Quantifying rates of dome-and-keel formation in the Barberton granitoid-greenstone belt, South Africa

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ABSTRACT

The Barberton granitoid-greenstone belt is a classic dome-and-keel province, characterized by kilometer-scale gneiss domes and elongate keels of largely folded supracrustal rocks. Combined U–Pb SHRIMP data and structural mapping demonstrate that the geometry of the Barberton belt reflects events that occurred over ~30 million year interval, from ca. 3230 and 3203 Ma. Early deformation with NW–SE shortening in the upper crust was accompanied by emplacement of tonalite-trondhjemite-granodiorite TTG magmas at 3234 ± 12 and 3226 ± 9 Ma. Much of the structural grain of the greenstone belt relates to a long episode of post-orogenic extension, with NE-directed extension in the lower crust leading to exhumation of high-grade gneisses in the southern Barberton terrane. Advective heat transfer during emplacement of kilometer-scale (TTG) plutons around the margins of the greenstone belt facilitated the infolding of the relatively denser and colder greenstone sequence. The end of this tectonic cycle is punctuated by the emplacement of the undeformed, 3203 ± 7 Ma Dalméin pluton, which sharply truncates not only anticlines and synclines in the greenstone belt but also the dominant fabric in the high-grade gneisses.

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1. Introduction

The Archean Barberton granite-greenstone belt in the Mpumalanga Province of South Africa (Fig. 1) is a classic dome-and-keel province that has featured in many debates over the role of vertical vs. horizontal tectonics in the Archean (Visser, 1956; Anhaeusser, 1969, 1984, 2001; Anhaeusser et al., 1981; De Wit et al., 1992; De Wit, 1998; Van Kranendonk et al., 2009). The geometry of the belt has been attributed to overturn of an inverted crustal density profile, in which vertical movement of the granitoid-gneiss domes was triggered primarily by magmatic processes in the mid-to upper-crusts (Viljoen and Viljoen, 1969; Anhaeusser et al., 1981; Jackson and Robertson, 1983; Van Kranendonk et al., 2009). This emphasis on vertical motion follows from the supposition that heat flows were 2–6 times higher in the Archean compared to modern environments, due to both the retention of primordial heat and higher radioactivity of ancient crust (e.g., Lambert, 1976). Other workers maintain that the dome-and-keel geometry reflects a more conventional tectonic regime dominated by horizontal motions during a major collisional event at 3230 Ma (e.g., De Wit et al., 1992; De Ronde and De Wit, 1994; Lowe, 1994; Schoene et al., 2008, 2009). In this interpretation, the steep structural fabrics

are the result of refolding of the greenstone sequences around rigid granitoid ramparts during progressive shortening (e.g., De Wit, 1983, 1998; De Wit et al., 1992; De Ronde and De Wit, 1994; Lowe, 1994, 1999; Lowe et al., 1999; De Ronde and Kamo, 2000).

A number of recent structural studies provide evidence that the granitoid-gneiss domes in the southern part of the belt were exhumed along a low-angle detachment surface (Figs. 1 and 2), following the main orogenic phase. These studies focus on the high-*P* and low-*T* metamorphic assemblages found in association with the granitoids, which suggest exhumation from depths of 30–50 km. (Kisters et al., 2003; Diener et al., 2005; Moyen et al., 2006; Schoene et al., 2008; Lana et al., 2010). The relatively low strain intensities and low apparent geotherms (<15–20 °C/km) recorded in the high-grade rocks in this portion of the belt point to the presence of cold and rigid continental blocks that could, at least temporarily, sustain significant crustal stacking and overthickening (e.g., Diener et al., 2005; Lana et al., 2010). However, while the new kinematic and metamorphic data from the Barberton belt are consistent with a core-complex scenario, geochronological constraints to illustrate the correct sequence of events are lacking. Consequently, the processes that shaped the geometry of the granitoid-greenstone belt have not been integrated with an accurate chronology of events.

This study presents new SHRIMP ages of key structural/magmatic events that elucidate the formation of the dome-and-keel geometry in the Barberton belt. The new dataset provides

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Fig. 1. Geological map of the Barberton region showing the geometry of greenstone belts and surrounding granitoid plutons and gneiss domes. Note that the greenstone belt is subdivided into a northern and southern terrane.

important time constraints on the intrusive relationships between the gneisses and granitoids that were emplaced during the post-collisional phase. The timespan derived from our new data is consistent with contemporaneous exhumation of the high-grade gneisses and infolding of the greenstone supracrustal rocks. Integrating this new data with existing thermochronological constraints demonstrates that much of the dome-and-keel geometry is caused by sequential exhumation and related magmatism following the ca. 3230 Ma assembly of the Barberton belt.

