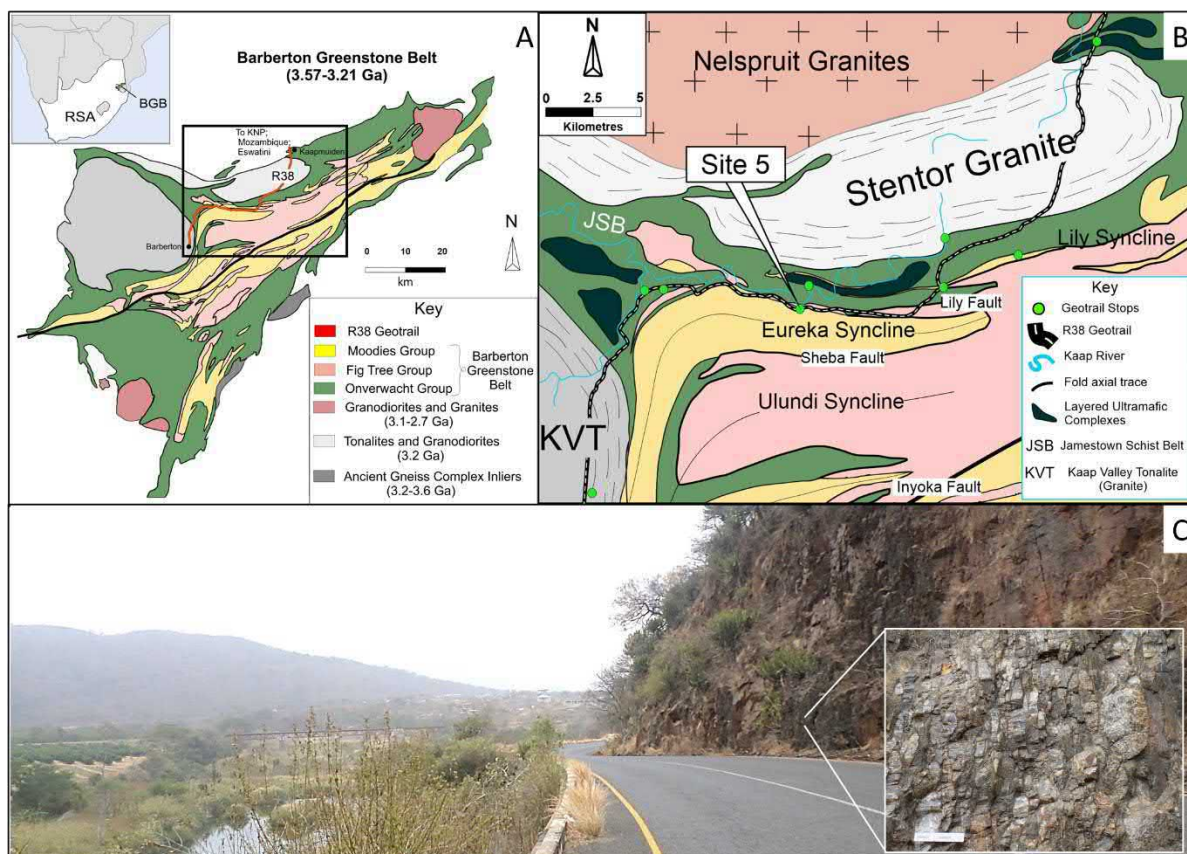


Figure 5.1 Main features of Site 5. (A): Simplified geological map of the R38 Geotrail in the context of the BGB. (B): a detailed view of the black rectangle in (A) shows the location of Site 5 at the southern margin of the Stentor Pluton. (C): View towards the northeast shows Site 5 and the Kaap River in the north. Scale in insert is c. 4.5 cm long. Maps modified after Visser *et al.* (1956) and Anhaeusser *et al.* (1981).

5.2 Aspects of the Moodies Basal Conglomerate and Site 5 that may interest laypersons

Site 5 is a well-known stop among geologists interested in quantifying strain from deformed clasts. Almost all granite clasts from this site were sampled for radiometric dating because they were found to imply the possibility of a pre-3.2 Ga potassic granite terrain that has since been completely eroded (see Kröner *et al.*, 2018 and references therein).

The proximity of the Kaap River to Site 5 provides an opportunity to use present-day sedimentary processes to convey the significance of Moodies Group sediments. A key point of discussion would be the usefulness of sedimentary rocks in recording some of the lithologies and environmental conditions on the surface of the Earth at the time of deposition.

The sub-vertical orientation of strata contrasts sharply with the flat sediment deposits of the Kaap River below the road. This presents an opportunity to discuss the sub-vertical folds and faults of the Barberton Greenstone Belt. Such a discussion would include the fault zone along which the Kaap River flows.

5.2 Aim

To collate information about key features of the Moodies Basal Conglomerate at Site 5.

5.4 Geology

5.4.1 Background

Sedimentary units are one of the best repositories of remnants from terranes no longer preserved on the Earth's surface (e.g., Froude *et al.*, 1983). Some of the oldest known crustal materials on Earth are in the form of detritus within Archean metasedimentary rocks of the Yilgarn Craton (Compston and Pidgeon 1986; Wilde *et al.* 2001; Valley *et al.*, 2005; Bell *et al.*, 2015; Harrison *et al.*, 2017; Wang and Wilde, 2017). Similarly, metasedimentary rocks of the Palaeoarchean Barberton Greenstone Belt (BGB) preserve remnants of the oldest *sensu stricto* granites not only of the Kaapvaal Craton, but the world at large (Van Niekerk and Burger 1978; Tegtmeier & Kröner, 1987; Kröner & Compston, 1988; Sanchez-Garrido *et al.* 2011; Heubeck *et al.*, 2013; Zeh *et al.*, 2013; Agangi *et al.*, 2018; Kröner *et al.*, 2018).

Exposed at Site 5 are units of the two upper lithostratigraphic units of the Barberton Greenstone Belt (BGB) viz. the c.3225-3.217 Ma Moodies Group and the underlying 3.260-3.225 Ma Fig Tree Group (Van Eden, 1941; Visser *et al.*, 1956; Anhaeusser, 1975; Anhaeusser, 1976;

Eriksson, 1977, 1979; Reimer *et al.*, 1985; Heubeck and Lowe, 1994, 1999; Byerly *et al.*, 2018; Heubeck, 2019).

Metasedimentary rocks of the Moodies Group are mainly comprised of terrestrial to shallow-marine lithologies divided by a regional basaltic lava into lower, upward fining sequences, and upper, unsystematic or upward coarsening sequences (Heubeck and Lowe, 1994; Heubeck *et al.*, 2013, Heubeck *et al.*, 2016; Heubeck, 2019). Further divisions of this Group are tectonically based (Lowe and Byerly, 1999; Byerly *et al.*, 2018), with the regional Inyoka-Saddleback fault system separating the Moodies Group into a microcline-bearing facies in the northern domain, and a correlatable microcline-lacking facies in the southern domain (Heubeck and Lowe 1994, Byerly *et al.*, 2018; Heubeck, 2019). Situated in the outer arc of the northern limb of Eureka Syncline (Ramsay, 1963; Anhaeusser, 1965; 1976; 1984), this roadcut exposure displays steeply inclined Moodies strata of the northern domain. The strata here dip, fine and young towards the south (Anhaeusser, 1976; Lowe *et al.*, 2016).

A walk from the parking spot to the outcrop (25°41'4.68"S; 31° 9'56.25"E) is a traverse across a c.3225 Ma transitional contact between Fig Tree strata in the northeast, and basal Moodies strata in the southwest. The contact separates felsic volcanoclastics of the uppermost unit of the Fig Tree Group in the northeast (Schoongezicht Formation), from relatively quartz-rich sandstones interbedded with the basal conglomerate in the southwest (Anhaeusser, 1966, 1976; Heubeck *et al.*, 2016).

5.4.2 Moodies Basal Conglomerate

5.4.2.1 Deformed clasts

A northwest-southeast direction of clast flattening-with a constrictional component, was determined from the basal conglomerate (Anhaeusser and Viljoen, 1965; Anhaeusser, 1966). This shortening produced a linear fabric of aligned, flattened and elongated clasts. The clasts plunge steeply towards the southeast, which is the dip direction of bedding and slaty cleavage of the outcrop (Ramsay, 1963; Anhaeusser, 1966; Gay 1969). In contrast, the same conglomerate unit shows little to no strain ca. 20 km south of this exposure (Table 5.1, Fig. 5.2.) (Anhaeusser, 1966; 2014, 2019). This has been interpreted to suggest the predominance of granite emplacement in the generation of regional stresses required to produce the increased clast strain observed at margins of the Barberton Greenstone Belt (Anhaeusser 1966; 1976; 1984; 2014; 2019; Tegtmeier & Kröner, 1987).

Mineralogical influences on the distribution of strain in the BGB and the surrounds have been observed in both regional and outcrop scales (Van Eeden, 1941; Anhaeusser, 1966). Clasts of the volcano-sedimentary units are most deformed whereas feldspar porphyry, sandstone and granitic clasts bear lesser flattening and elongation strain (Ramsay, 1963; Gay 1969; Kisters and Anhaeusser, 1995). The same relationship is evident on a regional scale, where the granitoids that surround the BGB appear to have served as competent bodies onto which the BGB supracrustals were thrust (Van Eeden, 1941; Anhaeusser, 1966, 1972, 1976, 2019; Fripp, 1980; De Wit *et al.*, 1987).

Table 5.1 Variable clast deformation in the Moodies Basal Conglomerate of the Eureka Syncline (Anhaeusser, 1976).

Position in Eureka Syncline	Sample locality	Percentage deformation of the three principal axes (a>b>c) of conglomerate pebbles in the Eureka Syncline		
		a	b	c
Northern limb	Royal Sheba Gold Mine area	19	9	-23
	Royal Sheba Gold Mine Area	26	2	-28
	Honeybird Creek area	29	7	-23
	Honeybird Creek area	32	0	-21
Southern limb	Hislop's Creek area	62	-4	-35
	Elephant's Kloof area	51	-9	-26
	Clutha Siding	30	10	-30
	Clutha Siding area	35	-24	-3
	Clutha Gold Mine	43	10	-29
	Joe's Luck Siding Area	47	11	-39
	Lilydale area	48	-10	-25
	Honeybird Creek area	64	7	-43
	Honeybird Creek area	67	31	-44
	Ezzy's Pass-Eureka Siding area (includes Site 5)	49	12	-40
Positive values represent percentage elongation and negative values represent percentage decrease in clast axes.				

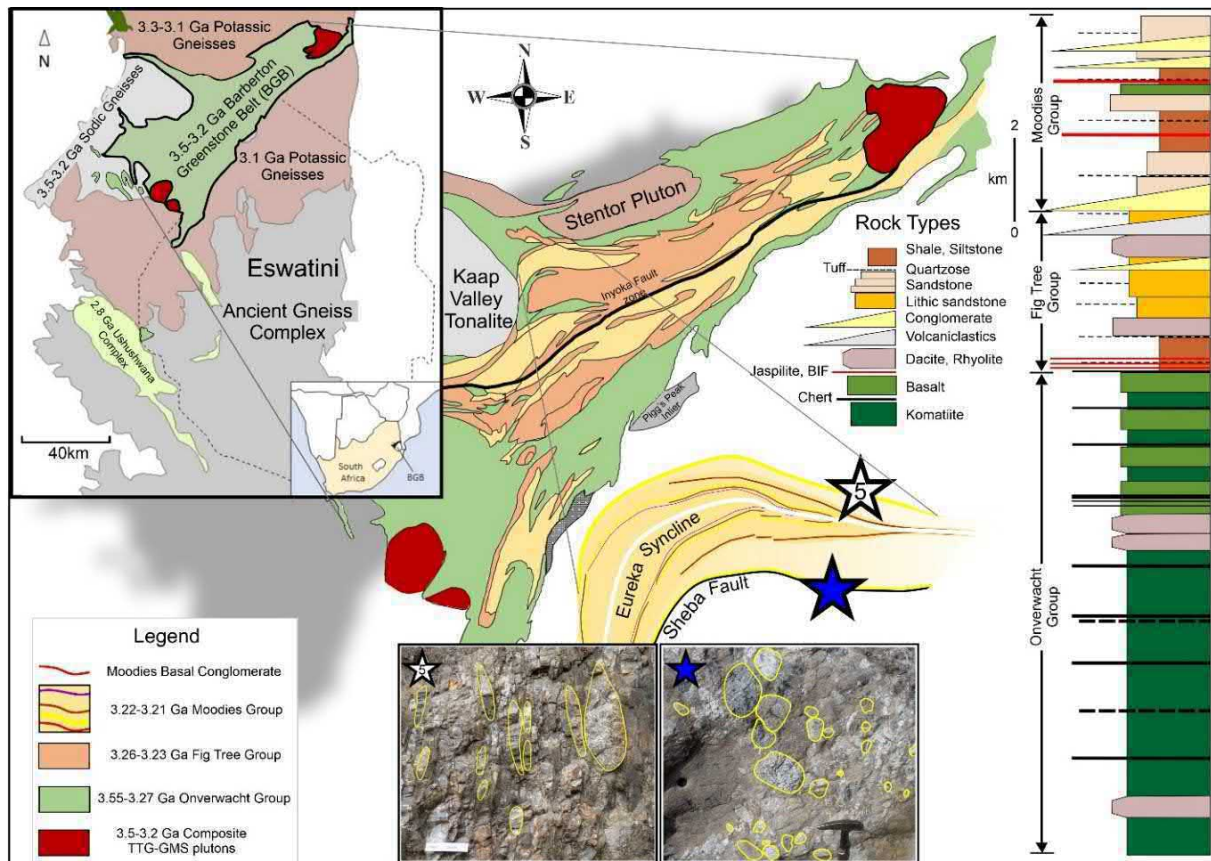


Figure 5.2 Geological context of the Moodies Basal Conglomerate from which >3.3 Ga potassic granite clasts and detrital zircon grains were liberated (Kröner *et al.*, 2018 and references therein). From the stratigraphic column of the Barberton Greenstone Belt it can be seen that the source region of Moodies Group detritus was more quartz-rich than that of the underlying lithostratigraphic units. Maps modified after (Kröner *et al.*, 1989) and Anhaeusser *et al.* (1983); stratigraphic column modified after Schmitz and Heubeck (2021).

5.4.2.2 Composition

The basal Moodies conglomerate is a polymict, mostly clast-supported pebble-cobble conglomerate (Fig. 5.1, 5.2) interbedded with greywacke, as well as various types of sandstones that become increasingly quartz-rich up the stratigraphy (Anhaeusser, 1976; Heubeck and Lowe, 1994). One of the distinguishing features of Moodies clastics is an abundance of monocrystalline quartz within the matrix of conglomeratic beds, and as a framework grain in sandstone beds dominating up to 3.7 km thickness of Moodies strata in the northern BGB (Heubeck and Lowe, 1994; Heubeck *et al.*, 2019). This suggests the initial access of river systems to uplifted felsic plutonic bodies and their volcanic equivalents, at the onset of Moodies-time deposition (Anhaeusser, 1966, 1976; Eriksson, 1980; Jackson *et al.*, 1987; Heubeck and Lowe, 1994; Hessler and Lowe, 2006; Zeh *et al.*, 2013; Heubeck 2019). Lateral

variation in clast composition, thickness and the abundance of sandy beds have been noted at different exposures (Anhaeusser and Viljoen, 1965; Anhaeusser, 1966, 1976; Eriksson, 1980; Heubeck and Lowe, 1994, Heubeck 2019). This is understood to suggest the existence of varied sources of Moodies sediments deposited either in various contemporaneous Moodies basins or one large Moodies basin that received partially mixed deposits of detritus from rivers that drained different sources to produce localised differences in clast lithologies (Heubeck and Lowe, 1994, Lowe and Byerly, 2007; Byerly *et al.*, 2018; Heubeck, 2019).

Mineralogically, variants of chert dominate the clast population of the conglomerate (Ramsay, 1963; Anhaeusser, 1966, 1976; Gay, 1969; Reimer *et al.*, 1985; Sleep and Hessler, 2006). Though substantial amounts of chert variants occur within the BGB, these are overrepresented due to the higher weathering susceptibility of other lithologies, particularly the mafics and ultramafics (Ramsay, 1963), which, by volume, dominate BGB stratigraphy (Lowe and Byerly, 1999; Byerly *et al.*, 2018; de Wit *et al.*, 2018). The rest of the clast population is made up of iron-formation, red-banded jasper, feldspar porphyries, sandstone, conglomerate, shale, and volcanics (Ramsay, 1963; Anhaeusser 1966, 1976; Gay, 1969; Reimer *et al.*, 1985; Sleep and Hessler, 2007). As much as 5% of the pebble population is made up of granitoids and metamorphic rocks (Reimer *et al.*, 1985).

5.4.2.3 Well-studied granite clasts

Though a minority, the granite pebbles of the Moodies basal conglomerate have been the focus of many publications (Kröner *et al.*, 2018 and references therein). This is because they are texturally and chemically distinct from known granites of the BGB (Kröner *et al.*, 2018 and references therein). They contain quartz and feldspar intergrowths, are of a potassic, calcium-poor composition, and were found to contain >3.33 Ga magmatic and xenocrystic zircons within a terrane known to contain only <3.33 Ga sodic plutons and felsic volcanics (Anhaeusser, 1966; Van Niekerk and Burger 1978; Tegtmeier & Kröner, 1987; Kröner & Compston, 1988; Kamo and Davis, 1994; Sanchez-Garrido *et al.* 2011; Zeh *et al.*, 2013; Agangi *et al.*, 2018; Kröner *et al.*, 2018; Byerly *et al.*, 2018). The antiquity of these zircon ages suggests that Moodies-time fluvial systems may have had access to a source terrain that is no longer accessible and much older than any of the sodic granitoids in the immediate surrounds (Van Niekerk and Burger 1978; Tegtmeier & Kröner, 1987; Kröner & Compston, 1988; Sanchez-Garrido *et al.* 2011; Heubeck *et al.*, 2013; Zeh *et al.*, 2013; Kröner *et al.*, 2018).

The lack of an existing granitic terrane compatible with the Moodies granitic clasts and detrital zircons raised a provenance issue (Kröner *et al.*, 2018 and references therein). This followed field observations of south-directed paleocurrent indicators within units of lower Moodies clastics, which demands a north-based source (Anhaeusser 1976, Reimer *et al.* 1985). In addition to that, localised south-dipping alluvial fan deposits within the upper Moodies Group also supported a southerly transport of detritus from a northerly source (Heubeck and Lowe, 1994). A suitable north-based source was proposed by Anhaeusser (1976) to have been the ca 3.3 Ga Nelspruit Migmatite Terrane (NMT) (Robb, 1978; Kröner *et al.*, 2018). This terrain is composite in nature, comprising tonalitic to *sensu stricto* granitic phases. However, the NMT is not only younger than most of the granite clasts but was also found to be geochemically incompatible with them (Hao Wang cited in Kröner *et al.*, 2018).

Another probable granitic source terrane for Moodies granite clasts is the 3.6-3.2 Ga Ancient Gneiss Complex (AGC) in Eswatini (Eriksson, 1980; Reimer *et al.*, 1985; Jackson *et al.*, 1987; Compston and Kröner, 1988). If the AGC, located ca. 80 km southwest of Site 5-with a closer multiphase inlier (Kröner *et al.*, 2014) exposed about 30 km southeast of Site 5, is the source terrain, northwards transportation of sediments would be implied. This transport direction was reported by previous workers (Eriksson, 1980; Jackson *et al.*, 1987; Reimer *et al.*, 1985; Compston and Kröner, 1988). However, variable palaeocurrent directions in Moodies strata have also been documented (Reimer, 1967; Eriksson, 1977; Heubeck and Lowe, 1994; Homann *et al.*, 2015). These variable palaeocurrent directions of the Moodies Group were understood to reflect an uplifted source terrane drained by variably oriented river systems (Heubeck, 2019).

After observing geochemical and lithological similarities between the Moodies and Fig Tree Groups, Zeh *et al.* (2013) noted that the two Groups not only have the occurrence of granite clasts in common (Ramsay, 1963; Eriksson, 1980), but they contain detritus of U-Pb-Hf isotopic geochemistry matching that of the AGC. The reworking of Fig Tree Group sediments-that may have been extensively exposed in the northern BGB, would explain why Moodies detritus originally derived from a southern source terrane (AGC) would have palaeocurrent indicators requiring a northern source terrane (Zeh *et al.*, 2013). Additionally, the rarity of pre-BGB and pre-AGC zircons in Moodies granite clasts and other Moodies and Fig Tree clastic units supported the unlikelihood of there ever being an extensive, fully eroded, pre-BGB, potassic granitic source terrane in the northern BGB from which Moodies Group detritus was derived (Drabon *et al.*, 2017). This was cemented by findings of Drabon *et al.* (2017), who after dating >3400 zircons from various BGB clastic sediments and impact spherules beds,

found that less than 0.5 % of them were older than felsic magmatic rocks of the Barberton Granite Greenstone Terrane.

5.5 Conclusion

Though an ancient, evolved granitic terrain with pre-3.3 Ga potassic granites may not have served as a basement for the earliest strata of the BGB as suggested by Ramsay (1963), the role of regional granites in the deformation of BGB margins remains widely accepted (Anhaeusser, 2019 and references therein).

Chapter 6: Lily Syncline

6A Parking: 25°40'27.64"S; 31° 14'28.46"E; 6B Parking: 25°39'38.24"S; 31°17'9.68"E

6.1 Site description

This site offers a view of the steep, narrow ridge at the core of the Lily Syncline. Parking space is available near the “Kaapmuiden 20” signpost shown below (Fig. 6.1). West of this signpost is a steep-sided valley along which flows the Honeybird Creek. For this reason, vehicles should be parked on grassy plains clear of the edge of dense tree cover to the west. A decimeter-scale, pebble clast, matrix-supported conglomeratic bed occurs at 25°40'28.74"S; 31° 14'30.18"E. This is a vantage point for viewing the Lily Ridge.

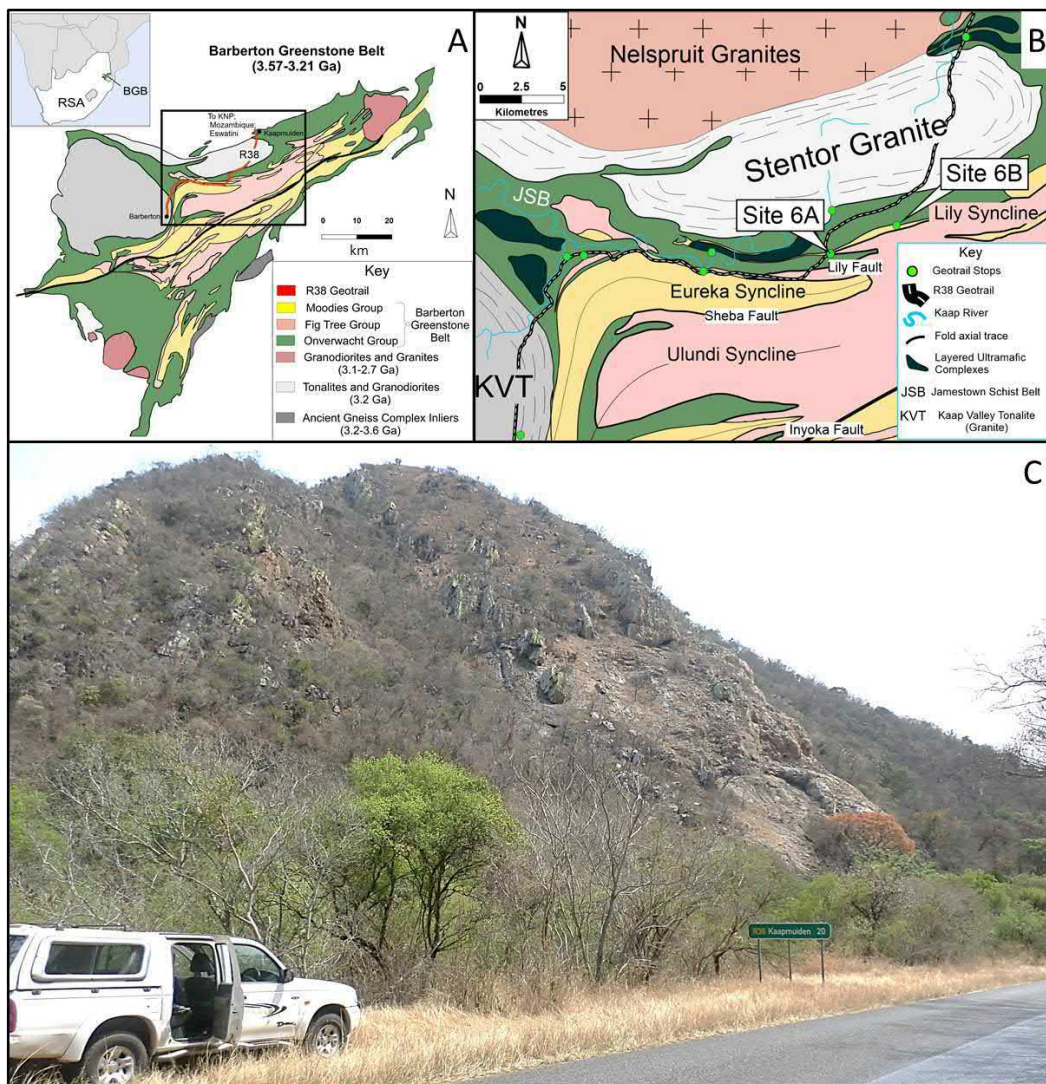


Figure 6.1 Main features of Site 6A. (A): Simplified geology of the R38 road in the context of the Barberton Greenstone Belt. Modified after Visser *et al.* (1956) and Anhaeusser *et al.* (1981). (B): A detailed view of the rectangular border outlined in (A) shows the location of Site 6 within the narrow Lily Syncline. (C): View of the core of the Lily ridge from a conglomeratic bed on the eastern side of the road.

6.2 Aspects of the Lily Syncline and Site 6A that may interest laypersons

Following the collapse of a crown pillar in 2016 (Barberton Times, 2016), the Lily Mine became well-known to South Africans. At Site 6A and B, visitors will be introduced to the Lily Syncline in order to provide a structural context for the Lily Mine area and a ridge that is a prominent feature between Site 5 and 7 of the R38 Geotrail. This site will be used to discuss the steepening of sedimentary beds and increased intensity of flattening strain at the margins of the Greenstone Belt that lie proximal to granitic bodies.

A short walk towards the nearest road bend north of the parking site offers a view of a crowned eagle's nest perched on a branch on the eastern side of the road (Figure 6.2). Of appeal to birding enthusiasts will be a potential sighting of the majestic bird of prey. Visitors will be asked to submit pictures of sightings on social media to contribute to social media posts about the R38 Geotrail once various online pages have been set up.

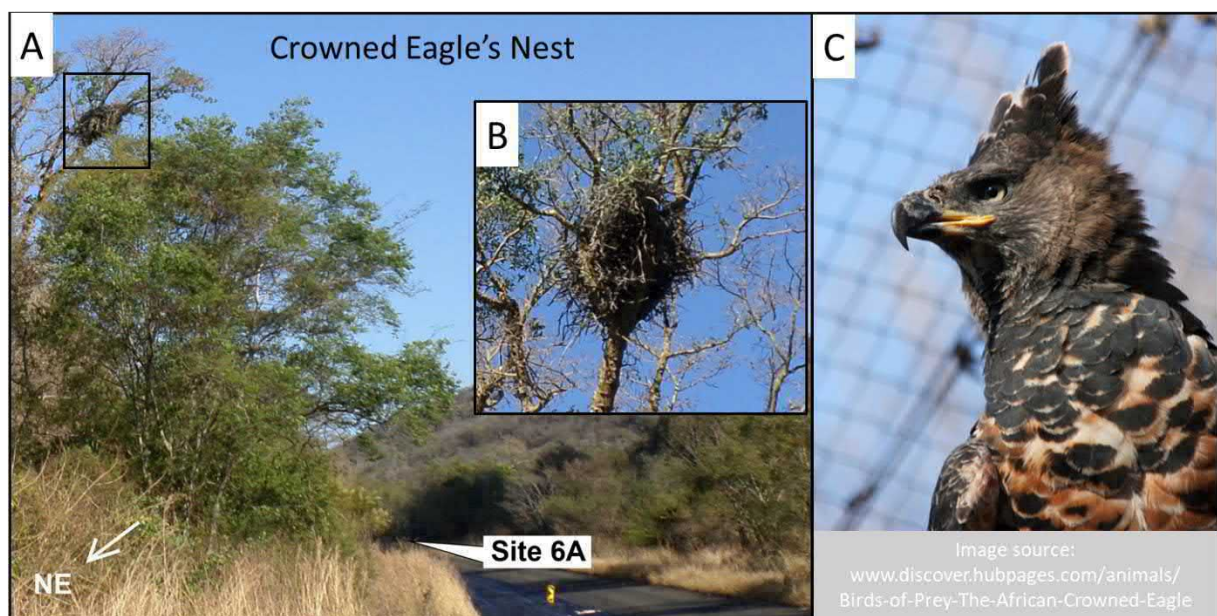


Figure 6.2 Nest of a bird of prey at Site 6A. (A): View of a crowned eagle's nest viewed from the north. The ridge exposed at Site 6 can be seen in the middle centre of the image. (B): Close view of the nest from below. (C): Image of an African crowned eagle, image source included in the picture.

6.3 Geology

6.3.1 Background

6.3.1.1 Lily Syncline

The Lily Syncline is the northernmost fold in a series of northeast-trending major folds of the BGB (Visser *et al.*, 1956). Like the other regional synclines, the Lily Syncline contains a core of clastic sediments of the Moodies Group (Visser *et al.*, 1956; Anhaeusser and Viljoen, 1965). Though the basal conglomerate of the Moodies Group was found, by previous workers, to be continuous along the Lily Syncline, characteristic repetitive sequences of the Group comprised of basal clastic sediments that grade vertically into suspension or chemical sedimentary rocks were not developed (Visser *et al.*, 1956; Viljoen, 1963; Anhaeusser, 1963). Instead, a single unit of clastic sedimentary rocks is commonly reported (Visser *et al.*, 1956; Viljoen, 1963; Anhaeusser, 1963). Also reported is the presence of a feldspathic quartzite and banded cherts that occur at different strike lengths of the core of the syncline (Keenan, 1986; Kohler, 2003). Outer parts of the syncline are composed of Fig Tree clastic and chemical sedimentary rocks or felsic schists underlain by mafic and ultramafic schists of the Onverwacht Group (Visser *et al.*, 1956; Viljoen, 1963; Anhaeusser, 1963; Kohler, 2003).

According to Visser *et al.* (1956), the Lily Syncline is an ENE-WSW-trending tight to close fold that extends for c. 40km along the northern margin of the Barberton Greenstone Belt. The central part of the syncline is marked by steep, narrow ridges collectively referred to as the Lily Ridge (Visser *et al.*, 1956; Anhaeusser, 1963; Kohler, 2003) (Fig. 6.1, Fig. 6.3). The Lily Ridge extends from the Noordkaap Consort Mine area in the west and the Scotman's Mine area southeast of the Three Sisters. At the eastern end, the syncline is a steep, tight fold of a southwards or northwards dip (Visser *et al.*, 1956). Towards the west, including at Sites 6A and B, the syncline becomes a southerly-dipping isoclinal fold. Further to the west - in the vicinity of the Scotia Talc Mine (Site 4) - the syncline is segmented into three ridges separated by the Bon Accord layered ultramafic complex (Visser *et al.*, 1956; Anhaeusser and Viljoen, 1965; Keenan, 1986).

Later work by Kohler (2003) proposed a distinction between the western and eastern segments of the Lily Syncline. The two segments are separated by the layer-discordant Louville Fault 5km east of Site 6B and are comprised of different stratigraphic successions (Kohler, 2003).

The bounding faults of the Lily Syncline are part of a series of regional high-angle faults that separate major NE-SW-trending synclines formed during the first regional shortening event of the BGB viz. D₁ of Ramsay (1963). Secondary fractures of these regional faults served as loci for gold deposits that have been mined for more than 140 years (Anhaeusser, 2013; Pearton and Viljoen, 2017). In particular, the Lily Mine occurs within the Lily Fault Zone, with mineralised zones ranging from lithological boundaries that present a competence contrast exploited by the Lily Fault, or secondary fractures within the main fault zone (Anhaeusser, 1963; Pearton and Viljoen, 2017).

Previous reports are consistent in reporting the coincidence of the Revolver Fault with the northern margins of the western and eastern segments of the Lily Syncline (Visser *et al.*, 1956; Anhaeusser and Viljoen, 1965; Kohler, 2003). This places the Lily Syncline at the southern margin of the Stentor Granite Pluton.

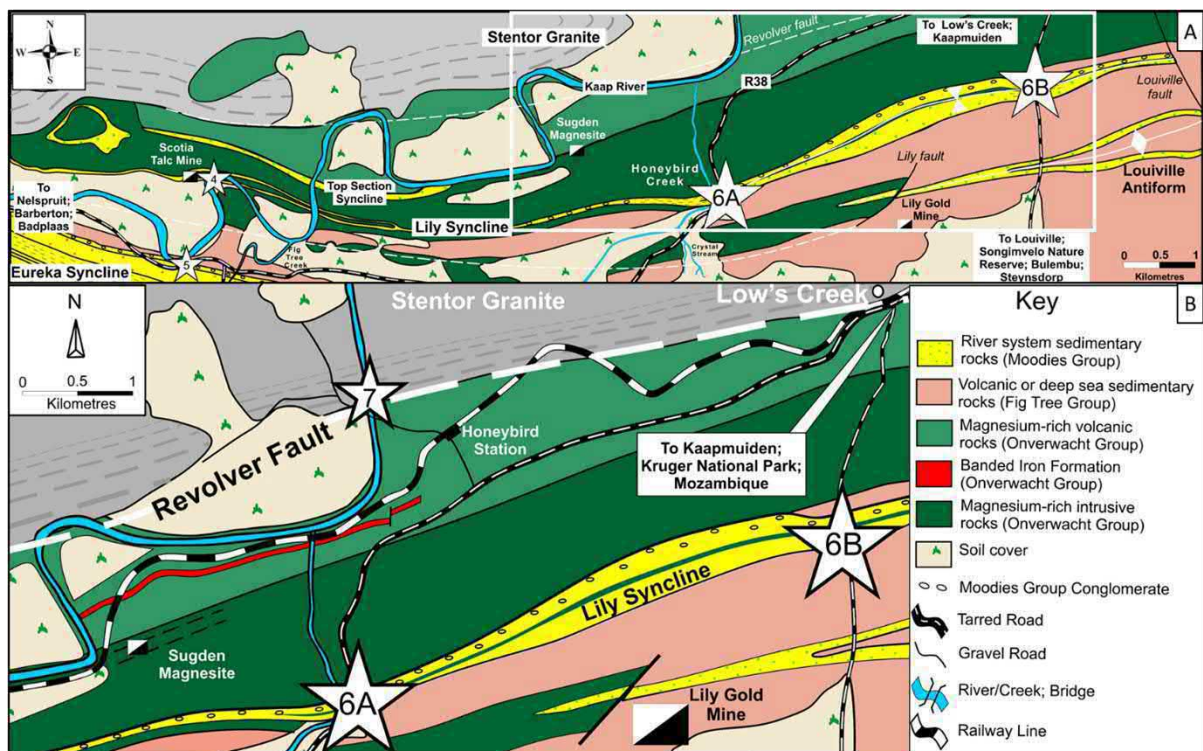


Figure 6.3 Simplified geological map of the northern margin of the Barberton Greenstone Belt. (A): Extent of the Lily Syncline along the southern margin of the Stentor Pluton. (B): The Moodies Group in the core of the Lily Syncline modified after Visser *et al.* (1956) to show the localisation of the Moodies Basal conglomerate within the Moodies Group on the northern extent of a portion of the Lily Syncline referred to by Kohler (2003) as the West Lily Ridge (that is, part of the segment of the syncline shown) between Site 6A and 6B.

The southern margin of the syncline is traced by different fault zones. In the eastern extremity, the syncline is bound by the high-angled Main Southern Fault (Fig. 6.4) and separated from the Eureka Syncline by the Louville Antiform (Kohler, 2003). However, the western extremity of the Lily Syncline does not include a southern limb. Instead, the northern limb is truncated in the south by the Southern Fault, on which rests the northern limb of the Eureka Syncline (Viljoen, 1963) (Fig. 6.4).

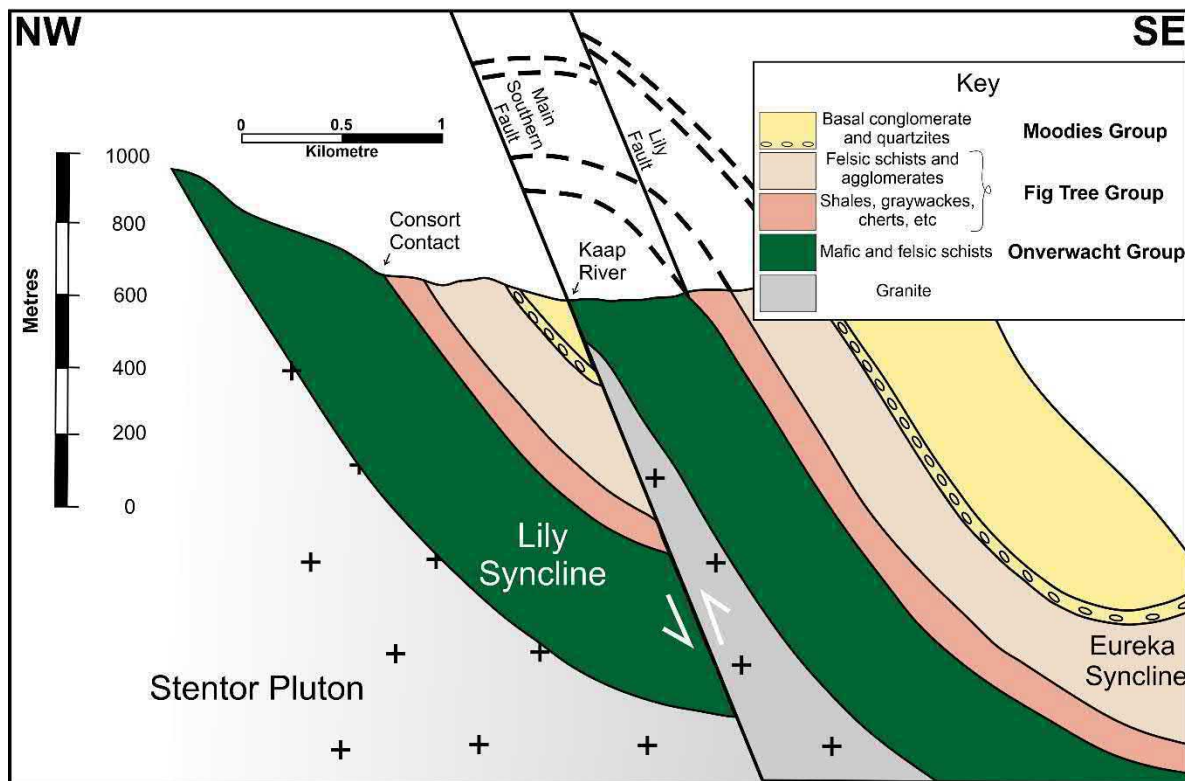


Figure 6.4 Cross section of the Lily and Eureka synclines near the western extent of the Lily Syncline. Redrawn after cross-section C-D of Figure 2 in Viljoen (1963).

The location of the Lily Syncline at the northern edge of the BGB within the contact aureole of the Stentor Pluton has resulted in a higher metamorphic grade and strain than in synclines such as the Eureka and Saddleback further south. Site 6A presents silicified lithologies that are difficult to correlate with Moodies clastic sedimentary rocks. Previous maps of the same area represent these as hornfels with a conglomeratic bed (Anhaeusser and Viljoen, 1965) or a banded chert (Keenan, 1986). Strain analysis by previous authors focused on conglomerate clasts to quantify the reported flattening and elongation of clasts at margins of the BGB (Ramsay, 1963; Viljoen, 1963; Anhaeusser, 1966a; Gay, 1969). Invariably, these studies found that the flattening strain of clasts, and the enclosing synclines, decreased drastically from 66% at the margin of the belt in contact with the Stentor pluton to 40% in central regions over a distance of ~15 km.

Previously identified marker beds of the Lily Syncline include a clast-supported basal conglomerate (shown in Fig. 6.6) and a feldspathic sandstone (Visser *et al.*, 1956; Anhaeusser and Viljoen, 1965).

6.3.2 Present study

6.3.2.1 Research question

The lithological succession of the core of the Lily ridge at the scenic roadside exposure designated as Site 6 is yet to be described.

6.3.2.2 Aim

To identify sites where deformed Moodies Group lithologies of the Lily Ridge are exposed along frequently used roads so that visitors can relate the scenic landscape to lithological differences.

6.3.2.3 Objectives

To describe the lithological units that make up Moodies Group roadside exposures of the Lily Syncline along the R38 road (Site 6A) and the road between Low's Creek and Kaapmuiden (Site 6B).

6.3.2.4 Methods

Where possible, traverse sections across strike of roadside lithologies exposed at Site 6A and 6B were accessed and described to identify roadside outcrops that would appeal to visitors. Where roadside outcrops of conglomeratic beds had exposed ZX planes to be seen by visitors, the principal clast axes *a* and *c* were measured across strike. This was to determine variation in clast size and axis ratios from the basal to upper parts of the conglomeratic bed. To preserve roadside outcrops for visitors, conglomerate clasts were not liberated from outcrops to measure lengths of the three principal axes used to describe strain of clasts at Site 6A. Also, clasts from this eastern extent of the Lily Syncline are silicified together with the matrix. In this state, previous workers found it impossible to liberate any clasts (Anhaeusser, 1964, 1966a).

Field descriptions were made, and thin sections of selected regional markers were produced to determine the mineral compositions. The markers of interest are a feldspathic quartz and chert wacke and a mafic schist underlying the extensive narrow valley at the centre of the Syncline. To describe the shape of grains in the Lily Syncline, thin sections of the wacke were cut along the ZX and XY planes.

Elemental maps of well-preserved parts of regional markers (from Site 6B) were produced to determine the elemental composition of characteristic minerals and identify element associations. The selected regional markers include a feldspathic sandstone and a chlorite schist that occupies the core of the Lily Ridge from Scotia Talc Mine in the west, to the eastern extremity beyond Site 6B.

6.4 Results

6.4.1 Field observations

The lithological succession of the core of the Lily ridge at the scenic roadside exposure designated at Sites 6A and B have not so far been described. A northwest to southeast traverse across sub-vertical strata of the Lily Syncline at Site 6A and Site 6B revealed a range of lithologies. The various lithologies are listed with coordinates in the accompanying appendix (Appendix D, Table 2). Respective distances of traverses across the Moodies core of the Lily Syncline (Lily Ridge) are c. 200 m at Site 6A and c.260 m at Site 6B. These traverses are segmented by a 50-100 m wide valley of dense tree cover and poor outcrop. This valley is a continuous longitudinal feature along the syncline and is the site of talc mineralisation and serpentinite carbonation at the Scotia Talc Mine (Site 4).

Site 6A

Roadside outcrops are sparse at this locality. The following descriptions account mostly for lithologies intercepted along a c.500 m long traverse mostly through an area northwest of Site 6A. The traverse was from the northwest (25°40.236'S; 31°14.417'E) to the southeast (25°40.479'S; 31°14.503'E) through dense forest cover in which the multiply-fractured, ENE-trending, sub-vertical (60-80° dip), lichen-ridden strata of the Lily Syncline occur. The lithologies are mentioned according to their appearance from the lowermost and oldest unit exposed north of the c. 200 m Lily Ridge, to strata just south of the core of the Lily Ridge i.e., the designated view site at Site 6A.

- The northernmost outcrop is a fuchsitic sericite schist interlayered with cm-scale zones of mud cracks and white quartz (Fig. 6.5 A). Layer-discordant veins of white quartz are ptygmatically folded about northeast-trending fold axes. The southern extent of this outcrop is interlayered with segmented or continuous cm-scale bands of carbonaceous shale (Fig. 6.5 A).

- Further south, a strike-parallel valley separates this outcrop from a sub-vertical grey, banded chert (Fig. 6.5 B) overlain by an iron-rich banded chert. Further south is a ferruginous shale with isolated, lenticular chert clasts (Fig. 6.5 C) that range in length from 2-5 cm and width between 0.3-1.5 cm. This unit occupies the low-lying area immediately north of the Lily Ridge.

From northwest to southeast along the younging direction of the northern strip of the Lily Ridge at Site 6A are the following lithologies:

- The northernmost unit is a wavy-bedded quartzite comprised of stacked, laminae of a 0.2 to 1.5 cm thickness range (Fig. 6.5 C). The quartzite beds are coated by <2 mm shale laminae and are overlain by >10 cm beds of carbonaceous shale.
- The carbonaceous shale is interlayered to the southeast by a quartzite along erosional contacts characterized by carbonaceous shale rip-up clasts within the quartzite (Fig. 6.5 D). The shale clasts are angular and mainly of clast lengths between 3 and 8 cm and widths between 0.8 to 1.5 cm.
- This interlayered sequence grades upwards into a thin-bedded quartzite <2 mm interlayered with thin ferruginous and carbonaceous shale layers. (Fig. 6.5 E). The latter sequence is separated from the adjacent strata in the southeast by a densely vegetated narrow valley filled with quartzite and shale boulders.

At the southern edge of the narrow valley is a grey quartzite in which dm-scale gravelly and pebbly conglomerate beds occur on a roadside outcrop east of the R38 road (25°40'29.46"S; 31°14'31.02"E). The few clasts identified in this unit range in length between 2 and 3.8 cm but clast widths are not exposed. All Site 6A lithologies are extensively altered. The clastic sediments are silicified while the chemical sediments appear to be devitrified.

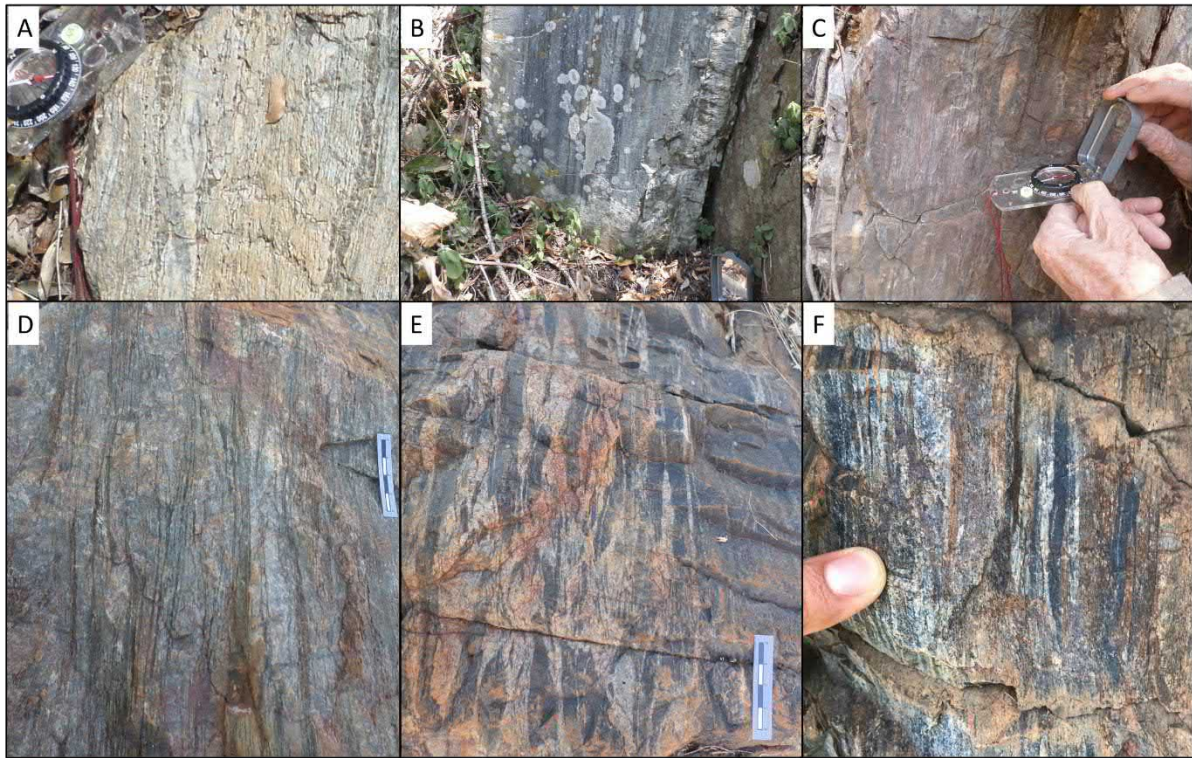


Figure 6.5 Lithologies of the Lily Syncline exposed across a north-to-south traverse through the Lily Ridge in the vicinity of the Honeybird Gorge (Site 6A). The younging direction is towards the left of each photograph. (A): Fuchsitic quartz-sericite schist interlayered with segmented mud layers (with mudcracks) and continuous white chert layers. (B): Banded chert (C): Ferruginous shale with isolated chert clasts. (D): Quartzite lenses coated by mud laminae. (E): Carbonaceous shale rip-up clasts in quartzite. (F): Chert interlayered with ferruginous shale.

Site 6B

Similar to Site 6A, a northeast-southwest roadside traverse at Site 6B allows access to older strata in the north and younger strata at the core of the Lily Ridge (Fig. 6.6). The lithologies are detailed in order of appearance from north to south along the traverse:

- The northernmost outcrop exposed is a massive, clast-supported boulder-cobble conglomerate. The clast population of this unit is dominated by chert variants and rare granitic lithologies. Conglomerate clasts are aligned with a sub-vertical cleavage that may dip steeper or shallower than bedding planes of the northern or southern sectors of the Lily Ridge. Texturally, the conglomerate clasts are poorly sorted, and clasts are sub-rounded to sub-angular (Fig. 6.6 A).
- Further south is a more matrix-supported cobble-pebble conglomerate (Fig. 6.6 B) that grades southwards into a feldspathic sandstone with isolated sub-angular clasts, interlayered with ferruginous shale (Fig. 6.6 C). The decreasing grain size from basal

to upper sections of the conglomeratic bed is coupled with a decrease in clast lengths and widths (c and a principal axes of chert clasts in Table 1).

- At the southern margin of the conglomeratic unit is a fault zone marked by a 100 m wide steep, densely-vegetated valley occupied by Quaternary cover and an isolated outcrop of a biotite-chlorite-muscovite-albite schist (Fig. 6.6 D).
- Beyond the valley is a sequence of black and grey and ferruginous chert with slump structures that indicate layer-parallel shortening.
- Further south is a fine to medium-grained quartz and chert wacke with a matrix rich in sericite and kaolinite (Fig. 6.6 E). Isolated, dm-scale lenticular zones of angular chert gravel (0.5-3 cm) are sparsely distributed within bedding planes of the sandstone.
- A valley separates the gravel-bearing wacke from thin-bedded grey shales that constitute the adjacent ridge that bears no conglomeratic bed (Fig. 6.6 F).

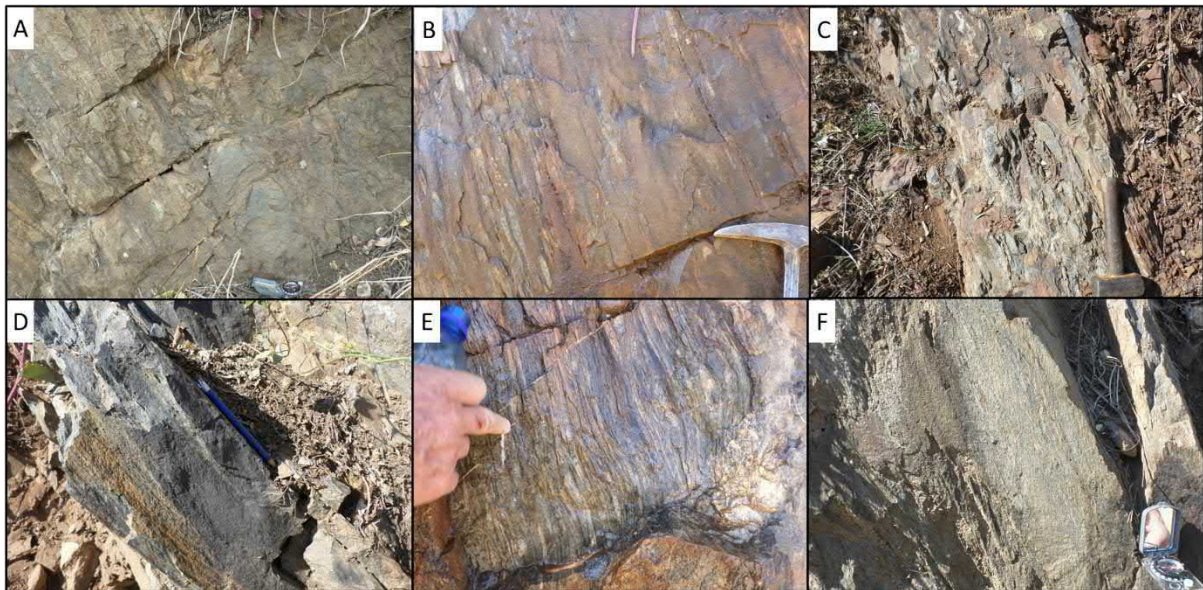


Figure 6.6 Lithologies of the Lily Syncline exposed at Louville from north to south. (A): Clast-supported cobble-boulder conglomerate. (B): Matrix-supported cobble-pebble conglomerate (C): Ferruginous shale interlayered with feldspathic sandstone. (D) Biotite-chlorite-muscovite-albite schist. (E): Slump-folded banded chert. (F): Layers of quartz and chert gravel in a sericitic wacke.

Table 6.1. Decreasing flattening strain in a conglomeratic bed along a 5 m north-to-south traverse across the northern limb of the Lily Syncline at Site 6B.

