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APPENDIX P:
SELECTED GEOLOGICAL FIELD GUIDES

BARBERTON – MAKHONJWA MOUNTAINS WORLD HERITAGE SITE PROJECT

Selected Geological Field Guides

by

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Version 1.1

Selected Geological Routes Used for Field Trips and Guided Tours in and around the World Heritage Site

1 Bulembu Road Section – "Barberton-Makhonjwa Geotrail Geosites and View Points" by Tony Ferrar and Christoph Heubeck (2013)

For a full account and detailed descriptions of this route the reader is referred to the published guide book, included in the Nomination Dossier.

2 Komati-River Valley

From Guide Book by Carl R. Anhaeusser

Overview. The drive east from Badplaas takes this excursion across generally flat-lying granitic terrane made up almost exclusively of trondhjemitic gneisses of the Badplaas and Stolzburg plutons. 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.* 3,445 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 Swart Rand dyke (Black Ridge), which trends in a NNW-SSE direction traversing the southern part of the Tjakastad township. The housing seen at Tjakastad 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 the water reservoirs seen on the hill immediately above the town (Fig. 1).

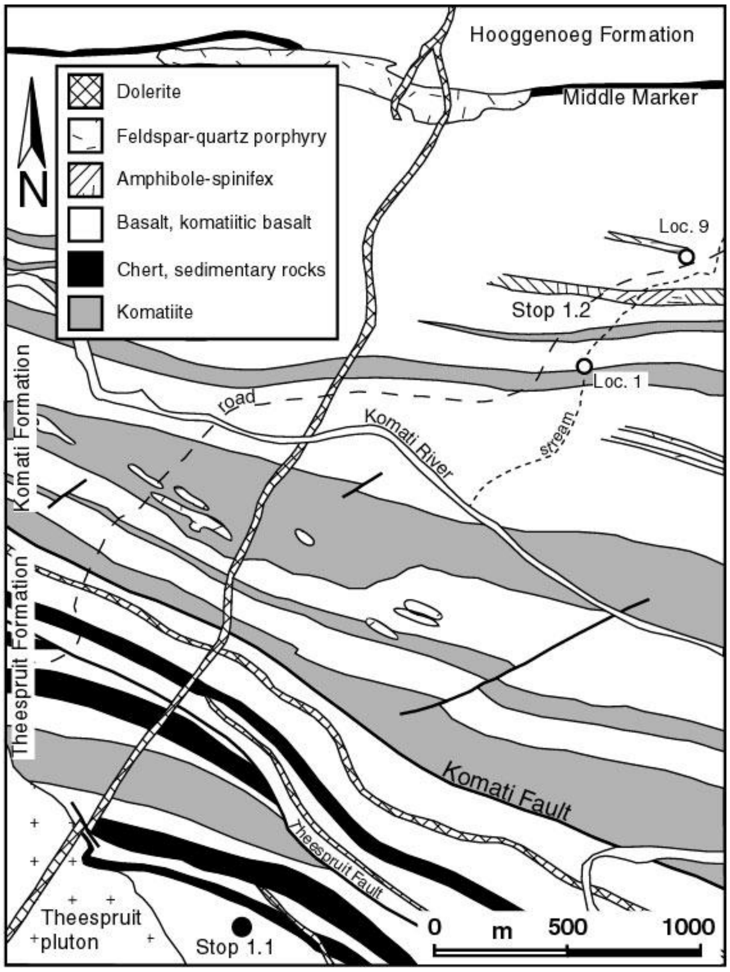


Fig. 1 Simplified geological map of the Theespruit and Komati Formations northeast of Tjakastadt (modified after Viljoen and Viljoen, 1969b).

Tjakastad water reservoir exposures of the Theespruit Formation and vantage point to view the geological setting of the type locality of the Onverwacht Group.

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.

The view northwards from the reservoir stop shows the main type locality sections. In the immediate foreground and at the reservoir stop are rocks of the Theespruit Formation. Down the slope towards the Komati River a white vein quartz filling of the Theespruit Fault can be seen trending east and west internally within the succession. The Komati Fault near the river separates the Theespruit Formation from the structurally overlying Komati Formation. The stratigraphic layering of the type section is seen looking to the northeast. Good exposures may be seen 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 reservoir northwards (and including also the Sandspruit Formation to the south) the entire section, which consists of a subvertical rock succession, is estimated to be approximately 15 km thick (Viljoen and Viljoen, 1969a).

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 2.

The second objective at Stop 1 is to examine one of the characteristic units of the Theespruit Formation, viz., one of the many interlayered felsic units that led to Viljoen and Viljoen (1969b) separating the Theespruit succession into a separately recognizable entity or formation. The type locality section is located approximately 2 km to the east of the reservoir 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 dominantly 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 alga-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 alga-like structures in the Onverwacht as being “dubiomicrofossils”.

Locality: S25º59.456’, E30º49.500’

Stream section northeast of Tjakastad displaying 3,480 Ma komatiites and komatiitic basalts of the Komati Formation.

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 therefore commences over 1.5 km from the base of the 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. 2) 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 in the flow unit. This is overlain by a zone of bladed or plate spinifex-textured olivine together with some clinopyroxe (A2) which, in turn, is overlain by a zone of random spinifex (Fig. 3a) that grades 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 while 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 the younging direction to the north be determined.

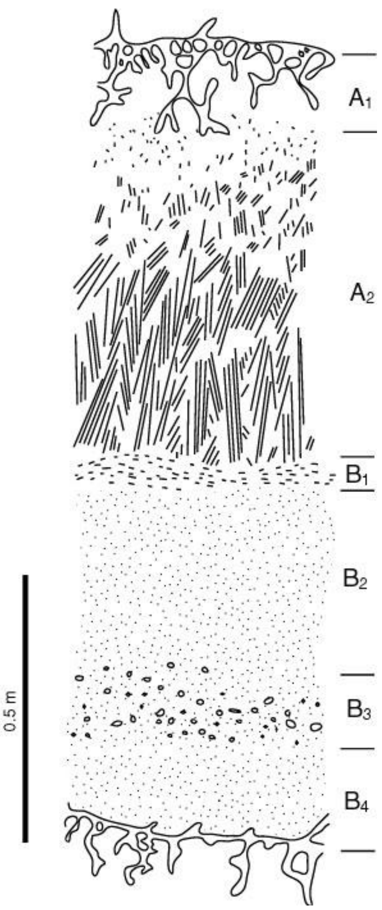


Fig. 2. Diagrammatic section of a typical komatiite flow (after Pyke et al., 1973).

Locality 2. Approximately 30m north of locality 1, on the west bank of the stream, are exposures of komatiitic pillow basalts and somewhat 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 komatiic pillow basalts (Fig. 3b). 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 young direction to the north to be ascertained. The rocks have a typical green colour (greenschist metamorphic grade) and show very limited or hardly any effects 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. 3c) 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 in the type locality. The spherulitic structures are considered to represent immiscibility features and are commonly associated with pillow-flow transitions.