2. Regional geology

The ancient crust within the Barberton granitoid-greenstone belt records magmatic, metamorphic and structural events ranging from 3550 to 3100 Ma (Armstrong et al., 1990; Kröner et al., 1989; Kröner et al., 1991, 1996; Kamo and Davis, 1994). Elongate keels of supracrustal rocks and broad, kilometer-scale granitoid-gneiss domes characterize much of the Barberton geology (Fig. 1) (Viljoen and Viljoen, 1969; Anhaeusser et al., 1981; Jackson and Robertson, 1983). Early studies focussed on the main lithostratigraphic groups of the low-grade metamorphic rocks in the greenstone belt (Viljoen and Viljoen, 1969; Anhaeusser, 1969), proposing a largely intact and structurally uninterrupted stratigraphy that is subdivided into (bottom to top): (1) mafic and ultramafic volcanic sequence of the Onverwacht Group, (2) sedimentary, volcanic and volcanoclastic

rocks of the Fig Tree Group, and (3) predominantly coarse-grained clastic sedimentary rocks of the Moodies Group.

The original “layer cake” model with an uninterrupted stratigraphy for the Barberton greenstone belt was subsequently challenged by geochronological studies of U–Pb in zircon (e.g., Armstrong et al., 1990; Kröner et al., 1991). It is now known that compositionally similar units and formations cannot be correlated throughout the belt (Armstrong et al., 1990; Kröner et al., 1991; Kohler and Anhaeusser, 2002; Byerly et al., 1996). For example, the metavolcanic rocks of the Onverwacht Group, exposed along the northern flank of the greenstone belt, were shown to be some 200 Ma younger than similar rocks in the southern type locality (Byerly et al., 1996). Rock units thought to be characteristic of the lowermost Theespruit Formation at the base of the greenstone belt in the south, were shown to be almost 300 Ma younger in the north (e.g., Kohler and Anhaeusser, 2002). The geochronological differences, together with abrupt facies changes recorded in both the Onverwacht Group and Fig Tree Group, have since been interpreted as indicating the presence of two distinct island arcs, namely, a northern terrane and southern terrane (Fig. 1) (De Wit et al., 1992; De Ronde and De Wit, 1994; De Ronde and Kamo, 2000).

While an early D1 deformation phase is preserved in the lower, ca. 3400 Ma stratigraphic units of the Onverwacht Group (Fig. 3), a common metamorphic history beginning at ca. 3230 Ma for both the northern and southern terranes has given rise to a collisional

Fig. 2. (a) Geological/structural map showing the southern part of the Barberton greenstone belt and surrounding gneiss complexes (Steynsdorp, Theespruit, Steynsdorp) and magmatic batholiths/plutons (e.g., Mpulizi, Dalmeir, Nelshoogte and Kaap Valley). The map also shows the contours for the bedding in the greenstone belt and for the foliation in the gneiss complexes and margins of the batholiths. The Landsat image in the background is 15-m resolution, panchromatic-band provided by the GLCF – Earth Science Data Interface. Map compiled from Anhaeusser (1984), Kisters and Anhaeusser (1995), and Westraat et al. (2005). Ages are from (#) Schoene and Bowring (2007), (*)Schoene et al. (2008), (**) Layer et al. (1992, 1998) and (***) Kamo and Davis (1994).

Fig. 3. Chart diagram illustrating the main tectonic models proposed for the evolution of the Barberton region.

interpretation for the Barberton belt. The main collisional phase (D2) was accompanied by the emplacement of TTG batholiths (e.g., Kaap Valley pluton; Fig. 2) and regional metamorphism in the Barberton region (Armstrong et al., 1990; Kröner et al., 1991; Kamo and Davis, 1994; Dziggel et al., 2002, 2005, 2006; Kisters et al., 2003; Diener et al., 2005). Most workers (e.g., Fig. 3) attribute the dominant regional strain to this event, although the details remain controversial; e.g., the exact number of distinct terranes, and the location of suture zones (see summary in Lowe and Byerly, 2007). This D2 collision also marks a significant change in the depositional style of the supracrustal rocks, from turbiditic sedimentation of the Fig Tree Group to coarse, molassic sediments that comprise the Moodies Group (Fig. 3) (Eriksson, 1979; Lamb, 1987; Heubeck and Lowe, 1994).

