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EXCURSION GUIDE TO THE GEOLOGY OF THE BARBERTON GREENSTONE BELT

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EXCURSION GUIDE TO THE GEOLOGY OF THE BARBERTON GREENSTONE BELT

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FOREWORD

The Barberton granite-greenstone terrane has emerged as one of the classic Archaean geological environments worldwide in which to study the nature and processes of early Earth history. Over the years numerous scientists, covering a wide spectrum of earth science disciplines, have visited the region for the purposes of undertaking research and also to view the exceptionally well-preserved and well-exposed geology the area has to offer. Many issues pertaining to Archaean geology, such as crust forming processes, geodynamics, atmosphere and seawater composition, early impact history, and the origin of life, have been or are extensively studied in the Barberton terrane. The 2004 Field Forum on Processes on the Early Earth, jointly sponsored by the Geological Society of America and the Geological Society of South Africa, aims at providing a vehicle for investigation and discussion of these problematics. Consequently, the excursion route described herein (Fig. 1) was designed in such a way that it provides both an introduction to the stratigraphy and lithological variation of this greenstone belt and information about its tectonic history, and covers prime exposures where the aforementioned main issues can be examined.

A number of informal as well as formal excursions have been organised at different stages to enable visitors and scientists to view some of the classic exposures that have contributed to the understanding of the tectono-magmatic, sedimentary and metamorphic processes that occurred in the area over 3 billion years ago. This guidebook is intended to supplement existing guides to the geology of the region, which were prepared by different groups of scientists for the purposes of highlighting specific topics, or of providing a general overview of the geology. Some of the earlier guidebooks are listed below, but may not be readily available as they were usually prepared in small numbers specifically for delegates registered and participating on some of the formal excursions led through the area from time to time. The more significant guidebooks include:

1) Pretorius, D. A. (Compiler) (1965). *Guide-Book to the Barberton Mountain Land*. Post-Congress Excursion, 11-18 July, 1965. Econ. Geol. Res. Unit, University of the Witwatersrand, Johannesburg, June 1965, 33 pp.
2) Anhaeusser, C. R. (Ed.) (1981). *Barberton Excursion - Guide Book, Archaean Geology of the Barberton Mountain Land.* South African Geodynamics Project, Geological Society of South Africa, Johannesburg, July 1981, 78 pp.
3) Ashwal, L. D. (Ed.) (1991). *Two Cratons and an Orogen. Excursion Guidebook and Review Articles for a Field Workshop Through Selected Archaean Terranes of Swaziland - South Africa - Zimbabwe.* IGCP Project 280, The Oldest Rocks on Earth. Department of Geology, University of the Witwatersrand, Johannesburg, 312 pp.
4) Anhaeusser, C. R. (Compiler) (1999). *Field Excursion to the Barberton Mountain Land, South Africa*. 62nd Annual Meeting of the Meteoritical Society, Johannesburg, South Africa, 17-20 July, 1999,142 pp.
5) Lowe, D.R. and Byerly, G.R. (2003). *Field Guide to the Geology of the 3.5-3.2 Ga Barberton Greenstone Belt, South Africa.* Guidebook prepared for Field Conference, Archean Surface Processes, June 23-July 2, 2003, 184 pp.

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ITINERARY

Day One	**Sunday, 4 July 2004**
Route	Johannesburg - Ogies - Carolina - Badplaas - Tjakastad - Badplaas (*c*. 350 km)
Stop 1.1	Tjakastad water reservoirs Overview of the geology of the Onverwacht Group 3.5 Ga volcano-sedimentary sequence of the Theespruit Formation
Stop 1.2	Stream section northeast of Tjakastad 3.48 Ga komatiites and komatiitic basalts of the Komati Formation
Stop 1.3	Roof pendant in Theespruit pluton Sandspruit Formation
Stop 1.4	Spinifex-textured ultramafic rocks of the Sandspruit Formation
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Stop 2.2	Brandybal farm Sandspruit Formation komatiites, komatiitic basalts and migmatites
Stop 2.3	Nederland metasedimentary rocks 3.5 Ga volcano-sedimentary rocks of the Sandspruit Formation and granite-greenstone contact relationships
Day Three	**Tuesday, 6 July 2004**
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Stop 3.2	Barite Valley Syncline Barite, impact spherule layers, bedded cherts and chert dykes, contact between the Mendon Formation (Onverwacht Group) and the Fig Tree Group
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Option 4.2	Msauli Chert, Barberton greenstone belt Accretionary lapilli, altered ultramafic rocks and chert dykes, contact between the Mendon Formation and the Fig Tree Group
Option 4.3	Stromatolitic dolomites of the 2.5 Ga Malmani Subgroup, Transvaal Supergroup

Transfer to Vredefort Dome for second part of Field Forum

EXCURSION ROUTE

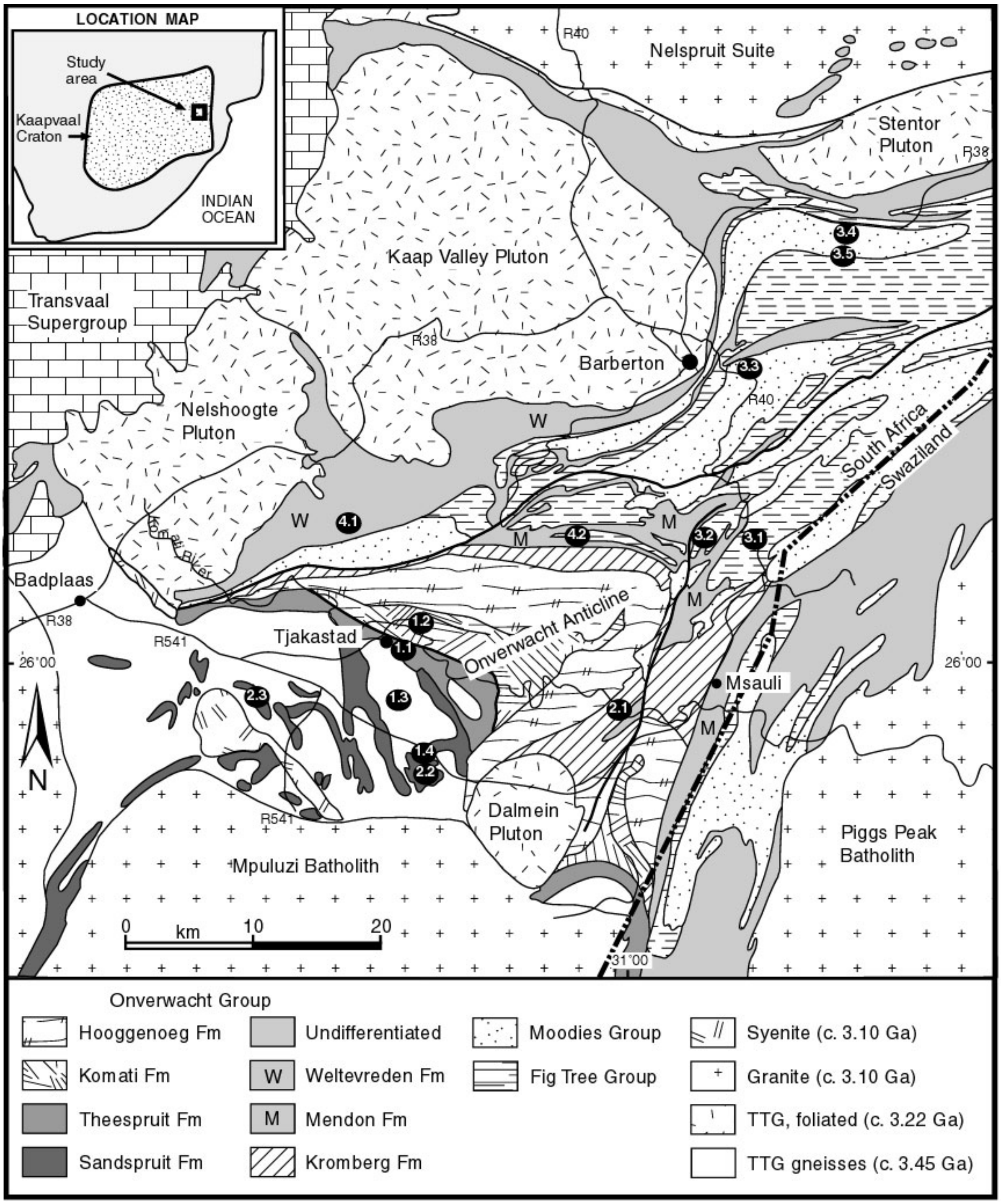


Figure 1: *Simplified geological map of the western and southern part of the Barberton greenstone belt and surrounding granitic terrane showing the excursion route and principal stops planned to show some of the key exposures relating to the theme of the Field Forum.*

THE GEOLOGY OF THE BARBERTON GREENSTONE BELT

Introduction

Greenstone belts are elongate outcrops of volcano-sedimentary sequences that are wedged in between granitoid-gneiss domes. They are a characteristic feature of the Archaean and contain much of the Earth´s mineral wealth (De Wit and Ashwal, 1997). Greenstone belts are dominated by thick successions of commonly pillowed submarine basalts that are complexly intercalated with ultramafic to felsic volcanic and volcaniclastic rocks and siliciclastic and chemical sediments. These sequences provide the only clues to early atmospheric and hydrospheric processes and to the origin of life on Earth. The thickness of the supracrustal successions is typically several kilometres, and the oldest and youngest rocks can be separated by hundreds of millions of years in age, suggesting a complex, multistage tectono-magmatic evolution. Evidence for polyphase, mainly compressional deformation associated with greenschist to amphibolite facies metamorphism is widespread. The evolutionary history of Archaean greenstone belts is controversial, and, to date, no tectonic model has been agreed upon (papers in De Wit and Ashwal, 1997).

The Barberton greenstone belt (Figs. 1, 2) is one of the key belts for greenstone studies and represents a type locality of mid-Archaean supracrustal sequences. The belt consists of a NE-SW striking succession of supracrustal rocks, termed the Swaziland Supergroup, which ranges in age from *c.* 3550 to 3220 Ma. The belt has a strike length of *c.* 130 km, width of 10-35 km, and an approximate depth of 4-5 km, and is surrounded by granitoid domes and intrusive sheets, ranging in age from *c.* 3500 to 3100 Ma.

Stratigraphy

The volcano-sedimentary sequence of the Barberton greenstone belt (Fig. 2) has been subdivided into three stratigraphic units, the Onverwacht, Fig Tree and Moodies Groups (SACS, 1980). The Onverwacht Group formed between 3550 and 3300 Ma and consists predominantly of ultramafic and mafic volcanic rocks (komatiites, komatiitic basalts, basalts), with minor felsic volcanic and sedimentary rocks that formed in a deep to shallow marine environment. Ultramafic-mafic igneous complexes also occur. The Fig Tree Group, 3260-3230 Ma in age, consists of deep to shallow marine sandstone and shale with minor jaspilitic banded iron formation and felsic volcanic rocks. The Moodies Group was deposited at *c.* 3227 Ma ago and consists of shallow-marine to fluvial sandstone and conglomerate with minor shale and banded iron formation.

Onverwacht Group

South of the Inyoka Fault, the Onverwacht Group has been subdivided into six formations, the Sandspruit, Theespruit, Komati, Hooggenoeg, Kromberg and Mendon/Swartkoppie Formations (Viljoen and Viljoen, 1969a; Lowe and Byerly, 1999). The formations are best developed in the southwestern part of the belt northeast of Tjakastad (Fig. 3). Metamorphic grade is mainly greenschist facies, but locally reaches amphibolite facies close to the contact with the surrounding granitoid domes, in particular in the Sandspruit and Theespruit Formations. Onverwacht Group rocks north of the Inyoka Fault have been grouped together as the Weltevreden Formation (Lowe and Byerly, 1999). Differences in the stratigraphy, ages and depositional environments of rocks north and south of the Inyoka Fault indicate that the fault zone represents a tectonostratigraphic boundary.

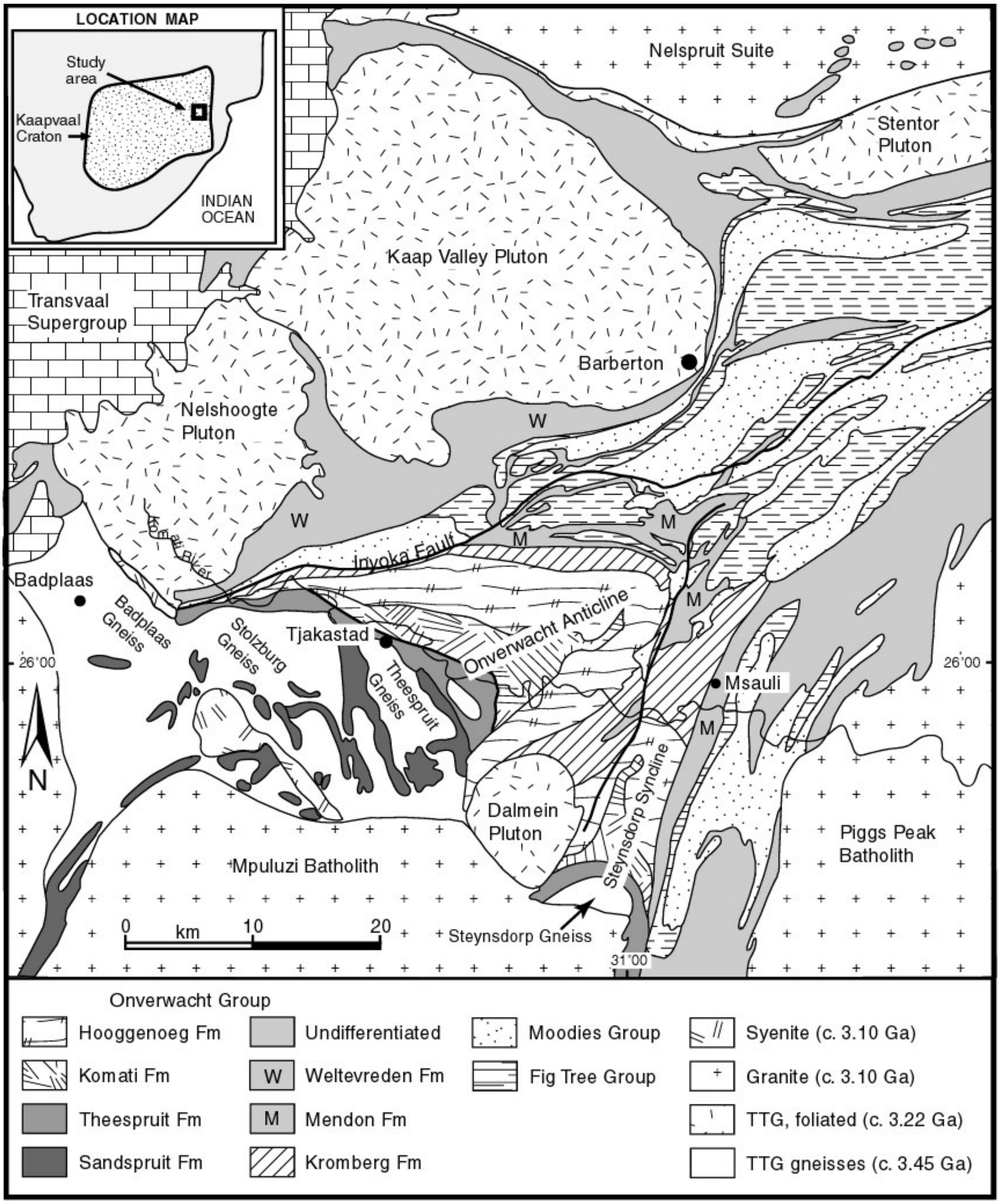


Figure 2: *Geological map of the Barberton greenstone belt and immediately surrounding granitic terrane (modified after Anhaeusser and Robb, 1980; Kamo and Davis, 1994; Lowe and Byerly, 2003).*

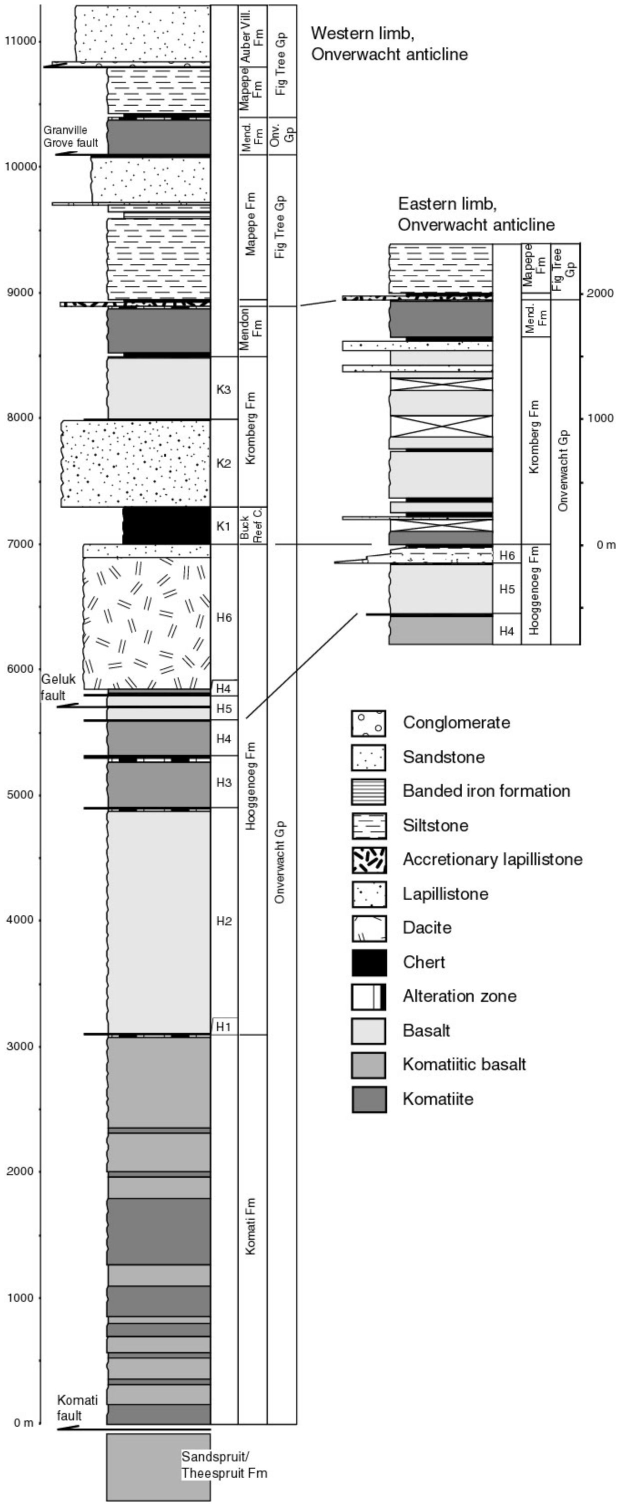


Figure 3: *Stratigraphic logs of the western and eastern limbs of the Onverwacht Anticline, southwestern part of the Barberton greenstone belt (mainly after Lowe and Byerly, 1999).*

Sandspruit Formation *(c. 2100 m)* –The Sandspruit Formation (Fig. 2) is represented by large rafts or xenoliths between or within TTG plutons. The formation consists largely of deformed and metamorphosed ultramafic rocks (serpentinite, talc schist) and metabasalt, now amphibolite. Thin metasedimentary layers are present locally. The stratigraphic relationship to the Theespruit Formation is unclear.

Theespruit Formation *(c. 1900 m)* –The Theespruit Formation (Fig. 2) consists predominantly of basalt (locally pillowed), komatiitic basalt, altered felsic volcanic and volcaniclastic rocks, and their deformed equivalents. Thin layers of banded black chert are present locally. Tectonic slices of tonalitic gneiss, dated at 3538 Ma (Armstrong et al., 1990), are tectonically intercalated with the supracrustal rocks. The Theespruit Formation is separated from the bulk of the greenstone belt by a high-strain zone, the Komati Fault, along which highly sheared mafic and ultramafic rocks occur. The oldest dated supracrustal rocks of the Barberton belt are schistose felsic volcanic rocks in the Steynsdorp Anticline that are attributed to the Theespruit Formation (3547 Ma, Kröner et al., 1996).

Komati Formation *(c. 3100 m)* –The Komati Formation (*c*. 3480 Ma, Armstrong et al., 1990) consists of spinifex-textured komatiite, komatiitic basalt and pillowed and massive basalt (Figs. 2, 3). Komatiites and komatiitic basalts are common in the lower part, whereas komatiitic basalts predominate in the upper part of the formation (Dann, 2000). Interflow sedimentary layers are absent, suggesting high eruption rates. The Komati Formation is the type locality for komatiites, which were first described by Viljoen and Viljoen (1969b).

Middle Marker *(1-10 m)* –The Middle Marker is a regionally extensive sedimentary horizon (4-5 m average thickness) that separates the Komati Formation from the Hooggenoeg Formation. It consists of silicified cross-bedded volcaniclastic sandstone, including accretionary lapilli, and green and black chert (Lanier and Lowe, 1982). Volcanic rocks directly underlying the marker bed are strongly altered. The marker horizon is the locus of shearing and intrusion of feldspar porphyries. Zircons from the Middle Marker have been dated at 3472±5 Ma (Armstrong et al., 1990).

