Age, lithology and structural evolution of the c. 3.53 Ga Theespruit Formation in the Tjakastad area, southwestern Barberton Greenstone Belt, South Africa, with implications for Archaean tectonics

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A B S T R A C T
A field and petrographic re-assessment of rocks from the Theespruit Formation near Tjakastad, southwestern Barberton Greenstone Belt, is combined with new zircon U–Pb ages and whole-rock Sm–Nd isotopic results to reveal several important inconsistencies with a previous thrust-accretion model for this area. It was found that faults previously interpreted as ‘thrusts’ are extensional, ‘thrust slices of basement orthogneiss’ are little deformed quartz-feldspar porphyries or sheared felsic volcaniclastic sedimentary rocks, and ‘sedimentary diamictites’ are felsic agglomerates. These observations, combined with consistent facing directions of bedding, the recognition of distinct stratigraphic variations across strike, and U–Pb zircon dates from 13 felsic metavolcanic samples, all suggest that the Theespruit Formation is a moderately strained, coherent stratigraphic succession deposited at c. 3530 Ma and affected by extensional shear deformation at c. 3230 Ma. A 3453±6 Ma population of zircons analysed previously from an amphibolite-facies Theespruit Formation felsic schist is interpreted as being of metamorphic, rather than detrital, origin, and arising from heat associated with felsic plutonic rocks that stitch the Theespruit Formation to the overlying Komati Formation at ca. 3450 Ma, 220 Ma earlier than the proposed thrusting event.

1. Introduction
A major debate in the Earth sciences concerns the style of early Archaean tectonics, whether dominated by vertical movements due to vigorous mantle convection, plume-derived magmatism, and crustal diapirism, or horizontal plate interactions akin to the modern tectonic paradigm (e.g. de Wit, 1998 versus Hamilton, 1998). Although both tectonic processes have separately been identified as affecting different Archaean cratons, or different parts of such cratons (e.g. Myers and Kröner, 1994; Van Kranendonk et al., 2007), it is the relative contribution of these two processes that is still widely debated. This is particularly true regarding the formation of Palaeoarchaean granitoid-greenstone terrains (GGTs) that have a characteristic dome-and-basin map pattern, the likes of which do not occur in modern geological environments (Anhaeusser et al., 1969).

This debate is particularly vigorous regarding the formation of Archaean crust in the Barberton Mountain Land part of the Kaapvaal Craton, southern Africa (Fig. 1). On the one hand, Viljoen and Viljoen (1969) first interpreted the evolution of this area in terms of a little-deformed volcano-sedimentary (greenstone) succession, the Barberton Greenstone Belt, intruded by granitoid magmas and then tilted and folded by the vertical rise of tonalitic diapirs (Anhaeusser, 1984). However, the diapiric model was challenged by de Wit and colleagues who identified nappes, reverse faults, high strain zones, and local repetitions of stratigraphy in the greenstones (de Wit, 1982, 1983, 1991; de Wit et al., 1983, 1987a). These authors deduced that the Barberton Mountain Land part of the Kaapvaal Craton formed through horizontal thrust-accretionary plate tectonic processes in which granitoid diapirism played no part (de Wit et al., 1992).

Whereas nappes and reverse faults may form in a variety of environments, including diapirism (e.g. Ramberg, 1967; Dixon and Summers, 1983), the single critical piece of geological evidence supporting the thrust-accretion hypothesis is the interpreted large-scale occurrence of older-over-younger rocks in the Theespruit area, located in the southwestern part of the Barberton Greenstone Belt (Fig. 1) (de Wit et al., 1983, 1992; Armstrong et al., 1990). In this area, ca. 3.49–3.48 Ga (Lopez-Martinez et al., 1992; Dann, 2000) komatiites, basalts and a layer of felsic tuff of the Komati Formation occur structurally above interlayered pillow basalt, sedimentary ‘diamictite’, and ‘orthogneiss’ of the Theespruit Formation, which has been...
interpreted to be ≤ 3458 Ma — and therefore younger than the Komati Formation — based on a population of zircons from a felsic ‘diamictite’ within the Theespruit Formation (Armstrong et al., 1990). The kilometre-wide Komati schist zone (KSZ) that separates the two formations in the southwestern part of the Barberton Greenstone Belt was interpreted as a plate-boundary thrust (de Wit et al., 1992) (Fig. 1). In this model, the Komati Formation was interpreted to represent structurally thickened oceanic crust (“Jamestown Ophiolite Complex”; de Wit et al., 1987a), subducted onto a “tectono-stratigraphic assemblage” of Theespruit Formation rocks across the KSZ at ca. 3230 Ma (de Wit et al., 1983, 1987a, 1992; Kamo and Davis, 1994). Tectonic thickening was interpreted to have given rise to partial melting of the lower parts of the overthickened crust and the generation of granitoid rocks which were emplaced into active thrusts (de Wit et al., 1987b). Evidence of local high-pressure metamorphism from the granitoid domains flanking the belt has also been used to support models of continental collision (e.g., Dziggel et al., 2001, 2002; Diener et al., 2006; Stevens and Moyen, 2007).

However, several papers have cast doubt on the thrust-accretion hypothesis for this part of the Barberton Greenstone Belt. First, detailed volcanological studies have shown that a section with proposed sheeted dykes within the Komati Formation, which de Wit et al. (1987a) suggested was part of an ophiolite, is rather a series of thick flows with primary spinifex textures and sills or massive flows (Cloete and King, 1991; Cloete, 1999; Dann, 2000). In addition, more detailed volcanological studies indicate that the Komati and Hooggenoeg Formations, the latter of which contains felsic volcanic rocks, represent largely intact, autochthonous volcanic stratigraphy, more than 6 km thick (e.g., Cloete, 1999; Dann, 2000; Dann and Grove, 2007), and is thus very different from modern ophiolite sections. A gabbro interpreted as part of the ophiolite stratigraphy was shown to be some 100 Ma younger than the Komati volcanics (Kamo and Davis, 1994). These data preclude an ophiolite interpretation for the lower Onverwacht Group.

Second, Byerly et al. (1996) showed that the greenstones above the KSZ are a contiguous, progressively upward-younging succession (cf. Viljoen and Viljoen, 1969) which formed over more than 200 Ma of Earth history. This implies that much of the succession is not a tectono-stratigraphic assemblage as implied in the thrust-accretion hypothesis (de Wit et al., 1987a) — at best, an accretion component is reduced to a single thrust (the KSZ).

Third, Kisters and Anhaeusser (1995) showed that granitoid rocks flanking the greenstones show several features that are consistent with emplacement of granitoids as diapirs, thus indicating a more complicated story than proposed by the thrust-accretion hypothesis alone.

The above, when combined with more detailed observations presented below, are incompatible with the proposed thrust-accretion hypothesis, as it has previously been described. The cumulative weight of these arguments, together with our new data, requires a re-assessment of the critical Tjakastad area, the results of which are presented here.

The new results indicate that the published thrust-accretion model is unlikely to have been the primary tectonic mechanism responsible for development of the Palaeoarchaean rocks in the Tjakastad area. Instead, the structural features of this area are more consistent with formation through partial convective overturn of a dense succession of continually upward-younging, dominantly volcanic, supracrustal rocks and a structurally lower, more buoyant layer of intraplated granitoid sheets, following previously published diapiric models. The conclusions have important implications for early Archaean tectonic models, as the Barberton area has been widely cited as a key example in support of modern plate-tectonic processes operating in the ancient past (cf. Windley, 1995).

Fig. 1. Simplified geological map of southwestern part of the Barberton Mountain Land, Kaapvaal Craton, showing the location of Figs. 3 and 4 (rectangle), and of Fig. 7a, b, and c (dots). Inset shows location of the Barberton Greenstone Belt in the Kaapvaal Craton, within southern Africa. Age data for granitic rocks cited in text.
2. Regional geology and previous geochronology

The Barberton Mountain Land part of the Kaapvaal Craton is a Palaeoarchaean (3.55–3.20 Ga) granite–greenstone terrain that forms part of the eastern Kaapvaal Craton (Brandl et al., 2006). The volcano-sedimentary strata, known as the Barberton Supergroup, are preserved in the tightly folded, ENE trending Barberton Greenstone Belt (BGB), which is surrounded by granitoid rocks of a variety of ages (Lowe and Byerly, 1999). The most recent overview of the BGB is by Lowe and Byerly (2007), who also provide an extensive literature survey.

2.1. Barberton Supergroup

The Barberton Supergroup comprises a basal Onverwacht Group of komatiitic, basaltic and felsic volcanic rocks, and conformably to unconformably overlying argillaceous and arenaceous metasedimentary and minor felsic volcanic rocks of the Fig Tree and Moodies Groups, respectively (Fig. 2) (Viljoen and Viljoen, 1969; Brandl et al., 2006). The Onverwacht Group has been broadly subdivided into a lower ultramafic unit and an upper mafic to felsic unit, separated by a prominent unit of chert known as the Middle Marker (Viljoen and Viljoen, 1969). The lower and upper units are each composed of three formations. The structurally lowermost is the undated Sandspruit Formation composed of strongly deformed and metamorphosed mafic–ultramafic schists adjacent to, and infolded with, TTG rocks in the southwestern part of the BGB. Apparently conformably overlying the Sandspruit Formation is the heterolithic Theespruit Formation which regionally contains komatiites and basalts and characteristic units of metamorphosed felsic lava and tuff and which, in the Tjakastad area, has been interpreted to also include slivers of orthogneiss and diamictite (de Wit et al., 1983). The Theespruit rocks occur in two areas separated by younger granitoid intrusives,

![Generalized stratigraphic section of Barberton Supergroup](image-url)

Fig. 2. Generalized stratigraphic section of Barberton Supergroup, showing available age constraints (see text for references). SF = Sandspruit Formation. Modified from Lowe and Byerly (2007) and Byerly et al. (1996); age data cited in text.
namely around and east of the village of Tjakastad and partly in the Songimvelo Game Reserve (in this paper referred to as the Tjakastad area), and farther east in the area around Steynsdorp, close to the border between South Africa and Swaziland; here referred to as the Steynsdorp area (Fig. 1).

Single zircon $^{207}\text{Pb}/^{206}\text{Pb}$ evaporation ages of 3548 ± 3 to 3544 ± 3 Ma for four separate samples of felsic tuff from the Theespruit Formation in the Steynsdorp area and flanking the ca. 3510 Ma Steynsdorp Pluton provide a well-constrained date of deposition for this unit (Kröner et al., 1996). Zircon xenocrysts in the Steynsdorp Pluton vary in age from 3535 to 3531 Ma, providing further evidence on the age of the underlying Theespruit and Sandspruit Formations (op cit.). In the Tjakastad area, along strike to the west, a sample from what was interpreted as a thin sliver of basement orthogneiss (but see below) in the Theespruit Formation yielded similar results of 3538 ± 6 Ma and 3538 ± 4/−2 Ma, respectively, in two separate zircon dating efforts (Armstrong et al., 1990; Kamo and Davis, 1994). A felsic volcaniclastic rock from immediately above the so-called basement gneiss sliver, but still within the Theespruit Formation, yielded zircons with identical ages of 3533 ± 10 Ma (Armstrong et al., 1990). However, this sample also contained a second population of zircons with an age of 3453 ± 6 Ma, and it is this result that has been widely cited as a maximum age for the deposition of the Theespruit Formation (see below: Armstrong et al., 1990). An arkosic rock from the Sandspruit Formation to the west of our study area contained concordant zircons at 3521 ± 6 Ma, with older grains up to ca. 3540 Ma (Dziggel et al., 2001, 2002).

The Sandspruit and Theespruit Formations are separated from the overlying, lower-grade, Komati Formation by a broad zone of ductile shear, 600–800 m wide, and known as the Komati schist zone (KSZ). The overlying Komati Formation is composed of komatiitic and interlayered basaltic to fine-grained tuffaceous rocks dated at ca. 3490 Ma (Lopez-Martinez et al., 1992) and 3481 Ma (Dann, 2000). The Middle Marker at the top of the lower ultramafic unit is a persistent horizon of siliciclastic feldspar porphyry dykes and sills intruding the Komati Formation (Kamo and Davis, 1994).