Ratios of principal clast axes (a/c) from ZX planes			
Entry no.	Site 1 (Coordinates: 31, 17.158 E; 25, 39.613 S)	Site 2 (Coordinates: 31, 17.153 E; 25, 39.647 S)	Site 3 (Coordinates: 31, 17.153 E; 25, 39.649 S)
1	3.47	7.69	4.55
2	4.78	4.93	5.63
3	5.50	5.76	3.33
4	32.00	6.94	13.00
5	5.07	5.45	7.71
6	3.29	5.78	2.08
7	1.47	5.63	3.95
8	5.50	2.11	9.60
9	3.81	9.17	4.77
10	4.53	10.25	5.29
11	5.35	3.25	4.98
12	12.80	5.17	
13	5.33	1.81	
14	1.93	5.00	
15	15.42	8.86	
16	9.91	3.16	
17	8.91	4.44	
18	3.15	10.47	
19	3.41	2.00	
20	8.91	24.89	
Mean	7.23	6.64	5.90

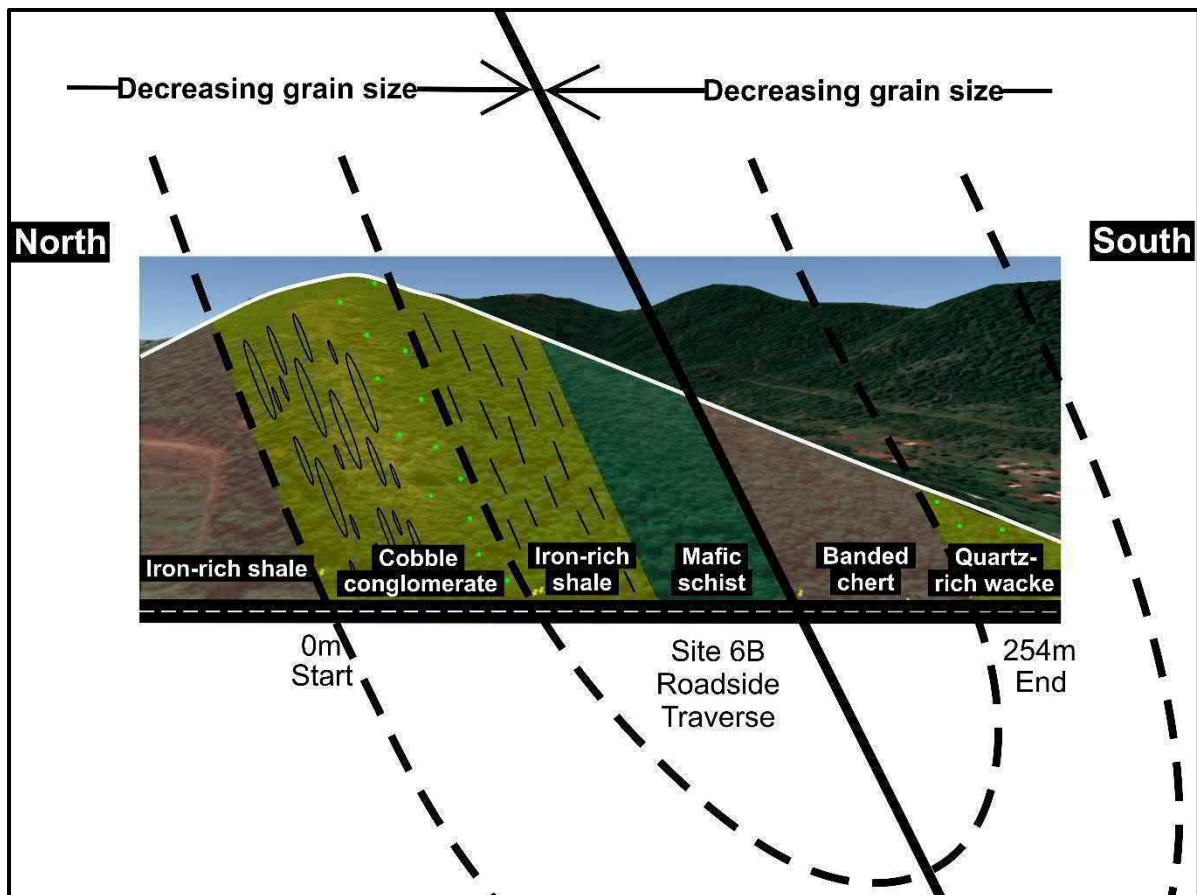


Figure 6.7. Lithologies exposed along the roadside traverse at site 6B. See text for descriptions. Dashed fold outlines are an interpretative model based on descriptions by Visser *et al.* (1956) and Kohler (2003).

6.4.2 Petrographic analysis

Thin sections of representative samples of the regional markers reveal the following. At Site 6B, the feldspathic sandstone is a chert, with angular to sub-rounded quartz clasts that are 2 to 5 mm long. It is comprised of 30% monocrystalline quartz with recrystallised margins, 15% polycrystalline quartz and 15% microcrystalline quartz (chert) in a 40% matrix. The matrix is made up of biotite, muscovite, sericite and monocrystalline quartz. The biotite is chloritised, mylonitised and interlayered with carbonaceous shale, forming drapes over grains. The mica-rich matrix is associated with tabular fragments of carbonaceous matter of up to 1 cm lengths, as well as skeletal nodules of ilmenite and fractured subhedral crystals of chromite.

The mafic schist is comprised of chloritised biotite, muscovite and feldspar interlayered with zones richer in muscovite and partially sericitised feldspar than biotite. Fine, euhedral crystals of ilmenite and chromite are uniformly distributed within the sample. The schistose fabric of the mafic schist is sheared into asymmetric microfolds (Fig. 6.8 D, E).

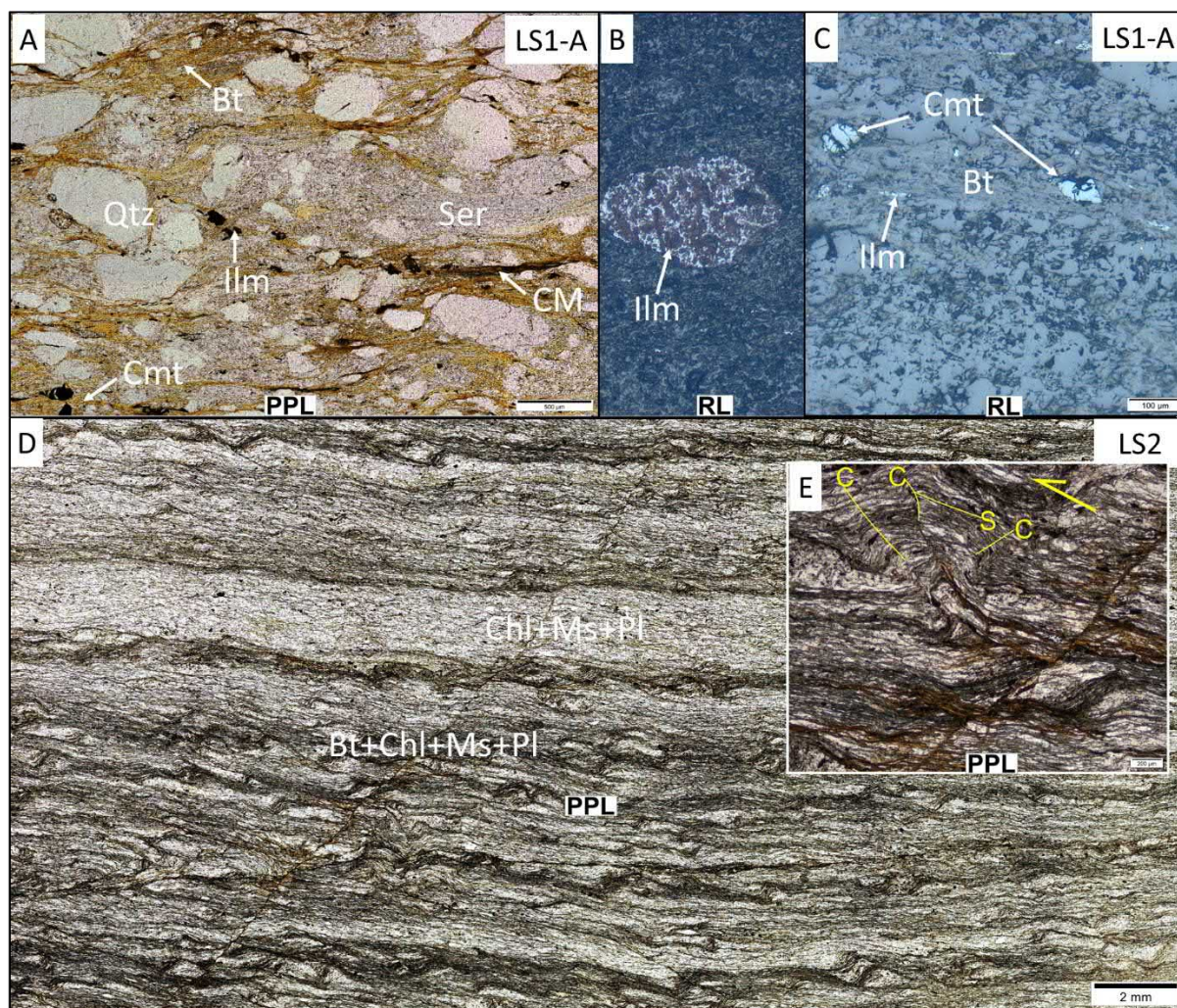


Figure 6.8 Mineral composition of markers in the Moodies Group core of the Lily Syncline. Sample names are in the top right corners of images and PPL, XPL and RL labels are in photomicrographs produced using plane-polarised, cross-polarised or reflected light, respectively. (A): Chert and quartz grains within a matrix of chloritised biotite and sericite. (B-C) Resorbed grains of ilmenite and chromite entrained in the matrix. (D): Compositionally layered mafic schist located at the core of the Lily Syncline. (E): Close up view of (D) shows asymmetrical, non-cylindrical microfolds in a compressional top-to-the-northeast shear sense. Note the radial crenulation lineations (C) associated with overfolded foliation (S). Bt: biotite, Qtz: quartz, Ilm: ilmenite, Cmt: chromite, Ser: sericite, CM: carbonaceous matter, Chl: chlorite, Ms: muscovite, Pl: plagioclase.

6.4.3 Elemental maps

Elemental maps of the regional markers show that the quartz-rich wacke (Fig. 6.9) is principally comprised of quartz and chert grains corresponding to discrete silica concentrations. The clasts occur within a biotite-sericite matrix comprised of higher aluminium, potassium and iron content than the enclosed clasts. Discrete patches of higher chromium, titanium and nickel concentrations in the matrix are related to the occurrence of spinels.

The mafic schist (Fig. 6.10) shows ubiquitous magnesium within chloritised biotite of the sample. Diffuse occurrences of chromium, iron, nickel and sulphur coincide with chloritised biotite. Discrete concentrations of chromium, nickel, iron and titanium represent uniformly distributed spinel crystals. Lenses of sheared plagioclase coincide with streaks of high calcium and aluminium content. Sulphur (within fine pyrite crystals) and silica occur mainly as diffuse components respectively within chloritised biotite and quartz-rich bands in the schist.

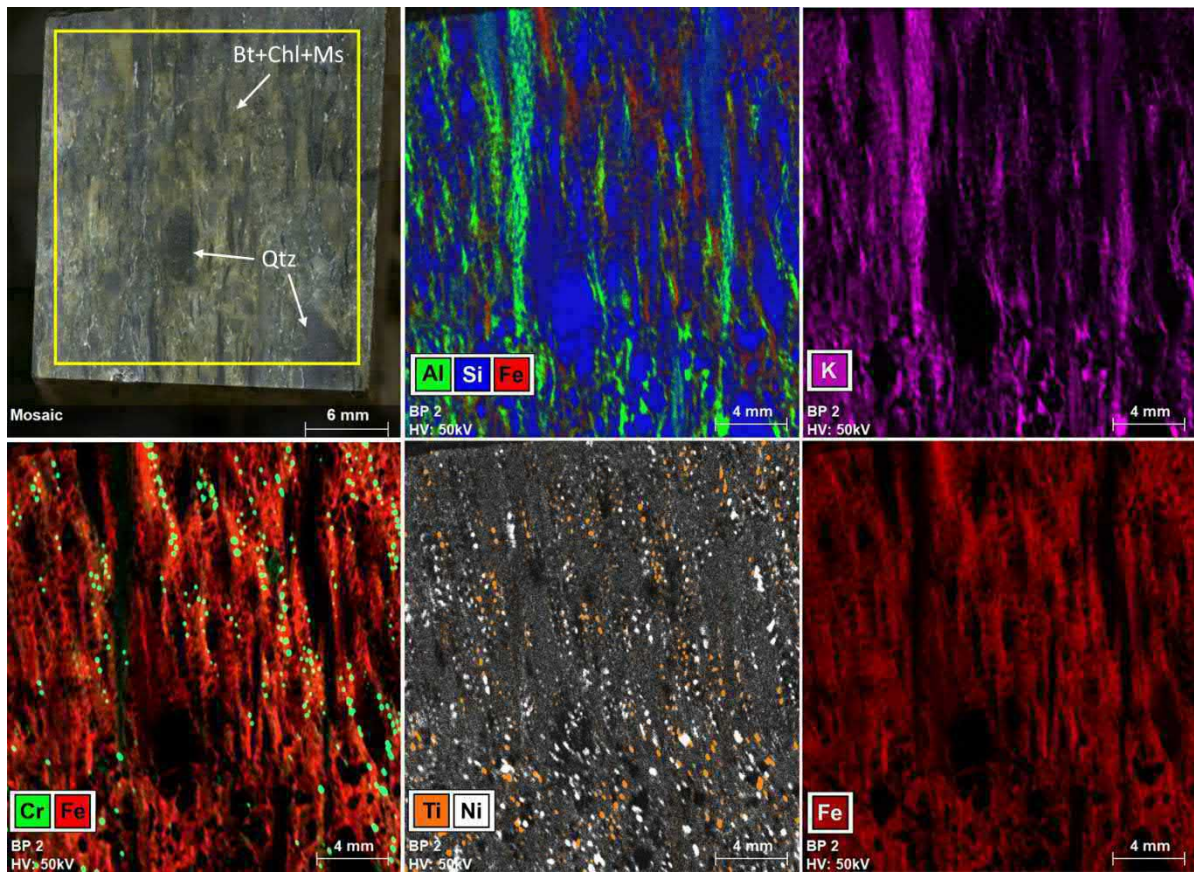


Figure 6.9 Element distribution maps of a representative sample of the altered feldspathic quartz-chert wacke marker (Sample NP-LS1C=BP). Maps show uniform distribution of discrete zones rich in chromium, nickel and titanium within the biotite-chlorite-muscovite/sericite matrix. Discrete concentrations of silica correspond to quartz and chert grains.

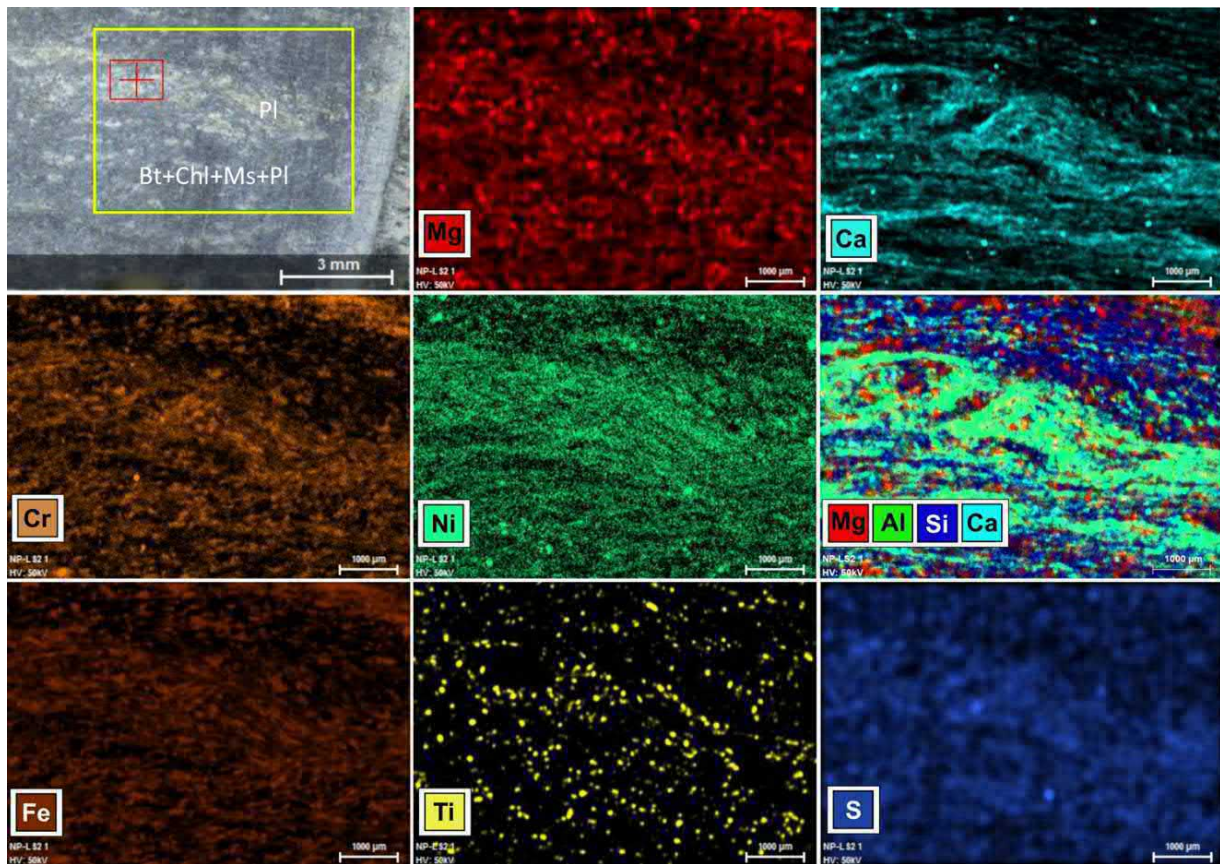


Figure 6.10 Element distribution maps of a representative sample of the mafic schist marker (NP-LS2). It shows that the uniformly distributed magnesium within chloritised biotite of the sample is associated with diffuse and discrete concentrations of chromium, nickel, iron and titanium spinels. Streaks of high calcium and aluminium concentrations correspond to sheared plagioclase lenses. Sulphur and silica only occur as diffuse components respectively within chloritised biotite and quartz-rich bands in the schist. The sulphur may represent fine pyrite crystals within the schist.

6.5 Discussion

In comparison to other regional synclines, there is only around 250 m of outcrop of the c.3.1-3.5 km Moodies Group strata (Anhaeusser, 1976b; Lowe and Heubeck, 1994) in the core of the tightly folded Lily Syncline at sites 6A and B. At both sites 6A and 6B the upward-fining sedimentary facies are marked by a succession of clastic units on the outer margins of the Lily Ridge that are overlain by fine-grained or chemical sedimentary strata.

At Site 6A, the sedimentary succession of the Lily Ridge overlies a sequence of fuchsitic schists (possibly metasomatised Onverwacht Group mafic tuffs) overlain by thin bedded grey and ferruginous shales presumably of the Belvue Road Formation of the Fig Tree Group. The Lily Ridge is comprised of quartzite lenses coated by mud drapes (Fig. 6.5. D) overlain by a rhythmic sedimentation sequence of shale and quartzite with shale rip-up clasts. The cyclic

shale-quartzite sequence grades into a micro-layered sequence of quartzite, ferruginous and carbonaceous shale/mudstone, reminiscent of rhythmites (Fig. 6.5 F). The former sequences indicate frequent changes in flow energy that would allow cyclic sedimentation of clastic and suspended sediments as is often observed in sub-tidal (Visser, 1980) and intertidal (Fenies *et al.*, 1999) depositional environments. The latter sequence presents quartzite-filled troughs in shale layers, indicating the limited settling of sand grains in a low flow energy environment where no dessication cracks were observed. Therefore, a likely depositional environment for this succession is a sub-tidal zone where strong ebb channels could have contributed clastic material to an aggrading sequence of suspended sediments (Williams, 1998). The occurrence of finely laminated ferruginous shale, carbonaceous shale and quartzite layers at the core of the Lily Ridge (i.e., stratigraphically uppermost position) may indicate basin deepening during landward transgression on an ancient coastline.

Duplication of lithological successions could not be established between the northwestern and southeastern sectors of the Lily Ridge due to poor outcrop south of the longitudinal fault zone at the core of the ridge. However, the highly silicified roadside shale outcrops at the southern extent of the Lily Ridge contain a rare and thin (<50cm) pebble conglomeratic bed. This further supports the concept that clastic sediments form the outer margins of the ridge and are overlain by fine-grained sedimentary strata.

Similarly, the outer margins of the Lily Ridge at Site 6B are comprised of clastic sedimentary rocks. The basal unit is a polymict boulder-cobble conglomerate that grades upward into a progressively more matrix-supported pebble conglomerate. All conglomeratic beds observed contain poorly sorted, sub-angular to sub-rounded clasts. This suggests deposition as a gravity-flow deposit such as an alluvial fan (Boggs, 2011). The conglomeratic bed in the northern limb of Site 6B does not occur at the same stratigraphic level of the southern limb at Site 6B, or at Site 6A. Also, no other conglomeratic bed was noted between Site 6B and the adjacent Louville antiform at the southern margin of the Lily Syncline (Fig. 6.3). Instead, shales and cherts dominate sequences south of the conglomeratic bed observed at the northern margin of the Lily Ridge at Site 6B. This suggests that Site 6B may be the distal zone of a northerly-based alluvial fan. Detritus from the fan may have initially prograded into a south-based water body prior to a landward transgression of the water body, as indicated by the upward fining trend in the conglomeratic beds (Table 6.1) and overall succession observed. This transgression may have been caused by northeast-up-southwest-down extensional deformation

at the southern margin of the Stentor Pluton (Chapter 7) as this contact zone forms the northern boundary of the Lily Syncline.

Grey and black chert, and ferruginous chert towards the south occupy the core of the ridge. The southernmost extent of the ridge is a gravelly quartz-rich, chert-biotite-sericite wacke (Fig. 6.6 F) with no underlying conglomeratic bed. This indicates that the texturally immature boulder- to pebble conglomeratic beds at the northernmost extent of the ridge are not duplicated at this locality. Also indicated are the upward and lateral fining trends of the lithological succession from north to south. This, according to Boggs (2011) represents a hiatus in the supply of boulders. Alternatively, the retrograding sequence could indicate transgression of the basin due to basin deepening (e.g., Heubeck and Lowe, 1994). Also implied is a northerly source terrain for the conglomeratic beds of this locality which is here confined to the northern limb of the ridge. The latter is in direct contrast with observations of Eriksson (1978) who identified southward sediment transport directions in lower Moodies Group strata. This suggests that lower Moodies Group sediments may have been deposited in different basins (Heubeck and Lowe, 1994; Heubeck, 2013, 2018).

The upward-fining trends of basal Moodies Group successions of the Lily Ridge at Site 6A and 6B are consistent with observations at other regional synclines (Anhaeusser, 1963, 1976b, 2019; Heubeck and Lowe, 1994; Heubeck, 2019). This consistent trend led previous workers to conclude that the deposition of the lower units of the Moodies Group occurred within one basin prior to syndepositional uplift that led to the isolation of mountain-bound basins (Heubeck and Lowe, 1994; Heubeck, 2013, 2018). This was understood to have led to the local variations in lithological and structural characteristics of upper Moodies Group sediments in different regional synclines (Heubeck and Lowe, 1994).

The mineralogy of the two regional markers, the feldspathic quartzite and the talc-bearing serpentinite, differ slightly at Site 6B to the existing descriptions by Visser *et al.* (1956). The feldspathic quartzite noted in the western Lily Ridge (Visser *et al.*, 1956) occurs at Site 6B as a quartz-bearing wacke with a sericite-rich matrix (NP-LS1A-C). If the quartz-bearing arkose reported at other localities correlates with the quartz-rich, sericite wacke at Site 6B, sericitisation would be the cause of the difference in mineral assemblage. If, however, these two clastic units are not related, the high sericite component in the matrix of the wacke may have been derived from the reworking of felsic schists and volcaniclastics of the uppermost formations of the Fig Tree Group (Bien Venue or Schoongezicht).

The other marker, a talc-bearing serpentinite which is mined at Scotia Talc Mine only occurs as a biotite-chlorite-muscovite-albite schist (NP-LS2) at Site 6B where it is sheared sub-vertically in a compressional top-to-the-northeast sense. This shear sense is consistent with that noted in sheared metavolcanics observed by Kohler (2003) along the Scotsman's Fault at the northern margin of the eastern Lily Syncline.

If the interpretative fold structure drawn in Fig. 6.7 is valid, the boulder conglomerate in the northernmost outcrop of Site 6B may be an alluvial fan deposit in the northern limb that grades southwards into a gravelly wacke in the southern limb. The same fan deposit grades southwestwards into a basal quartzite at the northern and southern margins of the Lily Ridge at Site 6A.

The lack of bedding-cleavage relationships that systematically define folded and overfolded limbs of the isoclinal, south-dipping western Lily Syncline (Visser *et al.*, 1956; Kohler, 2003) may indicate the presence of axial planar, as well as post-folding slaty cleavage. The latter may have developed obliquely to bedding and axial planes of regional synclines (e.g., Eureka Syncline, Ramsay, 1963; Anhaeusser, 1976b). For this reason, the present study cannot confirm whether the observed strata constitute the northern and southern limbs of tightly folded Moodies Group strata in the core of the Lily Syncline. What can be deduced, however, is the upward fining succession at the core of the Lily Syncline as observed from the exposed margins and cores of the Lily Ridge.

6.6 Conclusion

In the absence of preserved palaeocurrent indicators, textural maturity and lithological successions were guiding factors in making deductions about basin characteristics suggested by the strata preserved at Sites 6A and B. Generally, the basal gravel deposits observed were subsequently overlain by extensive fine-grained or chemical sediments due to basin deepening or a change of material in the source region(s).

6.7 Recommendation

Though a northerly source terrain for Moodies Group detritus is suggested for Site 6B, more detailed studies could conduct traverses across the Lily Ridge to provide certainty.

Chapter 7: Stentor Pluton contact zone

Bridge: (25°38'57.79"S; 31°14'33.36"E); Parking: (25°38'57.75"S; 31°14'32.35"E)

7.1 Site description

Site 7 is in Honeybird Siding on both sides of a bridge over the Kaap River (Fig. 7.1). Good parking spots are on the western side of the river where recent land-clearing has provided more flat ground. Large trucks also use this road, so it is best to park vehicles away from road edges. Rock exposures of interest are mostly >1.5 m above the flood water level of the river and are accessible from 70 m south, to 200 m north of the Kaap River bridge. With permission from farm owners (dortan@soft.co.za), visitors may walk down from the bridge towards outcrops exposed along the western side of the river. The eastern side is fenced off, but gate entry is possible by prior arrangement. Visitors are to be vigilant and maintain a safe distance and elevation from the river as crocodiles and monitors are known to occupy this locality. Littering is prohibited.

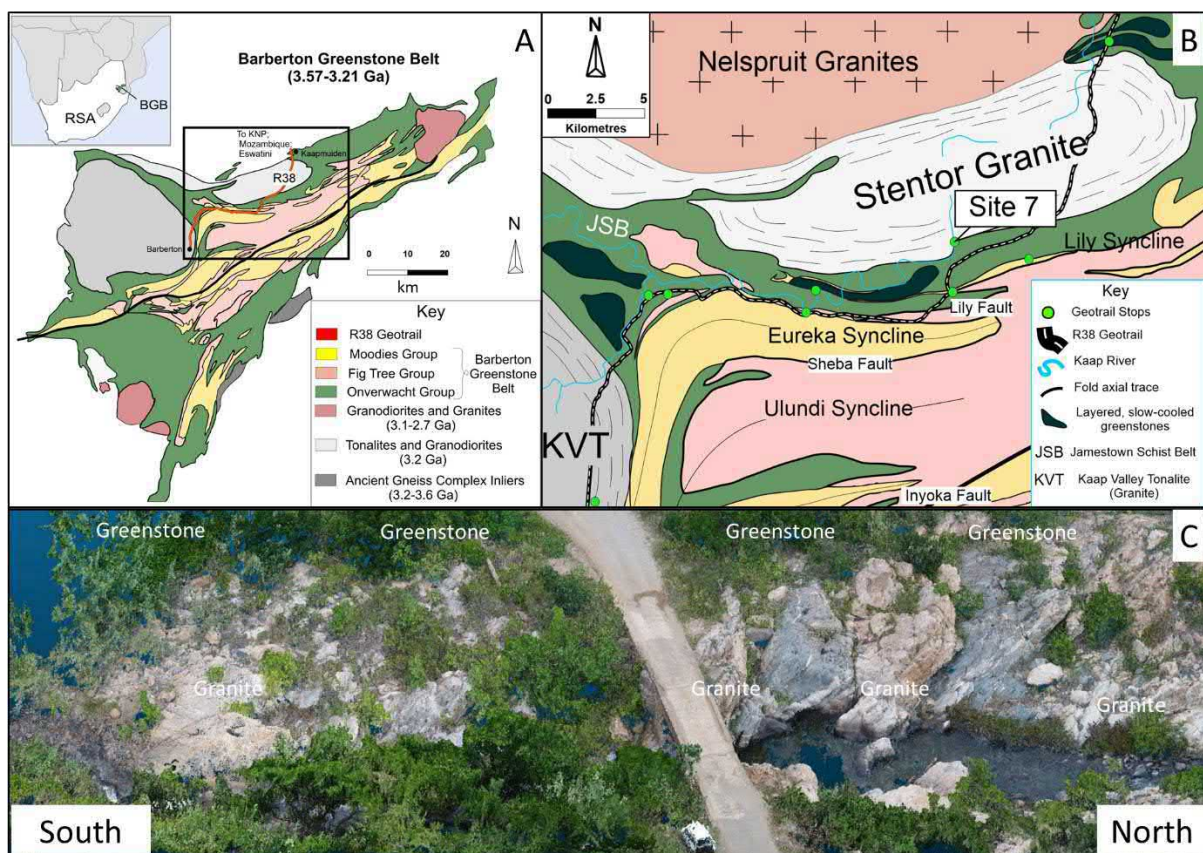
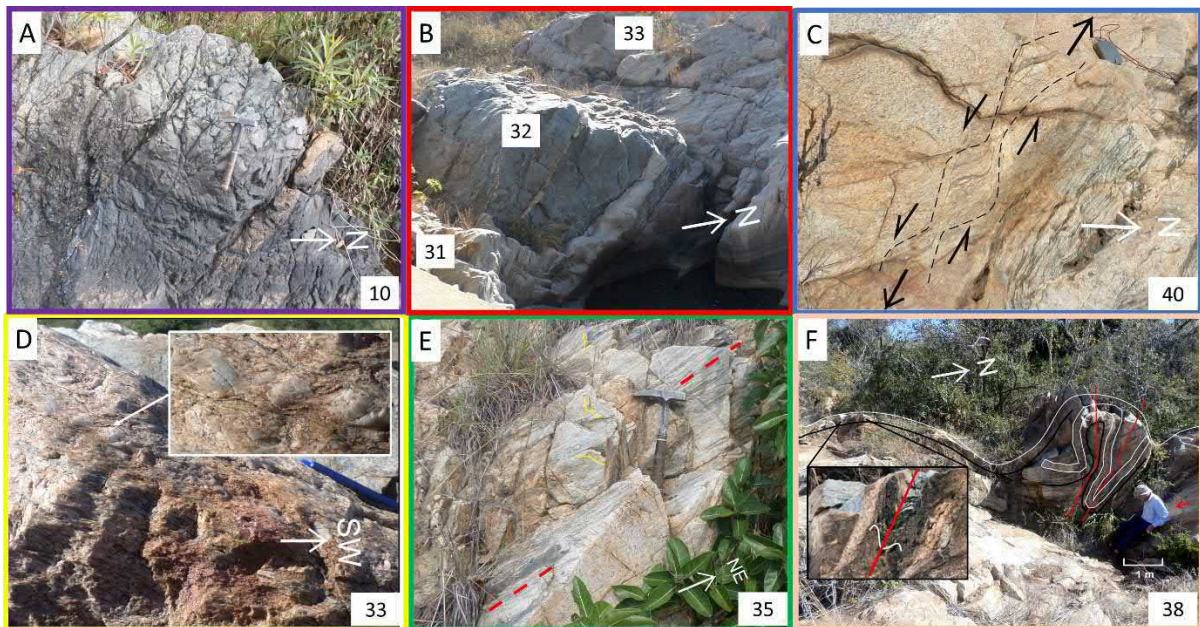


Figure 7.1. Main features of Site 7. (A): Simplified geological map of the R38 Geotrail in the context of the BGB. (B): a detailed view of the black rectangle in (A) shows the location of Site 7 at the southern margin of the Stentor Pluton which is contact zone spanning at least 200 m shown in (C). (C): View of riverside outcrops on the Stentor Pluton contact zone exposed on the western side of the Kaap River (note bakkie in lower centre for scale). Maps modified after Visser *et al.* (1956) and Anhaeusser *et al.* (1981); image captured from a 3-D model generated from drone images acquired by Thomas Jones.

7.2 Aspects of the Stentor Pluton and Site 7 that may interest laypersons

Preserved at this locality are rock fabrics visible from the main bridge and from easily accessible outcrops (Fig. 7.3). These fabrics and the phases that they occur in may be presented to visitors as an example of clues that tell us about the stabilisation of early landmasses of the Earth (Robb and Anhaeusser, 1980; Tegtmeier and Kröner, 1987; De Ronde and De Wit, 1994; Poujol et al. 2003).



Sites where rocks above can be seen



Figure 7.2 Characteristic rock fabrics at Honeybird Siding. Colours of photograph borders are coordinated with stars marking locations of outcrops photographed. Numbers in bottom right corners of images A-F indicate unit numbers used in a map included in later sections of this chapter (Fig. 7.6). Aerial image captured from a 3-D model created by Thomas Jones.

Zones of weakness in the Stentor Granite generated through fracturing have provided easier access for agents of weathering and erosion to generate boulders through spheroidal weathering

(Ollier, 1971). As early as the 1960s and 70s, art enthusiasts were visiting boulders of the Stentor Pluton exposed along the R38 road opposite a turnoff to Esperado (Site: 25°36'54.55"S, 31°18'26.76"E; Parking: 25°36'52.94"S, 31°18'27.24"E). They were drawn by designs painted on granite boulders by a self-trained artist named Nukain Mabuza (died in 1981). Nukain transformed his home and a granite hill on the side of his residence into a 'Stone Garden'. This he did by painting artistic figures and patterns on his hut, nearby bedrock and loose boulders of the Stentor Pluton. For doing so, he became one of two South Africans whose self-taught artworks would be recognised as one of few (44 in 1992) environmental artworks of the world (Bornman, 2002). Unfortunately, Nukain had already passed on when his work attained international acclaim. In his honour, Barberton gateways were painted in designs inspired by the original 'Stone Garden' he created. Aspects of his life were also included in a play by internationally acclaimed playwright, Athol Fugard titled 'Painted Rocks at Revolver Creek'. Coincidentally, the play was shown in Cape Town during the 35th International Geological Congress in 2016.



Figure 7.3 Painted boulders of the Stentor Pluton as seen from the R38.

7.3 Geology

7.3.1 Background

7.3.1.1 Stentor Pluton

The Stentor Pluton is a composite pluton that intrudes the northern margin of the Barberton Greenstone Belt (Anhaeusser and Viljoen, 1965; Anhaeusser, 1966b; Tegtmeyer and Kröner, 1987; Kohler, 2003). It extends for c. 30 km and has a width of c. 7 km. Though elusive in the north, the margins of the pluton occur as south-dipping shear zones along which strata in the eastern margin of the pluton were displaced to the east (Kohler, 2003; Anhaeusser, 2019). The southern margin coincides with the Revolver Shear Zone, while the northern margin is poorly resolved but partially coincides with the Stentor Fault in the northeastern margin (Visser *et al.*, 1956; Viljoen and Viljoen, 1970; Kohler, 2003; Anhaeusser, 2019). The northern extent of the pluton has not been well defined but is marked by migmatites and granodioritic to granitic phases of the Nelspruit Batholith (Visser, 1956). The southern margin of the Stentor Pluton at Site 7 is a scenic contact zone (Fig. 7.1, 7.2) of interleaving blocks of granite and country rock exposed along the perennial Kaap River (Anhaeusser and Viljoen, 1965; De Ronde and De Wit, 1994; Kohler, 2003; Dziggel *et al.*, 2006).

7.3.1.2 Contact Zone Phases

The southern contact of the Stentor Pluton at Honeybird Siding forms part of the top-down-and-to-the-south dextral Honeybird Shear Zone (HSZ) (De Ronde and De Wit, 1994; Dziggel *et al.*, 2006; Schoene *et al.*, 2008; Anhaeusser, 2019). Ortho- and paragneiss phases of the Stentor Pluton form the footwall of the HSZ while the northern sections of the Lily Syncline form the hangingwall (Schoene *et al.*, 2008). At Site 7, the HSZ is characterized by sheets of gneissic phases interleaved with blocks of schistose amphibolites of the Onverwacht Group (Anhaeusser and Viljoen, 1965). From south to north, the schists occur as bedrock, and gradually become sparsely distributed enclaves that phase out with increasing distance from the contact zone (Anhaeusser, 1968).

Geochemical analyses by Kohler (2003) classified the felsic intrusive sheets of the pluton into a spectrum of *sensu stricto* granites and granodiorites (Fig. 7.4.). Texturally, the granite sheets were found to range between medium-grained banded and foliated gneisses to coarse-grained feldspar augen gneisses.

Two main granitic phases of the Stentor Pluton were identified at Site 7 (Viljoen, 1964; Schoene *et al.*, 2008). The older phase is a composite unit comprised of a banded gneiss with

amphibolitic, tonalitic and granodioritic layers and sheets. The younger phase is a compositionally homogeneous granodiorite that intrudes the older composite phase (Viljoen, 1964; Anhaeusser, 1966b; Schoene *et al.*, 2008).

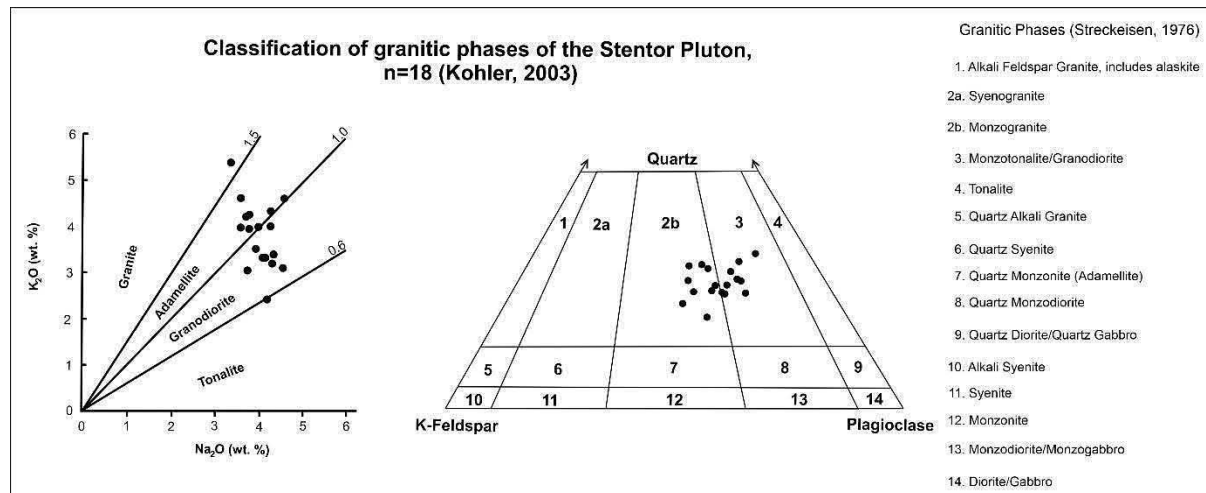


Figure 7.4 Range of compositions and relative proportions of felsic minerals of the Stentor Pluton granite sheets. These compositions were worked out using XRF chemical data (left) and modal abundances noted during petrographic analysis (right). Modified after Kohler (2003).

7.3.1.3 Age

U-Pb ages of c. 3347 Ma and c. 3250 Ma were determined from an equigranular phase and an augen gneiss phase, respectively (Tegtmeyer and Kröner, 1987). These ages were understood to indicate the composite nature of the pluton or the presence of xenocrystic zircon from a c.3347 Ma basement (Tegtmeyer and Kröner, 1987). Significantly younger U-Pb ages (c.3107-3097 Ma) were reported by Kamo and Davis (1994) and Zeh *et al.* (2009). These younger ages overlap with phases of the c.3106-2740 Ma Nelspruit Batholith situated in the northern and western contacts of the pluton (Kamo and Davis, 1994). Subsequently, a joint study of field relationships and radiometric dating by Schoene *et al.* (2008) identified a c.3258 Ma banded gneiss phase cross-cut by a c.3106 Ma homogeneous gneiss phase at Site 7. Both phases are intruded by c.3104 Ma folded pegmatitic dykes (Schoene *et al.*, 2008).

7.3.1.4 Main Fabrics

Work by Fripp *et al.* (1980) described this contact zone as a northeast-verging ductile thrust zone. Subsequently, De Ronde and De Wit (1994) proposed a sinistral transtensional emplacement regime. Thereafter, Dziggel *et al.* (2006) suggested an unroofing and ascent of the Stentor Pluton along an extensional detachment aligned with a northeast-southwest-

trending orogeny associated with the c.3230 Ma terrane accretion of northern and southern BGB domains. According to Dziggel *et al.*, (2006), the granitoids of the northern terrane of the BGB were emplaced after peak metamorphism of the BGB, which is associated with the c.3230 Ma terrane accretion event. The same authors proposed that the accretionary event resulted in an overthickened northeast-southwest-trending orogen that underwent extensional collapse leading to exhumation of the Stentor Pluton along an extensional detachment. The unroofing facilitated solid-state emplacement of northern-terrane granitoids against BGB supracrustals (Dziggel *et al.*, 2006). It is during this last phase of three TTG magmatic events of the Barberton Mountain Land that the c.3258-3104 Ma Stentor, c.3227 Ma Kaap Valley and c.3230 Ma Nelshoogte Plutons were emplaced (Kamo and Davis, 1994).

Marginal sectors of the Stentor Pluton and mafic schist country rocks are LS-tectonites (Dziggel *et al.*, 2006). Similar to the pluton, these schists preserve contact-parallel, south-to-southeast-dipping foliation formed by flattened biotite and quartzofeldspathic aggregates (Fripp, 1980; De Ronde and De Wit, 1994; Kohler, 2003; Dziggel *et al.*, 2006; Schoene *et al.*, 2008). According to the same authors, this schistosity is accompanied by mineral stretching lineations of a moderate to shallow southwest plunge. In a study area mainly based in the southwestern margin of the pluton, but inclusive of Site 7, Dziggel *et al.* (2006) found that lineations plunge southwestwards, parallel to fold axes of intrafolial isoclinal folds that indicate sinistral, top-down-and-to-the-south extension associated with the emplacement of the pluton. Later work by Schoene *et al.* (2008) also documents southwest-plunging stretching lineations in all Site 7 lithologies. However, their assessment of rare 3-D outcrops indicated a dextral top down-and-to-the-south shear sense. This variation in shear sense along the same contact zone was understood to result from localised heterogeneities (Schoene *et al.*, 2008). Images of folded foliation described above are here labelled as S₂ folded about southwest-plunging F₁ fold axes (Fig 7.5). These labels are not a reference to deformation events of the BGB. They are used here to denote present fabrics at Site 7 that are understood to predate igneous layering (S₀) and unfolded gneissosity or schistosity (S₁).

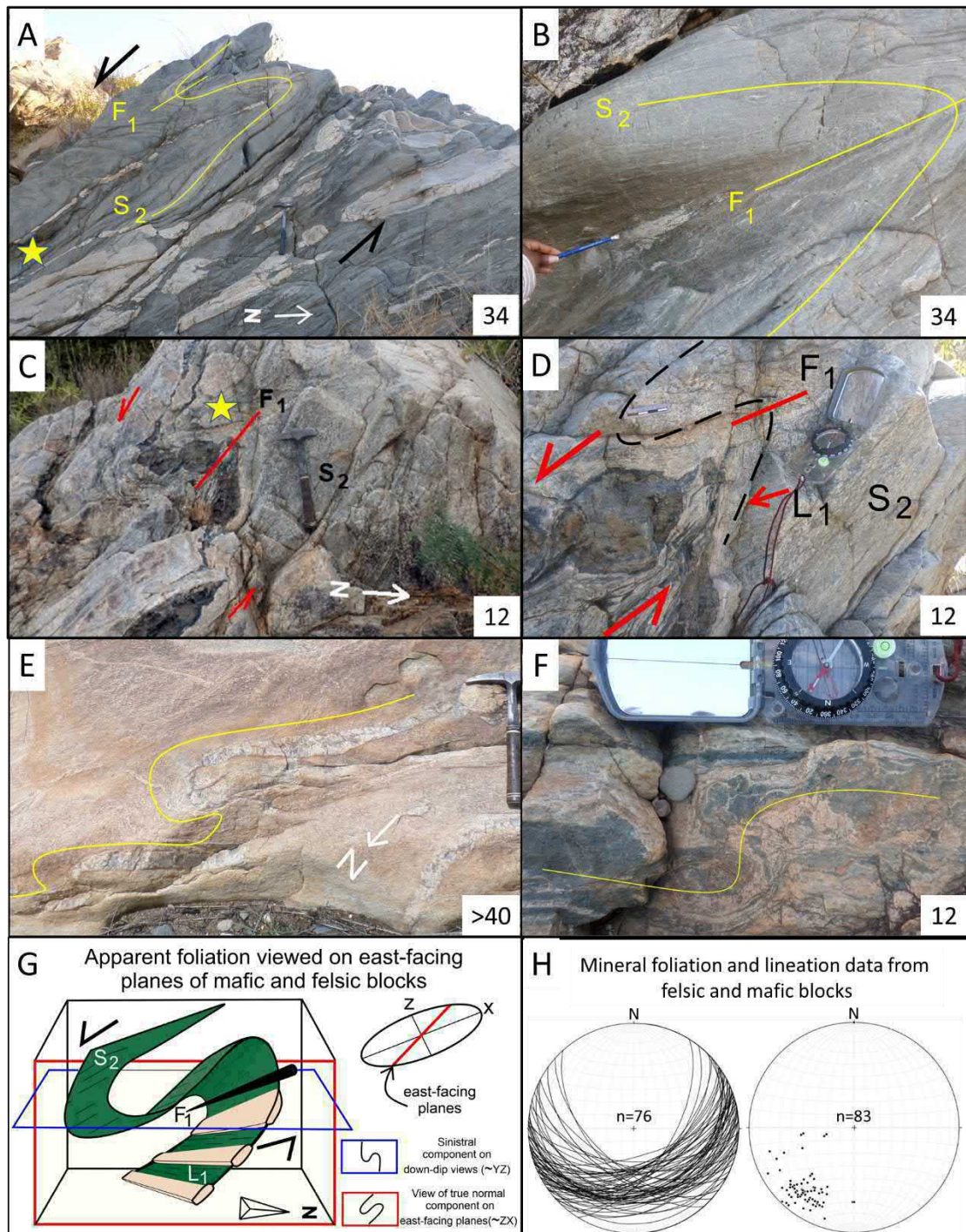


Figure 7.5 Strain associated with the emplacement of the Stentor Pluton at Honeybird Siding. Content based on work of some previous authors which was confirmed by observations of the present study. Unit numbers are shown in the bottom right corner. Subscripts on feature labels relate to a local deformation sequence. (A-D): Approximate ZX planes of metavolcanic and granitoid sheets both show top-to-the-SE extension and the parallel relation of fold axes and mineral lineations. Yellow stars in (A) and (C) mark sites of images (B) and (D) respectively. Note blue pencil approximating the fold axis in (B). (E-F): Examples of top-to-the-southeast displacement on down-dip views (YZ planes) of folds at Site 7, suggesting a (local) sinistral component to the top-down-and-to-the-southeast extension seen in A-D. (G): Block diagram of ductile strain in mafic and felsic sheets shows context of SW-plunging boudins and fold axes. (H): Orientation data of main fabrics on equal-area stereonet plots.

7.3.1.5 Metamorphism

The c. 1km contact zone from the southern margin of the Stentor Pluton to northern sectors of the BGB contains abrupt changes in metamorphic grade, with amphibolite to greenschist facies transitions over a few hundred meters (Dziggel *et al.*, 2006). Peak metamorphic assemblage of the mafic metavolcanic schists adjacent to the pluton includes hornblende, clinopyroxene, plagioclase, garnet $\pm$ quartz (Dziggel *et al.*, 2006). Retrogression observed by the same author is marked by the symplectite intergrowths of epidote and quartz after clinopyroxene. A clockwise P-T path determined by the same author was understood to indicate contact-parallel exhumation of the mafic schists from a depth between 15 and 22 km. This followed prograde metamorphism at c. 600–700° C and 5 ± 1 kbar (Dziggel *et al.*, 2006).

7.3.2 Present study

7.3.2.1 Research question

Previous geological descriptions of Site 7 were focused on the Stentor Granite. However, no petrographic descriptions of the amphibolite sheets interleaved with the pluton have been produced. Such descriptions coupled with existing knowledge are yet to be translated into a phase map with marked locations of key outcrops that preserve defining features of the southeastern contact zone of the Stentor Pluton.

7.3.2.2 Aim

To compile new and existing information about properties of the Stentor Pluton contact zone at Site 7 in order to produce a phase map and diagrams charting the locations of key features to be seen by visitors.

7.3.2.3 Objectives

Literature review and petrographic analysis of grey-green sheets to add to existing descriptions; field mapping to locate key outcrops that preserve characteristic fabrics of the pluton and the production of collages that summarise features of the site.

7.3.2.4 Methods

Phase mapping of a c.200 x 100 m² area suitable for visitors was carried out on Google Earth aerial photographs and later refined using high resolution aerial photographs (1 pixel: 7.7mm

on the ground) obtained from drone imaging. Orientations of mineral stretching lineations and mineral foliation and banding were recorded to populate the phase map and inform interpretative diagrams.

Mineral composition and modal abundances of main mineral components were described using a petrographic microscope as well as full-thin section photomicrographs acquired in the Wits Imaging Lab. Phase maps were generated for altered phases using the TESCAN-integrated mineral analysis instrument at the Wits Automated Lab. Elemental maps were generated using a micro-XRF instrument at the Friedrich-Schiller University of Jena, Germany. Details of analytical techniques are included in the ‘General Methods’ section.

7.4 Results

7.4.1 Field observations: Characteristic Phases

Key outcrop exposures of this site occur along a 300 m north-south riverside traverse across various lithological boundaries. This traverse extends from c. 70 m south of the main bridge, and c.200 m north thereof, providing access to shear-sense indicators of ductile strain associated with the emplacement of the Stentor Granite.

All variants of mafic lithologies to be seen at Site 7 can be accessed along a south to north riverside traverse on the western side of the Kaap River (Fig. 7.6). From south to north, they occur as highly fractured green-grey chlorite gneisses (10, 13) and amphibolites (32, 34, 38F). These range from bedrock interleaved with granitic sheets in the south, to variably sized folded enclaves of decimetre to metre scale in the north. Granitic sills are commonly folded with the schistose fabric of the mafic sheets (Fig. 7.5 A-B). Also, planar mafic enclaves in the granitic sheets are folded together with the gneissic fabric of granitic sheets (Fig. 7.5 C-D). The folds in both cases are isoclinal and parallel to lithological contacts. In rare 3-D exposures, the folds appear to indicate local sinistral, top-to-the southeast transcurrent shearing (Fig. 7.6 G).

Felsic lithologies exposed along the same traverse mentioned above include a trondhjemitic augen gneiss (12) with a migmatitic southern margin. The migmatitic horizons are comprised of boudinaged or tabular bands of amphibolite within quartzofeldspathic layers dominated by albite. Northern horizons of this gneiss contain millimetre-scale augens. Further north of 12 is a massive, recrystallised granodiorite sheet (31). Further northwards, and interleaved with schistose amphibolites (32 and 34), is a foliated granodiorite sheet with centimetre-scale augens (33). North of 33, exposures are trondhjemitic-to-granodioritic sheets with strongly-developed southwest-plunging quartzofeldspathic stretching lineations smeared with biotite layers (35, 37). Further north, units 35 and 37 are intruded by moderately foliated granodiorite sheets (36/38), both of which host partially migmatized (37) or brecciated and digested (39) mafic sheets. In terms of strain distribution, the homogeneous potassic granite-to-granodioritic phases (31, 36 and 38 in Fig. 7.6) have weakly- to moderately-developed biotite foliation and quartzofeldspathic stretching lineations compared to the strongly lineated trondhjemitic-to-granodioritic heterogeneous gneisses they intrude (35, 37). Mafic lithologies at this locality coincide with higher strain zones and commonly coincide with local stream channels (Fig. 7.6).

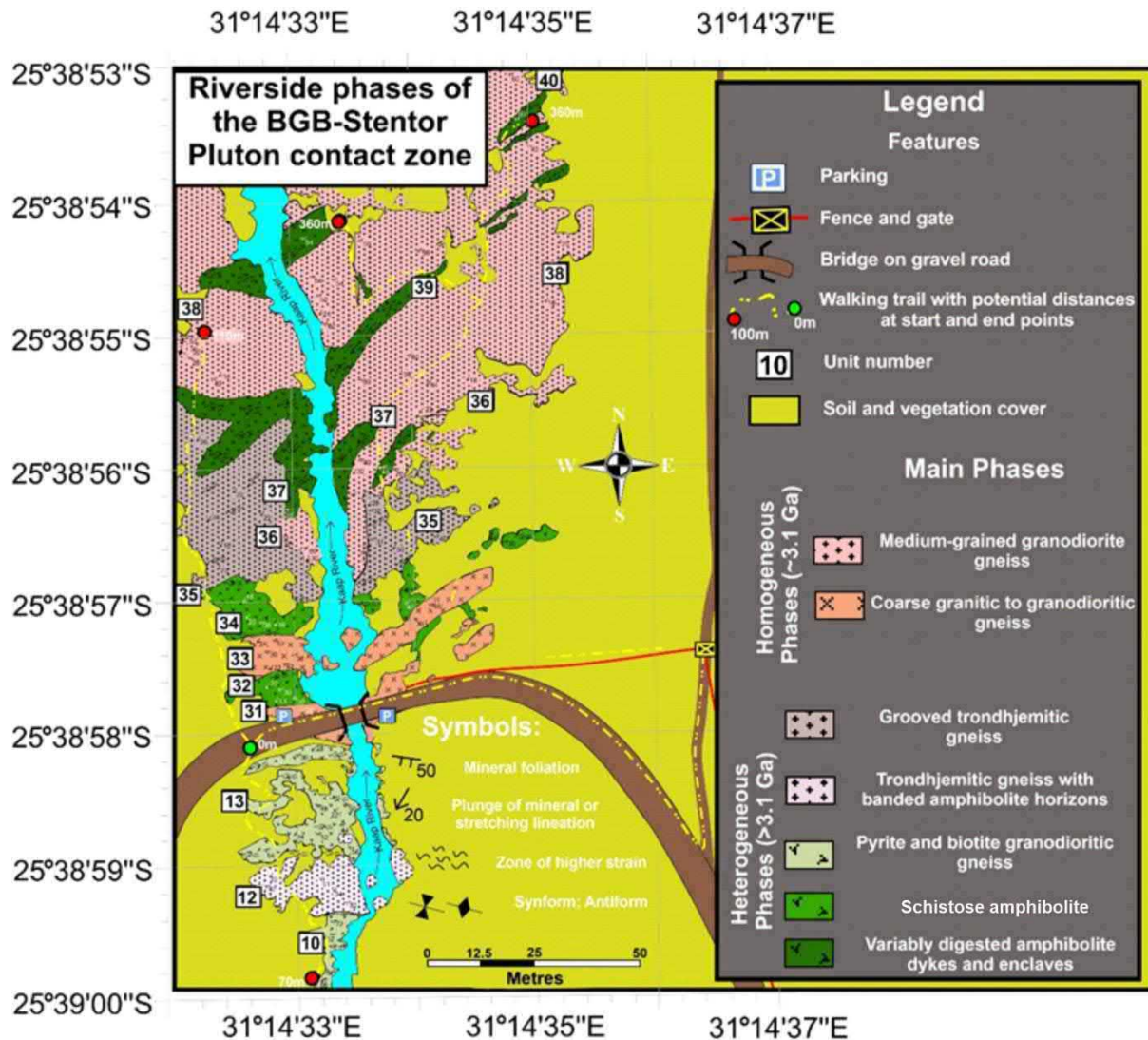


Figure 7.6 Phase map of the southern margin of the Stentor Pluton at the Honeybird Siding locality c. 6 km west southwest of Low's Creek. Numbered units are individual south-dipping blocks bound by south-dipping faults. Ages sourced from Schoene *et al.* (2008).

7.4.2 Petrographic analysis of mafic phases

Southernmost outcrops are comprised of sub-hedral to anhedral chlorite aggregates partially resorbed into granoblastic albite and minor quartz. Associated with chlorite margins are millimetre-scale anhedral magnetite. Internal sectors of some chlorite crystals are occupied by pyrite aggregates. Notably, pyrite from a fracture-dense pyrite-biotite gneiss (10) is of a pock-mark texture (Fig. 7.7 B) whereas that from less fractured pyrite-biotite gneiss (13) has smooth surfaces. Texturally, the pyrite aggregates consist of sub- to anhedral pyrite grains growing from margins and cores of advancedly altered hexagonal minerals. In some cases, the pyrite growth front is truncated by anhedral albite (Fig. 7.7 C), suggesting post-pyrite shearing in the

pyrite-biotite gneiss. Further north are sheared schistose amphibolites (32 and 34) with only magnetite as an opaque phase associated with hornblende. Northwards of these sheets are variably digested amphibolites that occur as mafic bands of a folded banded gneiss (38F) or as deci- to metre-scale gneiss enclaves. These amphibolites are comprised of hornblende aggregates and altered clinopyroxene resorbed into a matrix of zoisite, feldspar and quartz (Fig.7.7 F).