The evolution of the belt in the aftermath of the collision remains less clear, with much disagreement hinging on the absolute ages of various post-collisional (D3) features. In identifying the inflection point between compressive tectonics and extension, some workers favor an extremely rapid period of collision and compression (ca. 15 Ma – Kamo and Davis, 1994; or 3 Ma – De Ronde and Kamo, 2000), while others propose a >100 Ma transitional period (e.g., De Ronde and De Wit, 1994; Schoene et al., 2008) (Fig. 3). The age of the post-orogenic plutons such as the Dalmein intrusion (in the southern part of the Barberton greenstone belt; Fig. 2) is critical to resolving this question, as it clearly post-dates the main phase of compressional tectonics and was emplaced at relatively shallow, upper crustal conditions in the wake of regional metamorphism and deformation. The final episode of magmatism in the region is the intrusion of granodiorite-monzogranite-syenite (GMS) batholiths such as the Mpuluzi batholith, to the south of the Dalmein pluton, at ca. 3100 Ma (Fig. 2) (Jackson and Robertson, 1983; Westraat et al., 2005; Belcher and Kisters, 2006; Schoene et

al., 2008). This plutonism accompanied the formation of intracratonic basins in the central parts of the Kaapvaal Craton (Fig. 3) (De Ronde and De Wit, 1994; De Ronde and Kamo, 2000; Schoene et al., 2008).

The question we address in this contribution is whether the dome-and-keel architecture of the granitoid-greenstone belt (Fig. 3) was achieved during continuous, long-lived deformation, possibly diachronous throughout the belt (D2/D3, De Ronde and De Wit, 1994; D3–D4 Lowe et al., 1999) or whether the seemingly coaxial D2 and D3 events form distinct episodes during which older pre-existing structures were reactivated (e.g., Kamo and Davis, 1994; De Ronde and Kamo, 2000).

3. Field relationships

3.1. Lithologies and contact relationships

Granitoid rocks exposed between the Mpuluzi and Dalmein plutons (Fig. 4a and b) comprise tonalitic, trondhjemitic and granodioritic (TTG) gneisses that are intruded by multiple horizontal sheets and vertical dykes of potassic granitoids (Anhaeusser et al., 1981; Robb et al., 1983; Westraat et al., 2005). Outcrops immediately to the south of the Theespruit gneiss and Dalmein pluton expose remnants of a strongly foliated trondhjemitic gneiss, which has been intruded to the south by a banded granodiorite gneiss, and subsequently by potassic leucogranite sheets (Figs. 4 and 5). The trondhjemitic gneiss, hereafter referred to the Southern Theespruit Gneiss (Fig. 5a), is a medium-grained grey gneiss consisting dominantly of quartz, plagioclase, orthoclase, biotite and hornblende. The foliation (or gneissosity) is mainly defined by hornblende and biotite clots that wrap around recrystallized K-feldspar, quartz and

Fig. 4. (a) Lithological and structural map of the study area showing the dominant peak- to post-peak metamorphic fabric and mineral lineation. Note that the metamorphic fabric is folded in the Steynsdorp area and that small and large-scale folds are coaxial with the main NE-trending structural grain of the greenstone belt (Map partly modified from Robb et al., 1983). The Dalmein pluton in the middle truncates the gneissosity in the south and supracrustal rocks in the north. Sample localities for the Southern Theespruit gneiss (GG-1), Mooihoek banded gneiss (SG-1) and Dalmein (DL) are indicated. The Landsat image in the background is 15-m resolution, panchromatic-band provided by the GLCF – Earth Science Data Interface. (b) NE–SW cross-section across the study area illustrating the structural arrangement between the gneisses, supracrustals and younger intrusives.