Hooggenoeg Formation *(c. 3900 m)* –The Hooggenoeg Formation (Figs. 2, 3) consists of pillowed and massive basalt, spinifex-textured komatiitic basalt, thin silicified sedimentary horizons, and, at the top of the sequence, dacitic volcanic rocks and an epiclastic sedimentary unit. The volcanic rocks have been dated at 3445 Ma (Armstrong et al., 1990). Lowe and Byerly (1999) subdivided the succession into several stratigraphic units (H2-H5), each of which is represented by a mafic volcanic interval (H2v-H5v) capped by a 2-20 m-thick chert horizon (H2c-H5c). The Middle Marker is regarded as H1, and dacitic volcanic and volcaniclastic rocks are denoted as H6. Each chert horizon is underlain by a metasomatic alteration zone that is characterized by silicification, the presence of chromium-bearing sericite (subsequently denoted as fuchsite), stratiform chert, quartz and carbonate veins, and, locally, dykes of black chert. Chert H4c contains spherule layer S1, interpreted to represent quenched liquid silicate droplets of large meteorite impact origin (Lowe et al., 2003). Along the eastern limb of the Onverwacht Anticline, H6 is represented by a fining-upward sedimentary sequence of dacite-clast conglomerate and turbiditic sandstone; intrusive rocks are absent.

Buck Reef Chert *(c. 350 m)* – Along the western limb of the Onverwacht Anticline the Buck Reef Chert overlies Hooggenoeg Formation volcaniclastic sandstones (H6) along a gradational contact (Fig. 3). It is a homogeneous sequence of black-and-white banded chert and banded ferruginous chert cut by dykes and sills of ultramafic and mafic volcanic rock. The lowermost 5-40 m of the Buck Reef Chert consists of silicified shallow-water sediments and evaporites represented by pseudomorphs after nahcolite (Lowe and Fisher Worrell, 1999). A felsic tuff at the base of the Buck Reef Chert has been dated at 3416±5 Ma (Kröner et al., 1991).

Kromberg Formation *(c. 1700 m)* –The Kromberg Formation consists of pillowed and massive basalt, komatiite, mafic lapillistone and thin horizons of banded black chert (Fig. 3). The formation is capped by a horizon of black-and-white banded chert. A *c.* 700 m sequence of partly cross-bedded lapillistone of mafic to ultramafic composition occurs at the base of the sequence along the western limb of the Onverwacht Anticline. Lowe and Byerly (1999) regarded the Buck Reef Chert as the lowermost part of the Kromberg Formation. The lower contact of the volcanic Kromberg sequence is locally sheared and intruded by ultramafic sills (Hofmann, unpubl. data), whereas at one locality the contact between the Buck Reef Chert and Kromberg lapillistone is unconformable (Ransom et al., 1999). A felsic tuff associated with the capping chert has been dated at 3334±3 Ma (Byerly et al., 1996).

Mendon Formation *(c. 400 m)* –The Mendon Formation (*c*. 3298 Ma, Byerly et al., 1996) consists of komatiites that are overlain by a *c.* 20 m-thick unit of black-and-white banded chert and/or massive black chert (Fig. 3). Beds of silicified accretionary lapilli of komatiitic composition (Msauli Chert, *c.* 20 m thick) underlie the chert unit locally. Komatiites below the silicified sedimentary horizons are strongly altered to a chert-carbonate-fuchsite rock, commonly showing a gneissic fabric, which may be transected by dykes of black chert.
In the older literature, the Swartkoppie Formation represents the uppermost unit of the Onverwacht Group and groups together a variety of problematic rocks, including cherts and

altered and sheared ultramafic rocks. This unit is probably equivalent to the upper part of the Mendon Formation.

Weltevreden Formation **–** Onverwacht Group rocks north of the Inyoka Fault have recently been termed Weltevreden Formation (Lowe and Byerly, 1999), because they cannot be unambiguously correlated with Onverwacht Group rocks south of the fault. The stratigraphic position of Weltevreden rocks (Fig. 2) below Fig Tree sedimentary rocks may suggest correlation with the Kromberg and Mendon Formations. The unit consists of komatiite and komatiitic basalt, komatiitic tuff, minor basalt and black and black-and-white banded chert. Layered ultramafic intrusive complexes also occur.

Fig Tree Group

South of the Inyoka Fault the Fig Tree Group has been subdivided into the Mapepe and Auber Villiers Formations by Lowe and Byerly (1999). North of the Inyoka Fault four formations have been distinguished, the Ulundi, Sheba, Belvue Road and Schoongezicht Formations (e.g., Condie et al., 1970). The northern facies formed in a deep-water environment, whereas the southern facies formed in a deep- to shallow-water to alluvial environment.

Southern facies **–** The Mapepe Formation (Fig. 3) is several hundred metres thick and coarsens upward. The lower unit consists of shale with intercalated jaspilitic banded iron formation. The middle unit consists of tuffaceous shale and laminated felsic tuff, whereas chert-clast conglomerate interbedded with shale occurs in the upper unit. Sedimentary barite beds occur locally. Two distinct spherule horizons (S2, S3), interpreted to represent quenched liquid silicate droplets of meteorite impact origin (Lowe et al., 2003), occur at the base and in the middle unit. The Mapepe Formation was deposited in a variety of sedimentary environments, ranging from deep- to shallow-water, fan delta and alluvial environments (Lowe and Nocita, 1999). Zircon dates from the Mapepe Formation range from 3260 to 3230 Ma (Kröner et al., 1991).

The Auber Villiers Formation (Fig. 3) is approximately 1 km thick and consists of dacitic, plagioclase-phyric volcaniclastic rocks in the lower part, dacitic tuff, tuffaceous turbidites and chert-clast conglomerate in the middle part, and massive tuff and tuffaceous sandstone in the upper part.

Northern facies **–**The Ulundi Formation is tens of metres thick and occurs between black chert at the top of the Onverwacht Group and sandstones of the Sheba Formation. It consists of carbonaceous and pyritic shale, chert and jaspilite. Spherule bed S3 (Lowe et al., 2003) locally occurs at the base of the formation. The Sheba Formation is *c*. 1-2 km thick and consists of coarse-grained turbiditic sandstones with minor siltstone and shale interbeds. The Belvue Road Formation is several hundred metres thick and consists mostly of carbonaceous shale with minor turbiditic sandstone intercalations. Banded ferruginous chert occurs at the base of the sequence. Kohler and Anhaeusser (2002) have recently introduced a new stratigraphic unit, the Bien Venue Formation, which overlies the Belvue Road Formation in the northeastern part of the greenstone belt. The succession (0.7-3 km thick) consists of quartz-muscovite schist derived from dacitic to rhyodacitic volcaniclastic protoliths dated at *c*. 3256 Ma. Subordinate rock types include banded chert, phyllite, and biotite-plagioclase and chlorite schists, derived, in turn, from dacitic and basaltic precursors. The Schoongezicht Formation, which is *c*. 30 Ma younger than the Bien Venue Formation, overlies the Belvue Road Formation in the central and northwestern part of the greenstone belt, where it is several hundred metres thick and coarsens upward. It consists of plagioclase-rich turbidites intercalated with shale at the base and cross-bedded volcaniclastic sandstones and dacite clast conglomerates at the top. Intercalated felsic volcanic rocks have been dated at 3226±1 Ma (Kamo and Davis, 1994).

Moodies Group

The Moodies Group (Fig. 2) refers to quartz-rich (*c*. 50 vol.%), predominantly arenaceous rocks in contrast to the quartz-poor Fig Tree sandstones. A conglomerate horizon frequently forms the base of the Moodies Group. The Moodies Group is approximately 3 km thick and has been subdivided by Anhaeusser (1976a) into three formations, the Clutha, Joe´s Luck and Baviaanskop Formations. Each formation is a fining-upward sequence ranging from conglomerate or pebbly quartzose sandstone at the base to a thick sandstone unit to capping siltstone and shale. An amygdaloidal basalt flow overlain by shale and jaspilite occurs in the Joe´s Luck Formation. Moodies Group rocks south of the Inyoka Fault are generally more proximal in character (increase in average grain size, decrease of quantity/thickness of shale horizons). Braided-alluvial facies types predominate in the Moodies Group (Eriksson, 1978), but locally, in the northern part of the greenstone belt, the oldest record of tides on Earth is preserved in sandstones and mudstones of the upper Clutha Formation (Eriksson, 1977a; Eriksson and Simpson, 2000). The Moodies Group was deposited prior to the emplacement of the Kaap Valley tonalite pluton, which has been dated at 3227±1 Ma (Kamo and Davis, 1994).

Associated granitoids

The Barberton greenstone belt was intruded by a variety of granitoids (Fig. 2) during several episodes of magmatism. The granitoids with an age range from 3500 to 3200 Ma belong to the tonalite, trondhjemite, granodiorite (TTG) suite and have a prominent gneissic fabric, whereas the younger, *c*. 3100 Ma granitoids, are potassium-rich and form prominent sheets. Pre-Onverwacht Group tonalitic gneisses (Ancient Gneiss Complex) occur southeast of the greenstone belt in Swaziland and have been dated at 3644 Ma (Compston and Kröner, 1988). Prominent granitoids and their approximate emplacement ages are summarized in Table 1.

The emplacement of the plutons was shown to have occurred during four major magmatic episodes, at *c*. 3500, 3445, 3230 and 3100 Ma. The oldest pluton identified so far is the Steynsdorp pluton at the southern margin of the greenstone belt (*c*. 3510 Ma, Table 1). Available age data for the Doornhoek, Theespruit and Stolzburg plutons indicate that the majority of these trondhjemite plutons at the southern margin of the belt were emplaced during a relatively short period of time at *c*. 3445 Ma (Table 1). In contrast, the tonalite and trondhjemite plutons at the northern margin of the belt record ages at least 200 million years younger than those exposed at the southern margin. The Kaap Valley pluton, the only tonalite pluton in the area, has been dated at 3227 ± 1 Ma (Kamo and Davis, 1994). A similar age of 3236 ± 1 Ma was established for the trondhjemitic Nelshoogte pluton (De Ronde and Kamo, 2000). Age estimates on the Stentor trondhjemite pluton (Fig. 1), which has been dated at 3107 ± 5 Ma, indicate an even younger pulse of sodium-rich magmatism along the northern margin of the Barberton greenstone belt (Kamo and Davis, 1994). Age estimates on potassium-rich granites and granodiorites, including the Nelspruit and Mpuluzi batholiths as well as the Boesmanskop and Salisbury Kop plutons indicate identical emplacement ages between *c*. 3105 and 3107 Ma (Kamo and Davis, 1994, Table 1).

Table 1. Estimated U-Pb ages for the major plutonic rocks in the Barberton Mountain Land, South Africa

Pluton	Description of sample analysed	Mineral phase(s) analysed	Age (Ma)	*Reference
1) Steynsdorp	Trondhjemitic gneiss	Zircon	3509 + 8/-7	a)
	Trondhjemite	Zircon	3509 ± 4	d)
2) Doornhoek	Undeformed trondhjemite	Zircon, monazite	3448 ± 4	a)
3) Theespruit	Massive trondhjemite	Zircon, sphene	3443 + 4/-3	a)
	Trondhjemite	Zircon	3437 ± 6	b)
	Banded and foliated tonalitic gneiss	Zircon	3440 ± 5	e)
4) Stolzburg	Foliated trondhjemitic gneiss	Zircon	3459 +35/-23	a)
	Foliated trondhjemitic gneiss	Zircon	3445 ± 4	e)
4) Nelshoogte	Foliated trondhjemite	Zircon	3236 ± 1	c)
5) Kaap Valley	Hornblende tonalite	Zircon, sphene	3227 ± 1	a)
	Hornblende tonalite	Zircon	3226 ± 14	b)
6) Dalmein	Massive quartz-monzonite	Sphene	3215 ± 2	a)
7) Stentor	Heterogeneous granodiorite	Zircon	3107 ± 5	a)
8) Salisbury Kop	Granodiorite	Zircon	3105 ± 3	a)
9) Mpuluzi	Granodiorite	Zircon, sphene	3107 + 4/-2	a)
10) Nelspruit	Porphyritic potassic granite	Zircon, sphene	3106 + 4/-3	a)
11) Boesmanskop	Syenite	Zircon, sphene	3107 ± 2	a)

*a) Kamo and Davis (1994), b) Armstrong et al. (1990), c) De Ronde and Kamo (2000), d) Kröner et al. (1996), e) Kröner et al. (1991). The minerals listed here were all interpreted to be of magmatic origin, possible metamorphic ages have not been included. Errors are given as 2σ, except for b) (1 σ).

Tectonic Evolution

Early studies regarded the Barberton greenstone belt as a relatively simple synclinorium with a layer-cake stratigraphy. In this model, Onverwacht Group rocks in the northern and southern parts of the belt were thought to represent laterally correlative units. The surrounding granitoids were considered separate intrusive entities into the greenstone belt, lacking any connection to the greenstone sequence. Crustal shortening was considered to have affected the combined stratigraphy of the Onverwacht, Fig Tree and Moodies Groups in basically one major compressional episode, involving upright folding and steep thrusting, possibly as a result of diapiric granite emplacement (Visser, 1956; Ramsay, 1963; Anhaeusser, 1975).

A different tectonic scenario, consisting of a combination of early subhorizontal thrusting and tectonic stacking, followed by later upright folding, was proposed by Jackson et al. (1987), and De Wit and co-workers. According to De Ronde and De Wit (1994), the earliest, well-recognized tectonothermal events (3490-3450 Ma) in the evolution of the Barberton greenstone belt represent mid-ocean ridge-like processes (formation of the Komati and Hooggenoeg Formations) with seafloor-style metamorphism, the formation of chert and ubiquitous hydration and metasomatism of the mafic rocks (De Wit and Hart, 1993). These were followed by two periods of arc-related and trench-related processes, separated by 160 Ma.

The first period (3453-3416 Ma) recorded island-arc processes, as indicated by felsic magmatism, and the earliest regional compressional deformation (D_1), including the formation of layer-parallel shear zones, recumbent folds, inverted stratigraphy and the emplacement of possible ophiolite sequences (De Ronde and De Wit, 1994). North-directed thrusting during D_1 saw stacking of ophiolitic allochthons and juxtaposition of unrelated tectonic domains, such as felsic volcanic and volcaniclastic sequences (Theespruit Formation) and mafic rocks of the Kromberg Formation. Coeval granitoid magmas intruded along the thrust zones within a fore-arc or back-arc environment (De Ronde and De Wit, 1994). Tectonic slices of older tonalitic gneiss (3538 Ma) in the Theespruit Formation signify the involvement of older continental crust during D_1.

The second period (3259-3222 Ma) recorded intra-arc and inter-arc-related processes, culminating in the amalgamation of the northern and southern part of the greenstone belt along the Saddleback-Inyoka Fault system (De Ronde and De Wit, 1994). Calc-alkaline felsic volcanics and classical turbidites in an upward coarsening succession make up most of the Fig Tree Group. The sediments have been interpreted as fan delta or marine sediments in a fore-arc or back-arc basin (*cf.* Lamb and Paris, 1988). Accretion-like convergent processes dominated between 3230 and 3080 Ma. During this period two coaxial deformation events occurred, D_2 and D_3. D_2 (*c.* 3227 Ma) caused juxtaposition of units along thrust zones, tight folding, and the syntectonic deposition of coarse clastic rocks (sandstone, conglomerate) in an emerging marine to subaerial environment (e.g. Lamb and Paris, 1988), possibly in piggy-back sedimentary basins riding on thrust sheets (Lamb, 1984). Early D_3 produced NE-SW striking, open syncline/tight anticline pairs with related thrust and strike-slip components and is regarded as a continuum of D_2. Late D_3 shear zones, associated with gold mineralization, have been dated between 3130 and 3080 Ma (De Ronde and De Wit, 1994).

A competing tectonic model for the evolution of the Barberton greenstone belt was developed by Lowe (1994, 1999a). In this model, felsic volcanic rocks of the Theespruit Formation formed in a volcanic arc at 3510 Ma and were intruded by the Steynsdorp pluton to form the Steynsdorp protocontinental block. Rifting along the margins of the block resulted in the formation of the ultramafic and mafic volcanic rocks of the Komati and Hooggenoeg Formations. A second subduction-related arc system during 3440 Ma resulted in the deposition of dacitic volcanic rocks near the top of the Hooggenoeg Formation and the intrusion of the Theespruit/Stolzburg tonalite plutons to form the Songimvelo protocontinental block. Subsequent rifting resulted in the eruption of Kromberg Formation basalts. A break in volcanic activity was followed by the development of a third subduction-related volcanic arc system, represented by 3260-3230 Ma volcanic rocks in the southern facies of the Fig Tree Group. Arc magmatism was coeval with a thrusting event, during which most of the Fig Tree and Moodies sedimentary rocks were deposited in fore-arc and foreland basins. Continued compression resulted in the incorporation of the sedimentary units in the Barberton fold-and-thrust-belt.

Evidence for Life
(by Frances Westall, CNRS, Orleans, France)

Sedimentary chert horizons in the Onverwacht and Fig Tree Groups have been studied since the 1960s for microfossils and carbon isotope ratios. Most of the early finds of microfossils were dismissed as artifacts or dubiofossils (Schopf and Walter, 1983; Schopf, 1993). However, stratiform and domical microbial mats were described by Byerly et al. (1986), Walsh (1992), Westall and Gerneke (1998), Walsh and Lowe (1999), Westall and Walsh (2000), Westall et al. (2001), Walsh and Westall (2003), and Westall (2003, 2004) in cherts having carbon isotope ratios ranging from -14 to -40 ‰ (Hayes et al., 1983; Schidlowski et al., 1983; Walsh, 1992; Walsh and Lowe, 1999; Westall et al., 2001; Westall et al., unpubl. data). These mats may be visible macroscopically in outcrop (e.g. the Byerly et al., 1986 sites). Although some rare, relatively large microbial filaments (>1 µm in diameter and tens to

>100 μm in length) associated with the mats can be identified in thin section (Walsh, 1992; Westall and Walsh, 2000; Westall, 2004), most individual microfossils are <1 μm in size and have been identified by electron microscope techniques (Westall and Gerneke, 1998; Westall and Walsh, 2000; Westall et al., 2001; Westall, 2003, 2004). Microorganism morphologies include filaments, short rods, coccoids and vibroids that generally occur in colonial associations together with copious quantities of polymer (extracellular polymeric substances, EPS). The colonies formed mats at the surfaces of the volcaniclastic sediments deposited in shallow water to subaerial, evaporitic conditions. Moreover, evidence for microbial corrosion of pillow lava rinds is described by Furnes et al. (2004). Detrital fragments of mats occur in deeper water deposits.

Recent discussions of microfossil remains from similarly old cherts of the Pilbara Craton in Australia (Schopf, 1993; Brasier et al., 2002; Schopf et al., 2002) have raised questions regarding the morphological identification of fossil microorganisms in rocks of this age. The existence of natural and laboratory-produced siliceous and carbonaceous bacteriomorph structures makes identification of microbial fossils even more complicated (Yushkin, 1999; Garcia-Ruiz et al., 2003). Hopefully, investigations presently underway by various groups on the cherts from the Barberton and Pilbara regions will resolve the problem.

FIELD FORUM EXCURSION DETAILS

DAY ONE, SUNDAY, 4 JULY 2004

Guide: Carl R. Anhaeusser

Overview: The drive east from Badplaas on R541 (Fig. 1) takes the excursion across generally flat-lying granitic terrane made up almost exclusively of trondhjemitic gneisses of the Badplaas and Stolzburg plutons (Fig. 2). Approximately 10 km east of Badplaas the road crosses a shear zone, followed by a zone of north-south-trending greenstone remnants and associated migmatites that separate the Badplaas from the Stolzburg pluton, the latter having an age of *c*. 3445 Ma (Kamo and Davis, 1994).

En route the terrane seen to the north consists of the mountainous Barberton greenstone belt, while to the south the Boesmanskop syenite pluton, dated at *c*. 3107 Ma (Kamo and Davis, 1994) forms a prominent feature of the landscape. The road northeast to Tjakastad winds its way across the Stolzburg pluton and eventually crosses the north-south-trending Tjakastad schist belt, which exposes schistose mafic, ultramafic and felsic rocks of the Theespruit Formation. The Theespruit Formation was originally defined as one of three lower formations making up the Tjakastad Subgroup of the Onverwacht Group (Viljoen and Viljoen, 1969a), the others being the Sandspruit Formation at the base and the Komati Formation overlying the Theespruit Formation. However, it has to be kept in mind that the Komati Formation is separated from the lower units by a major shear zone, the Komati Fault.

The road crosses the Tjakastad schist belt and emerges onto the northwest rim of the Theespruit pluton dated at *c*. 3443 Ma by Kamo and Davis (1994). It also cuts across a prominent Proterozoic-aged dyke, known locally as the Swartrand (Black Ridge) dyke, which trends in a NNW-SSE direction traversing the southern part of the Tjakastad township. The Tjakastad village partly overlies the trondhjemite gneisses of the Theespruit pluton in the south and rocks of the Theespruit Formation in the north. The excursion heads towards two water reservoirs seen on the hill immediately above the town (Fig. 4). This lunch stop will be combined with the first geological stop of the afternoon.

Stop 1.1. Tjakastad water reservoirs exposures of the Theespruit Formation (S25°59.456′, E30º49.500’) and vantage point to view the geological setting of the type locality of the Onverwacht Group.

Objectives: The Tjakastad area provides the most complete geological section of the formations that make up the Onverwacht Group of the Barberton Supergroup (formerly Swaziland Supergroup). The early work in the area by Viljoen and Viljoen (1969a, b), led to the subdivision of the Onverwacht Group into six formations, the lower three (Sandspruit, Theespruit, Komati) making up the Tjakastad Subgroup and the upper three (Hooggenoeg, Kromberg and Swartkoppie) collectively forming the Geluk Subgroup. Subsequent work has suggested numerous changes to this scheme and the interested reader is referred to the summary by Lowe and Byerly (1999), who outline the more recent ideas on the geologic evolution of the Barberton greenstone belt.