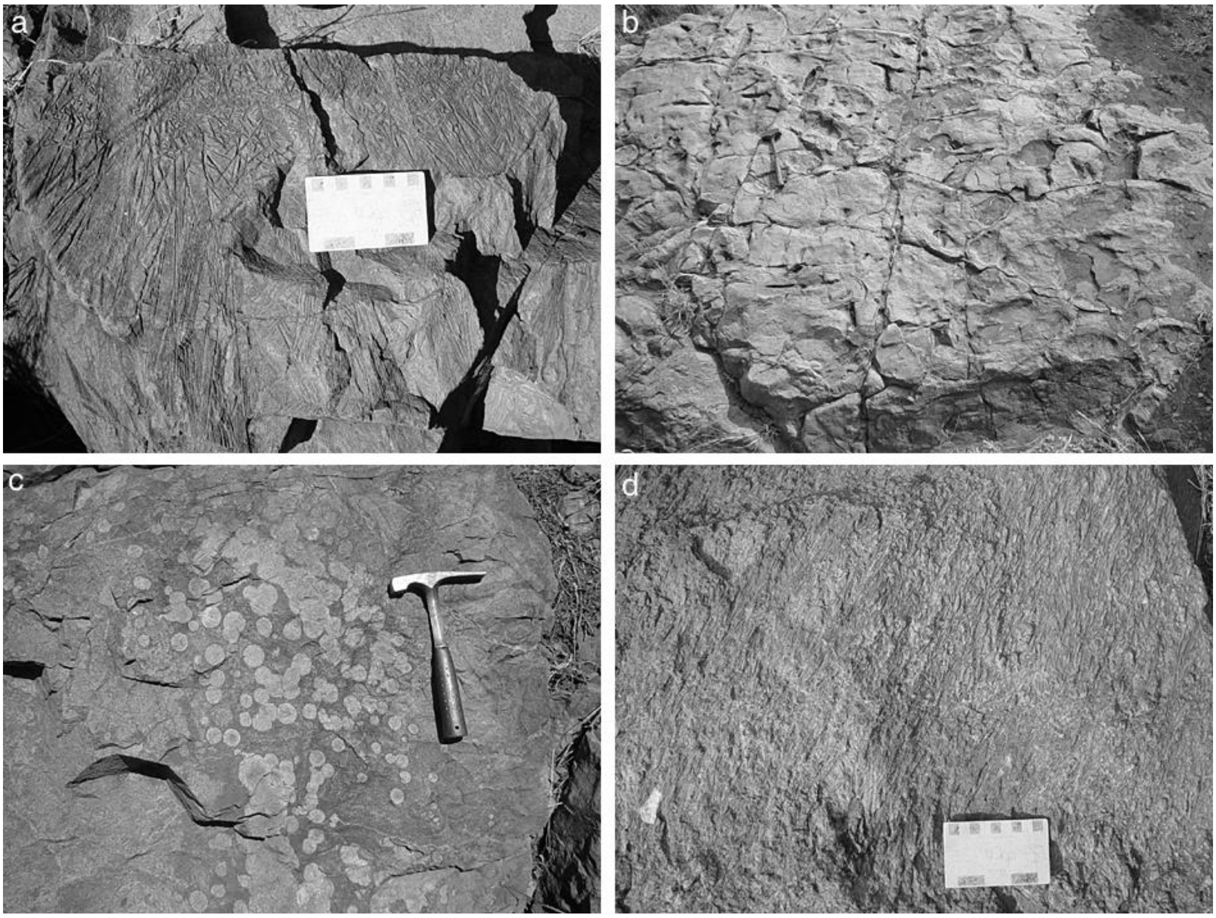


Fig. 3 (a) Komatiite showing bladed spinifex-texture (base) and random spinifex-texture (top). Photo from Spinifex Stream. (b) Rock pavement of komatiitic pillow basalt at locality 3. (c) Ocelli in massive basalt or lomatiitic basalt. Photo from Spinifex Stream. (d) Amphibole spinifex-texture in komatiitic basalt at locality 6.

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 sometimes carbonate-filled 'drain-away cavities' (cavities formed when the lavas were horizontal and when the lava was able to flow out of the pillows leaving behind tubular gas chambers), which are often stacked above one another in 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 are do not consist of olivine and clinopyroxene, but are rather made up of radiating, generally actinolitic, amphibole needles (Fig. 3d).

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.

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 away 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 1). The favoured interpretation is that the dyke represents a plagiogranite similar to those reportedly encountered in modern oceanic settings, and more particularly with plutonic parts of ophiolites.

Table 1 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

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 represent 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 1800s 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 that was sunk. 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. The next stop will be made on the farm Aarnhemburg 155IT.

Locality: S 25º 58.188', E30º 50.065'

Roof pendant in the Theespruit pluton on the farm Aarnhemburg 155IT.

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 succession and act as roof pendants or greenstone rafts. Outcrop 1 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., Sandspruit Formation).

Outcrop 1. The exposure 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 which show a range of internal features, including spherulitic or ocelli structures, vesicles, chilled margins and re-entrant rims (Fig. 4). The pillows vary considerably in size, some as big as a football and others 1-2 metres in length. Inter-pillow quartz and carbonate is well displayed 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 linked to invasive diapiric trondhjemite gneiss plutons like those found elsewhere in the neighbouring 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.

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.

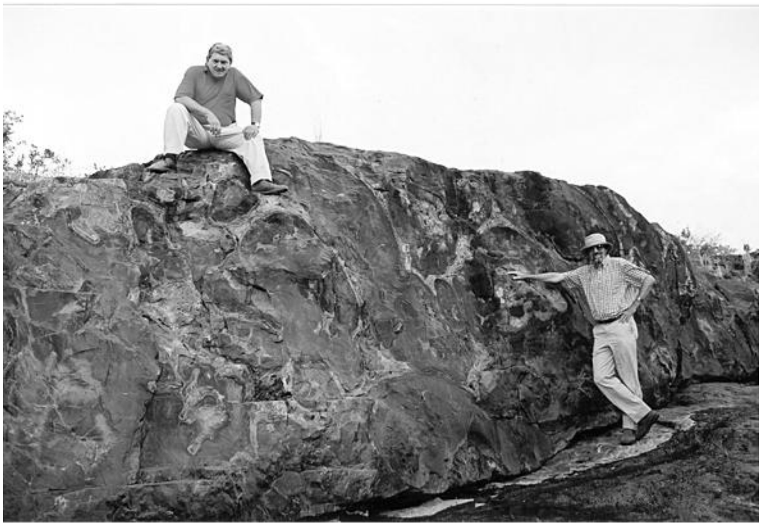


Fig. 4 Amphibolite facies komatiitic basalt with pillow structures, Theespruit pluton roof pendant.

Locality: S26º 01.587', E30º 50.230'

Spinifex-textured ultramafic rocks of the Sandspruit Formation (3.5 km due south of Outcrop 1 on the farm Brandybal 171 IT).

Objective. 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 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 associated 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.

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-actinolte 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' canbeseen at this locality.

Sandspruit Formation komatiites and komatiitic basalts and migmatite exposures on the farm Brandybal 171 IT

The area south of the main mass of the Barberton greenstone belt consists of greenstone remnants surrounded and intruded by trondhjemitic grey gneisses similar to that seen in the Theespruit pluton. One such remnant is that best developed on the farm Brandybal 171 IT, which has been regarded as the type locality for the Sandspruit Formation (Viljoen and Viljoen, 1969b). The traverse selected for the excursion 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 the changes that take place to the amphibolites and serpentinites where they are invaded by granitoid rocks. This granite-greenstone interaction has produced a spectacular set of migmatite exposures seen along the banks of the river (Fig. 5).

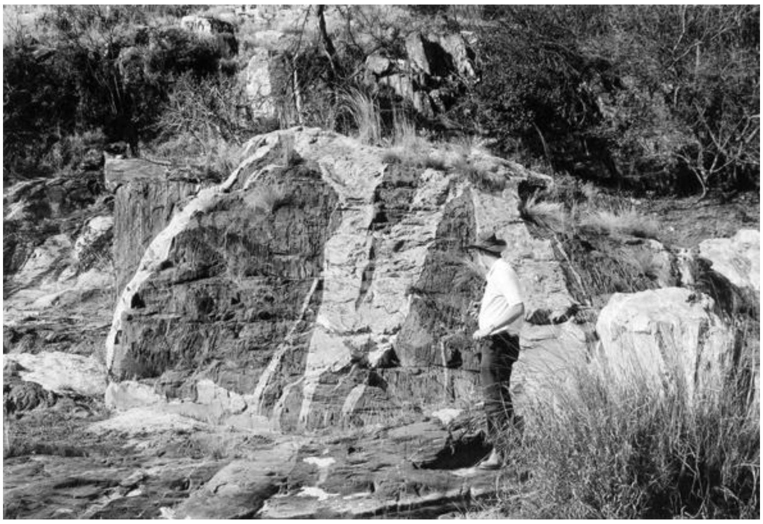


Fig 5 Tightly folded trondhjemite dyke in Sandspruit Formation amphibolite.

The traverse starts 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 hundreds of 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 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 is encountered which shows horizontally orientated columnar jointing. A little further on a second diabase dyke can be seen cutting through the 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. They have not experienced any of the metamorphism and structural disturbance found throughout the district and most 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 about the time the Bushveld Complex was emplaced.

Locality: S26º 04.565', E30º 50.327'

Nederland metasedimentary rocks

Examination of high-pressure amphibolite facies volcano-sedimentary rocks of the Sandspruit Formation and granite-greenstone contact relationships. Geochronological data combined with PT-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-section west of the dirt road (and immediately north of the old Theespruit bridge, Fig. 6), shows 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. The well-equilibrated mineral assemblages comprise diopside, plagioclase, quartz, potassium feldspar and hornblende. Locally, poikiloblastic garnet up to several mm in diameter can be observed. Retrogression in these rocks is dominated by an occasionally extensive replacement of peak metamorphic minerals by epidote and sphene. PT- estimates for the peak metamorphic mineral assemblages vary between 650-700 °C and ca. 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 ca. 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 ca. 3230 Ma.