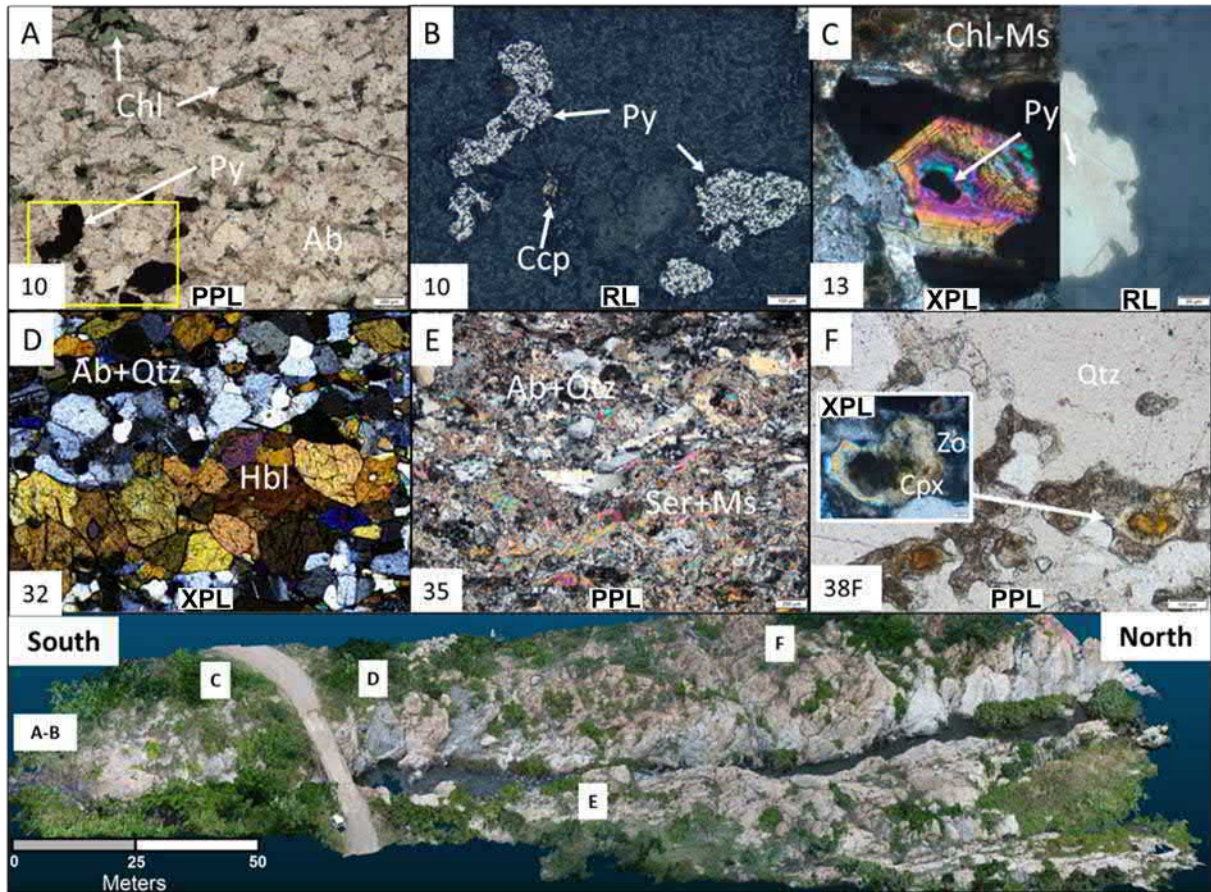


Figure 7.7 Characteristic lithologies of the Honeybird Siding contact zone; numbers at bottom left corners are suffixes of sample names with an NP- prefix. PPL, XPL and RL labels are in photomicrographs produced using plane-polarised, cross-polarised or reflected light, respectively. (A-C): Pyrite aggregates associated with chlorite partially resorbed into granoblastic albite and minor quartz. Note pyrite replacement of a hexagonal alteration zone truncated by albite in (C). (D): Basal sections of hornblende aggregates that form lineations observed in outcrops. (E): Quartz ribbons aligned with muscovite in sericite matrix form quartzofeldspathic lineations observed in outcrops. (F): Altered clinopyroxene in digested amphibolite is partially resorbed into a quartz-albite-zoisite groundmass. Sampled blocks marked on image captured from a 3-D model created by Thomas Jones.

7.4.3 Elemental maps of pyrite-biotite gneiss (unit 10)

Elemental maps of the southernmost pyrite-bearing chlorite-muscovite schist (Fig. 7.8) show homogeneous distribution of S corresponding to pyrite crystals. Though Ca also occupies the

whole map, discrete clusters Ca-bearing minerals like calcite coincide with chlorite-rich zones, while fine, uniformly distributed Ca occupies albite-oligoclase zones. In addition to chlorite-forming elements (Fe, Mg, Al), chlorite lenses coincide with zones rich in Ti-bearing minerals such as rutile, while albite-oligoclase zones only bear higher concentrations of albite-oligoclase-forming elements.

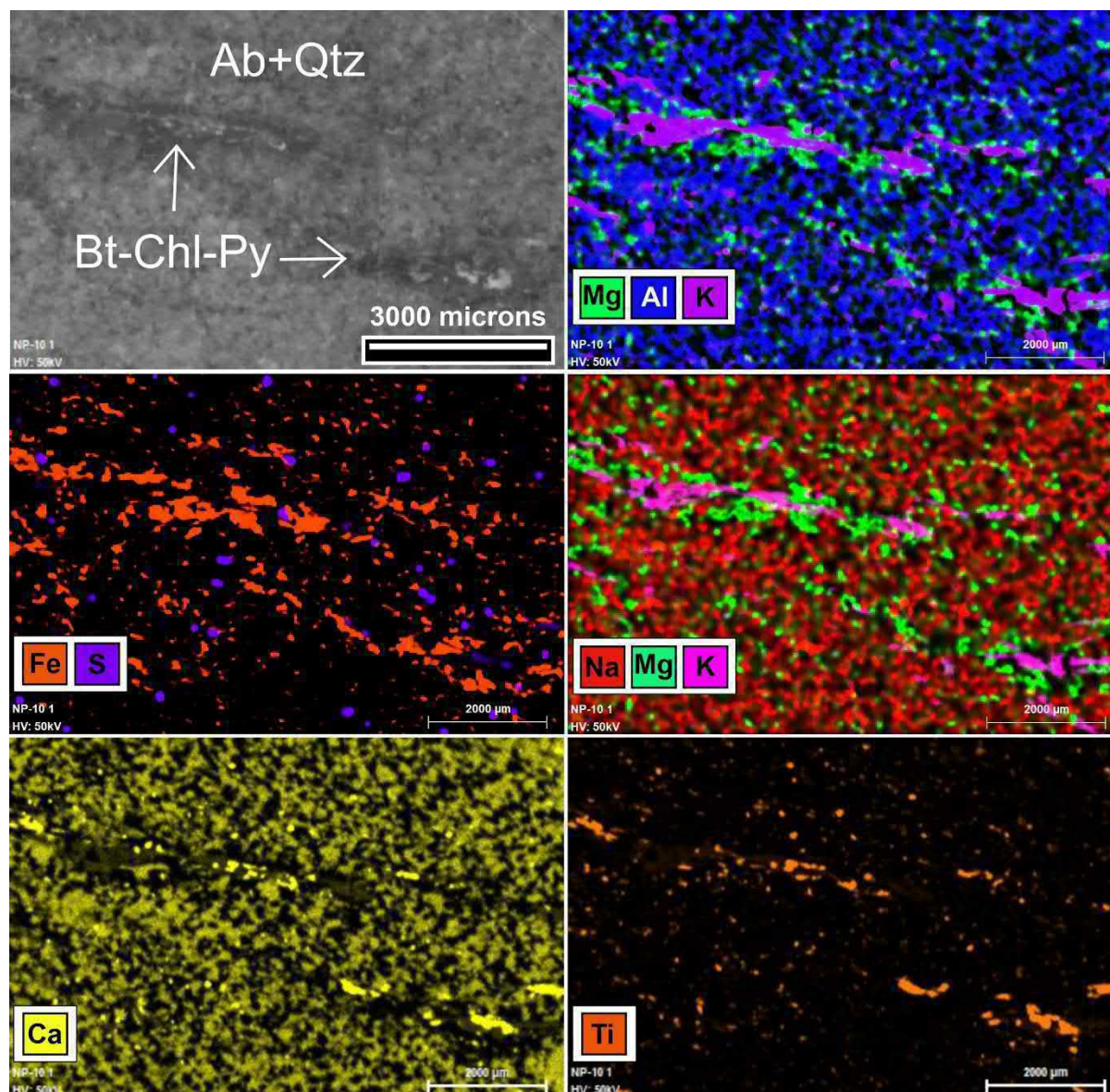


Figure 7.8 Elemental maps of sample NP-10. The maps show an association of chlorite lenses with coarse clusters of fine Ca- and Ti-bearing minerals in addition to chlorite-forming

7.4.4 Phase maps of pyrite-biotite gneiss (unit 13)

A phase map of the pyrite-bearing chlorite-muscovite schist closest to the bridge shows pyrite growth from a highly altered phase which is the only Ca-rich zone in the analysed part of sample (Fig. 7.9). The altered phase lacks sulphur and shows a very faint signal for Al and Fe (dark and medium grey respectively). As in the other pyrite-bearing pyrite-biotite gneiss, a faint signal for Ca occurs even in the albite-rich zones, indicating an albite composition towards oligoclase. Also similar in both chlorite schists is the higher concentration of Fe, Mg, Ti, Ca, Al and K in the chlorite. This is consistent with the observed chloritised biotite and associated Ca-rich altered phase.

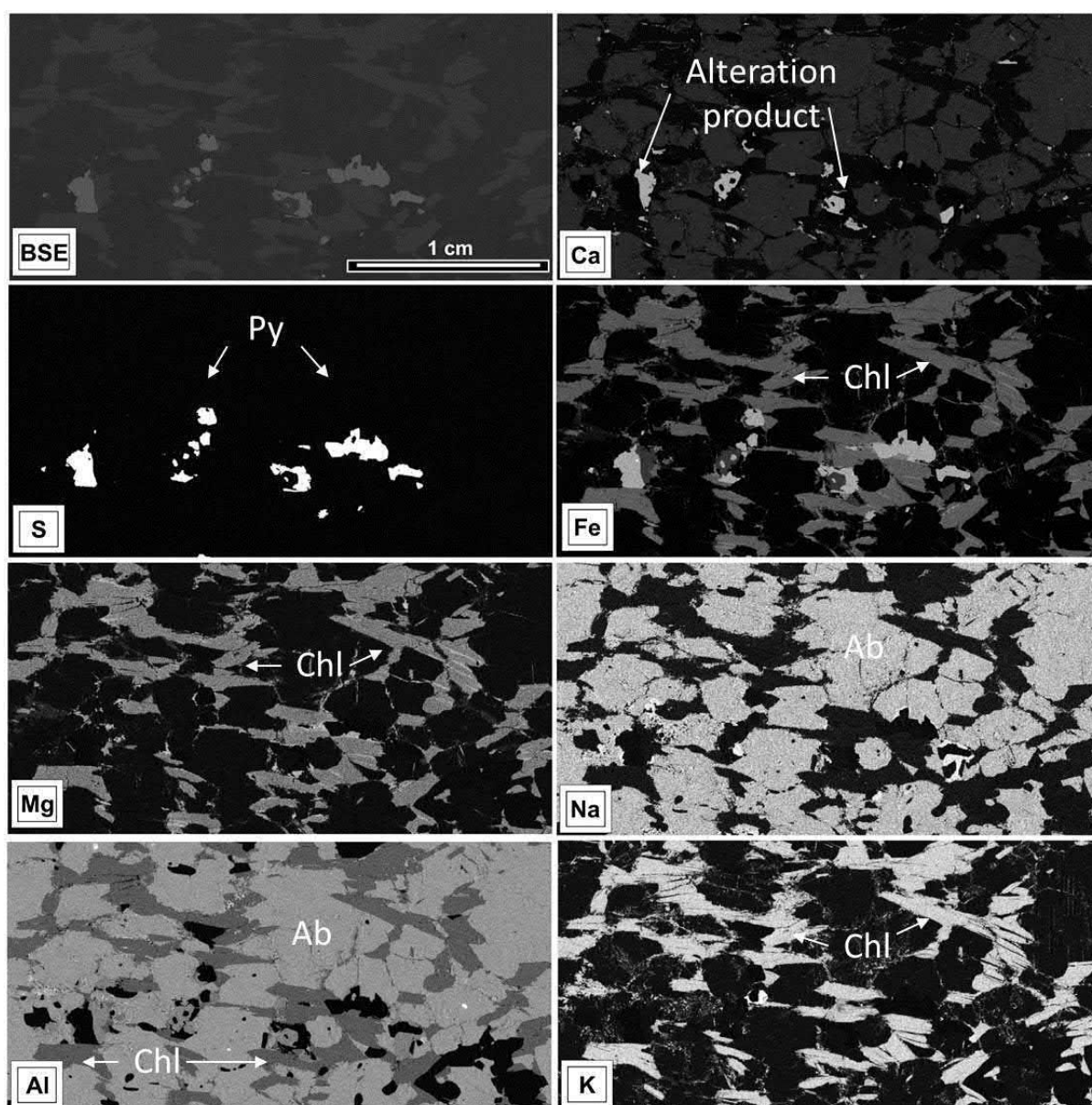


Figure 7.9 Elemental maps of Sample NP-13. BSE map shows albite crystals (Ab) as dark grey, chlorite (Chl) as faint grey and pyrite (Py) as medium grey. The single element maps show that the advancedly altered mineral replaced by pyrite occupies zones richer in Ca and lesser Fe and Al.

7.5 Discussion

Deformation and retrograde metamorphism of amphibolites and biotite schists produced a greenschist assemblage of albite + chlorite + muscovite $\pm$ quartz. Later sericitisation and saussuritisation of feldspars formed micas and zoisite.

The anhedral shape of pyrite aggregates and crystals appears to be a result of pseudomorphic growth after a sheared, rarely-preserved hexagonal phase. However, the occurrence of pockmarks on pyrite from the fracture-dense unit (10) and smooth-textured pyrite from a less fractured unit (13) suggest deformation, dissolution to be the cause of the former texture.

In a study focussed on most of the southern margin of the Stentor Pluton, Dziggel *et al.* (2006) report andradite and pyrope as the main garnets found in metabasic wallrocks of the pluton. However, partially preserved minerals in the limited number of samples of mafic phases of the present study preserve hornblende + plagioclase + clinopyroxene $\pm$ quartz as (remaining) components of a peak metamorphic assemblage.

Although Schoene *et al.* (2008) found evidence for dextral transtension in sporadic three-dimensional exposures, sporadic outcrops observed during the present study present shear sense indicators of a combined normal and sinistral component. Similar observations of sinistral transtension were reported by Dziggel *et al.* (2008). This may indicate that the dextral Revolver Shear Zone and the sinistral Honeybird Shear Zone c.1 km south thereof constitute a complex southern contact zone of the Stentor Pluton that was subjected to localised controls on strain orientation (Schoene *et al.*, 2008).

7.6 Conclusion

The structural fabrics of the Honeybird Siding contact zone records extensional ductile shearing contemporaneous with the emplacement of the Stentor Pluton. Amphibolite and biotite schists invaded by quartzofeldspathic sills and dykes occur as chloritised mafic schists and (para-)gneisses, depending on the abundance of quartzofeldspathic components that invaded them. The growth of pyrite from an altered phase predated at least one deformation event that resulted in shearing and partial dissolution of secondary pyrite observed in the mafic sheets studied.

Chapter 8: Magnesite Veins

Parking: (25°33'23.53"S; 31°19'32.52"); Representative outcrop: (25°33'23.47"S; 31°19'33.25"E)

8.1 Site description

Site 8 occurs near Kaapmuiden, c. 2 km south of where the R38 intersects the N4 national toll route that leads to various tourist attractions (see introduction). The site presents roadcut exposures of magnesite veins at a locality known among locals as Pettigrew's Neck/Pettigrew Nek (Fig. 8.1).

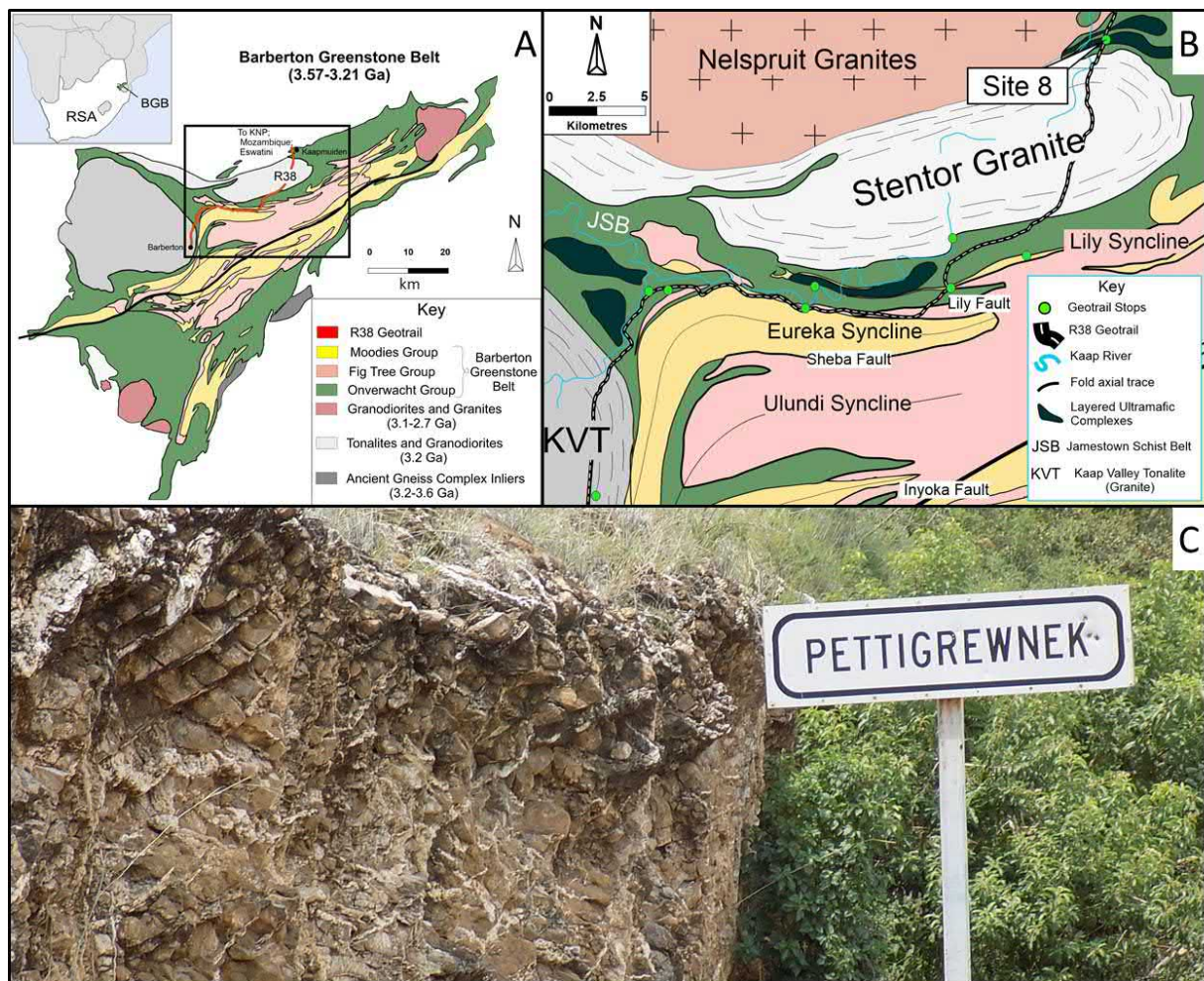


Figure 8.1 Main features of Site 8. (A): Simplified geological map of the R38 Geotrail in the Barberton Greenstone Belt. (B): A detailed view of the black rectangle in (A) shows the location of Site 8 in a layered ultramafic complex situated at the northeastern margin of the Stentor Pluton. (C): View of site 8 from northeast of the roadcut surface. Maps modified after Visser *et al.* (1956) and Anhaeusser *et al.* (1981).

8.2 Aspects of magnesite and Site 8 that may interest laypersons

The name Kaapmuiden, which translates to Cape (River) mouth in Dutch, bears reference to the confluence of the Kaap River and the Crocodile River (Fig. 8.2) just northwest of the town (Raper, 1987). Two other important junctions occur near Kaapmuiden, they are the intersections of the road and railway from Barberton, with a major road and railway that links Pretoria to Mozambique. The transport systems were developed during the Barberton goldrush in the 1880s to connect the growing town of Barberton with prominent transport routes (See appended guidebook). Presently, the stretch of road between Kaapmuiden and Mozambique has high tourist traffic due to attractions such as the Kruger National Park, a variety of cultural sites as well as being (part of) a direct, well-maintained route to Mozambique and Eswatini.

Across the road from site 8 is an old quarry where 13 063 tonnes of magnesite were extracted from 1909 to 1928 by the Budds Magnesite Mine (Groeneveld, 1975). After that, only intermittent attempts at mining produced a combined mass of less than 200 tonnes between 1945 and 1955 (Groeneveld, 1975).

At site 8, visitors will see different features of magnesite veins that give clues about how the mineral formed. Also important is the heterogenous distribution of strain and metasomatism that can be appreciated from the outcrop by walking along the road. This will introduce visitors to concepts of strain partitioning within a single lithology. By learning how Site 8 magnesite formed, visitors will gain insight into one of the reactions that transform gaseous CO₂ into a component of the lithosphere.

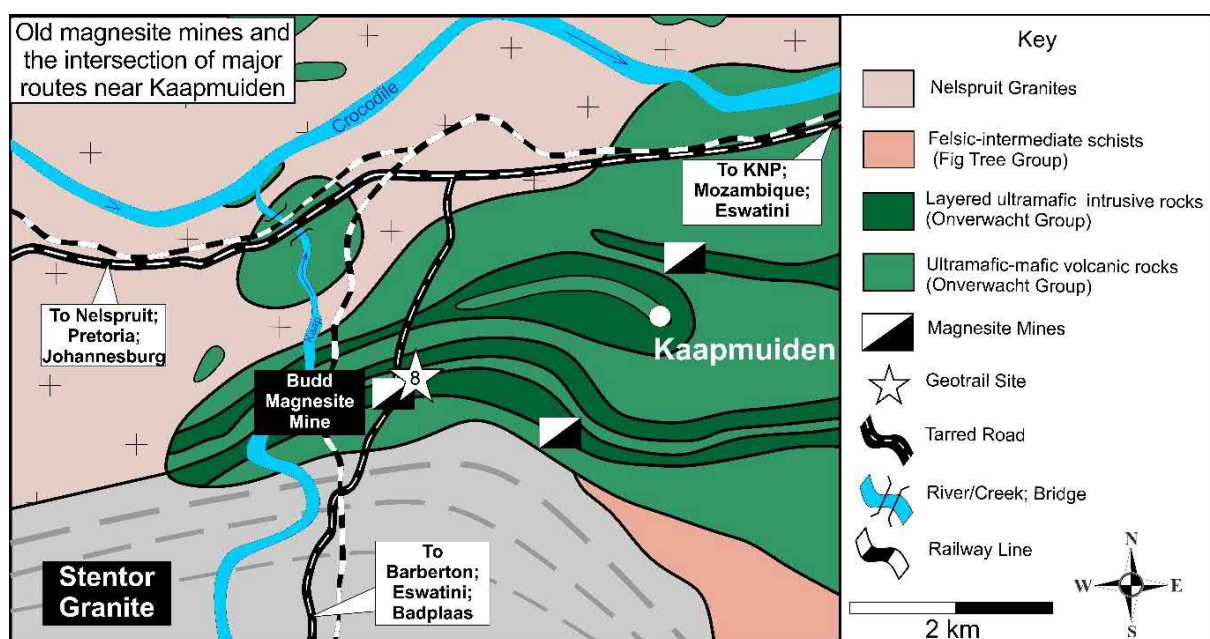


Figure 8.2 Simplified map shows the intersection of rivers, roads and railway lines northeast of Kaapmuiden.

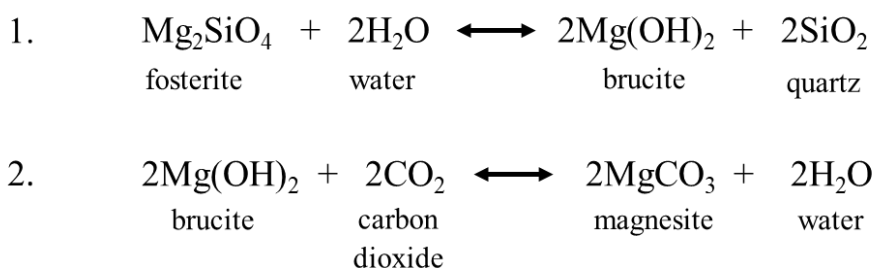
8.3 Geology

8.3.1 Background

Magnesite is a magnesium carbonate (MgCO_3) of a hardness range between 3.5-5 (Siegal, 1967). Uses of magnesite include the manufacturing of refractory bricks and in cement, antacid, and fertilizer products. Once calcined to periclase (MgO), magnesite becomes a useful source of magnesium metal used to make metallic alloys for products such as vehicle parts (Van Zyl *et al.*, 1942; Evans, 1993).

Magnesite occurs in various geological settings and originates from hydrothermal and sedimentary processes (Pohl, 1990; Abu-Jaber and Kimberley, 1992). Typically, economic magnesite deposits are shallow-marine or lacustrine stratabound lenses of a sugary or coarse texture (Veitsch Type) or cryptocrystalline veins in altered ultramafic lithologies (Kraubath Type) (Pohl, 1990). In South Africa, hydrothermal deposits associated with Archean layered ultramafic complexes are a significant natural source of magnesite (Toulkeridis *et al.*, 2010).

Formation processes of vein magnesite have been the focus of research since the beginning of the last century. Dominant genetic models include hypogene and supergene processes reviewed in Pohl (1990) and Abu-Jaber and Kimberley (1992). Generally, previous studies show that magnesite deposits occur within the uppermost 300 metres of the Earth's crust (Pohl, 1990; Abu-Jaber and Kimberley, 1992; Zedef *et al.*, 2000 and references therein). Though the nature of the carbonate-rich fluid has been a topic of contention, the source of magnesium ions is widely considered to be from altered ultramafic rocks (Pohl, 1990; Abu-Jaber and Kimberley, 1992). This has been confirmed by various experiments where serpentinites or forsteritic dunites (e.g., reactions 1 and 2 below) were used in the synthesis of magnesite during CO_2 sequestration experiments (Saldi *et al.*, 2009; Montes-Hernandez *et al.*, 2012; Moore *et al.*, 2014).



Economic magnesite deposits of the Barberton region are in the northeastern margin of the Barberton Greenstone Belt (Van Zyl *et al.*, 1942; Groeneveld, 1975; Viljoen and Viljoen, 1970;

Anhaeusser, 2006; Toulkeridis *et al.*, 2010). The biggest of these is situated in the Magnesite-Canal layered ultramafic complex c. 10 km east southeast of Kaapmuiden. This deposit is the only one in the region to have a production history that extends from the beginning of the 20th century to the present day (Groeneveld, 1975).

The Budd magnesite of Site 8 occurs within a layered ultramafic complex comprised of differentiated sequences (Viljoen and Viljoen, 1970). From base to top, these comprise a peridotitic chill margin, dunite, orthopyroxenite, websterite and anorthositic gabbro/leucogabbro (Viljoen and Viljoen, 1970). Invariably, magnesite occurrences occur in the dunite layers of the differentiated sequences of the region within shallow ground (<300 m) (Groeneveld, 1975). The shallow occurrence led previous authors to propose meteoric water as a critical factor in the development of magnesite veins in mafic lithologies (Groeneveld, 1975).

Host rocks to the layered sequences include metabasalts of a tholeiitic or komatiitic nature (Viljoen and Viljoen, 1970). These, according to Viljoen and Viljoen (1970), are interlayered with olivine-bearing pyroxenite cumulates or amphibolite-grade banded iron formations. The Ship Hill layered intrusive complex c. 1 km north of Site 8 reveals the effects of general north-northwest shortening experienced by the area in which Site 8 occurs (Viljoen and Viljoen, 1970) (Fig 8.3). The same deformation is preserved in the Koedoe layered sequence c. 7 km northeast of the Ship Hill layered sequence. In both sequences, lithologies are folded into synclines with steep fold axes that plunge westwards, parallel to strike-slip faults that truncate southern limbs (Viljoen and Viljoen, 1970). The Ship Hill Syncline is truncated by the steep Ship Hill Fault (Fig. 8.3). Though no synclinal structure was reported for the Budd layered sequence, the unit is bound to the south by the Stentor Fault which is understood to be partly responsible for the slight bending of layers (Viljoen and Viljoen, 1970).

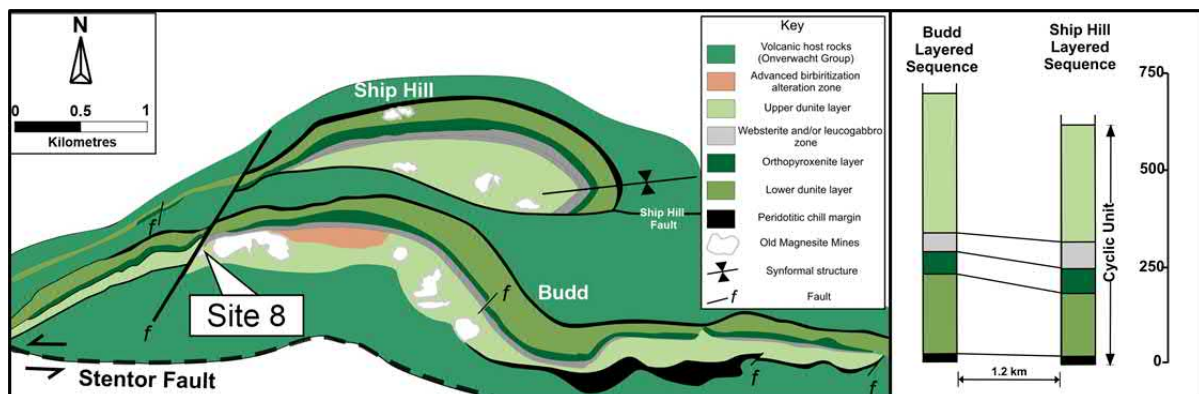
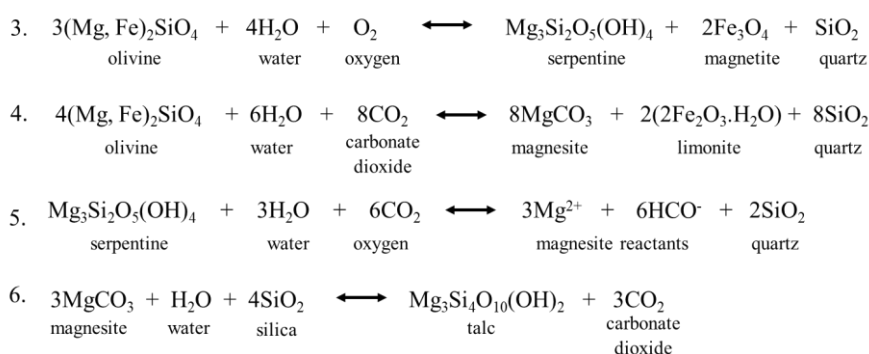


Figure 8.3. Old magnesite mines in the Budd and Ship Hill layered ultramafic complexes. Map and stratigraphic units modified after Viljoen and Viljoen (1970).

Two main vein types of magnesite are recorded in the Barberton Greenstone Belt: (i) white, amorphous magnesite and (ii) magnesite of a nodular texture (Van Zyl *et al.*, 1942).

The white amorphous magnesite is more common (Van Zyl *et al.*, 1942 and Groeneveld, 1975). It occurs as sub-horizontal, quartz-poor veins and sub-vertical veins with a higher quartz content. These variants are sometimes referred to as “zebra ore” and “mompapa” veins, respectively (Van Zyl *et al.*, 1942; Groeneveld, 1975).

Hall (1918) proposed a two-step reaction whereby the serpentinisation of olivine precedes the carbonation of serpentine by meteoric water with dissolved CO₂ (Reaction 3). An additional mechanism proposed by Van Zyl *et al.* (1942) postulated that excess CO₂ in an aqueous solution, also bearing dissolved O₂, would facilitate direct carbonation of magnesium ions liberated from the hydrolysis of forsterite (Reaction 4). The latter mechanism would produce magnesite and limonite or magnetite depending on redox conditions of the system (Taylor, 1974). Notably, most BGB magnesite deposits occur in serpentinised dunites thought to have been altered according to reaction 5 below (Van Zyl *et al.*, 1942). According to Groeneveld (1975), the resulting magnesite may form talc through silicification (reaction 6). Interestingly, a study of magnesite deposits of the northern Barberton Greenstone Belt (BGB) reported talc mineralisation beneath a horizontally continuous, termination zone of magnesite mineralisation at c. 15 m depth, suggesting the overall shallow nature of some BGB magnesite deposits (Van Zyl *et al.*, 1942). However, the possible origin of the above-mentioned underlying talc through the silicification of magnesite may be worth investigating. This would be to assess the apparent confinement of magnesite formation to shallow land horizons as a lack of deep quarries or borehole data prevented conclusive statements from previous workers (Van Zyl *et al.*, 1942; Groeneveld, 1975).



Plausible sources of carbonate ions required for the widespread carbonation of ultramafic rocks in the BGB were listed in Anhaeusser (1969). These include ultramafic wallrocks through

decarbonation or dissolution of carbonate amygdales and calcareous oozes which may be associated with pillow lava. In addition to these, a general review of magnesite deposits by Zedef *et al.* (2000) lists heat-induced decarbonation of limestone; decomposition of organic material; exhalative volcanic activity; ascending CO₂-rich fluids from a deeper source; or a combination of the listed sources. Additionally, Beard and Hopkinson (2000) as well as Keleman and Matter (2008) propose the oxidation of methane-bearing fluids by mixing with oxygenated water to form carbonate ions.

In the case of deposits in the Budd Layered Ultramafic Complex, magnesite is reported to have formed after brucite (Toulkeridis *et al.*, 2010) through the carbonation of serpentinised dunitic layers in flat topographic layers where pervasive percolation by meteoric water could take place (Viljoen and Viljoen, 1969; Anhaeusser, 1976). The localised occurrence of magnesite in shallow-level, magnesium-rich dunite layers is common among similar layered ultramafic complexes of the BGB (Van Zyl *et al.*, 1942). However, evidence that the transportation of magnesite ionic components from source regions to brittle structures of various lithologies was also noted (Van Zyl *et al.*, 1942).

Whether magnesite formed in situ or was transported and emplaced in current structures remained unclear to some previous workers (Van Zyl *et al.*, 1942). However, the occurrence of magnesite veins in lithologies that are not compositionally favourable for substantial magnesite formation (hornblende-chlorite schist, dolerite dyke, quartz vein) suggested to Van Zyl *et al.* (1942) the transportation of magnesite-forming constituents into fractures.

Cryptocrystalline magnesite veins are typically purer than crystalline or nodular sedimentary equivalents (Pohl, 1990; Abu-Jaber and Kimberley, 1992). Bulk chemical analysis of six cryptocrystalline magnesite from the Budd deposit reflects a predominance of magnesium (46 wt. %) and volatiles (51 wt. %) with calcium (< 0.36 wt. %) and silica (< 0.03 wt. %) as the main contaminants (Toulkeridis *et al.*, 2010). A Pb-Pb age of 3043 ± 49 Ma was determined from bulk chemical analysis data of the cryptocrystalline veins of the Budd magnesite deposit. This age, according to Toulkeridis *et al.* (2010), suggested magnesite formation occurred shortly after uplift of the Budd Layered Ultramafic Complex during the intrusion of the c.3109 Ma Nelspruit Batholith north and east of the deposit. Further support for this came from Ce, $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ geochemical data interpreted to indicate precipitation of magnesite in an oxidising epithermal system similar to that proposed by Pohl (1990) for the Kraubath Type deposit of Archean vein magnesite (Toulkeridis *et al.*, 2010).

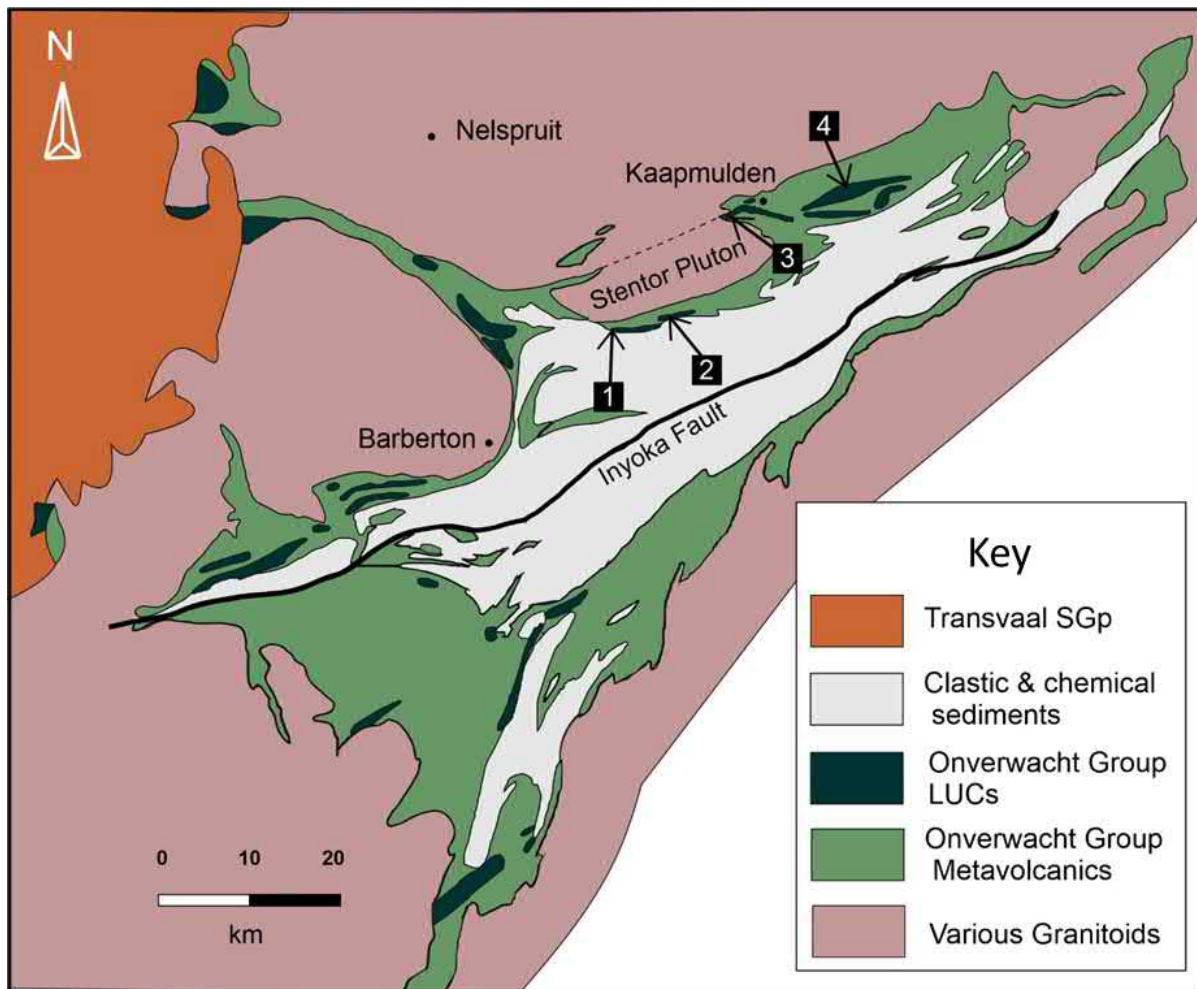


Figure 8.4 Layered ultramafic complexes (LUC) of the Barberton Greenstone Belt. Numbered are magnesite deposits described in the present study: 1. Bon Accord LUC, 2. Sugden LUC, 3. Budd LUC, 4. Magnesite-Canal LUC. These complexes host the Bon Accord and Sugden magnesite deposits in the southwest, and the Budd and Strathmore deposits in the northeast. Map modified after Anhaeusser *et al.* (1981) and Anhaeusser (2006). SGp: Supergroup.

The growth of the steel industry in South Africa resulted in an increased demand for refractory magnesite blocks used in the production of steel as refractory blocks for furnaces (Van Zyl *et al.*, 1942). Mining and beneficiation of BGB magnesite is sub-economic due to its low iron content (a sintering agent) and the consequent higher processing temperatures to form refractory magnesite blocks (Van Zyl *et al.*, 1942). However, increased steel production led to a rise in magnesite mining of the northern BGB occurrences during the early 20th century. Without high demand, the sub-economic mining and beneficiation processes of most vein magnesite occurrences of the BGB deposits (Van Zyl *et al.*, 1942) remains unfavourable in the present day.

8.3.2 Present study

8.3.2.1 Research Question

Descriptions of various magnesite deposits of the world and their associated host rocks are abundant in the literature (Abu-Jaber and Kimberely, 1992 and referenced therein). Similarly, a number of publications have reported the carbonatisation of ultramafic lithologies of the Barberton Greenstone Belt (Van Zyl *et al.*, 1942; Groeneveld, 1975; Toulkeridis *et al.*, 2010). Previous workers have addressed the composition of magnesite, mode of occurrence, sources of components that form magnesite, and the age of precipitation. However, there is almost no documentation of key vein-type magnesite structures such as the sinusoidal habit observed as a pervasive fabric of veins at site 8. Whether the sinusoidal habit of the veins is common among regional deposits, and whether it relates to deformation or elemental distribution within the vein has been one of the focusses of this study.

8.3.2.2 Aim

To describe selected physical and chemical characteristics of magnesite veins in the roadside exposure of the Budd magnesite deposit for the purpose of identifying primary and secondary characteristics.

8.3.2.3 Objectives

To determine if veins of different orientation have a sinusoidal nature. To compare the nature of magnesite veins of the Budd deposit at Site 8 with magnesite veins of three other localities within a similar geological context. To produce elemental maps of magnesite veins to chart the distribution of selected elements to determine if a paragenetic sequence can be observed from crystal zonation.

8.3.2.4 Methods

General features of magnesite veins such as colour, habit, distribution, and hardness were recorded from the Budd magnesite deposit at Site 8. These features were compared with those of magnesite deposits in the Sugden, Strathmore and Bon Accord localities. All selected sites occur within the northern margin of the Barberton Greenstone Belt and have a similar host rock and mode of occurrence. Orientation measurements of veins were recorded from veins exposed in three dimensions. All samples were of the sub-horizontal group of magnesite veins as these are purer (Van Zyl *et al.*, 1942; Groeneveld, 1975). Four samples of magnesite within host rocks were obtained from Site 8 (2 samples) and the Strathmore magnesite deposit (2 samples)

to supplement existing descriptions of vein and host rock relationships. Material from the Sugden locality was too friable to yield cohesive samples of the host rock and associated veins. Thin sections from the Bon Accord magnesite locality were not produced due to the discovery of the site at the end of the present study. Five representative samples of magnesite were selected for elemental mapping of samples with characteristic features. These include a zoned magnesite micro-vein of low sinuosity (NP-MJ2) and a highly sinuous, thicker vein (NP-MJ3). A representative map of a sample from the Sugden locality (NP-MJS) was produced to compare it with those from the Budd deposit as both localities occur within 1 km of the margins of the Stentor Pluton. A representative sample from the Strathmore Magnesite Mine (CH-S19) which lies more than 1 km from the Stentor Pluton was mapped. This was to compare elemental distribution of key elements in magnesite of the study area with a representative sample from an economic deposit of the region. A phase map was obtained for sample NP-MJ2 to identify the zoned crystals. In the absence of a thin section of Sugden magnesite (NP-MJS), Raman spectra were generated to identify components within the core of a representative magnesite vein from the opposite margin of the Stentor Pluton to Site 8.

8.4 Results

8.4.1 Field Observations

Site 8 magnesite veins occur in highly fractured dunite layers within the sub-vertical Budd Layered Ultramafic Complex. The roadside exposure presents fracture-bound dunite blocks that have been variably serpentinised and birbitised to iron hydroxides. Host rock alteration is advanced in zones where the fracture density is higher and the protolith is unrecognisable. Predominant fractures are sub-horizontal. These are cut by diagonal and sub-vertical veins that typically occur as conjugate pairs. Some veins bifurcate and taper out while others form continuous layers of equal thickness which may step up or down. These observations were found to be consistent with the mode of occurrence of other magnesite localities at the northeastern edge of the BGB.

Apart from those that appear to be recrystallised into lumpy aggregates, magnesite veins observed are of a cryptocrystalline texture. They range in width from mm to dm scale and are white in colour. Vein margins are encrusted with quartz and thin strips of chloritised dunite or unrecognisable wall-rock fragments that have been highly altered to iron hydroxides, as is the case at the Sugden locality.

Composite magnesite veins are a common occurrence. They may be composed of a series of short, on-lapping, tapering layers (tension gashes) or of tabular, continuous layers separated by voids. In both cases, thin remnants of serpentinised, chloritised and/or birbiritised dunite occur between individual magnesite veins. The thickest veins observed (few tens of centimetres) were all composite veins but not all were separated by vestiges of host rock material. At the Bon Accord locality, some tabular, parallel veins of equal thickness were separated by a continuous, layer-parallel void (Fig. 8.5 E). Asymmetric folding on those veins suggests a top-to-the-southeast displacement caused by northwest-southeast extension. Within the same Bon Accord locality, as well as at the Budd deposit, compressional deformation of a top-to-the-southeast sense can be seen in tabular, thicker veins (Fig. 8.5, A and D respectively). No consistent link between vein orientation and thickness was observed at all four localities.

At outcrop scale, it was apparent that magnesite veins form along fractures within serpentinised, chloritised and sometimes birbiritised dunite. Additionally, magnesite of the Bon Accord locality was either found to occupy fractures in serpentinised dunite which was chloritised at vein margins, or in fractures through existing chrysotile asbestos veins. Not all chrysotile asbestos veins in the carbonated dunite of the Bon Accord locality were directly associated with magnesite.

A tightly sinusoidal habit can be seen in all vein orientations except in the thicker (5-10 cm), tabular variants that are often composites of cryptocrystalline or nodular veins. Tight folding of veins did not appear to be confined to any horizon of the exposures studied.

Thinner magnesite veins (0.5-1 cm) are generally more weakly sinusoidal than thicker equivalents (2 cm-5 cm). The thicker and tightly folded veins occur within dunite blocks of no apparent ductile strain close to vein margins (Fig. 8.5 A). Some of the fractures fully-filled with magnesite are tabular and continuous, while others have tapered ends and are reminiscent of tension gashes (Fig. 8.5 A). The vergence of microfolds is consistent in magnesite layers that fully occupy fractures. This is not the case for magnesite layers that partly fill continuous, tabular fractures in the host rock. The latter magnesite layers are tightly folded and have different directions of fold vergence within a few centimetres of the same vein (Fig. 8.5, A). They are separated from the host rock by voids and vary in thickness along the vein.

A three-dimensional view of cryptocrystalline magnesite veins (Fig. 8.6 C, D) shows that the microfolds are not continuous in the Y-direction of a finite strain ellipse applicable to any of the folds. Instead, the real fold fabric indicates mostly layer-parallel shortening through the

development of microfolds of varying orientations within a single magnesite layer. Where the magnesite layer margins are free of host-rock material, syneresis cracks may be observed (Fig. 8.6 C).

Open folds observed in thick, tabular veins reported above are invariably sub-parallel or diagonal to vein lengths (Fig. 8.5, A, E, F). In contrast, tight folds are generally oriented normal to lengths of bounding fractures whereas the asymmetric folds vary in vergence. Therefore, it appears that magnesite veins of the study area fall into two main groups. Group 1 veins are characterised by tabular planar veins that may occur as single or composite veins. Group 2 veins are characterised by tightly sinusoidal veins of a consistent or inconsistent fold vergence. These may fully or partially fill open fractures. At least one vein margin of Group 1 veins lies in direct, consistent contact with the wall-rock. These folded tabular veins present open, asymmetric folds of a consistent vergence along vein lengths.

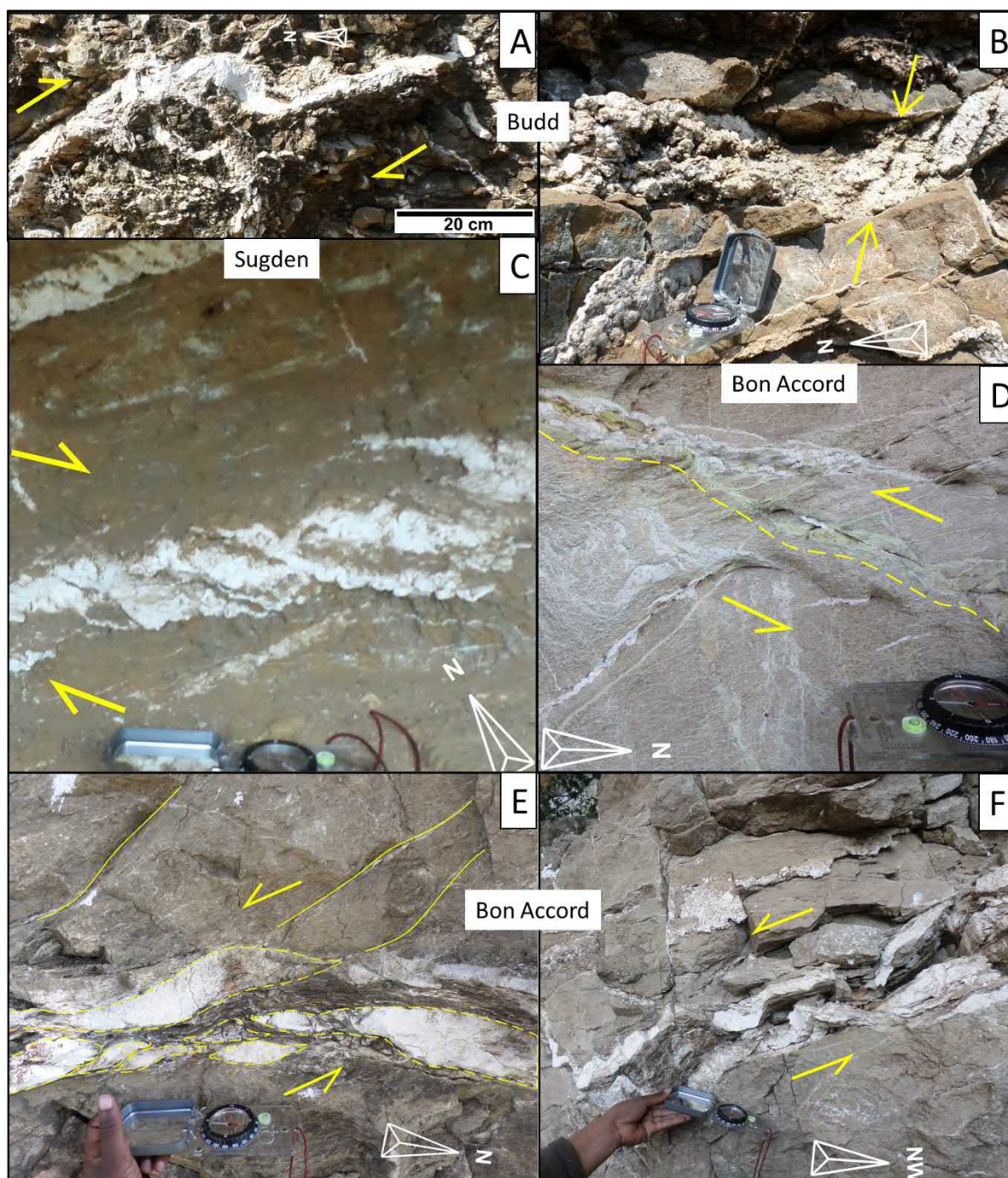


Figure. 8.5 Deformation features associated with magnesite from the northern margin of the Barberton Greenstone Belt. (A): Apparent rotation of a vein network in a top-to-the southeast shear sense (B): Northeast-dipping series of tension gashes filled with sparry magnesite. (C): 3-D exposure of southeast-dipping tension gash planes preserve microfolds oriented normal to vein lengths. (D) Fracture along an asbestos vein provided a site for magnesite growth. (E): Southeast-dipping planes of curved, tapering magnesite veins are separated by films of chloritized dunite. (F) Folded, parallel veins bound by films of host rock are separated by vein-parallel voids. Note vestiges of wall-rock on the bottom margin of the vein above the void.

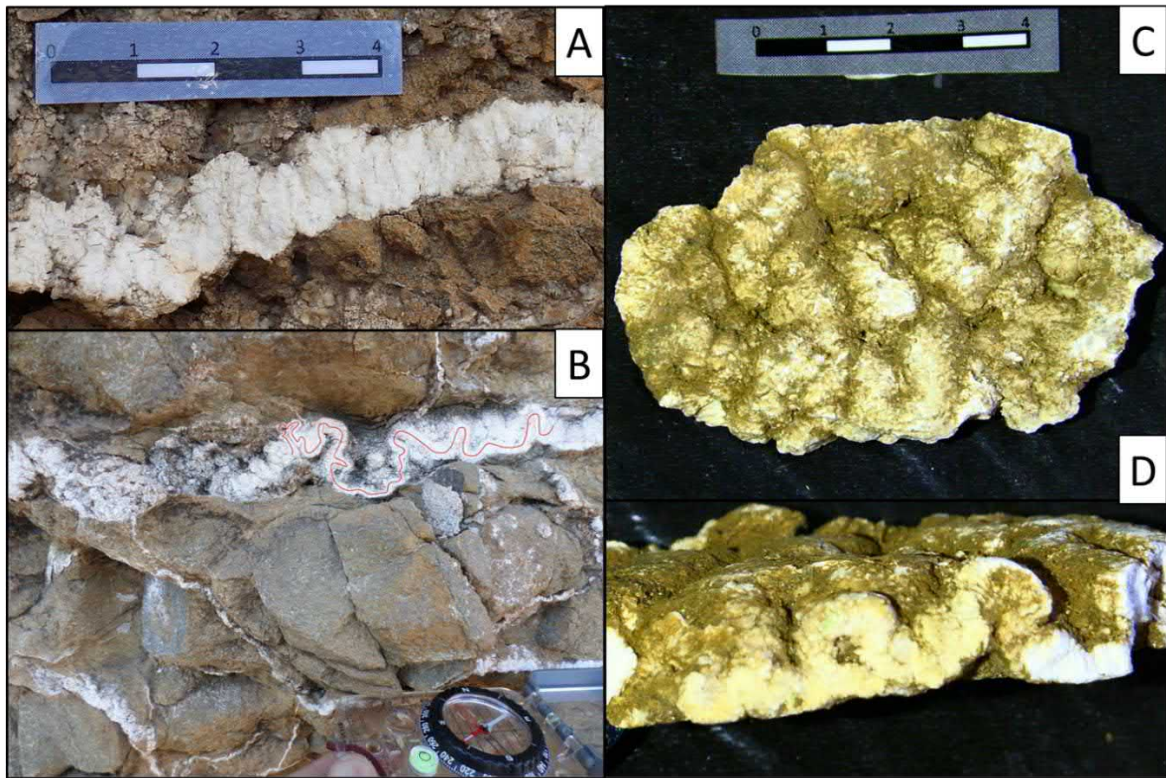


Figure. 8.6 Magnesite veins at site 8 (Budd deposit) occur as white layers shortened into microscale folds. (A): Some magnesite layers are folded tightly into microfolds of a generally consistent trend. (B): Other magnesite layers occur as open folds that verge in different directions. (C-D): Plan and cross-sectional view of a folded magnesite layer showing surfaces of multidirectional, non-cylindrical folds. Note the syneresis (dehydration) cracks on the folded surface of the magnesite layer.