Fig. 5. (a) Strongly foliated Southern Theespruit Gneiss. Note the solid-state gneissosity defined by hornblende and biotite minerals. (b) Layer-parallel boudinage of millimeter-wide felsic bands in the Southern Theespruit gneiss. (c) Rotated plagioclase along the fabric in the Southern Theespruit gneiss. (d) Banded Mooihoek gneiss showing alternating plagioclase-rich and biotite-rich bands. Both bands record a pervasive solid-state fabric defined by recrystallized quartz and feldspar and aligned biotite. (e) Xenoliths of the strongly foliated Southern Theespruit Gneiss in the Mooihoek gneiss. Note the presence of a segmented mafic layer in contact with Southern Theespruit Gneiss. (f) Solid-state fabric in the plagioclase-rich bands in the sheeted granitoid zone. (g) Rotated plagioclase along the fabric in the Mooihoek gneiss. (h) Complex contact/intrusive relationships between the banded (and foliated) gneisses, amphibolites (remnants of greenstones) and leucogranite sheets. There is often a color contrast between the white plagioclase-rich banding in the granodiorites and the rather pink K-feldspar-rich leucogranite sheets. Dashed white line marks the lower contact of the leucogranite sheet.

plagioclase. Locally, the gneisses show a compositional layering defined by alternating decimeter-wide grey, granodioritic bands and white felsic bands. The feldspar-rich bands record a strong NE-plunging lineation (Fig. 4a, inset i) defined by elongate aggregates of quartz and feldspar. In places, the felsic bands record layer-parallel extension and boudinage, with stretching parallel to the mineral lineation (Fig. 5b). The grain shape preferred orientation of these feldspar-quartz aggregates defines a pervasive S/L fabric.

The banded granodioritic gneiss, the Mooihoek banded granodiorite, intrudes parallel to, and locally crosscuts, the main foliation of the Southern Theespruit gneiss (Fig. 5d). In many places, xenoliths of the older Southern Theespruit gneiss are exposed (Fig. 5e) within the younger Mooihoek Gneiss intrusion. The compositional layering of the Mooihoek granodiorite is mainly defined by centimeter-scale felsic layers that alternate with bands of biotite and hornblende (Fig. 5d). The plagioclase-rich bands are characterized by strongly oriented, euhedral to subhedral K-feldspar megacrysts that reach lengths of up to 5 cm, evidence of an early magmatic fabric. This fabric has been partially overprinted, with fine- and coarse-grained granodioritic phases showing solid-state fabrics including asymmetric porphyroblasts of plagioclase (Figs. 5f and g). The geometry of these asymmetric fabrics is consistent with a normal sense of shear with greenstone material down and granitoid rocks up.

Supracrustal remnants are relatively rare and mainly occur as isolated, meter- to several decimeter-large, mostly amphibolite-facies mafic layers that are transposed along the main fabric (Fig. 5i). These supracrustal segments are intensely foliated, although primary intrusive relationships can still be discerned in the form of foliation, bedding parallel, and variably transposed trondhjemitic apophyses in the basal 10–15 m of the Theespruit Formation.

To the north, amphibolite-facies gneisses are cross cut by the Dalmein Pluton – a 5-km-wide homogenous, granodioritic pluton that also truncates the metavolcanic and sedimentary rocks of the greenstone belt. The granodiorite is characterized by por-

phyritic textures with randomly oriented K-feldspar megacrysts, zoned plagioclase, and quartz. Biotite and epidote are common in hand-specimen. Xenoliths are common throughout the pluton, with massive blocks of the greenschist facies metavolcanic and felsic schists from the Theespruit Formation evident (Fig. 6a and b), especially along the eastern and western margins of the pluton (cf. Robb et al., 1983). In the southern portions of the Dalmein pluton, these amphibolite-facies xenoliths take the form of folded and boudinaged granitic bands that alternate with the amphibolites (Fig. 6c).

3.2. Fabric development

The metamorphic foliation and compositional layering trend predominantly NE–SW, but with an E–W trend in some isolated locations (Fig. 4a). This fabric is the lateral extension of the amphibolite-facies foliation that wraps around the Theespruit, Steynsdorp and Stolzberg complexes (Kisters et al., 2003; Diener et al., 2005; Schoene et al., 2008; Lana et al., 2010) (Fig. 4a). At a larger scale, the strain fabric contains the fold axes of the kilometer-scale Kromberg antiformal syncline (De Ronde and De Wit, 1994) and Steynsdorp anticline (Kisters and Anhaeusser, 1995) (Fig. 2). Both of these regional features contain smaller, parasitic folds with wavelengths of 1–100 m. Many of these folds have tight, isoclinal geometries with fold axes that plunge to the NE at shallow to moderate angles (Fig. 5i).