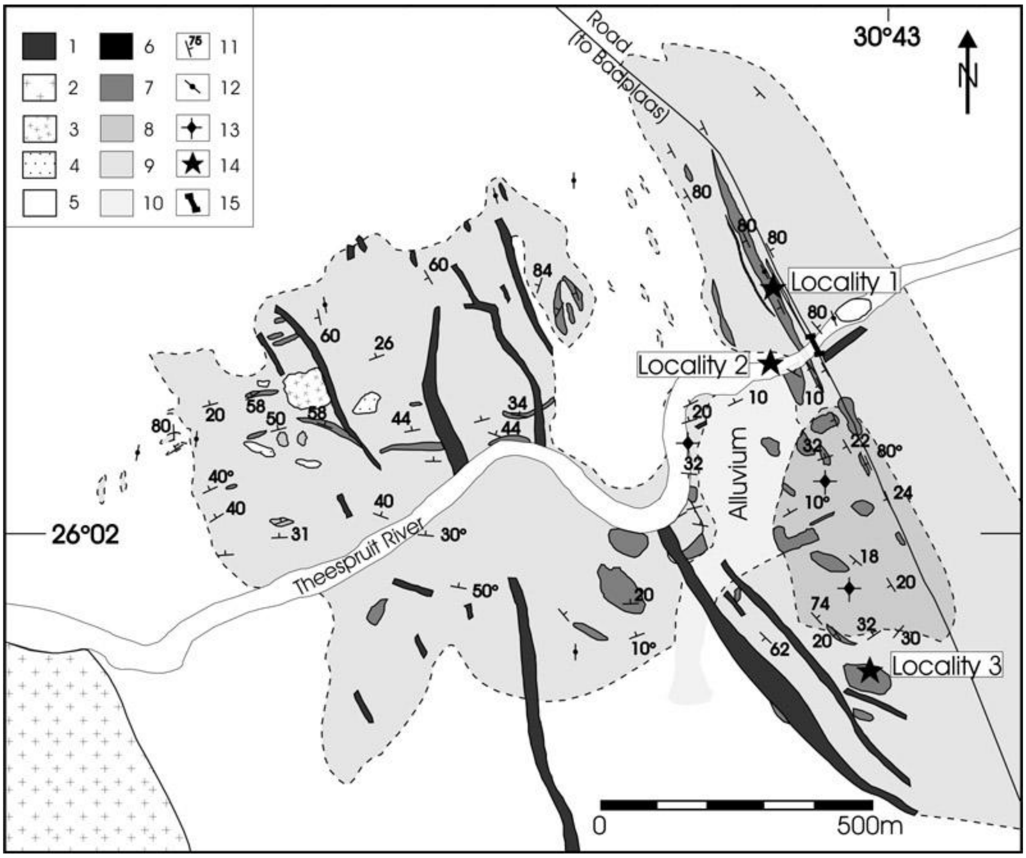


Fig 6 Geological map of the Nederland greenstone remnant and localities visited during the fieldtrip. 1- Mafic dykes; 2- Boesmanskop Syenite, 3- Alkaline intrusions; 4- Pegmatites; 5- Trondhjemitic gneisses; 6- Cherts; 7- Clastic metasedimentary rocks; 8- Ultramafic schists; 9- Amphibolites; 10- Alluvium; 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. 7) occur in the bed of the Theespruit River ca. 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 gives an age of 3431 $\pm$ 11 Ma 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 & Davis, 1994). Another explanation may be that the age of 3431 $\pm$ 11 Ma reflects a slightly younger pulse of magmatism in the area.

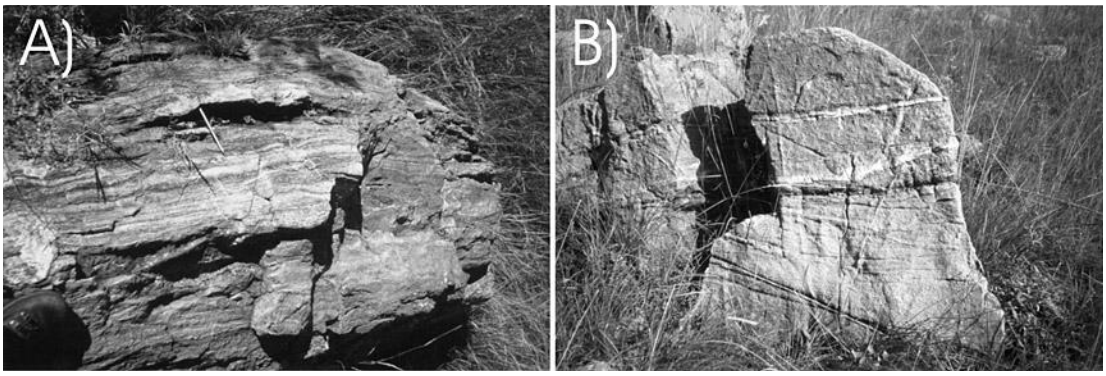


Fig. 7. A) Finely banded and boudinaged clastic metasedimentary rock at locality 1; B) Boulder of relatively undeformed, ungraded and cross-bedded meta-arkose at locality 3.

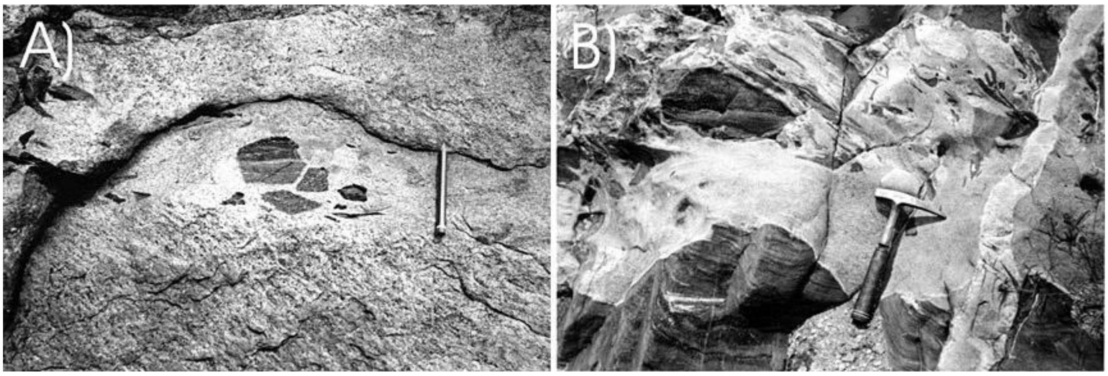


Fig. 8. A) Fine-grained, leucocratic granitoid phase with xenoliths of amphibolite schists intrusive into and crosscutting trondhjemitic gneisses; B) Magmatic brecciation of strongly schistose greenstone material some 100 m southeast of locality 2.

Locality 3. A prominent, oval-shaped body mainly composed of ultramafic schists occurs to the south of the Theespruit river. At its southern margin, we 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. 9). Only 5 grains are concordant within error and imply a range of rock ages in the source area for the clastic sediments. The dates range between ca. 3525 and 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 $\pm$ 11 Ma is given by the late-tectonic trondhjemite. Thus, these metasedimentary rocks were deposited between ca. 3521 and 3431 Ma, contemporaneously with the erosion of spatially associated older, and presumably potassium-rich, granitoid rocks.

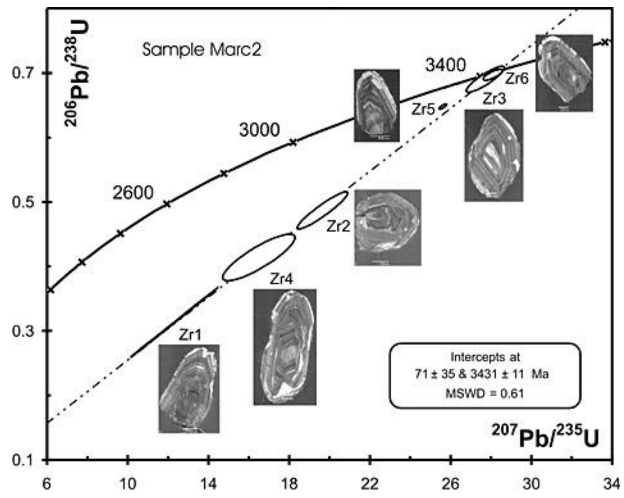


Fig. 9. U-Pb Concordia diagram for a pervasively foliated trondhjemitic gneiss at locality 2. Inset shows data that are 90% concordant or more. Error ellipses are at 2 sigma.

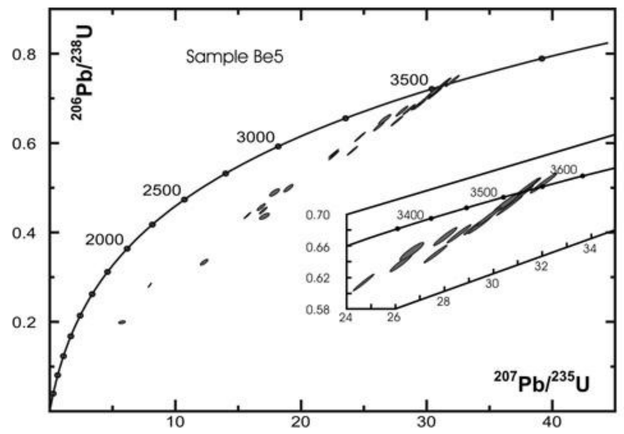


Fig. 10. U-Pb Concordia diagram for detrital zircons in sample BE 5. Error ellipses are at 2 sigma.

3 Songimvelo Nature Reserve (S26°02.170', E30°59.980')

Extracted from Guide Book by 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. 1). The section formed in a time span of ca. 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. 1). Some key issues include the observation of microfossils in some carbonaceous cherts, the origin of alteration zones as shear zones vs Archaean weathering phenomena, and the depositional and tectonic setting of the Kromberg Formation.