Conformably overlying the Middle Marker is the Hooggenoeg Formation of tholeiitic basalt, komatiitic basalt, komatiite, and felsic volcanic flows and sills, dated at 3470 Ma (Byerly et al., 2002). Intruding the top of the Hooggenoeg Formation is a 3457–3438 Ma subvolcanic sill and related felsic lavas and volcaniclastic rocks (de Wit et al., 1987b; Kröner and Tordt, 1988; Armstrong et al., 1990; Kröner et al., 1991; Byerly et al., 1996; de Vries et al., 2006). These rocks are locally eroded and conformably overlain by metamorphosed feldspar flow and volcaniclastic rocks of the Komberg and Mendon Formations, dated at ≤3416 and >3298 Ma (Byerly et al., 1996; Lowe and Byerly, 2007).

The Fig Tree Group conformably overlies the Onverwacht Group and is largely composed of clastic and pelitic metasedimentary rocks, but includes felsic volcanic and volcaniclastic rocks dated at 3258 and 3226 Ma, the same age as the Kaap Valley Tonalite and related intrusions to the north of the BGB (Kröner et al., 1991; Kamo and Davis, 1994; Byerly et al., 1996; Lowe and Byerly, 2007). The Fig Tree Group is conformably unconformably overlain by conglomeratic and arkosic units of the Moodies Group which, near the top of the section, contain several intraformational unconformities indicative of syntectonic deposition in a contractional regime (Lamb, 1987). Deposition of the Moodies Group commenced around 3230 Ma and ended some time before 3110 Ma (Huebeck and Lowe, 1994). Synchronous folding and tilting of the entire Barberton Supergroup to near vertical occurred before 3216 ± 2/−1 Ma, the age of the discordant and undeformed Dalmein Pluton (Fig. 1; Kamo and Davis, 1994).

2.2. Granitoid rocks

Tonalitic, trondhjemitic and granodioritic (TTG) rocks intrude and envelope the BGB and may be divided into five age groups. The oldest of these is the trondhjemitic Steynsdorp Pluton, dated at between 3510 and 3502 Ma (Kröner et al., 1996). The second, more voluminous group of plutons is represented by linked, foliated to gneissic TTG rocks of the Stolzburg, Theespruit and Doornhoek plutons along the southwestern flank of the BGB. These plutons are not isolated intrusions but represent three domal fold-thrust-accretion belts.

Anhaeusser (1984) and Kisters and Anhaeusser (1995) showed that the domal plutonic lobes were not formed by cross-folding but display characteristic features of diapiric emplacement. These plutonic rocks have been dated between 3469 and 3437 Ma, the same age as quartz-feldspar porphyry dykes intruding the Komati Formation and felsic volcanism in the Hooggenoeg Formation (Fig. 2; Kröner and Tordt, 1988; Armstrong et al., 1990; Kamo and Davis, 1994; Kröner et al., 1996). Titanite from a 3476 ± 10 Ma quartz-feldspar porphyry dyke cutting the Komati Formation in the Tjakastad area was dated at 3458 ± 2 Ma and interpreted to represent the minimum age of intrusion of the dyke (Kamo and Davis, 1994). A second zircon generation from the ca. 3445 Ma Stolzburg Pluton, considered by Kamo and Davis (1994) to be of metamorphic origin, is ca. 3237 Ma. Metamorphic titanite from the ca. 3445 Ma Doornhoek and Stolzburg plutons was dated at ca. 3215 Ma and ca. 3201 Ma, respectively (Kamo and Davis, 1994).

The third group of plutons is represented by the foliated Kaap Valley Tonalite and related granitoids along the northern flank of the BGB, dated at 3229 ± 5 Ma (Tegtmeyer and Kröner, 1987), the same age as felsic volcanism in the Fig Tree Group (Kröner et al., 1991). The undeformed Dalmein Pluton was emplaced at 3216 ± 2/−1 Ma (Kamo and Davis, 1994). Finally, a number of potassic granitoids, all ~3105 Ma in age, surround the BGB, among them the 3105 ± 3 Ma Mpuluzi Batholith in the south (Kamo and Davis, 1994). These coarse-grained, porphyritic granites are largely undeformed.

2.3. Metamorphism

Viljoen and Viljoen (1969) identified an amphibolite-facies metamorphic aureole in greenstones around granitoids that surround the BGB. de Wit (1983) noted that greenstones of the Theespruit Formation adjacent to the Theespruit Pluton in the Tjakastad area were transformed into garnet amphibolites, and aluminous felsic schists contained lakedy kyanite porphyroblasts. Cloete (1993, 1999) found that pressure estimates from fluid inclusion geobarometry in the Komati and Hooggenoeg Formations increased with stratigraphic depth to a maximum of 3.9 kbar and that the obtained values correspond to the lithological thickness of the overlying Barberton Supergroup, thus indicating that these rocks had only been subject to burial metamorphism. Temperature estimates using the plagioclase-amphibole thermometer vary from 320–420 °C in the Hooggenoeg Formation, to 490–530 °C in the Komati Formation. Rocks adjacent to the Theespruit Pluton were affected by a dynamic metamorphic overprint in addition to the static metamorphism, as evidenced by retrograde temperatures of 280 °C. Prograde P–T paths for Komati and Hooggenoeg Formation rocks have a slightly anti-clockwise shape, from which Cloete (1993, 1999) concluded that metamorphism of the BGB was not analogous to that in Phanerozoic thrust-accretion belts.

Van Vuren and Cloete (1995) studied the metamorphism of rocks in and around the Theespruit Pluton. These authors found that amphibolite-facies metamorphic conditions within the pluton were 3–5 kbar and 470 °C. Marginal greenstones recrystallised under...
conditions of 4.8–7 kbar, 500 °C. The highest metamorphic grade was found in the narrow septum of strongly deformed greenstones between the Theespruit and Stolzburg Plutons (Fig. 1), where garnet-clinopyroxene mafic granulites yielded \( P-T \) estimates of 9 kbar and 700 °C.

Dziggel et al. (2002) recorded high-\( P \), low-\( T \) metamorphic conditions of 8–11 kbar and 650–700 °C in remnants of the Theespruit and Sandspruit Formations within the western part of the Stolzburg Pluton. Two phases of titanite were dated from rocks of this area, at 3418 ± 8 Ma and 3236 ± 5 Ma, indicating two periods of metamorphism.
(and deformation) (Dziggel et al., 2001, 2002). Similar observations were made by Diener et al. (2006) for the area just west of Tjakastad, with prograde conditions of 5.5–6.3 kbar and 490 °C. Higher grade conditions in rocks closer to the granitoids gave peak P–T estimates of 7–7.7 kbar and 550 °C, with retrograde assemblages of 3.8 kbar and 543 °C, indicating nearly isothermal decompression. The estimates of peak metamorphic conditions indicate low geothermal gradients of 20 °C/km for the western part of the Barberton Mountain Land (Dziggel et al., 2002; Diener et al., 2006). A syn-tectonic tonalite dated at 3231 ± 5 Ma from the Schapenburg schist belt in the southwestern part of the Barberton Mountain Land gives a maximum age of amphibolite-facies metamorphism in this area, as does a 3219 Ma date on metamorphic zircon from an amphibolite in the western Tjakastad schist belt, only 10–20 Ma after sediment deposition (Stevens et al., 2002; Kisters et al., 2003; Diener et al., 2006).

3. Details of the proposed thrust-accretion model for the Tjakastad area

As evident from the above overview, there is only one exception to a generally conformable interpretation for the Barberton Supergroup. This occurs in the Tjakastad area, where purported evidence of large-scale tectonic inversion of older-over-younger rocks has been documented. de Wit et al. (1983) suggested that the Theespruit Formation in the Tjakastad area was a tectono-stratigraphic assemblage of tectonically intercalated mafic volcanic rocks, metasedimentary diamictite and associated quartz-sericite schist, quartz-feldspar porphyries, and slivers of banded granitoid basement gneiss. In this model, the Theespruit Formation was “...most readily explained as a set of imbricate slices... that represents an inclined (oblique?) section through an imbrication zone beneath a major thrust... represented by the Komati schist zone...” (de Wit et al., 1983, p. 27; word in italics added).

The most critical field evidence in support of a tectono-stratigraphic interpretation for the Theespruit Formation in the Tjakastad area was the recognition of tectonic slivers and cobbles of inferred basement orthogneiss (de Wit et al., 1983). Orthogneiss slivers were recognised in three distinct settings: (1) as a 500 m long wedge within quartz-sericite schist, diamictite, and quartz-plagioclase porphyries within the KSZ (locality E on Fig. 3); (2) as tectonic slivers within the Theespruit Formation, in one location being unconformably overlain by sedimentary diamictite (locality D on Fig. 3); (3) as cobbles in metasedimentary diamictite (locality A on Fig. 3). In setting 1, de Wit et al. (1983) suggested that the orthogneiss had experienced an earlier deformation history as indicated by isocinal folds of metamorphic layering, the axial planes of which were oriented at a high angle to the regional foliation (Fig. 6b and c of de Wit et al., 1983). Similarly, cobbles of orthogneiss within diamictite in setting 3 were interpreted to exhibit a pre-existing gneissosity oriented at a high angle to both bedding and the regional upright cleavage in the matrix diamictite.

That thrusting had occurred in the Tjakastad area was interpreted from the penetratively deformed nature of the rocks, the presence of faults that offset the stratigraphy and were interpreted to indicate a reverse sense of displacement, the presence of tectonic contacts as indicated by changes in the orientation and intensity of mineral elongation lineations across unit contacts, the occurrence of a south-facing bedding top direction within the KSZ, and along-strike differences in the thickness and components of the “tectono-stratigraphy”.

Supporting a tectono-stratigraphic interpretation for the Theespruit Formation were zircon ages of 3538 ± 4/–2 Ma and 3538 ± 6 Ma (Armstrong et al., 1990; Kamo and Davis, 1994) for the “basement gneiss sliver” at locality D. At the time these results were obtained, they were the oldest recorded from the BGB and apparently confirmed a basin interpretation for these rocks relative to the overlying Komati Formation and younger conformable stratigraphy. Furthermore, euhedral to subhedral zircons with well-developed facets from the unconformably overlying ‘diamictite’ (which Armstrong et al., 1990 interpreted as a volcaniclastic rock), about 2 m above the gneiss wedge, yielded two distinct age populations: an older group at 3531 ± 10 Ma, indistinguishable from the “gneiss sliver”, and a younger group at 3453 ± 6 Ma. Although “…core and rim structures are common...” (Armstrong et al., 1990, p. 94) in zircons from these greenschist- to amphibolite-facies schists, the younger ages were interpreted to be derived from detrital grains and thus to indicate a maximum age of deposition for the Theespruit Formation. These data apparently confirmed both a younger age of the ‘diamictite’ relative to the gneiss sliver, and that the KSZ represents a major thrust across which older rocks of the Komati Formation (3490 Ma) were tectonically emplaced over the younger rocks of the Theespruit Formation (<3453 Ma).

3.1. Controversial or unexplained elements of the previously proposed model

Several features of both the geological interpretation by de Wit et al. (1983; Fig. 3) and the geochronology (Armstrong et al., 1990) do not appear easily reconcilable with a thrust-accretion origin for the rocks in this region:

1) Why do the identified faults within the Theespruit Formation (f–f on Fig. 3) extend the lithological layering, rather than duplicate it as required for a thrust-accretion model?

2) What is the sense of shear across the KSZ? If, as proposed, the KSZ represents a crustal-scale thrust, then it should contain evidence of reverse (Komati-over-Theespruit Formation) displacement.

3) What is the origin of the ‘diamictites’ and how do they differ from volcaniclastic rocks or igneous intrusive breccias described elsewhere in the map area, or to felsic volcanic schist and meta-agglomerate along strike to the east and west as described by Kisters and Anhaeusser (1995) and Kröner et al. (1996)?

4) Why does a large unit of quartz-plagioclase porphyry with intrusion breccia textures pass laterally along strike into ‘diamictite’ (just east of locality E in Fig. 3)?

5) Why is there a discrepancy in the interpreted origin of felsic schists of the Theespruit Formation between the Tjakastad area (sedimentary diamictite and orthogneiss: de Wit et al., 1983) and those areas along strike to the west?