A number of vertical shear zones are found, also with NE–SW trends. Several kinematic indicators; e.g., asymmetric folds, S–C fabrics and mantled plagioclase (Fig. 5c), reveal a consistent greenstone-rocks-down/granitoid-up sense of motion (Fig. 4a, inset ii) (cf. Kisters et al., 2003; Diener et al., 2005; Schoene et al., 2008; Lana et al., 2010). A NE-plunging lineation (L1) defined by aligned biotite and hornblende grains and recrystallized feldspar and quartz grains is parallel to the trend and plunge of fold axes. The plastic deformation of the quartzofeldspathic material in these

Fig. 6. (a) Wide-angle view of the eastern margin of the Dalmein pluton showing numerous, meter-scale xenoliths of the Theespruit Formation. (b) Detail of a greenschist-facies xenoliths of the Theespruit formation. (c) Detail of a strongly layered, amphibolite-facies xenoliths in the southern margin of the pluton.

Table 1
SHRIMP data for the Southern Theespruit Gneiss, Moihoeck Gneiss and Dalmein.

#	Point Analyses	comm 206	ppm U	ppm Th	232Th/238U	207/206	1 σ^a	207/235	1 σ^a	206/238	1 σ^a	Rho	206Pb/238U	1 σ^b	207Pb/206Pb	1 σ^b	Disc. ^c
<i>Southern Theespruit Gneiss GG1</i>																	
20	CL2-GG2-6	0.01	162	50	0.32	0.29	0.0006	27.30	2.6122	0.68	2.6042	1.00	3330	68	3432	3	0.03
19	GG02-CL2-103	0.29	215	38	0.18	0.26	0.0008	21.87	1.6570	0.62	1.6252	0.98	3115	40	3218	5	0.03
18	GG02-CL2-109	0.03	346	244	0.73	0.29	0.0013	25.85	1.5840	0.64	1.5210	0.96	3204	38	3424	7	0.07
17	GG1-CL2-4	0.04	383	274	0.74	0.30	0.0006	26.62	1.3912	0.64	1.3789	0.99	3197	35	3474	3	0.09
16	GG1-CL2-5	0.07	329	59	0.19	0.25	0.0006	19.02	1.5160	0.55	1.4988	0.99	2832	34	3186	4	0.13
15	GG02-CL2-103.1	1.32	358	38	0.11	0.25	0.0012	18.64	1.4790	0.55	1.3973	0.94	2809	32	3169	8	0.13
14	CL2-GG2-4	0.05	253	11	0.05	0.25	0.0007	18.81	2.8503	0.55	2.8368	1.00	2812	65	3181	4	0.13
13	CL2-GG2-2	0.07	465	341	0.76	0.30	0.0005	25.18	2.5883	0.61	2.5835	1.00	3065	63	3470	2	0.13
12	GG02-CL2-101	0.06	408	17	0.04	0.25	0.0009	17.93	1.5824	0.53	1.5389	0.97	2725	34	3167	6	0.16
11	CL2-GG2-5	0.08	1118	6	0.01	0.21	0.0016	12.37	3.2152	0.43	3.1177	0.97	2309	61	2892	13	0.25
10	GG02-CL2-105	0.19	546	66	0.13	0.25	0.0010	16.96	1.5996	0.49	1.5512	0.97	2551	33	3206	6	0.26
9	GG02-CL2-6.1	0.07	494	4	0.01	0.24	0.0005	14.86	1.4780	0.45	1.4604	0.99	2417	29	3099	4	0.28
8	GG1-CL2-2	4.72	881	794	0.93	0.22	0.0029	12.26	1.9754	0.41	1.4399	0.73	2216	27	2957	22	0.33
7	GG02-CL2-110	0.13	835	492	0.61	0.26	0.0007	15.61	1.4128	0.43	1.3870	0.98	2315	27	3259	4	0.41
6	GG02-CL2-102	0.04	353	339	0.99	0.24	0.0011	13.37	1.5757	0.40	1.5133	0.96	2164	28	3141	7	0.45
5	GG1-CL2-1	0.25	889	151	0.18	0.21	0.0004	10.09	1.5150	0.35	1.4997	0.99	1942	25	2892	3	0.49
4	GG1-CL2-3	2.54	850	116	0.14	0.22	0.0040	11.18	2.2383	0.36	1.3736	0.61	1991	24	3010	28	0.51
3	CL2-GG2-3	0.13	847	563	0.69	0.21	0.0016	8.50	2.6862	0.29	2.5839	0.96	1636	37	2932	12	0.79
2	CL2-GG2-1	4.25	919	434	0.49	0.21	0.0021	5.18	2.8405	0.18	2.6441	0.93	1076	26	2881	17	1.68