8.4.2 Petrography

Thin sections of magnesite veins within their host rocks show that veins are either amorphous or syntaxial with symmetrical margins of fibrous crystals extending into a core of coarse, interlocking crystals (Fig. 8.7, E insert). The veins occur within massive, serpentinitised dunitic cumulates. In a highly altered, vein-bearing sample from the Budd locality (NP-MJ2), the serpentinitised dunitic cumulates were further altered to chlorite and interstitial spaces were steatitised to talc. In a better-preserved sample from the same locality (NP-MJ1), cumulus olivine grains are preserved together with minor orthopyroxene and rare clinopyroxene. These occur in a groundmass of serpentine intergrown with magnesite and tremolite replaced by talc. In Strathmore samples, magnesite veins occurred as single and composite veins associated with vein-parallel magnetite clusters. The term composite is used here to describe networks of parallel veins and not the presence of different minerals grown syntaxially or antitaxially in the same vein-as defined by Durney and Ramsay (1973). The composite Strathmore veins were observed in a serpentinitised dunite (NP-301). In a websteritic sample, magnesite occurred as a replacement of chrysotile asbestos within a band of prehnite and serpentine (NP-306).

A magnesite vein from the Budd locality showed that some vein cores are occupied by zoned, monoclinic crystals that lie in alternating orientations (Fig. 8.7, B, C). The low sinuosity of the vein and the recrystallisation of vein margins prevented an assessment of any recurring coincidence of fold hinges with a particular orientation of zoned monoclinic crystals. It was however clear that the pattern of alternating crystal orientations repeats along the vein and is segmented by coarse, recrystallised carbonate layers that truncate crystal zonation along vein margins (Fig 8.7 B, D; 8.8). The truncated zoned crystals oriented with c-axes normal to vein margins had euhedral crystal terminations attached either to the upper or lower vein margin (green arrows in Fig. 8.7 D). Invariably, magnesite veins observed were syntaxial veins of a single growth front and no zonal texture. Where composite veins had formed, the localisation of subsequent crack-seal events was mostly within the wall-rock or at the interface of older veins and the wall-rock.

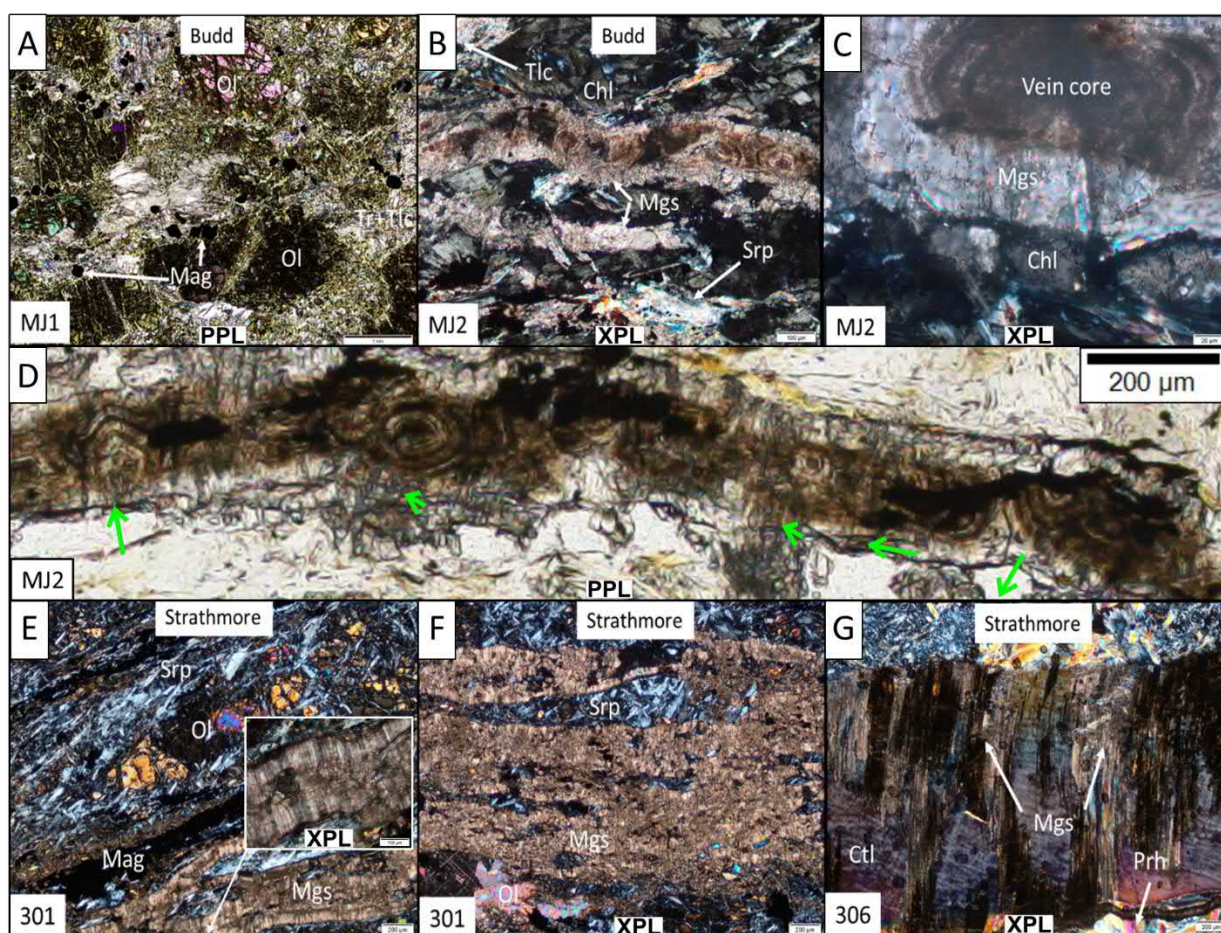


Figure. 8.7 Main microscopic characteristics of magnesite. Suffixes of sample names at bottom left corners are without an NP- prefix. PPL, XPL and RL labels are in photomicrographs produced using plane-polarised, cross-polarised or reflected light, respectively. (A): Serpentinised olivine (Ol) cumulates in a groundmass of serpentine (Srp), magnetite (Mag), talc (Tlc) and tremolite (Tr). (B): Serpentine altered to chlorite bears a magnesite (Mgs) vein made up of a zoned core and coarse outer margins. (C): Close up of a well-preserved, zoned monoclinic crystal from the vein shown in B. (D): Wider context of the crystal array of zoned, monoclinic crystals shown in C shows how individual crystals have their c-axes oriented in varying orientations (green arrows). (E): Magnetite layers aligned with magnesite veins adjacent to serpentinised olivine cumulates. Note syntaxial growth of magnesite crystals in insert. (F): Composite (stretching) vein indicative of multiple crack-seal events. (G): Fibrous magnesite growth after chrysotile (Ctl) asbestos. Prh: Prehnite.

8.4.3 Phase maps

The only phase map produced from a magnesite sample (NP-MJ2) of the present study (Fig. 8.8.) is accompanied by elemental maps (Fig. 8.12) and shows that the zoned vein core described above is comprised of magnesite. Also shown is the dolomitic composition of the marginal phase that surrounds and segments the vein along boundaries of zoned magnesite crystals.

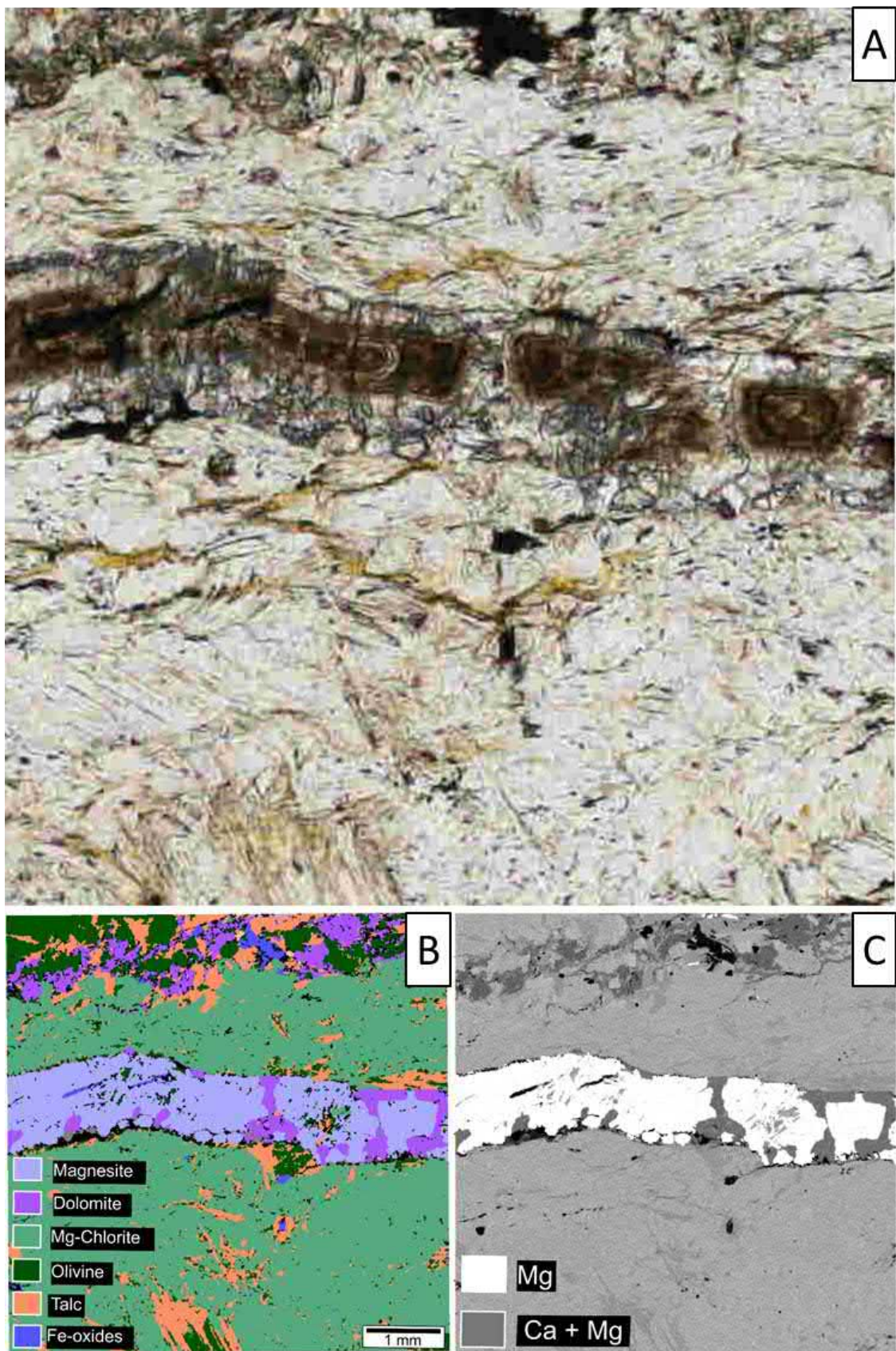


Figure. 8.8 Distribution of phases in a magnesite vein (sample NP-MJ2) from site 8 (Budd deposit). (A) PPL photomicrograph of a segment of a magnesite vein from site 8 respectively corresponding to phase and elemental maps (B). (B-C): The zoned magnesite crystals form a layer segmented by dolomitic phases within a dunite that was first chloritised and then steatised to talc.

8.4.4 Elemental maps

Elemental maps of representative magnesite samples from the northeastern (Fig. 8.9, MJ3, zoned MJ2) and southwestern (Fig. 8.10, MJS) margins of the Stentor Pluton reveal a heterogeneous distribution of calcium. The magnesite veins have a higher concentration of magnesium at cores and no detectable calcium. Higher calcium content occurs along vein margins and in vein-discordant features such as margins between zoned magnesite crystals seen in Fig. 8.8 A.

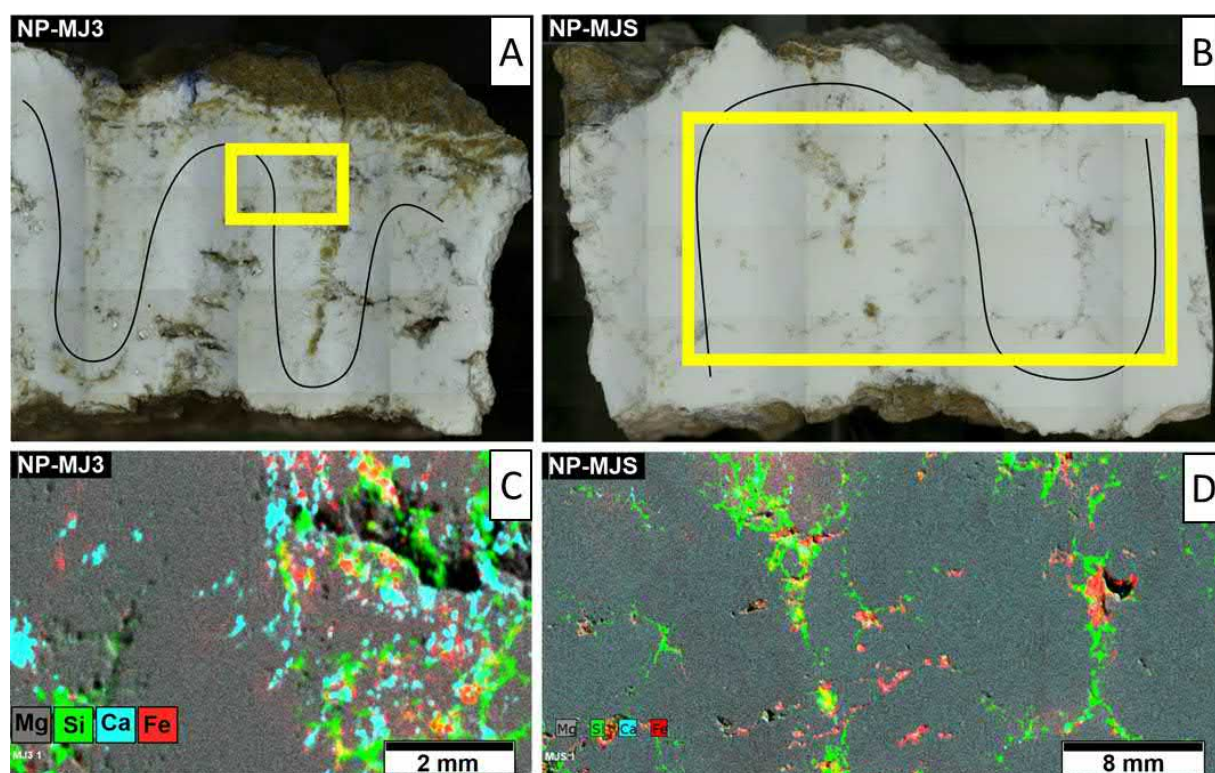


Figure. 8.9 Distribution maps of selected elements on sinusoidal magnesite veins from deposits on the northeastern (A) and southwestern (B) margins of the Stentor Pluton. (A): Well-defined microfolds of softer magnesite from site 8 (Budd) contain calcium, corresponding to a composition towards dolomitic magnesite. The silica content occurs only on the margins. (B): The hardest magnesite sampled (Sugden) is mainly made up of magnesite with minimal dolomitic (calcium) content. Silica is also confined to vein margins.

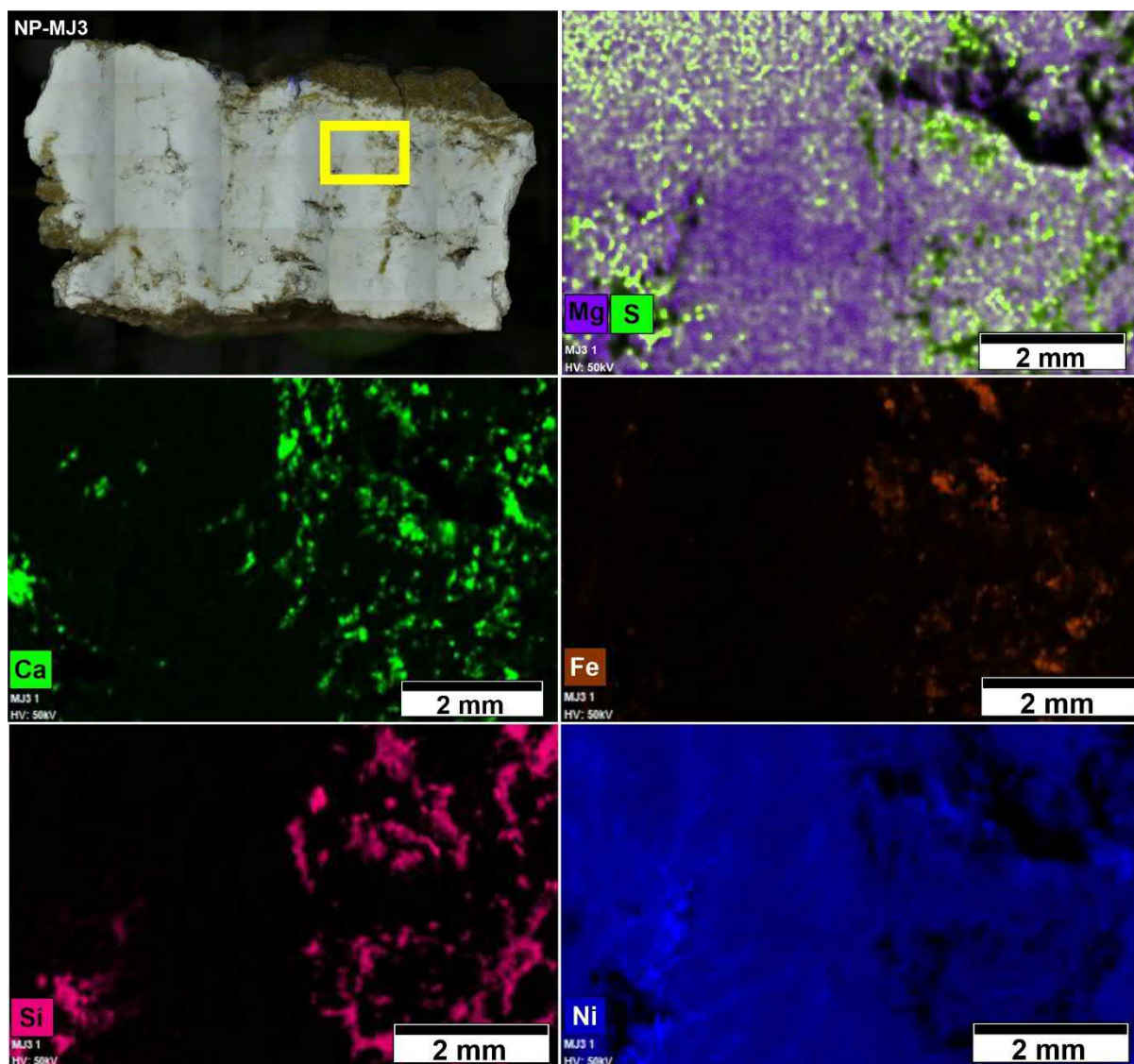


Figure. 8.10 Distribution maps of selected elements in a magnesite vein from Site 8. The maps show that the outer margins of the sinusoidal vein are occupied by concentrations of sulphur associated with calcium, iron, silica and nickel. Nickel and sulphur also occur as diffuse phases within the core of the vein.

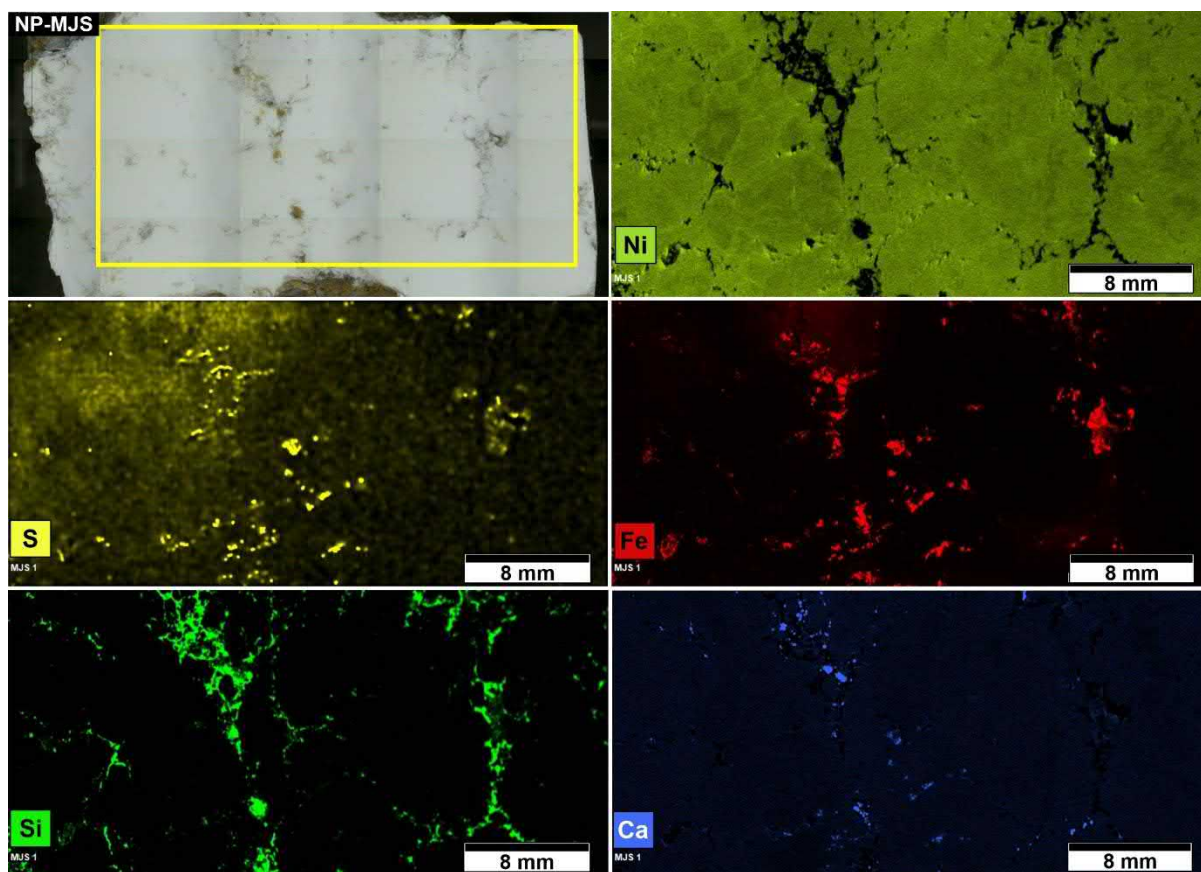


Figure. 8.11 Distribution maps of selected elements in a magnesite vein from the Sugden deposit. The maps show that the outer margins of the sinusoidal vein are occupied by concentrations of sulphur associated with calcium, iron, silica, and nickel. Nickel and sulphur also occur as diffuse phases within the core of the vein.

Two elemental maps of magnesite veins mapped together with the enveloping dunitic host rock (NP-MJ2 in Fig. 8.12; CH-S19 in Fig. 8.13) show that iron and chromium are more abundant in the dunitic host than in the magnesite vein. In addition to forming discrete chromite crystals coincident with higher iron abundance, chromium occurs as a diffuse phase uniformly distributed within a sample of serpentine, chlorite, talc and tremolite. Nickel and strontium are more concentrated within veins than in the host rock.

As noted in the hand specimen, element distribution maps show that the micro magnesite vein of NP-MJ2 is comprised of zoned crystals with margins richer in silica, iron and sulphur. Single element maps of NP-MJ3, NP-MJS and NP-MJ2 show that sulphur and iron occur as iron sulphides. Nickel and sulphur occur as uniformly distributed specks within the magnesite veins but it is difficult to correlate the two due to the microscopic size of the specks. Sulphur also occurs as discrete zones on vein margins that do not coincide with mapped metal elements, this might indicate the presence of a compound composed of sulphur and unmapped elements.

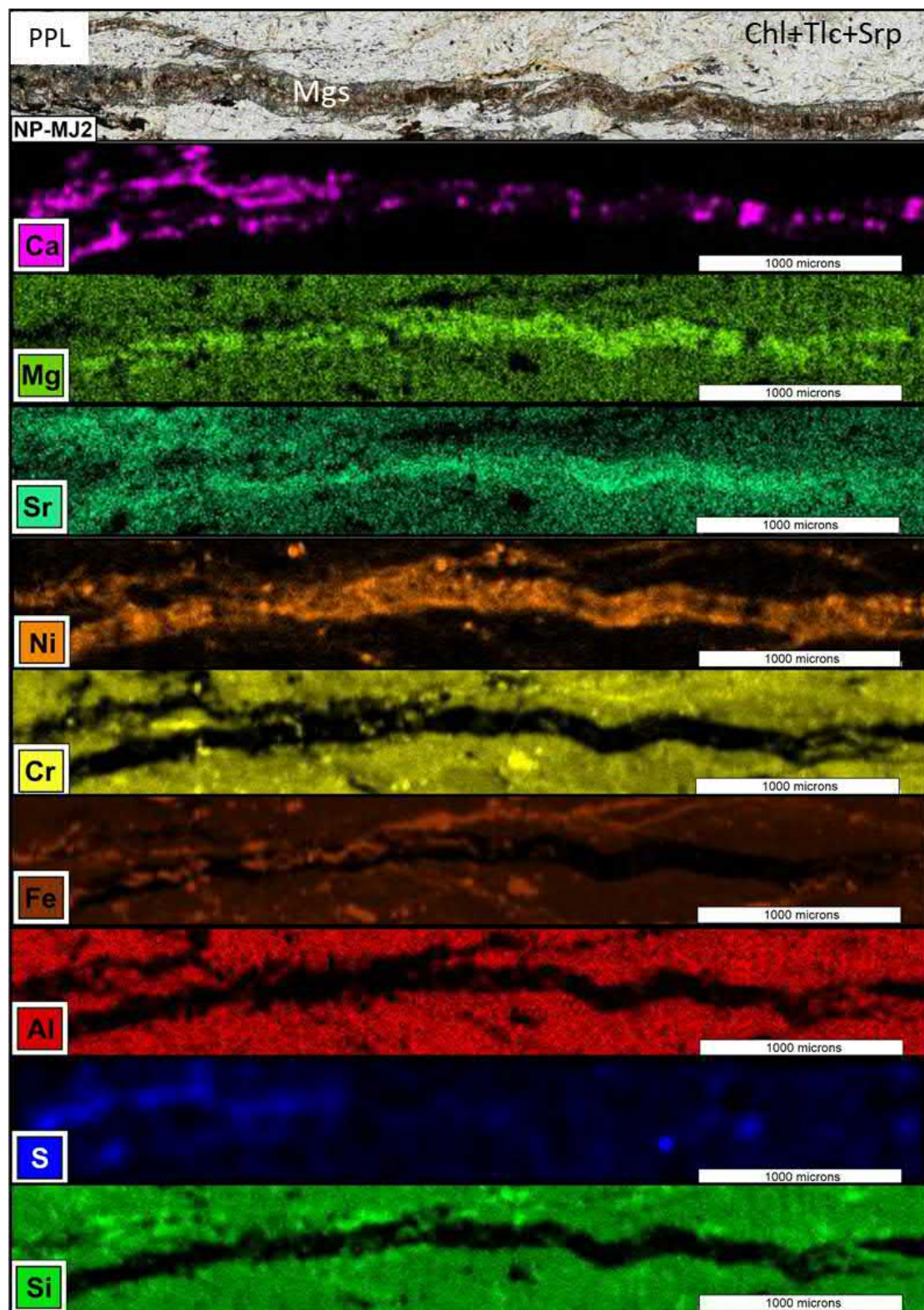


Figure. 8.12 Element distribution maps of a sample from site 8 (Budd deposit). The maps show a partly dolomitic magnesite vein that contains higher magnesite, calcium, nickel and strontium than the serpentinitised, steatitised and chloritised dunite host.

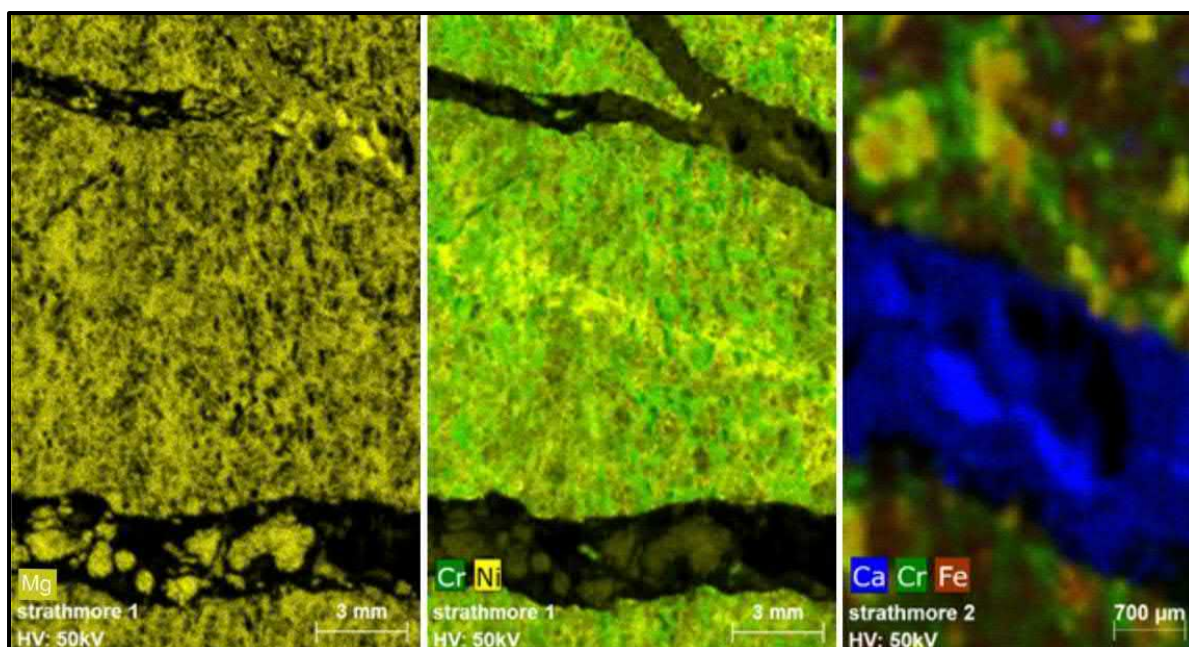


Figure. 8.13 Distribution of selected elements in a magnesite-bearing dunite from Strathmore Magnesite Mine. (A) Magnesium occupies the host rock as well as the cores of two veins at the top and bottom of the map. (B) Chromium and nickel are concentrated only in the host rock. (C) Detailed view of the bottom right corner of maps (A) and (B) shows that calcium occupies most parts of the vein-including in association with the magnesium-bearing clusters. Chromium and iron are only concentrated in the dunitic host rock.

8.4.5 Raman Spectra

All spectra generated from four spots on sample MJS from the Sugden locality have two main peaks at c.330 and c.1100 cm^{-1} respectively (Fig. 8.14). These peaks, according to Larkin (2011), occur within the characteristic range of carbon monoxide bonds and magnesium ions. This confirms that the magnesite in the core of a highly sinusoidal vein from the Sugden locality is indeed a magnesium carbonate compound. The two other spots were comprised of carbonates with peaks at c.1100 cm^{-1} and unconfirmed with peaks at c.320 cm^{-1} and 310 cm^{-1} that may correspond to the presence of calcium and iron, respectively.

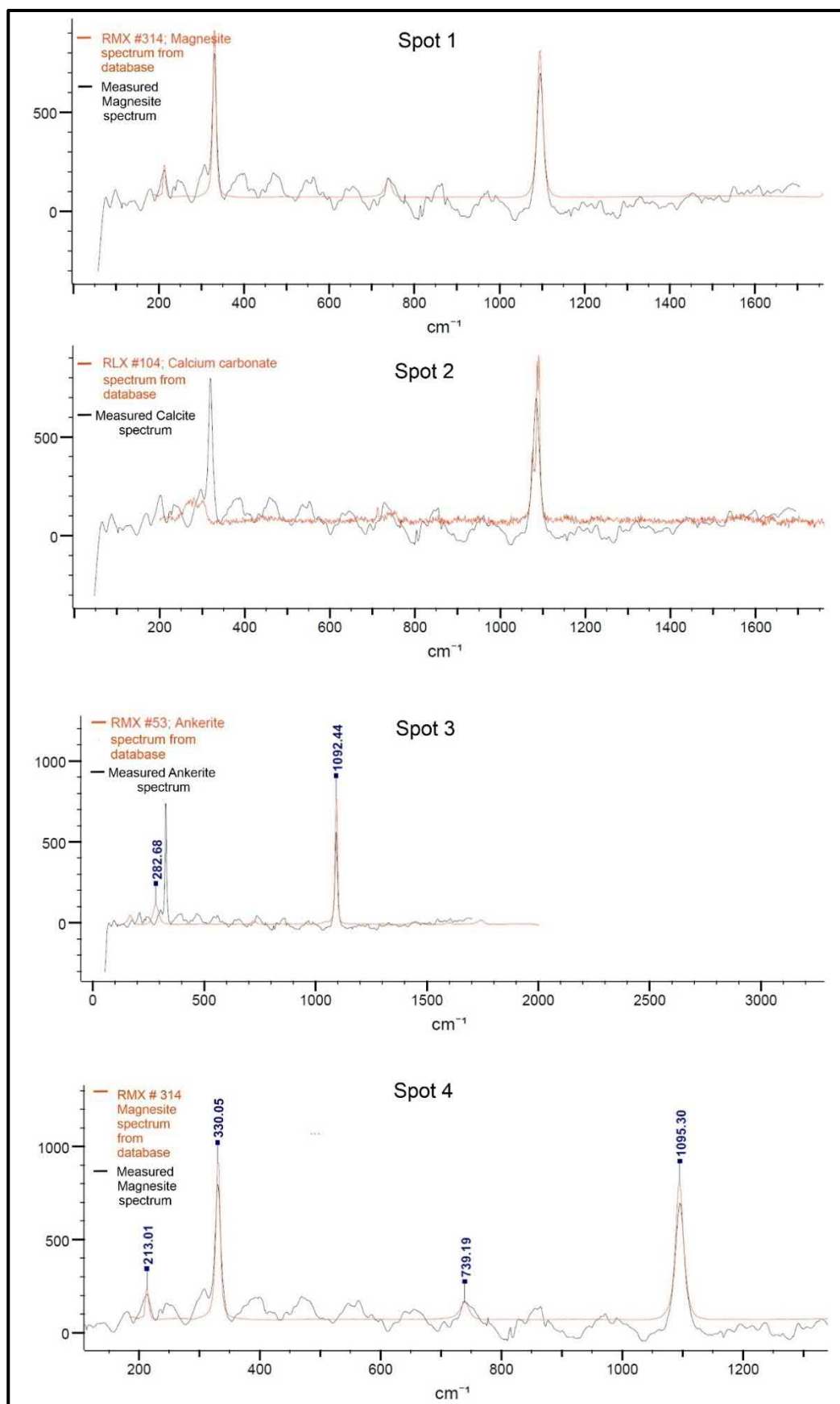


Figure. 8.14 Raman spectra of magnesite sample MJS from the Sugden locality. The spectra show the presence of magnesite, ankerite and calcium carbonate (presumably as a component of dolomite).

8.5 Discussion

The magnesite veins studied either show a syntaxial texture or are amorphous and may form a network of anastomosing, sub-parallel veins separated by thin vestiges of the host-rock. Some amorphous veins truncate zonal-textured magnesite veins, indicating different events of veining. According to the available data, magnesite veining began with an initial sealing of fractures in the wall-rock by a fluid that allowed zoned, euhedral magnesite to develop due to slight, episodic physiochemical changes during crystallisation (Ridley, 2013). Subsequently, syntaxial-textured and amorphous magnesite veins developed at or near interfaces of the wall-rock and earlier veins and were separated from other veins by thin vestiges of wall-rock or older veins (Fig. 8.7 B-F). According to Bons *et al.* (2012), such vein networks indicate episodic extensional fracturing and sealing which in this case occurred mostly at interfaces of older veins and the wall-rock.

Zoning within magnesite veins hosted in Archean ultramafic lithologies is not a widely reported feature. The rare occurrence of an array of variably oriented, zoned monoclinic crystals at the core of a slightly sinuous vein (Fig. 8.7 D) suggests syndepositional changes in the mineralising fluid or differences in crystallisation rate and the diffusion rate of magnesite-forming ions (Shore and Fowler, 1996). The truncation of crystal zoning in the zonal magnesite vein by dolomitic, amorphous veins confirms that zonal veins predate amorphous veins in some localities. No apparent coincidence of microfold limbs or hinges with a certain c-axis orientation of zoned crystals was observed along the sinusoidal vein in MJ2 (Fig. 8.7 D; 8.8A; 8.12). However, the crystal pattern in many places has been obscured by recrystallisation which makes it difficult to explain why the zoned monoclinic magnesite crystals occur in alternating orientations. More such zonal-textured veins need to be studied to investigate the prevalence of this crystal arrangement in sinuous and planar veins.

Also uncertain is the identity of the intermediate magnesite phase. Notably, the rhombohedral nature of magnesite and brucite crystals are well-documented (O'Donoghue, 2006; Montes-Hernandez *et al.*, 2013-and references therein). The Raman spectra, element and phase maps of the above-mentioned vein confirm the presence of a magnesium carbonate compound in the zoned monoclinic crystals. This suggests that some magnesite veins from the Budd deposit may have formed as a pseudomorph of an earlier hydrated or hydroxylated phase that precipitated as monoclinic crystals. This is supported by records of the kinetic inhibition of the direct precipitation of magnesite from an aqueous solution due to its highly hydrophilic nature

(Christ and Hostetler, 1970; Saldi *et al.*, 2009). This presents an inhibition that decreases with an increase in temperature of the precipitating fluid (Saldi *et al.*, 2009; Montes-Hernandez *et al.*, 2012 and references therein; Moore *et al.*, 2014).

Experimental studies of magnesite precipitation from aqueous solutions found that magnesite only precipitated as a dominant phase at temperatures exceeding 120° (Montes-Hernandez *et al.*, 2012 and references therein; Moore *et al.*, 2014; Tanake *et al.*, 2019). These experimental studies found that low temperatures favoured the precipitation of metastable forms of magnesium carbonate because of the kinetically inhibited precipitation of magnesite. The kinetically favoured precipitation of metastable phases produced an assortment of hydrated and/or hydroxylated magnesium carbonates that belong to different crystal systems. Variants of a monoclinic crystal system include lansfordite $\text{MgCO}_3 \cdot 5\text{H}_2\text{O}$, pokrovskite $\text{Mg}_2(\text{CO}_3)(\text{OH})_2$, hydromagnesite $4\text{MgCO}_3 \cdot \text{Mg}(\text{OH})_2 \cdot 4\text{H}_2\text{O}$, dypingite $4\text{MgCO}_3 \cdot \text{Mg}(\text{OH})_2 \cdot (5-8)\text{H}_2\text{O}$, artinite $\text{Mg}_2(\text{CO}_3)(\text{OH})_2 \cdot 3\text{H}_2\text{O}$ and nesquehonite $\text{Mg}(\text{OH})(\text{HCO}_3) \cdot 2\text{H}_2\text{O}$ (Frost *et al.*, 2008; Moore *et al.*, 2014; Lamérand *et al.*, 2020; mindat.org, 2020). Depending on prevailing physiochemical conditions at the time of precipitation, any of these metastable ionic compounds could have been formed initially and later pseudomorphed by magnesite.

The occurrence of highly sinuous magnesite veins in dunite blocks that appear massive may be a result of tectonic or non-tectonic processes (e.g., Shanmugam, 2017). According to Ramberg (1959 and references therein), reasons for a contrast in ductile strain between veins and host rock include ptigmatic folding due to (a) passive syndepositional folding in response to highly irregular surfaces, (b) shortening of a more competent layer within a less competent substrate or (c) forceful injection of fluid into less competent host rock. The lack of highly sinuous fracture planes on the wall rock observed during the present study refutes (a) as a mechanism for tightly folded layers (Fig. 8.6 A); however, this mechanism may apply to the tabular, slightly sinuous veins (Fig. 8.5 E, F). Further, the lack of axial planar cleavage and the multidirectional shortening of multiply oriented magnesite layers reduces the likelihood of a tectonic origin of the folds because of the multiple tectonic events that would be necessary to account for the unlimited number of microfold trends seen in Fig. 8.6 C-D.

Recalling evidence of aqueous magnesite transportation (within fractures) to some final emplacement sites (Van Zyl *et al.*, 1942, page 39-40), slump folding may be a possible cause for the microfolds observed. According to Alsop and Marco, (2013), slump folds may develop even on surfaces of a negligible slope ($<1^\circ$) similar to the predominantly sub-horizontal

magnesite veins studied. However, a cross-sectional view of folded veins does not show increased fold tightness in a certain direction to imply flow of all veins towards a particular direction (a pre-crystallisation down-dip direction).

Detachment folding is another mechanism by which “lifting” and folding of layers detached from bounding surfaces could be achieved (Wallace and Homza, 2004 and references therein). However, individual magnesite layers observed are detached and attached to upper or lower bounding surfaces at short intervals along a single vein length. This suggests that vein bounding surfaces in this case may not be viewed as décollements.

Notably, non-tectonic undulations of sub-spherical surfaces (e.g. botryoidal, nodular, spherulitic, reniform, mammillary) in colloform veins may form because of gel precipitation in the absence of deviatoric stress (Rogers, 1917). However, the tight, non-cylindrical nature of magnesite veins with similar fold amplitudes on either side of the folded layers observed suggests a different mechanism.

Enterolithic folding, viz. diagenetic folds observed, for example, in rehydrated layers of sulphates of shallow (West, 1965; Hussain and Warren, 1989) or deep-water environments (Dean *et al.*, 1975) presents features more reconcilable with observations of the present study (e.g., Gandin *et al.*, 2005, Fig. 15-16). This variant of non-tectonic folding produces tight to open folds and is widely observed in rehydrated layers of gypsum precipitates that were previously dehydrated during burial metamorphism (Murray, 1964; Boggs, 2011). According to early genetic models for enterolithic folding, gypsum nodules form in unconsolidated sediments underlying sabkhas that contain saline fluids concentrated in calcium sulphate (Hussain and Warren, 1989; Gandin *et al.*, 2005; Alçiçek *et al.*, 2017 and references therein). The nodules form due to capillary systems that deliver gypsum-mineralising fluids from the overlying sabkha into pore spaces of unconsolidated sediments in the phreatic zone. Burial metamorphism causes dehydration of the gypsum nodules to anhydrite. Subsequent uplift into a shallower sub-surface environment more accessible to aqueous fluids allows effective hydration of anhydrite nodules within a confined space. The hydration of the anhydrite nodules to produce the same volume of gypsum releases calcium and sulphate ions into surrounding aqueous fluid, allowing lateral growth of gypsum nodules to form a layer. Due to the growing calcium and sulphate ion content in the limited space where the gypsum layer continued to grow, a highly folded habit is assumed by the growing layer to reach chemical equilibrium within limited space. The resulting enterolithically folded layer appears as sinusoidal veins in

cross-sectional view, but in plan view, the layer resembles intestines, as the entero- prefix of the name describes (Alçıçek *et al.*, 2017; Escavy and Herrero, 2019).

Later work proposed other mechanisms for the development of enterolithic folds outlined in Escavy and Herrero (2019). Common among those models is the general occurrence of gypsum at shallower levels than anhydrite in the same deposit. Notably, there are cases of gypsum and anhydrite occurring in the same stratigraphic level due to unclear controls (Murray, 1964). An investigation of the spatial relation of magnesite and hydrated and/or hydroxylated equivalents in the BGB would yield diagnostic features of vein magnesite paragenesis.

From previous studies (Van Zyl *et al.*, 1942; Groeneveld, 1975; Toulkeridis *et al.*, 2010) and observations of the present study, it is known that the folded carbonate compound in veins found at sites of the present study is anhydrous magnesite, and not hydrous/hydrated magnesium compounds. This limits the extent to which the gypsum-anhydrite system may be used as an analogue for the mechanism of magnesite enterolithic folds because the enterolithic folds of the gypsum-anhydrite system occur in gypsum instead of the anhydrous version of the compound. Notably, one of the simpler models to describe the formation of enterolithically folded gypsum involves the dehydration of a gypsum layer/vein to form anhydrite which occupies a smaller volume. If the resulting water remains trapped and in contact with the newly generated anhydrite layer, this layer collapses into enterolithic folds (Flügel, 2004). Since this model allows for the anhydrous version of the CaSO_4 compound to be stable at the surface or shallow levels, it may be a better analogue for the hydrated/hydroxylated magnesite precursor-magnesite system. If this assertion is the case, the tight folds observed in magnesite layers described in the present study may be a result of the following sequence of events.

1. Pre-existing fractures in dunitic lithologies were sealed by a supersaturated, saline fluid concentrated in dissolved CO_2 and magnesium liberated through olivine hydrolysis along dunite fracture planes and crystal margins. Metastable hydrated and/or hydroxylated magnesium carbonates (or possibly sulphates) i.e., magnesite precursors precipitated in preference to magnesite due to thermodynamic controls (discussed above). The magnesite precursor veins are zonal in texture due to syndepositional changes in physiochemical environmental conditions.

Composite magnesite veins formed through a series of “crack-seal” events (Bons *et al.*, 2012, page 44). Cracks formed at or very close to interfaces of the dunite wall-rock and earlier zonal

or amorphous veins, producing veins coated with vestiges of wall-rock on at least one side (Fig. 8.5 E-F).

2. Burial and dehydration of hydrated magnesite precursors caused the formation of (anhydrous) magnesite. Where fractures remained sealed, the dehydration of fracture-sealing hydrated magnesite precursors generated more volume and trapped water that caused the collapse of the magnesite-after-magnesite precursor vein. This intralayer collapse generated variably oriented enterolithic folds (Fig. 8.6).

Suitable P-T conditions allowed the ductile asymmetric folding of tabular magnesite veins which are not enterolithically folded but occur as open folds of a consistent vergence (Fig. 8.5 A, E, F). This may have been due to regional events of sub-vertical shortening.

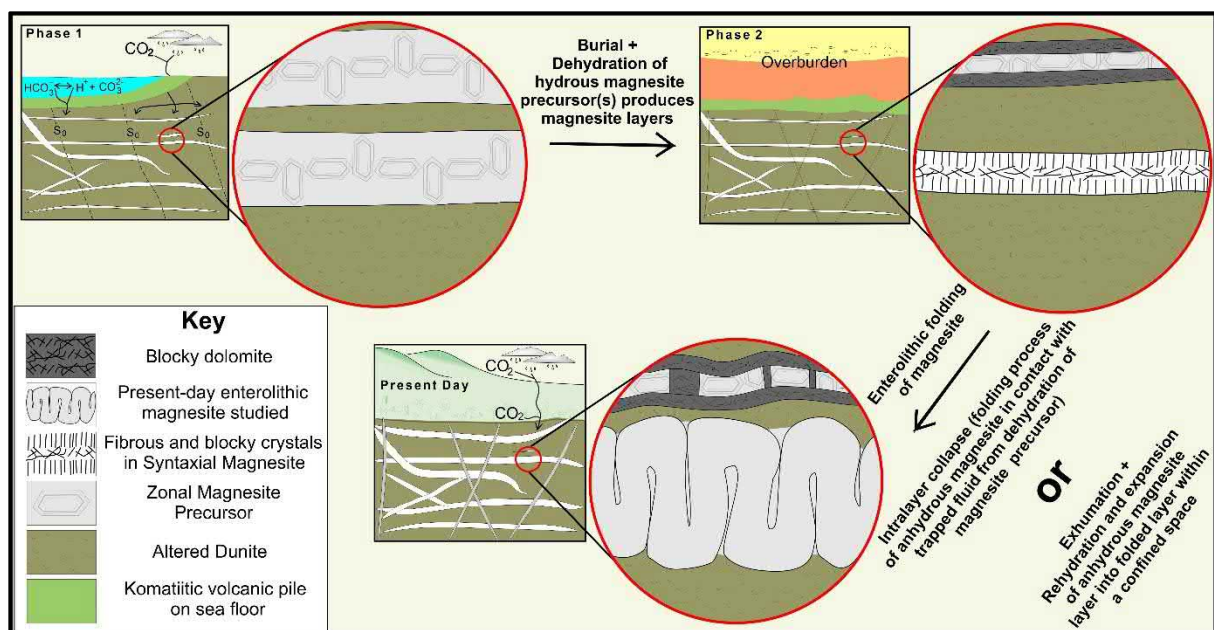


Figure 8.15 Proposed events associated with the development of enterolithic folds observed in selected BGB magnesite occurrences. Hydrated and/or hydroxylated magnesite (magnesite precursor) formed in fractures of altered ultramafic rock and is dehydrated to anhydrous magnesite upon burial. Newly generated water and volume in the original magnesite precursor vein caused the newly produced anhydrous magnesite to collapse into non-tectonic enterolithic folds through a mechanism used to describe enterolithic fold development associated with gypsum-anhydrite dehydration-hydration transitions (Flügel, 2004).

To a limited extent, calcium is known to substitute for magnesium in magnesium carbonates (Evans, 1993). The lack of calcium in the host rock adjacent to the magnesite vein (Fig. 8.12), indicates an external calcium source such as the clinopyroxene-bearing layers of the Budd Layered Ultramafic Complex. Notably, discontinuities such as vein-host rock interfaces, zoned magnesite crystal margins and brittle fractures across the NP-MJ2 magnesite vein are

recrystallised into amorphous dolomitic veins (Fig. 8.8). It is therefore conceivable that the recrystallisation of euhedral, hydrated and/or hydroxylated magnesite included the substitution of magnesium by calcium ions during the circulation of calcium-bearing fluids.

Elemental maps of magnesite veins MJ3 and MJS show silica in vein margins. The silica is accompanied by iron and sulphur of matching texture and spatial distribution. This suggests the presence of pyrite within vein margins. The same veins contain diffuse sulphur. Chromium and nickel sourced from ultramafic lithologies are not equally partitioned into the magnesite veins. Instead, nickel appears more concentrated in the magnesite veins while chromium appears to be concentrated in the host rock. This distribution may be due to the chalcophile nature of nickel and the lithophilic nature of chromium (Robb, 2005). For this reason, nickel and sulphur are observed in vein cores that are invariably poor in chromium.

8.6 Conclusion

Magnesite veins of the Budd, Sugden, Bon Accord and Strathmore localities formed during multiple stages within tensional fractures in altered dunitic layers at shallow crustal levels where brittle deformation could prevail. The fractures were either short and tapered (tension gashes) or continuous. Such fractures were filled by fluids saturated in carbonate ions, providing anions to react with magnesium ions liberated from the dunite host on fracture planes at low temperatures (e.g. $20^{\circ}\text{C} < T < 90^{\circ}\text{C}$ in Montes-Hernandes *et al.*, 2012 or similar in Moore *et al.*, 2014 and references herein). This allowed the initial precipitation of monoclinic-and possibly other forms of metastable, hydrated and/or hydroxylated magnesium carbonate(s) within dunite fractures. The metastable intermediates formed due to the kinetically inhibited reaction pathway required for the direct precipitation of magnesite from Mg^{2+} and CO_3^{2-} ions in an aqueous phase (Saldi *et al.*, 2009; Montes-Hernandez *et al.*, 2012 and references therein; Moore *et al.*, 2014). With all other factors unchanged, an increase in temperature of at least 120°C (in Moore *et al.*, 2014), resulted in the dehydration and/or dehydroxylation of the hydrated and hydroxylated magnesium carbonate(s). This allowed the formation of magnesite pseudomorphs after a monoclinic metastable intermediate. Alternatively, brucite formed during hydrolysis of dunite, may have been carbonated according to equation 1 above. This would have produced a hydrated and/or hydroxylated magnesium carbonate such as the monoclinic compounds listed above. In this way, magnesite would form from the dehydration of the latter compound. Final hydration of magnesite produced enterolithic folds observed at BGB magnesite quarries.

Though highly sinusoidal magnesite veins are of a non-tectonic origin, they fill hydrofractures and pre-existing tensional fractures. The nature of folding observed in the less sinusoidal, more tabular veins of consistent fold vergence remains uncertain but is likely linked to regional deformation events of the Barberton Greenstone Belt.

9. Main Conclusion

The diversity of outcrops exposed along the R38 road make it a useful destination for geology and tourism fieldtrips that aim to expose diverse groups of visitors to topics related to processes that have shaped the Earth and led to the development of certain minerals. With continued research, the subjects superficially discussed in chapters of the present work may reveal unanswered questions about their history, allowing an improved interpretation and easier translation into science outreach material that can be modified to suit various groups of society.

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Appendices

Appendix A: Science outreach geotourism activities of the present project

Appendix B: Chapter 2 chemical data and locations of key verdite occurrences

Appendix C: Chapter 3 structural data

Appendix D: Chapter 6 structural data

Appendix E: Chapter 7 structural data

Appendix F: Chapter 8 magnesite features

Appendix G: Science outreach/geotourism guidebook

Appendix A Science outreach activities

1. Activity: Radio interview of selected team members of the Barberton-Makhonjwa Mountains World Heritage Site project.

Date: 24 September 2020 (Heritage Day in South Africa)

Topic: Local tourism and development (Mr Louis Loock: Mpumalanga Tourism and Parks Agency); Geoheritage of the Barberton Greenstone Belt and the possibility of future discoveries (Phumelele Mashele); Historical and current aspects of the Barberton-Makhonjwa Mountains World Heritage Site (Mr Mark Ngwenyama; Barberton Tourism and Biodiversity Corridor)

Duration: ~ 5minutes

Host: RISEfm Mpumalanga provincial radio station

Reference: info@risefm.co.za



- 2. Activity:** Filming 2x c. 3-minute clips with primary school learners for a T.V. show with an audience of children and teenagers.

Date: 02 March 2021

Topic: Aspects of Barberton geology and the geology career

Duration: 8 hours

Host: SANHU productions for SABC3's 'Hectic on 3'.

Reference/evidence: <https://youtu.be/QhzVHG6QvV4>

- 3. Activity:** Highlighting the geoheritage and geotourism potential of the Barberton region to local-, provincial- and national-level tourism officials who are at liberty to implement developments proposed by the present study. Conveying to UP and Wits geoscience students and lecturers the importance of science outreach in the academic and economic development of communities who live near geological attractions.

Date: 30 April 2021

Topic: Extending the R40 Barberton-Makhonjwa Geotrail towards other tourist attractions.

Duration: 15 min

Host: National Department of Tourism

Reference/evidence:

<https://www.tourism.gov.za/AboutNDT/Publications/Bojanala%20%20June%202021.pdf>

N.B. The article linked above was not drafted or edited by the present author prior to publishing.

“THE MAKHONJWA MOUNTAINS ARE A MAJOR DRIVE FOR BOTH DOMESTIC AND INTERNATIONAL TOURISM”

- DEPUTY MINISTER FISH MAHLALELA



Deputy Minister Fish Mahlalela addressing the students.



University of Mpumalanga students with Deputy Minister Mahlalela and Chairperson of MTPA Board Thulani Nziwa.