6) What is the relationship between the proposed orthogneiss wedge (locality E in Fig. 3) and adjacent ‘diamictite’ and quartz-plagioclase porphyries that surround, or structurally overlie it; does the wedge represent the limb of an inverted nappe, or is it located within a tectonic mélange?

7) What role, if any, did the ca. 3450 Ma granitoid rocks play in the deformation of the Tjakastad area? Is it the same as for ca. 3230 Ma granitoids higher up in the stratigraphy which were emplaced during regional thrusting (cf. de Wit et al., 1987b), or do

Fig. 5. Outcrop and thin section features of Theespruit Formation rocks: a) Well-preserved pillow basalts near the base of the formation. View is to the west-northwest, and drainage cavities (arrowed) and pillow tails (i.e. at hammer head) indicate way-up to north-northeast; b) Well-rounded cobble of fine-grained siliceous metasediment in foliated and linedated felsic agglomerate with stretched lapilli (not visible in photo); class shows coarse layering defined by white and light grey colour banding; c) Cross-sectional view of crude layering parallel to bedding defined by thin, irregular sills of darker grey, weakly plagioclase feldspar-porphyrty dacite in fine-grained felsic volcaniclastic siltstone; d) Cross-sectional view of massive, coarse felsic volcaniclastic rock (agglomerate) with irregular, ragged-edged flammé of feldspar-pyhricty, devitrified rhyolite glass (arrow) and well rounded clasts of chert; e) View, looking down on a horizontal surface, of alkaline trachytic breccia: note angular, unstrained nature of clasts in this view; f) View, looking horizontally, on alkaline trachytic breccia, showing strongly down dip linedated nature of fragments; g) Cross-polarised thin section view of alkaline trachyte fragment in breccia, showing Carlsbad-twinned subhedral K-feldspar phenocryst in fine matrix of flow-aligned albite. Pen knife in b)–f) is 10 cm. All locations are shown in Fig. 4.

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they display evidence of diapiric emplacement as previously proposed (Anhaeusser, 1984; Kisters and Anhaeusser, 1995)?

8) What is the significance of high-grade metamorphic minerals (kyanite in pelitic schists and garnet in amphibolite) within the Theespruit Formation? Is the variation in metamorphic grade across the Tjakastad area consistent with, and exclusive to, a thrust-accretion model of formation, or can it be explained in other ways?

9) What is the significance that the ‘basement gneiss sliver’ in the Tjakastad area is identical in age to felsic volcanic schists of the Theespruit Formation along strike to the southeast?

10) What is the significance of the younger population of zircons in the dated, amphibolite- to greenschist-facies ‘diamictite’ (ca. 3453 Ma) which display core–rim overgrowth relationships and are identical in age to titanite in a quartz-feldspar porphyry sill cutting the Komati Formation (ca. 3458 Ma) and to widespread TTG plutonism (3470–3437 Ma) and felsic volcanism (3452–3438 Ma)?

4. Results of field mapping and petrography

Field mapping was undertaken in order to investigate the apparent inconsistencies in the thrust-accretion model for the Tjakastad area. The results are presented in Fig. 4, and below, under subheadings which individually address each of the apparent inconsistencies. The map differs in several key ways from that published by de Wit et al. (1983), particularly by way of different interpretations for the origin of several rock units in the Theespruit Formation, a stratigraphic interpretation for the Theespruit Formation, the nature of contact relationships between units, the position of the northeastern part of the Komati Fault, and the kinematics of faulting/shearing across the Tjakastad area.

4.1. Lithology

4.1.1. General stratigraphy

The map area spans the transition from amphibolite-facies schists of the Theespruit Formation, which lie adjacent to the trondhjemitic Theespruit Pluton, to lower greenschist-facies komatiites of the Komati Formation across the high-strain KSZ (Fig. 4). The best and most continuous exposures are in a small streambed in the hills east of Tjakastad. The Theespruit Formation consists of lower pillowed and ocellar metabasalts (Fig. 5a), three compositionally and texturally distinct horizons of felsic schists and metagabbros, and layer interlayered mafic–ultramafic schists of indeterminate extrusive or intrusive origin. Way-up indicators of bedding are defined by graded bedding in felsic volcaniclastic layers and pillow structures in basaltic. These consistently face to the northeast, away from the Theespruit Pluton (Fig. 4), except on the limbs of small-scale tight folds of bedding. Lithologic units of the Theespruit Formation continue into the KSZ, but then change across the Komati Fault into the base of the Komati Formation, which includes the type area of bladed olivine–spinel-textured komatiites, massive ultramafic rocks, talc-serpentinite schist, and intrusive quartz-feldspar porphyry.

4.1.2. Felsic volcanic rocks

Three compositionally and texturally distinct horizons of felsic rock occur in the Theespruit Formation (Fig. 4). The structurally and stratigraphically lowest felsic horizon is ≤5 m thick. It consists of irregularly layered felsic volcanic schist, grey-and-white layered, fine-grained, cherty metasediment (siltstone/mudstone), kyanite-bearing quartz-sericite schist derived from aluminous greywacke, subordinate felsic tuff, and some massive felsic volcanic rocks. This horizon contains the dated unconformity locality that is discussed in detail in Section 4.1.4.2. The kyanite–andalusite schists contain faint relict bedding and lineated kyanite porphyroblasts that locally retain fresh cores of andalusite (see Section 4.3). The cherry metasediment is characterised by irregular, wispy layering and common chaotic folds that were possibly developed during, or soon after sedimentation, as they contain no axial planar foliation and have highly non-cylindrical fold axes, in contrast to folds of clearly tectonic origin elsewhere. Locally, cherry metasediments occurs as mis-oriented blocks in massive, fine-grained quartz-sericite rock (Fig. 5b). Nearby, the layered metasediment has an irregular contact with massive, weakly feldspar-phric quartz-sericite rock that cuts bedding within the metasediment (Fig. 5c). These textures are interpreted as indicative of coeval sedimentation of felsic volcaniclastic detritus and felsic magmatism, the latter formed as either a felsic volcanic flow or a sill emplaced directly below the sediment–water interface.

The middle horizon of felsic schist, which de Wit et al. (1983) interpreted to contain extensive sedimentary diamictite with local clasts of basement granitoid orthogneiss, is similar in thickness to the lower horizon, but is texturally heterogeneous along strike. The best exposures of the coarse clastic rocks are in the small streambed mentioned above, where they comprise almost the entire thickness of the horizon. This unit contains subrounded to subangular, pebble to boulder size clasts of abundant grey to black chert, less abundant white and grey banded chert, clastic metasediment, and fine-grained felsic volcanic rock, and rare mafic volcanic rock and granite in a fine-grained quartz-sericite matrix (Fig. 5d). We interpret these rocks to be proximal felsic volcanic pyroclastic agglomerates and not sedimentary diamictite, because:

1. Well-preserved parts of the unit contain common, irregular, ragged-edged flans of felsisp-phryic, devitrified rhyolite glass (arrow in Fig. 5d), whose delicate shapes could not have survived transport by water together with the rounded boulders of chert, as they would have almost instantly been disaggregated.

2. The unit locally contains elliptical bombs of vesicular to pumiceous felsic volcanic rock in an otherwise massive, fine-grained felsic matrix that lacks bedding.

3. The coarsely fragmental rocks lack bedding, are unsorted, and laterally pinch and swell along strike, grading into felsic lavas and/or fine-grained volcaniclastic sediment.

4. In thin section, the matrix is seen to be generally fine-grained with no evidence of original detrital grains, and it contains elliptical to irregular-shaped patches of slightly coarser quartz–biotite–plagioclase with serrate, gradational contacts with the matrix, that are interpreted to represent flan.

The stratigraphically highest felsic horizon locally displays metre- to centimetre-scale bedding and is characterised by layers of quartz-phryic felsic porphyry with subordinate felsic schist, fine-grained quartz-biotite metasedimentary rock of possible volcaniclastic origin, and minor felsic volcanic breccia.
Fig. 7. a) Large, unstrained amphibolite-facies pillow basalt in a 100–200 m xenolith from core of Theespruit Pluton (see Fig. 1 for location): pillows face to the southeast; b) Vertical cross-sectional view of strongly lineated, amphibolite-facies pillows within narrow greenstone septum between lobes of the Theespruit Pluton; c) Sheeted, heterogeneous margin of Theespruit Pluton, which contrasts with the unstrained, homogeneous core of the pluton; d) Rodded I-tectonite felsic volcanic schists from the KSZ, used as building materials; e) Plane polarised light view of a vertically-oriented thin section, looking southeast, showing SW-side-up displacement in sheared metabasalt from the KSZ (NE = northeast, SW = southwest); f) Plane polarised thin section view of aluminous felsic schist of Theespruit Formation, showing kyanite (Ky = dark grey) overgrowing andalusite (And = light grey, with well-developed cleavage) along extension cracks, indicative of an increase in pressure during metamorphism and the development of linear fabrics in these rocks. Locations are shown in Figs. 1 and 4.
Two thin units of strongly sheared, texturally variable felsic schist in the western part of the KSZ are similar to the lower felsic unit in the Theespruit Formation. The topmost unit contains fine-grained tuff, grey chert, and fine-grained quartzite. In the southern-central part of the KSZ, a ∼100 m thick unit of felsic schist that de Wit et al. (1983) interpreted as sedimentary diamictite varies, from base to top, from a massive rock with plagioclase-phyric texture, a middle unit of felsic volcanic breccia with elliptical, ragged-shaped inclusions of carbonate-altered felsic material, and ≤50 m of grey and white layered chert.

In the eastern part of the KSZ in the study area, de Wit et al. (1983) described an intrusion breccia within the core of a quartz-plagioclase porphyry intrusion that they interpreted to pass along strike westwards into sedimentary diamictite (see Fig. 3). However, it was found that the rocks in this area comprise two distinct units (see Fig. 4): 1) an intrusive unit of quartz-feldspar porphyry to the north (described below); and 2) a fine-grained, highly siliceous felsic volcanic rock with closely packed, monomict, angular breccia fragments (Fig. 5e). Detailed outcrop mapping showed that parts of this unit display crude layering on a metre scale, with lenses or layers of massive rock alternating with layers of homogeneous, monomict, angular breccia, and other layers of less well sorted breccia with irregular-shaped blocks of massive rock up to a metre in size. Overall, bedding becomes more clearly defined at a finer-scale to the north, suggestive of graded bedding and a top to the north facing direction. Despite excellent preservation of igneous textures on horizontal outcrops, the rocks have been deformed into vertically-plunging L-tectonites (Fig. 5f). In thin section, the massive unit at this...
locality is characterised by euhedral K-feldspar phenocrysts with Carlsbad twinning in a matrix of fine, flow-aligned albite with trachytic texture (Fig. 5g). Quartz phenocrysts are lacking in this lithology, and thus the rock is classified as an alkali trachyte breccia.

The textural uniqueness of each of the felsic horizons in the Theespruit Formation and the KSZ indicate that they do not represent a tectonically-duplicated series of thrust slices as suggested by de Wit et al. (1983). Rather, the above data indicate distinct lithological variations across strike, and the presence of consistent bedding top indicators to the north in both pillowed mafic volcanics and felsic volcanic/volcaniclastic horizons suggests that the rocks of the Theespruit Formation in the Tjakastad area, through into the KSZ, represent a strained, but otherwise intact volcano-sedimentary succession. This hypothesis is discussed more fully in Section 4.2.

4.1.4.1. Northeastern wedge of granitoid gneiss. The wedge of or volcaniclastic metasedimentary rocks, for reasons detailed below. These rocks are re-interpreted as either quartz-plagioclase porphyries localities and that no such rocks occur in the Theespruit area. Instead, composition with characteristic euhedral, bipyramidal quartz and Fig. 3) consists of a homogeneous, medium-grained rock of trondhjemitic well preserved, consisting of subhedral to euhedral quartz and unstrained except for a very local, weak foliation developed within the KSZ (see Section 4.2 below); they are almost completely rocks contain none of the strong marked by a discrete, fault-

4.1.4.2. Granitoid gneiss at the ‘unconformity’ locality. The ‘unconformity’ locality described by de Wit et al. (1983: locality D in Fig. 3) is underlain by highly strained schists over a strike length of 10 m and 30 cm across. The interpreted unconformity in this outcrop consists of a lower unit (‘orthogneiss sliver’ of de Wit et al., 1983), 28 cm wide, of millimetre- to centimetre-scale interbedded quartz-sericite, quartz-sericite-tourmaline-chloritoid, quartz-clayite, sericite–quartz-chlorite-actinolite schist (Fig. 6f, g). The overlying schist unit (‘diamictite’ of de Wit et al., 1983) is 2 cm wide and displays a distinctive breccia of strongly lined felsic fragments (L = 62°→107°) on foliation surfaces (Fig. 6h). The fragments are composed of sericite (95%)–chlorite-titanite up to 120 × 40 × 2 mm in size (X/Y/Z axes) within a matrix of slightly darker quartz-sericite-biotite-chlorite schist. These rocks are overlain by quartz-sericite schist, finely layered grey and white banding rather than a metamorphic segregation, and that the rock is thus a quartz-plagioclase porphyry intrusion, similar to the type dated by Kamo and Davis (1994) at ca. 3470 Ma; it is not a basement orthogneiss.