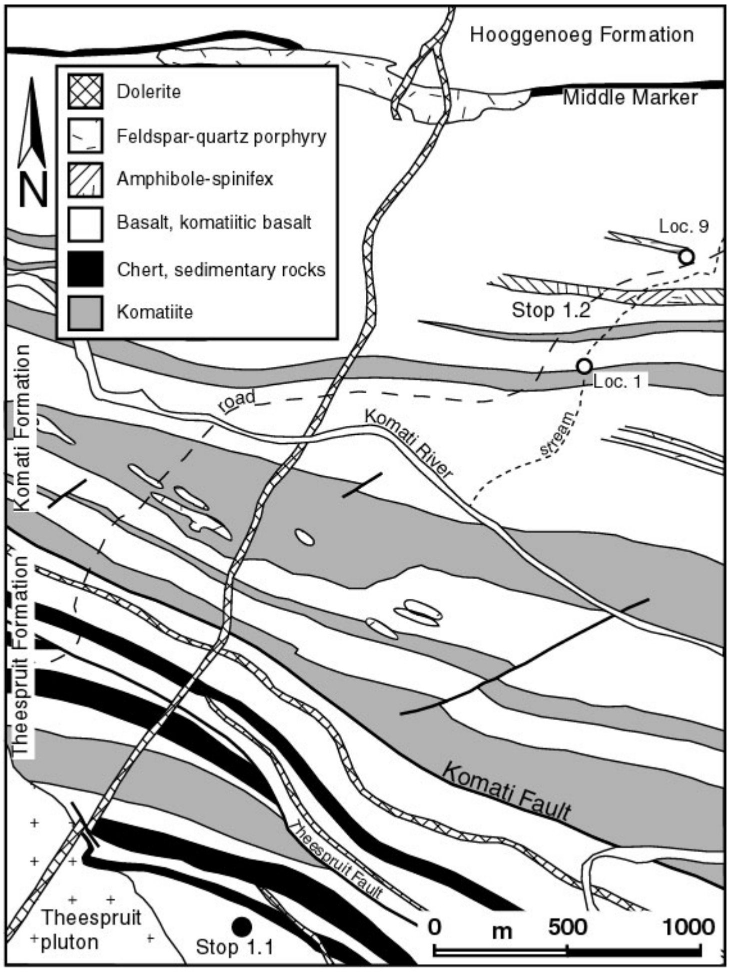


Figure 4: *Simplified geological map of the Theespruit and Komati Formations northeast of Tjakastad (modified after Viljoen and Viljoen, 1969b).*

The view northwards from the reservoirs stop shows the main type locality sections. In the immediate foreground and near the reservoirs are rocks of the Theespruit Formation (Fig. 4). Down the slope towards the Komati River a white quartz vein marks the trace of the Theespruit Fault, which can be seen trending east-west within the Theespruit succession. The Komati Fault near the river separates the Theespruit Formation from the structurally overlying Komati Formation. The stratigraphic layering of the type section may be seen looking to the northeast. Good exposures are present in a traverse over the crest of the hills, but the stream to the west (Spinifex Stream) has some of the best developed komatiite and komatiitic basalt flow features, including spinifex-textured komatiite, pillow lavas, pillow breccias, columnar joints, vesiculated pillows, pillows with spherulitic or ocelli structures, and pillows with re-entrant rims (see Cloete, 1999 and Dann, 2000 for more details).

The hills in the far distance represent the upper formations of the Onverwacht Group. From the reservoirs northwards (and including also the Sandspruit Formation to the south) the entire section, which consists of a subvertical succession of predominantly volcanic rocks, was estimated to be approximately 15 km thick by Viljoen and Viljoen (1969a).

Time will not permit a visit to the type section locality in the Spinifex Stream area on this excursion, but a short river section to the west provides a good alternative set of exposures of rocks and volcanic features typical of the Komati Formation. Following the examination of the exposures at the reservoir locality the plan is to undertake an approximately 1 km traverse northwards along the river section at stop 1.2 (Fig. 4).

The second objective at stop 1.1 is to examine one of the characteristic units of the Theespruit Formation, viz., one of the many interlayered felsic units that led Viljoen and Viljoen (1969b) to separate the Theespruit succession into an independently recognizable entity or formation. The type locality section of the Theespruit Formation is located approximately 2 km to the east of the reservoirs where it has a thickness of about 1890 m. The sequence consists of metamorphosed mafic lavas and tuffs (komatiitic basalts) and felsic interlayers. Some serpentinized ultramafic rocks occur mainly as pods and lenses, the latter having been formed as a result of flattening of the formations by the invading trondhjemites of the Theespruit pluton.

The felsic units are made up of coarse agglomerates and finer-grained tuffs. These rocks, which are commonly deformed to felsic schists, consist predominantly of quartz and sericite, and are often highly aluminous, containing variable amounts of pyrophyllite, andalusite, chloritoid and staurolite. Sillimanite may also be encountered close to granitoid intrusions. The felsic horizons are often immediately overlain by, or closely associated with, thin, impersistent bands and lenses of black carbonaceous and siliceous cherty sediments. The carbonaceous cherts reportedly host primitive algae-like fossil remains (Engel et al., 1968), but not all palaeobiologists are necessarily in agreement with this interpretation. For example, Schopf and Walter (1983) referred to many of the algae-like structures in the Onverwacht as being “dubiomicrofossils”.

Stop 1.2. Stream section northeast of Tjakastad displaying 3480 Ma komatiites and komatiitic basalts of the Komati Formation (S 25° 58.188’, E30° 50.065’).

Objectives: The stream section displays a number of excellent examples of the types of volcanic rocks, and their textures, found in the Komati Formation. The full succession of the Komati Formation, which was estimated to be about 3500 m thick (Viljoen and Viljoen, 1969b), occurs approximately 3 km to the east of this locality. The stream section to be traversed at stop 1.2 (Fig. 4) commences approximately 1.5 km from the base of the Komati Formation and extends up-section for another 1 km. Nine localities have been singled out for special attention on the traverse.

Locality 1. The exposures on the east bank of the stream show a series of komatiite lava flows which strike east-west and have a vertical dip. One flow in particular clearly demonstrates the unique textures first described in detail at Munro Township in Ontario, Canada (Fig. 5) by Pyke et al. (1973). The flow shows a basal zone consisting of a lower chill contact, an overlying cumulate olivine zone (B2-B4), followed by a foliated “hopper” olivine zone (B1) roughly halfway up the flow unit. This is overlain by a zone of bladed or plate spinifex-textured olivine together with some clinopyroxene (A2), which, in turn, is overlain by a zone of random spinifex (e.g. Fig. 6a) grading upwards into a fine-grained flow-top or chill-top breccia (A1). This internal zonation is repeated in all the komatiite flows in the Komati Formation, some in the type locality being only 20 cm thick, whilst others may be as much as 1.5 - 2 m thick. On average, the komatiite flows are generally about 1 m thick and the textures enable one to determine the northward way-up or younging direction.

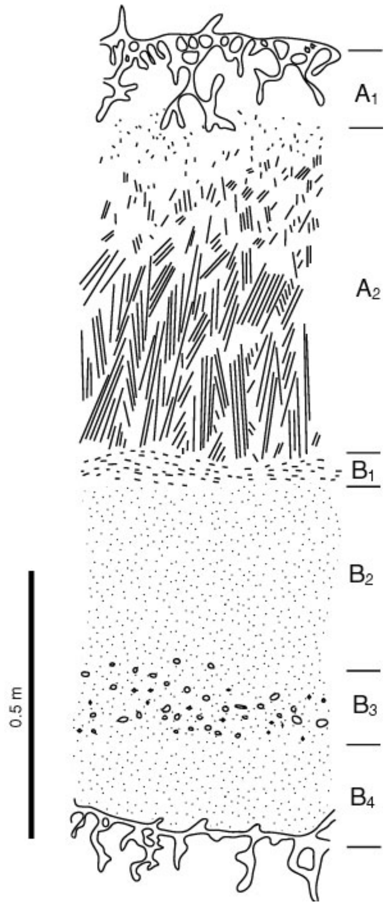


Figure 5*: Diagrammatic secton of a typical komatiite flow (after Pyke et al., 1973*).

Locality 2. Approximately 30m north of locality 1, on the west bank of the stream, are exposures of komatiitic pillow basalts and rubble-like pillow breccia material in a zone about 10 m wide. The pillow selvages are clearly evident and represent chilled rims formed as the lavas were erupted into a body of water.

Locality 3. About 50 m north of locality 2, on the east bank of the stream, are excellent exposures of komatiitic pillow basalts (Fig. 6b). The pillows can be seen both in plan view on the flat pavements and in section on a nearby cliff face. The pillows are bulbous, and show smooth curved upper surfaces and downward protuberances enabling confirmation of the younging direction to the north to be ascertained. The rocks have a typical green colour (greenschist metamorphic grade) and show very limited or hardly any affects of structural deformation.

Locality 4. Continue upstream on the east bank for an additional 100 m. Good exposures of massive mafic lava can be seen in the stream channel. Several flow units can be identified, with each flow terminating with zones of spherulitic structures or lighter coloured (bleached?) felsic lava. Some large spherulitic (ocelli) structures (Fig. 6c) occur within the grey-green basalts, particularly on the west bank, where some can be seen to range from golf-ball size to tennis-ball size. Cloete (1999) recorded spherulites over 15 cm in diameter at the type locality. The spherulitic structures are considered to represent immiscibility features and are commonly associated with pillow-flow transitions.

Locality 5. Continuing upstream on the east bank for approximately 400 m there are few exposures to be seen until the stream makes a bend. Large outcrops of locally deformed pillowed basalt and pillow breccia occur, with the pillows elongated in a subvertical orientation. Some pillows show quartz- and, in places, carbonate-filled 'drain-away cavities' (cavities formed when the lavas were horizontal and when the molten lava was able to flow or drain out of the pillows leaving behind tubular gas chambers), which are often stacked above one another in some of the larger pillows.

Locality 6. Another stroll for about 60 m upstream of locality 5 permits the examination of some spinifex-textured komatiitic basalt exposures. These differ from those seen earlier in the komatiite flows at locality 1 in that the blades and needles do not consist of olivine and clinopyroxene, but are rather made up of radiating, generally actinolitic, amphibole needles (Fig. 6d).

Locality 7. About 30 m northwards is an exposure in the stream that consists of blue-black serpentinized komatiite. The ultramafic rock has been altered to talcose serpentinite, probably by reaction with hydrothermal fluids making their way into the rock along fissures and veins that can be seen anastomosing throughout the exposure.

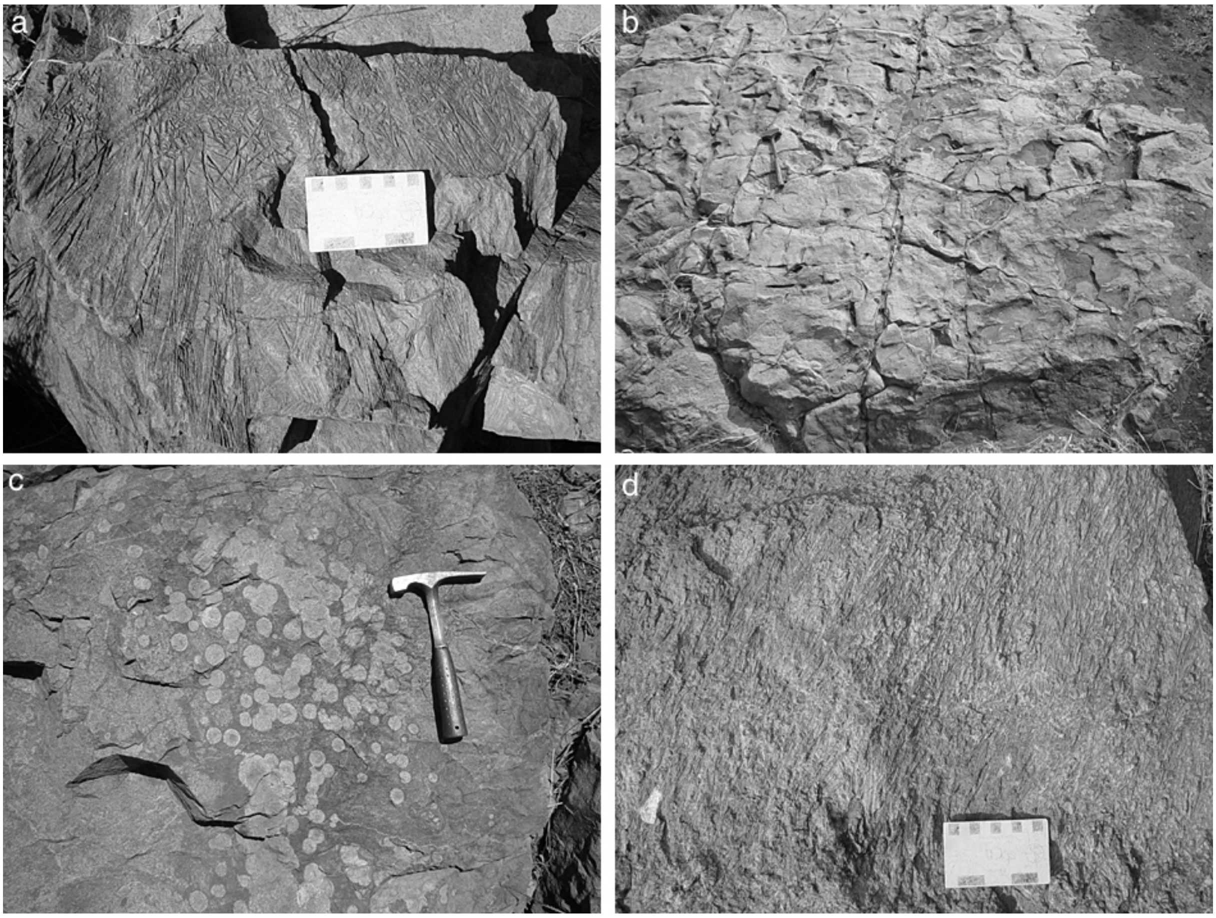


Figure 6 : *(a) Komatiite showing bladed spinifex-texture (base) and random spinifex-texture (top). Photograph from Spinifex Stream. (b) Rock pavement of komatiitic pillow basalt at locality 3. Hammer for scale. (c) Ocelli in massive basalt or komatiitic basalt. Photograph from Spinifex Stream. (d) Amphibole spinifex-texture in komatiitic basalt at locality 6.*

Locality 8. A short climb up the slope from locality 7 towards the roadway (and locality 9) crosses a narrow (30-40 cm wide), east-west trending, granitoid dyke. This granitoid rock is stratigraphically about 3.75 km from the nearest granite pluton (the Theespruit pluton) and is deemed to be unrelated to this body. The rock has a porphyritic texture and has an anomalously high Na_2O content (see Table 2). The favoured interpretation is that the dyke represents a plagiogranite similar to those reportedly encountered in modern oceanic settings, and more particularly with the plutonic components of ophiolites.

Locality 9. Approximately 50 m north of the plagiogranite dyke and immediately alongside the roadway are outcrops of komatiitic basalt that show the development of long, needle-like crystals of amphibole (actinolite-tremolite), some measuring upwards of 30 cm. The exposure has been disturbed during the construction of the road and the remains are not all *in situ*. The amphibole represents another variety of spinifex texture not too dissimilar to that seen earlier at locality 6. These rocks have MgO contents that range from about 18 to 23% and were initially subdivided by Viljoen and Viljoen (1969b) into what they termed 'Geluk-type' komatiitic basalts. They differ from the other komatiitic basalts they defined, which were referred to as 'Barberton-type' basalts (9-12% MgO) and 'Badplaas-type' basalts (14-17% MgO). This terminology was not accepted in later years as it was shown, as additional analyses became available, that the three fields are not separate entities, but form a continuum from high-Mg tholeiitic basalts to high-Mg komatiitic basalts.

From this locality, and looking northwards, it is possible to see the position of the Middle Marker chert horizon on the hillslopes above. Some gold exploration in the late 1880s was carried out along the more prospective chert horizons and an old dump seen cascading down the slope marks the position of one of the shafts. No mines were ever established in the region. The Middle Marker represents the approximate halfway point in the Onverwacht volcanic pile. The mountainous terrane beyond forms part of the Geluk Subgroup – a sequence approximately 7 km thick and consisting of additional volcanic and volcanoclastic rocks (as well as interlayered chemical sedimentary rocks) that have been likened to calc-alkaline volcano-sedimentary successions found more typically in modern-day island arc geotectonic settings.

The excursion then proceeds south through the Tjakastad village, crossing once again the Swartrand Dyke and continuing across the Theespruit pluton to the main Badplaas road (R541). The next stop will be made on the farm Aarnhemburg 155IT.

Table 2. Major and trace element comparison between trondhjemite gneiss from the Theespruit pluton and the plagiogranite dyke intruded into the Komati Formation on the farm Hooggegnoeg 7341JT

Wt%	Plagiogranite dyke (sample PD1)	Trondhjemite gneiss (sample BTP 12)
SiO_2	69.96	71.60
TiO_2	0.37	0.27
Al_2O_3	14.69	14.84
Fe_2O_3	2.86	2.41
MnO	0.04	0.02
MgO	0.86	1.19
CaO	0.64	2.52
Na_2O	7.49	5.00
K_2O	0.93	1.42
P_2O_5	0.14	0.09
LOI	0.71	0.58
ppm		
Rb	20	47
Sr	326	518
Y	10	7
Zr	128	99
Nb	11	8
Co	8	9
Ni	15	38
Cu	10	36
Zn	68	57
V	34	22
Cr	19	69
Ba	398	338

Unpublished data - C. R. Anhaeusser

Stop 1.3. Roof pendant in the Theespruit pluton (S26° 01.587’, E30° 50.230’) on the farm Aarnhemburg 155IT.

Objectives: Numerous greenstone remnants or xenoliths occur in the granitoid terrane surrounding the Barberton greenstone belt proper. Detailed mapping allows some of these remnants to be clearly linked to the main greenstone belt, but others are isolated or disconnected making it difficult to be precisely sure of the relationship to any particular formation. The absence of distinctive marker units makes this connection even more tenuous.

The structural history of the Theespruit pluton suggests that the granitoid body is a relatively shallow-seated occurrence and that many of the greenstone xenoliths found in the pluton have been stoped from the base of the overlying successions and act as roof pendants or greenstone rafts. Stop 1.3 provides a good example of one such roof pendant. However, its position relative to the surrounding greenstone terrane and the absence of any distinctive marker unit makes correlation difficult. The favoured view is that the komatiitic pillow basalts seen at the locality are part of the lowermost unit in the Barberton stratigraphic column (viz., the Sandspruit Formation).

Stop 1.3. The exposure to be visited is located on the eastern side of the Theespruit pluton, not too far from the centre of the body. The xenolith is approximately 100 m^2 in areal extent, is entirely surrounded by leuco-trondhjemitic gneisses, and is cut by a stream on the eastern side of the xenolith, which provides excellent exposures. The only greenstone rock type represented consists of hornblende amphibolite (derived from komatiitic basalt).

The traverse commences on the southern side of the xenolith where the contact consists of hornblende tonalite-trondhjemite intruded into pillow basalts. Proceeding northwards along the stream trondhjemite dykes can be seen intruded into the amphibolites and in places agmatitic textures are evident. Near the centre of the exposure numerous pillow structures are evident (Fig. 7), which show a range of internal features, including spherulitic or ocelli structures, vesicles, chilled margins and re-entrant rims. The pillows vary considerably in size, some as big as a football and others 1-2 m in length. Inter-pillow quartz and carbonate is evident, as are ‘drain-away’ cavities also partially or completely filled with quartz and carbonate. The most notable feature of the exposure is the high metamorphic grade of the rocks and the almost total absence of deformation. The roof pendant has, thus, escaped the influence of flattening strains that are invariably found close to the margins of invasive diapiric trondhjemite gneiss plutons, like those found elsewhere in the surrounding granitoid terrane. In other places on the Theespruit pluton xenolithic remnants occur that have undergone partial melting and metasomatic replacement to form a variety of dioritic to amphibolitic migmatites.

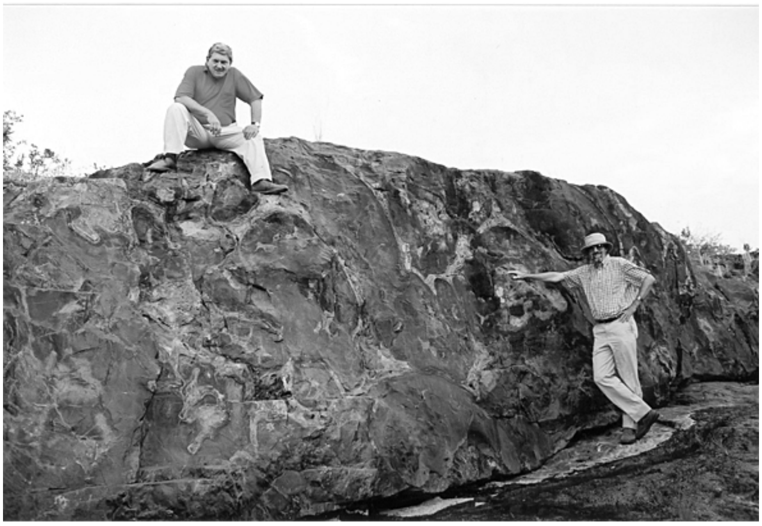


Figure 7: *Amphibolite facies komatiitic basalt with pillow structures, Theespruit pluton roof pendant on the farm Aarnhemburg 155IT.*

The trondhjemitic granitoid rocks surrounding the greenstone remnant are leucocratic biotite-quartz-plagioclase rock showing little or no gneissic fabric suggesting that the level of exposure, of what has been interpreted as a diapiric pluton, is close to the position of no finite strain in the crestal part of the pluton. Only nearer the pluton contacts with surrounding greenstones does a gneissic fabric make an appearance and the rocks display a prolate fabric which produces a vertical lineation.