PROGRAMME

GEOLOGY AND TOURISM STUDENTS NETWORKING DINNER

DATE: Wednesday, 29 April 2021
VENUE: African Rest Hotel Mbombela, Mpumalanga Province
TIME: 18:00

PROGRAMME DIRECTOR: Mr Thulani Nziwa
CHAIRPERSON: MTPA Board

TIME	ITEM	LED BY
18:00 – 18:20	Registration and hand over of promotional memorabilia	All
18:20 – 18:30	Everyone seated and Arrival of Principals	All
18:30 – 18:35	Welcome Remarks	Executive Mayor: Cllr Sibusiso Mathonisi Mbombela Local Municipality
18:35 – 18:50	Extending the existing Geotrail towards other tourists attractions	Msc Candidate: Ms Phumelele Mashele Wits University
18:50 – 19:15	New discoveries of the future – Based on the drilling project	PHD Candidate University of Pretoria
19:15 – 19:20	Video Play: Scenic Tourism Attractions: Mpumalanga Destination of Choice	MTPA
19:20 – 19:30	Remarks by the MEC	MEC: Hon Vusumuzi Prince Mkhathshwa Economic Development and Tourism
19:30 – 19:45	Address by the Deputy Minister	Hon Fish Mahlalela Deputy Minister of Tourism
19:45 – 19:50	Closing Remarks	Executive Mayor: Cllr Jenda Sidell Ehlanzeni District Municipality
19:50 – 20:00	Photo Opportunity & Media Interaction	Students, Deputy Ministers, MEC, Executive Mayors, MTPA Board Chair, CEO
20:00	Dinner	All
	Social Entertainment	Students

4. Activity: Conveying the geoheritage of the Barberton-Nelspruit region to heritage, tourism and development students from the University of Mpumalanga prior to showing them main lithologies of the Barberton Greenstone Belt along the R40 Barberton-Makhonjwa Geotrail.

Date: 07 May 2021 (presentation); 15 and 22 May 2021 (Geotrail tours)

Topic: Part 1: South Africa’s 10th World Heritage Site

Duration: ~ 2hrs (presentation); 2 x 8hrs (Geotrail tours)

Host: University of Mpumalanga

Reference/evidence: Dr Penny Bernard (penny.bernard@ump.ac.za); Dr Dalifa Ngobese (dalifa.ngobese@ump.ac.za)

Appendix B

Table 1. Abundances of selected oxides of immobile elements in verdite, associated rocks and other alteration products of ultramafic rocks.

Entry details		Relative proportions of Immobile element oxides		
Locality	Label	Al	Mg	Fe(tot)+Ti
Barberton GSB Verdite*	NP-004	95,56	1,25	3,19
Barberton GSB Verdite (Antenen, 1991)	Ant91	86,89	10,05	3,06
Nelspruit Verdite, quarry A	NBV6	89,55	5,61	4,84
Nelspruit Sericite-rich verdite, quarry B	NBN4C	89,04	5,73	5,22
Salisbury GSB Verdite : O'Brien's Claim deposit (Sample 8995, Schreyer et al., 1981), n=1	Sch80S	94,55	1,56	3,90
Salisbury GSB Verdite : O'Brien's Claim deposit (Kerrick et al., 1987), n=6	Ker87SV	81,71	11,85	6,44
Murchison GSB Verdite: Mashishimala (Patridge, 1946), n= No data	Pat46M	87,25	2,42	10,33
Murchison GSB Verdite: Mashishimala (Schreyer et al., 1981) n=2	Sch80M	94,82	0,49	4,69
BGB Verdite chlorite schist verdite host*	HUTH	36,42	54,19	9,38
Nelspruit Verdite chlorite schist verdite host (NBVHP)*	NBVHP	30,22	37,60	32,18
Average of O'Briens chloritite verdite host (Kerrick et al., 1987), n=2	Ker87SV	81,71	11,85	6,44
Average of altered spinifex-textured komatiite (Duchac et al., 1987), n=7	Dch87Spx	76,63	7,09	16,28
Average of BGB rodingites (hibschite) samples, Anhaeusser, 1979), n=5	Anh79Rd	46,87	23,62	29,51
Fuchsitic Listvenites (Hinksen et al., 2017), n=13	Hnk17FL	2,01	82,48	15,51
Proposed altered BGB verdite protolith inside quarry*	HUVH	11,04	57,71	31,26
Proposed altered BGB verdite protolith outside quarry*	HUHP	7,51	59,97	32,52
Noordkaap Type ultramafic bodies, cited in Viljoen and Viljoen (1970), n=No data	Anh69NT	7,42	74,33	18,25
Nodular serpentinised pyroxenite overlying verdite zone*	CBV	6,57	75,20	18,23
Harzburgite dyke intrusive to verdite-host HUVH and HUHP*	SPYH1	11,68	67,81	20,51
Average Pyroxenitic Komatiite, Sawmill Complex LUC, Weltevreden Fm, Onverwacht Gp (Huber and Byerly, 2018) n=6	Hbr18PK	6,61	76,41	16,99
*Individual samples analysed during the present study.				

Table 1 continued.

Average Peridotitic Komatiite, Sawmill Complex LUC, Weltevreden Fm, Onverwacht Gp (Huber and Byerly, 2018), n=8	Hbr18PdK	6,17	77,71	16,12
Average Dunitic Komatiite, Sawmill Complex LUC, Weltevreden Fm, Onverwacht Gp (Huber and Byerly, 2018), n=6	Hbr18DK	4,67	82,70	12,63
Average Komatiitic Basalt, Sawmill Complex LUC, Weltevreden Fm, Onverwacht Gp (Huber and Byerly, 2018), n=2	Hbr18KB	29,18	36,64	34,18
Average Spinifex Komatiite, Sawmill Complex LUC, Weltevreden Fm, Onverwacht Gp (Huber and Byerly, 2018), n=2	Hbr18SK	12,59	67,17	20,24
Average Spinifex komatiite, Pioneer Complex LUC, Weltevreden Fm, Onverwacht Gp (Stiegler et al., 2012), n=5	Sgl12S	35,31	34,44	30,26
Average komatiitic cumulate, Pioneer Complex LUC, Weltevreden Fm, Onverwacht Gp (Stiegler et al., 2012), n=10	Sgl12C	6,65	76,57	16,78
Peridotitic komatiites: Sandspruit Fm (Viljoen and Viljoen, 1969a), n=3	VV69KS	9,25	72,14	18,61
Peridotitic komatiites: Komati Fm (Viljoen and Viljoen, 1969a), n=8	VV69KK	6,88	77,97	15,15
Barberton Type basaltic komatiites (Viljoen and Viljoen, 1969a), n=3	VV69BT	44,66	45,89	9,45
Badplaas Type basaltic komatiites (Viljoen and Viljoen, 1969a), n=5	VV69BP	24,40	68,66	6,93
Geluk Type basaltic komatiites (Viljoen and Viljoen, 1969a), n=4	VV69GL	23,56	70,75	5,69
Metatholeitic basalts (Viljoen and Viljoen, 1969a), n=4	VV69MT	57,93	27,60	14,47
Porphyritic dyke intrusive to BGB verdite zone*	CDWV	13,91	70,78	15,31
Feldspar porphyry intrusive into Komati Fm (Viljoen and Viljoen, 1969b, unit 10), n= 9	VV69FPK	81,19	7,17	11,65
Feldspar porphyry intrusive into Hooggenoeg Fm (Viljoen and Viljoen, 1969b, unit 15), n= 1	VV69FPH	68,07	10,70	21,23
JSB aluminous schist interlayered with metabasalts (Anhaeusser, 1973, 1971; unit 9) n=7	Anh69AS	88,56	2,15	9,29
Carbonatized serpentinitized from verdite quarry*	HUT	8,39	85,10	6,51
Carbonatized serpentinite from Scotia Talc Mine*	NP-STM	0,49	90,98	8,53
Host of fuchsitic listvenite (Hinksen et al., 2017) n=3	Hnk17Srp	2,59	80,58	16,83
*Individual samples analysed during the present study.				

Table 2. Major element bulk composition of Barberton verdite and associated wallrocks from north-eastern quarry in the Handsup LUC determined using XRF spectrometry.

wt%	Barberton GSB Verdite (NP-004)	Nelspruit Verdite, quarry A (NBV6)	Nelspruit Sericitic- rich verdite, quarry B (NBN4C)	Salisbury GSB Verdite : O'Brien's Claim deposit, sample 8995, Schreyer et al., (1980), (Sch80S)	Salisbury GSB Verdite : O'Brien's Claim deposit, Kerrich et al. (1987), (Ker87S), n=6	Murchison GSB Verdite: Mashishimala, Patridge (1946), (Pat46M)	Murchison GSB Verdite: Mashishimala, Schreyer et al. (1980), (Sch80M), n=2	BGB Verdite chlorite schist verdite host (HUTH)	Nelspruit Verdite chlorite schist host, quarry A (NBVHP)	Average of O'Briens chloritite verdite host, Kerrich et al. (1987), (Ker87C), n=2	Proposed altered BGB verdite protolith inside quarry (HUVH)	Proposed altered BGB verdite protolith outside quarry (HUHP)	Average altered BGB spinifex komatiite, Duchac et al. (1987), (Dch87Spx), n=7	Average of BGB rodingites (hibschite) samples, Anhaeusser (1979), (Anh79Rd), n=5	Fuchsitic Listvenites Hinksen et al. (2017), (Hnk17FL), n=13
SiO ₂	59,94	48,54	66,36	44,80	47,27	42,55	40,65	34,00	55,63	38,15	46,06	45,15	85,43	43,08	30,72
Al ₂ O ₃	26,68	32,58	20,97	36,40	31,33	36,05	50,35	18,63	11,21	22,00	4,41	2,95	7,33	12,51	0,70
Fe total	0,05	1,02	0,67	0,30	1,25	2,00	0,14	4,06	11,30	9,27	12,10	12,38	1,15	7,17	5,40
MnO	0,01	0,01	0,01	0,00	0,00	N.D.	0,29	0,07	0,07	0,03	0,16	0,22	0,02	0,12	0,07
MgO	0,35	2,04	1,35	0,60	4,55	1,00	0,26	27,72	13,95	18,85	23,06	23,55	0,68	6,30	28,74
CaO	0,01	-0,02	0,01	0,00	0,81	0,54	1,24	0,69	0,14	0,12	6,83	5,26	0,08	27,81	0,52
Na ₂ O	0,53	0,15	0,09	0,90	0,74	1,31	2,18	1,06	0,00	0,00	0,02	0,01	0,09	0,08	0,03
K ₂ O	7,52	11,25	7,34	9,60	6,89	7,93	1,23	0,01	0,02	1,27	0,01	0,00	2,23	0,03	0,15
TiO ₂	0,84	0,74	0,56	1,20	1,22	2,27	2,35	0,74	0,64	1,32	0,39	0,39	0,41	0,70	0,01
P ₂ O ₅	0,01	0,03	0,01	0,01	0,18	N.D.	0,01	0,01	0,09	0,02	0,03	0,03	0,04	0,13	0,01
Cr ₂ O ₃	1,17	0,27	0,47	1,00	0,67	0,83	0,51	1,74	0,36	722,50	0,34	0,46	0,20	0,00	0,20
NiO	0,00	0,01	0,00	N.D.	0,01	N.D.	N.D.	0,18	0,10	409,50	0,18	0,21	0,02	0,00	0,15
LOI	3,43	4,49	2,92	5,01	5,23	5,66	1,00	11,40	6,85	9,16	6,56	9,38	2,32	0,45	33,55
Total	100,54	101,11	100,77	P.D.	P.D.	P.D.	100,01	100,31	100,36	100,40	100,14	99,98	P.D.	99,84	99,90

N.D.: No data; P.D.: Partial data from previous analysis where the set of elements analysed were different to that of the present study; n: Number of samples analysed, equal to 1 where omitted from row title.

Table 3. Major and trace element bulk composition of Barberton and Nelspruit representative verdite and wall rocks determined using XRF spectrometry.

Major element composition of XRF bulk analysis	Sample name	Elements																
	wt%	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	TiO ₂	P ₂ O ₅	Cr ₂ O ₃	NiO	LOI	TOTAL			
	SPYH1	50,54	4,42	7,57	0,08	25,66	4,40	0,05	0,00	0,19	0,03	0,44	0,24	5,84	99,45			
	CBV	39,91	3,05	8,38	0,15	34,93	0,99	0,03	0,00	0,09	0,02	0,76	0,24	11,40	99,94			
	NVHV	98,39	0,09	0,58	0,01	0,04	0,03	0,01	0,01	0,01	0,01	0,00	0,00	0,27	99,44			
	NBVHP	55,63	11,21	11,30	0,07	13,95	0,14	0,00	0,02	0,64	0,09	0,36	0,10	6,85	100,36			
	HUHP	45,15	2,95	12,38	0,22	23,55	5,26	0,01	0,00	0,39	0,03	0,46	0,21	9,38	99,98			
	NVBG	72,49	14,77	2,30	0,04	0,54	1,61	4,69	2,90	0,32	0,09	0,00	0,00	0,79	100,54			
	CDWV	50,92	5,44	5,77	0,03	27,69	0,34	0,05	0,01	0,22	0,03	0,41	0,25	6,77	97,94			
	HUTH	34,00	18,63	4,06	0,07	27,72	0,69	1,06	0,01	0,74	0,01	1,74	0,18	11,40	100,31			
	NVCV2	75,27	15,15	0,81	0,00	1,08	0,01	0,06	5,48	0,16	0,02	0,00	0,01	2,25	100,30			
	HUVH	46,06	4,41	12,10	0,16	23,06	6,83	0,02	0,01	0,39	0,03	0,34	0,18	6,56	100,14			
	NP-004	59,94	26,68	0,05	0,01	0,35	0,01	0,53	7,52	0,84	0,01	1,17	0,00	3,43	100,54			
	NP-STM	30,02	0,20	3,49	0,06	37,33	0,11	0,09	0,05	0,01	0,01	0,14	0,30	28,66	100,48			
	HUT	39,43	3,56	2,64	0,09	36,09	2,32	0,08	0,03	0,12	0,02	0,23	0,11	15,57	100,30			
Trace element composition of XRF bulk analysis	NBN4C	66,36	20,97	0,67	0,01	1,35	0,01	0,09	7,34	0,56	0,01	0,47	0,00	2,92	100,77			
	NBV6	48,54	32,58	1,02	0,01	2,04	-0,02	0,15	11,25	0,74	0,03	0,27	0,01	4,49	101,11			
		Sc	V	Cr	Co	Ni	Cu	Zn	Ga	Rb	Sr	Y	Zr	Nb	Mo	Ba	Pb	Th
		(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)
	NBN4C	20,62	205,6	2996,77	1,84	35,09	1,51	4,76	23,53	180,31	5,7	3,38	106,73	10,06	0,48	905,35	7,9	3,49
	HUT	15,28	80,2	1502,56	24,14	858,97	27,99	120,55	3,62	1,17	18,97	5,12	5,1	d.l.	d.l.	12,37	2,39	0,44
	NPSTM	0	3,31	816,49	52,22	2115,99	d.l.	14,58	d.l.	d.l.	0,38	0,19	0,94	d.l.	d.l.	0,27	6,05	d.l.
	NP004	20,88	196,28	7497,95	0	12,04	3,39	d.l.	d.l.	70,97	7,55	0,59	39,41	1,25	d.l.	170,01	3,51	d.l.
	NBV6	25,01	208,07	1683,26	5,27	98,48	2,36	5,08	46,51	282,48	10,55	8,39	355,63	28,52	1,3	1564,39	8,67	23,45
	HUVH	25,27	136,42	2308,37	85,8	1528,66	139,59	82,38	6,39	0,55	6,14	6,08	20,89	3	1,95	2,77	4,69	d.l.
	NVCV2	1,99	34,84	16,51	11,58	61,23	1,77	2,79	17,14	154,5	5,82	5,03	119,63	8,25	0,62	691,28	6,27	4,97
	HUTH	68,17	459,74	12532,78	34,5	1426,49	0,85	28,12	8,4	0,24	61,53	6,35	37,27	0,93	d.l.	1,23	4,45	d.l.
	CDWV	25,81	133,52	2828,78	90,46	2000,99	10,59	30,99	5,48	0,75	0,24	6,97	10,77	1,03	0,63	2,31	3,68	d.l.
	NVBG	3,76	16,6	d.l.	4,4	3,83	8,59	38,05	20,3	95,38	270	17,25	256,61	14,3	0,86	491,3	21,35	24,48
	SPYH1	21,14	109,79	3080,26	97,35	1887,15	37,92	107,13	3,51	0,27	3,02	0,78	8,55	0,02	0,78	d.l.	4,44	d.l.
	CBV	10,89	55,11	3838,87	110,1	2016,27	8,72	37,72	1,33	1,89	1,66	1,36	4,21	d.l.	0,16	7,56	3,61	d.l.
	NVHV	0,85	d.l.	3,38	0,71	5,6	5,17	3,7	d.l.	0,34	15,29	-0,13	0,22	d.l.	0,19	13,04	8,77	d.l.
	NBVHP	23,25	168,52	2423,17	64,3	789,88	d.l.	119,22	22,91	1,92	7,25	17,02	139,5	10,38	d.l.	1,91	6,4	5,89
	HUHP	25,24	92,69	3251,36	96,37	1754,31	66,69	67,67	5,6	-0,74	5,64	2,92	19,45	1,57	d.l.	4,98	7,15	d.l.

Table 4. Known verdite quarries within and north of the northern Barberton Greenstone Belt.

Entry no.	Locality	Geological setting	Coordinates (d m s)	
			Latitude	Longitude
1	BGB: Handsup 305 JU east (samples NP-002, NP-003, NP-004, NP-005, NP-006, NP-007, NP-008, HUTH, HUVH, CDWV, SPyH1, HUHP, CBV)	Sheared serpentinites and talc schists of a layered ultramafic sequence (Antenen, 1991).	25°40'59.32"S	31° 3'47.52"E
2	BGB: Clarendon Vale 308 JV east	Sheared serpentinites and talc schists of a layered ultramafic sequence (Antenen, 1991).	25° 41' 24" S	31° 03' 45" E
3	BGB: Clarendon Vale 308 JV west	Sheared serpentinites and talc schists of a layered ultramafic sequence (Antenen, 1991).	25° 40' 48" S	31° 03' 13" E
4	Nelspruit Batholith: Kaapse Hoop 483 JT	Altered serpentinite (Antenen, 1991).	25° 37' 11" S	30° 45' 40" E
5	BGB: Hillside 459JT	Sheared serpentinites and talc schists of a layered ultramafic sequence (Antenen, 1991).	25° 35' 54" S	30° 58' 09" E
6	Nelspruit Batholith: Kangwane (Ngodini)	Sheared ultramafic xenolith in Nelspruit (Antenen, 1991).	25° 19' 13" S	31° 06' 26" E
7	Nelspruit Batholith: Kangwane (2,3 km east of Kruger Mpumalanga International Airport)	Sheared ultramafic xenolith in Nelspruit (Antenen, 1991).	25° 22' 08" S	31° 07' 43" E
8	Nelspruit Batholith: (A) Maggiesdal 456JT (samples NBV1, NBV6, NBVHP, NVCV2, NVFV, NVBG of this study)	Sheared ultramafic xenolith in Nelspruit Batholith (Antenen, 1991).	25°32'30.37"S	30°57'52.47"E
9	Nelspruit Batholith: (B) Maggiesdal 456JT (samples NBN4C, NBN3, NBN4A of this study of this study)	Sheared ultramafic xenolith in Nelspruit Batholith	25°32'21.51"S	30°57'47.10"E
10	Lot 87, 97 section D, 101 of Kaap Valley Block; Handsup 305 JU west; Hillside 459 JT; Oorschot 692JT	Sheared serpentinites and talc schists (Groeneveld, 1975).	No data	No data

Figure 1. Lithologies proximal to the verdite-bearing zone at the Handsup Quarry.

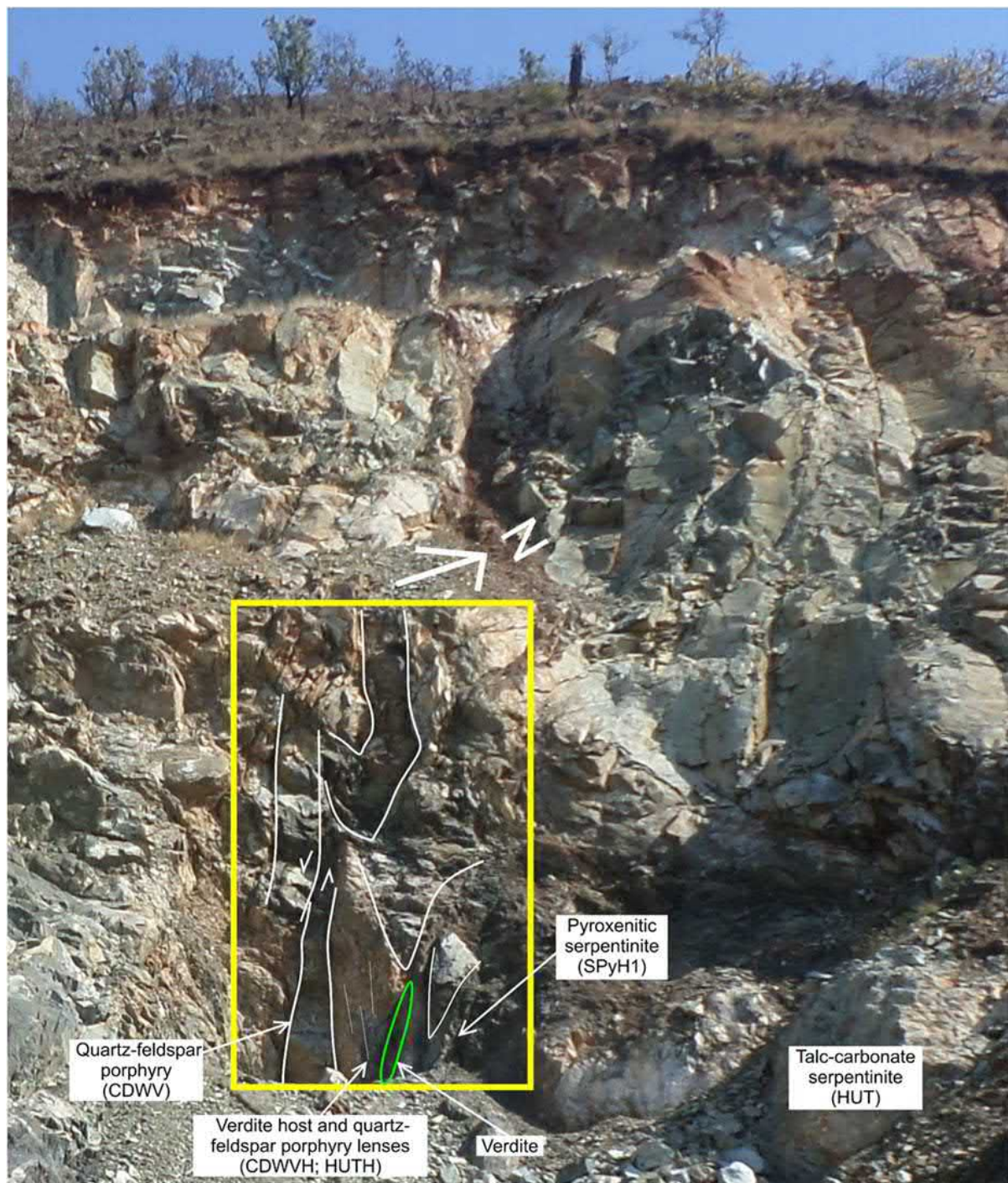


Figure 1. Spatial relationship of lithologies associated with the verdite-bearing shear zone that was mined at the Handsup Quarry. The site of in-situ verdite (green) occurs adjacent to a porphyritic quartz-feldspar dyke that appears to be off-set in a top-to-the-southeast sense.

Appendix C

Table 1. Structural data from Site 3.

Entry no.	Strike of schstosity	Dip	Dip direction	Longitude (dms) E	Latitude (dms) S
1	72	68	SE	31 05 20.76	25 40 25.32
2	60	76	SE	31 05 21.12	25 40 25.32
3	58	64	SE	31 05 21.48	25 40 25.08
4	84	86	SE	31 05 21.42	25 40 25.08
5	94	78	SE	31 05 21.72	25 40 25.08
6	96	82	SE	31 05 21.90	25 40 25.08
7	114	82	SE	31 05 22.14	25 40 25.02
8	84	90	SE	31 05 23.10	25 40 24.54
9	118	86	SE	31 05 23.70	25 40 24.36
10	120	80	SE	31 05 23.82	25 40 24.30
11	82	80	SE	31 05 23.94	25 40 24.24
Entry no.	Azimuth of crenulation fold axis	Plunge	Plunge direction	Longitude (dms) E	Latitude (dms) S
1	74	8	E	31 05 20.76	25 40 25.32
2	62	12	E	31 05 21.48	25 40 25.08
3	90	14	E	31 05 21.42	25 40 25.08
4	82	14.5	E	31 05 21.72	25 40 25.08
5	90	14	E	31 05 21.90	25 40 25.08
6	110	12	SE	31 05 23.7	25 40 24.36
7	89	16	E	31 05 23.82	25 40 24.30
8	80	12	E	31 05 23.94	25 40 24.24

Appendix D

Table 1. Decreasing flattening strain in a conglomeratic horizon along a north-to-south traverse across the northern limb of the Lily Syncline at Site 6B.

Site 1 XZ Plane Clast Axes Lengths (cm)			Clast axes ratio		Site 2 XZ Plane Clast Axes Lengths (cm)			Clast axes ratio		Site 3 XZ Plane Clast Axes Lengths (cm)			Clast axes ratio		Ratios of principal clast axes (a/c) from ZX planes			
Long (a)	Short (c)	(a)/(c)			Long (a)	Short (c)	(a)/(c)			Long (a)	Short (c)	(a)/(c)			Entry no.	Site 1 (Coordinates: 31, 17.158 E; 25, 39.613 S)	Site 2 (Coordinates: 31, 17.153 E; 25, 39.647 S)	Site 3 (Coordinates: 31, 17.153 E; 25, 39.649 S)
56,2	16,2	3,47			27,7	3,6	7,69			5	1,1	4,55			1	3,47	7,69	4,55
8,6	1,8	4,78			39,9	8,1	4,93			4,5	0,8	5,63			2	4,78	4,93	5,63
17,6	3,2	5,50			26,5	4,6	5,76			6	1,8	3,33			3	5,50	5,76	3,33
38,4	1,2	32,00			12,5	1,8	6,94			7,8	0,6	13,00			4	32,00	6,94	13,00
7,1	1,4	5,07			18	3,3	5,45			5,4	0,7	7,71			5	5,07	5,45	7,71
5,6	1,7	3,29			10,4	1,8	5,78			2,7	1,3	2,08			6	3,29	5,78	2,08
2,8	1,9	1,47			18	3,2	5,63			7,5	1,9	3,95			7	1,47	5,63	3,95
3,3	0,6	5,50			3,8	1,8	2,11			9,6	1	9,60			8	5,50	2,11	9,60
8	2,1	3,81			11	1,2	9,17			6,2	1,3	4,77			9	3,81	9,17	4,77
8,6	1,9	4,53			8,2	0,8	10,25			3,7	0,7	5,29			10	4,53	10,25	5,29
10,7	2	5,35			6,5	2	3,25			40,8	8,2	4,98			11	5,35	3,25	4,98
6,4	0,5	12,80			6,2	1,2	5,17								12	12,80	5,17	
11,2	2,1	5,33			4,7	2,6	1,81								13	5,33	1,81	
2,9	1,5	1,93			10,5	2,1	5,00								14	1,93	5,00	
29,3	1,9	15,42			31	3,5	8,86								15	15,42	8,86	
33,7	3,4	9,91			6	1,9	3,16								16	9,91	3,16	
31,2	3,5	8,91			8	1,8	4,44								17	8,91	4,44	
4,1	1,3	3,15			33,5	3,2	10,47								18	3,15	10,47	
12,6	3,7	3,41			7	3,5	2,00								19	3,41	2,00	
9,8	1,1	8,91			22,4	0,9	24,89								20	8,91	24,89	
															Mean	7,23	6,64	5,90

Table 2. Orientation data of mostly roadside outcrops of the Lily Syncline at Site 6A.

Entry no.	Locality	Limb	Strike	Dip	Dip Direction	Longitude E (D,DM)	Latitude S (D, DM)	Lithology	Cleavage Strike	Dip	Dip Direction	Cleavage Type
1	Site 6A (Honeybird Gorge)	Northern Limb	65	72	SE	31, 14.427	25, 40.240	Devitrified grey banded chert	No data	No data	No data	No data
2			68	60	SE	31, 14.425	25, 40.243	Fuchsitic sericite schist	No data	No data	No data	No data
3			76	62	SE	31, 14.412	25, 40.253	Fuchsitic sericite schist with pygmatically folded white quartz veins of an apparent ENE trend (Fig. 6.3 A)	No data	No data	No data	No data
4			72	76	SE	31, 14.462	25, 40.316	Interlayered ferruginous and white banded chert	No data	No data	No data	No data
5			78	72	SE	31, 14.390	25, 40.343	Interlayered ferruginous and white banded chert with grey schist layers increasing towards the north	No data	No data	No data	No data
6			No data: Tools had to be left at lower elevations due to steepness and narrowness of outcrop ledges. Only camera photos and gps data from images could be captured. See images of sub-vertical outcrops in chapter 6.			Between entry 5 and 7		Quartzite with thin shale laminae. Is interbedded with carbonaceous shale and bears rip-up clasts thereof. See chapter 6 for details.	No data	No data	No data	No data
7			75	70	SE	31, 14.433	25, 40.450	Thinly laminated feruginous shale interlayered with white and black devitrified chert	No data	No data	No data	No data
8		Southern Limb	81	72	SE	31, 14.485	25, 40.457	Grey quartzite	78	70	SE	Fold axial planar cleavage
9			74	76	SE	31, 14.517	25, 40.491	Only quartzite with conglomeratic horizon identified along R38 at this locality	67	78	SE	Slaty cleavage
10	Site 6B (Louville Road)	Northern Limb	60	54	SE	31, 17.152	25, 39.626	Quartzite in conglomeratic horizon	No data	No data	No data	No data
11			62	74	SE	31, 17.153	25, 39.635	Quartzite in conglomeratic horizon	58	58	SE	Slaty cleavage
12			69	54	SE	31, 17.150	25, 39.641	Quartzite in conglomeratic horizon	78	66	SE	Fold axial planar cleavage
13			66	68	SE	31, 17.143	25, 39.642	Quartzite in conglomeratic horizon	89	62	SE	Slaty cleavage
14			54	70	SE	31, 17.144	25, 39.643	Quartzite in conglomeratic horizon	72	56	SE	Slaty cleavage
15			69	60	SE	31, 17.155	25, 39.649	Ferruginous banded chert	66	63	SE	Fold axial planar cleavage
16		Southern Limb	62	58	SE	31, 17.154	25, 39.715	Grey and black banded chert	No data	No data	No data	No data
17			68	59	SE	31, 17.155	25, 39.722	Grey and black banded chert	No data	No data	No data	No data
18			71	64	SE	31, 17.154	25, 39.733	Grey and black banded chert	76	73	SE	Slaty cleavage
19			78	65	SE	31, 17.154	25, 39.737	Ferruginous and grey banded chert	80	68	SE	Slaty cleavage
20			84	66	SE	31, 17.158	25, 39.754	Ferruginous shale	60	52	SE	Fold axial planar cleavage
21			83	64	SE	31, 17.163	25, 39.637	Gravelly, quartz and chert-rich sericite wacke	65	53	SE	Fold axial planar cleavage

Appendix E

Table 1 Stretching lineation data from Site 7 (DDM coordinates)

Reading	Azimuth	Plunge	Direction	DDM Coordinates X	DDM Coordinates Y	Details	Phase number
1	222	38	SW	31 14, 557	25 38,991	Banded migmatite	11
2	204	20	SW	31 14, 553	25 38,987	Biotite on quartzofeldspathic grooves	12
3	210	22	SW	31 14, 550	25 38,986	Biotite on quartzofeldspathic grooves	12
4	227	18	SW	31 14,549	26 38,986	Biotite-Quartzofeldspathic interlayered foliations	12
5	250	30	SW	31 14,550	25 38,983	Biotite-Quartzofeldspathic interlayered foliations	13
6	230	12	SW	31 14,547	25 38,982	Biotite-Quartzofeldspathic interlayered foliations	13
7	228	20	SW	31 14,558	25 38,978	Quartzofeldspathic lineation	13
8	222	18	SW	31 14,561	25 38,978	Quartzofeldspathic lineation	13
9	248	22	SW	31 14,558	25 38,977	Quartzofeldspathic lineation	13
10	214	28	SW	31 14,551	25 38,974	Biotite-Quartzofeldspathic interlayered foliations	13
11	181	30	SW	31 14,553	25 38,972	Biotite-Quartzofeldspathic interlayered foliations	13
12	220	18	SW	31 14,555	25 38,960	Hornblende lineation	32
13	204	38	SW	31 14,552	25 38,955	Quartzofeldspathic lineation	33
14	208	22	SW	31 14,551	25 38,955	Quartzofeldspathic lineation	33
15	210	22	SW	31 14,552	25 38,952	Quartzofeldspathic lineation	33
16	223	28	SW	31 14,553	25 38,951	Hornblende lineation	34
17	206	32	SW	31 14,554	25 38,937	Biotite	35
18	206	21	SW	31 14,560	25 38,916	Biotite-Quartzofeldspathic interlayered foliations	36
19	208	30	SW	31 14,556	25 38,948	Biotite	37
20	218	38	SW	31 14,567	25 38,921	Biotite-Quartzofeldspathic interlayered foliations	38
21	240	20	SW	31 14,539	25 38,916	Quartzofeldspathic lineation	38
22	218	24	SW	31 14,560	25 38,906	Biotite-Quartzofeldspathic interlayered foliations	38

Table 1 continued (UTM coordinates)

Reading	Azimuth	Plunge	Direction	UTM Coordinates X	UTM Coordinates Y	Details	Phase number
23	248	46	SW	323591	7161920	Biotite lineation	11
24	263	50	SW	323593	7161918	Biotite lineation	11
25	254	66	SW	323596	7161925	Biotite lineation	11
26	222	18	SW	323595	7161937	Biotite lineation	11
27	218	20	SW	323594	7161927	Biotite lineation	11
28	222	18	SW	323592	7161935	Biotite lineation	11
29	214	32	SW	323592	7161935	Biotite lineation	11
30	208	34	SW	323591	7161936	Biotite lineation	12
31	256	68	SW	323591	7161920	Quartzofeldspathic lineation within fine-grained granitic phase of a banded TTG phase	12
32	248	46	SW	323591	7161920	Quartzofeldspathic lineation within fine-grained granitic phase of a banded TTG phase	12
33	230	22	SW	323596	7161943	Biotite lineation	13
34	212	30	SW	323634	7161991	Biotite lineation	31
35	213	32	SW	323640	7161992	Biotite lineation	31
36	212	10	SW	323636	7161993	Biotite lineation	31
37	263	50	SW	323593	7161918	Quartzofeldspathic lineation within granitic dyke	Not applicable
38	222	18	SW	323595	7161937	Hornblende lineation	32
39	226	22	SW	323584	7161972	Quartzofeldspathic lineation	33
40	227	14	SW	323574	7161990	Hornblende lineation	34
41	220	32	SW	323615	7162021	Hornblende lineation	34
42	222	20	SW	323608	7162027	Hornblende lineation	34
43	212	22	SW	323566	7161997	Hornblende lineation in xenolith	35
44	208	18	SW	323566	7161997	Quartzofeldspathic groove/lineation	35
45	226	20	SW	323562	7161998	Quartzofeldspathic groove/lineation	35
46	224	20	SW	323564	7162000	Biotite lineation	35
47	220	24	SW	323578	7162012	Quartzofeldspathic groove/lineation	35
48	218	10	SW	323575	7162012	Quartzofeldspathic groove/lineation	35
49	198	10	SW	323579	7162010	Quartzofeldspathic groove/lineation	35
50	204	20	SW	323621	7162016	Quartzofeldspathic lineation/groove	35
51	208	12	SW	323568	7162019	Quartzofeldspathic groove/lineation	35
52	206	18	SW	323568	7162026	Quartzofeldspathic groove/lineation	35
53	216	26	SW	323566	7162030	Quartzofeldspathic groove/lineation	35
54	262	48	SW	323577	7161991	Hornblende lineation	35
55	222	18	SW	323577	7161991	Quartzofeldspathic Grove/ lineation	35
56	220	4	SW	323566	7162036	Quartzofeldspathic lineation	36
57	222	12	SW	323557	7162022	Quartzofeldspathic lineation	
58	224	12	SW	323565	7162005	Quartzofeldspathic lineation	
59	202	32	SW	323599	7162034	Hornblende lineation	37
60	212	20	SW	323604	7162027	Hornblende lineation	37
61	212	18	SW	323614	7162034	Hornblende lineation	37
62	216	22	SW	323570	7162003	Quartzofeldspathic lineation	38/40

Table 1 continued (UTM coordinates)

63	225	24	SW	323571	7162001	Quartzofeldspathic lineation	38/40
64	224	32	SW	323575	7162060	Quartzofeldspathic lineation	38/40
65	224	2	SW	323574	7162065	Quartzofeldspathic lineation	38/40
66	222	4	SW	323560	7162053	Quartzofeldspathic lineation	38/40
67	242	8	SW	323563	7162055	Quartzofeldspathic lineation	38/40
68	254	18	SW	323558	7162053	Quartzofeldspathic lineation	38/40
69	206	22	SW	323618	7162040	Biotite lineation	38/40
70	202	20	SW	323602	7162036	Biotite lineation	38/40
71	180	30	SW	323559	7162041	Hornblende lineation	39
72	219	24	SW	323615	7162049	Biotite lineation	38/40
73	202	24	SW	323619	7162047	Biotite lineation	38/40
74	224	20	SW	323629	7162045	Biotite lineation	38/40
75	204	20	SW	323630	7162046	Biotite lineation	38/40
76	218	18	SW	323627	7162048	Biotite lineation	38/40
77	214	18	SW	323627	7162048	Biotite lineation	38/40
78	219	16	SW	323616	7162052	Biotite lineation	38/40
79	214	18	SW	323614	7162049	Biotite lineation	38/40
80	218	22	SW	323604	7162052	Biotite lineation	38/40
81	222	38	SW	323610	7162300	Biotite lineation	38/40
82	222	40	SW	323612	7162295	Biotite lineation	38/40
83	222	18	SW	323561	7162047	Quartzofeldspathic lineation	38/40

Table 2 Mineral foliation data from Site 7

Reading	Strike	Dip	Direction	DDM Coordinates X	DDM Coordinates Y	Details	Phase number
1	132	60	SW	31 14,557	25 39,002	Biotite-Quartzofeldspathic interlayered foliations	10
2	118	60	SW	31 14,555	26 39,002	Biotite-Quartzofeldspathic interlayered foliations	10
3	92	64	SW	31 14,561	26 38,993	Biotite-Quartzofeldspathic interlayered foliations	10
4	103	60	SW	31 14,562	26 38,998	Biotite-Quartzofeldspathic interlayered foliations	11
5	70	54	SE	31 14,553	25 38,986	Gneissic banding (biotite±hornblende, clinopyroxene interlayered with quartzofeldspathic aggregates)	12
6	66	44	SE	31 14,550	26 38,986	Biotite-Quartzofeldspathic interlayered foliations	12
7	78	35	SE	31 14,549	26 38,986	Biotite-Quartzofeldspathic interlayered foliations	12
8	90	32	S	31 14,549	26 38,986	Biotite-Quartzofeldspathic interlayered foliations	12
9	46	56	SE	31 14,556	25 38,983	Biotite-Quartzofeldspathic interlayered foliations	13
10	138	56	SE	31 14,554	25 38,981	Fold axial plane of microfold formed by folded biotite and quartzofeldspathic layers	13
11	94	56	SE	31 14,561	25 38,978	Biotite-Quartzofeldspathic interlayered foliations. Plan of view 1m east of here shows sinistrally folded foliation	13
12	74	48	SE	31 14,559	25 38,979	Biotite-Quartzofeldspathic interlayered foliations	13
13	102	44	SW	31 14,558	25 38,974	Biotite-Quartzofeldspathic interlayered foliations	13
14	118	34	SW	31 14,553	25 38,972	Biotite-Quartzofeldspathic interlayered foliations	13
15	81	36	SE	31 14,551	25 38,974	Biotite-Quartzofeldspathic interlayered foliations	13
16	54	52	SE	31 14,550	25 38,983	Biotite-Quartzofeldspathic interlayered foliations	13
17	85	38	SE	31 14,547	25 38,982	Biotite-Quartzofeldspathic interlayered foliations	13
18	66	38	SE	31 14,559	25 38,979	Biotite-quartzofeldspathic interlayered foliations	13

Table 2 continued

19	80	10	SE	31 14,562	25 38,978	Biotite foliation in thick, weakly foliated aplitic dyke in unit	13
20	114	50	SW	31 14,558	25 38,974	Biotite-quartzofeldspathic interlayered foliations	13
21	106	42	SW	31 14,558	25 38,977	Quartzofeldspathic foliation	13
22	79	30	SE	31 14,555	25 38,960	Hornblende-quartzofeldspathic interlayered foliation	32
23	63	42	SE	31 14,552	25 38,955	Feldspar augens	33
24	62	38	SE	31 14,551	25 38,955	quartzofeldspathic foliation	33
25	71	36	SE	31 14,552	25 38,955	Orientation of feldspar augen in potassic gneiss	33
26	96	40	SW	31 14,554	25 38,954	quartzofeldspathic foliation immediately south of contact with 34	33
27	62	38	SE	31 14,551	25 38,955	Biotite-quartzofeldspathic interlayered foliations	33
28	74	30	SE	31 14,552	25 38,955	Biotite-Quartzofeldspathic interlayered foliations	33
29	60	52	SE	31 14,552	25 38,952	Biotite-quartzofeldspathic interlayered foliations	33
30	86	36	SE	31 14,553	25 38,951	Hornblende foliation interlayered with quartzofeldspathic dykes. Both planar features are folded isoclinally into southeast-dipping extensional folds. Northern layer segments moved up and to the northeast while southern segments moved down and to the south/southeast. Quartzofeldspathic dykes are boudinaged into southwest-plunging boudins that are aligned with hornblende mineral lineations.	34
31	70	44	SE	31 14,583	25 38,944		34
32	72	34	SE	31 14,585	25 38,943		34
33	68	40	SE	31 14,581	25 38,944		34
34	126	50	SW	31 14,556	25 38,948	Biotite-quartzofeldspathic interlayered foliations	36
35	55	32	SE	31 14,560	25 38,916	Biotite-quartzofeldspathic interlayered foliations	36
36	82	34	SE	31 14,560	25 38,914	Biotite-quartzofeldspathic interlayered foliations	36
37	56	38	SE	31 14,560	25 38,913	Biotite-quartzofeldspathic interlayered foliations	36
38	126	50	SW	31 14,556	25 38,948	Biotite-quartzofeldspathic interlayered foliations	36
39	94	30	SW	31 14,554	25 38,937	Biotite-quartzofeldspathic interlayered foliations	37
40	94	30	SW	31 14,554	25 38,937	Biotite-quartzofeldspathic interlayered foliations	37

Table 2 continued

41	100	20	SW	31 14,539	25 38,916	Folded migmatite	38/40
42	44	26	SE	31 14,540	25 38,917		
43	116	24	SW	31 14,538	25 38,920		
44	92	40	SW	31 14,588	25 38,890	Biotite foliation in migmatitic unit	38/40
45	70	62	SE	31 14,594	25 38,883	Biotite foliation in migmatitic unit	38/40
46	54	60	SE	31 14,567	25 38,921	Biotite-quartzofeldspathic interlayered foliations	38/40
47	66	38	SE	31 14,558	25 38,914	Biotite-quartzofeldspathic interlayered foliations	38/40
48	38	56	SE	31 14,562	25 38,909	Biotite-quartzofeldspathic interlayered foliations	38/40
49	59	52	SE	31 14,561	25 38,908	Biotite-quartzofeldspathic interlayered foliations	38/40
50	110	56	SW	31 14,561	25 38,906	Biotite-quartzofeldspathic interlayered foliations	38/40
51	66	62	SE	31 14,560	26 38,906	Biotite-quartzofeldspathic interlayered foliations	38/40
52	81	54	SE	31 14,561	26 38,904	Biotite-quartzofeldspathic interlayered foliations	38/40

Appendix F

Table 1. Key features of magnesite veins at localities studied.

Locality Name	Geographic location (D,M,S)	Distance from the Stentor Pluton	Significance	Host rock	Magnesite Features	Characteristic deformation Features
Budd deposit (Site 8)	25, 33, 23.46 S; 31, 19, 33.18 E	539 m in the northeastern margin	Is the most accessible historical mine of the four deposits described. Has a local name and occurs nearby landmarks linked to cultural and industrial heritage of the area.	Serpentinised and chloritized dunite with abundant iron hydroxides on rock surfaces.	Dirty white and moderately hard. Predominantly sub-horizontal with lesser sub-vertical and diagonal veins. Millimeter- to centimeter-scale veins are folded in an open or tight manner. Thicker sub-vertical or diagonal veins crosscut thinner, sub-horizontal veins. Few 3-D exposures of tension gashes identified dip northeastwards. Late-stage ofolding of some thick veins suggests top-to-the-southeast shortening during northwest-southeast compression.	Later shortening produced north-dipping structures that verge towards the southeast.
Sugden	25, 40, 6.36 S; 31, 13, 22.32 E	1.67 km in the south-southwestern margin	Historical, abandoned magnesite mine in a remote locality south of Sugden train station in Sugden Siding.	Nodular, advancely altered host almost completely made up of iron hydroxides and silicified nodules of serpentine.	White and hardest veins observed (sparked when struck with a geopick). Predominantly sub-horizontal with lesser sub-vertical and diagonal veins. Millimeter- to centimeter-scale veins are folded in an open or tight manner. Thicker sub-vertical or diagonal veins crosscut thinner, sub-horizontal veins. Abundant tension gash arrays that step down towards the southeast are filled with magnesite that occurs as tightly folded layers of a northeast-southwest trend.	Abundance of southeast-dipping tension gashes filled with folded magnesite layers of a northeast-southwest trend.
Bon Accord	25, 40, 11.16 S; 31, 09, 52.26 E	464 m in the southwestern margin	Previously undocumented locality at the western wall of a dam of the Kaap River, c. 500m west of Scotia Talc Mine and 6 km west-southwest of the Sugden deposit. Access made possible by the currently low dam water level.	Best preserved dunite of an olive green-brown colour has well-preserved asbestos veins that predate magnesite veins. Hand specimens display delicate micro-scale magnesite veins can be seen around crystal boundaries.	White, moderately hard. Predominance of sub-horizontal veins intercepted by sub-vertical veins and diagonal veins. Thick diagonal veins are comprised of individual tapered magnesite veins separated by chlorized dunite host rock. Other thick veins are comprised of continuous parallel magnesite veins pulled apart to produce a layer-parallel void through extensional top-to-the-southeast shearing that produced northeast-southwest-trending folds on the separated magnesite veins. Subvertical asbestos veins share a southeast dip direction with sub-vertical magnesite veins that cut across them at a low angle. Other asbestos veins share a northwestwards dip with longitudinal magnesite veins at the core. This composite veins system was thrust in a top-to-the southeast sense during an episode of brittle deformation.	Southeast-dipping extensional folds in tabular, parallel magnesite layers separated by voids or hostrock. Northwest-dipping thrust fault through asbestos vein, and lined with magnesite.
Strathmore	25, 31, 49.62 S; 31, 26, 26.22 E	8.65 km from the northeastern margin; 4.08 km south of the Nelspruit Batholith	Largest deposit of the region. Mined for more than a century and is still operational.	Serpentinised dunitess with asbestos veins coinciding with prehnite-bearing zones.	White and moderately hard. Predominance of sub-horizontal veins intercepted by diagonal veins and-to a lesser extent-sub-vertical veins.	Predominance of sub-horizontal veins. Veins mostly open folded.

Table 2. Key petrographic features of magnesite veins at localities studied.

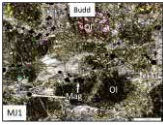
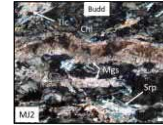
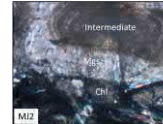
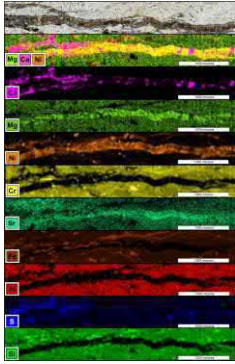
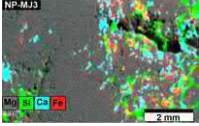
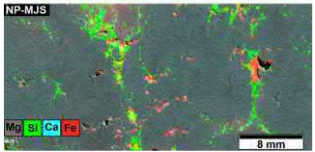
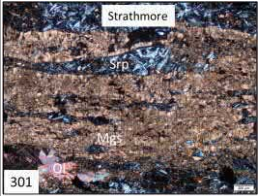
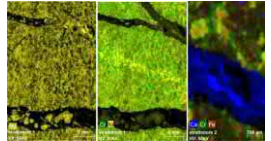
Site 2: Barberton Verdite									
Entry no.	Sample name	Sampling location (D.D.M)	Field description	Field image	Thin section description	Image of characteristic features	Modal abundances of minerals (%)	Elemental map or phase map?	Map(s)
1	NP-M11	Budd, 25, 33.376 S; 31, 19.553 E	Dark grey-green, medium-grained serpentized dunite with iron oxide and hydroxide powder on olivine crystals. Sample free of magnesite veins.	See thesis chapter 8	Partly serpentized olivine orthocumulate with unaltered and stratified interstitial minerals replaced by tremolite intergrown with talc. Magnesite occurs at olivine margins and along mineral fractures. Subhedral magnetite forms clusters at margins of serpentized olivine and in the interstitial spaces.		Olivine: 55%; Matrix of intergrown Tremolite and Talc: 30; Magnesite along fractures and margins of olivine cumulate: 15	No	N/A
2	NP-M12	Budd, 25, 33.388 S; 31, 19.552 E	Dark green medium-grained chloritized dunite with a thin (micrometer scale) magnesite vein.	See thesis chapter 8	Pervasively chloritized serpentine interstitial to irregular patches of talc (see phase map). The magnesite vein in this sample is segmented into short amorphous segments that separate long segments that contain zoned monoclinic crystals of alternating orientations. Adjacent zoned crystals may have the c-axis aligned with- or normal to vein margins. Focusing on the two well-preserved crystals near the left and right edges of the third picture, note the euhedral terminations of crystals attached to the upper and lower margins of the same vein.	  	Chlorite: 30; Serpentine: 20; Magnetite: 2; Talc: 18; Magnesite veins: 10	Elemental maps and phase map	
3	NP-M13	Budd, 25, 33.391 S; 31, 19.553 E	Tightly folded cryptocrystalline magnesite vein with silica-bearing vein margins at the top and bottom part of the vein.	See thesis chapter 8	No thin section	N/A	N/A	Elemental map	 
4	PN	Budd, 25, 33.395 S; 31, 19.554 E	Cryptocrystalline magnesite vein with quartz encrusted on vein surfaces. The vein is folded in three dimensions and fold axes suggest shortening in several directions within the same sample without the development of interference reminiscent of orogenic folds.	See thesis chapter 8	No thin section	N/A	N/A	No	

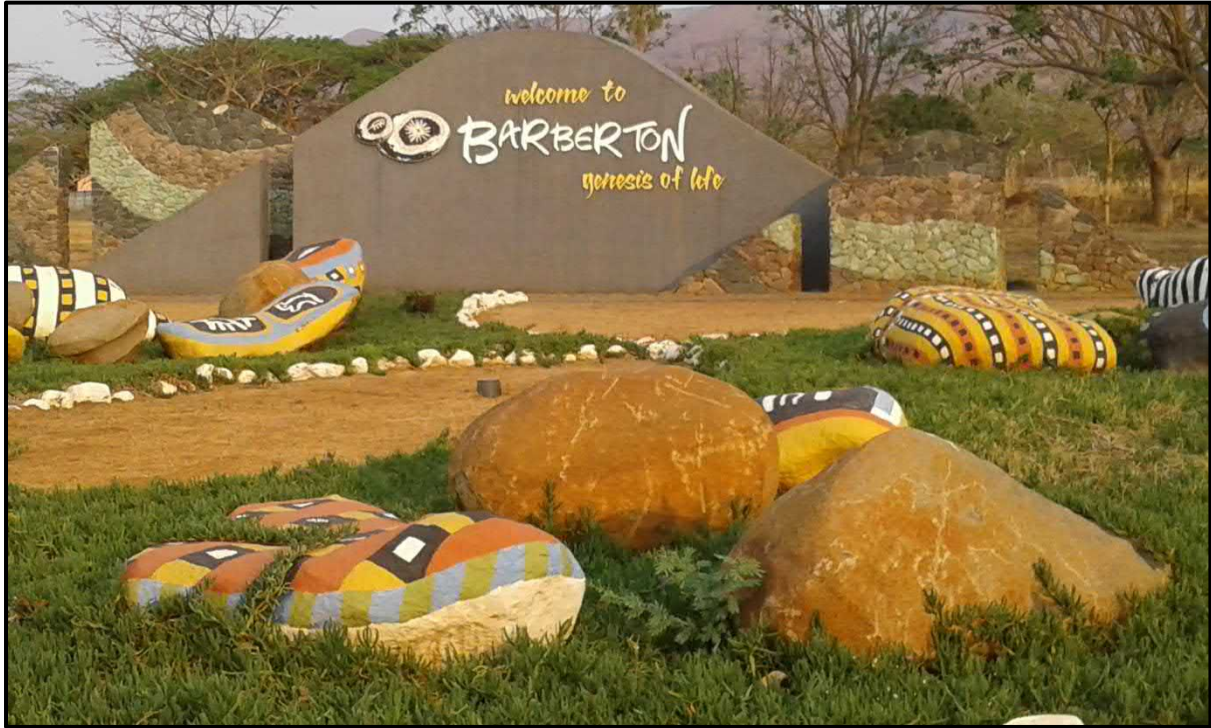
Table 2 continued.

5	NP-MJ5	Sugden; 25, 40.106 S; 31, 13.372E	Cryptocrystalline magnesite vein that is tightly folded by has poorly defined vein margins that are made up of nodular magnesite	See thesis chapter 8	No thin section	N/A	N/A	Elemental map	 
6	NP-301	Strathmore Mine; 25, 31.827 S; 31, 26.437 E	Serpentinized dunite cut across by multiple magnesite veins		Advancely serpentinized olivine cumulate invaded by composite magnesite veins that are separated by thin vestiges of serpentine and rarely by partially serpentinized olivine. Magnesite vein margins are sometimes lined by linear aggregates of euhedral magnetite crystals		Olivine: 30%; Serpentine: 47; Magnetite: 3; Magnesite: 20	No	
7	NP-305		Serpentinite cut across by a vein of prehnite that it in turn cut across by a vein of chrysotile asbestos	Float sampled to describe a suspected presence of magnesite in the chrysotile asbestos vein.	Moderate serpentinization of cumulates along margins, fractures and cleavage planes. Pseudomorphic replacement of chrysotile asbestos by magnesite along a vein.		Olivine: 35%; Enstatite: 10; Diopside: 5; Serpentine: 30; Prehnite vein: 10%; Chrysotile asbestos+ Magnesite: 8%; Magnetite: 2%;	No	
8	CH-519		No data	No data	No data	No data	No data	No data	

Reading Rocks along the R38 Geotrail

Gateway to the Barberton-Makhonjwa Mountains

World Heritage Site



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WITS
UNIVERSITY



BACKGROUND

English

UNESCO World Heritage Sites are world-class places of cultural, historical, and/or natural heritage. An increasing number of UNESCO World Heritage Sites are those recognised mostly for their geological significance. In 2018, part of the Barberton region was granted UNESCO World Heritage Site status because of the important information contained in rocks of this region. They contain the oldest best-preserved rock sequences ever formed on the Earth. They also contain some of the earliest signs of life visible to the naked eye. These rocks preserve clues about the composition of the early atmosphere and oceans, their temperature, the role of volcanism, meteorite impacts and so much more! All this is known because of the many years spent by dedicated scientists who learned how to read rocks of this region like a book! Experience the world in a different way by learning how to read rocks along the R38 geotrail!

siSwati

Tindzawo letingemagugu emhlaba wonkhe ngekugcotjwa nguMhlab'uhlangene nguleto letibaluleke ngekwemasiko, umlandvo noma indalo letfolakala kuto. Kantsi kuleminyaka, netindzawo letigcame ngekucuketsa ematje labalulekile setiyavama kuniketwa leso sihloko. Kusukela emnyakeni wa 2018, incenye yetintsaba tase Mjindini nemaphetselo setatiwa ngekuba ngulelinye lemagugu emhlaba wonkhe. Loku kungoba lawomatje ayincenye yematje ekucala lakheka uma lwandle nemhlaba kuhlukana, lokuya kusengakabi khona tilwane noma bantfu. Nanoma lamatje amadzala kundlula lamaningi eveni lonkhe, matje akulenzawo agcineke ngendlela lengakavami ngobe asacuketse tinkhomba letisifundzisa lokunengi ngesimo semhlaba ngalesikhatsi wona akheka. Eminyakeni leyendlulile, tifundziswa letati kabanti ngematje tabona tinkhomba tetilwanyana lebetitingushe tekucala letagcineka ngekhati kwematje akulenzawo. Kundlula lapho, kwabese kutfolakala kutsi ematje akulenzawo asifundzisa ngekwakheka kwemoya nemhlaba lesiphila kuwo. Tibonele kutsi ematje afundzwa kanjani ngakulomgwaco loyi-R38!