Finally, the “evidence” cited by de Wit et al. (1983) that the isoclinally folded ‘granitoid gneiss’ at this locality had experienced an earlier phase of deformation on the basis that the folds were oriented at a high angle to the regional foliation, cannot be substantiated since the rock containing the folds is a misoriented boulder that has partly slid down the side of a hill.

4.1.3. Granitic dyke in the Theespruit Formation A 7–8 m wide sheared granite dyke occurs at one locality in the small stream bed to east of Tjakastad, about 100 m north of the Theespruit Formation/Theespruit Pluton contact. The foliation in this granite is parallel to that in the adjacent Theespruit schists, and the dyke could not be followed along strike due to poor exposure. The dyke is similar in appearance to granitic to aplitic dykes and veins emplaced into the Theespruit and Stolzburg plutons in the Tjakastad region. The dyke was sampled for zircon dating, and the results are reported below.

4.1.4. “Granitoid gneiss” The detailed map of the Theespruit area by de Wit et al. (1983) shows the precise locations where exposures of granitoid gneiss were thought to occur in three distinct tectonic settings. These included a large wedge of orthogneiss in the northeastern part of the KSZ, tectonic slivers in the lower part of the Theespruit Formation, and cobbles within sedimentary “diamictite”. Re-examination of these easily recognisable sites and detailed petrographic studies of collected samples indicate that granitoid rocks were misidentified at these localities and that no such rocks occur in the Theespruit area. Instead, these rocks are re-interpreted as either quartz-plagioclase porphyries or volcaniclastic metasedimentary rocks, for reasons detailed below.

4.1.4.1. Northeastern wedge of granitoid gneiss. The wedge of ‘granitoid gneiss’ in the northeastern part of the map area (locality E in Fig. 3) consists of a homogeneous, medium-grained rock of trondhjemitic composition with characteristic euhedral, bipyramidal quartz and plagioclase phenocrysts (<3 mm) scattered throughout a finer-grained quartz-feldspathic matrix (Fig. 6a). In thin section, igneous textures are well preserved, consisting of subhedral to euhedral quartz and plagioclase phenocrysts in a fine-grained quartz-feldspathic matrix that is overprinted by radiating sheaths of riebeckite needles (Fig. 6b).

In contrast to the map of de Wit et al. (1983), this unit occurs outside, to the north of the KSZ, the northern margin of which is clearly marked by a discrete, fault-filled, 2 m wide quartz vein (Fig. 4). The rocks contain none of the strong L-5 tectone fabric characteristic of the KSZ (see Section 4.2 below); they are almost completely unstrained except for a very local, weak foliation developed within 10 m of the KSZ. These rocks contain no leucosome veins, migmatitic textures, or gneissic layering with palaeosomes–leucosome relations to indicate that they were derived through metamorphic segregation processes related to an earlier thermo-tectonic event, as previously interpreted (de Wit et al., 1983). The compositional layering which does occur in this unit is only locally developed in the southern part of the unit, and in only one outcrop does it outline isoclinal folds (Fig. 6c; same as Fig. 6b and c of de Wit et al., 1983). This instantly recognisable outcrop with the isoclinally folded layering contains undeformed, euhedral bipyramidal quartz and plagioclase phenocrysts throughout, even across the hinge regions and limbs of the folds (Fig. 6d, e). The presence of these igneous phenocrysts indicate that the folded compositional layering represents igneous flow

Table 1

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
<th>Rock type</th>
<th>Field relationship</th>
</tr>
</thead>
<tbody>
<tr>
<td>BA36</td>
<td>$25°51.17'$ E30°50.15'</td>
<td>Felsic tuff</td>
<td>Bottom of 30 m wide layer</td>
</tr>
<tr>
<td>BA57</td>
<td>$25°50.24'$ E30°51.14'</td>
<td>Felsic tuff</td>
<td>Top of 50 m wide layer</td>
</tr>
<tr>
<td>BA62</td>
<td>$25°50.02'$ E30°50.02'</td>
<td>Felsic tuff</td>
<td>Close to Theespr. tonalite contact</td>
</tr>
<tr>
<td>BA65</td>
<td>$25°50.04'$ E30°49.59'</td>
<td>Granite</td>
<td>Sheared dyke cutting Theespruit Fm.</td>
</tr>
<tr>
<td>BA66</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Amphibolite</td>
<td>Interlayered with felsic tuff</td>
</tr>
<tr>
<td>BA69</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Amphibolite</td>
<td>Interlayered with felsic tuff</td>
</tr>
<tr>
<td>BA70</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Felsic tuff</td>
<td>30 m wide layer</td>
</tr>
<tr>
<td>BA71</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Felsic tuff</td>
<td>6 m wide layer</td>
</tr>
<tr>
<td>BA73</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Felsic tuff</td>
<td>10 m wide layer, very fresh</td>
</tr>
<tr>
<td>BA74</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Basaltic komat.</td>
<td>10 m wide layered pillow</td>
</tr>
<tr>
<td>BA75</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Gabbro</td>
<td>Sill in felsic tuff</td>
</tr>
<tr>
<td>BA79</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Felsic tuff</td>
<td>60 m wide layer</td>
</tr>
<tr>
<td>BA84</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Felsic tuff</td>
<td>Interlayered with chert</td>
</tr>
<tr>
<td>BA85</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Felsic tuff</td>
<td>Layer 10 mN of BA 85</td>
</tr>
<tr>
<td>BA104</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Fsp-porphyry</td>
<td>25 thick layer in felsic tuff</td>
</tr>
<tr>
<td>BA105</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Fsp-porphyry</td>
<td>ca. 25 m wide layer, fresh</td>
</tr>
<tr>
<td>BA107</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Felsic tuff</td>
<td>Closure of isoclinal fold</td>
</tr>
<tr>
<td>BA108</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Felsic tuff</td>
<td>ca. 50 m thick layer, fresh</td>
</tr>
<tr>
<td>BA109</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Felsic tuff</td>
<td>Same layer as BA 105, bio-rich</td>
</tr>
<tr>
<td>BA110</td>
<td>$25°50.00'$ E30°50.00'</td>
<td>Felsic tuff</td>
<td>Thick, coarse felsic volcaniclastic conglomerate</td>
</tr>
</tbody>
</table>

All samples listed in the table are plotted on Fig. 4.
Table 2
Isotopic data from evaporation of single zircons from rocks of the Theespruit Formation and one intrusive granite, Tjakastad–Songimvelo area, Barberton Greenstone Belt.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Zircon colour and morphology</th>
<th>Grain #</th>
<th>Mass scans</th>
<th>Evaporation temp. in °C</th>
<th>Mean $^{207}\text{Pb}/^{206}\text{Pb}$ ratio$^2$ and 2-$\sigma$ m error</th>
<th>Mean $^{207}\text{Pb}/^{206}\text{Pb}$ age and 2-$\sigma$ m error</th>
</tr>
</thead>
<tbody>
<tr>
<td>BA 56</td>
<td>Clear to yellowish,</td>
<td>1</td>
<td>109</td>
<td>1599</td>
<td>0.310606±0.64</td>
<td>3523.9±0.3</td>
</tr>
<tr>
<td></td>
<td>Long-prismatic,</td>
<td>2</td>
<td>150</td>
<td>1601</td>
<td>0.310751±0.67</td>
<td>3524.6±0.3</td>
</tr>
<tr>
<td></td>
<td>Neat-idiomorphic</td>
<td>3</td>
<td>83</td>
<td>1598</td>
<td>0.310486±0.13</td>
<td>3523.3±0.7</td>
</tr>
<tr>
<td></td>
<td>Near-idiomorphic</td>
<td>4</td>
<td>102</td>
<td>1599</td>
<td>0.310533±0.93</td>
<td>3523.6±0.5</td>
</tr>
<tr>
<td>Mean of grains 1–4</td>
<td></td>
<td>1–4</td>
<td>444</td>
<td></td>
<td>0.310593±0.44</td>
<td>3524.0±0.2</td>
</tr>
<tr>
<td>BA 57</td>
<td>Yellow-brown,</td>
<td>1</td>
<td>128</td>
<td>1603</td>
<td>0.310214±0.13</td>
<td>3521.9±0.5</td>
</tr>
<tr>
<td></td>
<td>Long-prismatic,</td>
<td>2</td>
<td>110</td>
<td>1602</td>
<td>0.310412±0.13</td>
<td>3523.0±0.8</td>
</tr>
<tr>
<td>Mean of grains 1–3</td>
<td></td>
<td>1–3</td>
<td>320</td>
<td></td>
<td>0.310337±0.13</td>
<td>3522.6±0.7</td>
</tr>
<tr>
<td>BA 62</td>
<td>Clear to light</td>
<td>1</td>
<td>98</td>
<td>1598</td>
<td>0.310602±0.11</td>
<td>3523.9±0.5</td>
</tr>
<tr>
<td>Mean of grains 1–4</td>
<td></td>
<td>1–4</td>
<td>364</td>
<td></td>
<td>0.310612±0.62</td>
<td>3524.0±0.3</td>
</tr>
<tr>
<td>BA 65</td>
<td>Clear to yellowish,</td>
<td>1</td>
<td>64</td>
<td>1602</td>
<td>0.288251±0.16</td>
<td>3408.2±0.7</td>
</tr>
<tr>
<td></td>
<td>Brown, long-</td>
<td>2</td>
<td>64</td>
<td>1600</td>
<td>0.288265±0.81</td>
<td>3408.3±0.4</td>
</tr>
<tr>
<td>Mean of grains 1–3</td>
<td></td>
<td>1–3</td>
<td>193</td>
<td></td>
<td>0.288238±0.54</td>
<td>3408.1±0.3</td>
</tr>
<tr>
<td>BA 70</td>
<td>Clear, thin, long-</td>
<td>1</td>
<td>66</td>
<td>1605</td>
<td>0.310381±0.86</td>
<td>3522.8±0.4</td>
</tr>
<tr>
<td></td>
<td>Prismatic,</td>
<td>2</td>
<td>88</td>
<td>1599</td>
<td>0.310536±0.64</td>
<td>3523.6±0.6</td>
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<tr>
<td>Mean of grains 1–4</td>
<td></td>
<td>1–4</td>
<td>220</td>
<td></td>
<td>0.310548±0.56</td>
<td>3523.6±0.3</td>
</tr>
<tr>
<td>BA 71</td>
<td>Dark-red-brown,</td>
<td>1</td>
<td>75</td>
<td>1604</td>
<td>0.311338±0.69</td>
<td>3527.6±0.3</td>
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<tr>
<td>Mean of grains 1–4</td>
<td></td>
<td>1–4</td>
<td>252</td>
<td></td>
<td>0.311470±0.35</td>
<td>3528.2±0.2</td>
</tr>
<tr>
<td>BA 73</td>
<td>Clear to yellowish,</td>
<td>1</td>
<td>52</td>
<td>1599</td>
<td>0.310174±0.02</td>
<td>3521.8±0.5</td>
</tr>
<tr>
<td>Mean of grains 1–3</td>
<td></td>
<td>1–3</td>
<td>220</td>
<td></td>
<td>0.310241±0.53</td>
<td>3522.1±0.3</td>
</tr>
<tr>
<td>BA 72</td>
<td>Clear, thin, long-</td>
<td>1</td>
<td>95</td>
<td>1595</td>
<td>0.312116±0.87</td>
<td>3531.4±0.4</td>
</tr>
<tr>
<td></td>
<td>Prismatic,</td>
<td>2</td>
<td>44</td>
<td>1599</td>
<td>0.312437±0.87</td>
<td>3531.0±0.4</td>
</tr>
<tr>
<td>Mean of grains 1–3</td>
<td></td>
<td>1–3</td>
<td>221</td>
<td></td>
<td>0.312273±0.71</td>
<td>3532.2±0.4</td>
</tr>
<tr>
<td>BA 74</td>
<td>Clear,</td>
<td>1</td>
<td>74</td>
<td>1608</td>
<td>0.312382±0.12</td>
<td>3532.7±0.6</td>
</tr>
<tr>
<td>Mean of grains 1–6</td>
<td></td>
<td>1–6</td>
<td>581</td>
<td></td>
<td>0.312468±0.43</td>
<td>3531.2±0.2</td>
</tr>
<tr>
<td>BA 85</td>
<td>Clear,</td>
<td>1</td>
<td>128</td>
<td>1597</td>
<td>0.310002±0.37</td>
<td>3525.9±0.2</td>
</tr>
<tr>
<td>Mean of grains 1–3</td>
<td></td>
<td>1–3</td>
<td>302</td>
<td></td>
<td>0.310867±0.58</td>
<td>3525.2±0.3</td>
</tr>
<tr>
<td>BA 86</td>
<td>Grey-brown,</td>
<td>1</td>
<td>127</td>
<td>1600</td>
<td>0.312022±0.48</td>
<td>3531.0±0.2</td>
</tr>
<tr>
<td>Mean of grains 1–4</td>
<td></td>
<td>1–4</td>
<td>382</td>
<td></td>
<td>0.311919±0.37</td>
<td>3530.4±0.2</td>
</tr>
<tr>
<td>BA 105</td>
<td>Clear,</td>
<td>1</td>
<td>110</td>
<td>1601</td>
<td>0.311212±0.45</td>
<td>3526.9±0.2</td>
</tr>
<tr>
<td>Mean of grains 1–5</td>
<td></td>
<td>1–5</td>
<td>656</td>
<td></td>
<td>0.311207±0.18</td>
<td>*3526.9±0.2</td>
</tr>
<tr>
<td>BA 106</td>
<td>Clear,</td>
<td>1</td>
<td>104</td>
<td>1608</td>
<td>0.311834±0.74</td>
<td>3530.0±0.4</td>
</tr>
<tr>
<td>Mean of grains 1–4</td>
<td></td>
<td>1–4</td>
<td>356</td>
<td></td>
<td>0.311852±0.40</td>
<td>3530.1±0.2</td>
</tr>
</tbody>
</table>