Stop 1.4. Spinifex-textured ultramafic rocks of the Sandspruit Formation (3.5 km due south of Stop 1.3 on the farm Brandybal 171 IT).

Objectives: This locality provides relatively easy access to a zone of komatiite lavas that alternate with komatiitic basalts correlated with the lowermost unit of the Onverwacht Group, viz., the Sandspruit Formation. This correlation is presently under review (Anhaeusser, 2004), as unusual clastic metasedimentary rocks (similar to the 3525-3540 Ma metasediments described by Dziggel et al., 2002, and by Annika Dziggel in this guidebook) occur with the metavolcanic rocks. These new ages place this formation approximately 60-90 Ma older than the overlying Theespruit and Komati Formations and may suggest an earlier stage of greenstone belt development than that associated with the Onverwacht event.

Stop 1.4. The exposures in this area consist mainly of metabasaltic rocks and komatiites. Good examples of spinifex-textured komatiite flows are in evidence, and in places multiple flow sequences can be determined. Some of the khaki- to dark-green, massive rocks at this locality consist of high-magnesian komatiitic basalts with tremolite-actinolite as the principal amphibole. A short distance from the komatiitic rocks are black hornblende amphibolites showing well-formed pillows with distinct pillow rims and re-entrant structures as well as ocelli, gas vesicles, drain-away cavities and agmatitic textures. Some of the serpentinized komatiite (now a poor quality soapstone) is used by some local inhabitants for making ornamental carvings, and one such 'quarry' can be seen at this locality.

DAY TWO, MONDAY, 5 JULY 2004

Stop 2.1. Komati River section, Songimvelo Nature Reserve (S26°02.170', E30°59.980')

Guides: Axel Hofmann, Carl R. Anhaeusser

Objectives: A classic section of the uppermost Hooggenoeg Formation, Kromberg Formation type section, and lowermost Mendon Formation of the western limb of the Kromberg Syncline is exposed along the Komati River (Fig. 8). The section formed in a time span of *c.* 100 Ma and includes a wide variety of rock types, including dacite-clast conglomerates and turbidites, carbonaceous cherts, pillowed and massive basalt, and komatiitic basalt flow units, ultramafic flows or sills and lapillistone, and unusual alteration zones of volcanic rocks (Fig. 9). Some key issues include the presence of microfossils in some carbonaceous cherts, the origin of alteration zones as shear zones versus Archaean weathering phenomena, and the depositional and tectonic setting of the Kromberg Formation (Note: *The section is located within the Songimvelo Nature Reserve, and permission is required to visit these outcrops. Please avoid hammering of key outcrops*).

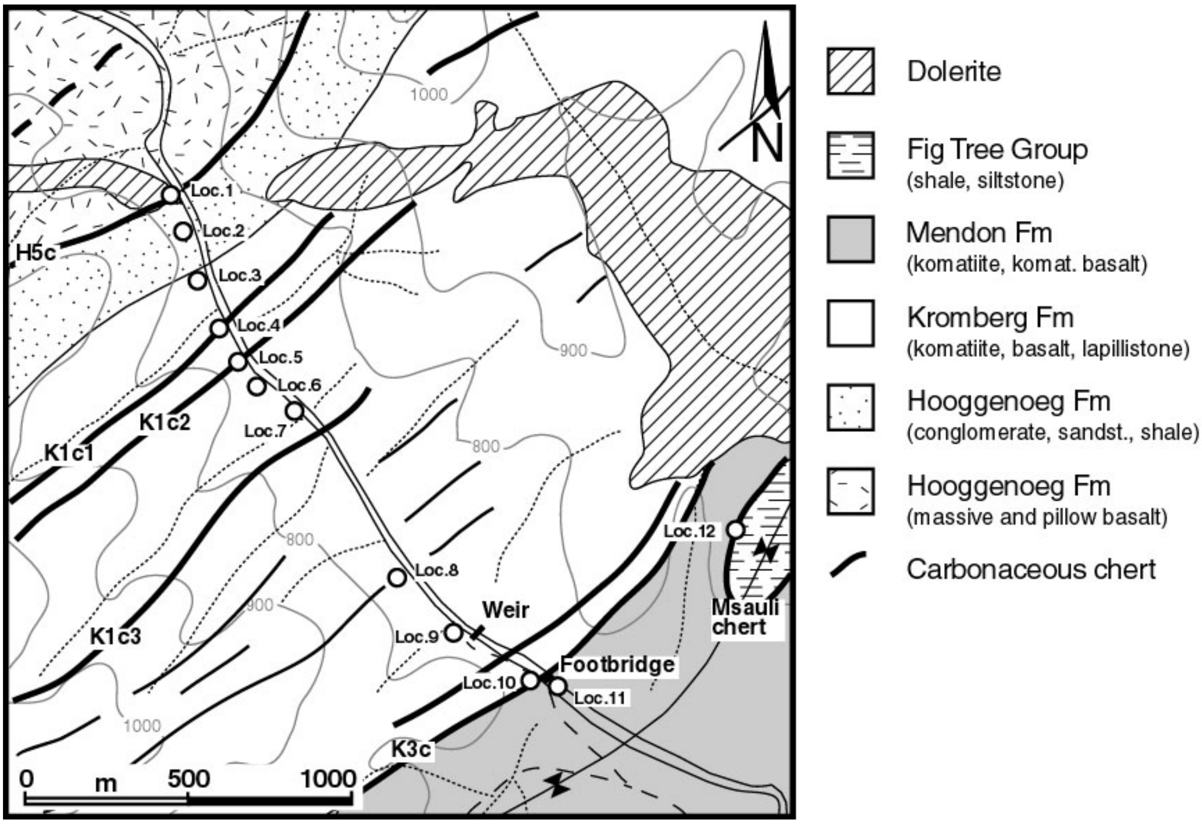


Figure 8: *Geological map of the Komati River section (modified after Viljoen and Viljoen, 1969c).*

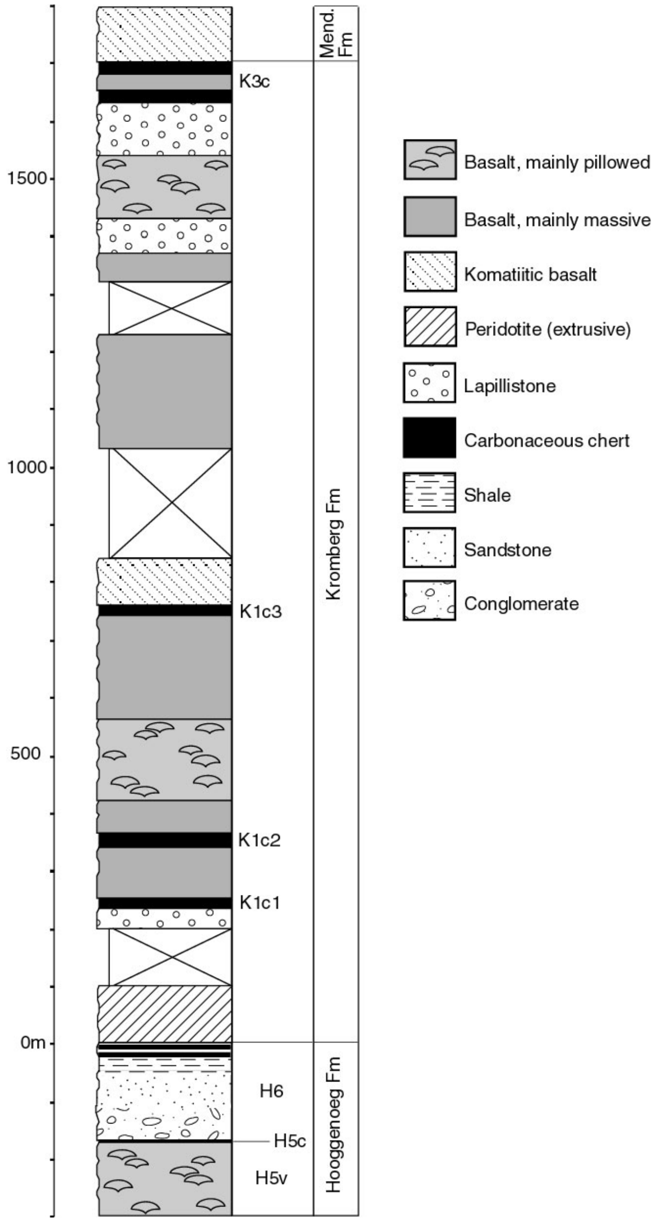


Figure 9: *Stratigraphic log of the Komati River section (modified after Viljoen and Viljoen, 1969c; Lowe and Byerly, 2003).*

Locality 1. A sequence of pillow basalt and minor massive basalt, a few hundred metres thick, forms the uppermost part of the Hooggenoeg Formation. The basalt sequence is capped by a thin chert horizon (H5c of Lowe and Byerly, 1999). Pillow basalt containing abundant ocelli and, commencing from *c.* 50 m below the chert bed, becomes silicified upsection. Silicification is generally associated with a colour change from greenish grey to light grey. This silicification is associated with the replacement of igneous minerals by quartz, carbonate and sericite, and an increase in SiO_2 and K_2O (Viljoen and Viljoen, 1969c; Byerly and Lowe, 1991). Silicified basalt is transected by massive black chert veins in the uppermost few metres (Fig. 10a). Chert-veined basalt is capped along a sharp contact by a *c.* 1 m thick horizon (H5c) of massive to thinly laminated black chert that is, in turn, overlain by laminated grey chert (Fig. 10b). Grey chert contains normally graded laminae with accretionary lapilli. Microfossils in black chert have been reported from this horizon (Walsh and Lowe, 1985; Walsh, 1992). The chert horizon and the underlying chert dykes are best exposed on the northern river bank.

Locality 2. The chert horizon is overlain along a sharp and planar contact by a *c.* 170 m thick, upward-fining sedimentary sequence. This unit, termed member H6 of the Hooggenoeg Formation, has been correlated with dacitic volcanic rocks of the west limb of the Onverwacht Anticline (Lowe and Byerly, 1999). The sequence starts with massive, poorly sorted, cobble to boulder conglomerate (Fig. 10c). The clasts consist predominantly of silicified dacitic volcanic rocks with minor carbonated volcaniclastic sandstone, grey and black chert, and feldspar-porphyry, in a coarse-grained, carbonated sandstone matrix. Dacite clasts have been dated at 3445±3 Ma (Kröner et al., 1991). This age is identical to ages obtained from intrusive/extrusive dacitic rocks of H6 (Kröner and Todt, 1988; Armstrong et al., 1990). Massive conglomerate is overlain by very thick beds of normally graded and massive conglomerate and very coarse-grained sandstone, followed by massive and parallel-laminated sandstone with minor intercalations of pebble conglomerate. The upper part of the sequence

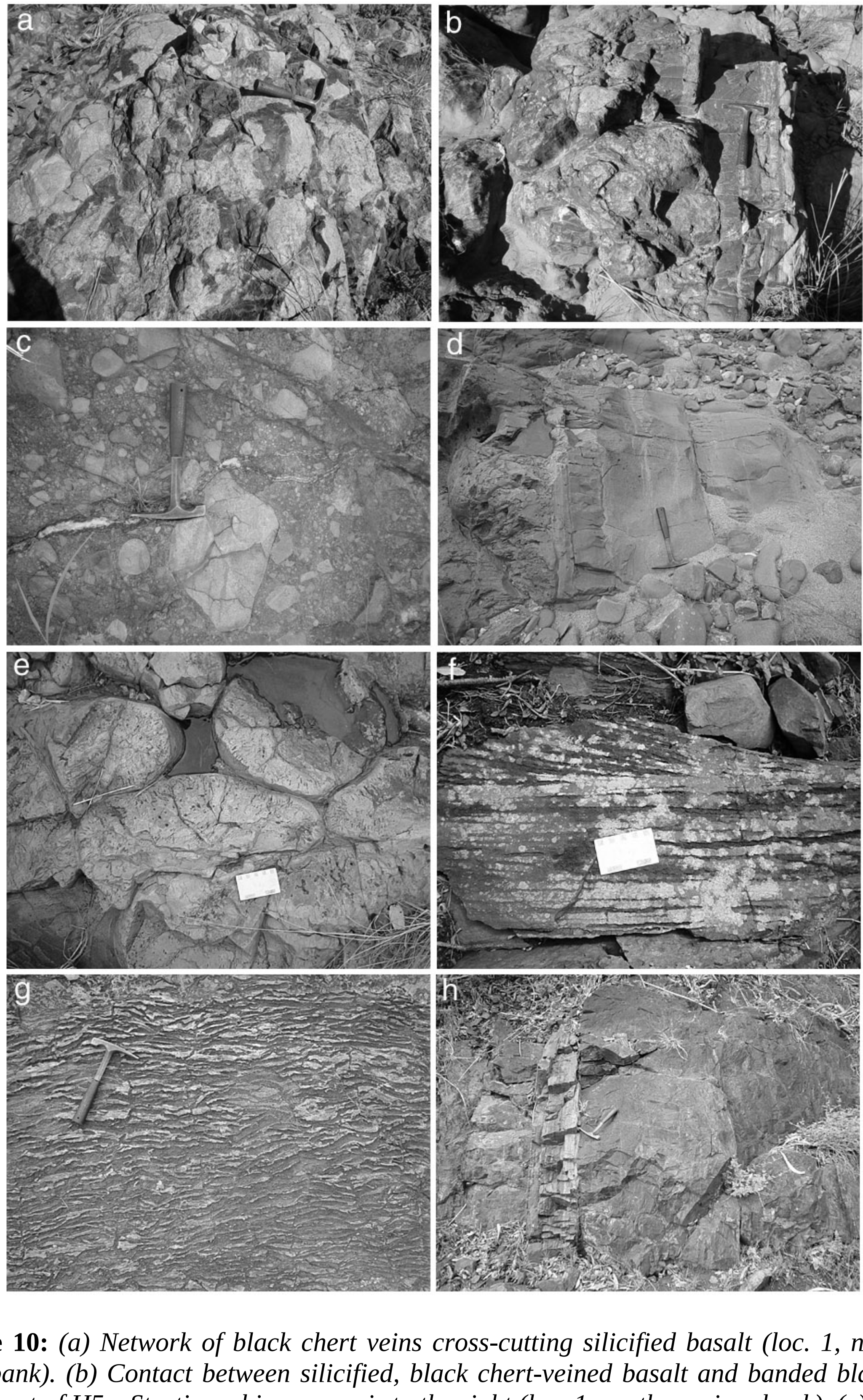


Figure 10: *(a) Network of black chert veins cross-cutting silicified basalt (loc. 1, northern river bank). (b) Contact between silicified, black chert-veined basalt and banded black and grey chert of H5c. Stratigraphic way-up is to the right (loc. 1, northern river bank). (c) Dacite cobble conglomerate (loc. 2). (d) Thick-bedded turbiditic sandstone and silicified shale. Stratigraphic way-up is to the right (loc. 2). (e) Pillow basalt with pipe vesicles (loc. 6). (f) Parallel- and trough cross-stratified ultramafic lapillistone (loc. 8). (g) Fuchsite-chert-carbonate zone representing altered ultramafic rocks (loc. 9). (h) Massive and banded black carbonaceous chert (loc. 10).*

consists of thick-bedded turbidites (Fig. 10d) of normally graded sandstone overlain by silicified shale (now chert), which grades upsection into silicified shale.

Locality 3. The sedimentary sequence H6 (Fig. 9) of the Hooggenoeg Formation is overlain along a sharp, but poorly exposed contact by a body of serpentinzed dunite that forms the base of the Kromberg Formation. Near the Komati River, the serpentinite has been trenched during an asbestos exploration programme. Similar ultramafic bodies are widespread in the Barberton greenstone belt and frequently represent intrusive sills. At this locality the ultramafic body is overlain by carbonated ultramafic lapillistone along a gradational contact. According to Viljoen and Viljoen (1969c) this relationship signifies an extrusive origin for the ultramafic rocks. Byerly and Lowe (1991) also favoured an extrusive origin, reporting the presence of spinifex-textured ultramafic rocks from a correlative section several kilometres to the north. On the other hand, silicified shales immediately below the ultramafic rocks along the Komati River are locally bleached, which may indicate a contact-metamorphic overprint.

Locality 4. A distinct banded chert horizon near the base of the Kromberg Formation, termed K1c1 (Fig. 9) and correlated with the Buck Reef Chert by Lowe and Byerly (1999), is exposed in the Komati River section. The exposed sequence starts with a lenticular-banded fuchsite-chert-carbonate alteration zone. The alteration zone contains lenses of less altered carbonated lapillistone, suggesting that the lapillistone represents the protolith of the alteration zone. This is overlain along a gradational contact by strongly carbonated, stratified lapillistone. The upper unit is represented by lapillistone that is transected by mostly stratiform black chert veins. The veins become more numerous upsection, resulting in massive black chert that contains matrix-supported lapillistone fragments near the top. In this unit, lapillistone becomes progressively more silicified upsection. This is followed by a chert horizon (K1c1), several metres thick, that consists of black and white banded chert.

Locality 5. The second chert horizon in the Kromberg section (K1c2, Fig. 9; Lowe and Byerly, 1999) is *c.* 25 m-thick and is intercalated with silicified variolitic basalt below and massive basalt above. It consists of black and white banded carbonaceous chert with minor bands of carbonate and green chert, the latter containing pseudomorphs after stellate crystals, possibly gypsum (Lowe and Knauth, 1977; Lowe and Byerly, 2003). Microfossils and fossil microbial mats have been reported from carbonaceous chert at this locality (Engel, 1968; Walsh and Lowe, 1985; Walsh, 1992). A shallow-marine environment of deposition has been proposed for this chert horizon (Lowe and Knauth, 1977).

Locality 6. Well-exposed rock pavement of pillow basalt *c.* 500 m above the base of the Kromberg Formation. Pillows show characteristic convex-upward tops and cusped bottoms and contain well-developed radial pipe vesicles and contraction cracks (Fig. 10e). Rims of pillow basalts from the Kromberg Formation contain micrometre-scale tubular structures, interpreted to represent corrosion features that formed by ancient microbes (Furnes et al., 2004).

Locality 7. Zone of silicified pillow basalt, now represented by a mottled rock of green and somewhat translucent black chert. Black chert veins also occur. This zone is underlain by carbonated, but not silicified pillow basalt.

Locality 8. A *c.* 20 m-thick, carbonated unit of parallel-laminated and trough cross-bedded ultramafic lapillistone (Fig. 10f) crops out 100 m upstream from the weir. The lapillistone is intercalated with massive and pillow basalt, but is part of a much thicker lapillistone unit.

Sedimentary structures may be related to reworking of volcaniclastic deposits in shallow water (Byerly and Lowe, 1991) or to pyroclastic transport and deposition during phreatomagmatic eruptions (De Wit et al., 1987).

Locality 9. Fuchsite-chert-carbonate zone representing strongly altered ultramafic rocks. Lenticular wavy bands of fuchsitic, apparently schistose, silicified volcanic rocks and associated dark grey, translucent chert occur in a brown-weathering carbonate matrix (Fig. 10g). Irregular veins of black chert oriented subparallel to the fabric are present. The rock is locally transected by shear zones that post-date the fabric-forming event. The alteration zone is overlain, along a sheared contact, by a 0.5 m-thick horizon of black chert that has been attenuated by shearing. Near the base of the unit, close to the weir (Fig. 8), occur slightly less altered rocks that show pillow structures and ocelli. Similar alteration zones are widespread in the Barberton greenstone belt and are derived from spinifex-textured komatiites and, less commonly, ultramafic lapillistone and komatiitic basalt. The origin of these zones is unclear. Lowe and Byerly (1986a) regarded them as low-temperature flow-top alteration zones that formed during submarine to subaerial weathering. Duchac and Hanor (1987) suggested subsurface hydrothermal alteration of ultramafic rocks. De Wit (1982) regarded such rocks as mylonites and interpreted them as major décollement zones along which extensive horizontal movements took place.

Locality 10. Above the chert-carbonate rock occur two units of massive basalt overlain by banded chert. The cherts may represent a tectonically duplicated single horizon (Lowe and Byerly, 2003), although two horizons of banded chert separated by volcanic rocks are common at this stratigraphic level at other localities (Hofmann, unpubl. data). The banded chert forms the top of the Kromberg Formation and has been denoted as K3c (Fig. 8) or Footbridge Chert by Lowe and Byerly (1999). Very good exposures of the chert can be found a few metres northwest of the footbridge that crosses the Komati River. The chert consists predominantly of massive black chert and black and white banded chert (Fig. 10h). Zircons from a dacitic tuff layer in the chert have yielded an age of 3334±3 Ma (Byerly et al., 1996). Possible microfossils were described from this layer by Westall and Gerneke (1998) and Westall et al. (2001).

Locality 11. A rock pavement in the Komati River just above the Footbridge Chert exposes the basal part of the Mendon Formation (Fig. 8). The rocks are mostly massive komatiitic basalts that contain interflow layers of massive to laminated black chert. The chert layers are disrupted, probably as a result of lava extrusion.