Afrikaans

Gebiede wat as wêrelderfenisgebiede verklaar deur die Vereenigde Nasies se OWKOVN is gekies vir hulle uitstekende kultuurerfenis, geskiedenis, en/of biologiese verskiedenis. 'n Groeiende gedeelte van OWKOVN wêrelderfenisgebiede word erken vir die waardevolle gesteente wat hulle bevat. Sedert 2018 was die Barberton-Makhonjwa berge verklaar as 'n wêrelderfenisgebied vir hulle unieke gesteentes. Die rede daarvoor is hierdie berge is een van die wêreld se eerste stabiele landvorms wat op die Aarde gevorm het. Binne in dié berge se gesteente sit die wêreld se oudste bewyse van lewendige organismes. Ook daarin lê verskillende aanduidings wat voorstel hoe ons landvorms en suurstof-ryk lug gevorm het. Ons eer die wetenskaplikes wat hierdie belangrike ontdekkings gemaak het. Hulle het gewys hoe die geheime in ons gesteente kon geles word soos woorde in 'n boek. Kom ons lees gesteentes soos boeke langs die R38 nes hulle!

Simplified glossary of key terms (0A; 0B):

(0A) An example of a code used after certain sentences in this guidebook that is connected to a source of information (books, journals) listed against their codes at the end of this guidebook.

Element A group of charged and neutral particles that make up the simplest form of matter.

Matter A substance made up of elements. Matter may be in the form of a solid, liquid, gas.

Mineral A type of solid matter made up of a single or group of elements arranged in a set way.

Crystal A mineral with well-developed faces.

Rock A hardened cluster of one or more types of minerals and/or pieces of stones, shells, or plants.

Composition What something is made up of.

Fluid Matter that can flow.

Lava Melted rock from the depths of the Earth which flows up towards the solid outer layer of the Earth through deep cracks and flow out to the surface of the Earth through of volcanoes on dry land or the ocean floor.

Magma Melted rock from the depths of the Earth which can flow up towards the solid outer layer of the Earth through deep cracks but could not flow out to the surface of the Earth

Volcano A vent through which fluid rock flows from deeper parts of the Earth to cool and form new rocks on the surface of the Earth.

Intrusive A fluid that flows into solid or partly solid matter.

Granite/Granitic Used here to describe an intrusive rock mainly made up of coarse crystals of black or dark green platy or tabular crystals surrounded by translucent, white and/or pink minerals.

Craton One of the stable portions of a continent which are mostly made up of granitic bodies.

Greenstone A rock that contains green minerals which grew from volcanic fluids or during later changes in temperature, pressure or overall composition of a rock after exposure to an external fluid.

Sediments Crystals or Pieces of hard material (rocks, shells, bones) which may be picked up, moved and dropped by agents of erosion such as water, wind or streams of ice.

Fold Folded stack of rock layers or mineral layers within a single rock layer.

Syncline Folded stack of rock layers where the youngest layer is found at the core of the fold.

Geologist Scientists who study rocks and minerals, and the processes that form or change them.

Syncline Concave-up structure of folded rock layers where the youngest layer occurs at the core of the fold.

Anticline Convex-up structure of folded rock layers where the oldest layer occurs at the core of the fold.

Synform A general concave-up structure of folded rocks.

Antiform A general convex-up structure of folded layers.

Fault A crack in rocks along which movement of rock has occurred.

Ga giga annum, that is, 1 000 000 000 years, which is a thousand million years.

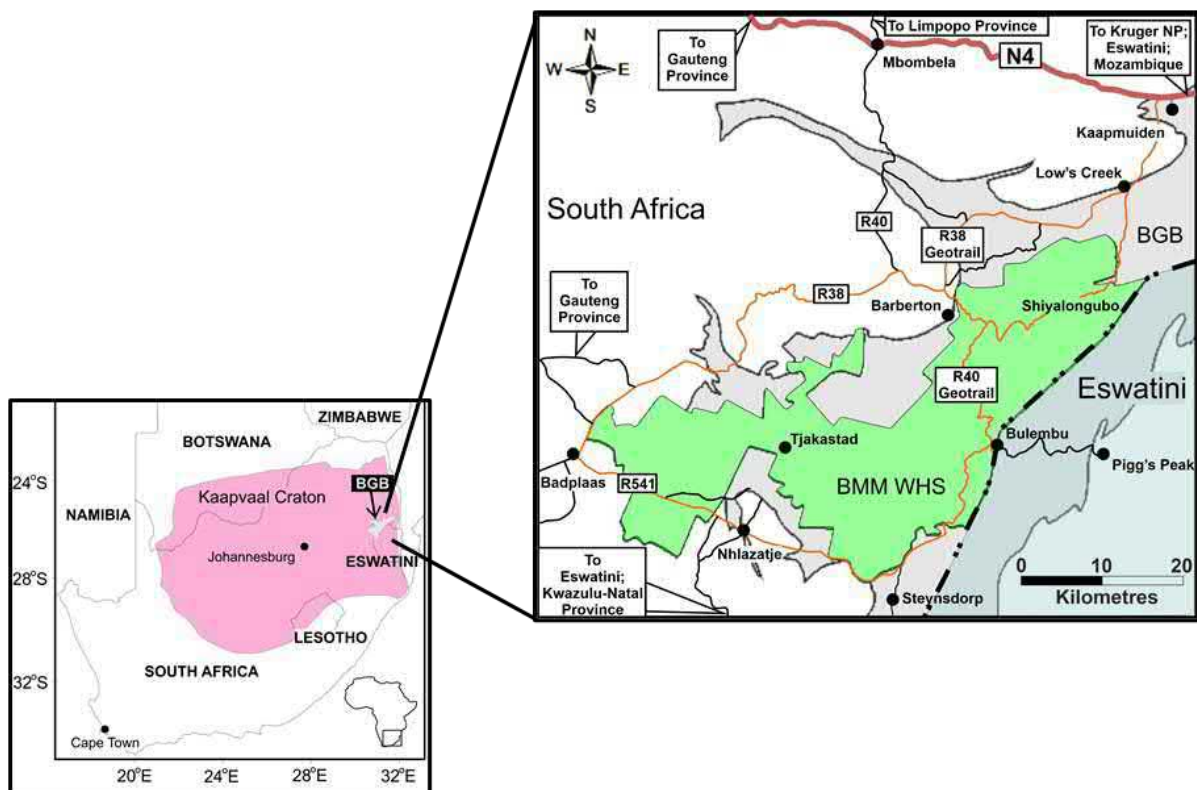
Vertical forces Push or pull forces that act along an upright or almost-upright direction.

Horizontal forces Push or pull forces that act along a flat or almost-flat direction.

Introduction

The town of Barberton was developed in 1884 during a gold rush that predates the Witwatersrand goldrush by two years (0-1A). The discovery of world class accumulations of gold during the 1880s brought the world's attention to the town of Barberton (0-1B). Once again, the town gains international attention as nearby mountains have become a UNESCO World Heritage Site (2018).

The fully developed R40 Barberton-Makhonjwa geotrail cuts across most of the important rocks of the region. Along this road, visitors can view rocks and information panels which describe the early Earth. Rocks along the developing R38 geotrail display structures and minerals formed when ancient rocks were exposed to different forces and fluids from deeper parts of the Earth. Situated just outside the Barberton-Makhonjwa Mountains World Heritage Site (BMM WHS), the R38 provides an educational gateway into the BMM WHS. It is also a connecting route between the BMM WHS and other popular tourist attractions.



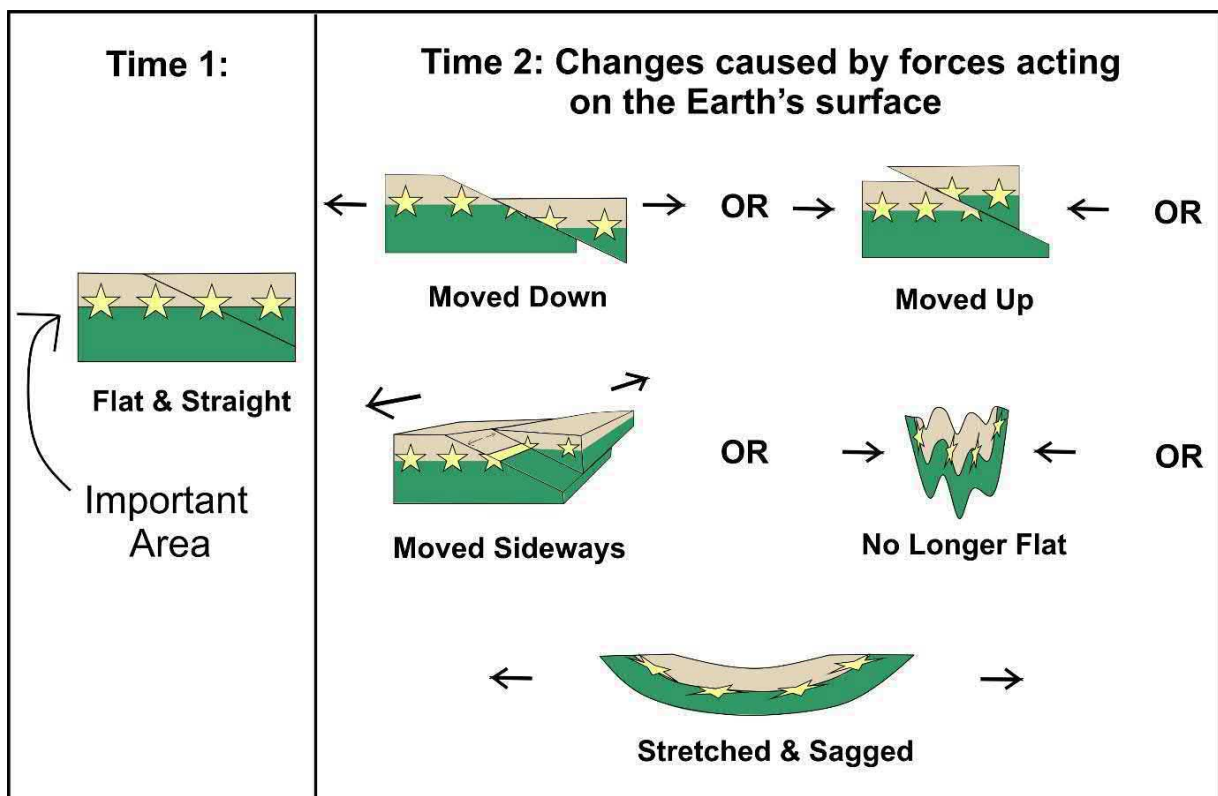
Linked to R40 and R541 roads, the R38 forms a road network locally known as the 'Genesis Route' which forms a figure of 8 (orange route above). This route allows access to most of the important rocks in the Barberton region and intercepts undeveloped cultural and natural sites with potential to draw visitors of diverse interests (0-1C).

Mental Tools for the R38 Geotrail Experience

1. Why would looking at rocks be fun?

Like crime scene detectives, geologists look for clues to find out how and where important rocks were formed. This is not always easy because we live in a planet that has had a long history of landscape changes caused by active processes in the deeper parts of the Earth (0-1D). Those processes have caused push or pull forces on the solid outer surface of the Earth, long before humans occupied the Earth's surface (0-1D). Active push or pull forces continue to shape our planet by causing rocks to break, move, stretch and fold. In some cases, these changes have caused rock blocks of interest to move to a hidden area in the crust of the Earth. This puts a geologist on a "hunt" for the hidden block of interest, especially if the block contains an important area. Depending on the type of geologist, the important area may be a layer with important fossils, a dangerous weak zone of rocks to be avoided or a mineral or sediment layer that tells of important processes of the early Earth.

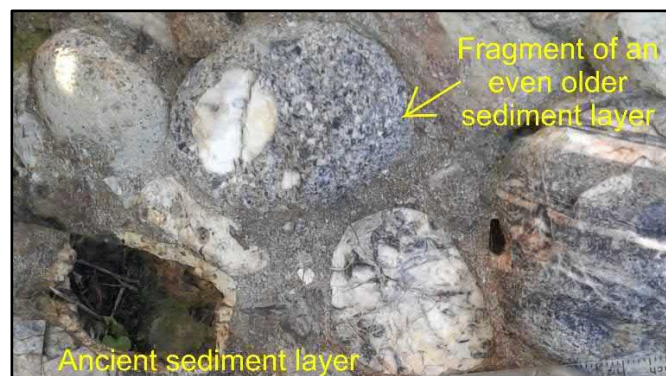
Challenge: The structures drawn below will be seen in sites of the R38 geotrail. As you visit different sites, see if you can spot more than one change in the same rock! Post your findings at <https://www.facebook.com/R38Geotrail/>.



2. How do geologists get information from rocks?

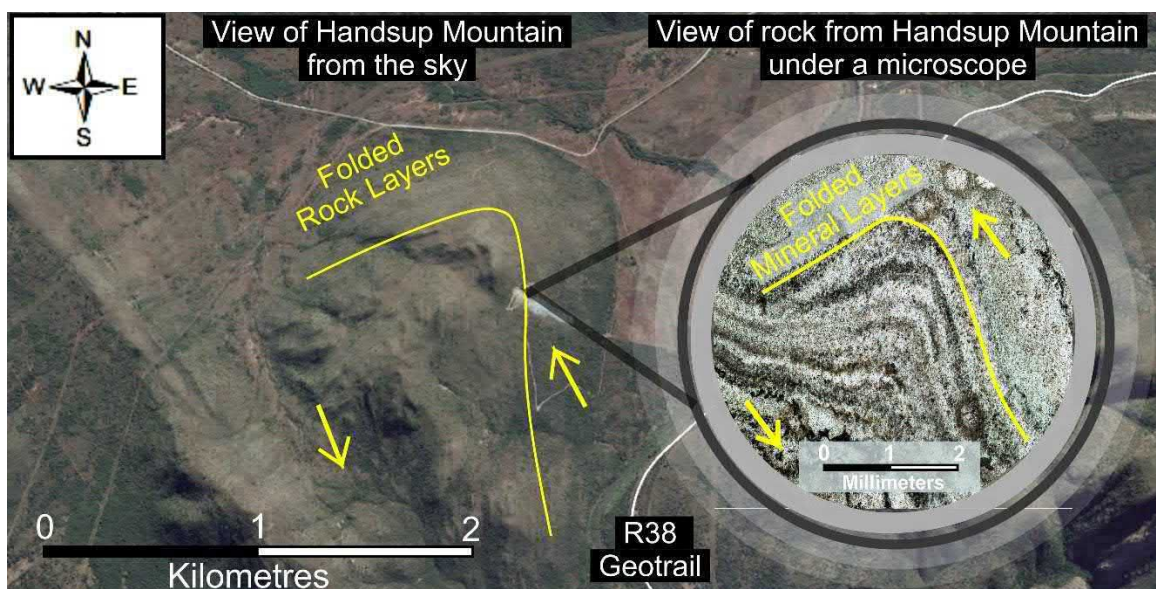
A. By noting rock composition, texture and relationships with other rocks.

Useful information comes from noting the composition and texture of rocks (0-1E). For example, ancient river gravel transformed into sediment rock layers-by burial and related changes-may contain fragments that tell us a lot. Rounded rock fragments may indicate far source regions while angular fragments may indicate a nearby source region. Ancient gravel beds may contain fragments of even older gravel beds. This tells us that weathering and erosion were shaping the surface of the early Earth in a somewhat similar way to the present day.

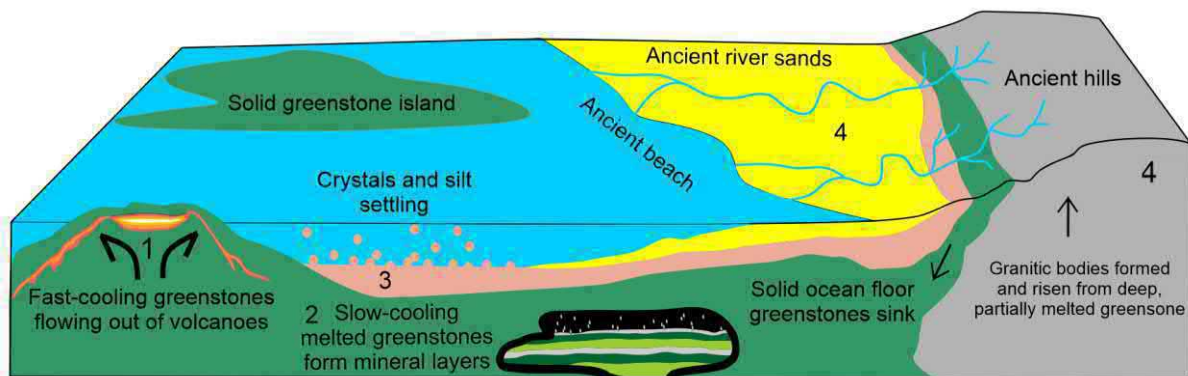


B. By studying rocks at different scales.

When preserved, large rock shapes and patterns seen in the environment-for example as mountains-can also be present in smaller rocks from the same mountain (0-1F). If the same shapes and patterns are seen in both large-scale and micro-scale images of rocks, a geologist becomes more confident in their understanding of structures within the area of interest.



Formation of rocks in the Ancient Barberton Region (0-0A)



1. **Magnesium-rich volcanic rocks:** Extremely hot lava (melted greenstones) flowed from deep in the Earth through deep cracks on the Earth's solid outer layer. Once on the water-covered surface of the early Earth, the lava cooled underwater to form solid greenstones.
2. **Magnesium-rich intrusive rocks:** Where the melted greenstones found no cracks leading up to the surface of the Earth, they pooled and cooled slowly. The slow cooling allowed more complete formation of crystals, with heavier crystals forming first and settling into a layer. Later, lighter crystals formed and settled into another layer together with fewer heavier crystals. In this way, the melted greenstones that cooled underground turned into a stack of rock layers made up of different crystals. Many economic minerals occur in such layered rocks.
3. **Sea-floor rocks:** Deep, quiet waters in vast ancient seas of the early Earth contained dissolved rock-forming minerals. When changes in either temperature, pressure, pH and caused dissolved minerals to settle out of the seawater to form layers on the seafloor. Other layers on ancient sea floors were those of very fine silt which settled out of deep seawater.
4. **Granites and river system rocks:** When greenstones thicken, they partially melt to form lava poor in magnesium but rich in sodium and silica. These elements combine to form minerals less heavy than those in greenstones. When the lava made up of lighter minerals pools and cools underground, solid, lighter bodies of unsinkable density become stable rock. These spread around, hold up and glue together heavy, sinking rock layers mainly made up of volcanic greenstone. Such granite-glued rock layers are building blocks of continents.

Once uplifted, and exposed to agents of weathering and erosion, mineral crystals in granites broke down to form sands grains. The sand grains were transported by rivers to low-lying areas where they settled into layers. Once buried, sand layers hardened into gravel rocks.

What we learn from rocks of the Barberton Region (0-0B)

Rocks of the Barberton region record Earth processes from one of the earliest segments in the timeline of the Earth (Figure 1.A). At the time they were formed, the planet could not support multicellular plants or life forms that depend on oxygen because the atmosphere was made up of toxic gases. Only local accumulations of oxygen were available in oceans. After continued oxygen production by single-cellular life forms and/or other means, oxygen became available in the atmosphere. This transformed the toxic composition of oceans and the atmosphere into compositions that support a wider variety of plants and life forms. Human beings only appeared on the Earth towards the end of the timeline. No human being was present to record processes of the early Earth. For that information, scientists come to Barberton to study the world's oldest, best-preserved rock layers.

Unlike ancient rocks elsewhere, original materials in most rock layers of the region only experienced slight changes to their composition. Because of low-temperatures and low pressures experienced, green minerals developed in the rocks. Other green minerals in the rocks are of a primary origin. They form from the cooling of high-temperature fluids that rise from the depths of the Earth to the surface or near-surface levels of the Earth. Collectively, these rock layers are referred to as the 'Barberton Greenstone Belt'. This describes the narrow strip that regional greenstones occur as within surrounding granites (Figure 1. B).

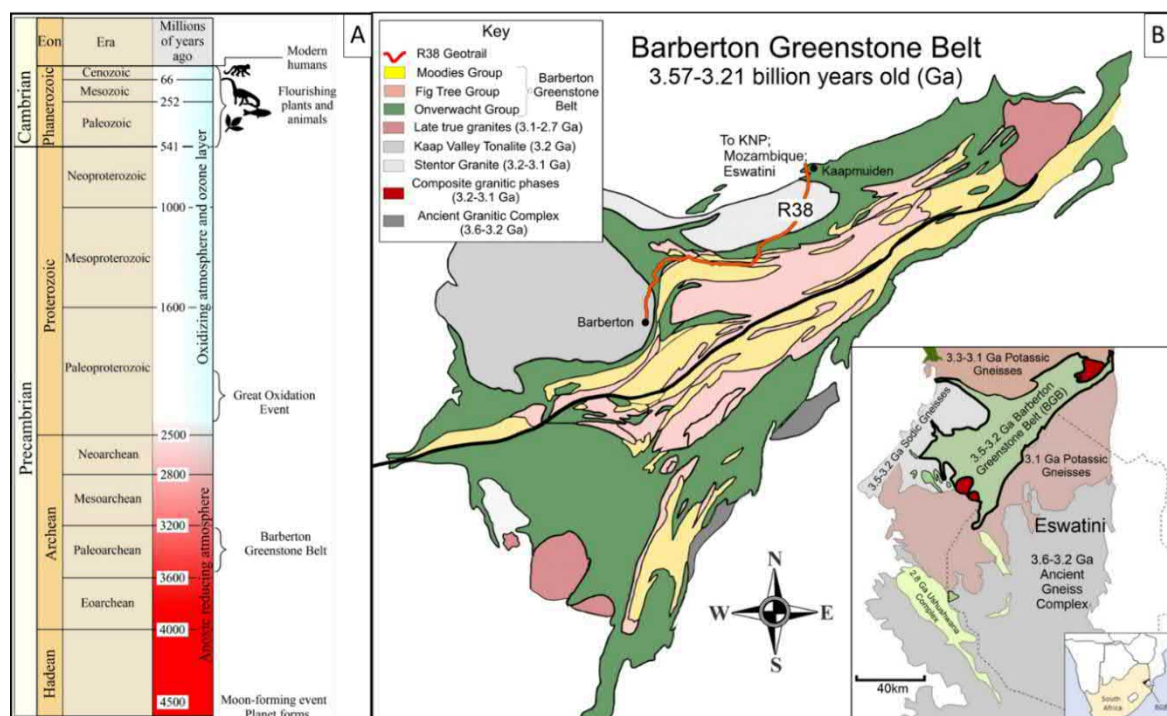


Figure 1. (A): Geological timeline showing the period in which rocks of the Barberton region were formed. (B): Ages of main rock types that contribute to the outstanding universal value of the region. Timeline adapted from (1AA); Maps adapted from (1AB) and (1AC)

The international value in rocks of the region comes from the following clues found in them (0-0B):

- ❖ Some greenstone layers and silica layers above them contain the world's earliest body fossils of bacteria visible under a microscope. These occur in greenstones formed under water and show that early life forms were present in ancient seas.
- ❖ Other greenstones display special textures which show that the early Earth was hotter, producing melted rock hotter than what we find flowing out of volcanoes today.
- ❖ The sea floor rocks contain gravel and volcanic rocks, as well as iron, silica and sulphur mineral layers. These show that deep, quiet seawater of the early Earth used to contain a higher amount of dissolved iron, silica and sulphur. From this, it was understood that the atmosphere of the early Earth was very different to the present atmosphere.
- ❖ Sea floor rocks that contain volcanic hailstones that tell of violent volcanic eruptions in the early history of the Earth.
- ❖ Granitic bodies of the region represent one of the earliest continent fragments.
- ❖ Layers of melted rock droplets that tell of massive meteorite impacts that melted rocks on the surface of the Earth, causing them to rise, cool, rain down and settle into layers.
- ❖ The river system rocks preserve the earliest mats of bacteria on sandstones. These show that early life forms were present in early rock layers outside of the ancient seas. These ancient sand layers also preserve raindrop impressions, ripples, mud cracks and signs of tidal currents. These suggest that some present-day processes also took place on surfaces of the early Earth.

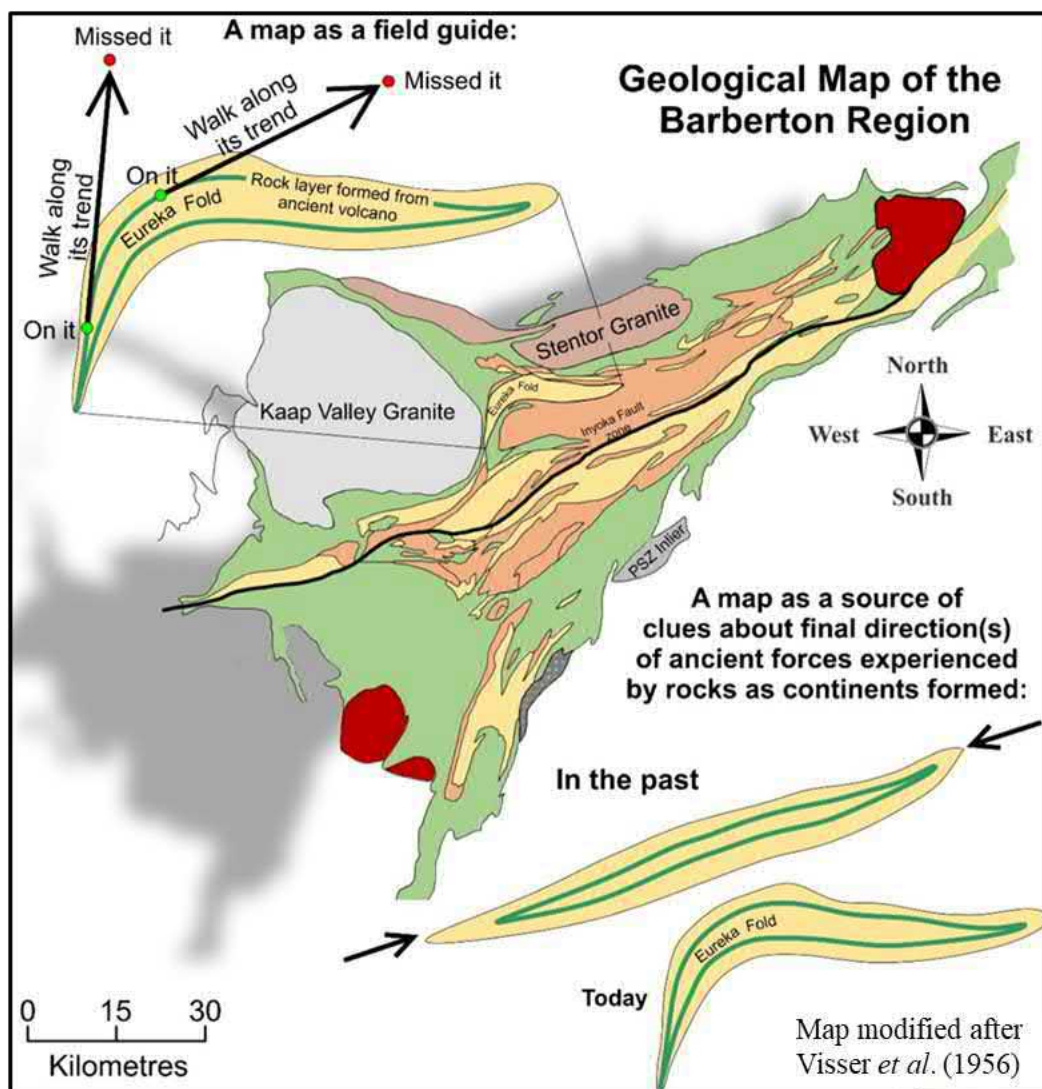


Please note: Best-preserved examples of rocks described above are mainly found along the R40 geotrail and in the Barberton-Makhonjwa Mountains World Heritage Site. Along the R38 are examples of physical and chemical changes that have taken place in rocks of the region.

What are geological maps?

Geological maps are drawings that show the position of rock types in an area. Like maps used to direct hikers to scenic viewpoints in a nature reserve, geological maps help geologists find important rock layers in a wide area or region. Finding certain rock layers is not always easy because rocks do not always extend in one direction over a long distance. Rock layers are often folded and/or broken into segments that have moved past each other.

To make it easier for others to find certain rock layers in an area, geologists draw shapes of rock layers and code them with colours or patterns that represent certain rock types. Symbols are added to show the steepness, trend (direction in which rock layers extend), texture of rock layers and many other features. Such drawings make it easier to see if there are patterns formed by rocks of a region. From studying rock layer patterns, geologists can work out a likely direction of ancient push and/or pull forces that helped stitch together pieces of early stable land masses that stuck together to form continents of this planet (0-0C).



R38 Geotrail Sites

The R38 geotrail road winds along the northern edge of the Barberton Greenstone Belt. It cuts across two granitic bodies, various greenstones and two sequences of ancient sediment layers.

As a developing geotrail, this section of the R38 between Barberton and Kaapmuiden contains no information panels, developed parking, bins, shops or public bathrooms. Visitors should note that Site 2 is located outside Diggers' Retreat-a restaurant and hotel where bathrooms can be used after buying refreshments. Parking space at roadside sites of the busy R38 (3, 5, 6A, 7, 8) is only available on the side of the road closest to the Kaap River. Future development of the R38 should include the development of parking space where the roadcut surfaces are at present. This will also expose less weathered surfaces of rocks described in this guidebook. While this has not happened, visitors who intend to see all geotrail sites are advised to start the trail in Barberton rather than Kaapmuiden (see Figure 2).

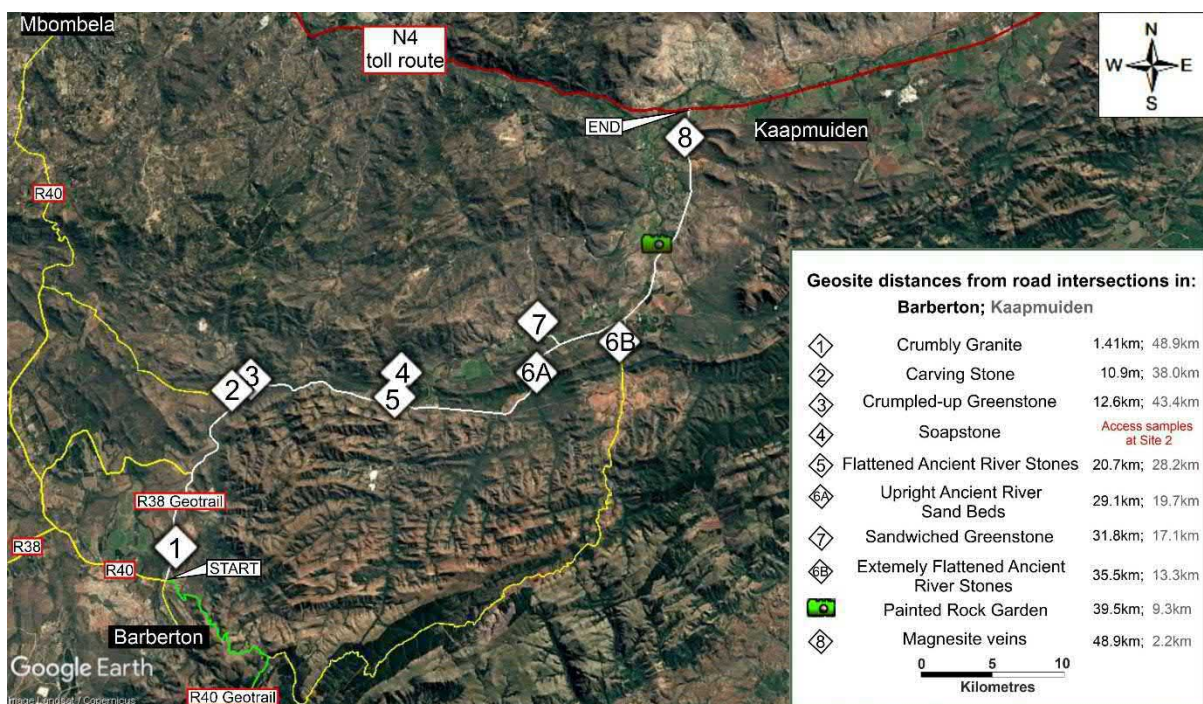


Figure 2. Sites to see along the R38 road.

Disclaimer

Please ensure your safety when visiting the R38 geotrail. The authors, landowners and all associated parties are not liable for damages or theft which visitors-or their property-might be exposed to along the R38 geotrail. Common threats at the different sites include humans, fast-moving cars, cows on the road, insects, and reptiles. Reminders of these threats are included in site descriptions of higher risk sites.

Upcoming Stop: Site 1 at the Kaap Valley quarry/sand pit

Right turn at: 25°45'0.28"S; 31° 3'12.07"E



Directions: Once at the R40-R38 four-way stop, proceed northwards (north arrow in bottom right corner) until you reach the right turn to site 1. (A): Google Earth image of a section of the R38 leading to the next site. (B): View of the site.

Did you know?

Early Dutch settlers likened the Barberton region to the Cape-Kaap in Dutch. Because of this, some towns and features of the region have names with “Kaap” in them(0-1A). These include:

- ❖ **De Kaap Valley** (The Cape Valley): Valley in which the town of Barberton is situated.
- ❖ **Noordkaap** (North Cape River): Train station and settlement about 12 km northeast of Barberton, named after the northern tributary of the Kaap River.
- ❖ **Kaap River** (Cape River): River that runs along the R38 geotrail. In this region, the Kaap river was the precolonial divide between Eswatini territory in the south and Sotho territory in the north⁰⁻¹. The discovery of a gold nugget where the Suidkaap (South Cape River) and Noordkaap Rivers join to form the Kaap River started the Barberton Goldrush.
- ❖ **Kaapmuiden**: Cape Mouth-where the Kaap River flows into the Crocodile River.

Site 1: Crumbly Granite (Kaap Valley Tonalite)

Turn into quarry entrance: 25°45'0.28"S; 31° 3'12.07"E

The municipal sand pit occurs in a type of granite called the Kaap Valley Tonalite (KVT) (1.1). Meter-scale domes of the KVT are surrounded by outer layers of crumbly granite fragments. These fragments are extracted for different construction and road maintenance uses.

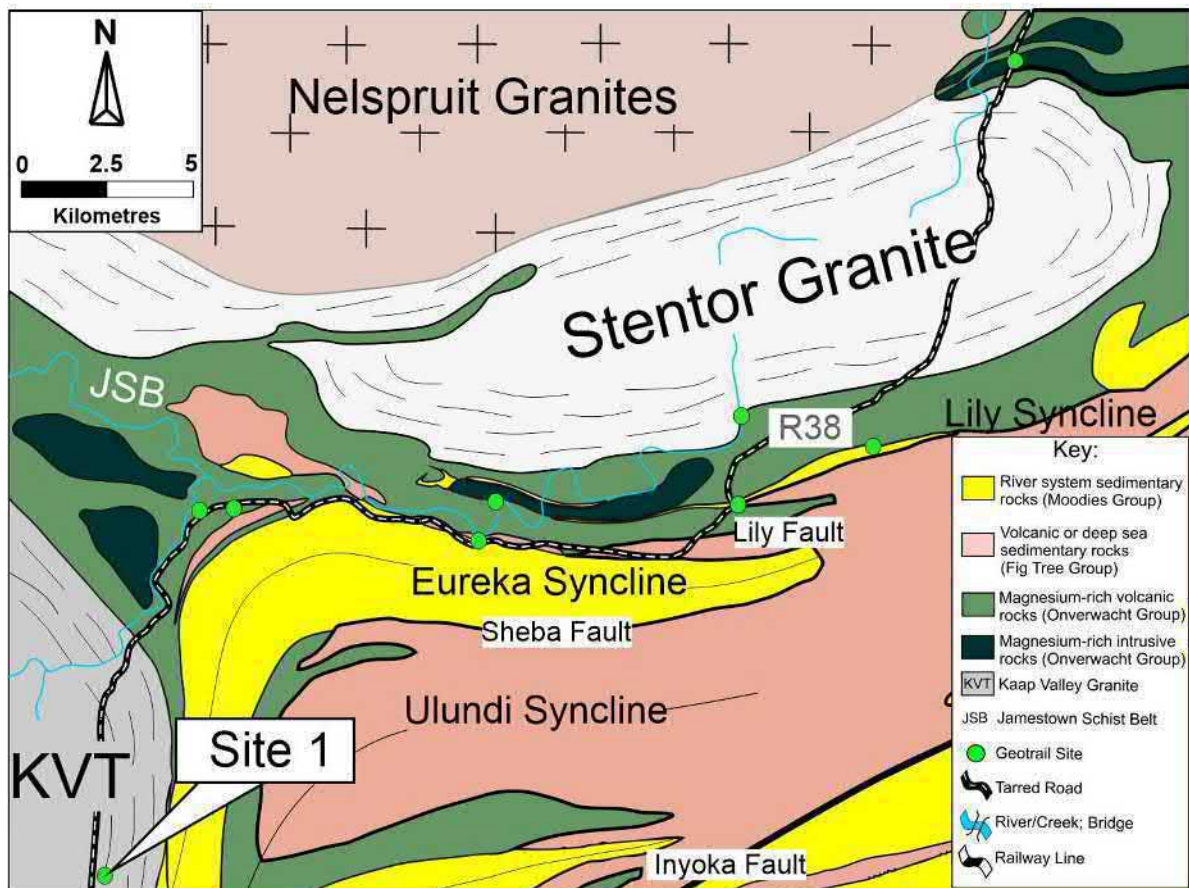


Figure 1.1. Simplified map showing the location of the Kaap Valley Tonalite (KVT) at the western edge of the Barberton Greenstone Belt.

Stabilizing early rock layers

Greenstones formed from heavy magnesium and iron minerals are the earliest and thickest rock layers found in the region (1.2). They piled up due to the addition of new volcanic and intrusive greenstone layers that cooled from fluids that rose from deeper parts of the Earth (1.3). With cooling from exposure to sea water, greenstone layers made up of heavy magnesium and iron minerals became unstable on the surface of the early Earth because of their heaviness (1.3). They sagged into deeper, warmer parts of the Earth where they began to melt in part (1.4). The product of that melting produced thick granitic fluids of a lower density (1.4). These fluids rose towards the surface of the Earth and either cooled slowly underground or flowed through deep

fractures to reach the surface of the Earth through volcanoes (1.5.). Once partially solid, the granitic fluids that cooled underground became granitic bodies of lower density than the heavy and cool greenstone layers at the surface of the early Earth (1.3). This caused the granite bodies to rise towards the surface of the Earth and the heavier greenstone layers to sink into the granitic body rising beneath them (1.3). As a result, fragments of stable granite-greenstone landmasses were formed by sinking greenstone layers that are supported from below by unsinkable granitic bodies (1.3). Due to binding processes of the Earth that may include magma (1.6) or horizontal pushing forces (1.7), isolated fragments of stable land masses were joined together to form larger stable continents.

Aside from physically supporting greenstones from beneath, granites prevent the recycling of surface rocks back into warmer, fluid parts of the Earth. This is understood to result from the removal of heat-producing elements in the warmer depths of the Earth as they flow into granitic melts (1.8). By rising towards the surface to form solid granitic bodies, the heat-producing elements are no longer available to add to the heat-production of deeper parts of the Earth that trigger recycling of the Earth's surface (1.9). As a result, the deeper parts of the Earth become more stable as granitic melts form and flow upwards (1.8).

Granitic bodies of the Barberton region

Granitic bodies of the early Earth are usually made up of coarse white, clear and dark mineral crystals. They form underground from a hot, thick fluid of melted rock that cools slowly enough for well-developed crystals to form (1.1). The forceful rising of early granitic bodies of the Earth caused flattening and stretching of rocks that were on the surface of the Earth. This produced an alignment of minerals within the rising granites and in the invaded greenstone layers (1.3).

Once weathering and erosion have removed rock layers above a cooled and solid body of granite, granite bodies become exposed as rocks on the surface of the Earth. This is the history of the earliest granites that form a large part of our continents.

Exposed near popular roads within the Lowveld region are different types of granite formed through melting of different sources and depths. Scenic routes close to granite exposures near the South African part of the Barberton Greenstone Belt include the N4 national route between Kaapmuiden and Nelspruit, the R40 secondary road between Nelspruit and Barberton and the R541 between Badplaas and Oshoek.

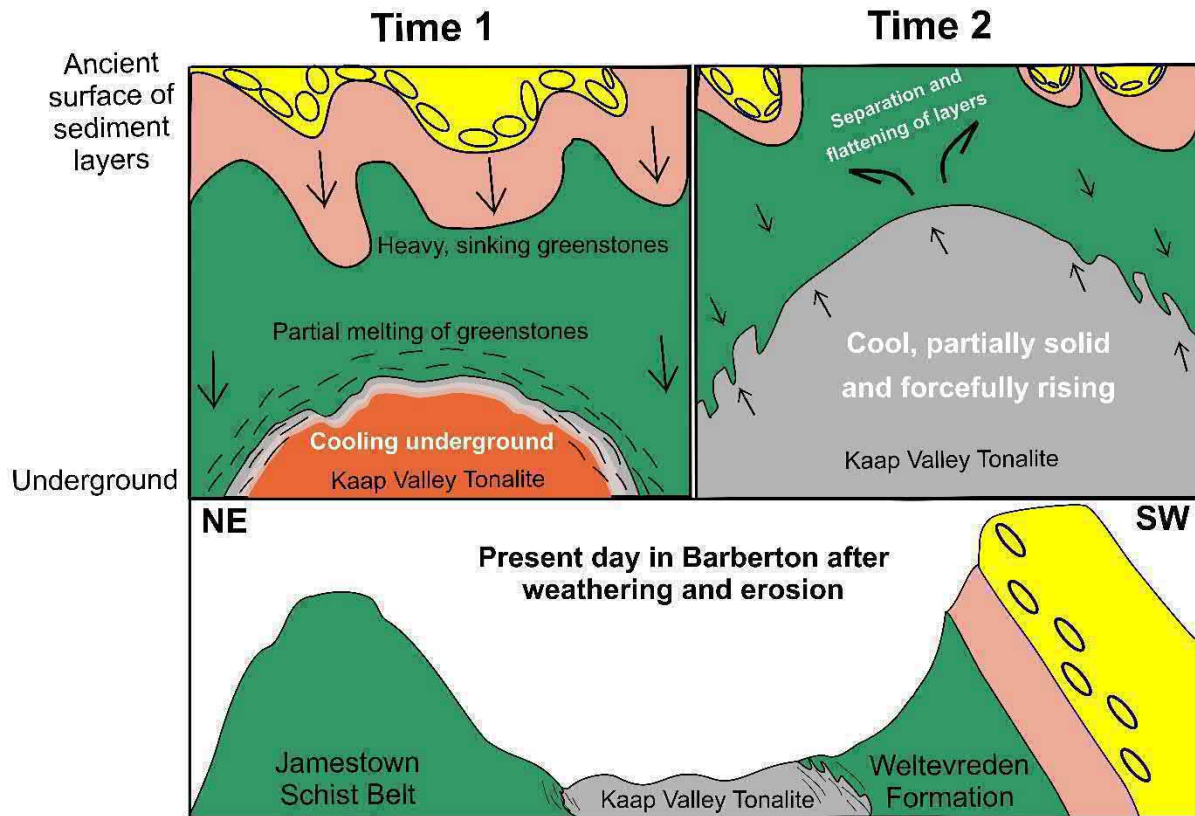


Figure 1.2. Drawing of how the Kaap Valley Tonalite (granitic body) rose to separate greenstones found at the edge of the granite just outside of the town of Barberton. Diagrams are not drawn to scale.

Features of the Kaap Valley Tonalite at the Barberton sand quarry

Many pieces of greenstone are found in the Kaap Valley Tonalite (KVT). Some greenstone pieces are very dark, with sharp edges, and contain cracks filled by the tonalite (Figure 1.3 A). They may be aligned with weak mineral layers of the Kaap Valley Tonalite. This relationship of greenstones and granite tells us that the greenstone pieces must be older than the granite. They were solid fragments that flowed and lined up with minerals of the Kaap Valley Tonalite as it forcefully rose towards the surface of the Earth (1.1.).

Other greenstones in the Kaap Valley Tonalite are oval-shaped with margins that blend into the tonalite (Figure 1.3 B). These may have formed together with the tonalite at deeper parts of the Earth where greenstones can exist as fluids (1.1.).

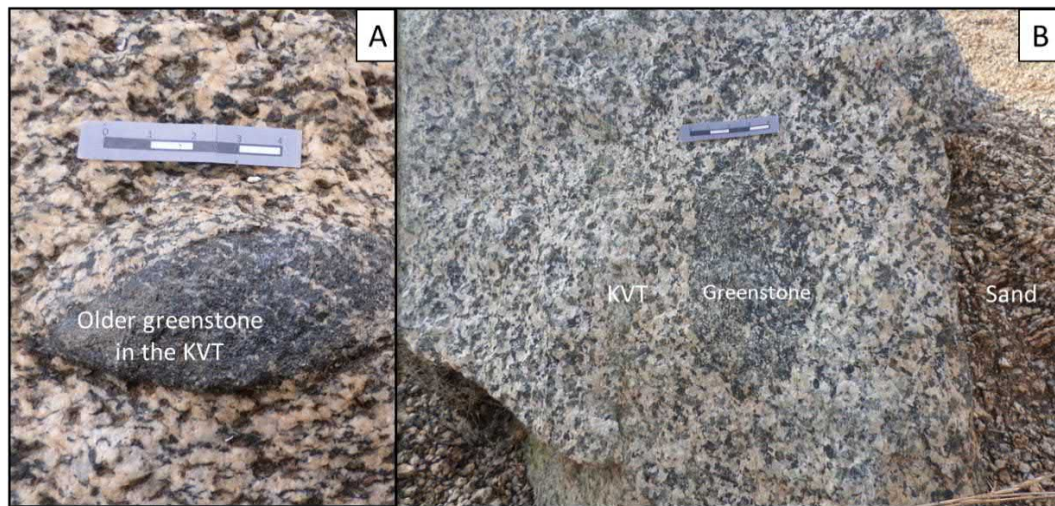


Figure 1.3. Features to see at Site 1. (A): Angular greenstone fragment invaded by fluids of the Kaap Valley Tonalite (KVT). (B): Oval-shaped greenstone in the KVT may have formed together with the KVT through mixing of different fluids. Note the sand grains at weathered edges of the solid core of the KVT. These sand grains may be carried by rainwater into nearby rivers where they will settle to form sand layers that will become sedimentary rock layers of the future.

How can we tell which rocks are older than others in the field?

Once the Kaap Valley Tonalite was solid, deep cracks in it were filled with melted greenstones from deeper in the Earth (1.1). These greenstones are younger than the granite because they contain pieces of the granite and they cut through mineral layers of the granite (Figure 1.3. b).

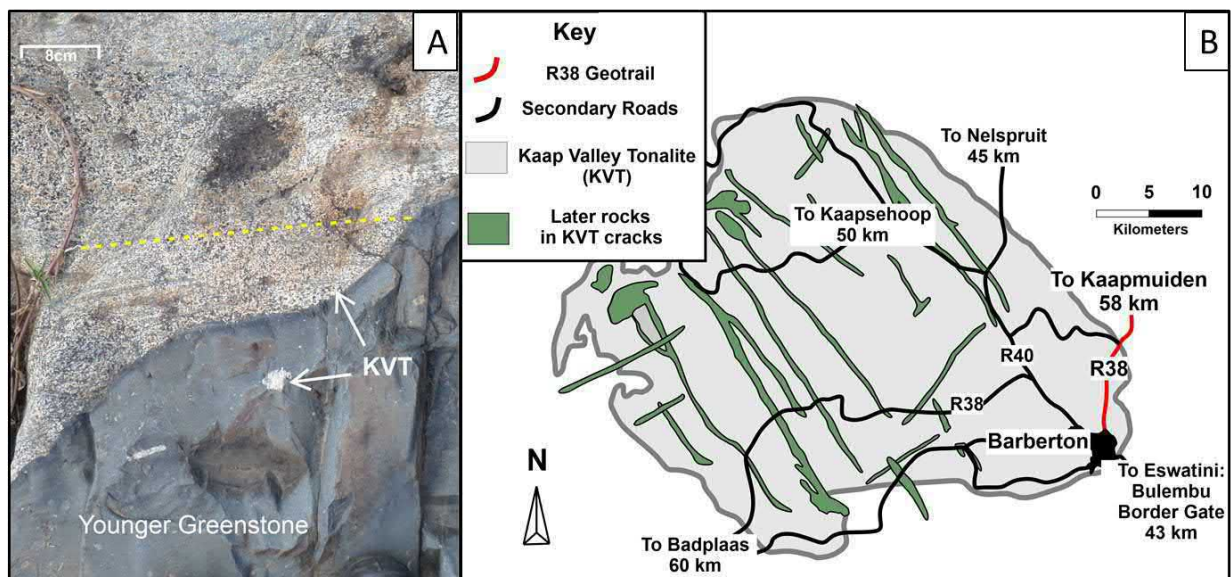


Figure 1.4. Rock relationships of the Kaap Valley Tonalite (KVT) and younger greenstones that form high ridges within the low-lying granitic body. (A): Pieces of the KVT in a younger, invading body of greenstone tells us that the KVT must be older than the younger greenstone. Notice how the younger greenstone cuts through mineral layers of the KVT (yellow dashed line). (B): Series of later greenstones that filled deep cracks in the Kaap Valley Tonalite form narrow ridges between Nelspruit and Barberton.

Barbertonians live on one of the best-preserved, easily
accessed, oldest rocks in the whole world!
#WhatsInYourBackyard



Fig. 1.5. Rock exposures of the Kaap Valley Tonalite in public spaces of Barberton. Tell us what rock types occur in or beneath your backyard at <https://www.facebook.com/R38Geotrail/> #R38GeotrailChallenge1.1

Upcoming Stop: Site 2 at Digger's Retreat

Left turn at: 25°40'45.96"S; 31° 4'46.60"E; Stop at: (25°40'43.60"S; 31° 4'47.69"E)



Directions: Exit the quarry and turn right onto the R38. Site 2 occurs just before the second railway intersection along the R38. (A): Google Earth image of a section of the R38 leading to the next site. (B): View of the site.

Did you know?

Barbertonite ($\text{Mg}_6\text{Cr}_2(\text{OH})_{16}[\text{CO}_3]\cdot 4\text{H}_2\text{O}$), also known as Stichtite-2H, is a pink or purple-coloured mineral named after the town of Barberton (1-2A). It was discovered in 1941 within the Barberton region, occurring in intrusive rock layers rich in magnesium. This, and other colourful minerals formed in similar rocks get their vivid colours from the chromium element they contain.

Waste rock of the Handsup Mine will be an obvious feature on your left as you drive from Site 1 to site 2. This mine is one of many historical mines where colourful minerals of chromium were once extracted.



Culture treat: Umjindi is the vernacular name for the town of Barberton. Though narratives differ, some published sources describe the name as a reference to Swati regiments choosing to go no further than the Barberton area as they dreaded returning to Eswatini after losing a battle for Pedi territory in the 1860s (1-2B).

Site 2: Carving Greenstone at Diggers' Retreat (Verdite)

Site 2 is a roadside stop outside a hotel and restaurant called Diggers' Retreat. It is situated near the R38 junction with the road to Consort Mine. Rocks to be seen here are in a cocopan container filled with green verdite hosted in a grey-green host rock. Also in the container is a white-to-light green, slippery greenstone called talc or soapstone. Both verdite and talc were transported to Site 2 from the Handsup Verdite Mine and the Scotia Talc Mine after both mines closed in 1999 and 2019 respectively. A lot of parking space is available around the cocopan on either side of the road. A wider collection of labelled rocks from the Barberton Greenstone Belt are available inside Digger's Retreat. Visitors may use bathroom facilities there after purchasing refreshments as there are no fuel stations, restaurants, or bathrooms along the remaining ~40km of the R38 geotrail.

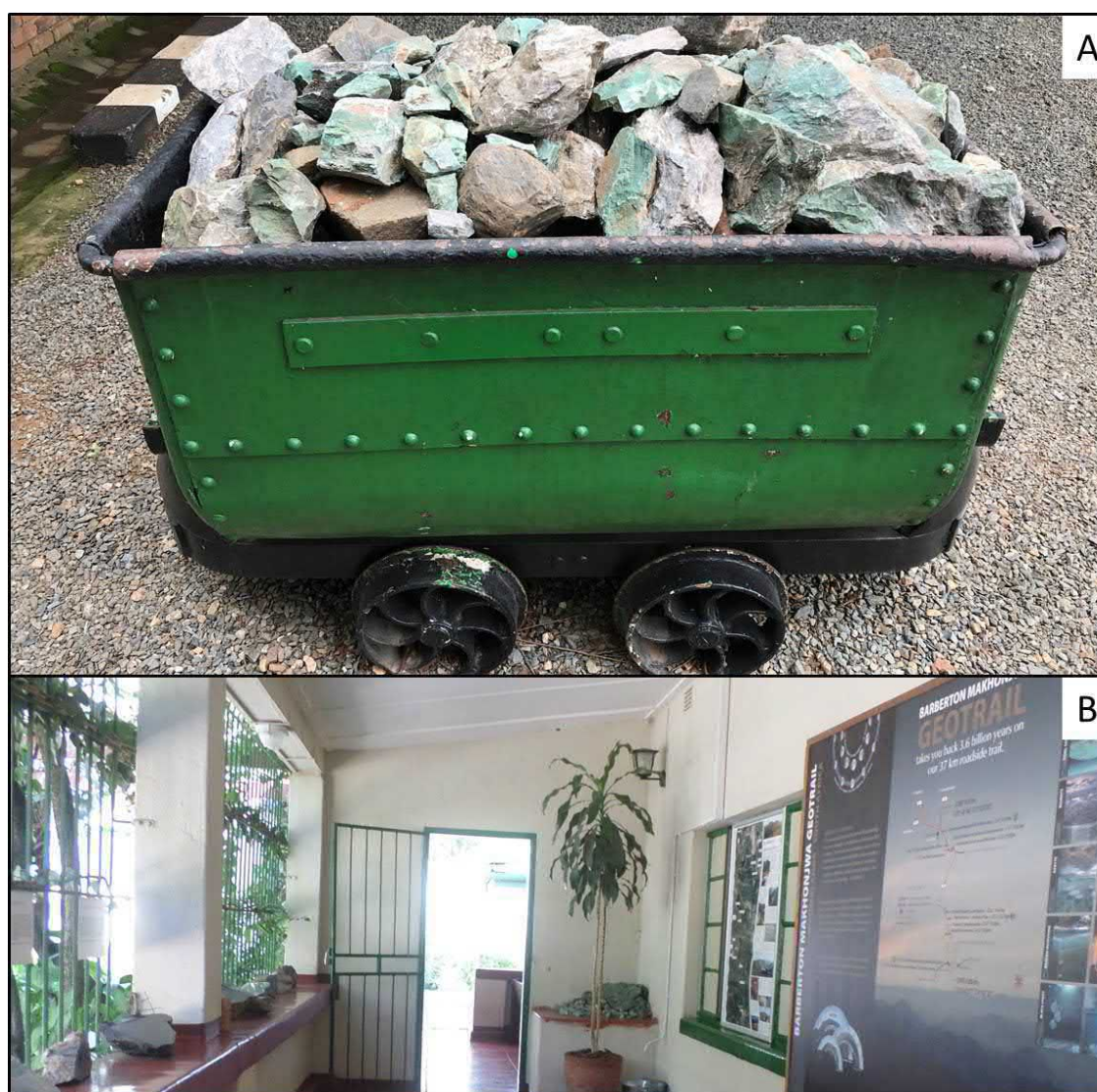


Fig. 2.1. A: Cocopan with verdite hosted in a grey-green host rock. B: Other rocks from the Barberton Greenstone Belt displayed in the Digger's Retreat reception area.