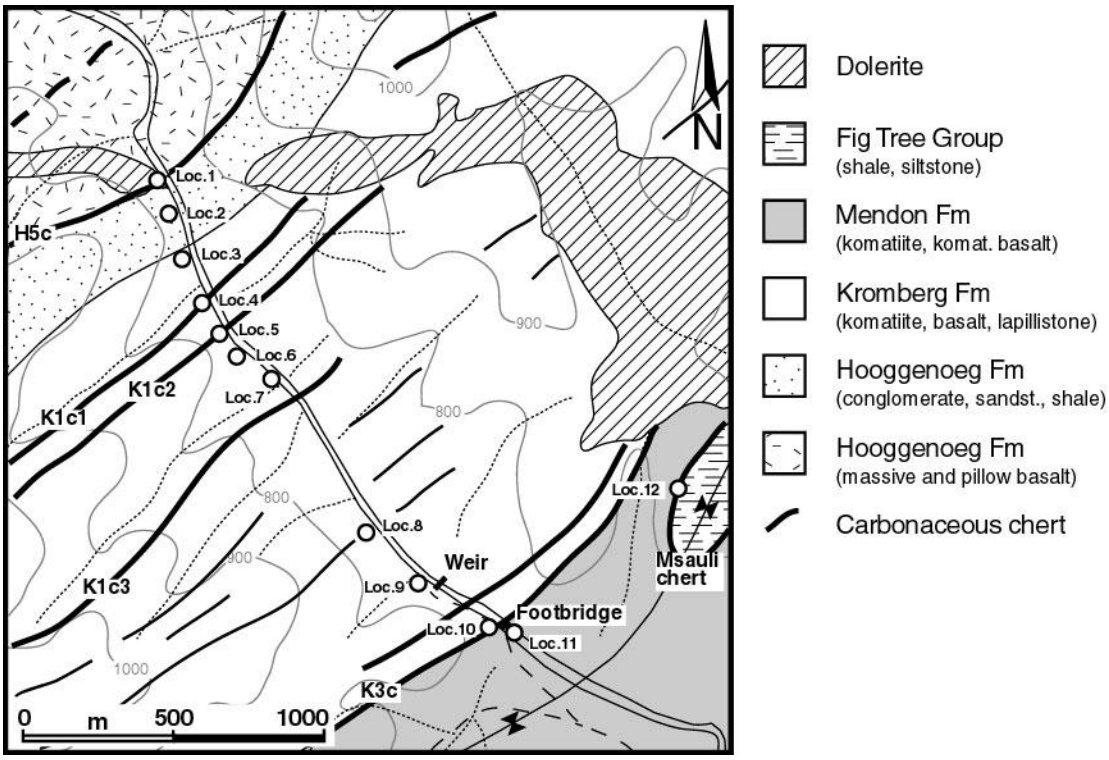


Fig. 1. Geological map of the Komati River section (modified after Viljoen and Viljoen, 1969c).

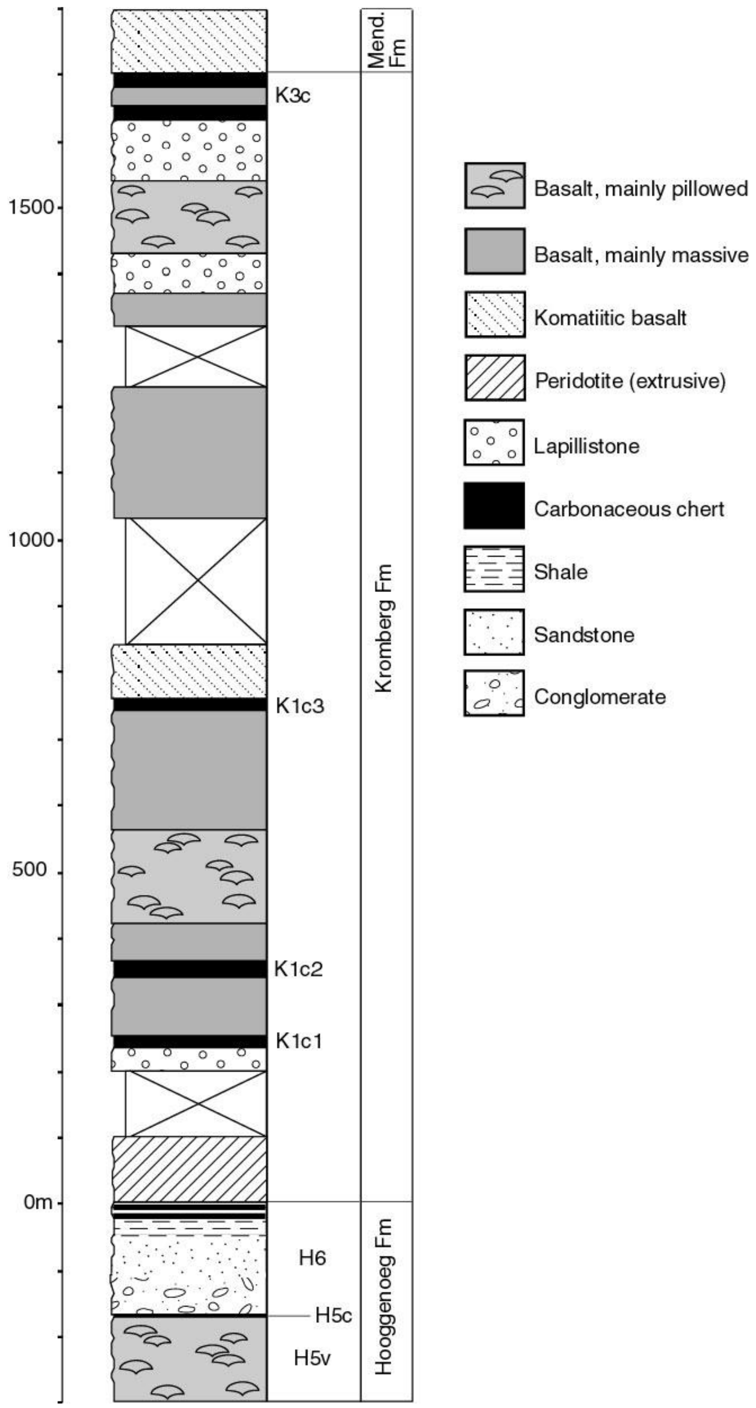


Fig. 2. Stratigraphic log of the Komati River section (modified after Viljoen and Viljoen, 1969c; Lowe and Byerly, 2003).

Locality 1. A few hundred metres thick sequence of pillow basalt and minor massive basalt 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 contains abundant ocelli and, starting from ca. 50 m below the chert bed, become 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). Silicified basalt is transected by a network of massive black chert veins in the uppermost few metres (Fig. 2). Chert-veined basalt is capped along a sharp contact by a ca. 1 m thick horizon (H5c) of massive to thinly laminated black chert that is overlain by laminated grey chert (Fig. 3a). 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 chert dykes underneath are best exposed on the northern river bank.