Locality 12. The contact between the Onverwacht and Fig Tree Group is exposed on a hill *c.* 800 m northeast of the footbridge (Fig. 8). The section starts with a fuchsite-chert-carbonate alteration zone that locally contains remnant spinifex-textures. The altered ultramafic rocks are sharply overlain by the Msauli Chert, an approximately 20 m-thick unit of graded beds of silicified accretionary lapilli. The graded beds show sedimentary features reminiscent of turbidites and have been interpreted as such by Stanistreet et al. (1981) and Heinrichs (1984). On the other hand, Lowe (1999b) interpreted the graded beds to represent shallow-water deposits. The Msauli Chert is overlain by black and white banded chert, followed by shale and siltstone of the Fig Tree Group.

Stop 2.2. Sandspruit Formation komatiites and komatiitic basalts and migmatite exposures on the farm Brandybal 171 IT (S26° 04.565’, E30° 50.327’)

Guide: Carl R. Anhaeusser

Objectives: The area south of the main Barberton greenstone belt consists of greenstone remnants surrounded and intruded by trondhjemitic grey gneisses similar to those seen in the Theespruit pluton. One such remnant occurs on the farm Brandybal 171 IT, and has been regarded as the type locality for the Sandspruit Formation (Viljoen and Viljoen, 1969b). The traverse selected for the excursion occurs further south and crosses a well-layered sequence of alternating schistose hornblende amphibolites (originally komatiitic basalts) and interlayered serpentinites and talcose schists (originally komatiites), and continues into the leuco-trondhjemitic gneisses of the Uitgevonden pluton exposed in a river section. The remainder of the traverse is designed to illustrate changes to the amphibolites and serpentinites where they are invaded by granitoid rocks. The intermingling of granitic rocks and greenstones has produced a spectacular set of migmatite exposures seen along the banks of the river (Fig. 11).

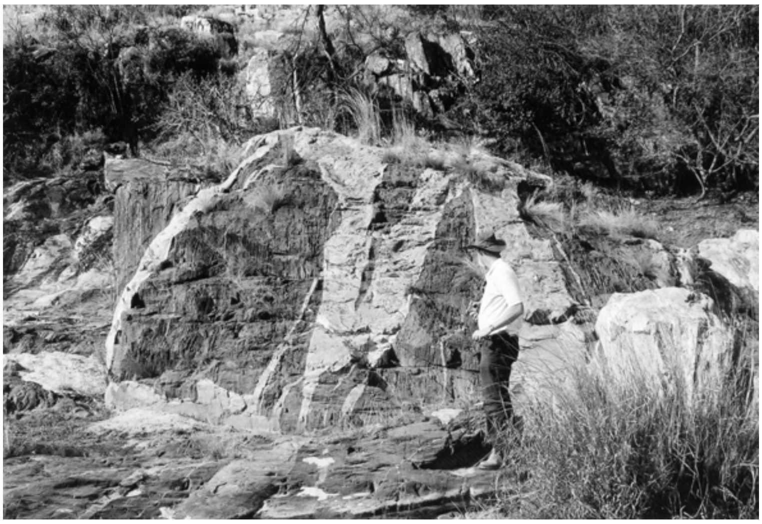


Figure 11: *Trondhjemite dykes in Sandspruit Formation amphibolite seen in the upper reaches of the Sandspruit stream on the farm Brandybal 171IT.*

Stop 2.2. The traverse begins on trondhjemitic gneisses developed on the northern flank of a large greenstone remnant that trends approximately NE-SW. A number of mafic and ultramafic lava units, all in amphibolite metamorphic grade, are crossed in the traverse before reaching the Uitgevonden pluton on the southeast side of the greenstone remnant. The trondhjemites show a gneissic fabric and steep lineation. Exposures of pseudotachylite are present where the river bends and flows in a northwesterly direction. This is also the only locality known to date in the Barberton Mountain Land where pseudotachylite has been found and the full significance of the discovery has yet to be determined.

The traverse continues along the river where spectacular migmatite exposures are evident over a distance of several hundred metres. Work still in progress along the section is aimed at detailed mapping and recording the variations and events seen in the outcrops, including the timing of the different granitoid rocks emplaced into the area.

Further north along the river a succession of finely laminated mafic schists and calc-silicate rocks is exposed, with some hornblende amphibolite layers showing the development of garnets. The traverse leaves the river bed and heads back along a small tributary stream towards the starting point. Along the way, a Proterozoic mafic dyke can be seen which shows horizontally orientated columnar jointing. A little further on, a second diabase dyke can be seen intruded into trondhjemite gneisses, the exposure providing a good example of the physical contact relationships of the dyke with the sidewall granitoid rocks. Chilled contacts and 'horn-like' structures splaying from the dyke are also in evidence. Dykes are prominent in the Barberton Mountain Land, particularly in the more 'brittle' granitic terrane. These dykes have not experienced any of the metamorphism and structural disturbance found throughout the district in the older rocks and most of the dykes are thought to be approximately the same age as the Bushveld Complex (i.e. *c*. 2060 Ma). Many thousands of dykes occur in the basement terrane extending north to the Limpopo Belt indicating major crustal extension on the Kaapvaal Craton around the time the Bushveld Complex was emplaced.

Stop 2.3. Nederland metasedimentary rocks

Guide: Annika Dziggel

Objectives: Examination of high-pressure amphibolite facies volcano-sedimentary rocks of the Sandspruit Formation and granite-greenstone contact relationships. Geochronological data combined with P-T estimates allow inferences to be made on the tectonic setting of the lowermost formations of the Onverwacht Group.

Locality 1 (S26°01.413', E30°42.458'). Along a short cross-sectional traverse west of the dirt road (and immediately north of the old Theespruit bridge, Fig. 12), the excursion will examine the field occurrence of medium-pressure amphibolites, interlayered recrystallized glassy cherts and finely banded clastic metasedimentary rocks. Lithologically, these rocks have been correlated with the lowermost formations of the Onverwacht Group. The strong layer-parallel foliation is subvertical and can be traced into the adjacent trondhjemitic gneisses. The occasionally boudinaged rocks display abundant quartzo-feldspathic veins, which are commonly tight- to isoclinally folded. Of particular interest are the clastic metasedimentary rocks that are characterized by a strong chemical banding (Fig.13). The well-equilibrated mineral assemblages comprise diopside, plagioclase, quartz, potassium feldspar and hornblende. Locally, poikiloblastic garnets up to several millimetres in diameter can be observed. Retrogression in these rocks is dominated by an occasionally extensive replacement of peak metamorphic minerals by epidote and sphene. P-T estimates for the peak metamorphic mineral assemblages vary between 650-700 °C and 8-11 kbars, indicating that these rocks were buried to at least 30 km during the peak metamorphic event, along an apparent geothermal gradient of *c*. 20°C/km. This implies a tectonic setting comparable to some modern orogenic belts, and that the exposed rocks possibly represent part of an exhumed mid- to lower-crustal terrane that formed a "basement" to the Barberton greenstone belt at the time of the peak metamorphic event, dated at *c*. 3230 Ma.

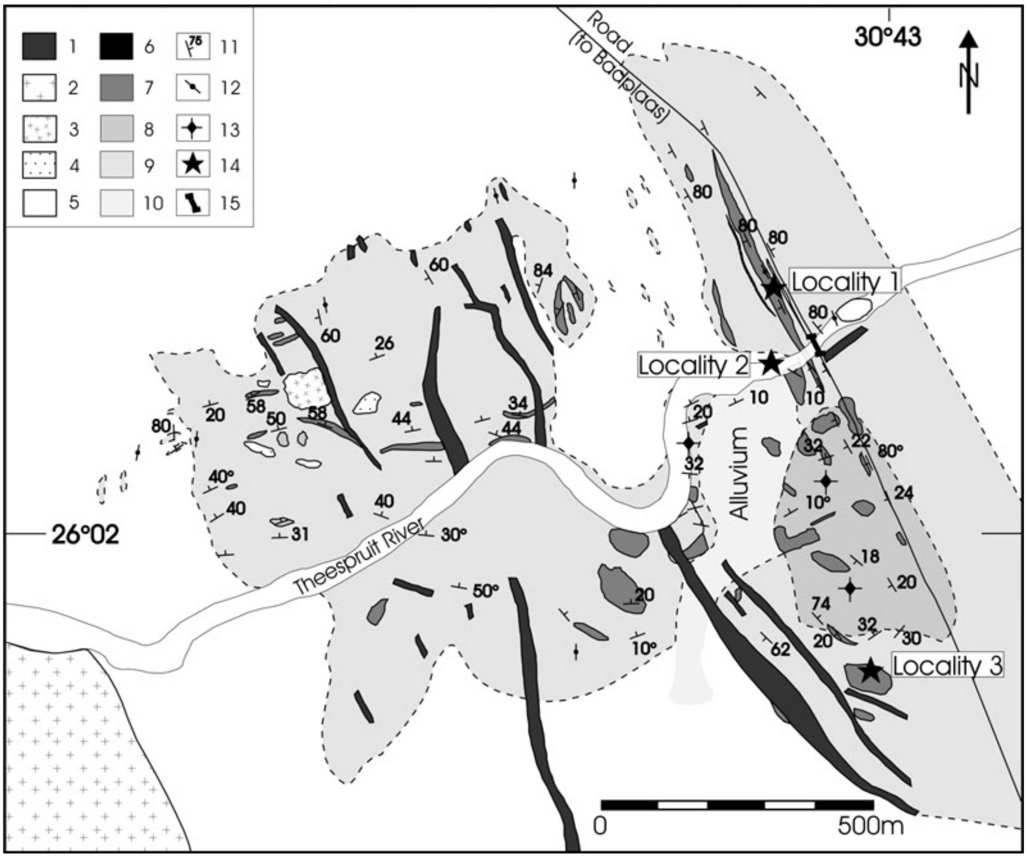


Figure 12: *Geological map of the Nederland greenstone remnant and localities visited during the field trip. 1- Mafic dykes; 2- Boesmanskop Syenite, 3- Alkaline intrusions; 4- Pegmatites; 5- Trondhjemitic gneisses; 6- Cherts; 7- Clastic metasedimentary rocks; 8- Ultramafic schists; 9- Amphibolites; 10- Alluvium; 11- Strike and dip of foliation*; *12- Strike of vertical foliation; 13- Horizontal foliation; 14- Field trip stop; 15- Bridge.*

Locality 2. Well-exposed pavements of trondhjemitic gneisses (Fig. 14) occur in the bed of the Theespruit River *c.* 100 m south of the previous locality, 80 m west of the Theespruit bridge. The trondhjemitic gneisses are intrusive into strongly banded and mylonitic amphibolites that dip at low angles to the north. The gneisses contain a pervasive foliation coplanar with the layer-parallel foliation observed at the previous locality. Some 100 m further southeast, the granite-greenstone contact is marked by an extensive brecciation of greenstone material by a fine-grained granitoid phase that crosscuts the pervasive fabric in the trondhjemitic gneisses and amphibolites. Conventional single-zircon dating yielded an age of 3431 ± 11 Ma (Fig. 15) for the emplacement of the trondhjemitic gneisses (Dziggel et al., 2002), and thus provides a minimum age for the greenstone remnant. Although slightly younger, this age is indistinguishable from other age estimates of the Stolzburg pluton (e.g. 3459 +35/-23 Ma, Kamo and Davis, 1994). Another explanation may be that the age of 3431 ± 11 Ma reflects a slightly younger pulse of magmatism in the area.

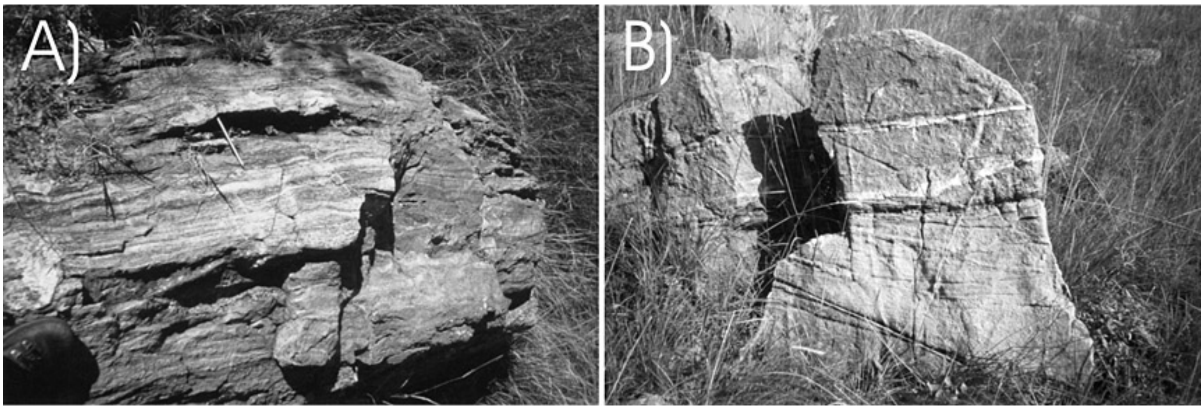


Figure 13: *(A) Finely banded and boudinaged clastic metasedimentary rock at locality 1; (B) Boulder of relatively undeformed, ungraded and cross-bedded meta-arkose at locality 3.*

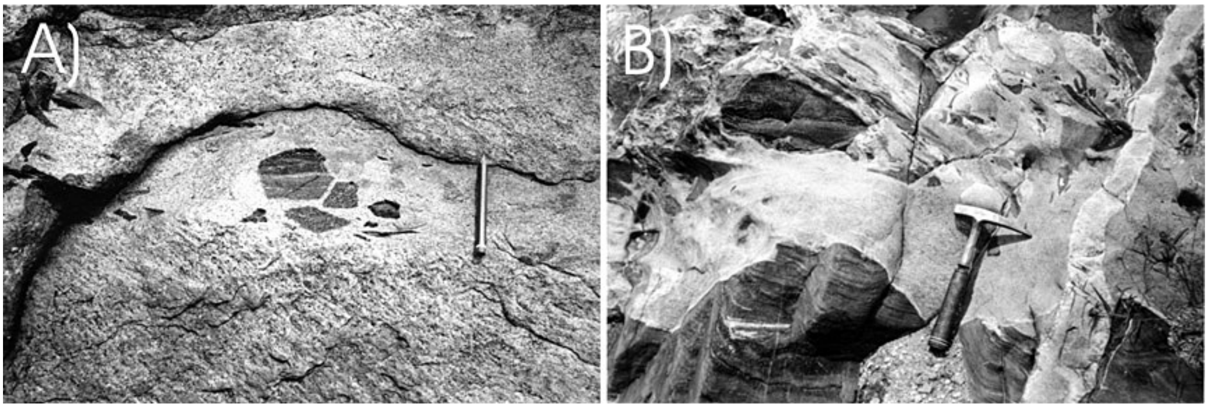


Figure 14: *(A) Fine-grained, leucocratic granitoid phase with xenoliths of amphibolite schists intrusive into and crosscutting trondhjemitic gneisses; (B) Magmatic brecciation of strongly schistose greenstone material some 100 m southeast of locality 2.*

Locality 3. A prominent, oval-shaped body, mainly composed of ultramafic schists, occurs to the south of the Theespruit River. At its southern margin, the excursion will examine boulders of relatively undeformed, medium- to coarse-grained and ungraded meta-arkoses, which are characterized by well-defined bedding-planes, and locally, by trough cross-beds. Bedding-parallel quartzo-feldspatic veins also occur. This is the only locality in which primary sedimentary features in these rocks have been preserved. The rocks contain up to 4.5 wt.% K_2O, and consist of potassium feldspar, quartz, plagioclase, clinopyroxene, and minor amounts of muscovite, epidote, actinolite and chlorite. Single zircon ages were obtained from these meta-arkoses, using ID-TIMS and SHRIMP dating (sample BE 5, Fig. 16). Only 5 grains gave concordant ages within error and imply a range of rock ages in the source area for the clastic sediments. The dates range from *c.* 3525 to 3540 Ma, indicating that at least two protoliths for the sedimentary rocks predate the formation of the bulk of the Barberton greenstone belt. A minimum age of 3431 ± 11 Ma is given by the late-tectonic trondhjemite. Thus, these metasedimentary rocks were deposited between *c.* 3521 and 3431 Ma, contemporaneously with the erosion of spatially associated older, and presumably potassium-rich, granitoid rocks.

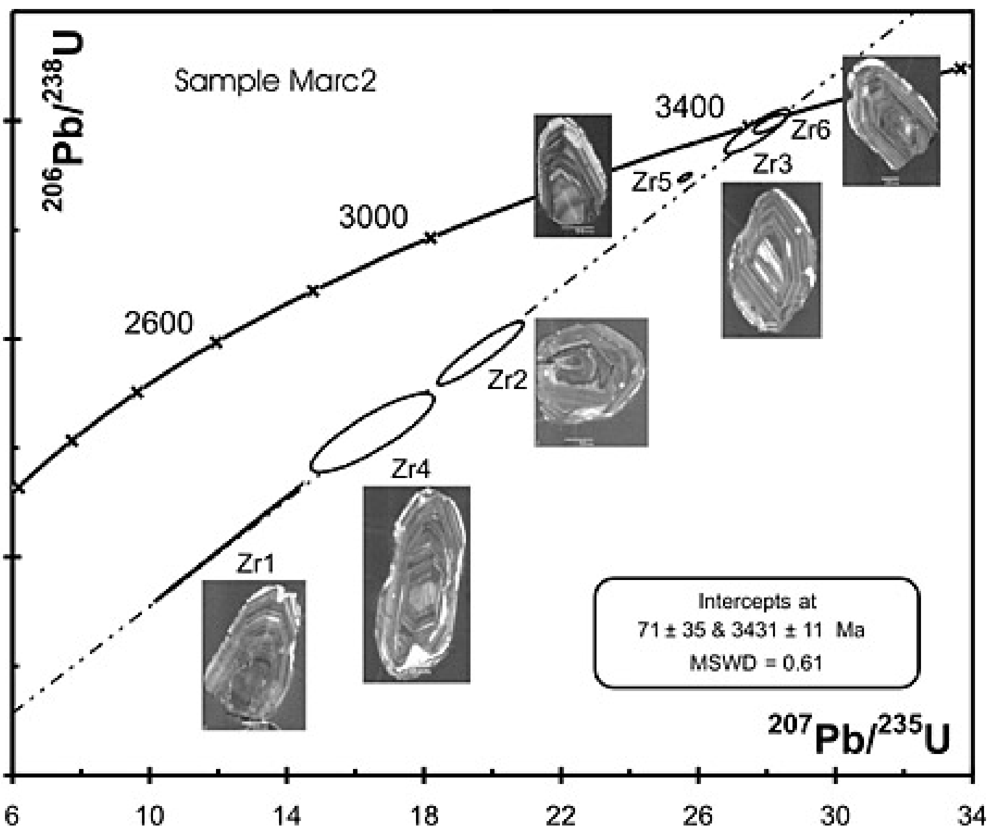


Figure 15: *U-Pb concordia diagram for a pervasively foliated trondhjemitic gneiss at locality 2. Inset shows data that are 90% concordant or more. Error ellipses are at 2 sigma (Dziggel et al., 2002).*

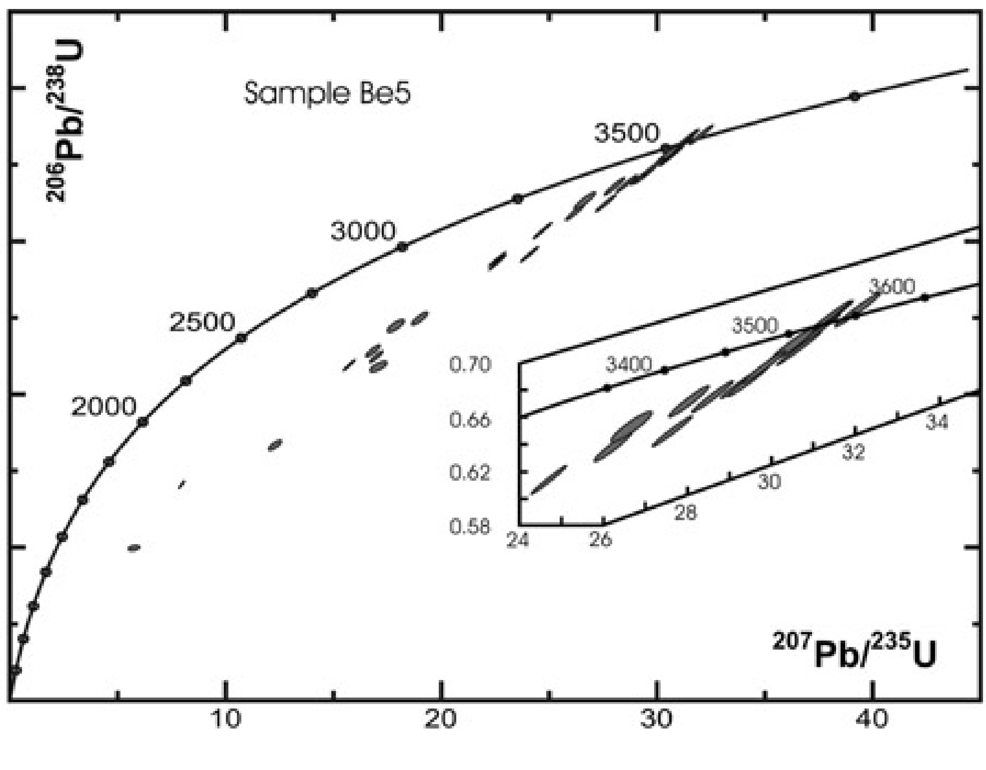


Figure 16: *U-Pb concordia diagram for detrital zircons in sample BE 5. Error ellipses are at 2 sigma (Dziggel et al., 2002).*

DAY THREE, TUESDAY, 6 JULY 2004

Stop 3.1. Manzimnyama Jaspilite (S25°54.335', E31°06.204')

Guide: Axel Hofmann

Exposures of folded, "oxide facies" jaspilitic banded iron formation. The jaspilite forms a continuous, a few tens of metres thick horizon, termed the Manzimnyama Jaspilite Member (Heinrichs, 1980). It occurs in the lower part of the Fig Tree Group in the Mapepe Formation. The iron formation consists of intercalated bright red jaspilite and haematite and/or magnetite-rich layers. The rock is finely laminated and lacks sedimentary structures produced by current activity suggesting a quiet, probably deep-water depositional environment. Primary iron oxides in chemical sediments in the Fig Tree Group may suggest local oxygen-rich conditions on an otherwise oxygen-poor mid-Archaean Earth, such as near the surface of a stratified Archaean ocean. Other explanations exist, such as the biologically mediated oxidation of ferrous iron by anoxygenic phototrophs (e.g., Ehrenreich and Widdel, 1994). Haematitic iron ore was mined from supergene-enriched Manzimnyama Jaspilite at Ngwenya haematite deposit in Swaziland (Ward, 1999).