Rocks that contain verdite

Verdite is a semi-precious stone once mined from the Handsup quarry. It occurs where fluids made up of similar minerals as granite flow through greenstones rich in magnesium and chromium (Figure 2.2.). This invasion causes the growth of new minerals that form verdite (2.1.). Minerals in verdite are soft and are the reason why verdite is easily carved into ornaments. Zones where verdite has formed are commonly narrow and irregular (2.2). This makes verdite less attractive as a decorative stone for buildings, decreasing the number of uses for it (2.2). As a result, large-scale verdite mining in the Barberton region stopped around the late 1990s (2.3).

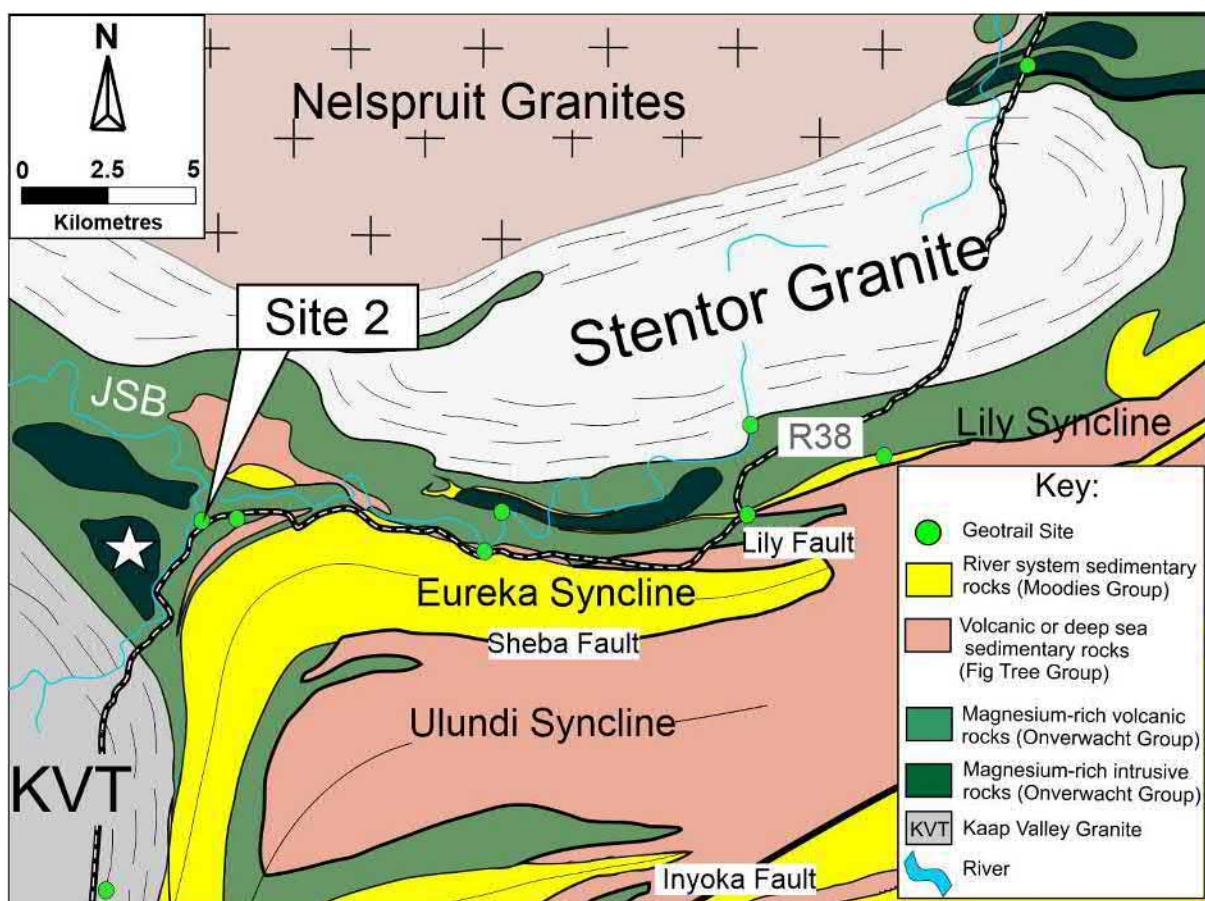


Fig. 2.2. A: Simplified geological map shows the location of Site 2 and the Handsup magnesium-rich intrusive rocks that host verdite (white star).

Verdite

High quality verdite is vividly green, soft and contains colourful layers of different minerals (Fig. 2.3. A). Though green verdite from Zimbabwe and the Limpopo Province of South Africa may contain red rubies (2.4), the only colours in verdite from the Barberton region are different shades of green, white, and yellow (2.1). The green layers contain a platy mineral called mica.

When mica absorbs a certain amount of the chromium element, it becomes a green mineral that makes up most of verdite (2.5). The other layers in verdite are made up of white sodium-rich minerals and yellow-red-brown titanium minerals (2.1).

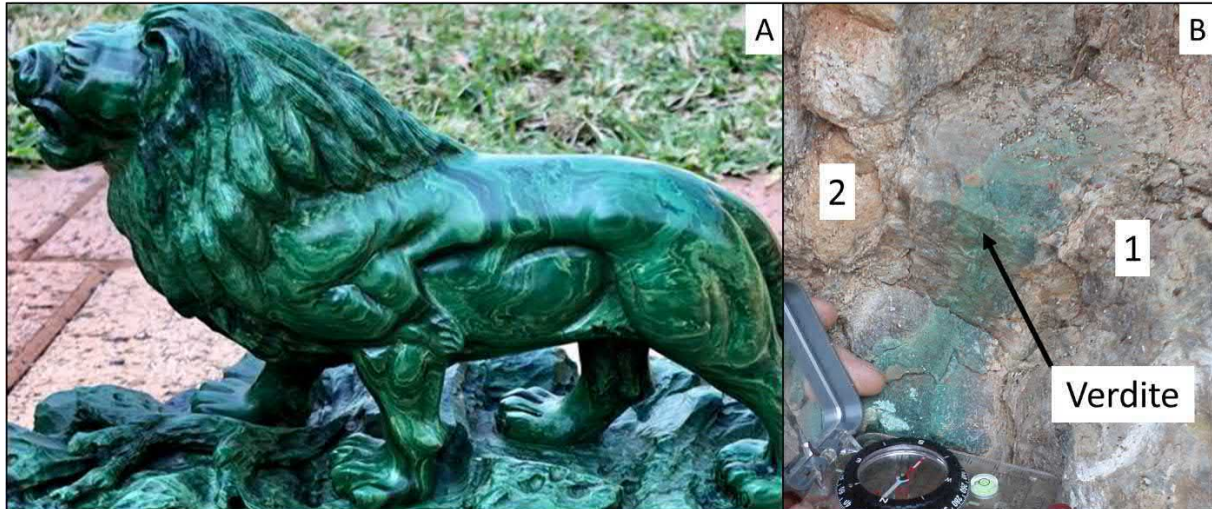


Figure 2.3. Verdite mineral layers (A): Close view of high-quality verdite statue reveals the green, yellow and white mineral layers of verdite. (B): Verdite which has not been moved from where it formed reveals the two main ingredients needed to form verdite (1 and 2).

Carving greenstone such as verdite is sometimes carved into statues of protected animals of the region.

#SaveTheRhino #SaveTheElephants #StopPoaching



Fig. 2.4. Verdite sculptures. (A): Mature African elephants reaching for marula fruit from a tall marula tree. (B): Rhino. (C): Leopard cubs with their mother. There is something unusual about the sculpture(s) above. Showcase your attention to detail by telling us what it is at <https://www.facebook.com/R38Geotrail/> #R38GeotrailChallenge2.1

Upcoming Stop: Site 3 south of Joe's Luck Siding

Stop at: 25°40'24.75"S; 31° 5'21.26"E



Directions: Head back onto the R38 and drive towards the Site 3. (A): Google Earth image of a section of the R38 leading to the next site. (B): View of the site.

Did you know?

During the Barberton Gold rush in the early 1880s, bars and hotels like Digger's Retreat were common along the old wagon route between Barberton and Kaapmuiden. They served a growing number of hopeful gold-seekers from all over the world. Legend says this route became known as "Leopard Crawl Road" because the beverages served caused miners to have to crawl to the next bar! (2-3A). The old route runs parallel to the R38 and is one of the routes used by Sir Percy FitzPatrick in a popular book he wrote that documents experiences he had in the Lowveld during the 1880s. The book is titled 'Jock of the Bushveld', which is a reference to his dog, Jock (2-3A).

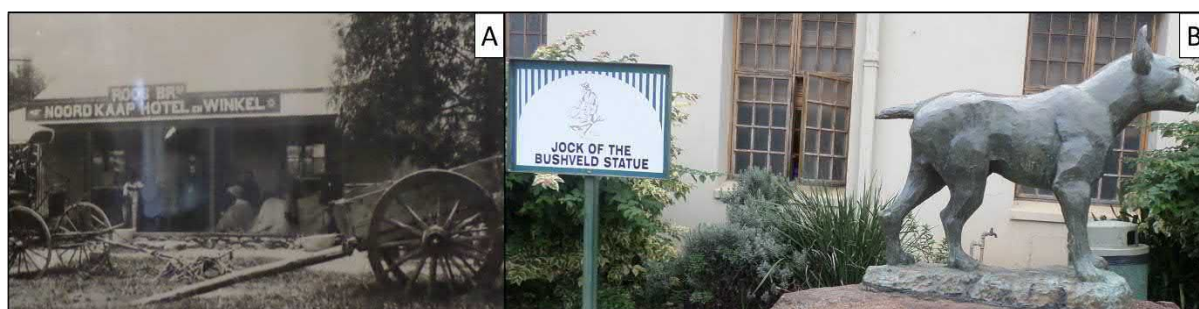


Figure 2-3B. (A): Early hotel once situated at present-day Digger's Retreat. Image source: Digger's Retreat. (B): Statue of Jock in front of municipal buildings in Barberton.

Site 3: Crumpled-up Greenstone (Crenulated Phyllite)

This is a roadside exposure of wavy greenstone along a busy road south of the junction of the Noordkaap and Suidkaap Rivers that form the Kaap River. Just beneath the northern edge of the road is a railway line that used to transport goods between Pretoria and Mozambique in the late 1890s (3.1.).

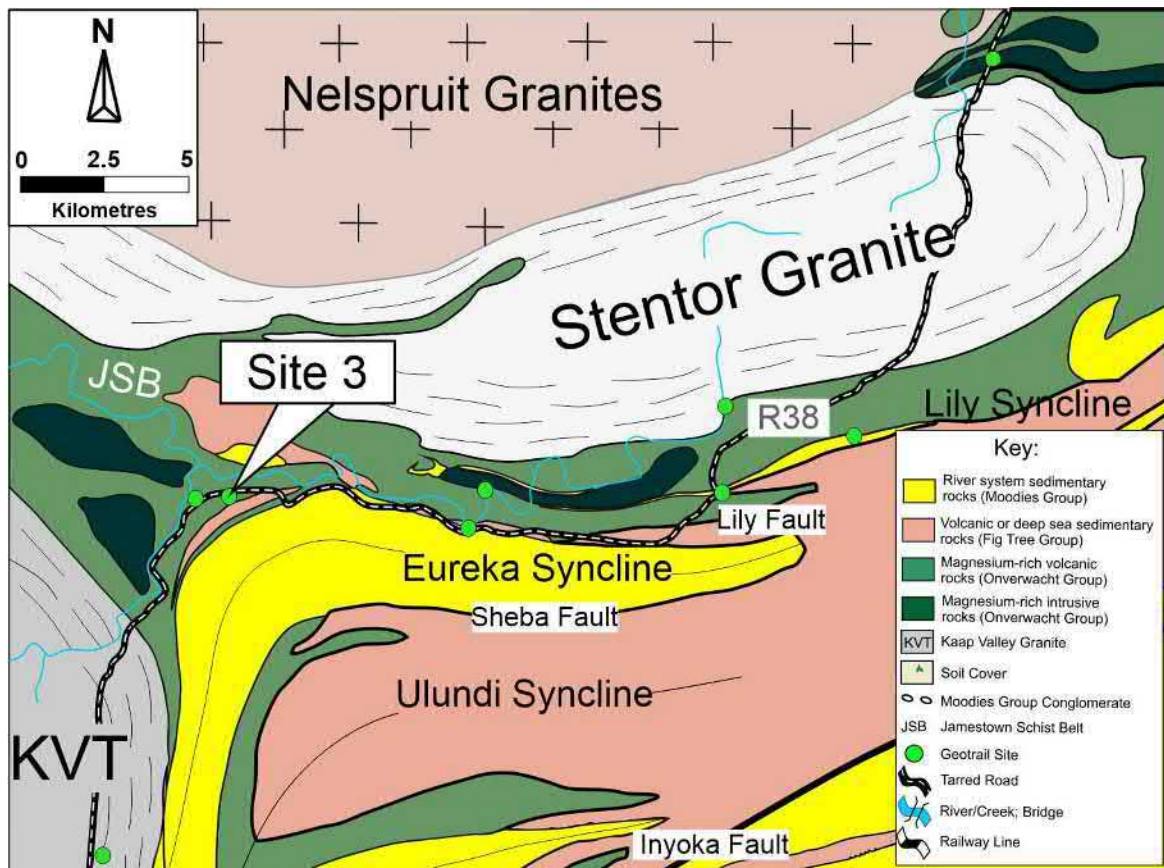


Fig. 3.1. (A): Simplified geological map of the R38 geotrail shows the locality of Site 3 in magnesium-rich volcanic rocks of the Onverwacht Group.

Crumpled-up Greenstone

The greenstone at Site 3 is completely made up of wavy surfaces. To crumple rocks into wavy surfaces, a series of changes took place (3.2). Time 1: Blocky grains or mineral crystals were laid down horizontally as magnesium-rich rocks of an ancient seafloor. Phase 2 to 3: The rocks were later flipped to vertical position and flattened by forces acting on the surface of the Earth. This flattening caused flat blocky minerals to change to flat minerals which are a more stable form as rocks become flatter and longer than before. Phase 4: Later vertical pushing forces acted at an angle to the rocks. This caused the newly formed flat minerals to crumple up into the wavy rock surfaces seen today.

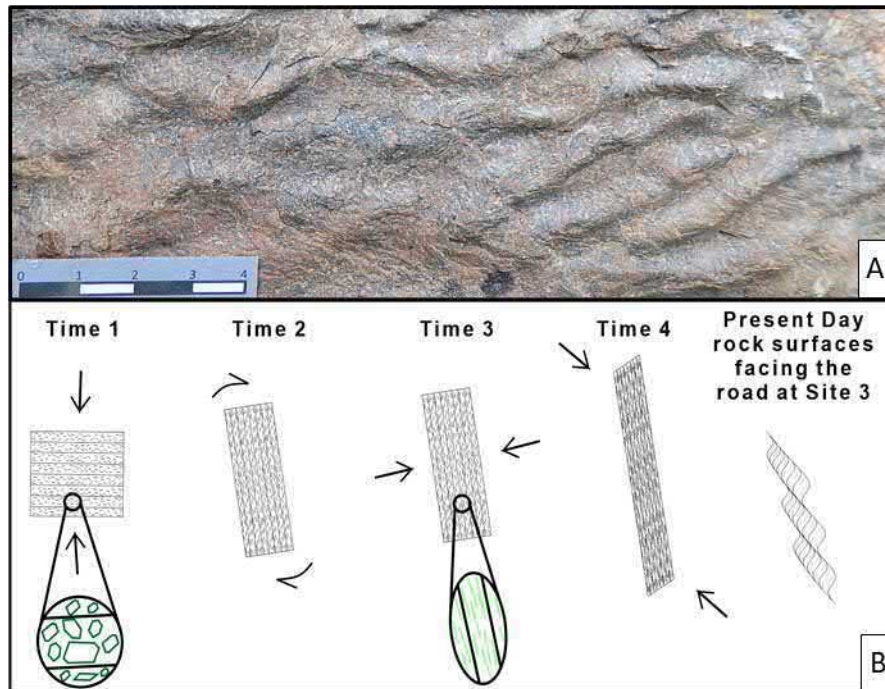


Figure 3.2. (A): Wavy surface at Site 3 is made up of layers of platy minerals that formed from blocky crystals which were flattened and folded during a series of changes shown in (B). Scale bar is 4cm long.

Not all rock layers at Site 3 are crumpled into wavy surfaces formed by small, horizontal folds (3.3). Other rock layers the rocks without the horizontal folds shown in Figure 3.2 were formed after the horizontal folds developed. These later rocks without horizontal folds were later folded to produce vertical folds. This tells us that horizontal pushing forces were part of the forces that shaped rock layers of the early Earth. Knowing the direction of ancient forces helped scientists form ideas about processes that led to the formation of stable land masses of the early Earth (3.4).

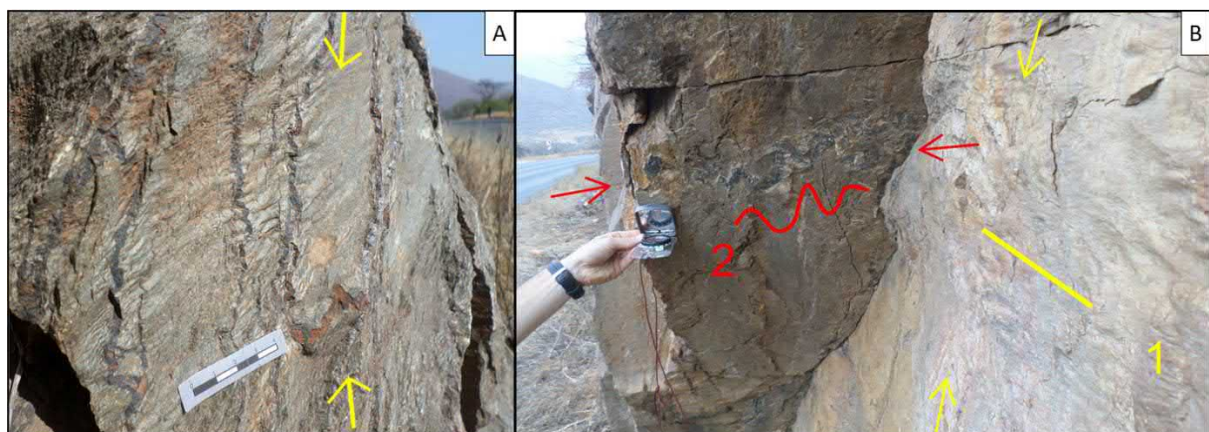


Figure 3.3. (A-B): Two directions of ancient pushing forces that acted on one of the earliest rock layers of the Earth (yellow and red arrows). Numbers refer to a local sequence of folding events. Scale bar is 4cm long.

Upcoming Stop: Site 5 at Ezzey's Pass

No Access at Site 4: Mine closed. Talc Samples placed at site 2.



Figure 3-4. Heaps of talc and waste rock outside the main entrance to the Scotia Talc Mine. The mine closed in early 2019 following labour disputes.

Did you know?

The Bon Accord layered rock hosts of talc from the Scotia Talc Mine are widely known for being the discovery site of Trevorite, a rare nickel-rich mineral typically found in meteorite rocks (3-4A). Other rare minerals were identified in the same rock group, including a recent one (2017) which was named Tredouxite after Professor Marian Tredoux who studied Bon Accord rocks for decades (3-4B).

Site 4: Soapstone (Talc-rich rock)

Due to the closure of Scotia Talc Mine, talc/soapstone will only be accessed at site 2.

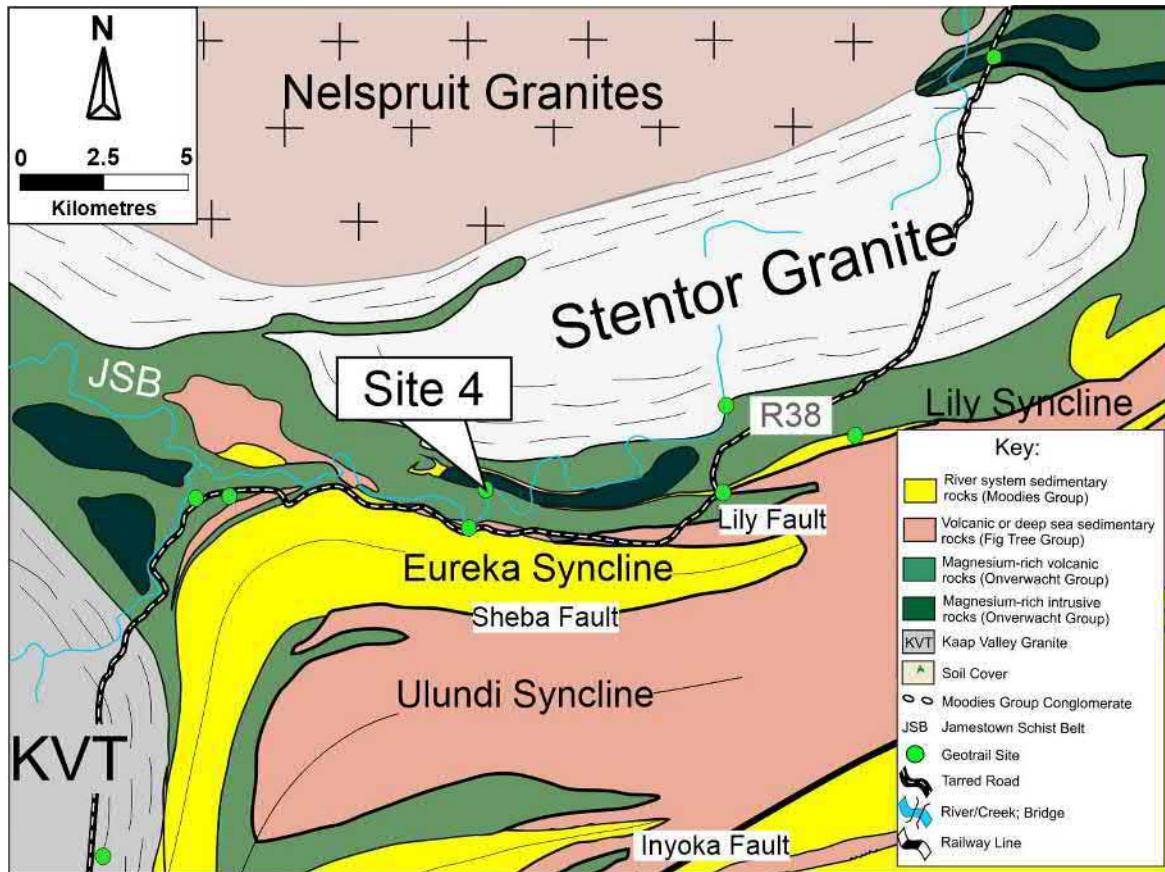


Fig.4.1. Simplified geological map the R38 geotrail shows the location of Site 4 in intrusive magnesium-rich rocks at within sediment layer of the Lily Syncline.

Soapstone

Soapstone is a slippery rock found in cracks and between mineral layers of intrusive magnesium-rich greenstones (4.1). It forms when fluids rich in silica introduce silica to magnesium-rich greenstones (4.1). One of the proposed sources of silica-rich fluids is the Stentor Granite (4.1).

Table 1. Quality of soapstone according to various features (4.2.)

Talc Quality:	Description
Best	Pure white colour, smooth texture, grit-free.
High	Tinted, smooth feel, grit-free.
Average	Off-colour, smooth feel, grit-free.
Low	White or coloured, contains grit.

Soapstone is one of the softest minerals in the world (4.3). On a scale of hardness from one (softest) to ten (hardest), talc occupies the first position because it can be scratched by human fingernails which are ranked slightly higher than 2 on the scale. Diamond occupies the tenth position as it can scratch many more minerals than common minerals can (4.3).

Rocks that contain soapstone

Pieces of soapstone displayed at Site 2 occur in greenstones that were first invaded by fluids rich in dissolved carbon dioxide (4.4). This first invasion formed white powdery carbonate minerals that are not slippery to the touch. Because carbonates and the magnesium-rich host rock of soapstone are of a higher hardness than one, their presence in soapstone may increase the hardness of samples. This lowers the value of soapstone because it has to be a fine powder when used in the production of products like paint and cosmetic powders (4.5).

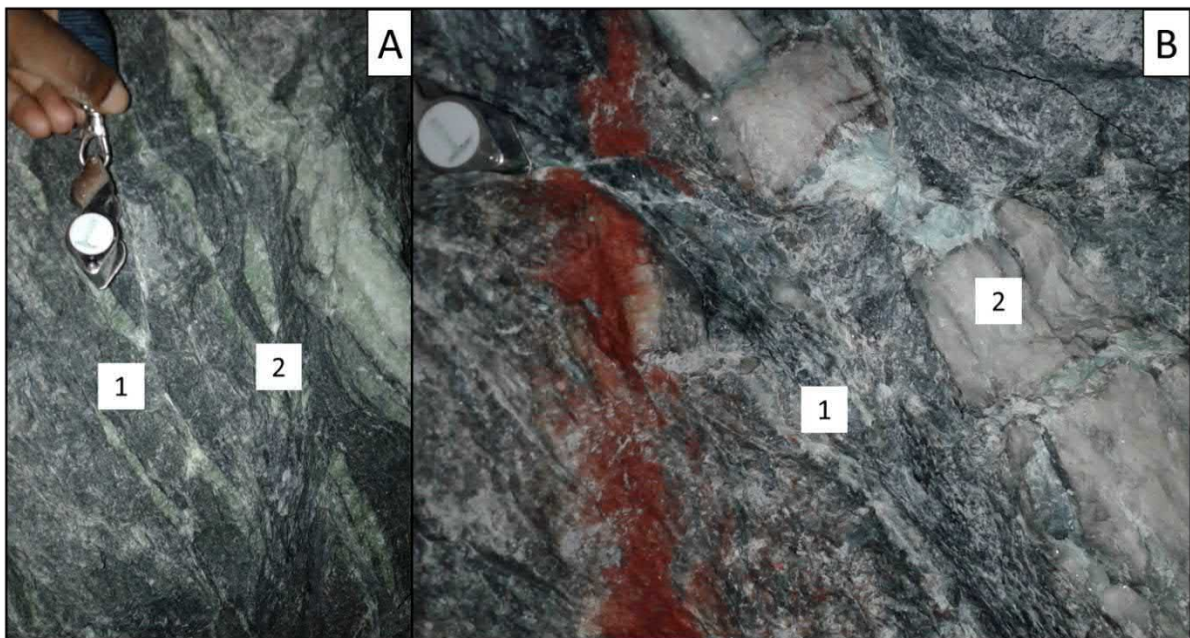
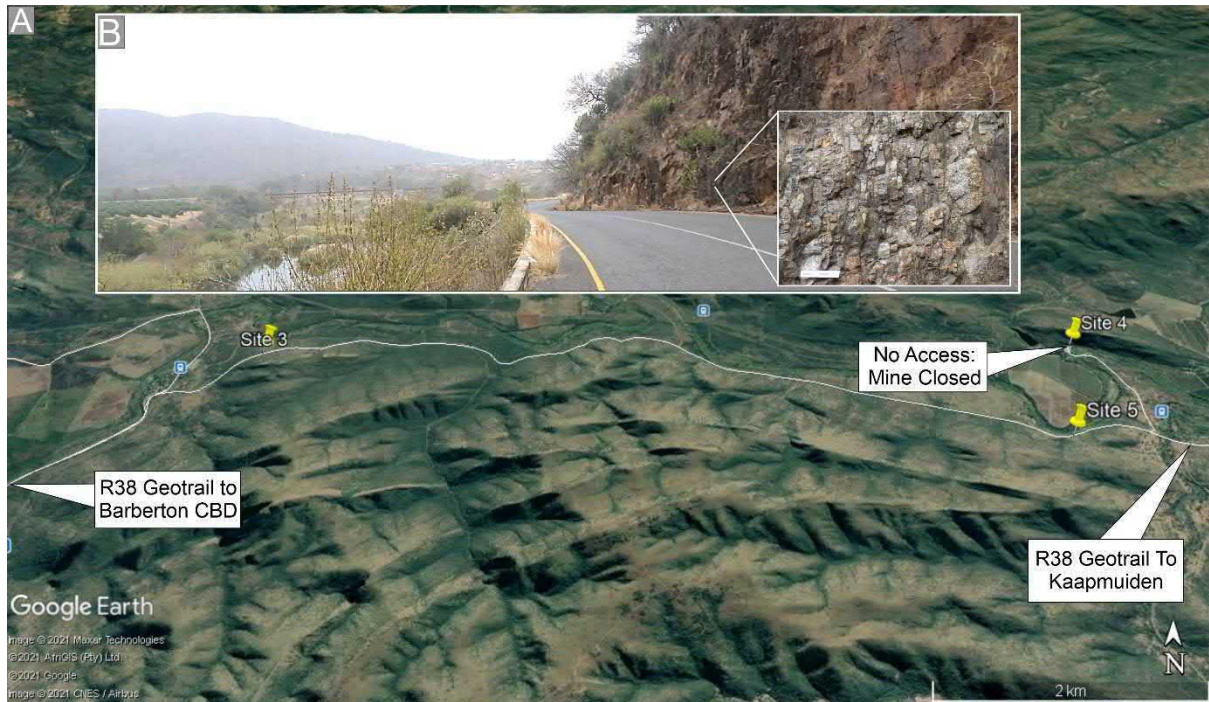


Figure 4.2. A: Original coarse greenstones (1) rich in magnesium first developed vertical cracks. The cracks were invaded by fluids rich in carbon dioxide and/or silica (2). This changed the coarse greenstones into weaker minerals called serpentine and a magnesium carbonate called magnesite. B: Rocks made of serpentine and magnesite were later changed to soapstone when fluids rich in silica flowed through later cracks in these rocks. The numbers in each image tell of relative ages (1= oldest; 2= youngest).

Upcoming Stop: Site 5 at Ezzey's Pass

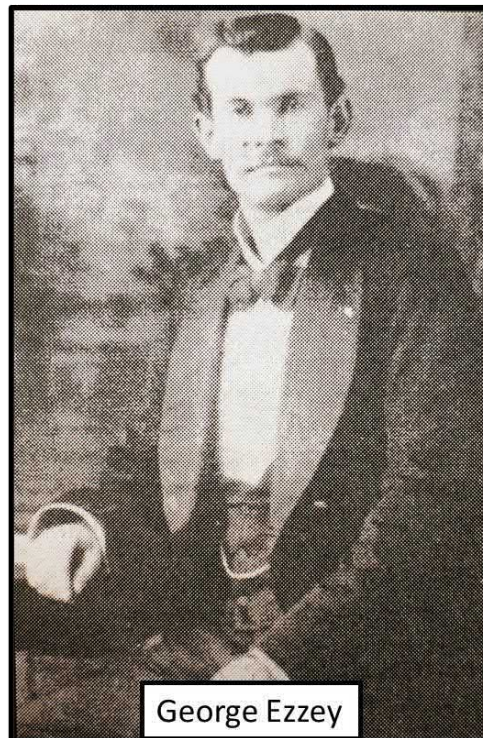
Stop at: 25°41'1.45"S, 31°10'3.09"E; Walk to: 25°41'4.68"S, 31° 9'56.25"E



Directions: Continue along the R38 towards Site 5. (A): Google Earth image of a section of the R38 leading to the next site. (B): View of the site. This stretch of road lies between two bends. Care should be taken when accessing or leaving the site.

Did you know?

The R38 section between Site 2 and Site 6 is locally known as Ezzey's Pass. It was a wagon route developed by George Ezzey to avoid crossing the Kaap River and railway line multiple times (3-5).



George Ezzey

Image source: Bornman (2017).

Site 5: Flattened ancient river gravel (Moodies Basal Conglomerate)

Parking: (25°41'1.45"S; 31°10'3.09"E), Outcrop: (25°41'4.68"S; 31° 9'56.25"E).

Sediment layers of the northern Barberton Greenstone Belt

Site 5 is a scenic roadside stop where the Kaap River winds closest to the road. It is a well-known stop among geologist interested in working out the amount of flattening and lengthening experienced by rocks in the northern Barberton Greenstone Belt (5.1).

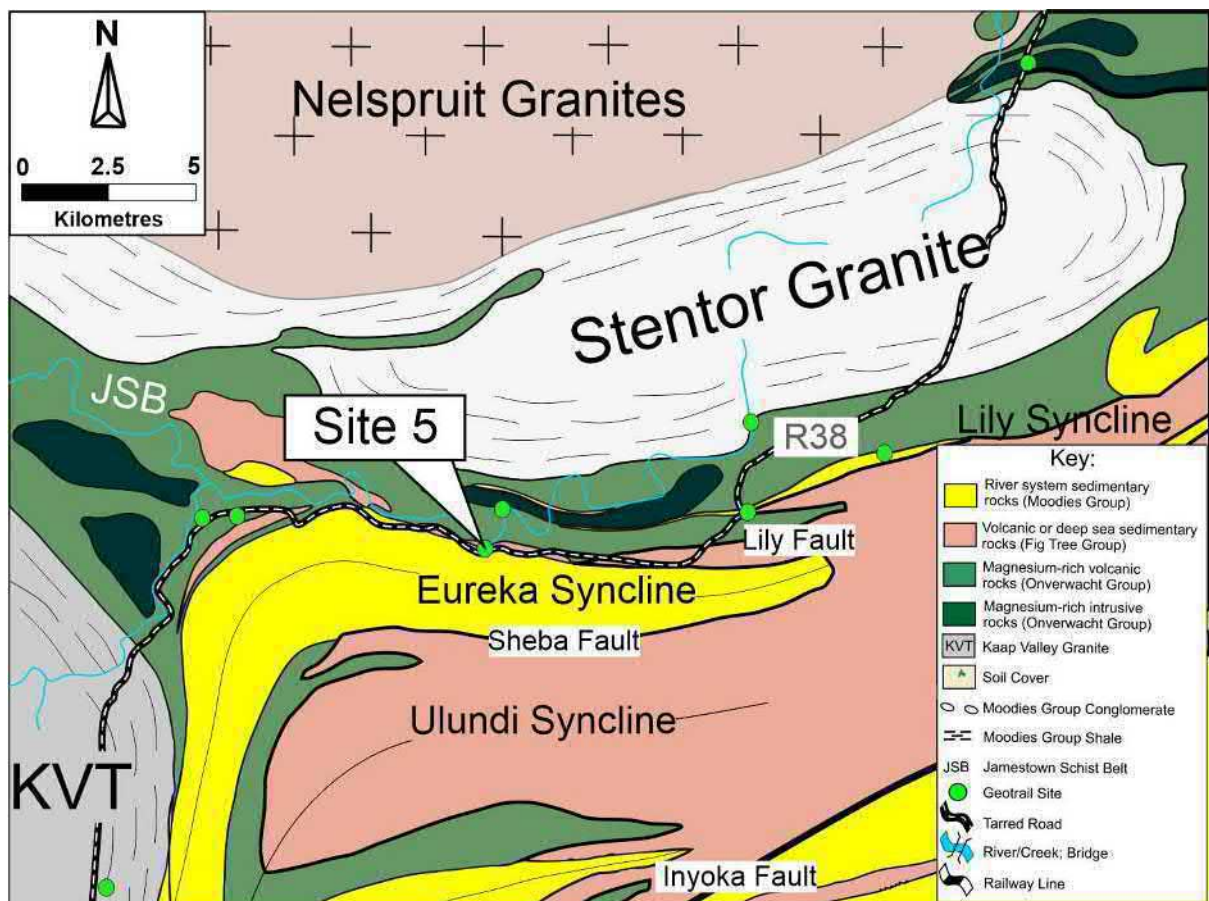


Fig.5.1. Simplified geological map the R38 geotrail shows the location of Site 5 in the northern edge of the Eureka Syncline.

Ancient River Sediments

Sandstones and pebble layers of ancient river systems are useful sources of information about rock-types that were present on the surface of the Earth when the sediment layers were formed (for example, 5.2). This is because river systems flow through landmasses, picking up different rock types along the course (5.3) in a similar way as the Kaap River is doing today.

Looking carefully at ancient pebbles from Site 5 teaches us a lot about the history of the Early Earth. For instance, the different rock types tell of an early Earth surface of diverse rock layers

formed by different processes-some of which take place today (for example 5.4). The presence of true granite pebbles tells us that part of the solid outer layer of the Earth had already started partially melting to form true granite components of future stable continents when the pebble layers were formed (5.5). The presence of flattened clasts aligned with planes of weakness along which sandstones have broken into tabular blocks indicates flattening of the area by horizontal pushing forces (5.6). The same layers of flattened pebbles exposed at Site 5 show no signs of flattening when studied further away from granitic bodies like the Stentor Granite of Kaap Valley Tonalite (5.1; see Figure 5.2). This suggests that the Stentor Granite may have played an important role in the flattening of the ancient pebble layers at Site 5 (5.1).

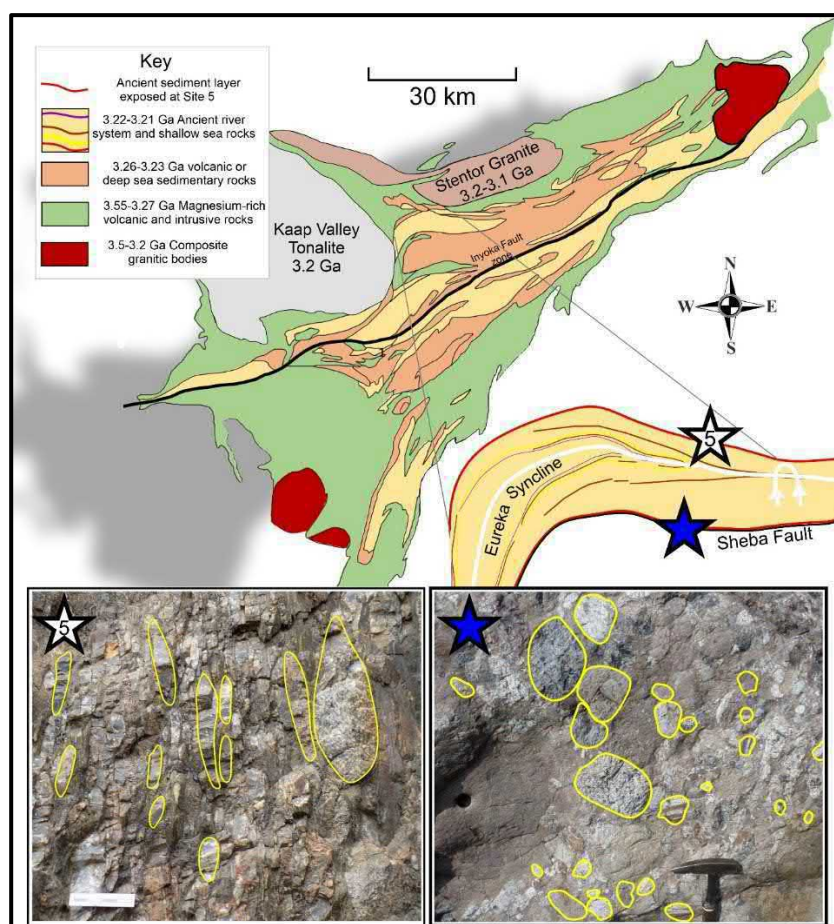


Figure 5.2. Difference in shape between rock fragments (sediments) of an ancient gravel bed. Sediments that occur closer to the Stentor Granite are flatter and generally shaped like lenses (white star). The same gravel bed only contains circular and oval-shaped sediments at a location that is further away from the granite. This may imply that the granite is the reason for the flattening of the gravel bed at Site 5.

The sandstone layers that occur on top of ancient pebble layers preserve clues about the flow direction of some ancient rivers (Figure 5.3). Since rivers flow from high-lying areas to low-lying areas (5.3), the flow direction of an ancient river tells of the general area where ancient

highlands and ancient valleys or coastlines occurred (5.3). This allows us to describe landscapes of the early Earth without having lived in it to observe and record what it looked like.

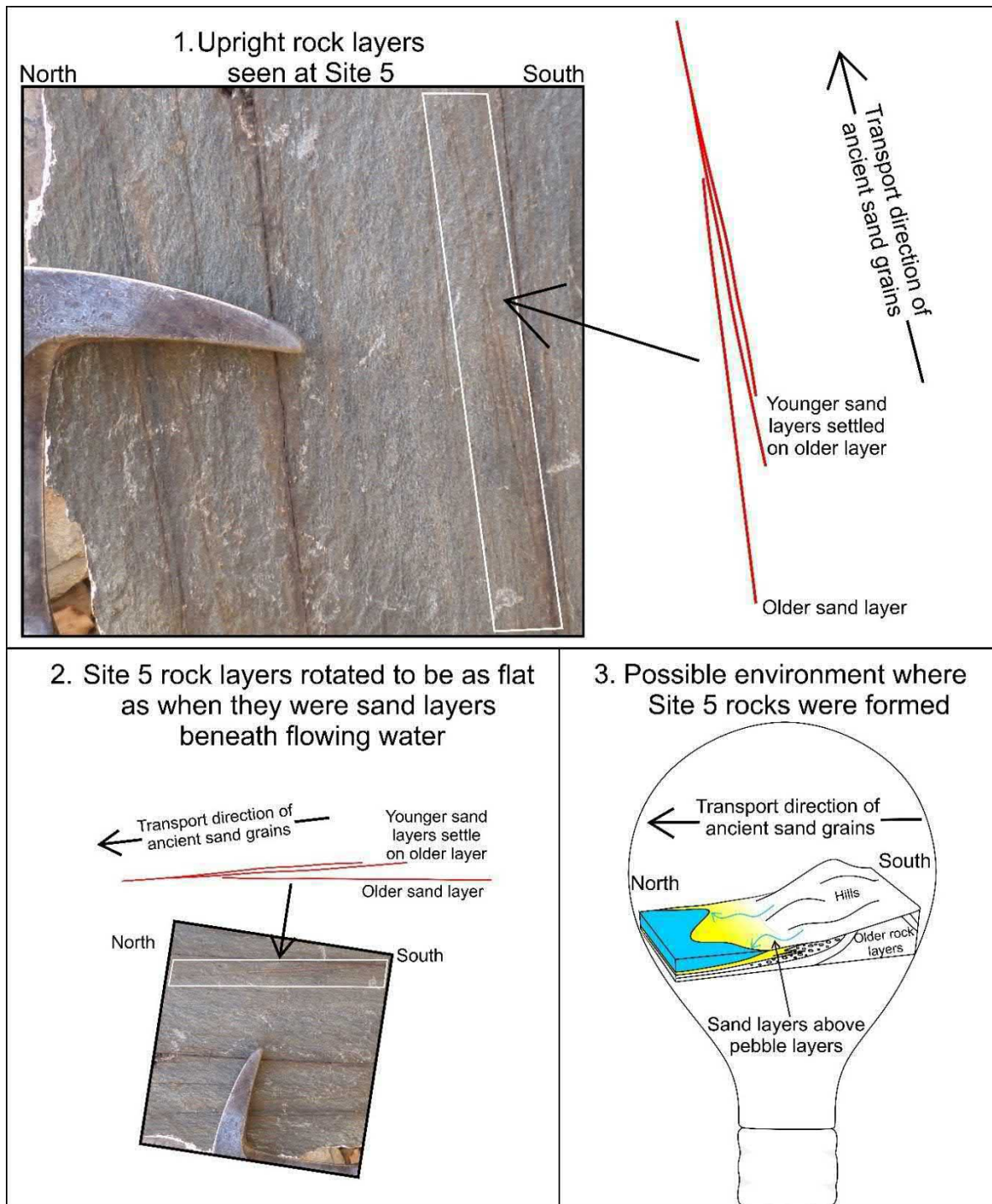
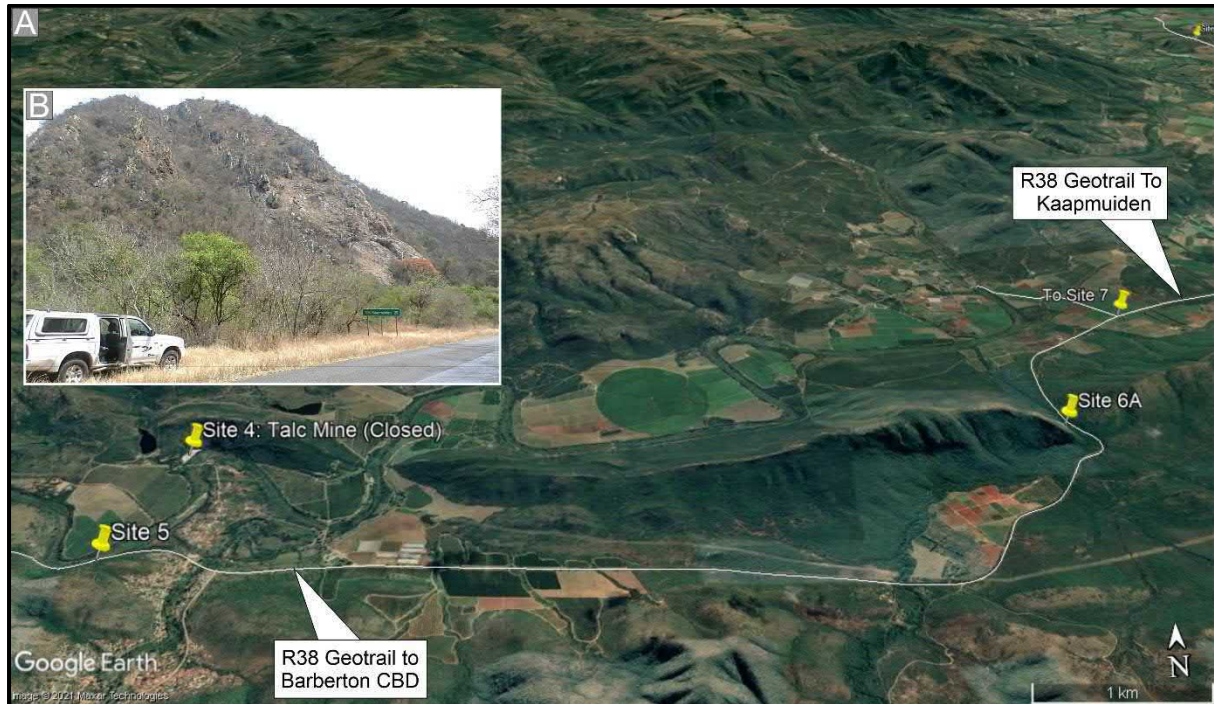


Figure 5.3. A cross section through sandstone of Site 5 reveals ancient sand layers in rocks that preserve clues about the flow direction of the water they settled from. These stacked sand layers were formed as sand grains were dragged by flowing water from the riverbed and then dropped to form little sloped sand piles called ripples. Such layering, and the type of pebble layers at Site 5 tells us that some sandstone layers of the early Earth settled along rivers that were flowing from high-lying areas in towards northern areas of the present-day Barberton area.

Upcoming Stop: Site 6A opposite Honeybird Creek

Stop at: 25°40'27.64"S; 31°14'28.46"E



Directions: Continue along the R38 towards site 6A and park near the ‘Kaapmuiden 20’ signpost. West of this signpost is a steep-sided valley along which flows Honeybird Creek. For this reason, vehicles should be parked on grassy plains clear of the edge of dense tree cover to the west. (A): Google Earth image of a section of the R38 leading to the next site. (B): View of the site.

Did you know?

Low’s Creek was named after Dave Low, a gold prospector who around 1885 discovered gold in Louiville-a township at the southern edge of Low’s Creek (5-6A). The gold Mine developed to exploit that deposit was named after Lily, the daughter of Dave Low.

Site 6: Extremely flattened ancient river stones (Lily Syncline)

Site 6A: Park at 25°40'27.64"S; 31° 14'28.46"E; Site 6B: Park at 25°39'38.24"S; 31°17'9.68"E

This site offers a view of the steep, narrow ridge at the core of the Lily Syncline. A thin gravel bed of ancient pebbles occurs at 25°40'28.74"S; 31° 14'30.18"E. This is a vantage point for viewing the Lily Ridge.

Following the collapse of supporting structures in 2016 (6A1), the Lily Mine became well-known to many South Africans. Visitors may have a moment of silence in honour of the three workers who could not be rescued. Their memory lives on in the hearts of many South Africans.

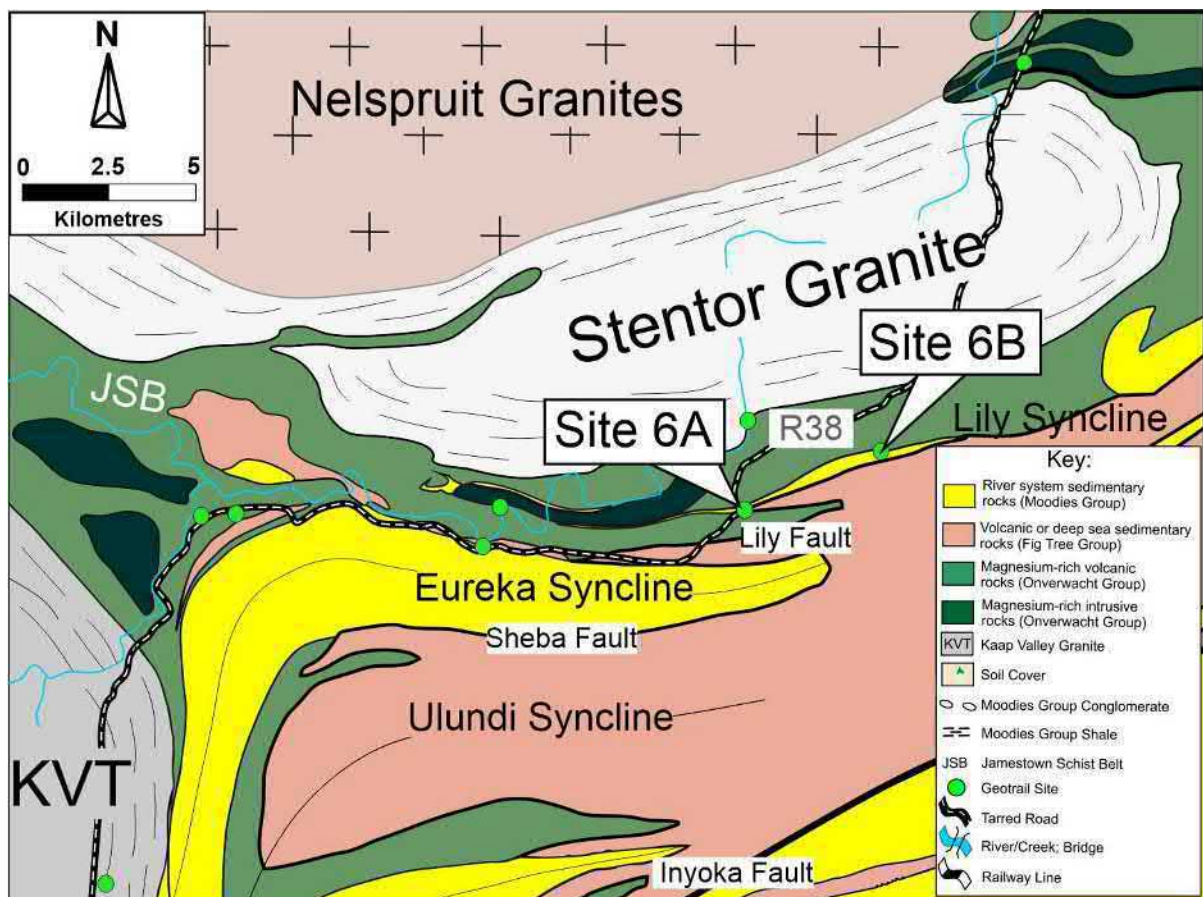


Figure 6A.1. Simplified geological map of the R38 shows the location of Site 6A and 6B within Lily Syncline.

Lily Ridge

The Lily Ridge is a steep narrow structure visible from the R38 between Sites 5 and 6A. It is the core of a larger concave-up fold called the Lily Syncline (6A1). Though the Lily Mine and the Lily Syncline share a name, the mine occurs on a neighbouring fold called the Louville

Antiform (Figure 6A). Both folds are made up of rock layers that were folded and tilted to an almost vertical position (6A2).

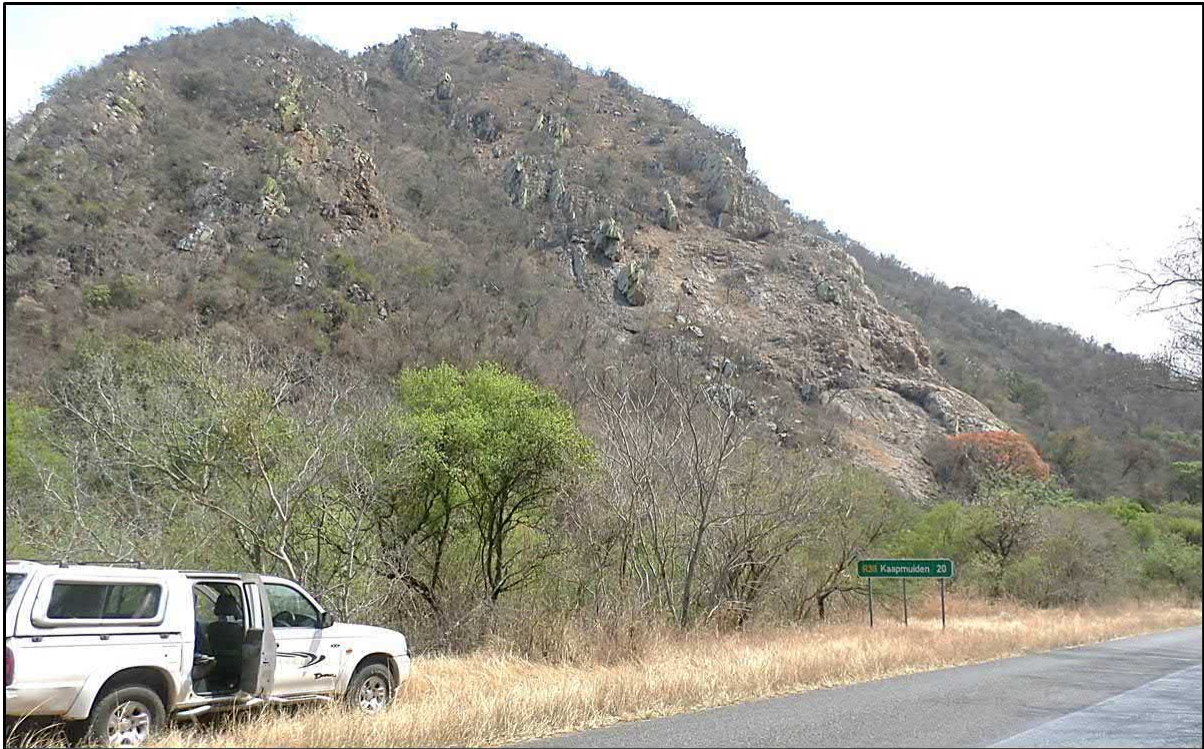


Figure 6A.2. View of the steep, narrow ridge at the core of the Lily Syncline. The dense vegetation along the left-hand side of the road traces the location of Honeybird Creek.

The folding and tilting are understood by some scientists to have been partly caused by sinking of thick rock layers into warmer, deeper parts of the Earth due to their heaviness and position next to and on top of lighter, rising granitic bodies (6A3). Other scientists consider the folding and tilting to be a result of horizontal pushing forces that caused flat rock layers to fold and lean towards the north (6A4). Regardless of the cause, folding tightened rock layers into thick masses of rock that began to sink into deeper parts of the Earth where rocks melt partially to form granitic fluids (6A5).