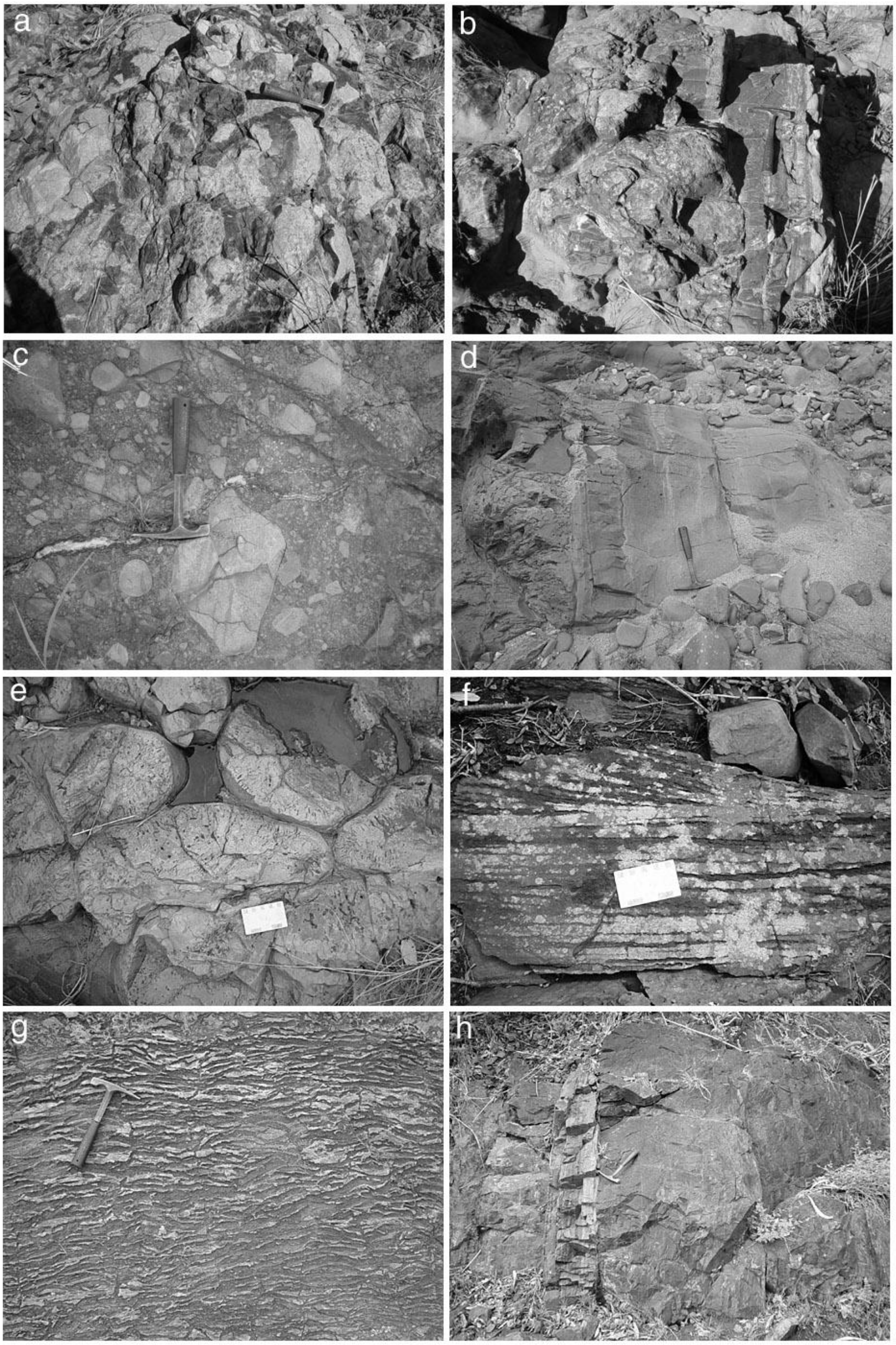


Fig. 3 (a) Network of black chert veins cross-cutting silicified basalt (locality 1, northern river bank). (b) Contact between silicified, black chert-veined basalt and banded black and grey chert of H5c. Stratigraphic way-up is to the right (locality 1, northern river bank). (c) Dacite pebble and boulder conglomerate (locality 2). (d) Thick-bedded turbiditic sandstone and silicified shale. Stratigraphic way-up is to the right (locality 2). (e) Pillow basalt with pipe vesicles (locality 6). (f) Parallel- and through cross-stratified ultramafic lapillistone (locality 8). (g) Fuchsite-chert-carbonate zone representing altered ultramafic rocks (locality 9). (h) Massive and banded black carbonaceous chert (locality 10).

Locality 2. The chert horizon is overlain along a sharp and planar contact by a ca. 170 m thick, fining-upward sedimentary sequence. This unit, termed member H6 of the Hooggenoeg Formation, has been correlated with intrusive and extrusive 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. 3c). 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 volcanic 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 consists of thick bedded turbidites (Fig. 3d) of normally graded sandstone overlain by silicified shale, now chert, which grades into silicified shale upsection.

Locality 3. The sedimentary sequence H6 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 program. Similar ultramafic bodies are widespread in the Barberton greenstone belt and frequently represent intrusive sills. At this locality the ultramafic horizon is overlain by carbonated ultramafic lapillistone, according to Viljoen and Viljoen (1969c) along a gradational contact, signifying an extrusive origin. Also in favour of an extrusive origin, Byerly and Lowe (1991) reported 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 and correlated with the Buck Reef Chert by Lowe and Byerly (1999), is exposed in the Komati River section. The exposed section starts with a lenticular-textured fuchsite-chert-carbonate alteration zone. The alteration zone contains lenses of less altered carbonated lapillistone, suggesting that 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 several metres thick chert horizon (K1c1) that consists of black and white banded chert.

Locality 5. The second chert horizon in the Kromberg section (K1c2, Lowe and Byerly, 1999) is ca. 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 ca. 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. 3e). 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 ca. 20 m thick, carbonated unit of parallel-laminated and through cross-bedded ultramafic lapillistone (Fig. 3f) 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. 3g). Irregular veins of black chert oriented sub-parallel to the fabric are present. The rock is locally transected by shear zones that postdate 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 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.

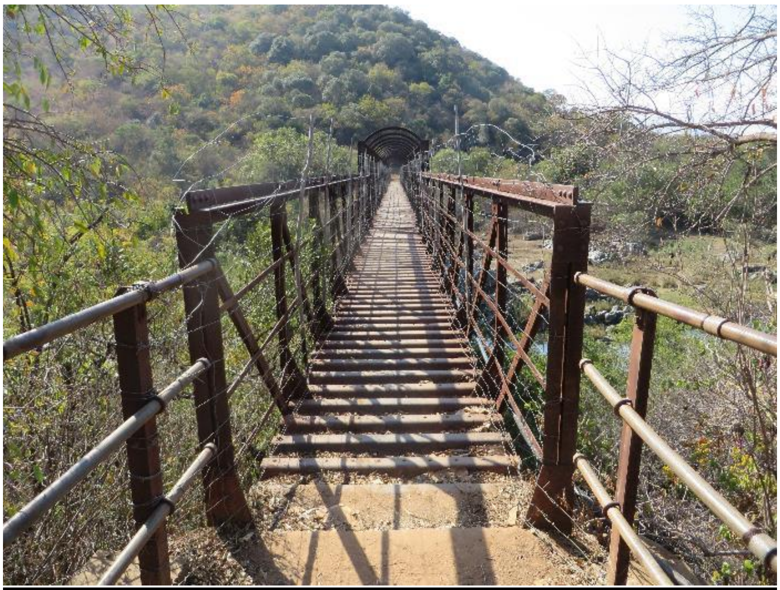
Locality 10. Above the chert-carbonate rock occur two units of massive basalt overlain by banded chert. The cherts may represent a tectonically duplicated single horizon (Lowe and Byerly, 2003), termed K3c or Footbridge Chert by Lowe and Byerly (1999). Very good exposure of the chert can be found a few metres NW of the footbridge that crosses the Komati River. The chert consists predominantly of massive black chert and black and white banded chert (Fig. 3h). 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 et al. (2001) and Westall and Gerneke (1998).

Locality 11. A rock pavement in the Komati River just above the Footbridge Chert exposes the basal part of the Mendon Formation. 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 ca. 800 m NE of the footbridge. 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, a ca. 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.

Field Guide by Dave Mourant

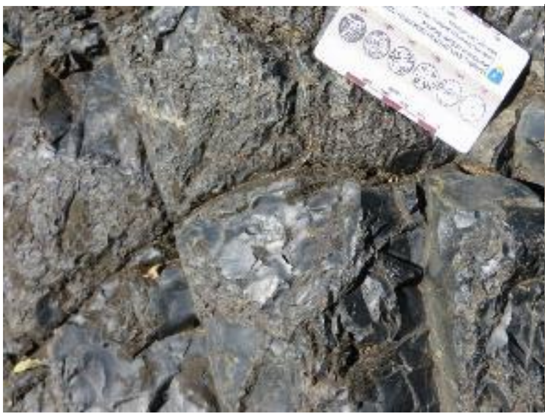
<u>Kromdraai Komati River Traverse</u>



<u>SITE 1</u>

(26°02’13.1”S, 31°00’01.0”E)

A rock pavement in the Komati River just above the Footbridge Chert exposes the basal part of the Mendon Formation. 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.




SITE 2

(26°02’10.2”S, 30°59’58.8”E):

The Footbridge and the Footbridge Chert. Near the path that leads to the footbridge across the Komati River, are two cherts that most likely represent a single unit repeated by local faulting. Viljoen and Viljoen (1969b) suggest that these cherts cap the upper mafic-to-felsic volcanic cycle in the Kromberg Formation. However, Byerly et al. (1983 and unpublished data) suggest that the “felsic” units below these cherts are additional examples of low-temperature flow-top alteration zones at the tops of basaltic lava flows and beneath sedimentary chert layers.

Good exposures of the chert occur a few metres NW of the footbridge crossing the Komati River. The rock consists predominantly of massive black chert and black and white banded chert. Zircon from a dacitic tuff layer in the chert yielded an age of 3334±3 Ma (Byerly et al., 1996). Possible microfossils were described from this layer by Westall et al. (2001)

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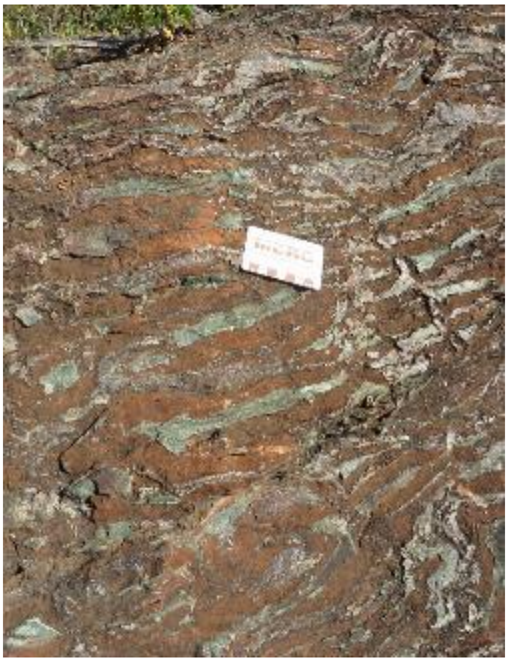
SITE 3

(26°02’10.62”S, 30°59’51.6”E):

Quartz-carbonate-fuchsite zone. An interval of complex and distinctive rock composed of alternating stratiform bands of carbonate and silicified volcanic or volcaniclastic rock. The silicified volcanic rock is green in color due to the presence of fine chrome-rich sericite or fuchsite. Stylolites are common as are stratiform and cross-cutting veins of megaquartz. Magmatic chromite is present in some layers and several septa of silicified spinifex texture suggest this zone was once a komatiitic lava flow.