(continued on next page)
The origin of the overlying, brecciated unit is more obscure. However, the monolithic nature of the ‘clasts’ and the petrographic evidence that the ‘clasts’ are similar in composition to the matrix suggests that this rock is a felsic agglomerate, as with other rocks in the Theespruit Formation. Such an origin is supported by the fact that this unit passes along strike to the northwest, as well as across strike, into the felsic volcanic rocks described above.

4.1.4.3. ‘Basement orthogneiss’ cobbles in ‘diamictite’. The third setting in which ‘basement orthogneisses’ were identified by de Wit et al. (1983), was as boulders within ‘sedimentary diamictite’. Although the boulder depicted in Fig. 6a of de Wit et al. (1983) could not be relocated, boulders of identical appearance and texture, with a similar misoriented compositional layering with respect to the regional foliation, were observed at several places (e.g. Fig. 5b). All such boulders consist of fine-grained, siliceous, weakly and irregularly colour-layered rock, identical to the grey-white cherty metasediment (siliﬁed tuf?) in the lowermost felsic unit. The irregular layering in the boulders is defined only by a change in colour, not in grain size or composition, as both grey and white layers are composed of >95% fine-grained silica. Locally, the darker layers are formed by remobilisation of grey-black chert into fractures within a white-cream quartz siltstone/chert. Again, no evidence of leucosome veins, metamorphic differentiation, or even a granitic composition was observed in the boulders to confirm an origin as orthogneisses. As such, and because of the volcanic origin of the host agglomerate, the layered boulders are re-interpreted here as fragments of underlying sediment entrained in felsic magma during explosive volcanism.

4.2. Strain

4.2.1. Granite-greenstone contact

In the map area, the Theespruit Pluton is a leucocratic, medium-grained trondhjemite that intrudes the greenstones as a series of little-deformed, bedding-subparallel sheets (sills). Here it represents the structurally highest and least deformed part of the pluton and has intruded the highest stratigraphic level in the greenstones. Anhaeusser (1984) and Kisters and Anhaeusser (1995) described contact agmatites and ring dikes in other low-strain, high-level parts of the pluton; the ring dikes are parallel to bedding and thus probably represent sills turned on edge together with greenstones, during doming.

Central parts of the pluton to the south of our map area are little deformed, medium- to coarse-grained leucocratic trondhjemite. Large xenoliths of amphibolite-facies pillow basalt in the core of the pluton are almost completely undeformed, although they are gently tilted and face radially away from the core of the pluton, indicating its domical shape (Fig. 7a). Greenstones along the margins of the pluton were transformed into strongly foliated and linearized, medium- to coarsely recrystallised amphibolites and host a transposed network of granite veins emplaced in part during folding. In the narrow septum of greenstones between the Theespruit and Stolzburg plutons (Anhaeusser, 1984; Kisters and Anhaeusser, 1995), all rocks have been intensely deformed into subvertical L-tectonites (Fig. 7b) under metamorphic conditions up to granulite-facies (9 kbar, 700 °C; Van Vuren and Cloete, 1995). The steeply dipping margins of the Theespruit Pluton are strongly foliated and heterogeneous, with multiple foliated sheets of variable composition and textural complexity (Fig. 7c). Such variation was explained by Anhaeusser (1984) as due to progressive and differential assimilation of marginal greenstones by successive intrusive trondhjemite sheets and comcomitant deformation of earlier sheets during successive intrusive pulses under ductile conditions imposed by the heat of the magmas.

4.2.2. Theespruit Formation and KSZ

Rocks in the Tjakastad region vary greatly in strain. In the southwest, adjacent to the Theespruit Pluton, little deformed rocks with well-preserved sedimentary and/or volcanic textures are preserved (e.g. pillows, ocelli, intra-pillow hyaloclastite breccia, and pillow shelves). Strain increases progressively towards the southern contact of the KSZ, which is marked by the Theespruit Fault, filled by a ≤12 m wide quartz vein. The southern 300 m of the KSZ is underlain by high strain ultramylonites derived from mafic, felsic and cherty precursors (Fig. 8a). Strain then decreases again northeast through the remainder of the KSZ until the Komati Fault, a locally sharply deformed structure (3:2:1) in which elongation lineations plunge moderately to the northwest (Fig. 8b). A narrow high strain zone is developed along the southern contact of the first felsic horizon, as noted by de Wit et al. (1983). North of this contact, lineations become more steeply plunging and more northerly trending in rocks of variable preservation, but generally the succession becomes more strongly deformed towards the KSZ. The trend and plunge of extension lineations across this zone vary between different rock types, as shown in Fig. 8b, only becoming fully transposed within the southern part of the KSZ where chloritic ultramylonites contain subvertical, NNW-SSE striking schistosities and downplunge
mineral elongation lineations. In the northeastern part of the study area, rocks of the KSZ are pure L-tectonites, with strongly developed vertical rodding lineations (Fig. 7d). In thin sections of the most highly strained schists, kinematic asymmetry could not be identified due to intense flattening strain. However, in slightly coarser-grained porphyroblastic and originally porphyritic rocks along the northern margin of the KSZ, kinematic indicators of SW-side-up displacement were observed (Fig. 7e). Northeast of the ultramylonitic schists, in the western part of the KSZ, strain decreases and local sedimentary features can again be recognised within felsic tuffaceous horizons. Pillow facing directions to the northeast were identified in basalts. In the eastern part of the KSZ, the strain is more heterogeneous, with high strain shear splays near its northern boundary (Fig. 4).

The Komati Fault on the northern margin of the KSZ is exposed in the western part of the map area, to the south of which SW-side-up kinematic indicators were observed in serpentinized, schistose metakomatiite (Fig. 8). In the eastern part of the map area, the Komati Fault is represented by a quartz vein that separates L-tectonite mylonitic felsic agglomerates from undeformed quartz-feldspar porphyry at the base of the Komati Formation (Fig. 4).

4.3. Metamorphism of the Theespruit Formation in the Tjakastad area

Metamorphic grade decreases from amphibolite-facies adjacent to the Theespruit Pluton (garnet-actinolite pillowed mafic volcanics), to greenschist-facies in the KSZ, where mafic mylonites contain chlorite-carbonate assemblages, felsic schists consist of sericite and quartz, and quartz-plagioclase porphyries contain chloritised riebeckite. As described above, P–T estimates decrease up stratigraphy from the western margin of the Theespruit Pluton (≤9 kbar, 700 °C; Van Vuren and Cloete, 1995) to the top of the Hooggenoeg Formation (Fig. 2: 1.9 kbar, 320–420 °C; Cloete, 1993, 1999).

The stratigraphically lowest felsic horizon in the Theespruit Formation contains abundant kyanite porphyroblasts that are elongate in the lineation direction and syn- to late-kinematic with respect to the deformation. Andalusite porphyroblasts contain round inclusions of an altered mineral that now consists of fine-grained quartz-sericite and which may have originally been cordierite. Andalusite porphyroblasts are commonly overgrown by kyanite, along extension cracks, indicating replacement of the former by the latter under conditions of increasing pressure (Fig. 7f). The initial assemblage andalusite–cordierite indicates P–T conditions of ~3.9 kbar, 550 °C (Spear, 1993), whereas the presence of overgrowing kyanite indicates an increase in pressure under near-isothermal conditions along part of an anticlockwise P–T path. As discussed below, these data have important implications for the tectonic evolution of the BGB.

5. Zircon geochronology and Nd isotopic systematics

Single zircons were dated from 13 felsic Theespruit lithologies and from one intrusive granite in the Tjakastad area, using the evaporation technique (Kober, 1986, 1987). The sensitive high-resolution ion microprobe (SHRIMP II; DeLaeter and Kennedy, 1998) and conventional analysis by thermal ionisation mass spectrometry (TIMS)
techniques were used for zircons from one sample. Sample preparation for the evaporation technique followed methods described in Kröner et al. (2003). In addition, we determined the Sm–Nd whole-rock isotopic composition of selected samples in order to calculate Nd model ages and \( e_{Nd(t)} \) values. The different analytical procedures are summarized in Appendix A. Sample locations and short field characteristics are given in Table 1 and Fig. 4.

Samples BA56, 57, 62, 70, 71, 73, 79, 84, 85, 104, 105, 107, 108, 109 and 110 represent variably foliated felsic volcanic schist layers of the Theespruit Formation in the Tjakastad and Songimvelo areas (Fig. 4). We also collected mafic rocks from the base of the succession, including BA 66, 69 (massive foliated amphibolite), 74 (pillowed basaltic komatiite) and 75 (gabbroic sill). Samples BA 79, 84, 85, 109 and 110 represent Theespruit felsic tuff or trachytic breccia interlayered with chert, along strike to the east of the main Tjakastad samples. The zircons in all samples vary in colour from clear to yellow-brown and in shape from stubby to long-prismatic. Most are idiomorphic, occasionally with slight rounding at their terminations. No core-overgrowth relationships were seen under the microscope. Between 3 and 6 zircons per sample were evaporated, and the mean \( ^{207}\text{Pb}/^{206}\text{Pb} \) ages obtained on 13 samples vary within a narrow range between 3522 and 3533 Ma, with very low errors (generally <1 Ma; see Table 2 and Figs. 9–10).

Felsic pyroclastic sample 160932 was collected from the middle of the three main felsic volcaniclastic units identified in the Tjakastad area (Fig. 4). The horizon is a coarse felsic volcaniclastic rock that contains irregular-shaped fragments of fine-grained and feldspar-phlytic dacite, and rounded cobbles (to 20 cm diameter) of grey-black and white layered chert in a fine-grained felsic volcanic matrix (Fig. 5b). The reasoning behind dating this sample was that it is from the most heterogeneous felsic layer in the region and thus should have the most varied detrital zircon population, if derived from a sedimentary diamictite with basement orthogneiss clasts, as previously suggested.