Stop 3.2. Barite Valley Syncline (S25°54.300', E31°03.281')

Guide: Axel Hofmann

Objectives: The contact between the Onverwacht and Fig Tree Group is well exposed in the Barite Valley Syncline (BVS; Fig. 17), where a variety of interesting lithologies can be found in close proximity. These include sedimentary barite deposits, several impact spherule layers, and bedded carbonaceous cherts and chert dykes (*The section is located within the Songimvelo Nature Reserve, and access is via Sappi Forest roads; permission is required to visit these outcrops. Please avoid hammering of key outcrops*).

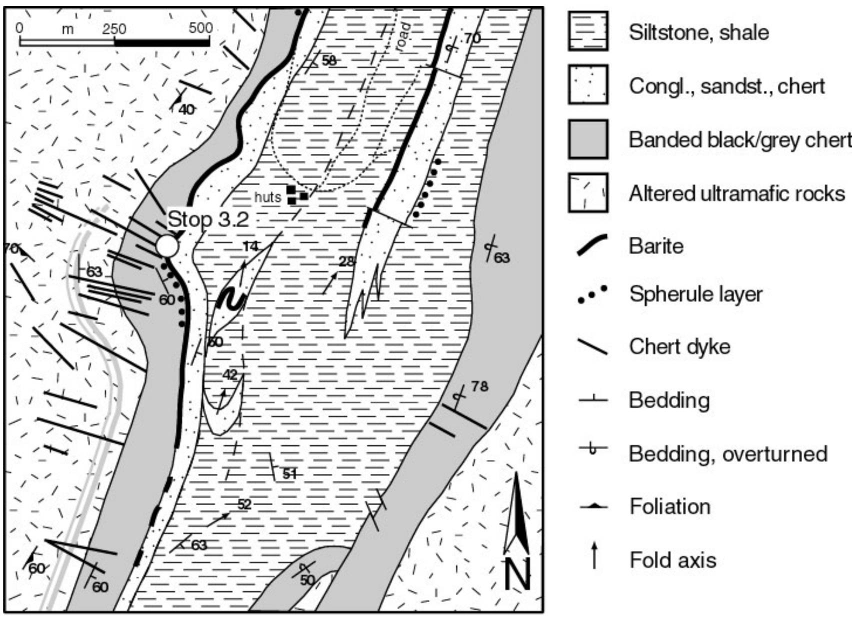


Figure 17: *Geological map of the southern part of the Barite Valley Syncline (Heinrichs and Reimer, 1977; Lowe and Byerly, 2003; Hofmann, unpubl. mapping) and locality of excursion stop 3.2.*

Stratigraphy. The stratigraphy and sedimentology of Fig Tree strata in the BVS (Fig. 17) have been described by Heinrichs and Reimer (1977) and Lowe and Nocita (1999). As pointed out by Lowe and Byerly (2003), the BVS is not a true syncline, as the two limbs are separated by faulting and show a different stratigraphy. The strata of the two limbs dip steeply to the east-southeast (Fig. 17). Open to tight folding is common in the central part of the synclinal structure and fold axes have a shallow to moderate plunge to the north-northeast.

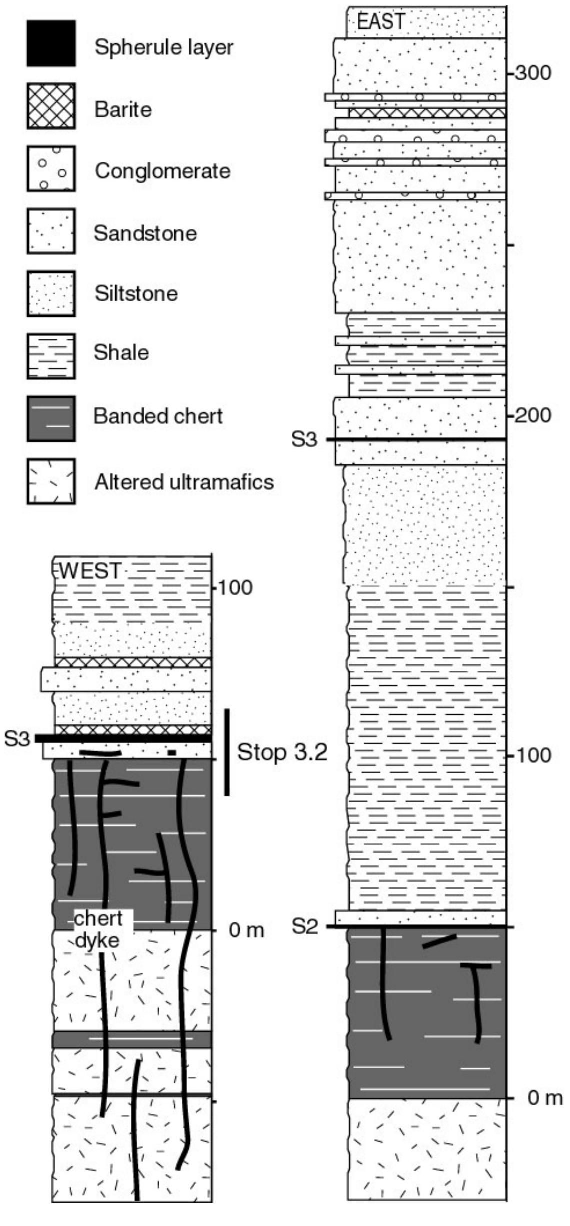


Figure 18: *Stratigraphic logs from the west and east limb of the Barite Valley Syncline (modified after Lowe and Nocita, 1999; Lowe and Byerly, 2003).*

On the west limb (Fig. 18) the lowermost stratigraphic unit consists of intensely altered komatiites of the Mendon Formation, represented by a fuchsite-chert-carbonate rock with a lenticular fabric, translucent chert veins and remnants of spinifex texture. A sequence of banded black and white and grey chert overlies the volcanic rocks. When weathered, this unit has the appearance of silicified shale and siltstone. Numerous chert dykes cut through the sedimentary and underlying volcanic rock units. Banded chert is overlain, possibly along an erosional unconformity (Lowe and Byerly, 2003), by a lithologically complex unit (*c.* 5-10 m thick) of massive greenish-grey chert, fine chert pebble conglomerate containing impact spherules, and bedded barite intercalated with silicified sandstones. Chert dykes do not penetrate this unit, which is overlain by about 20 m of parallel-laminated sandstone with intercalations of silicified sandstone, fine chert-pebble conglomerate, barite and jaspilite, followed by >100 m of parallel-laminated tuffaceous sandstone, siltstone and shale.

The stratigraphy of the east limb of the BVS is somewhat different (Fig. 18). A sequence of silicified fine-grained sedimentary rocks, now represented by black and white banded chert, overlies ultramafic rocks that are altered to a fuchsite-chert-carbonate rock. The sedimentary rocks are sharply overlain by a *c.* 5 m-thick, laterally continuous, lithological unit of black and white banded chert with jaspilite bands, chert-pebble conglomerate, massive grey chert, and rare stratiform veins of black chert. Rare chert dykes in the underlying units do not penetrate this zone. Chert is overlain by a coarsening-upward sedimentary sequence, *c.* 230 m thick, that consists of shale with some jaspilite beds in the lower part and a parallel-laminated tuffaceous siltstone to fine-grained sandstone facies in the middle part. Near the base of the sandstone facies is a bed of massive, densely packed impact spherules, approximately 20 cm thick. The upper part of the sedimentary sequence consists of interstratified sandstone, locally cross-bedded chert-pebble conglomerate, banded black and grey chert, and jaspilite, and is capped by barite. This is followed by a fining-upward sequence of chert-pebble conglomerate and sandstone, grading into parallel-laminated tuffaceous sandstone and siltstone, dated at 3227±4 Ma (Kröner et al., 1991). Detailed stratigraphic logs of this sequence have been reported by Lowe and Nocita (1999). These authors interpret the Fig Tree strata in the BVS to represent fan-delta and coastal deposits in an area of reduced subsidence or along the flanks of structural highs.

Spherule layers. Four spherule beds (two of which - S2 and S3 - are shown in Fig.18) have so far been described from the Barberton greenstone belt, S1 in the upper part of the Hooggenoeg Formation, and S2 to S4 in the Fig Tree Group (Lowe et al., 1989; Lowe and Byerly, 1999; Lowe et al., 2003). While S4 has been reported to occur at only a single locality, the other beds in the Fig Tree Group crop out at several localities in the central part of the greenstone belt. Difficulties have been encountered with the correlation and numbering of these spherule layers (Lowe et al., 2003; Hofmann et al., 2004). This is because the central part of the Barberton greenstone belt consists of several fault-bounded tectonostratigraphic units that structurally duplicate sections of the uppermost Onverwacht ultramafic volcanics and overlying Fig Tree strata (e.g. Lowe et al., 1985). Furthermore, zircon dating has suggested that the Onverwacht-Fig Tree contact in different tectonostratigraphic units is diachronous (Kröner et al., 1991; Byerly et al., 1996).

The spherules are interpreted to have formed by the condensation of global clouds of impact-generated rock vapour, based on spherule composition (Lowe et al., 2003), Ir anomalies (Lowe et al., 1989; Kyte et al., 1992; Lowe et al., 2003), spinel compositions (Byerly and Lowe, 1994), and Cr-isotope ratios, indicating an extraterrestrial Cr component and suggesting carbonaceous chondrite projectiles (Shukolyukov et al., 2000; Kyte et al., 2003).

Several spherule layers occur in the BVS. At stop 3.2 on the west limb of the BVS a layer near the base of the Fig Tree Group represents a spherule-bearing chert-pebble conglomerate (Fig. 19a). Locally, two layers occur separated by several metres of various chert beds. The spherule layer(s) has been correlated with bed S2 (Lowe and Nocita, 1999) or S3 (Lowe et al., 2003), the type localities of which occur a few kilometres to the southwest.

Other spherule layers on the east limb of the BVS (Fig. 18) include a poorly exposed bed at the base (S2) and a thin bed of almost pure spherules (S3) *c*. 140 m above the base of the Fig Tree Group (Lowe and Nocita, 1999; Lowe et al., 2003). Lowe et al. (2003) reported Ir values of 2-145 ppb for a bed on the west limb of the Barite Syncline (unclear which bed was analysed), and 5.8 ppb for S3 on the east limb of the BVS, a value close to Barberton komatiite values (< 5 ppb).

Tuffaceous rocks *c.* 300 m above the Onverwacht-Fig Tree contact on the eastern limb of the BVS have been dated at 3227±4 Ma (Kröner et al., 1991). This age is identical to ages obtained for the Schoongezicht Formation (Kröner et al., 1991; Kamo and Davis, 1994), which is the stratigraphically uppermost unit of the Fig Tree Group. This age indicates that the section of Fig Tree Group rocks in the BVS is condensed and may contain unconformities, as suggested by Heinrichs and Reimer (1977) and Lowe and Nocita (1999), making stratigraphic correlations between spherule layers at this locality difficult.

Barite beds. Layers of barite occur at particular stratigraphic levels within the Fig Tree Group and are well developed in the BVS where they can be traced for several kilometres. The barite is regarded as sedimentary in origin and has been studied by Heinrichs and Reimer (1977) and Reimer (1980). The barite beds (Fig. 19b) consist of detrital barite and coarse barite crystals, the latter forming cauliflower-like structures locally. Barite is admixed with

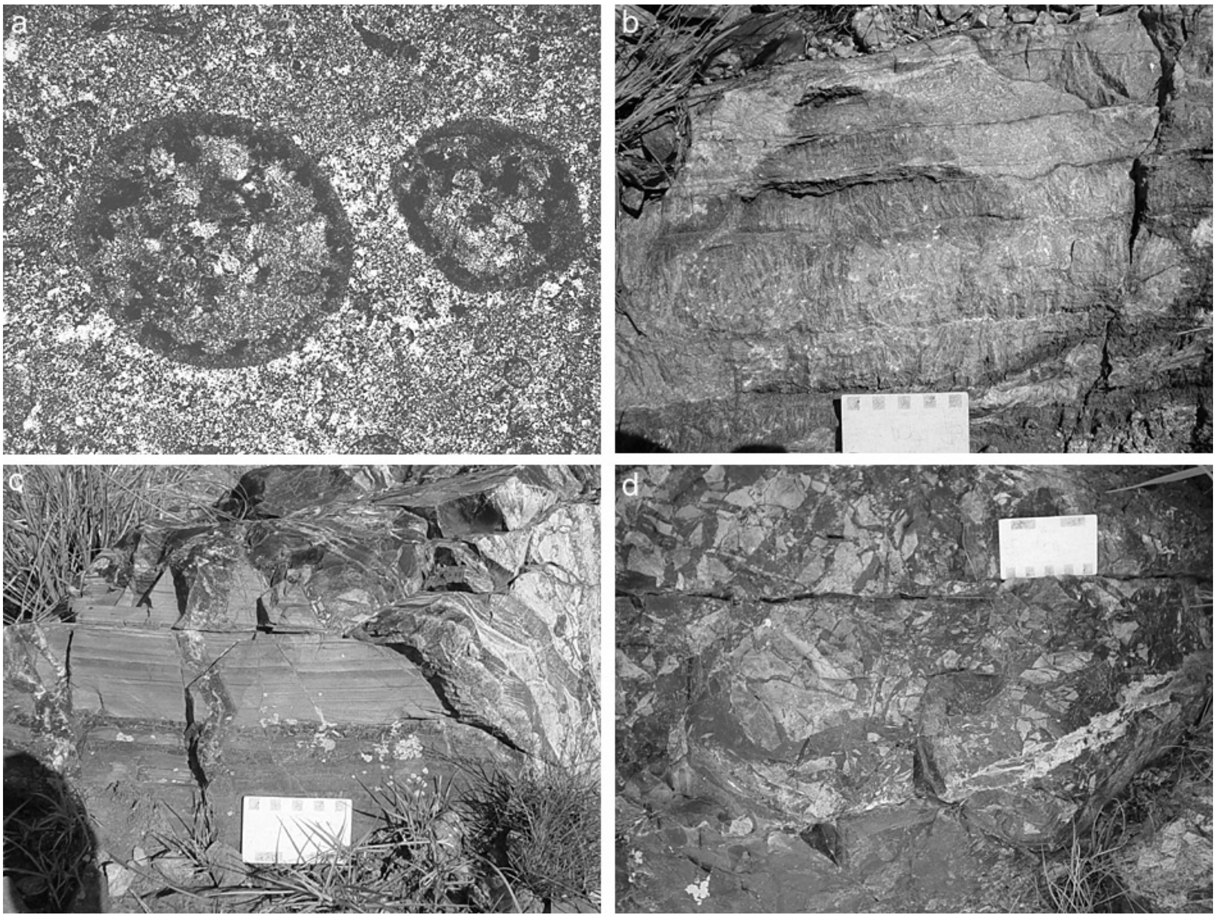


Figure 19: *(a) Photomicrograph (crossed polarized light) of spherules from spherule bed along the west limb of the Barite Valley Syncline. (b) Bedded barite. (c) Silicified sedimentary rock transected by chert dyke (upper right) showing internal layering. (d) Breccia of wall rock fragments in chert dyke. Note fragments with a jigsaw fit.*

minor amounts of volcanic quartz, muscovite, pyrite, zircon and chromite. The detrital barite is interpreted to represent reworked deposits of hydrothermal barite. Coarse-crystalline barite may represent primary precipitates and/or diagenetic replacements. Reimer (1980) reported $\delta^{34}S$ values of 2 to 8.4‰ for the barite. Barite has been recovered from a number of deposits in the BVS in the 1940s to the 1960s. The main barite workings are situated *c.* 2 km to the northeast of stop 3.2, where several barite beds occur in a *c.*10 m wide section (Heinrichs and Reimer, 1977; Ward, 1999).

Chert dykes. Chert dykes range in width from less than a cm to a maximum of *c.* 3 m. They dip steeply to the south-southwest (Fig. 17) and terminate along a greenish-grey chert horizon that is situated immediately below the impact spherule and barite beds. This chert horizon contains abundant diffuse patches of black translucent chert. Metre-scale chert dykes are frequently composite dyke-in-dyke structures, show a crude layering subparallel to the dyke walls (Fig. 19c), and are filled with massive black chert, botryoidal chert, and a very common black chert variety that, when weathered, reveals a fragmental texture. This texture is related to the presence of mostly granule- to pebble-sized, angular to rounded, wall-rock fragments floating in a matrix of black chert. Some fragments show evidence of *in situ* brecciation by thin chert veining, as indicated by the preservation of a jigsaw fit (Fig. 19d).

Black chert also forms bedding-parallel veins that branch off the dykes. These veins are up to 0.5 m wide and transect and brecciate the host rock subparallel to bedding. Several dykes contain equant to tabular, angular to rounded, pebbles of black, slightly translucent chert in a matrix of fragmental black chert. Rounding is commonly a result of marginal replacement by black chert. Furthermore, banded cherts are frequently replaced along the margin of chert dykes and veins. About 10 m below the impact spherule layer, spherules occur in several black chert dykes in a bedding-parallel, poorly exposed zone with a minimum lateral extent of 10 m and a thickness of *c.* 3 m. The spherules either float in black chert or form rock fragments composed entirely of spherules. They were regarded as having been derived from the overlying bed by downward movement into fissures, now represented by chert dykes (Lowe et al., 2003). However, spherules in spherule-bearing rock fragments are tightly packed, which is in contrast to spherules in the overlying bed that are disseminated in a chert-pebble conglomerate. On the other hand, chert clasts are lacking in the chert dykes. These relationships indicate that the spherules were not derived from this bed.

Stop 3.3. Dycedale Syncline (S25°47.415', E31°05.131')

Guide: Ken Eriksson

Upward-fining, fluvial channel deposits that record evidence for tidal modification are preserved in the Clutha Formation of the Moodies Group in the Dycedale Syncline. Facies are arranged in 45 to 140 cm-thick, fining-upward packages, in which the proportion of interlaminated sandstone, siltstone and mudstone increases upwards (Fig. 20). Basal conglomerates, up to 30 cm thick, are erosional and consist mainly of quartz pebbles and ripped-up clasts of laminated sandstone, siltstone and mudstone. Clast size decreases upwards within conglomerate beds. Overlying cross-bedded sandstone ranges in grain size from very coarse to fine sand. Locally, pebble stringers define set boundaries. Cosets vary from 20 to 210 cm thick. In several places, laminated sandstone, siltstone and mudstone, and wave and combined-flow ripple bedforms are preserved below coset boundaries. Within sets, foresets are tangential, planar or sigmoidal in shape, and, towards the top of upward-fining packages, commonly are draped with mudstone. In general, thin foresets have continuous mudstone drapes, whereas thicker foresets have no drapes, discontinuous drapes or are separated by mudstone chips. In bedding-plane views, these chips display polygonal desiccation cracks. Reactivation surfaces are present throughout the section. Laterally, within sets, a systematic thickening and thinning of foresets occurs with a corresponding increase in development of mudstone drapes associated with thinner foresets (Fig. 21). Some foresets contain internal ripple cross laminations directed up the foresets. These ripple cross laminations show a complex pattern of mudstone drapes. Interlaminated sandstone, siltstone and mudstone intervals cap the upward-fining packages and attain a maximum thickness of 25 cm, but commonly are absent at the tops of fining-upward packages as a result of erosion. Vertically within these intervals, thick-thin pairs and systematic thickening and thinning of laminations are developed. Desiccation cracks are ubiquitous. Where laminations are absent, mudstones are black and desiccation cracked.