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 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 attributed to low-temperature hydrothermal seafloor alteration (Hofmann and Harris, 2008). This zone is similar to those studied by Lowe and Byerly (1986b) who considered them to have formed initially as low-temperature flow-top alteration zones. Duchac and Hanor (1987), who studied a similar zone in the Mendon Formation, also suggested that they form at low temperatures but within hydrothermal systems.



SITE 4

(26°02'7.8"S, 30°59'48.0"E):

A ca. 20 m-thick, carbonated unit of parallel-laminated and trough cross-bedded ultramafic lapillistone occurs 100 m upstream from the weir. This is intercalated with massive and pillow basalt and is part of a much thicker lapillistone sequence about 100 m thick of poorly exposed mafic lapilli tuffs. Along the trail are several low-lying outcrops of well-bedded, locally cross-stratified lapilli tuff.

This unit probably correlates with a similar unit on the west limb of the Onverwacht Anticline that is up to 1000 m in thickness. The coarser grained layers are composed of carbonate-chlorite-quartz with minor rutile, and fresh magmatic chromite. The presence of chromites and very high chromium bulk rock compositions suggest these rocks may be derived from komatiitic rather than basaltic eruptions. The cross-bedding seen here is probably due to reworking by marine currents, though several workers have suggested that some of these cross beds could represent deposition within base surges associated with eruption and deposition during phreato-magmatic eruptions.

SITE 5

(26°02’3.48”S, 30°59’42.42”E):

“Felsic” top of a mafic-to-felsic volcanic cycle. Low outcrops are composed of black chert and green silicified mafic volcanic rock. Considered by Viljoen and Viljoen (1969b) to be the top of a mafic-to-felsic volcanic cycle, we interpret this rock to be flow-top alteration zone developed on a basaltic or komatiitic flow sequence (Lowe and Byerly, 1986b). At least 20 m of altered volcanic rock can be seen here. Fine quenched textures, vesicles, and silica pseudomorphs after olivine phenocrysts can be recognized with a hand lens. The green colour of the chert and altered volcanic rock is due to the presence of chrome-rich sericite or fuchsite. Unaltered magmatic chromite is also present, even in the most altered materials, including the chert.


<u>SITE 6</u>

(26°02'02.2"S, 30°59'42"E

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.


<u>SITE 7</u>

(26°02'00.4"S, 30°59'40.3"E)

Flow top breccia

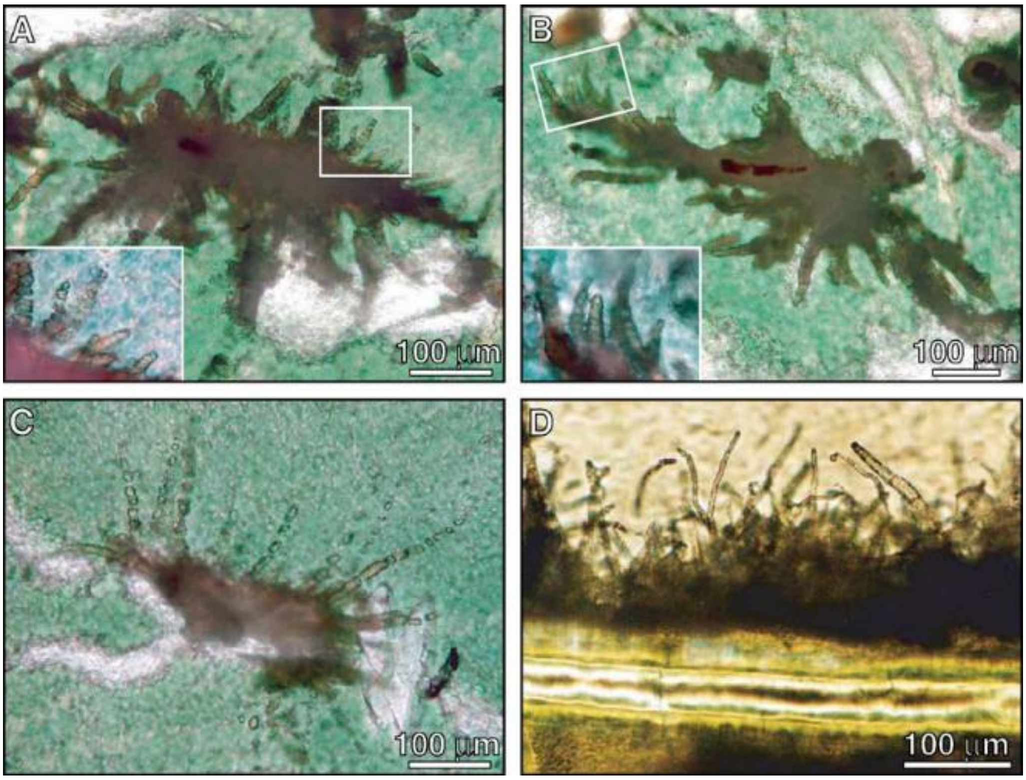
SITE 8

(26°01’45.24”S, 30°59’28.14”E):

Tholeiitic pillow basalts. Beautifully developed pillows occur about 500 m above the base of the Kromberg Formation within a sequence of basaltic lava flows. Minor massive and brecciated units are also present within this zone. All are composed of similar Fe-rich tholeiitic basalt. The pillows display classical structures and shapes used to recognize younging directions within flows.

Pillows show characteristic convex-upward tops and cusped bottoms and contain well-developed radial pipe vesicles and contraction cracks. Fine-grained textures of once-glassy selvages, radial cooling joints, and radial pipe vesicles are also present. Rims of pillow basalts and hyaloclasts from the Kromberg Formation contain micrometre-scale tubular structures interpreted to represent corrosion features that formed by ancient microbes (Furnes et al., 2004; Banerjee et al.,2006



Healed fractures within pillow rims displaying irregular patches consisting of fine-grained titanite (brown mineral in A–C). Detail of tubular structures within the white boxes in A and B are shown in the insets. Some of these tubular structures exhibit well-defined segmentation where they have been overprinted by chlorite, indicating that they predate the alteration process (C). Modern microbial tubular structures in basaltic glass are shown for comparison (D). Note the similarity in size, shape, and distribution between the modern and ancient tubular textures. Source: Banerjee et al. (2006).

SITE 9

(26°01’42.54”S, 30°59’24.84”E)

The second chert horizon in the Kromberg section (K1c2) is ca. 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 sparse fine pale green komatiitic volcaniclastic layers, 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 and features similar to modern bacteria and the organic mats they produce (Walsh and Lowe, 1985, 1999; Walsh, 1992). A shallow-marine environment of deposition has been proposed for this chert horizon (Lowe and Knauth, 1977).

SITE 10

(26°01’38.8”S, 30°59’22.6”E)

A distinct banded chert horizon near the base of the Kromberg Formation, termed K1c1 is exposed in the Komati River section. The exposed section begins with a green chert-carbonate alteration zone that is overlain by carbonated, stratified ultramafic lapillistone. Stratiform botryoidal and massive black chert veins less than 5 cm thick are common in both units. In the upper part of the section, the chert veins become more numerous, which results in massive black chert that contains matrix-supported lapillistone fragments near the top. Lapillistone becomes progressively more silicified upsection. This is followed by a black and white banded chert (K1c1) several meters in thickness. At least two generations of carbonaceous matter are present in the chert.

<u>SITE 11</u>

(26°01’41.4”S, 30°59’22.2”E):

Iron-rich tholeiitic basalts. At about 300 m above the base of the Kromberg Formation is a thick, massive basalt similar in composition to many other flows throughout the next 300 m of section. Thin units of coarse-grained pyroxene-rich rock within this interval may represent somewhat younger intrusive rocks.