The sample contained abundant zircon grains with a single population of dark- to medium-brown, elongate prisms that have common bipyramidal terminations and internal growth zoning. The recovered zircon grains are commonly cracked and often broken, but show no cores or rims. Seven zircon grains were analysed on SHRIMP II (Table 3 and Fig. 11a). Grain 1 is grossly discordant and probably lost Pb at unspecified periods in the past; it is therefore not considered for age assessment. Grains 2–7 are variably discordant but are well aligned along a chord (MSWD = 0.0025) with an upper Concordia intercept at 3539±16 Ma (2-sigma) and a lower intercept at 0±450 Ma. The weighted mean \( ^{207}\text{Pb}/^{206}\text{Pb} \) age of the six grains is 3539±2 Ma. We consider this to reflect the time of eruption of the felsic pyroclastic rock.

TIMS analysis of five single zircons from the same sample (Table 4) yielded three grossly discordant points (Z2, Z4, Z5) which are not further considered here, one near-concordant point (Z3) and one moderately discordant point (Z1) (Fig. 11b). A chord through grains Z3 and Z1 suggests recent Pb-loss (Fig. 11b), and we therefore consider both analyses meaningful. The mean \( ^{207}\text{Pb}/^{206}\text{Pb} \) age for both Z3 and Z1 is 3534±4 Ma, which is identical, within error, to the mean SHRIMP age. Combining all data points (SHRIMP and TIMS) yields a regression with an upper concordia intercept age of 3534±1 Ma.

In view of the slightly younger age of zircons from felsic volcaniclastic rocks analysed by evaporation as compared to the SHRIMP and TIMS results, we cannot completely rule out that non-zero lead loss has occurred in the younger zircons, although the grains from individual samples all show near-identical \( ^{207}\text{Pb}/^{206}\text{Pb} \) ratios, and our oldest evaporation age of 3533 Ma is identical, within error, to the mean SHRIMP U/Pb zircon age of 3531±10 Ma reported by Armstrong et al. (1990). We therefore adopt a mean age of c. 3530 Ma for the 14 samples of Theespruit felsic volcanic rocks analysed from the Tjakastad–Songimvelo area for the purpose of calculating Nd initial isotopic ratios and \( T_{IM} \) model ages. It is important to note that no ca. 3450 Ma zircons were identified during any of these analyses.

Evaporation of three idiomorphic zircon grains from granite dyke sample BA 65 yielded identical Pb isotopic ratios which combine to a mean \( ^{207}\text{Pb}/^{206}\text{Pb} \) age of 3408±0.3 Ma (Table 2, Fig. 9d). We consider this age to reflect the time of dyke emplacement, and the dyke may reflect either a late, K-feldspar-rich phase of the Theespruit Pluton, or
part of a phase of granite and aplite emplacement found in all granitic rocks of the southern Barberton Mountain Land. Importantly, the main phase of deformation in the dyke and the Theespruit Formation must be post-3408 Ma.

Whole-rock samples analysed for Sm and Nd isotopes (Table 5) were not, in all cases, the same as those used for zircon dating, since some rocks did not contain large or sufficient zircons for evaporation. In addition to felsic volcanic rocks of the Theespruit Formation, we also measured isotopic values for mafic/ultramafic samples from the Theespruit Formation, near outcrops of well-preserved pillow basalts that are interlayered with the felsic volcanic rocks.

The $\varepsilon_{\text{Nd}}(t)$ values for Theespruit felsic rocks (Table 5 and other data in Hamilton et al., 1979 and Kröner et al., 1996) show a moderate scatter from −1.4 to +1.3 (Fig. 12a), suggesting heterogeneous crustal sources, including juvenile mantle-derived material, as well as recycling of older crust. Although we cannot preclude disturbance of the Sm–Nd system in these samples, their high Sm and Nd concentrations probably inhibited drastic modifications resulting from fluid/rock interaction (e.g., Grou et al., 1996). A negative $\varepsilon_{\text{Nd}}$ value for the pillow metakomatiite sample BA 74 may be explained by sample alteration and possibly melt contamination by older crustal material.

The somewhat lower $\varepsilon_{\text{Nd}}(t)$ in the felsic samples than in the associated komatiites suggest that these rocks were not comagmatic, and that the felsic rocks were derived from a crustal source that was ultimately formed from the same mantle as the associated komatiites. This is consistent with previous Nd isotopic data for Theespruit felsic volcanics from the Steyndorp area farther east that also suggested involvement of older continental crust (Kröner et al., 1996).

A regression analysis of three komatiitic samples (excluding BA 74, which has negative $\varepsilon_{\text{Nd}}$ values of −0.4 at 3.53 Ga that we infer is due to melt contamination by older crust, or alteration of the Sm–Nd system by hydrothermal activity that affected the pillow basalts; e.g., de Wit et al., 1982; Duchač and Hanor, 1987), yields an age of 3.51 ± 0.07 Ga (2σ; MSWD = 0.02) (Fig. 12b). Enlarging the MSWD to a value of 1 would reduce the error to c. 10 Ma. This age is identical to the emplacement age of the Steyndorp Pluton (3510 Ma; Kröner et al., 1996) and not much different from the mean zircon age of c. 3530 Ma for the Theespruit Formation obtained herein. This agreement suggests that the Sm–Nd system in these rocks behaved as a closed system during their post-emplacement history. We therefore interpret the initial $\varepsilon_{\text{Nd}}$ values of c. +0.5 ± 1.4 (2σ) for the mantle-derived komatiites as representative of their mantle source. The Theespruit mantle value is not much different from that of other units of the Onverwacht Group, which yielded average $\varepsilon_{\text{Nd}}$ values of +2.5 (Lécuyer et al., 1994), +1.9 (Chavagnac, 2004), +1.1 (Lahaye et al., 1995), and +1.1 (Hamilton et al., 1979), with a total range in values from +3 to 0 (Fig. 12a).

A regression line calculated for the Theespruit mafic and associated felsic metavolcanic rocks yields an errorchron coefficient to an age of 3.63 ± 0.13 Ga (2σ; n = 9, MSWD = 16; initial $\varepsilon_{\text{Nd}}$ value = +1.1 ± 2.1, 2σ). Addition of four samples of the Theespruit Formation from the Steyndorp area (Kröner et al., 1996) produces an errorchron age of 3.60 ± 0.13 Ga (2σ; MSWD = 21, initial $\varepsilon_{\text{Nd}}$ value = +0.8 ± 2, 2σ) (Fig. 12b). The near-magmatic age derived from the composite regression line indicates that both the crustally-derived felsic rocks and mantle-derived komatiites were derived from overall isotopically similar sources. The inclusion of the felsite data in addition to the komatiitic samples in the regression analysis results in a rotation of the regression line to a higher age and increase in data scatter. The scatter of the felsite data may reflect open-system behaviour of the Sm–Nd system in the felsic rocks or their crustal protoliths.

Figure 11. a) Concordia diagram showing SHRIMP analyses of zircons from felsic tuff sample 160932 (Table 1). Data boxes for each analysis are defined by standard errors in 207Pb/206Pb, 206Pb/238U and 207Pb/206Pb; b) Concordia diagram showing TIMS analyses of zircons from sample 160932 (Table 4): recent Pb-loss line passes through zircons Z1 and Z3, with an upper intercept age of 3534 ± 4 Ma. Dated sample is shown in Fig. 5b from locality shown in Fig. 4.
system behaviour of the Sm–Nd system has been reported in many studies (Moorbath et al., 1997) and references therein; Gruau et al., 1996), and it may well be that fluid/rock interaction during metamorphism of the Theespruit Formation (see above) has affected the Sm–Nd system. However, any possible secondary disturbance of the Sm–Nd system had no drastic effects on the samples, as shown by the reasonable agreement between the whole-rock Sm–Nd ages and U–Pb zircon data.

6. Discussion

Remapping and dating of the Theespruit Formation and associated granitic rocks in the Tjakastad area indicates significant differences from previous tectonic interpretations of this area (de Wit et al., 1983). In particular, our data question the validity of the thrust-accretion model because:

1. No tectonic slivers or clasts of basement orthogneiss were found in the Tjakastad area. Those previously interpreted as such by de Wit et al. (1983) were found to consist either of undeformed quartz-feldspar porphyry with igneous flow banding, deformed metasediment, or clasts of cherty metasediment (metatuff?) within felsic agglomerate. Furthermore, there is no evidence to support the presence of an unconformity between “basement orthogneiss” and “sedimentary diamicite” at locality D in Fig. 3.

2. Rocks within the Theespruit Formation and KSZ that were interpreted as sedimentary diamicite and intrusion breccia by de Wit et al. (1983) are re-interpreted here to be felsic volcanoclastic rocks that form part of a series of felsic volcanoclastic horizons layered with mafic schists of indeterminate intrusive or extrusive origin. The unique compositional and textural features of each felsic horizon indicate that they are distinct from one another and thus do not represent slivers of a thrust-duplicated unit, as previously suggested (de Wit et al., 1983, 1992).

3. The presence of consistent top-to-the-northeast bedding indicators across the Theespruit Formation, in combination with the results of (1) and (2) above, strongly favour a coherent, though strained (see point 4, below) stratigraphic succession. The presence of local tight folds and of reversals of bedding in the KSZ (de Wit et al., 1983) is not considered to be a significant feature, as no large-scale folds of mappable units have been found.

4. Changes in the intensity and orientation of structures across the contact between mafic volcanic rocks and the first felsic horizon in the Theespruit Formation may readily be explained as an example of strain partitioning between the competent pillowed komatiitic unit and the well-layered, less competent, and more fully transposed felsic tuffaceous rocks (cf. figure 26.17 and related text of Ramsay and Huber, 1987; Hamner and Passchier, 1991 [pp. 22–24]). Northeast of this contact, the strain discontinuously increases towards the KSZ, in concert with more steeply-plunging and more northerly-trending lineations. These data suggest that the southern contact of the stratigraphically lowest felsic schist horizon represents the first major increase in strain approaching the KSZ and that the abrupt change in lineation orientations across this contact reflects varying degrees of transposition within a single shear system. Significantly, strain is highly variable within the KSZ, and largely dependent on lithology, with the highest strain localised in the softest (talc-carbonate) rocks, and areas of low strain localised in some pillowed basalts and felsic volcanic rocks (Fig. 4).

5. The distribution of strain, orientation of fabric elements, and kinematic data of discrete faults and shear zones within the Tjakastad area indicate a single set of penetrative fabric elements developed during greenstone belt–down, granitoid–up extensional deformation. No evidence of thrust kinematics, fold interference, or of multiple overprinting sets of structures was observed, and the period of earlier deformation within orthogneisses as interpreted by de Wit et al. (1983) has not been substantiated. Compilation of available data show that the extensional deformation occurred at ca. 3230 Ma. The KSZ is here interpreted as an extensional detachment zone, similar to that documented along strike to the west by Kisters et al. (2003) (see Fig. 1). As discussed below, we suggest that the most likely origin of the KSZ is as a dislocation plane to tight, upright, disharmonic folding between deep crustal levels dominated by the Stolzburg plutonic suite, and high level rocks of the rest of the Barberton Greenstone Belt.