The vertical sequence of strata within upward-fining packages records the increased influence of tidal currents with time at the expense of fluvial processes. Evidence for the change from fluvial to tidal processes includes an upward decrease in the proportion of conglomerate, the increase in abundance of mudstone drapes on foresets, the presence of cyclic foresets, and the occurrence of interlaminated sandstone, siltstone and mudstone at the top of upward-fining packages. Vertical transition from fluvial to dominantly tidal facies is considered to be related to sea level fluctuations rather than tectonics. Conglomerates reflect channel processes, whereas cosets of trough and tabular cross-bedded sandstone and the laminated sandstone, siltstone and mudstone were generated by flows modified by various tidal beats. Cosets of trough and tabular cross-bedded sandstone, with or without mudstone drapes, reflect lateral accretion of sediment, whereas interlaminated sandstone, siltstone and mudstone records vertical accretion. In both facies associations, mudstone developed during slack water phases, whereas sand and/or silt transport took place during the ebb or flood stages. Within both laterally and vertically accreting facies, alternating thick-thin laminations reflect diurnal twice-daily dominant and subordinate tides. Thinner groupings of foresets and thinner intervals of vertically stacked sandstone-siltstone-mudstone laminations formed during neap tides, whereas thicker groupings of foresets and laminations developed during spring tides. Desiccated mudstone drapes on foresets indicate that bedforms rarely were exposed during some portion of the tidal cycle. Evidence for exposure is best preserved at the top of upward-fining packages.

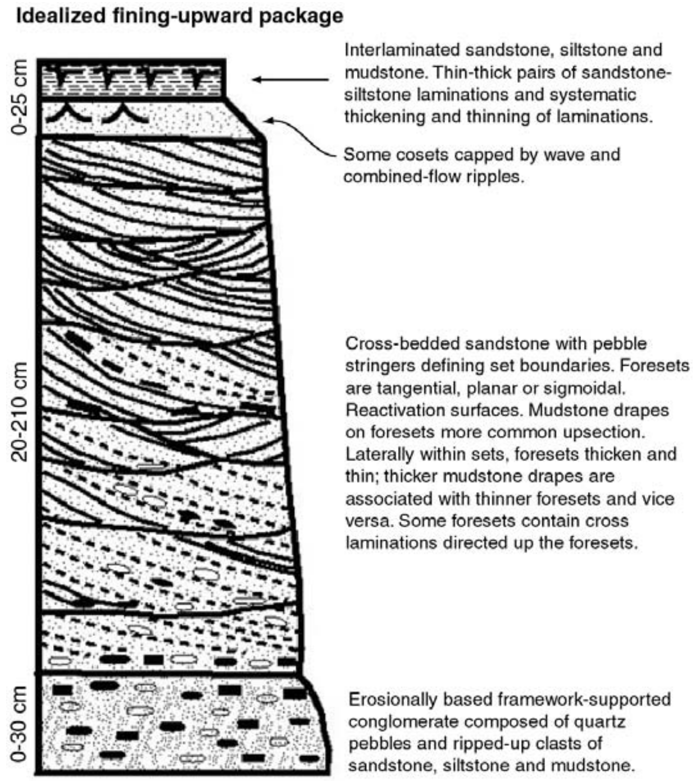


Figure 20: *Idealized vertical sequence of lithologies and sedimentary structures in upward-fining, tidally modified, fluvial deposits from the Moodies Group in the Dycedale Syncline located immediately east of Barberton.*

Figure 21: *Cross-bed set showing an increase in thickness of mudstone drapes from left to right corresponding with an increase in thickness of foreset bundle, Moodies Group, Saddleback Syncline (scale in cm).*

Stop 3.4. Sheba Creek, Eureka Syncline (S25°42.547', E31°09.671')

Guide: Ken Eriksson

Quantitative evidence for tides has recently been recognized for the first time in the Clutha Formation of the *c.* 3.25 Ga Moodies Group in the Barberton greenstone belt (Eriksson and Simpson, 2000), representing the oldest quantitative records of ancient tides. Tidal signatures in the Moodies Group along Sheba Creek in the Eureka Syncline are preserved as bundles of sandstone foresets separated by mudstone drapes in a tidal sand-wave deposit (Fig. 21). Detailed measurements of foreset-bundle thicknesses at a millimetre scale were made along traverses through the sand-wave deposit and plotted on a histogram of foreset bundle thickness versus foreset bundle number (Fig. 22A). Analysis of this plot, by analogy with modern tidal processes and records (Nio and Yang, 1980; Tessier et al., 1995), has led to the identification of a hierarchy of diurnal, semi-monthly, and monthly tidal periodicities (Eriksson and Simpson, 2000). Thick-thin pairs of foreset bundles are considered to reflect deposition from semidiurnal dominant and subordinate flood-tidal currents, respectively. Similar thick-thin diurnal pairs are widely developed in Holocene tidal sediments (De Boer et al., 1989). Cyclic variations in foreset bundle thicknesses (Fig. 22A) record longer period changes in strength of the dominant semidiurnal tidal currents consistent with semi-monthly neap-spring-neap tidal cyclicity. Alternating thicker and thinner neap-spring-neap cycles (Fig. 22A) are comparable to monthly anomalistic, perigean-apogean tidal signatures. Fast Fourier Transform analysis on the data set reveals strong peaks at 13.11, 9.83 and 2.18. The last 2 peaks are consistent with the interpretation of diurnal and neap-spring cyclicity discussed above, whereas the 13.11 peak is considered to record neap-spring-neap cycles in which both dominant and subordinate semi-diurnal bundles are developed (Eriksson and Simpson, 2000). Fast Fourier Transform analysis on the 4-5 month-long data set, from which inferred semi-diurnal, subordinate-tide foreset bundles had been removed (Fig. 22B), reveals only one well-developed peak at 9.33 that is interpreted as a strong semi-monthly signature (Eriksson and Simpson, 2000). Close inspection of Figure 22B reveals that monthly perigean-apogean cycles in the Moodies sand-wave deposit have a maximum number of 20 foreset bundles. These cycles suggest a lunar synodic orbital period of 18-20 days. This is considered to be an

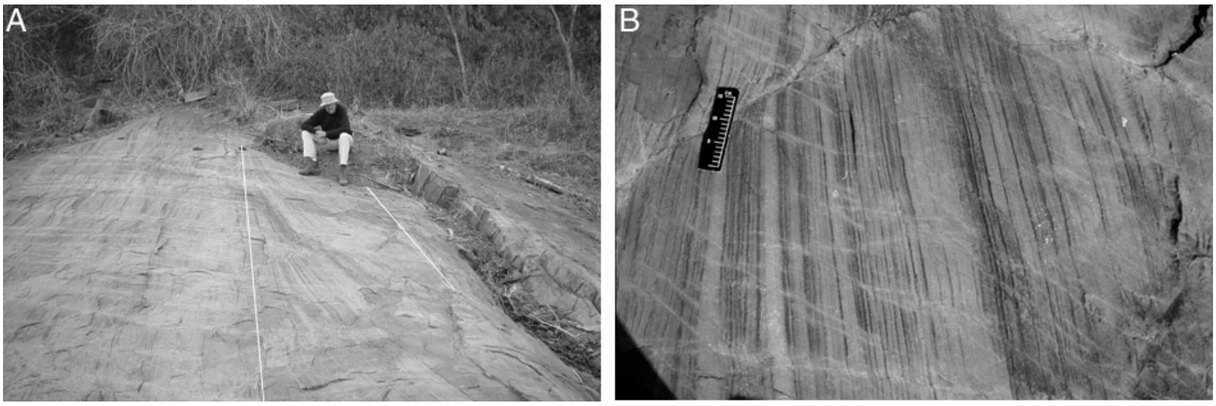


Figure 22: *(A) Tidal sand-wave deposit in the Moodies Group, Eureka Syncline. Bed is delineated by white lines; stratigraphic way-up is to the right. (B) Close-up view of (A) showing bundles of foresets separated by mudstone drapes (scale bar is 15 cm long). Note the thickening and thinning of foreset bundles.*

estimate of the minimum number of days in the synodic month during the middle Archaean because of the possibility of missing neap-tide foreset bundles especially within the apogean component of the monthly cycle when tidal current velocities are less than during perigee.

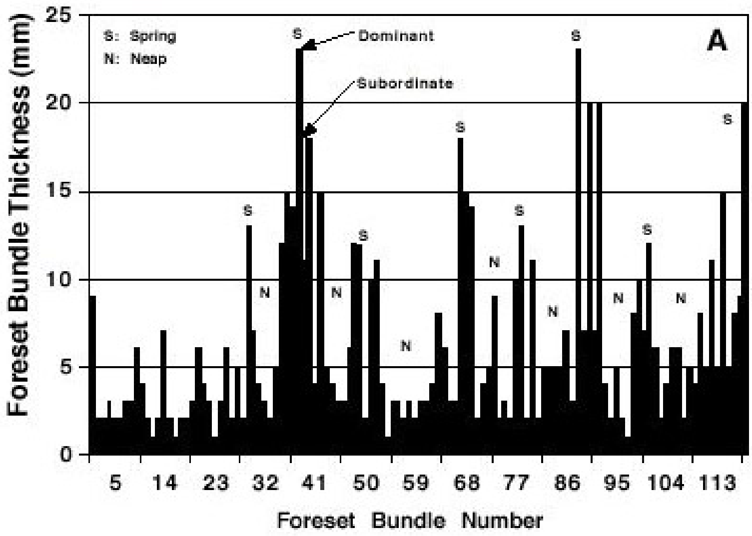
Stop 3.5. Basal conglomerate of the Moodies Group, Eureka Syncline

Guide: Carl R. Anhaeusser

Objectives: The Moodies Group is a dominantly sedimentary succession and has been subdivided into three formations in the Eureka Syncline type locality northeast of Barberton (Anhaeusser, 1976a). At the base is the Clutha Formation, which is overlain, in turn, by the Joe's Luck and Baviaanskop Formations. The sequence in the Eureka Syncline is approximately 3000 m thick, but varies considerably from place to place due to structural influences. The succession consists predominantly of conglomerates, quartzites, sandstones, subgreywackes, shales, banded and jaspilitic iron formations and minor intercalated alkalic lava units. Each of the formations commences with a conglomerate at the base and grades upwards into quartzites and sandstone-subgreywacke-shale units. The excursion will examine the basal conglomerates at two localities on the eastern side of the Eureka Syncline (stop 3.5, Fig.1). In the north (locality 1, north of stop 3.4) the conglomerates are located in close proximity to the granitic contact rocks of the Nelspruit batholith and have been extensively flattened and shortened as a result of the granite intrusion. By contrast the same conglomerate unit on the southern side of the syncline, near the Sheba Fault (locality 2, south of stop 3.4), shows very little structural deformation and the polymictic conglomerate pebbles can be seen in an almost pristine state compared to those in the north. The exposures are instructive from a structural point of view and they also provide an opportunity to examine the varied nature of the pebble population.

Locality1. Deformed basal conglomerates of the Moodies Group and underlying agglomerates of the Schoongezicht Formation of the Fig Tree Group. Ezzy's Pass locality, northern limb of the Eureka Syncline.

The road cutting at Ezzy's Pass on the Barberton-Kaapmuiden road slices through the basal conglomerate of the Clutha Formation at the base of the Moodies Group. It also exposes a portion of the deformed agglomerates of the Schoongezicht Formation at the top of the Fig



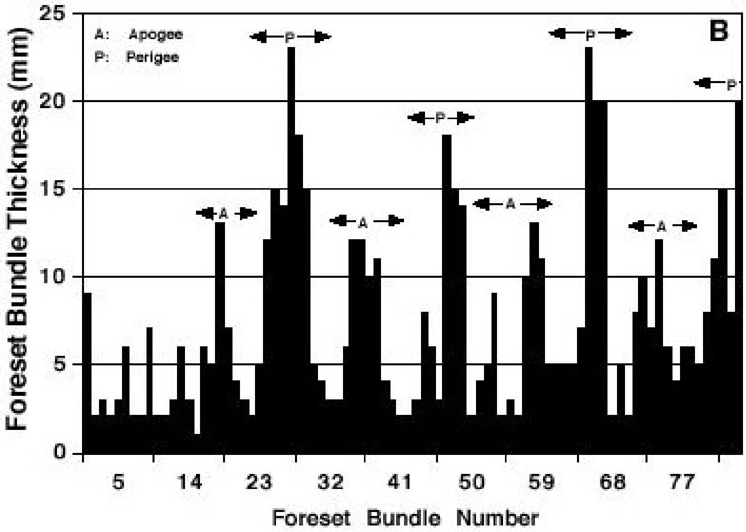


Figure 23: *(A) Bar chart of sandstone bundle thickness versus bundle number from a sandwave deposit in the Moodies Group, Eureka Syncline. Note the presence of thick-thin pairs of bundles considered to reflect the dominant and subordinate tides of a diurnal system, and the cyclic thinning and thickening of foreset bundles reflecting neap-spring-neap tidal cyclicity. (B) Plot of sandstone bundle thickness versus bundle number using same data set as A, but with inferred subordinate tidal bundles removed. Note the alternation of thicker and thinner neap-spring-neap cycles considered to represent, respectively, perigee and apogee records of an anomalistic tidal system.*

Tree Group. A traverse along the roadway shows flattened and elongated volcanic clasts in the Schoongezicht sequence, the latter overlain, in turn, by alternating Moodies conglomerates and subgreywackes. Features to note include the strongly flattened pebbles, mainly consisting of black chert, but also containing pebbles of granitic composition, as well as clasts of quartzite, lava, banded iron formation and shale fragments. The chert pebbles are strongly flattened ellipsoids and are drawn out to display a subvertical lineation. The mean pebble deformation at this locality was calculated to be 61% (Anhaeusser, 1969, 1976a). The granitic pebbles are unusual in that they display spectacular graphic intergrowth textures

(Anhaeusser, 1966) which may be an artefact of the deformation and mineralogical reconstitution of the rocks. Pebbles of granitic composition are rare in the Moodies Group, most of the recorded localities being in Moodies sediments occurring along the northern flank of the Barberton greenstone belt. Kröner and Compston (1988) reported single-grain U-Pb zircon ages for granitic clasts from this locality, which varied between 3570±6 Ma and 3518±11 Ma, and they suggested that the Moodies sediments had been derived from Ancient Gneiss Complex rocks similar to those preserved in parts of Swaziland, which is situated to the southeast of the Ezzy's Pass locality.

Locality 2. Relatively undeformed basal conglomerates of the Moodies Group on the southern side of the Eureka Syncline, near the Sheba Fault.

Approximately 3 km south of the Ezzy's Pass locality are exposures of relatively undeformed Moodies basal conglomerates exposed in Sheba Creek. Also seen in the creek are exposures of the Sheba Fault, which at this locality strikes approximately east-west and shows intricate "horsetail" structures linked to the shearing effects of the fault in a transcurrent or wrench fault tectonic regime (Anhaeusser, 1976a). Also seen in the creek are exposures of greywackes, shales and banded cherts of the Fig Tree Group developed on the northern flank of the Ulundi Syncline and south of the Sheba Fault (on the south bank of the creek).

The matrix-supported Moodies conglomerates show pebbles of chert, banded iron formation, granite, quartzite, lava, jaspilite and shale. As mentioned earlier, the rocks are only weakly deformed by comparison with the conglomerates seen at Ezzy's Pass, and they still show their well-rounded form. The influence on these rocks by the shearing along the Sheba Fault has also been minimal despite the close proximity of the fault zone.

DAY FOUR, WEDNESDAY, 7 JULY 2004

Option 4.1. Stolzburg Layered Ultramafic Complex, Barberton greenstone belt

Guide: Carl R. Anhaeusser

Objectives: The excursion to the Stolzburg Complex in the extreme southwestern part of the Barberton greenstone belt (Fig. 24, stop 4.1) is designed to give delegates an opportunity to view the plutonic mafic and ultramafic equivalents of the extrusive komatiitic lava sequences seen earlier on Day One.

The Stolzburg Complex is but one of at least 27 known layered ultramafic complexes developed in various segments of the greenstone belt. Most of the complexes occur in a zone extending from the far northeastern part of the belt to an area west of the belt where the Kalkkloof layered complex is partly covered by the Transvaal Supergroup rocks north of Badplaas – a distance of approximately 110 km (Fig. 24). The Stolzburg Complex is over 16 km in length and averages about 1 km in thickness. It is bounded on the northwest and southeast by two major regional faults that have been reactivated on numerous occasions. The layered body has a subvertical dip and consists of a Lower Division made up mainly of alternating cycles of serpentinized dunite and orthopyroxenite, and an Upper Division made up of cycles of serpentinized dunite, harzburgite, lherzolite, websterite gabbro-norite and anorthositic gabbro-norite. The two divisions are separated by a zone of calcium-enriched rodingite in the form of massive replaced gabbro and as rodingite dykes (Anhaeusser, 1979, 1985).

The layered ultramafic complexes have been interpreted as representing the penecontemporaneous plutonic differentiation products of komatiitic parent magmas rich in MgO (*c*. 28%). Fractionation of olivine, orthopyroxene and clinopyroxene from this komatiite liquid yielded a variety of cumulates and residual liquids, the latter probably represented by the komatiitic basaltic flow sequences which are spatially closely associated with the complexes (Anhaeusser, 1985).

The excursion will examine a number of localities showing the nature of the principal rock types (viz., dunite, orthopyroxenite, clinopyroxenite, and anorthositic gabbro-norite units, and some unusual rocks such as nodular harzburgites, rodingites, and calcium-metasomatised gabbros), and the geological and structural setting of the associated chrysotile asbestos deposits present in the area.

A recent proposal by Anhaeusser (2002), following a geotectonic reconstruction, suggests that many of the ultramafic complexes are developed in a zone, believed to be a suture or oceanic collisional zone, along the northern flank of the Barberton greenstone belt (Fig. 24). U-Pb dating of zircons from granitoid rocks in the Barberton region supports the idea that the greenstone belt formed as the result of the amalgamation of at least two arc-like crustal blocks (*c*. 3450 and 3230 Ma old; Kamo and Davis, 1994).

Locality 1. From Badplaas e*n route* to the Stolzburg area a stop may be made to examine the coarse-grained porphyritic phase of the *c.* 3107 Ma Boesmanskop syenite, located just east of the main road from Badplaas to Barberton (R38), and about 1.5 km from the T-junction turnoff to Machadodorp. At this locality the northwest-trending syenite body forms a low whaleback ridge on the farm Kees Zyn Doorns 708 JT. Large (up to 15mm) oscillatory zoned (rapakivi-like) crystals of perthitic feldspars occur in a matrix consisting of microcline, plagioclase, hornblende and biotite. Some quartz may be present in places, in addition to epidote, apatite, sphene, magnetite and zircon.

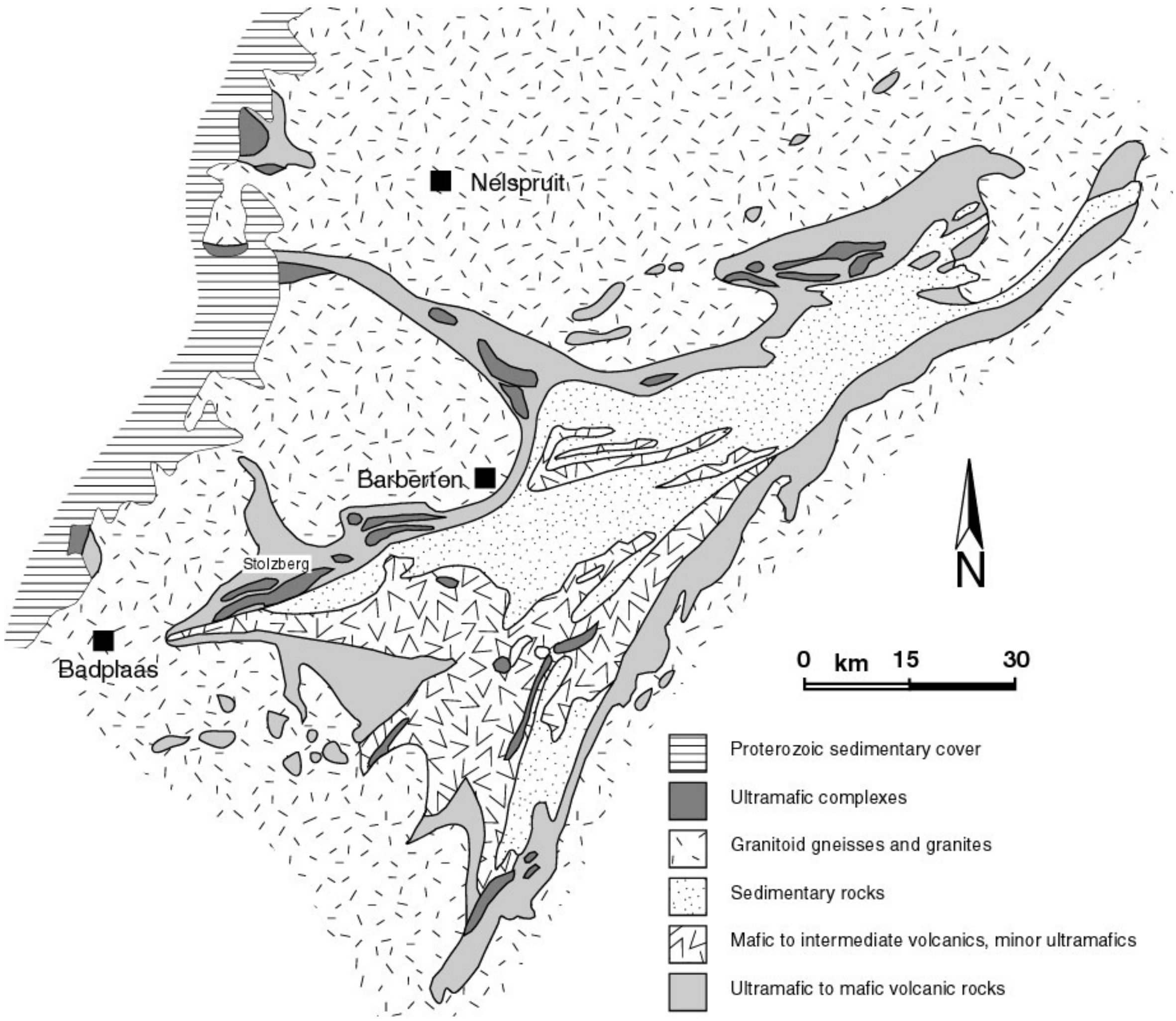


Figure 24: *Locality map showing the general geology of the Barberton greenstone belt and the distribution of layered ultramafic complexes (after Anhaeusser, 1985). The Stolzburg Complex, to be visited on this excursion, is situated approximately 16km northeast of Badplaas.*

Locality 2. A second stop along the main road, just beyond the Gladdespruit bridge and about 300m from a T-junction intersection (unpaved road signposted to Sterkspruit and heading east to Stolzburg), may be made on one of the more interesting mafic (diabase) dykes that trend in a NW-SE direction throughout this area. The dyke, which intrudes into the *c.* 3236 Ma (De Ronde and Kamo, 2000) Nelshoogte trondhjemite gneiss pluton, contains numerous inclusions of granitoid material stoped from the dyke sidewalls. This granitic material has undergone partial remelting resulting in the development of graphic textures in the fragments and, in places, the partial silicification of the mafic dyke.