As granitic fluids rise and cool underground, they may push apart rocks that are near the surface of the Earth (6A5). The Lily Ridge is understood to be the result of flattening caused by the Stentor Pluton as it forcefully rose toward shallower levels in the Earth (6A5).

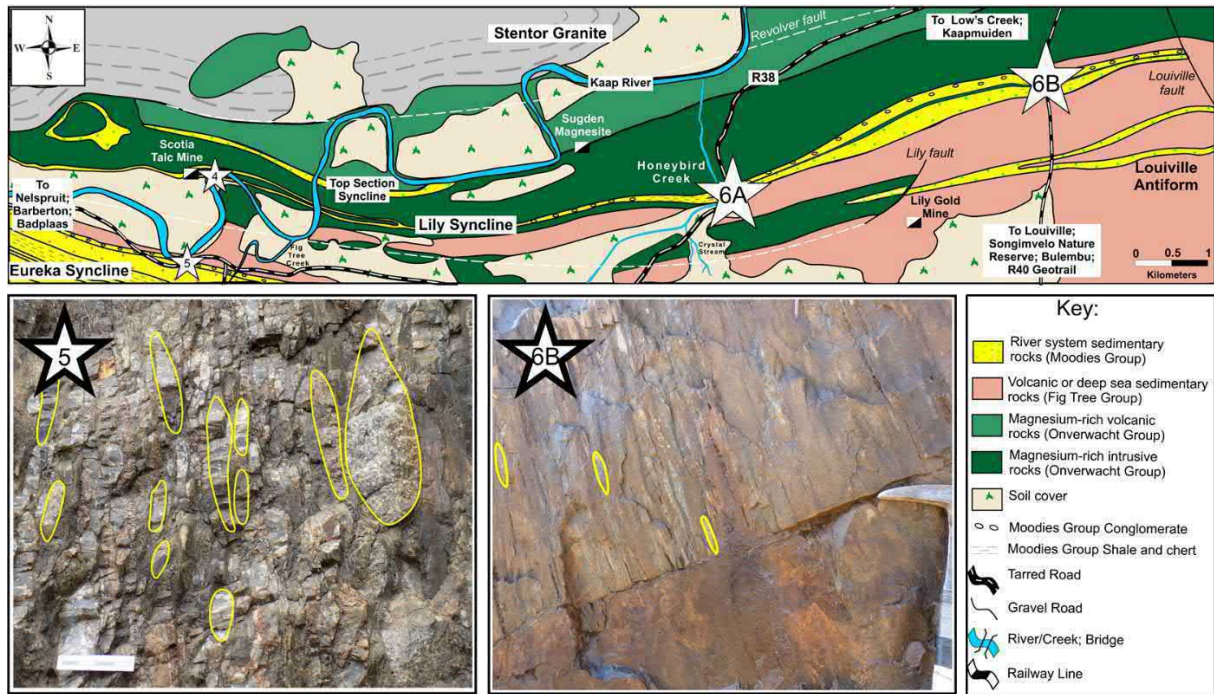


Figure 6A.3. Difference in intensity of pebble flattening in the Eureka (5) and Lily (6A) Synclines. Lense-shaped pebbles are more common in ancient gravel beds of the Lily Syncline regardless of rock type whereas pebble at Eureka Syncline may be lense-shaped or oval-shaped depending on the rock type. This may suggest more flattening was experienced by gravel layers that occur closer to the Stentor Granite.

Crowned Eagle's Nest

A short walk towards the nearest road bend north of the parking site offers a view of a crowned eagle's nest perched on a branch on the eastern side of the road (Figure 6A.4.). This bird of prey with a typical wingspan longer than 1.9 meters (6A6). It has the strongest talons (nails) when compared to other African eagles (6A7). This helps it hunt prey as big as small duikers (6A7). If you are lucky enough to get a good picture of the bird, share it at (<https://www.facebook.com/R38Geotrail/>).

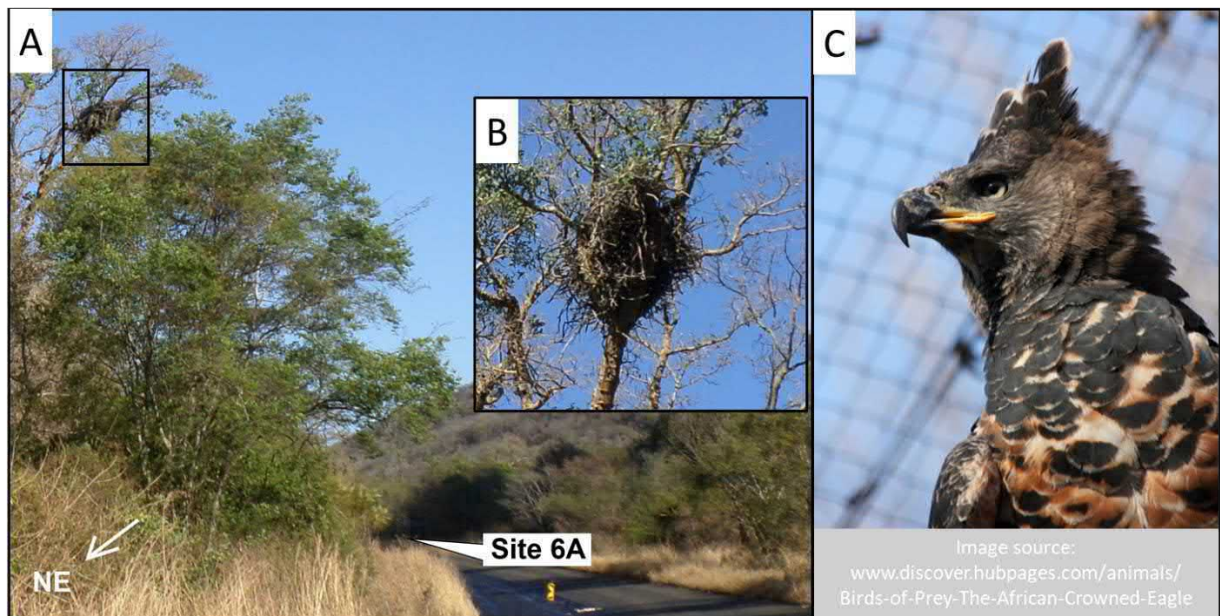


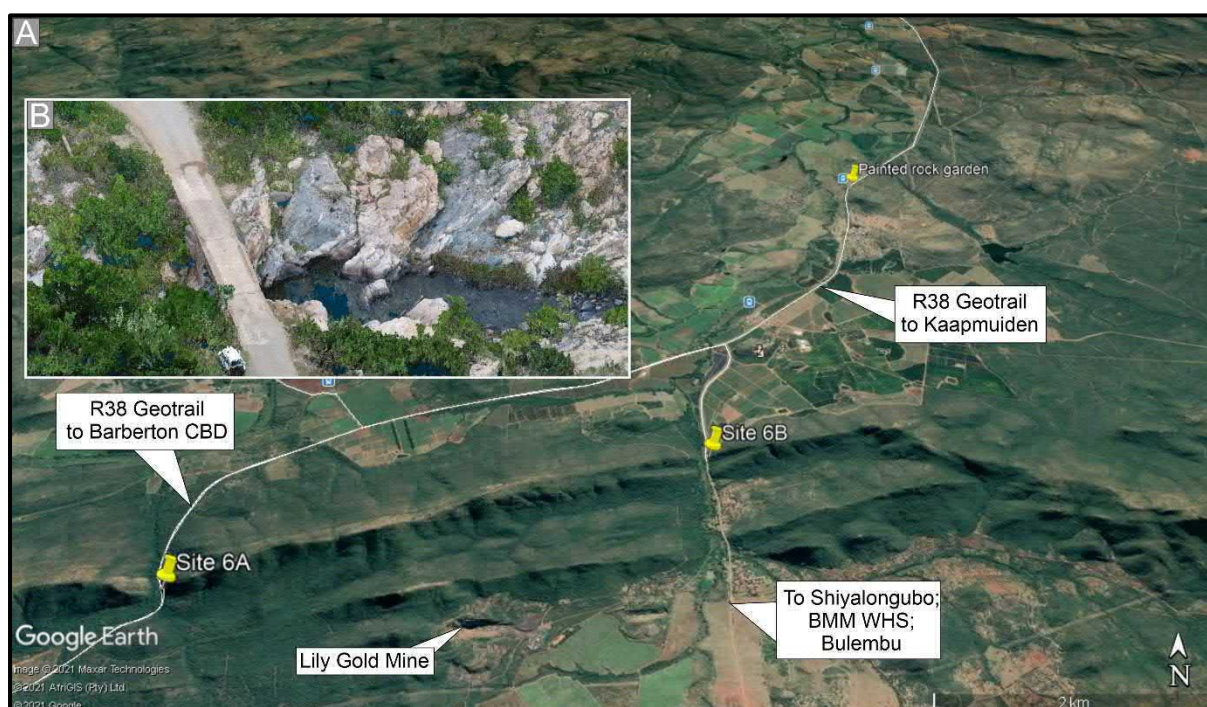
Figure 6A.4 Roadside nest of a local bird of prey. (A): View of a crowned eagle's nest viewed from the north. Part of the ridge exposed at Site 6 can be seen in the middle center of the image. (B): Close view of the nest from below. (C): Image of an African crowned eagle, image source

Marks on the famous fossil skull of the 'Taung Child' member of the *Australopithecus africanus* homonin species of prehistoric man indicate injuries caused by a Crowned Eagle (6A6). This 1924 discovery of the Taung Child fossil in Taung, Northwest Province of South Africa was the first such fossil discovery from a limestone cave in Africa (6A6). Features of the skull suggested that ape-men had been walking upright by at least 2 million years ago (6A7). However, new fossil discoveries continue to challenge existing ideas about the origin of humans, and when apes-and mammals in general-began walking upright on two feet (6A8).

Upcoming Stop: Site 7 at Honeybird Siding

Left turn leading to Site 7: $25^{\circ}39'25.33''\text{S}$; $31^{\circ}15'13.20''\text{E}$;

Parking: ($25^{\circ}38'57.75''\text{S}$; $31^{\circ}14'32.35''\text{E}$)



Directions: Continue along the R38 until you reach a rectangular ‘Honeybird’ signpost facing opposite traffic. Turn left onto a gravel road a few meters beyond the signpost and drive towards Site 7. Good parking spots are on the western side of the river near the ‘Honeybird’ signpost. Large trucks use this road, best to park vehicles away from road edges. (A): Google Earth image of a section of the R38 leading to the next site. (B): View of the site.

Did you know?

The name of Honeybird creek on the western edge of the road at site 6A comes from a legend of a honeybird said to have led a man to a snake which he killed near a rock with visible gold. The gold deposit at the site was later registered as Honeybird Reef (6A-7).

Site 7: Sandwiched Greenstone (Stentor Granite Contact Zone)

Site 7 is in Honeybird Siding near a bridge where a gravel road crosses the Kaap River. Rock exposures of interest are found from 50 meters south of the bridge to 200 meters north of the bridge along the Kaap River. With permission from farm owners (dortan@soft.co.za), visitors may walk down from the bridge towards high-lying rocks exposed along the river. Kindly avoid littering or walking close to the river as crocodiles are known to live in the area.

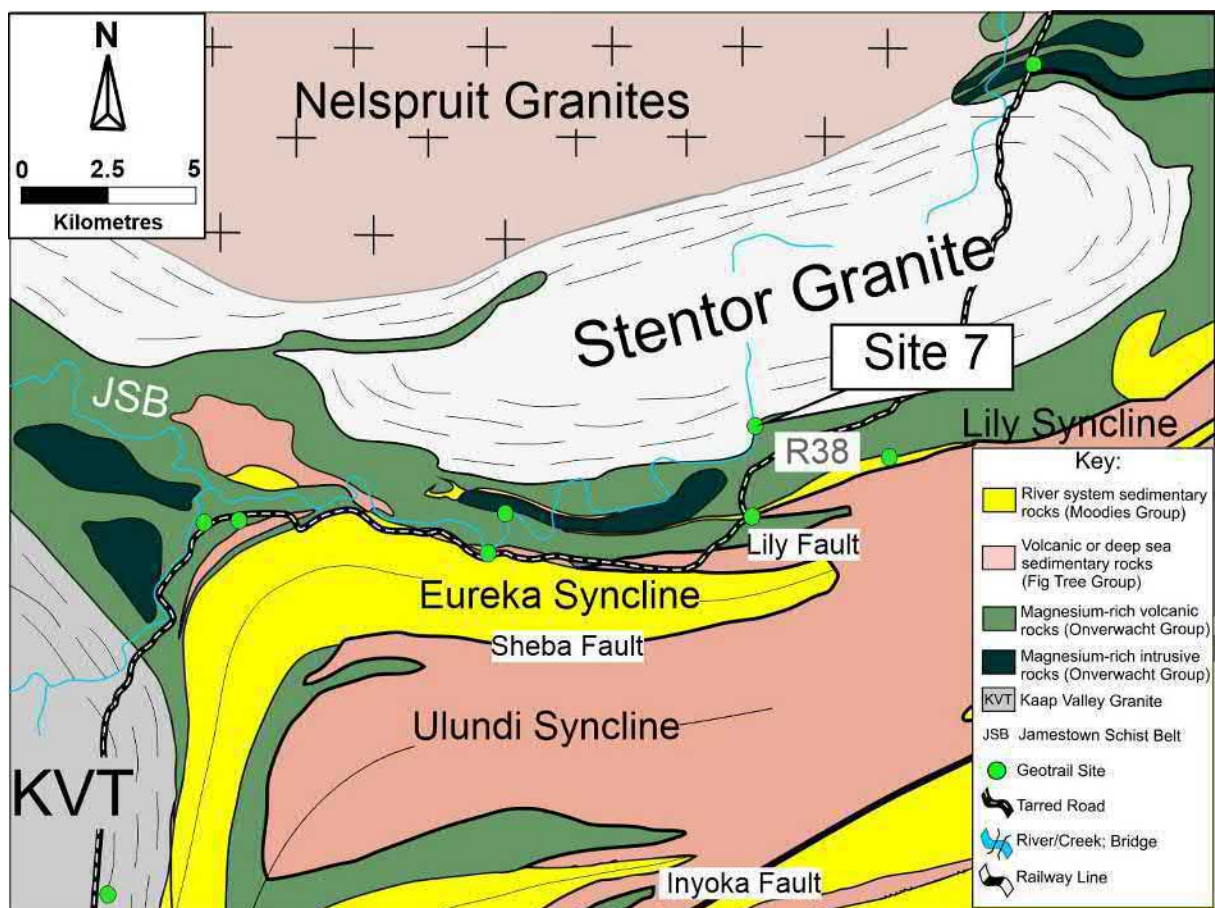


Figure 7.1. (A), (B): Simplified geological map of the R38 shows Site 7 as a sharp edge between the Stentor Granite and greenstones of the Onverwacht Group.

Sandwiched Greenstones

Rock exposures at Site 7 are of greenstone layers that are sandwiched between granite sheets. The arrangement of granitic sheets and greenstone sheets of Site 7 shows that partially solid fluids of the Stentor Granite forcefully rose against and into overlying greenstone layers (7.1). This occurred while unstable greenstone layers were sinking along edges of the Stentor Granite due to the heaviness of the greenstones when compared to the granite (7.2). As a result, folds and other structures related to stretching are preserved in both the granite and invaded

greenstones (Figure 7.3). The shapes and patterns preserved in rock layers of Site 7 show that the granite rose towards the north, rising against greenstones that sagged towards the south (Figure 7.1.). This granite-up-greenstone-down movement is common among other contact zones of the earliest preserved early granites of the region (7.3).

Looking closely at granitic sheets of Site 7 reveals a crushed texture formed by blocky granite fragments within fine granite fragments (7.1). This texture is one of the signs that tells us that the granite rose forcefully into the current position. As a result, minerals in it were ground into a fine matrix that wraps around uncrushed pieces (7.1).

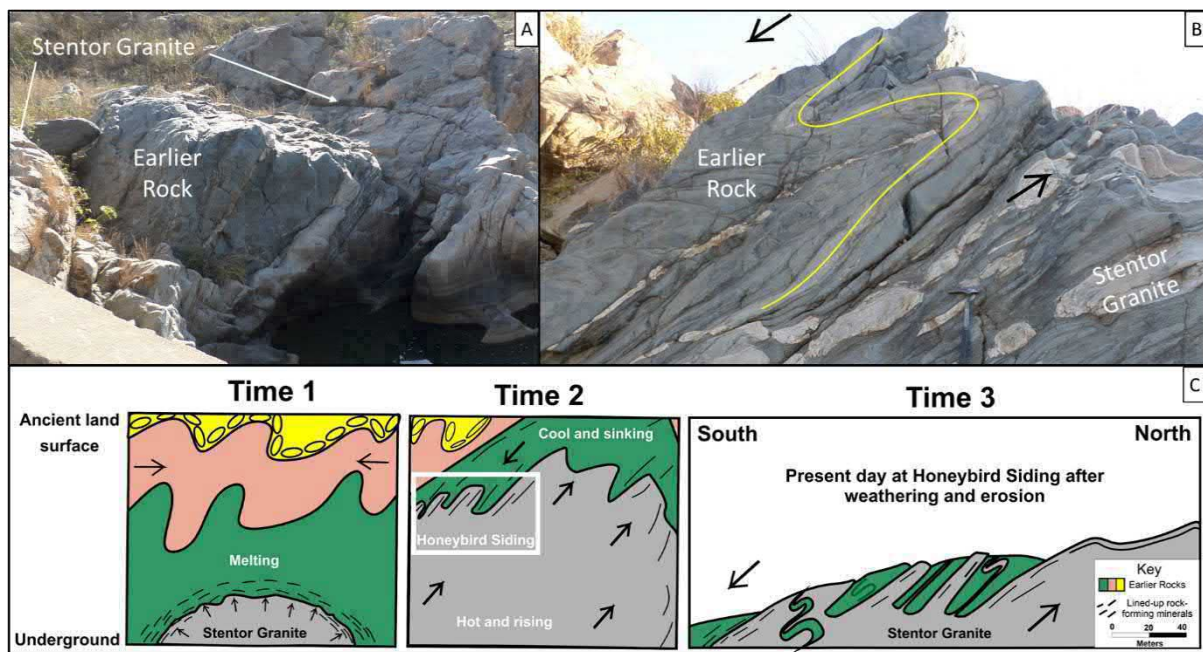
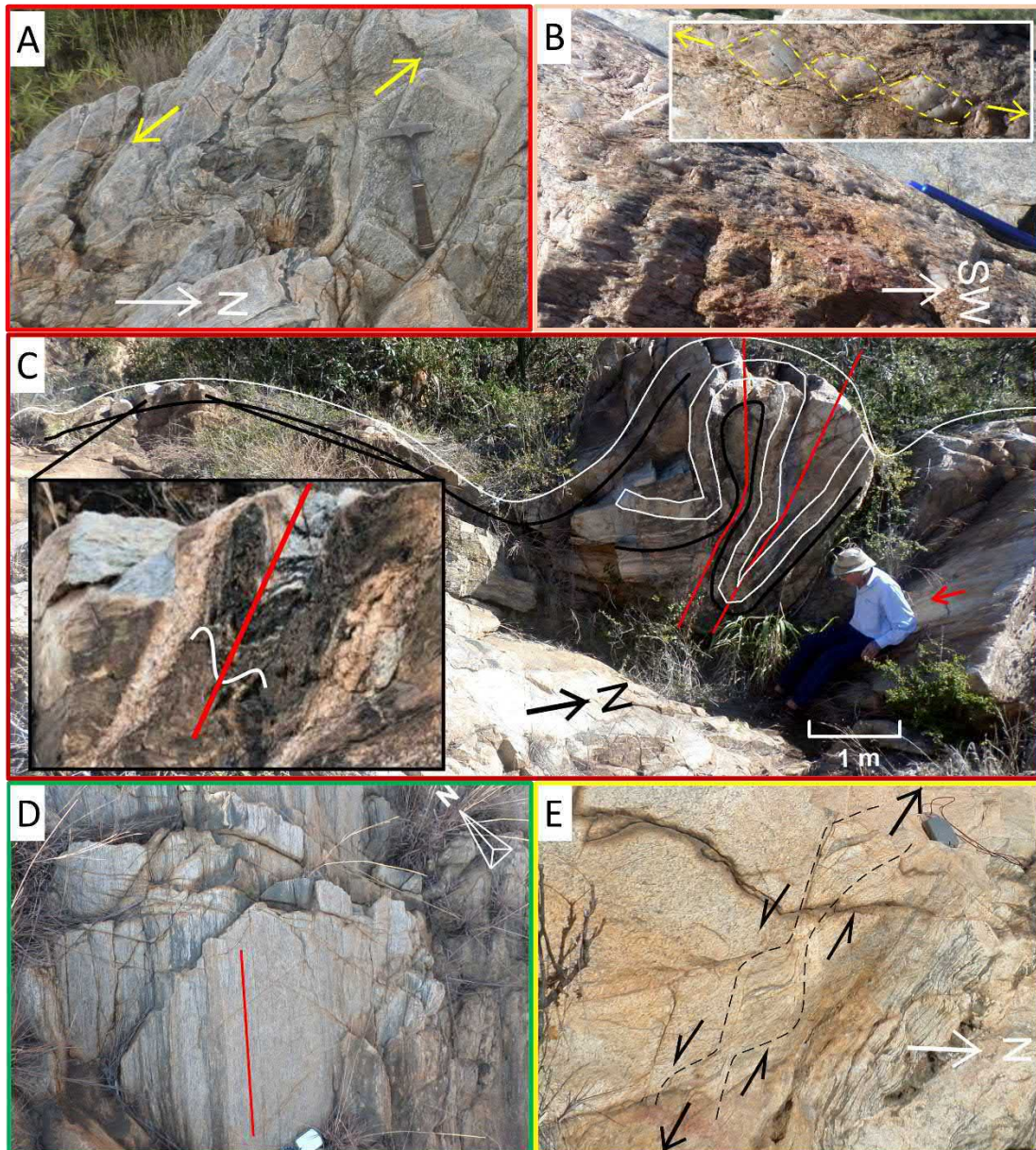


Figure 7.2. (A): Stentor Granite sheets against and within a sheet of greenstone. Some granitic fluid flowed along fractures of the greenstone sheets to form invasive layers called sills. (B): Common fold shape in both granite and greenstone sheets. Such folds tell of the direction of ancient pulling forces that were active when the Stentor Granite became solid.



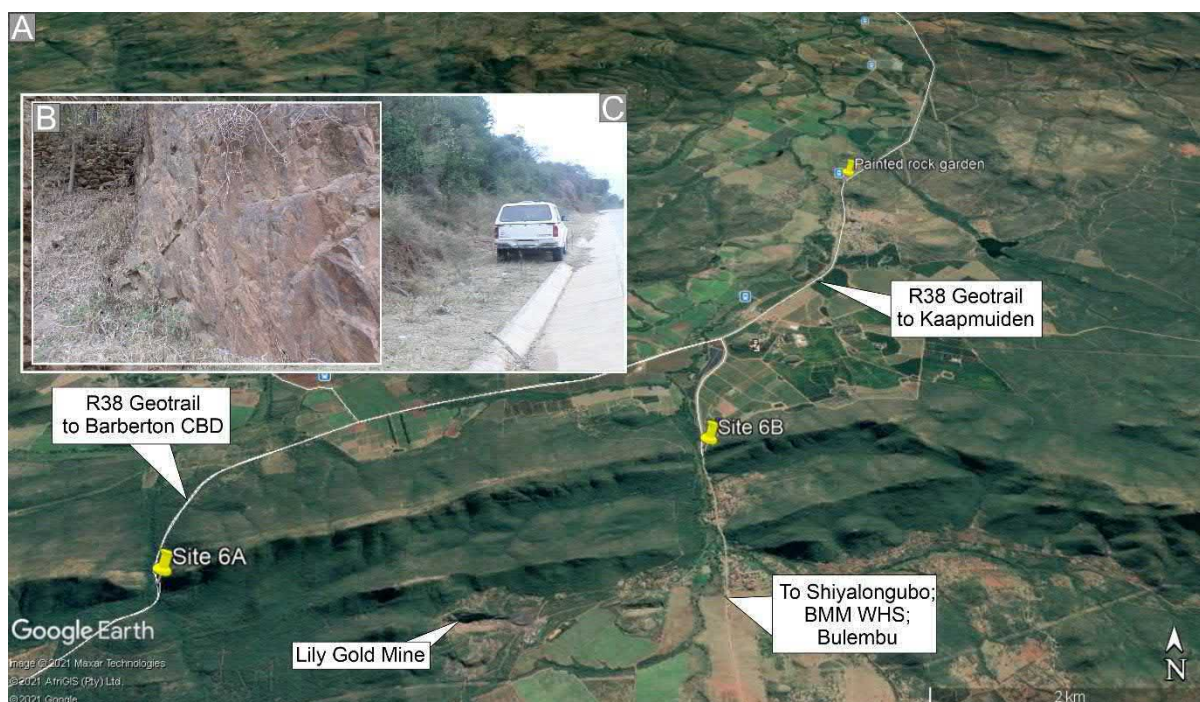
Sites where rocks above can be seen



Figure 7.3. Good rock exposures where one can see main structures of the Stentor Granite. All structures indicate up and to the north transport of the granite. (A): Folded granite sheet and greenstone layer. (B): Extended blocky granite fragments. (C): Folds in cores of domes responsible for the generally wavy landscape of Site 7. (D): Grooves caused by the extension of mineral clusters to form rod-like structures. (E): Extended layer. Direction arrows: N=North, SW= Southwest. Colours of star location markers match image borders of the associated image. Locations are marked on a snapshot image of a 3-D drone model produced by Thomas Jones.

Upcoming Stop: Site 6B north of Louville

Stop at: 25°39'38.24"S; 31°17'9.68"E



Directions: From Site 7, drive along the gravel road towards the R38 and turn left at the intersection. Proceed along the R38 and turn right at the Louville signpost. (A): Google Earth image of a section of the R38 leading to the next site. (B): View of the northernmost rocks at the start of the trail. (C) View of traverse towards the south along which Site 6B rocks are exposed.

Did you know?

The Louville road at Site 6B links Low's Creek to the Barberton-Makhonjwa Mountains World Heritage Site. It forms part of the local 'Genesis Route', a road network that extends across the Barberton Greenstone Belt, intercepting a number of tourist attractions. One of these is an old Mining town called Steynsdorp which was a site of a goldrush in 1885 until many gold-seekers left for Johannesburg in 1886 when the Wits goldfield was discovered (7-6B). Currently, Steynsdorp offers the Ebutsini Cultural Village where visitors may explore 4x4 routes, hiking trails, braai areas and aspects of Swati culture.

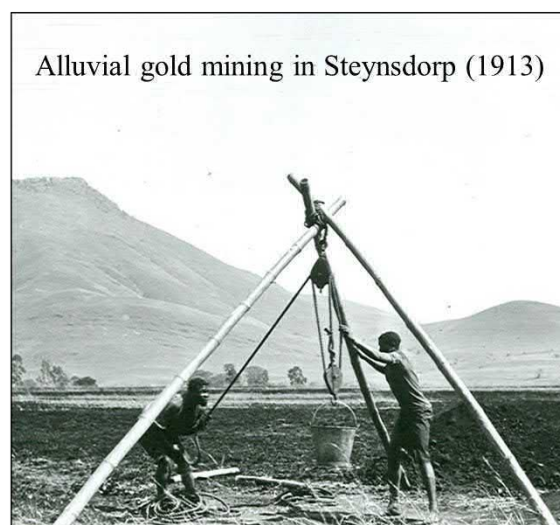


Figure 7-6B-2. Labourers at a sinking pit in the Steynsdorp alluvial goldfield. Image source: Barberton Museum Archives.

Site 6B: Extremely flattened ancient river stones

This 250m mini-trail provides the best view of folded rock layers which make up the Lily Syncline. From north to south, rocks exposed are made up of rock pieces which decrease in size, with the center exposing rocks made up of the smallest particles.

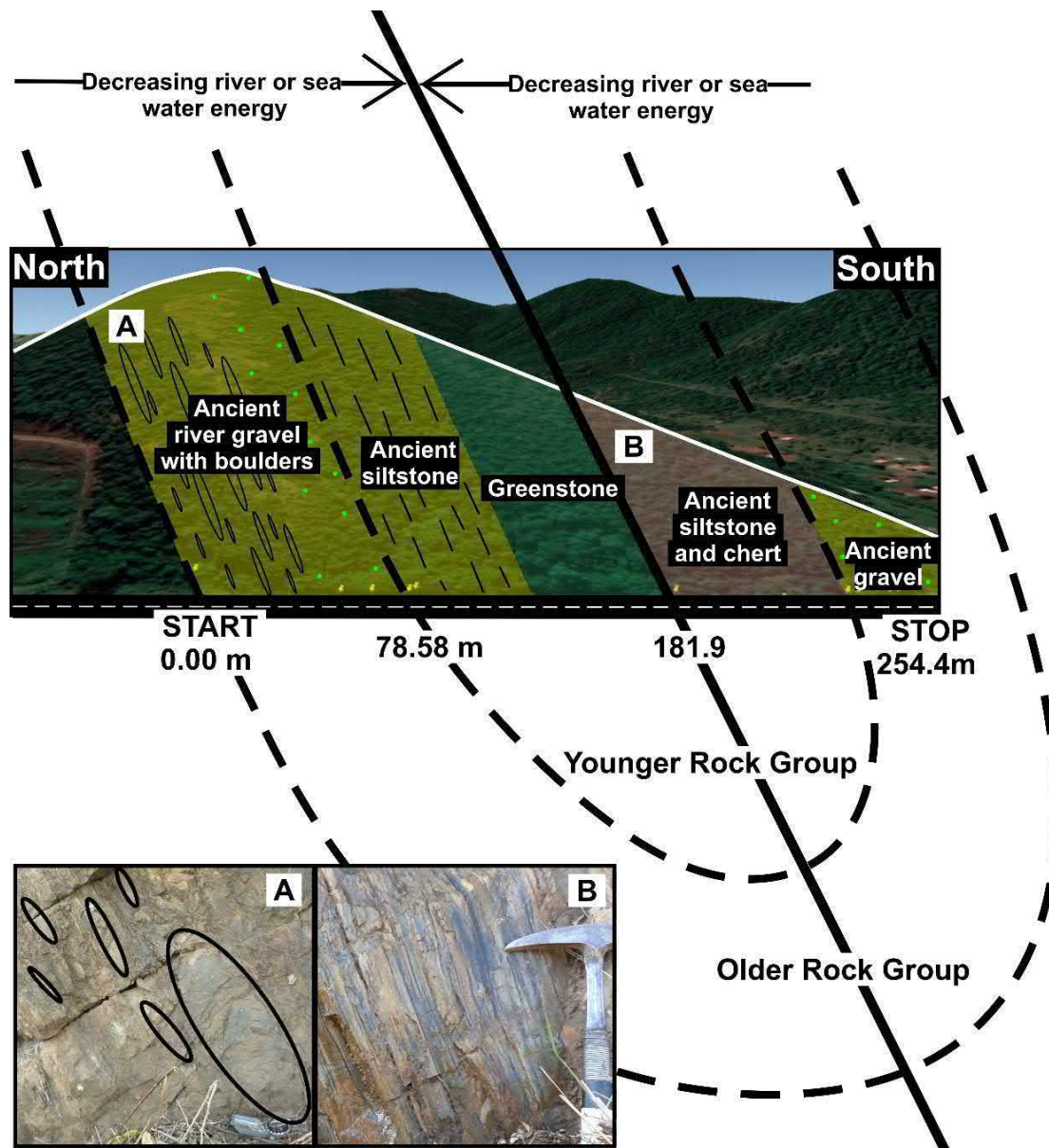


Figure 6B.1. Rocks that make up the core of the Lily Syncline in the Low's Creek become finer towards the south along. The largest ancient river stones in the north (A) give way to fine, layered rocks made up of ancient river silt and glassy rocks formed in deep sea water (B). The greenstone in the centre of (B) forms a valley between the steep edges of the Lily Syncline.

How did rocks of the Lily Syncline form? (6B)

Flowing water in a river channel, sea or a lake can pick up rock fragments (sediments) on the surface of the Earth, carrying them until a decrease in flow energy occurs. When this happens, rock fragments begin to settle out of the water and collect on the surface of the Earth to form layers of sediments. Because bigger rock fragments can only be picked up by fast-moving water, a decrease in flow energy causes bigger fragments to settle out first and settle into layers. As the flow energy of the water continues to decrease, even smaller rock fragments begin to settle out. In this way, sediments of bigger rock fragments may form one layer and smaller fragments form another depending on the flow energy of the water.

The rocks exposed at the start of the Site 6B trail are made up of bigger rock fragments mixed with smaller rock fragments. This mixed layer of big and small rock fragments tells of an ancient river system of fast-flowing water with high enough energy to pick up big and small rock pieces from nearby hills. These rock fragments may have settled out rapidly in response to a fast drop in flow energy of the water in which they were carried. Environments where such changes in flow energy occur include areas where rivers enter gentle slopes at the foot of steep hills. Rocks formed in this way are part of the older group of rocks drawn in (Figure 6B.1).

The younger group of rocks at Sit 6B are mainly made up of fine grains. Such rocks may have formed in deep sea or lake waters of very low flow energy. In those conditions, the low energy of the water cannot carry fine rock fragments, so they settle into silt or mud layers. When this water body contained high content of dissolved rock-forming minerals, these too began to settle into chert, that is, clusters of very fine silica-bearing crystals.

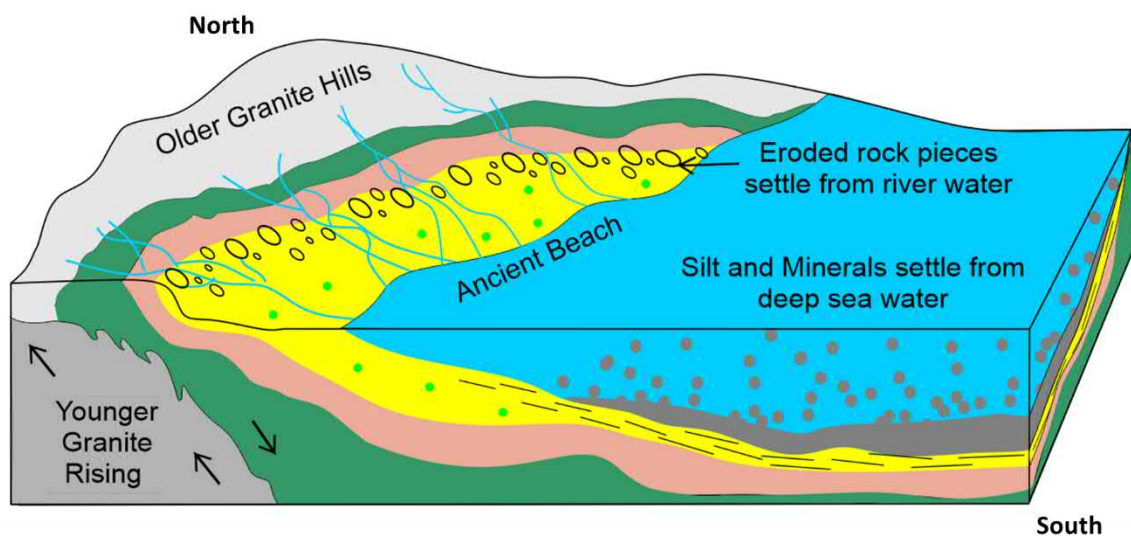


Figure 6B.2. Rocks at the core of the Lily Syncline appear to have been laid down between steep mountain ranges in the north and deep-sea water in the south.

Upcoming Stop: Painted Rock Garden near Revolver Creek

Stop at: 25°36'52.16"S; 31°18'27.97"E

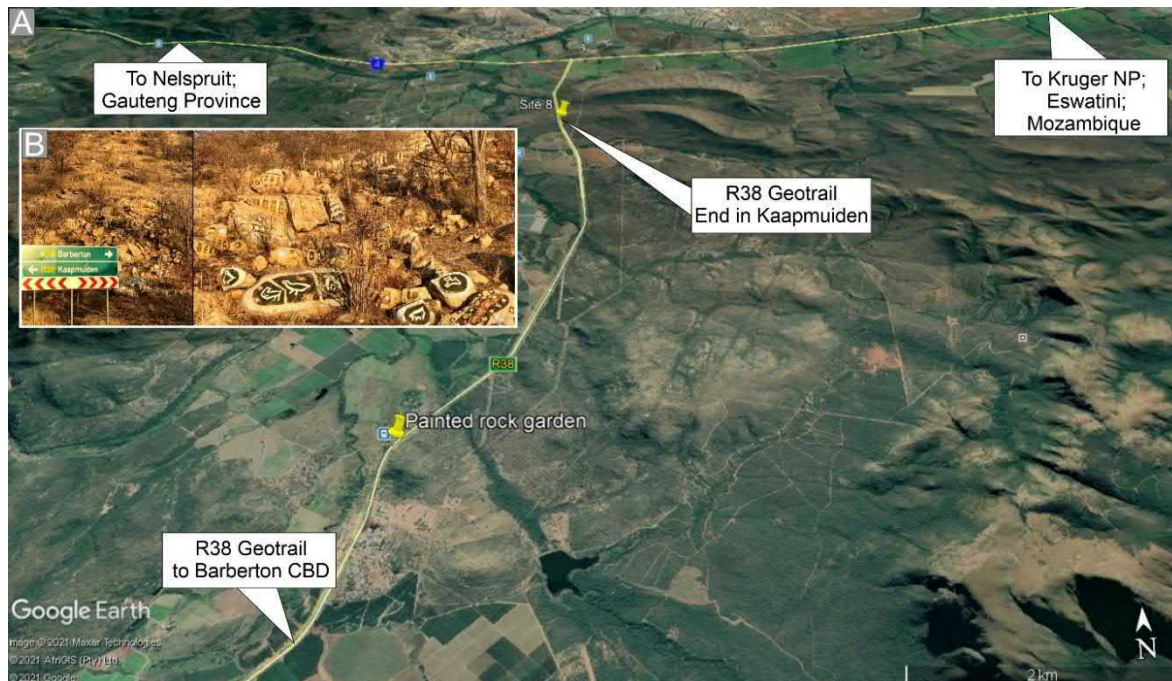


Figure 6B-NM-1. Head back to the R38 intersection and turn right towards Kaapmuiden. Stop near the 'Esperado' signpost on your left. The view site is on a granite hill on the opposite side of a left turn to Esperado. (A): Google Earth image of a section of the R38 leading to the next site. (B): View of the site.

Did you know?

Artists who produce work that differs from what is thought to be normal in western art are called "Outsider Artists". Common among such artists is the little or no training in art, and their social isolation (6B-NMA). South Africa boasts few such artists. They include a painter named Nukain Mabuza from near Low's Creek (6B-NMA).

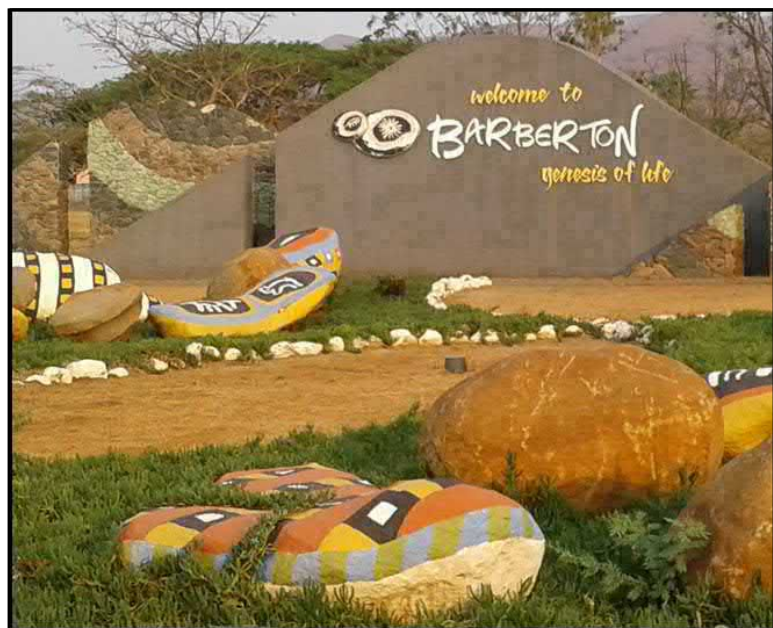


Figure 6B-NM-2. Designs inspired by the work of Nukain Mabuza were used to decorate Barberton Gateways in his honour.

View Site: Painted Rock Garden near Revolver Creek

For tourists in the 1960/70s, this site was a popular stop where visitors would come to admire painted huts and granite boulders arranged into a ‘stone garden’ (NMA1). Painted and arranged by a self-trained artist named Nukain Mabuza (NMA2), boulders in the rock garden display artistic drawings of animal figures, dots, squares and stripes. Widespread recognition of Nukain Mabuza’s talent came after his passing in 1981 (NMA2). This includes theatre plays about his life (NMA2), and the listing of his stone garden as one of few environmental works in the world which were created by self-trained artists (NMA1).



How do granite boulders form?

On the surface of the Earth where most agents of weathering and erosion are active, zones of weakness like cracks in the granite are widened over time (NMB1). Surface cracks may be formed as the top rock cover above a solid granite is eroded away (NMB1). This allows the granite to release pressure built up when the granite was buried beneath a thick cover of rock layers. Fractures formed in this way may be linked, forming in different directions that cut through each other. The cross-cutting of fractures provides good channels to be used by weathering agents. The more fractures a rock has, the easier it becomes for agents of weathering and erosion to transform the rock into blocks. The sharper edges of granite blocks are easier targets for weathering (NMB1). In this way, edges are first to waste away, leaving behind the rounded core of the granite boulder. In the present day, parts of the Stentor Granite exposed to more weathering and erosion have formed boulders on the surfaces of hills with solid granite masses at the core.

Upcoming Stop: Site 8 south of Kaapmuiden

Stop at: 25°33'23.53"S; 31°19'32.52"E



Directions: Continue along the R38 until you see a white, rectangular signpost on the opposite side of the road. The signpost reads 'PettigrewNek' and faces traffic of the opposite lane. (A): Google Earth image of a section of the R38 leading to the next site. (B): View of the site.

Did you know?

The R38 was a wagon route known as Pettigrew Road. This was in honour of Robert Pettigrew, a surveyor who developed the road linking Barberton to Mozambique around 1885 (NM-8). The local name for the upcoming site 8 is Pettigrew Nek. When wet, this portion of the R38 was difficult to cross because of the type of minerals in rocks of that area (NM-8).

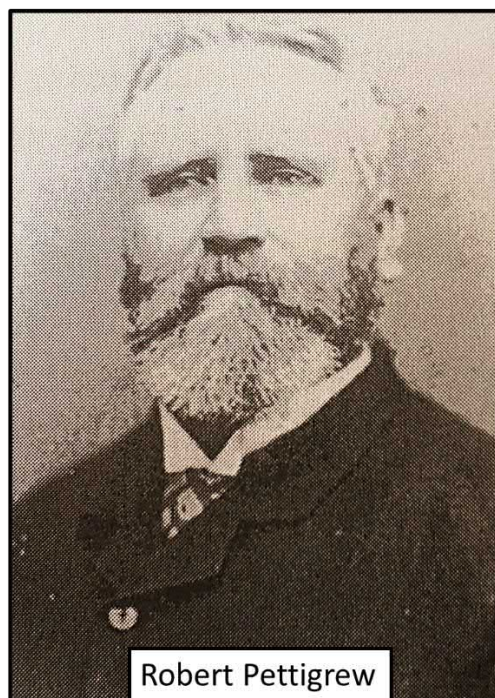


Image source: Bornman (2017).

Site 8: Magnesite Mineralisation (25°33'23.53"S; 31°19'32.52"E)

Site 8 is a roadside stop across from an old quarry where magnesite was mined in the early 1900s-be careful not to fall in.

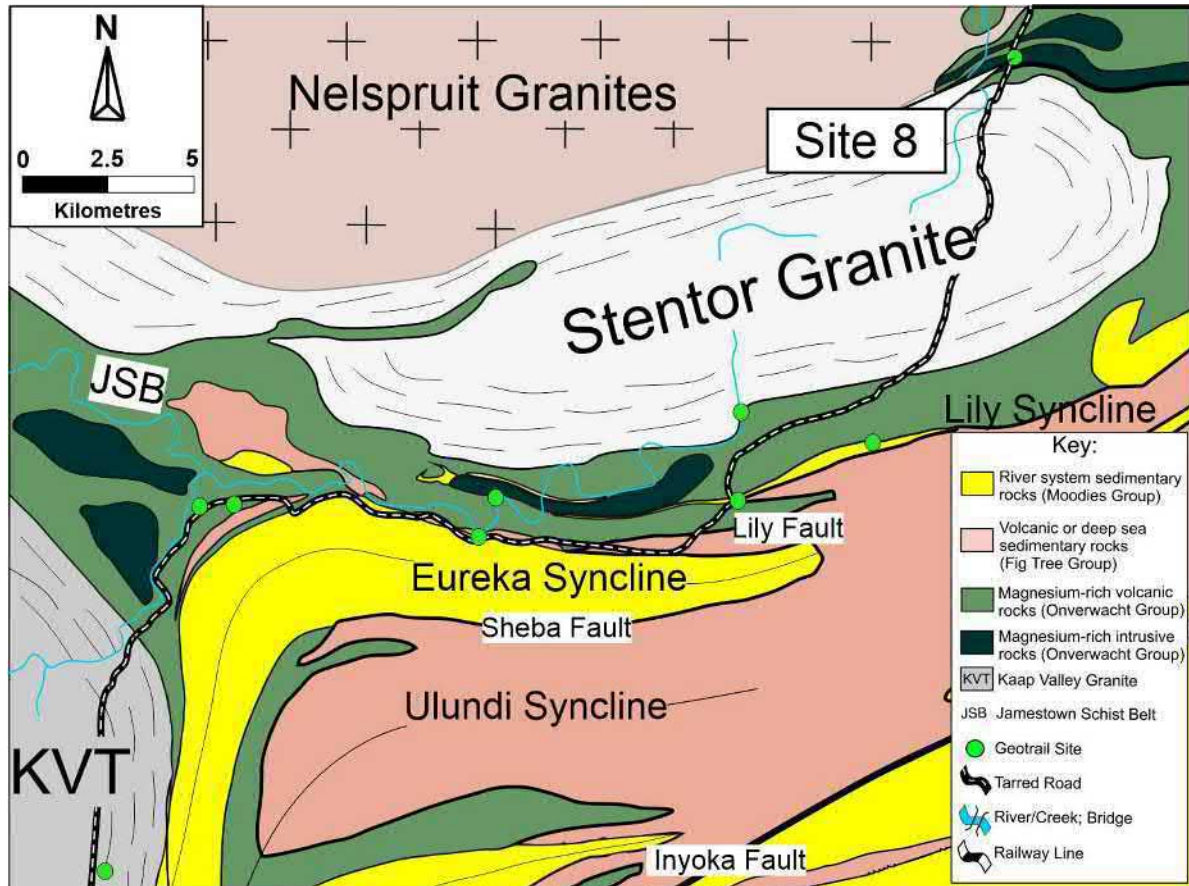


Figure 8.1. Simplified geological map of the R38 geotrail shows that Site 8 is located in intrusive magnesium-rich rocks on the margin of the Stentor Granite.

Rocks that contain magnesite

The focus at site 8 is on two main materials that form the roadcut surface. The main one of these is a brown-green rock with many cracks. The other is a white mineral within those cracks. The cracked rock is a coarse greenstone and the white mineral is called magnesite.

The coarse, green-gray rock at Site 8 made up of mineral crystals rich in magnesium (8.1). It formed when a pool of melted greenstone cooled slowly underground, forming mineral layers rich in different elements like magnesium, iron and calcium (Fig. 8.2). When the layered rocks were cool and solid, forces acting on the crust of the Earth caused the rocks to crack. These cracks were later filled with a fluid rich in dissolved carbon dioxide (8.2). The magnesium on

crack surfaces of the rock layers rich in magnesium combined with dissolved carbon dioxide in the incoming fluid. This produced different types of magnesium carbonate minerals that were later turned into magnesite (8.3.).

The source of dissolved carbon dioxide is not fully understood. Likely sources include the environment or fluids from buried greenstones or from the Stentor Granite (8.4). What is better understood is the usefulness of magnesite formation in using up carbon dioxide gas from the atmosphere to form a solid that is less harmful to the environment (8.5.). In this way, greenstones rich in magnesium play a similar role to plants that absorb carbon dioxide in the atmosphere to store it as a solid substance (8.5).

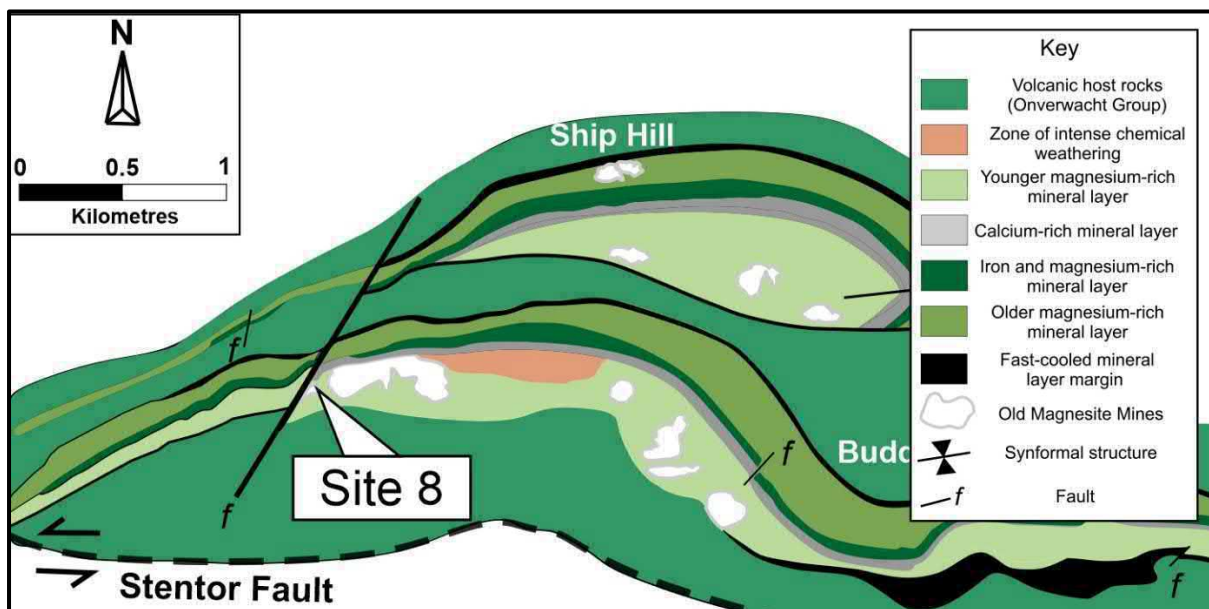


Figure 8.2. Geological map of the layered intrusive rocks near site 8 shows that old magnesite mines along the R38 geotrail occur in magnesium-rich layers within a stack of rock layers rich in other elements.

Magnesite

Magnesite from four sites near the margins of the Stentor Granite may occur as folded layers that lean towards different directions (Figure 8.2. A). When some magnesite layers are taken out of the host rock and all surfaces are viewed, we see that the folded layers form a special texture where each fold only extends for a short distance. We also see that each fold is not aligned with nearby folds of the same layer. Instead, magnesite layers appear to have folded in all directions, producing the creased texture seen in Figure 8.3. C, D. There are many reasons to why mineral layers fold (8.6). Reasons include pushing or pulling forces, the expansion or sliding of a layer, and many more.

Some magnesite veins at Site 8 only fill part of the crack they occur in. This suggests that the cracks now filled by magnesite were once filled by air, a rock or mineral that has since dissolved to form a void (8.7).

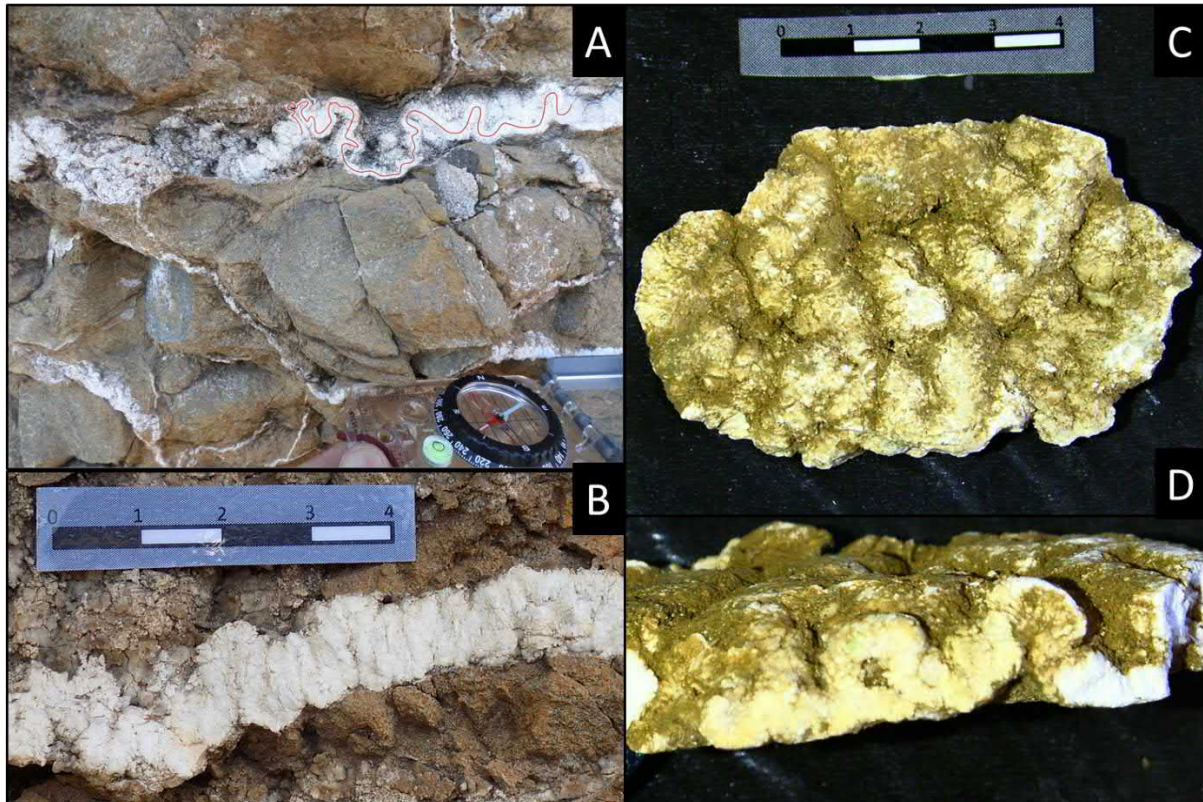


Figure 8.2. (A): Magnesite in a crack within a coarse greenstone is folded into structures that lean and extend for short distances in different directions. B: Magnesite bands that are folded tightly appear to curve in similar directions when seen from the side, but this is an illusion. (C&D): The best way to see the true shape of an object is by looking at more than one surface. The top surface (C) and side view (D) of a magnesite layer shows the true form of magnesite layers from this site. Such a texture suggests that the layer expanded in limited space.

Take-home message

As one of the earliest materials on planet Earth, rocks exposed along the R38 paint a picture of what the surface of the early Earth may have looked like. Here we also see the useful products that form when different fluids flow through magnesium-rich greenstones of the region.

We can access these greenstones and their products today because of the granitic bodies of the region (C1). These relatively unsinkable granitic rock masses found at most margins of the Barberton Greenstone Belt have kept the cooler and denser rock layers of the greenstone belt from sinking back into the depths of the Earth (C2). This association of rocks produced a stable landmass that gradually grew and combined with other early landmasses to eventually form what is now the oldest, best-preserved part of the African continent (C3).

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