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**Table 4**

| U-Pb TIMS data for sample 160932. |
|-----------------|-----------------|-----------------|-----------------|
| **Concentration** | **Measured** | **Corrected Atomic ratios**<sup>a</sup> | **Ages [Ma]** |
| **Weight** (mg) | **U** (ppm) | **Pb**<sup>206</sup>/<sup>207</sup>Pb | **Pb**<sup>206</sup>/<sup>204</sup>Pb | **Pb**<sup>207</sup>/<sup>204</sup>Pb | **Pb**<sup>208</sup>/<sup>204</sup>Pb | **Pb**<sup>206</sup>/<sup>238</sup>U | **Pb**<sup>207</sup>/<sup>238</sup>U | **Pb**<sup>208</sup>/<sup>238</sup>U | **U** (ppm) | **Sm**<sup>147</sup>/<sup>144</sup>Nd | **Sm**<sup>146</sup>/<sup>144</sup>Nd | **Nd**<sup>146</sup>/<sup>144</sup>Nd | **ɛ<sup>Nd</sup>(T) | **TOM** |
|-----------------|-----------------|-----------------|-----------------|
| Z1 brn euh frag | 0.001 236 126.7 2 | 8357 0.1994 | 0.70423 (156) 30.364 (698) | 0.31271 (26) | 2347 3499 3534 |
| Z2 sub prsm | 0.001 966 411.2 2 | 1284 0.2425 | 0.34605 (72) 8.8451 (176) | 0.18538 (26) | 1916 2322 2702 |
| Z3 sub prsm | 0.010 17 16.6 4 | 1905 0.2640 | 0.72643 (208) 33.1979 (884) | 0.31270 (38) | 3520 3529 3534 |
| Z4 prsm | 0.001 267 164.4 4 | 1600 0.2227 | 0.48660 (134) 16.6645 (460) | 0.24838 (32) | 2556 2916 3174 |
| Z5 prsm | 0.001 284 171.2 29 | 331 0.2628 | 0.46872 (104) 14.7999 (364) | 0.22901 (28) | 2478 2802 3045 |

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**Table 5**

| Sm–Nd isotopic data of metavolcanic rocks from the Theespruit Formation. |
|-----------------|-----------------|-----------------|-----------------|
| **Sample** | **Age (Ga)** | **Sm**<sup>147</sup>/<sup>144</sup>Nd | **Nd**<sup>146</sup>/<sup>144</sup>Nd | **ɛ<sup>Nd</sup>(T)** |
|-----------------|-----------------|-----------------|-----------------|

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<sup>a</sup> Ratios corrected for fractionation, laboratory Pb (1 pg) and U (0.25 pg) blank. Initial common Pb calculated using Pb isotopic compositions of Stacey and Kramers (1975). Two-sigma uncertainties on isotopic ratios, calculated with a modified unpublished error propagation program written by L. Heaman, are reported after the ratios (in brackets) and refer to the final digits.
7. The geometry of weakly strained crests and strongly sheared, dip-lined margins to the TTG domes to the south of the Theespruit Formation (e.g., Fig. 7a–c) is consistent with partial convective overturn of a thick, overlying greenstone succession and underlying, more buoyant and partially melted TTG, as originally proposed by Viljoen and Viljoen (1969), and subsequently supported by Anhaeusser (1984) and Kisters and Anhaeusser (1995), possibly within an overall extensional regime (Kisters et al., 2003). This structural geometry developed at between c. 3237 and 3216 Ma, which is the age range of metamorphic zircons and titanite, respectively, from the Stolzburg and Doornhoek Plutons, and contemporaneous with the age of the post-tectonic Dalmein Pluton (Kamo and Davis, 1994).

6.1. Significance of the available geochronology

Our new interpretation for the Theespruit–Komati Formations contact zone contrasts markedly with previous tectonic models based on geochronological data and this requires some discussion.

Previous tectonic models based on zircon dating suggested that thrusting occurred at ca. 3230 Ma (Kamo and Davis, 1994). However, this is negated by the fact that both the higher-grade Theespruit and Sandspruit Formations structurally beneath the KSZ, and the lower-grade greenstones above the KSZ (Komati Formation and younger formations), experienced a common felsic magmatic event at ca. 3450 Ma (Kröner et al., 1991; Kamo and Davis, 1994), and thus were in contact at, or before, this time. If thrusting did occur, then it must have been at, or before, ca. 3450 Ma. However, several lines of evidence preclude such an interpretation. First, only one phase of deformation is recognised in the area, and that is extensional, and/or related to partial convective overturn of greenstones and granitoids. Second, an early thrust scenario is improbable because of the conformable relations between the Komati Formation and overlying rocks of the Hooggenoeg (3445 Ma), Kromberg (≤3416–3334 Ma), and Mendon (3298 Ma) Formations, and Fig Tree (<3260 Ma) Group (Fig. 2: Byerly et al., 1996; Lowe and Byerly, 2007), which shows that the greenstone sequences were deposited progressively from at least 3490 to 3260 Ma. Third, the 3408 Ma age of the granite dyke that we dated from the Tjakastad area (sample BA 65) is significant because it cuts bedding in the volcanic rocks and is deformed, indicating deformation occurred after its emplacement.

The main deformation of the belt at ca. 3230 Ma was extensional, as evidenced by road lineations, extensional faults, and kinematic data from within the KSZ (this paper and Kisters et al., 2003). The 3418 ± 8 Ma titanite age cited by Dziggel et al. (2001) is most likely related to late stage felsic magmatism in the area, as indicated by the 3408 ± 2 Ma Pb-Pb zircon age for the foliated granite dyke sample BA 65 in the Tjakastad area.

The zircon ages of Armstrong et al. (1990) and Kamo and Davis (1994) indicate that the unit previously interpreted as a basement gneiss sliver, but which we re-interpret to represent a felsic volcaniclastic rock within the Theespruit Formation, was erupted at 3538 Ma. This is the same age as our felsic schist sample 18 samples (Tables 3, 4; Fig. 11) and the ages of 13 samples of felsic schist of the Theespruit Formation located along strike in the map area and to the east around the Steynsdorp Anticline (Kröner et al., 1996), as well as for the older population of zircons in the “diamicite” (herein re-interpreted as felsic agglomerate) from the Tjakastad area (3531 ± 10 Ma; Armstrong et al., 1990). Based on this remarkable similarity in age for the same lithostratigraphic unit now dated for 18 samples (this paper and Kröner et al., 1996), it is clear that 3530 Ma represents the true age of felsic volcanism in the Theespruit Formation.

The remaining question then becomes what is the origin and significance of the 3453 ± 6 Ma population of zircons within the
The description of zircon morphology by Armstrong et al. (1990), combined with subsequent information in Kamo and Davis (1994), allows for a reasonable alternative interpretation.

First, it must be recognised that the dated rock was metamorphosed under amphibolite-facies conditions, at which grade metamorphic recrystallisation of zircon and growth of new zircon rims on older cores are known to occur (e.g., Hoskin and Schaltegger, 2003). The observation by Armstrong et al. (1990) of core–rim structures in zircons from their dated sample suggests that the younger population of zircons may have grown during the contact metamorphic event associated with the emplacement of the ca. 3450 Ma Stolzburg plutonic suite (e.g., Van Vuren and Cloete, 1995) and associated felsic intrusions that span the Komati schist zone. Supporting evidence for such an event was presented by Kamo and Davis (1994) who identified major felsic igneous activity in the Barberton region at 3476–3440 Ma based on the following ages (listed from structural base to top): 3460 ± 5/−4 Ma zircon in the older phase of the Stolzburg Pluton, which is located immediately west of, and connected to, the 3443 ± 4/−3 Ma Theespruit Pluton (Fig. 1); 3458 ± 2 Ma titanite in a 3476 ± 10 Ma felsic dyke cutting the Komati Formation; 3457 ± 3 Ma zircon from a felsic sill emplaced into the base of the Komati Formation; and ca. 3445 ± 5 Ma from the Hooggenoeg Formation (de Witt et al., 1987b), which conformably overlies, and is genetically related to, the Komati Formation (Lowe and Byerly, 2007).

We therefore re-interpret the geochronological information from the Theespruit Formation to indicate that a bimodal, mafic and felsic volcanic succession was erupted at ca. 3530 Ma, and then intruded by the Steyndorp Pluton at 3510 Ma. Following eruption of the Komati Formation at 3490–3480 Ma, the emplacement of sheets of felsic plutonic rocks into the volcanic succession at between 3476 and 3443 Ma caused contact metamorphism and the growth of new zircons and titanite in the host rocks, particularly lower in the section where the volumetrically significant Stolzburg plutonic suite was emplaced and resulted in amphibolite-facies contact-style metamorphism.

6.2. Origin of the Komati schist zone and regional folding

The regional geology of the southwestern Barberton Greenstone Belt (BGB) is dominated by a set of folds that plunge inwards towards the core of the belt, and are defined by lobate, antiform granitoids and cuspate, generally synclinal greenstones (Fig. 13). The strike of fold axial traces rotates through 130° around the southwestern corner of the BGB, from almost due north-striking in the southeast, to northeast-striking in the southwest, to southeast-striking in the northwest (Fig. 13). The lobe-cusp fold geometry of this area indicates contemporaneous deformation of competent and incompetent lithologies, respectively. The age of this folding is tightly constrained by the fact that folds involve the 3229 ± 5 Ma Kaap Valley Tonalite, but are cut by the 3216 ± 2/−1 Ma Dalmein Pluton (Kamo and Davis, 1994: Fig. 1).

Within this broad fold context, the high-strain Komati schist zone is located within the hinge of a regional, upright anticline, along the boundary between a thick, upper succession of low grade, low strain, volcanic flow units of the post-Theespruit Formation rocks of the Onverwacht Group (≤ca. 3.49 Ga) and a thinner, lower succession of amphibolite-facies, older greenstones (c. 3.53 Ga Theespruit and Sandspruit Formations) and ca. 3.45 Ga trondhjemitic intrusions. Indeed, the KSZ is located directly along the amphibolite-green schist isograd (Fig. 4), where ultramafic rocks and greenstones are transformed into soft tical-carbonate and chlorite schists, respectively, mineral assemblages perfectly suited to accommodate large degrees of shear strain. Kinematic data indicate extensional, top to the east-northeast shearing along this zone. Kisters et al. (2003) documented an eastward-dipping, top-to-the-east-northeast extensional detachment plane at the same structural level as the KSZ along strike to the west (Figs. 1, 13), which can be considered the continuation of the same zone. A similar high strain zone occurs at the same structural level along strike to the east, separating the Steyndorp Pluton and its surrounding envelope of Theespruit Formation rocks from lower-grade, younger rocks of the upper Onverwacht Group in another hinge of a tight regional anticline (Fig. 1). Given that this zone separates younger rocks in the north from older, higher-grade rocks in the southwest, it is most likely that the eastern zone is also extensional; thus, we consider all of these zones to be a continuous structural dislocation zone across the southwestern Barberton Greenstone Belt, intimately related to fold development.

Significantly, these dislocation zones developed at the same time as folding of the low-grade greenstones of the post-Theespruit Formation units of the Onverwacht Group, and also at the same time as the dome-and-keel geometry and peak metamorphic conditions were being developed in the higher grade rocks (see 2.3; Metamorphism: Dziggel et al., 2001, 2002; Diener et al., 2006).

Studies of fold mechanics in buckled multilayers with units of different competencies show that high-strain dislocation zones commonly develop along unit interfaces and that folding may be accommodated by a variety of processes, including disharmonic folding if units are of significantly different thicknesses (e.g., Ramsay and Huber, 1987). Considering the contemporaneity of fold formation and shear zone development, the most likely origin of the KSZ and its equivalents along strike to the east and west, is as an extensional detachment surface during regional folding. That folds in the upper Onverwacht Group are tighter than folds of the KSZ and its equivalents along strike, but that folding also affected the shear zones themselves, suggests that folding was contemporaneous with shearing. As discussed below, the fact that the rocks above the detachment surface display one style (upright, large amplitude and large wavelength folds) and those below display another style (dome-and-keel geometry: see below) suggests disharmonic folding and/or a different response at different structural levels to similar overall driving forces.

In the upper-crustal, lower-grade rocks, the inward-plunging nature of the folds around the southwestern BGB into the core of the greenstone belt, together with the greenstone-down kinematics across the shear zones, strongly suggest that deformation was driven by greenstone sinking, as previously suggested for this area (Viljoen and Viljoen, 1969; Anhaeusser, 1984). Greenstone sinking was accompanied by the exhumation of deeper level rocks beneath the detachment zones, as indicated by the presence of higher pressure metamorphic mineral assemblages in these rocks. Exhumation along the Komati schist zone may have excised a considerable thickness of stratigraphy between the Theespruit and Komati Formations, although this is unconstrained.

In the higher-grade rocks, the presence of vertically-rodde L-tektonites in strongly deformed greenstone keels between low-strain, domed granitoid cores is consistent with deformation driven by greenstone sinking. This is in contrast to, and different from, deformation generated as a result of granite doming (diapirism), where greenstones in the roof zones of diapirc granitoids should be extensively flattened (e.g. Ramberg, 1967); however, this is clearly not the case, as shown in Fig. 7a. Greenstone sinking is confirmed by the occurrence of high-P metamorphic mineral assemblages of 9 kbar, 700 °C in highly transposed greenstone septae between domes (Fig. 13: Van Vuren and Cloete, 1995). The low geothermal gradients indicated by metamorphic studies in this area are also consistent with greenstone sinking (cf. Collins and Van Kranendonk, 1999).