1	GG02-CL2-104	0.18	954	20	0.02	0.17	0.0007	3.33	1.4731	0.14	1.4185	0.96	856	11	2555	7	1.98
<i>Moihoeck Gneiss SG-1</i>																	
1	SG11-CL1-103	5.38	198	43	0.22	0.24	1.9907	7.01	2.6650	0.21	1.7719	0.66	1255	20	3096	32	1.47
2	CL1-SG01-4	2.62	295	109	0.38	0.25	0.7307	17.54	2.7003	0.50	2.5995	0.96	2629	56	3201	12	0.22
3	CL1-SG01-2	0.48	386	228	0.61	0.25	0.2398	18.09	2.6008	0.52	2.5897	1.00	2714	57	3188	4	0.17
4	CL1-SG01-1	0.54	138	62	0.46	0.25	0.3423	18.79	2.7789	0.54	2.7578	0.99	2763	62	3214	5	0.16
5	SG11-CL1-101	0.04	523	458	0.91	0.25	0.3645	18.60	1.4343	0.54	1.3872	0.97	2795	31	3176	6	0.14
6	SG11-CL1-104	0.73	94	110	1.20	0.25	0.6492	19.27	1.6643	0.55	1.5324	0.92	2832	35	3206	10	0.13
7	CL1-SG01-3	1.15	176	100	0.59	0.25	0.4143	19.55	2.6401	0.56	2.6074	0.99	2866	60	3206	7	0.12
8	SG1-CL1-107	0.63	253	161	0.66	0.25	0.3418	19.97	1.4537	0.57	1.4129	0.97	2917	33	3204	5	0.10
9	SG1-CL1-106	0.01	237	123	0.54	0.26	0.2571	21.50	1.4346	0.61	1.4114	0.98	3067	34	3221	4	0.05
10	SG1-CL1-105	0.03	204	106	0.54	0.26	0.2687	21.65	1.4516	0.61	1.4265	0.98	3084	35	3222	4	0.04
11	CL1-SG01-6	0.08	656	568	0.01	0.24	0.3254	20.30	2.6641	0.60	2.6442	0.99	3050	64	3142	5	0.03
<i>Dalmein Pluton DL</i>																	
1	CL2-DL1-1	2.16	764	603	0.82	0.21	0.4403	6.63	2.7127	0.24	2.6028	0.96	1405	33	2805	13	1.00
2	DL2-24B	0.17	694	212	0.32	0.21	0.3478	8.49	2.0364	0.29	2.0065	0.99	1630	29	2937	6	0.80
3	DL2-31B	0.21	713	549	0.80	0.22	0.5507	10.41	2.3821	0.34	2.3176	0.97	1907	38	2975	9	0.56
4	DL1-CL2-104.1	1.59	661	275	0.43	0.21	0.5446	10.88	1.5127	0.37	1.4113	0.93	2028	25	2933	9	0.45
5	DL2-15B	0.06	542	159	0.30	0.23	0.2776	13.35	1.9227	0.42	1.9025	0.99	2239	36	3074	4	0.37
6	DL2-33B	0.18	527	270	0.53	0.24	0.2702	19.73	2.7634	0.60	2.7502	1.00	3012	66	3122	4	0.37
7	DL2-22B	0.05	587	258	0.45	0.24	0.2592	14.10	1.8027	0.43	1.7840	0.99	2328	35	3087	4	0.33
8	CL2-DL1-8	0.12	613	260	0.44	0.22	0.1602	13.25	2.6814	0.43	2.6766	1.00	2304	52	3007	3	0.31
9	DL1-CL2-6.1	0.64	398	111	0.29	0.23	0.3606	13.82	1.4302	0.44	1.3840	0.97	2350	27	3037	6	0.29
10	DL1-CL2-103	0.32	574	296	0.53	0.24	0.2530	15.06	1.5639	0.46	1.5433	0.99	2434	31	3108	4	0.28
11	DL2B-8	0.22	271	181	0.69	0.24	0.4094	16.69	2.0389	0.50	1.9974	0.98	2604	43	3141	7	0.21
12	DL1-CL2-105	0.40	483	188	0.40	0.24	0.3232	16.94	1.4220	0.51	1.3848	0.97	2647	30	3133	5	0.18
13	CL2-DL1-5	0.12	438	169	0.40	0.24	0.1620	17.74	2.6166	0.53	2.6116	1.00	2745	58	3136	3	0.14
14	DL2-23B	0.06	298	317	1.10	0.25	0.3260	18.85	2.5185	0.55	2.4973	0.99	2831	57	3172	5	0.12
15	DL1-CL2-108	0.15	447	215	0.50	0.24	0.2319	18.22	1.4074	0.55	1.3882	0.99	2818	32	3127	4	0.11
16	DL2B-7	0.17	203	144	0.73	0.25	0.5082	18.93	2.4023	0.56	2.3480	0.98	2862	54	3157	8	0.10
17	DL2-13B	0.01	383	568	1.53	0.25	0.2774	19.63	1.8095	0.57	1.7881	0.99	2900	42	3188	4	0.10
18	DL2-5.1B	0.06	463	278	0.62	0.25	0.3171	19.44	1.8276	0.57	1.7998	0.98	2908	42	3168	5	0.09
19	DL2-35.1B	0.05	443	143	0.33	0.25	0.2749	19.69	1.8557	0.57	1.8352	0.99	2928	43	3175	4	0.08