Locality 3. Altered serpentinites of the Lower Division of the Stolzburg Complex (Fig. 25). At this locality Proterozoic dykes may be seen cutting through the complex and the adjacent Nelshoogte schist belt, forming negative weathering features in the complex and positive ridges in the Nelshoogte schist terrane. The serpentinites have a yellowish-green colour, are strongly deformed with a flat-lying foliation, show cumulate olivine textures, signs of chrysotile asbestos veining, and veins of magnetite derived from the breakdown of the olivine

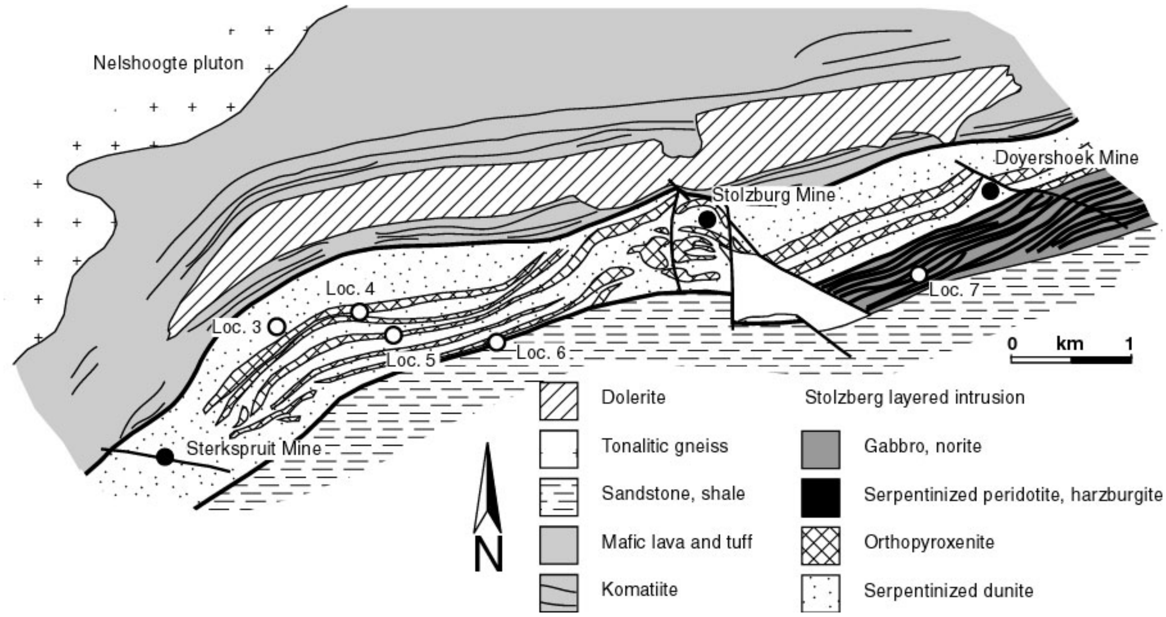


Figure 25: *Geological map of the Stolzburg layered ultramafic complex (after Anhaeusser, 1985) showing the cyclically repetitive nature of the subvertically dipping differentiated sequence which youngs to the southeast. The Lower Division of the complex (localities 3-6) comprises dunites and orthopyroxenites and the Upper Division (locality 7) consists mainly of harzburgites, ortho- and clinopyroxenites, gabbros and norites.*

following serpentinization. Abundant magnetite float can be seen in the roadway at this stop. Some late silicification has also produced localized patches of opal in the altered dunites.

Locality 4. Fresh, cumulate-textured orthopyroxenite. This locality along the roadway presents one of the few localities where fresh orthopyroxene can be guaranteed. Elsewhere in the complex the orthopyroxene has generally been steatized to talc and bastite, and finding freshly preserved orthopyroxene can be a hit and miss affair.

Locality 5. A section across the Lower Division showing alternating dunite and orthopyroxenite layers. The section can be viewed from the roadway looking in southwesterly direction. The dunites are grey in colour and are overlain in each cycle by reddish-brown orthopyroxenite units. At least six cycles of dunite and orthopyroxenite occur in this part of the complex which youngs to the southeast.

Locality 6. View point on the roadway near dumps of the now defunct Stolzburg Mine looking northeast. In the far distance can be seen alternating dunite-orthopyroxenite layers that form distinct ridges that extend northeastwards, parallel to the Mawelawela River valley. The traverse to be undertaken at Locality 7 extends to the east of these ridges. Closer at hand are the old workings of the Stolzburg chrysotile asbestos mine. The asbestos formed in the dunites close to the contacts with the orthopyroxenite layers, which acted as more competent massive rock units relative to the more ductile serpentinized dunites. The asbestos development can clearly be shown to be structurally controlled here and throughout the Barberton district (Anhaeusser, 1976b).

Locality 7. Traverse across the Upper Division of the Stolzburg Complex and rodingite zone. The traverse starts at the confluence of the Mawelawela River with a tributary stream draining eastwards from the Nelshoogte schist belt, which is developed to the northwest. The early part of the traverse crosses the dunite-orthopyroxenite layers seen earlier from Locality 6. For ease of walking the pathway leads across a wide zone of dunite (grey, cumulate-textured rocks with a subvertical pseudostratification, in which are developed chrysotile asbestos fibre veinlets). The contact with the overlying orthopyroxenite layer is amazingly sharp – one foot can be on the dunite and the other on the orthopyroxenite.

The overlying orthopyroxenite shows a coarse-texture and the rock consists of monomineralic cumulate orthopyroxene crystals, with very minor intercumulus plagioclase and the odd speck of chromite. The orthopyroxene is invariably steatized to talc and bastite and geochemically shows very little difference between samples taken from the base, centre or top of the approximately 100-150m thick unit.

The orthopyroxenite unit forms a prominent ridge, which then drops down eastwards onto more grey serpentinized dunite of the following cycle. In this dunite, dykes of rodingite are common, some being as thin as 5-10 cm and others a few metres wide.

The rodingites have a buff or pale flesh-pink coloured appearance and show chill contacts and garnet-rich layers (grossular and hydrogrossular), and contain a host of calcium-rich minerals including the garnets, vesuvianite, hibschite, diopside, nephrite, prehnite, zoisite and many others (Anhaeusser, 1979).

The traverse continues in a northeasterly direction (oblique to the trend of the layering) across calcium metasomatised rocks of pyroxenitic to gabbroic composition before reaching an unusual nodular harzburgite. The rock displays spectacular nodular features that consist of fresh olivine poikilitically enclosed in orthopyroxene that is partly altered to bastite.

East of the nodular harzburgites coarse-textured gabbroic rocks can be seen on a ridge before the ground drops away into the valley below. In this valley is the continuation of the Belvue Fault, one of the faults that bounds the Stolzburg Complex and separates it in this area from the overlying Fig Tree and Moodies sediments to the east.

Option 4.2. Msauli Chert, Barberton greenstone belt (S25°54.513', E30°55.513')

Guide: Axel Hofmann

Objectives: Accretionary lapilli beds, altered ultramafic rocks, and chert dykes at the Msauli Chert type section (*The section is located on the property of Taurus Estates, and access is via Sappi Forest roads; permission is required to visit these outcrops. Please avoid sampling in the gorge.*)

Msauli Gorge. The contact between the Onverwacht and Fig Tree Groups is locally occupied by an unusual chert unit that consists of silicified beds of accretionary lapilli, termed Msauli Chert (see also stop 2.1, locality 12). The type section is situated in a small gorge along the Msauli River. The section starts with a lenticular-banded, green fuchsite-chert-carbonate rock that represents altered ultramafic rocks at the top of the Mendon Formation (Fig. 26). The lenticular fabric is produced by lenses of translucent chert that are sub-parallel to the stratification of overlying sedimentary rocks and gives the rock a gneissic appearance (Fig.27a). The rock is transected by black and botryoidal chert dykes, which postdate formation of the lenticular fabric. The contact between altered volcanic rocks and the Msauli Chert is generally very sharp, but occupied by a vein of black chert at this locality.

The gneissic appearance of alteration zones at the top of ultramafic volcanic rock sequences led De Wit (1982) to interpret such horizons as shear zones. At this locality, however, the contact between the altered volcanic rocks and the chert horizon is undeformed and so are the overlying silicified sedimentary rocks, indicating a non-tectonic origin for the apparent gneissic fabric. Different interpretations for these zones include subaerial weathering of ultramafic flow tops (Lowe and Byerly, 1986), sea-floor alteration (Lowe et al. 1999) and post-depositional metasomatic alteration (Duchac and Hanor, 1987; Hanor and Duchac, 1990).

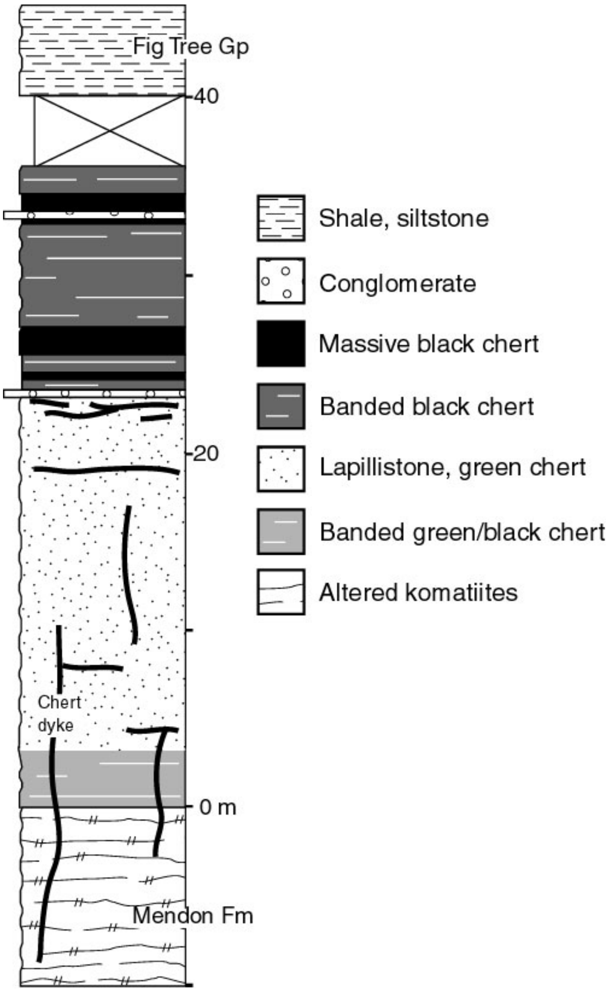


Figure 26: *Section of the Msauli Chert approximately 6 km east of the type section.*

The lower part of the Msauli Chert (*c*. 3 m) consists of thinly interbedded green chert, representing silicified volcaniclastic sand- and siltstone, and black carbonaceous chert (Fig. 27b). This unit is overlain by a c. 20 m thick sequence of accretionary lapilli-bearing, thin to medium, normally graded (Fig. 27c), and rarely cross-stratified beds intercalated with laminated green chert (refer to Heinrichs, 1980, and Lowe, 1999b) for a detailed description of this section). Accretionary lapilli (Fig. 27d) range from sand to fine pebble size, are massive or concentrically laminated, and consist of silicified ash (Lowe, 1999b). The graded beds commonly consist of accretionary lapillistone at the base and komatiitic tuff at the top. Current and climbing-ripple lamination, as well as cross-bedding, are common in the middle to upper parts of many beds. The graded beds show partial or complete Bouma sequences and have been interpreted as turbidites by Stanistreet et al. (1981) and Heinrichs (1984). On the other hand, Lowe (1999b) interpreted the graded beds to represent shallow-water deposits and suggested that the Bouma sequence reflects the declining rate of pyroclastic fall into flowing water rather than declining velocities of turbidity currents. A shallow-water interpretation was further based on the abundance of current structures, the lack of intraclasts and broken lapilli, the apparent lack of erosion of underlying beds, and the distinctiveness of accretionary lapilli within individual sedimentation units in different sections (Lowe, 1999b).

The Msauli Chert is transected by black and botryoidal chert veins and dykes mostly parallel and perpendicular to bedding (Fig. 27b). In many cases stratiform veins are botryoidal and formed later than the black chert dykes. Some dykes are represented by a network of mm- to cm-scale veins that surround angular host rock fragments, with large fragments being unrotated and almost *in situ*. Other dykes show an irregular lamination parallel to the dyke walls. The Msauli Chert is overlain by about 5 m of rather massive, but poorly exposed black chert, followed by variably silicified graphitic shale of the Fig Tree Group. Chert dykes apparently terminate at the massive black chert horizon. Microfossils showing apparent cell division have been reported from carbonaceous rocks from the Msauli Chert (Knoll and Barghoorn, 1977).

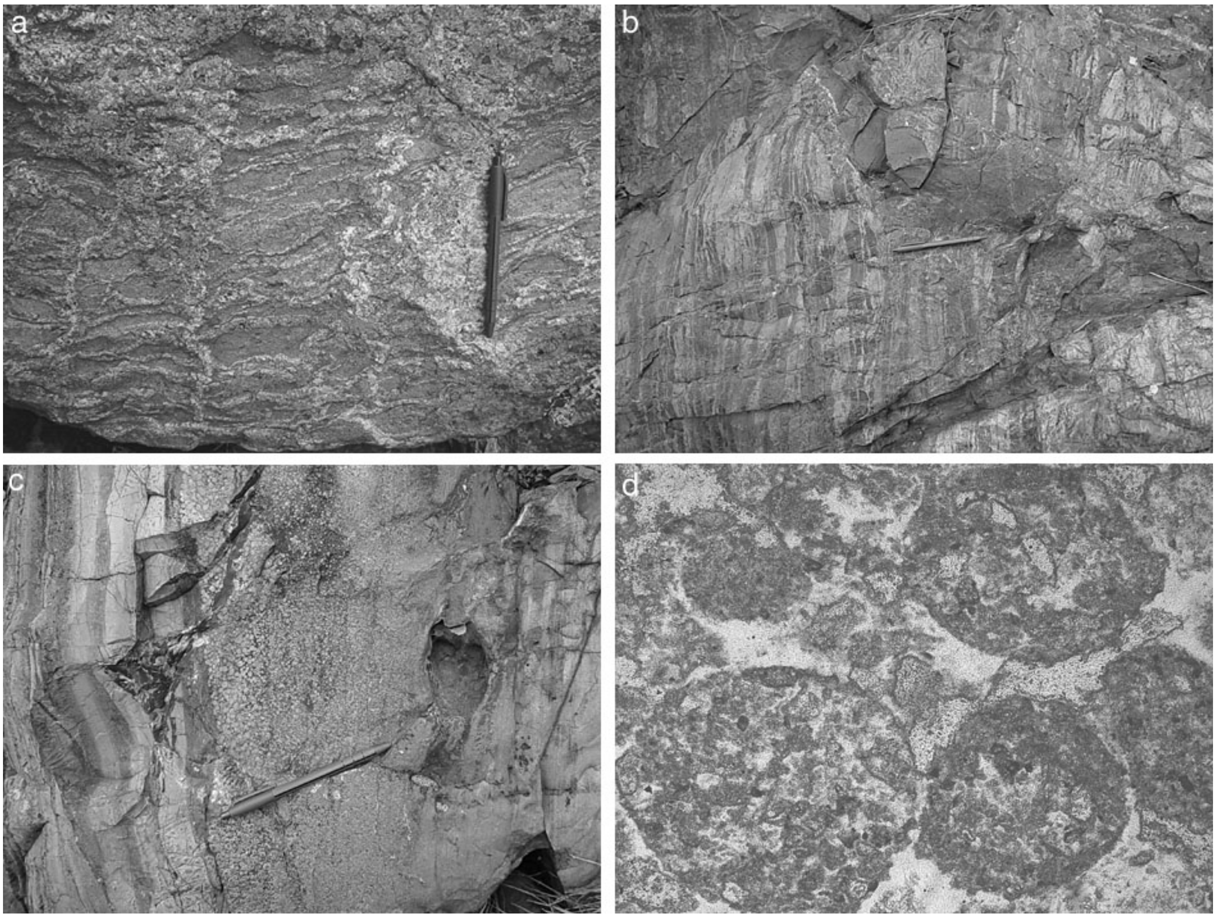


Figure 27: *(a) Altered ultramafic rocks of the Mendon Formation showing gneissic banding. (b) Banded chert of the lower part of the Msauli Chert cross-cut by a chert dyke. (c) Graded bed of accretionary lapilistone; stratigraphic way-up is to the right. (d) Photomicrograph (plane-polarized light) of silicified accretionary lapilli.*

Option 4.3. Stromatolitic dolomites of the 2.5 Ga Malmani Subgroup, Transvaal Supergroup

Guide: Ken Eriksson

Exposures along R539 to Sabie traverse through the Oaktree, Monte Christo and Lyttleton Formations of the Malmani Subgroup (Fig. 28). These exposures represent supratidal-intertidal lithofacies and consist of domal stromatolites, micrite, grainstones with and without ooids, intraclast breccias, wave and interference ripples, tepee structures, and crystal pseudomorphs. Giant elongate stromatolites of the Lyttleton Formation are exposed at the top of the pass and represent the deepest water facies of the Malmani Subgroup (Fig. 29); Truswell and Eriksson, 1974; Eriksson, 1977b).

Locally, layers of radiating crystal pseudomorphs between 3 and 30 cm thick are preserved in chert and less commonly in dolomite. The precursor mineral to the pseudomorphs nucleated as botryoids on the sea floor and grew upwards as radiating fans. These fans are interpreted as neomorphosed aragonite (Sumner and Grotzinger, 2000). Rare, 1 cm cubic pseudomorphs consisting of chert or void-filling dolomite indicate the former presence of halite that grew displacively in micrite. Gypsum pseudomorphs have not been observed in the Malmani Subgroup and rarely in stratigraphic successions younger than 1.8 Ga. The absence of any indication of the former presence of gypsum or anhydrite supports previous contentions of Grotzinger and Kasting (1993) that the Neoarchaean ocean was deficient in sulphate related to an anoxic atmosphere, or that it contained a high bicarbonate to calcium ratio such that with progressive evaporation, all calcium was consumed before the gypsum stability field was reached.

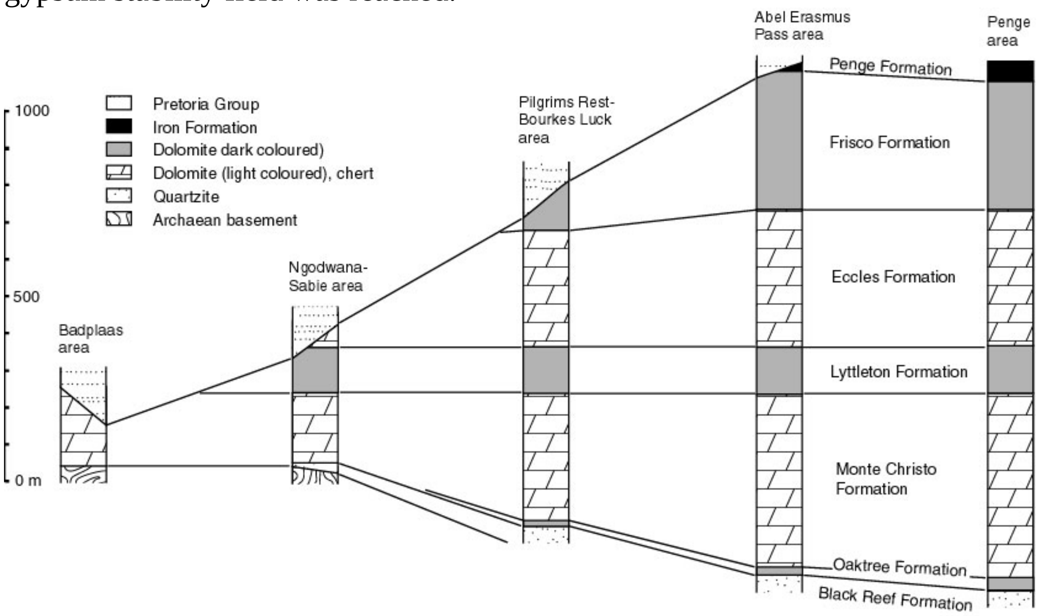


Figure 28: *Subdivision of the Malmani Subgroup along the eastern escarpment of Mpumulanga showing the erosional unconformity at the base of the Pretoria Group (from Button, 1973).*

Figure 29: *Giant, elongate stromatolitic domes from the Lyttleton Formation along R539. Note person for scale in bottom right of photograph.*

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