Some aspects of the geology of this area have similarities with metamorphic core complexes, as suggested by Kisters et al. (2003). However, the fact that folds and metamorphic mineral elongation and stretching lineations point inwards towards the core of BGB contrasts with the typically unidirectional trend of linear fabric elements within metamorphic core complexes, as does the dome-and-keel geometry of folding within the higher-grade footwall in the Tjakastad area. Other
differences between the geology of the Barberton Mountain Land and Phanerozoic metamorphic core complexes include the lack of evidence in the former for thinning of the lithosphere during extension, and for a prior episode of structural thickening of the lithosphere through orogenesis (see also Hickman and Van Kranendonk, 2004; Van Kranendonk et al., 2004).

6.3. Regional orogeny

Of critical importance relating to interpretations of the tectonic evolution of the BGB, is the contrasting evidence for contemporaneous extensional deformation (KSZ and associated zones along strike) and compressional deformation at c. 3230 Ma. Extension is indicated by the KSZ and associated zones along strike, and by the dome-and-keel architecture of c. 3.45 Ga trondhjemitic rocks along the southwestern margin of the belt. Compressional deformation is indicated by the presence of upright folds around the southwestern BGB, syn-sedimentary contractional folds (including recumbent isoclinal folds) in the Moodies Group (e.g. Lamb, 1987), and horizontal-tectonics style structures in the Fig Tree Group (e.g., de Wit, 1982). Significantly, extensional structures are restricted to the higher-grade, lower structural and stratigraphic parts of the BGB below the KSZ detachment, whereas compressional structures are restricted to higher structural and stratigraphic levels within the lower-grade core of the BGB, above the KSZ detachment. Whereas this contemporaneity of contrasting styles does occur in Phanerozoic plate-collisional orogens, extension in these terranes follows significant crustal thickening (crustal doubling) through thrusting, caused by the collision of tectonic plates (e.g. Searle, 2007).

We contend that our new data, combined with the evidence of an upward-younging, thick greenstone stratigraphy, precludes an origin of the BGB through subduction–accretion, as widely supposed (e.g. de Wit et al., 1992; Lowe, 1994; Stevens et al., 2002; Moyen et al., 2006). Rather, we suggest that the available data supports an origin of the BGB through partial convective overturn of a dense upper crust and more buoyant middle crust at c. 3230 Ma, similar to models previously proposed by Viljoen and Viljoen (1969) and Anhaeusser (1984).

We envisage a two-stage development of ancient Kaapvaal Craton crust.

1) Long-lived, predominantly mantle-derived magmatism from 3530–3416 Ma, which resulted in the construction of a thick greenstone pile, based on a substrate of older continental crust, perhaps including the ≤c. 3.64 Ga Ancient Gneiss Complex (AGC) in Swaziland, on the southeastern side of the BGB (Kröner, 2007). This was further thickened by the emplacement of c. 3470–3437 Ma TTG melts, for which geochemical data indicates melting under a range of pressures from a dominantly infracrustal basaltic source (Moyen et al., 2007); this implies that the Kaapvaal crust was already thick by this time. Emplacement of TTG into the mid-crust caused low-pressure metamorphism of adjacent greenstones (e.g. andalusite in the present study area).
2) Subsequent addition of mantle-derived melts at 3334–3298 Ma (Kromberg and Mendon Formations) resulted in magmatic over-thickening of the crust and development of a gravitational instability represented by a > 10 km thick, largely komatiitic volcanic pile on top of granitic (TTG and AGC) middle crust. Modelling studies show that this type of gravitationally unstable Palaeoarchaean crustal architecture, when combined with the input of conductive heat from the mantle melts and radiogenic heat in TTGs buried to mid-crustal levels by the associated eruptions, results in partial melting of the TTG middle crust, which, in turn leads to sinking of greenstones and doming of the partially melted TTG during partial convective overturn (Rey et al., 2003; Sandiford et al., 2004; Bodoros and Sandiford, 2006). We suggest that this mechanism, rather than subduction–accretion, was responsible for the deep burial of greenstones along the flanks of doming granitoids in the lower structural parts of the terrain, where high-pressure metamorphic assemblages have been recorded (e.g. Dziggel et al., 2002; Diener et al., 2006). Partial convective overturn was rapid, with associated isothermal decompression of the lower structural greenstones during extension-related exhumation at ca. 3230 Ma (Kisters et al., 2003; Dziggel et al., 2005). Uplift of the granitoid margins to the BGB shed detritus into the sinking greenstone core (Fig Tree and Moodies Groups), which became deformed as the sediments accumulated. Horizontal translation of the greenstones into the zone of sinking from adjacent to the rising granitoids was effected across mid-crustal detachment zones such as the KSZ and caused horizontal-style deformation of the upper greenstones (including recumbent isoclinal folds; e.g., Ramberg, 1967).

The model of partial convective overturn for the development of the BGB outlined above should be considered as a working, testable hypothesis and that it is not necessarily exclusive of subduction/accretion tectonics in the development of the craton as a whole, nor of Archaean plate tectonics in general (e.g., Van Kranendonk, 2004, 2007; Smithies et al., 2005).

7. Conclusions

Remapping of the southwestern part of the Barberton Greenstone Belt, combined with zircon dating and Sm–Nd data, has revealed several important inconsistences with the previous horizontal tectonic model of de Wit et al. (1983, 1992). Most significantly, the kinematics of shear deformation in the Theespruit Formation is extensional rather than contractional, as previously proposed. Rocks previously interpreted as tectonic slivers of basement orthogneiss were found instead to be composed of little deformed quartz-feldspar porphyry or sheared felsic volcaniclastic metasediment, and those interpreted as sedimentary diamictite were found to be felsic agglomerates. Consistent facing directions of bedding to the northeast, the recognition of distinct stratigraphic variations between several felsic volcanic and volcaniclastic horizons across strike, and strain data all combine to suggest that the Theespruit Formation in this area is a ca. 3530 Ma old, moderately strained, coherent lithostratigraphic succession and not a tectono-stratigraphic assemblage as proposed by de Wit et al. (1983, 1992).

We re-interpret a 3458 Ma population of zircons dated from an amphibolite-facies Theespruit Formation felsic schist by Armstrong et al. (1990) as being of metamorphic, rather than detrital origin, based on these authors’ observation of core–rim structures in zircons, the amphibolite-facies metamorphic grade of the dated rocks, and our observations for the widespread emplacement of felsic plutonic rocks of that age across the Komati Schist Zone, which we suggest provided heat for contact metamorphism of the greenstones.

In combination with previous work (Byerly et al., 1996; Lowe and Byerly, 2007), the data presented here show that the Barberton Supergroup is an upward-younging, autochthonous succession deposited over ~300 Ma (3530–3230 Ma). Deformation commenced during deposition of the Moodies Group at ca. 3230 Ma, and involved partial convective overturn of upper and middle crust.

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Appendix A. Laboratory procedures

A.1. Single zircon evaporation

Our laboratory procedures, as well as comparisons with conventional and ion-microprobe zircon dating, are detailed in Kröner et al. (1991) and Kröner and Hegner (1998). Isotopic measurements were carried out on a Finnigan-MAT 261 mass spectrometer at the Max-Planck-Institut für Chemie in Mainz.

The calculated ages and uncertainties are based on the means of all ratios evaluated and their 2σ mean errors. Mean ages and errors for several zircons from the same sample are presented as weighted means of the entire population. During the course of this study we repeatedly analysed fragments of large, homogeneous zircon grains from the Palabora Carbonatite, South Africa. Conventional U–Pb TIMS analyses of six separate grain fragments from this sample yielded a 207Pb/206Pb age of 2052.2 ± 0.8 Ma (2σ; W. T. Todt, unpublished data), whereas the mean 207Pb/206Pb ratio for 19 grains, evaporated individually over a period of 12 months, was 0.126634 ± 0.000027 (2σ error of the population), corresponding to an age of 2051.8 ± 0.4 Ma, which is identical to the U–Pb TIMS age, within error. The above 2σ error of the population of evaporated zircons is considered the best estimate for the reproducibility of our evaporation data and corresponds approximately to the 2σ (mean) error reported for individual analyses in this study (Table 2). In the case of combined data sets, the 2σ (mean) error may become very low, and whenever this error was less than the reproducibility of the internal standard, we have used the latter value (that is, an assumed 2σ error of 0.000027).

The analytical data are presented in Table 2, and the 207Pb/206Pb spectra are shown in histograms that permit visual assessment of the data distribution from which the ages are derived. The evaporation technique provides only Pb isotopic ratios, and there is no a priori way to determine whether a measured 207Pb/206Pb ratio reflects a concordant age. Thus, all 207Pb/206Pb ages determined by this method are necessarily minimum ages. However, many studies have demonstrated that there is a very strong likelihood that these data represent true zircon crystallisation ages when (1) the 207Pb/206Pb ratio does not change with increasing temperature of evaporation and/or (2) repeated analyses of grains from the same sample at high evaporation temperatures yield the same isotopic ratios within error. Comparative studies by evaporation, conventional U–Pb dating, and ion-microprobe analysis have shown this to be correct (Kröner et al., 1991, 1999; Cocherie et al., 1992; Jaekel et al., 1997; Karabinos, 1997).

A.2. SHRIMP II analysis

U–Pb ion microprobe data for zircons from sample 160932 were obtained on the SHRIMP II (B) of the John de Laeter Centre of Mass
Spectrometry at Curtin University, Australia (DeLaeter and Kennedy, 1998). Clear euhedral zircons some 80 to 150 µm in length were handpicked and mounted in epoxy resin together with chips of the Perch zircon standard CZ3. For data collection, six scans through the critical mass range were made, and details on the analytical procedure and data reduction can be found in Compston et al. (1992), Stern (1997), Nelson (1997), and Williams (1998). Primary beam intensity was about 2.0 nA, and a Köhler aperture of 100 µm diameter was used, giving a slightly elliptical spot size of about 30 µm. Sensitivity was about 26 cps/ppm/nA Pb on the standard CZ3. Analyses of samples and standards were alternated to allow assessment of Pb+/U+ discrimination. Common-Pb corrections have been applied using the 206Pb-correction method, and because of low counts on 204Pb it was assumed that common lead is surface-related (Kinny, 1986), and the isotopic composition of Broken Hill lead was used for correction. The analytical data are presented in Table 3. Errors on individual analyses are given at the 1-sigma level and are based on counting statistics and include the uncertainty in the standard U/Pb age (Nelson, 1997). Errors for pooled analyses are at the 2-sigma or 95% confidence interval. The data are graphically presented on a conventional concordia plot (Fig. 11a).

A.3. TIMS analysis of zircons

Rock samples were crushed to mineral size under clean conditions using a jaw crusher and disc pulveriser and initial heavy mineral concentrates were made using a Wilfley table at The University of Texas at Austin. Wilfley table concentrates were sieved to remove particles larger than 70 mesh and zircon was concentrated using standard heavy liquid and magnetic separation techniques. Zircons were characterised using a binocular reflected-light microscope, transmitted light petrographic microscope (with condenser lens inserted to minimise edge refraction) and a scanning cathodoluminescence (CL) imaging system on a JEOl 730 scanning electron microscope. Single grains of zircon were strongly abraded following Krogh (1982), subsequently re-evaluated optically and then washed successively in distilled 4 N nitric acid, water and acetone. They were loaded dry into TEFLON capsules with a mixed 205Pb/235U isotopic tracer solution and dissolved with HF and HNO3. Chemical separation of U and Pb from zircon using minicolumns (0.055 ml resin volume; after Krogh, 1973) resulted in total Pb and U procedural blanks that are estimated to be 1 and 0.25 pg, respectively. Pb and U were loaded together with silica gel and phosphoric acid onto an outgassed gel and phosphoric acid onto an outgassed

The Nd model ages were calculated using the parameters of DePaolo (1981). We interpret the Nd model ages in terms of mean crustal residence ages (Arndt and Goldstein, 1987).

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