<u>SITE 12</u>

(26°01’38.94”S, 30°59’22.12”E):

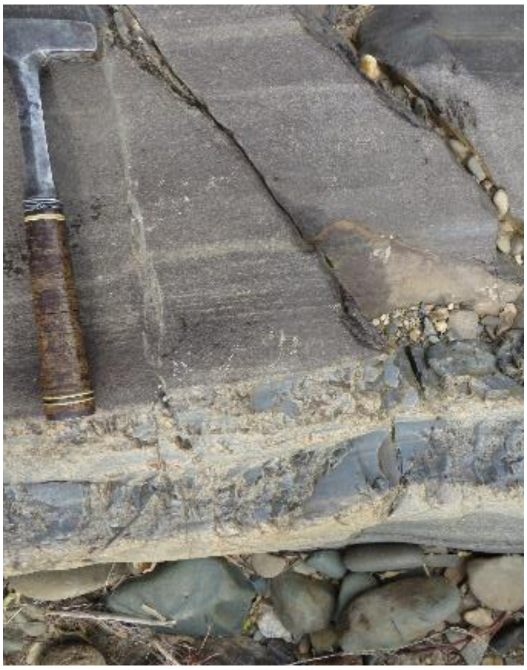
The sedimentary sequence H6 of the Hooggenoeg Formation is overlain along a sharp, but poorly exposed contact by serpentinzed dunite or peridotite that forms the base of the Kromberg Formation. Asbestos exploration pits along the river are the first clear indication of the presence of ultramafic rocks at this point in the sequence. This thick massive ultramafic unit represents an olivine cumulate at the base of a thick komatiitic flow. 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. Silicified shale immediately below the ultramafic rocks along the Komati River is locally bleached, which may indicate a contact-metamorphic overprint.

SITE 13

(26°01’32.34”S, 30°5919.62”’E):

Dacitic turbidites: a sequence of turbiditic sandstones of dacitic composition form distinctive ribbed outcroups. Individual beds display normal grading and sedimentary structures produced by deposition from turbidity currents. These include a sequence of current structures reflecting deposition by flows of declining velocity and load and flame structures. The fine-grained tops of these beds are highly silicified, cherty rock, also composed of dacitic debris, whereas the lower, coarse-grained portions contain significant amounts of diagenetic carbonate.



SITE 14

(26°01’27.24”S, 30°59’18.0”E):

The chert horizon is overlain along a sharp and planar contact by a ca. 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. 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. Zircons from a dacite clast have been dated at 3445±3 Ma (Kröner et al., 1991; Kröner et al., 2013). This age is identical to ages obtained from intrusive/extrusive dacitic rocks of H6 (Kröner and Todt, 1988; Armstrong et al., 1990; Kröner et al., 2013). Palaeomagnetic analyses of the clasts indicate the presence of a high temperature magneti-zation components that passes a conglomerate test, indicating the existence of a geodynamo 3445 million years ago (Usui et al., 2009).

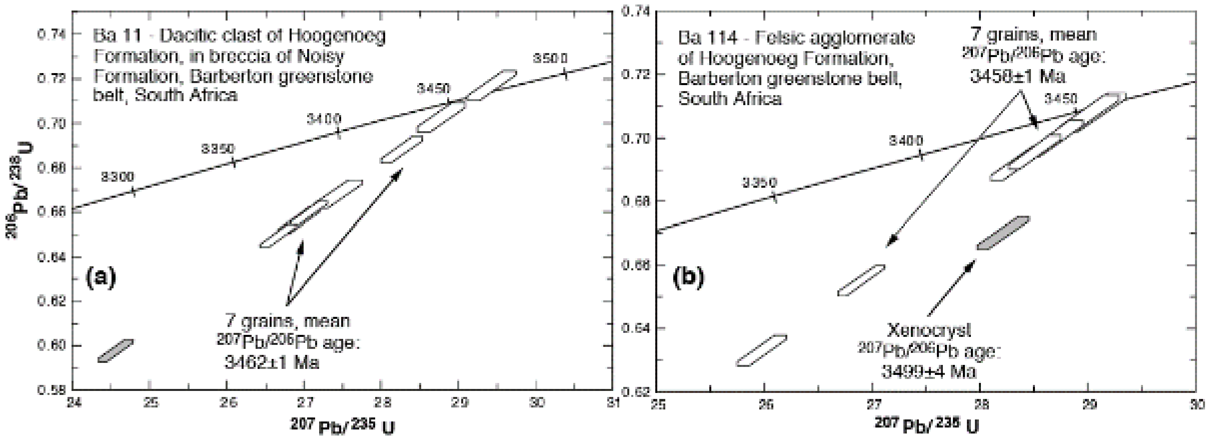


Fig. 1 Concordia diagrams showing SHRIMP II data for zircon from felsic volcanic rocks of the Hoogenoeg Formation, Komati River section. (a) Dacite clast from H6 boulder conglomerate. (b) Felsic agglomerate. Note that this sample contains an older xenocryst. From Kröner et al. (2013).

The conglomerate is overlain by very thick beds of normally graded and massive conglomerate and coarse-grained sandstone, followed by massive and parallel-laminated sandstone with minor intercalations of pebble conglomerate. The upper part of the sequence consists of thick-bedded turbidites of normally graded sandstone overlain by silicified shale (now chert), which grades upsection into silicified shale.

<u>SITE 15</u>

(26°01’22.8”S, 30°59’15.5”E)

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. Pillow basalt containing abundant ocelli and, commencing from ca. 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, an increase in SiO_2, K_2O, Eu/Eu*, $\delta^{18}O$ and $\delta^{30}Si$ values and a depletion of most mobile elements, including Ni, Co, Cu, and Zn (Hofmann and Harris, 2008; Abraham et al., 2011). Silicified basalt is transected by massive black chert veins in the uppermost few metres. Chert-veined basalt is capped along a sharp contact by a ca. 1 m thick horizon of massive to thinly laminated black chert that is, in turn, overlain by laminated grey chert. 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) as well as in chert veins underneath (Glikson et al., 2008). Hofmann and Harris (2008) attributed the origin of the silica alteration zones to low-temperature (100-150 °C) hydrothermal processes on the Archaean seafloor.

2.3.5 Sheba Mine Area, Brays GQ and return via Eureka City

STOP 1

Dycedale Syncline

Upward-fining, fluvial channel deposits that record evidence for tidal modification are preserved in the Clutha Formation of the Moodies Group in 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. 1). 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 instances, 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. 2). 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 the 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.

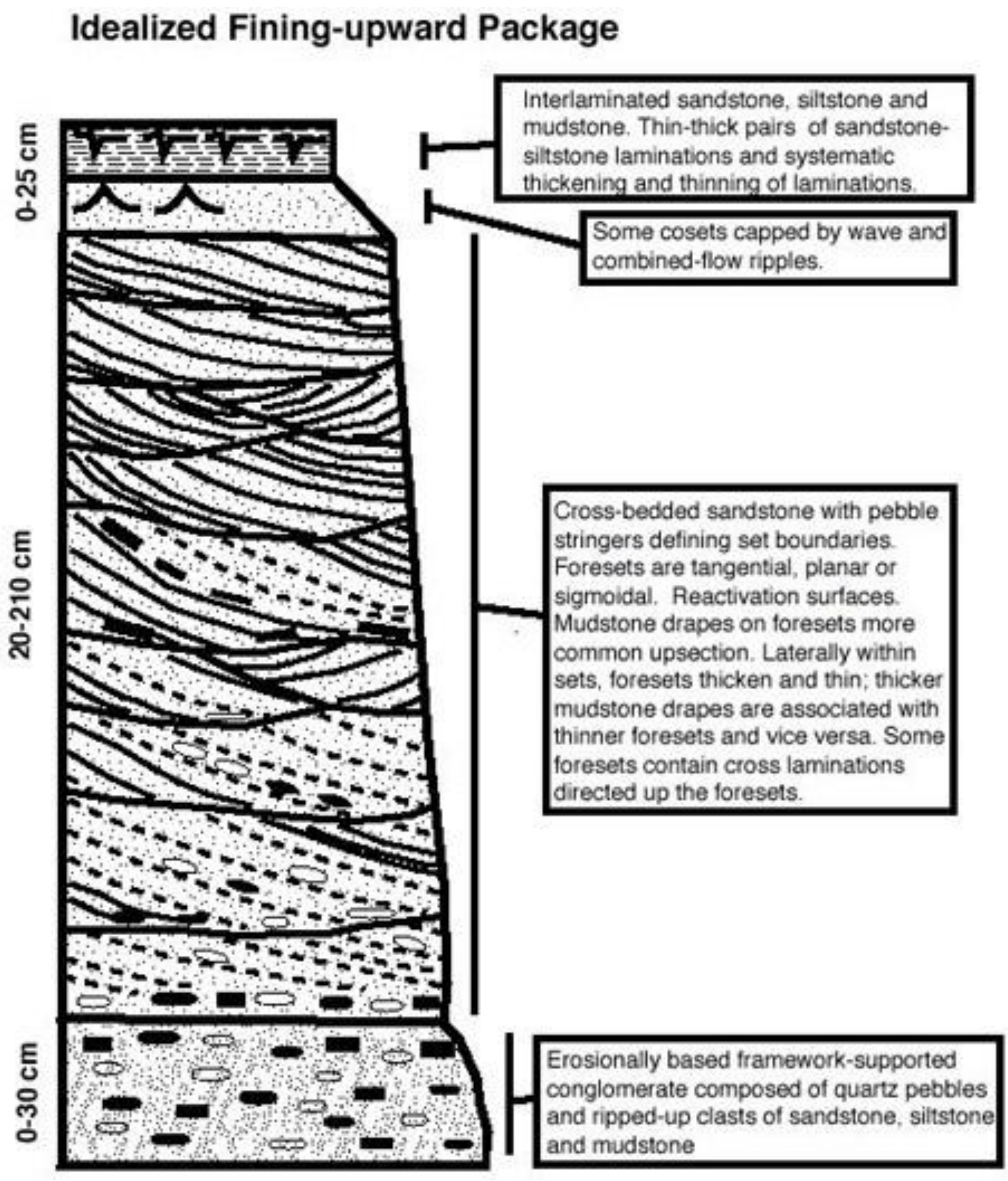


Fig.1 Idealized vertical sequence of lithologies and sedimentary structures in upward-fining, tidally modified fluvial deposits from the Moodies Group in the Dycedale Syncline.