Table 1 (Continued)

#	Point Analyses	comm 206	ppm U	ppm Th	232Th/238U	207/206	1 σ^a	207/235	1 σ^a	206/238	1 σ^a	Rho	206Pb/238U	1 σ^b	207Pb/206Pb	1 σ^b	Disc. ^c
20	DL2-16B	0.06	422	371	0.91	0.25	0.3279	20.07	2.4394	0.58	2.4173	0.99	2960	57	3183	5	0.08
21	DL2B-10	0.12	345	288	0.86	0.25	0.3002	20.70	1.9893	0.60	1.9665	0.99	3016	47	3195	5	0.06
22	CL2-DL1-4	0.17	270	143	0.55	0.25	0.2667	20.48	2.6087	0.59	2.5950	0.99	3009	62	3183	4	0.06
23	DL1-CL2-104	0.12	652	271	0.43	0.25	0.1894	20.03	1.4392	0.59	1.4267	0.99	2990	34	3161	3	0.06
24	CL2-DL1-10	0.09	450	193	0.44	0.24	0.4674	19.95	2.6764	0.59	2.6353	0.98	2996	63	3150	7	0.05
25	DL2-29.1B	0.03	360	109	0.31	0.25	0.5449	21.16	1.9082	0.61	1.8287	0.96	3051	44	3207	9	0.05
26	DL2B-12	0.07	229	166	0.75	0.25	0.3460	21.08	2.0300	0.61	2.0003	0.99	3066	49	3191	5	0.04
27	DL2B-6	0.13	264	428	1.67	0.25	0.3589	20.82	2.0303	0.61	1.9984	0.98	3066	49	3172	6	0.03
28	DL2-29B	0.04	354	462	1.35	0.25	0.5429	21.67	1.9165	0.62	1.8380	0.96	3112	45	3206	9	0.03
29	DL2B-11	0.08	230	369	1.66	0.25	0.3567	21.92	2.0234	0.63	1.9917	0.98	3142	50	3204	6	0.02
30	DL2-35B	0.00	162	191	1.22	0.26	0.4149	22.47	2.4026	0.64	2.3665	0.98	3174	59	3223	7	0.02
31	CL2-DL1-6	0.02	406	122	0.31	0.25	0.1432	22.38	2.5945	0.64	2.5906	1.00	3189	65	3208	2	0.01
32	DL2B-5	0.05	431	249	0.60	0.25	0.2757	22.31	2.0025	0.65	1.9834	0.99	3211	50	3188	4	0.01

^a Percent.^b Abstract.^c Conc. = ((207Pb/206Pb date)/(Pb206/Pb238 date)) – 1.

lineated aggregates indicates strain under peak amphibolite-facies conditions. Highly prolate aspect ratios (Fig. 5b) are consistent with stretching mineral lineation developed under constrictional strain. Significantly, this lineation is unidirectional, and has not been affected by subsequent folding events.