Fig 2 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).

Locality: S25°47.415', E31°05.131'

STOP 2

Sheba Creek, Eureka Syncline

Quantitative evidence for tides has recently been recognized for the first time in the Clutha Formation of the ca. 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. 3). 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. 3A). 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., 1999). Cyclic variations in foreset bundle thicknesses 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. 3a) 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. 4b), 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 4b 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 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.

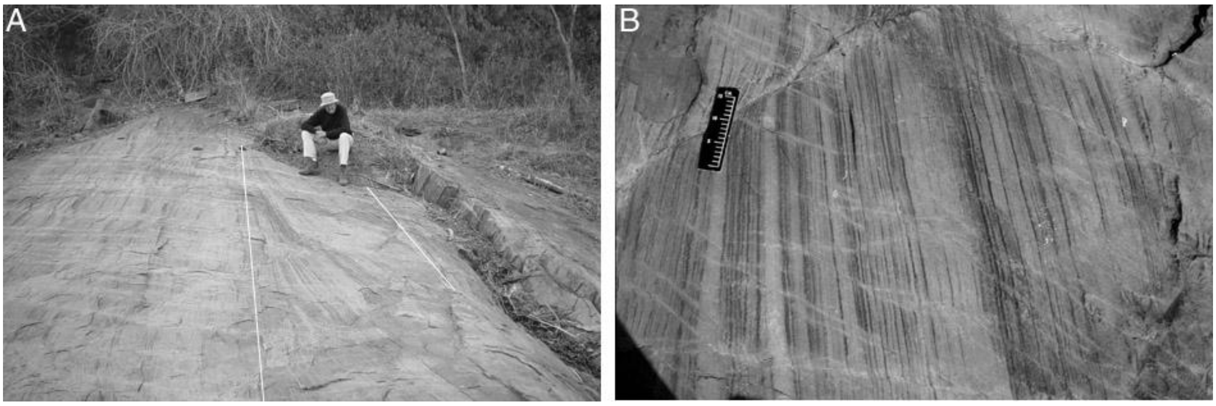
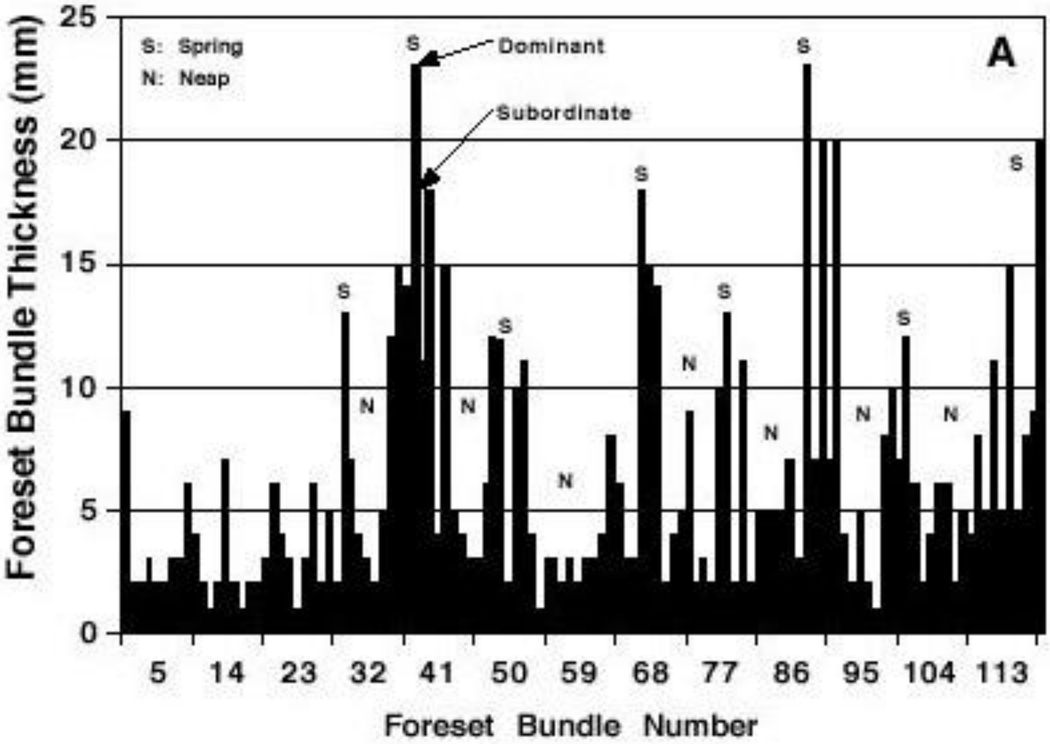


Fig 3 (A) Tidal sand-wave deposit in the Moodies Group, Eureka Syncline. Bed is delineated by white lines; stratigraphic way-up is to the right. (B) Close-up view of (A) showing bundles of foresets separated by mudstone drapes (scale bar is 15 cm long). Note the thickening and thinning of foreset bundles.



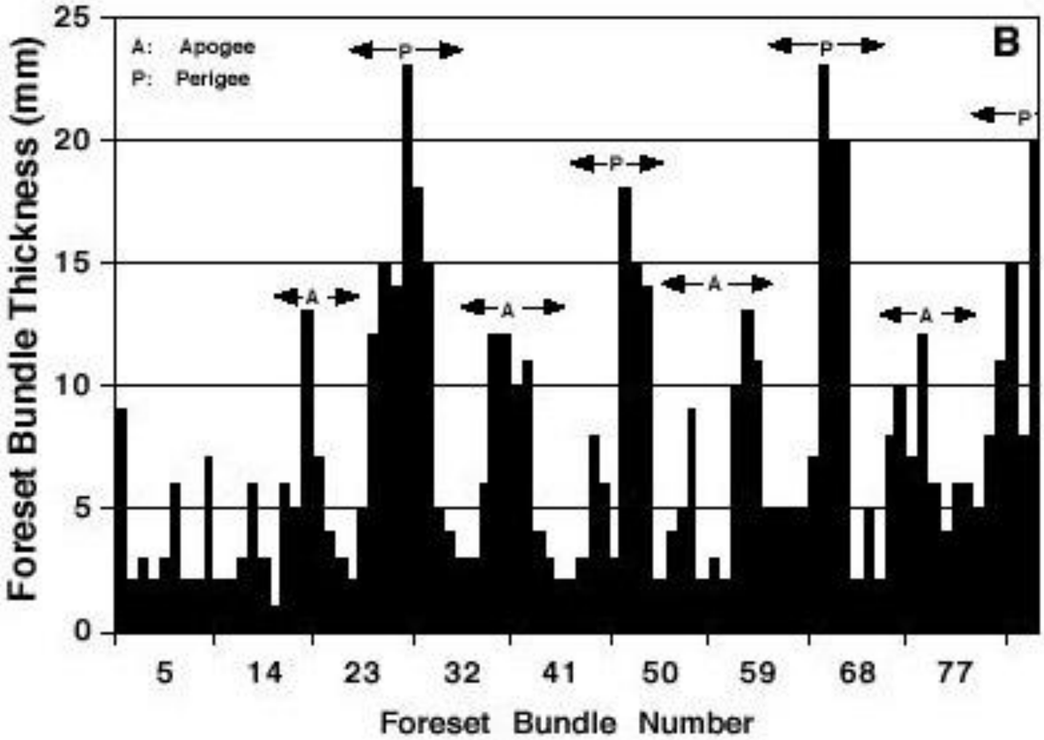


Fig 4 (a) Bar chart of sandstone bundle thickness versus bundle number from a sandwave deposit in the Moodies Group, Eureka Syncline. Note the presence of thick-thin pairs of bundles considered to reflect the dominant and subordinate tides of a diurnal system, and the cyclic thinning and thickening of 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.

Locality: S25°42.547', E31°09.671'

STOP 3

Basal conglomerate of the Moodies Group, Eureka Syncline

Objective. 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, 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 subgreywacke-shale units. The excursion will examine the basal conglomerates at two localities on the far eastern side of the Eureka Syncline. In the north the conglomerates are located in close proximity to the Nelspruit granite contact 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, 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.

Locality 1. 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 portion of the deformed agglomerates of the Schoongezicht Formation at the top of the Fig 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 to the 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.