In summary, the main fabric forming event in the southern Barberton terrane occurred during exhumation of the high-grade gneisses domes (Steynsdorp, Stolzburg and Theespruit; Figs. 1 and 2a) and emplacement of numerous granodiorite-granite sheets. The intrusion of the Dalmein pluton occurred at a relatively shallow level, in the waning stages of exhumation. Later emplacement of leucogranites such as the Mpuluzi batholith reactivated pre-existing structures under low-grade conditions (e.g., Westraat et al., 2005; Belcher and Kisters, 2006; Schoene et al., 2008). In the following section, we provide new U–Pb age data obtained from samples of the Southern Theespruit (GG-1) and Mooihoek gneisses (SG-1), and the Dalmein pluton (DL).

4. Geochronology

4.1. Analytical techniques

The U–Pb analyses were conducted using the WA Consortium SHRIMP II ion microprobe housed at Curtin University of Technology. Analytical procedures are described by Nelson (1997) and Williams (1998). Detailed cathodoluminescence images are shown in the online electronic Appendix (ES-1). Isotopic ratios were monitored with reference to Sri Lankan gem zircon standard BR266, with an age of 559.1 Ma, and Pb/U ratios in the samples were corrected using the $\ln(\text{Pb}/\text{U})/\ln(\text{UO}/\text{U})$ relationship as measured on zircons standard. All uncertainties in the calculated intercept or Concordia ages are reported at 95% confidence limits, unless stated otherwise. The results of the spot analyses are presented in Table 1.

4.2. Southern Theespruit gneiss (GG-1)

The granodioritic gneiss was collected 1.5 km south of the contact with the Theespruit pluton (26°06'12"S–30°51'07"E) (Fig. 4a). The sample is marked by a strong solid-state foliation (Fig. 5a) defined by peak metamorphic biotite and hornblende and recrystallized plagioclase, K-feldspar, and quartz. Accessory minerals include titanite, apatite, magnetite and zircon. Zircons extracted from this sample vary from translucent to milky, with pink and brown colors. Most of them are clearly zoned, some of which contain cores and/or metamorphic rims (ES-1). Other grains were homogeneous and particularly bright under CL. We targeted the centers of both zoned and homogeneous grains.

All the ion microprobe data from this sample are plotted on the concordia diagram (Fig. 7a). Two distinct zircon populations are present based on age grouping, Th/U ratios, and, to some extent, zircon morphology. Four analyses from the cores of relatively large euhedral zircons, with well-defined oscillatory zoning, yielded very old $^{207}\text{Pb}/^{206}\text{Pb}$ ages (>3400 Ma). This population is low in common Pb and has Th/U ratios between 0.3 and 0.8 (Table 1). The two oldest grains yielded an $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean age of 3471 ± 3 Ma, whereas one grain (#13) gave a concordant age of 3432 ± 10 Ma (2σ), similar, within error, to the age of the Theespruit Pluton (3437 ± 6 Ma) determined by Armstrong et al. (1990) from a sample taken a few hundred meters north of the GG-1 locality (Fig. 4a). The rims of these grains yield younger $^{207}\text{Pb}/^{206}\text{Pb}$ ages and lower Th/U ratios (<0.1) than the cores (Fig. 7a; Table 1) suggestive of metamorphic overgrowths, but discordance makes age determination difficult.

Zircons in the second population are characterized by bright, oscillatory zoning, probably of magmatic origin, and much lower Th/U ratios (0.2–0.01) than the first population. The seven most

Fig. 7. U–Pb Concordia diagrams for the zircons populations in the (a) Southern Theespruit gneiss and (b) Mooihoek banded gneiss. Errors are at the 2σ level. (c) U–Pb Concordia diagrams for the zircons populations in the Dalmein sample (DL). Errors are at the 2σ level.

concordant analyses from this population form a linear array that intercepts the concordia curve at 3234 ± 12 Ma (MSWD = 0.61; 2σ) (Fig. 7a), an age that is identical within error to the crystallization age of the 3236 ± 1 Ma Nelsoghtte pluton to the north of the Barberton greenstone belt (De Ronde and Kamo, 2000; Schoene et al., 2008), suggesting that the southern Theespruit gneiss was emplaced at the same time as the Nelsoghtte pluton.

4.3. Mooihoek gneiss (SG-1)

The sample of the Mooihoek granodiorite (SG-1) was collected ca. 50 m from the contact with the Dalmein pluton ($26^{\circ}07'01''\text{S}$ – $30^{\circ}52'31''\text{E}$) (Fig. 4a). Although mineralogically similar to sample GG-1 of the Theespruit gneiss, this sample is distinguished by alternating bands of biotite- and plagioclase. Zircons from this sample are of magmatic origin, with dominantly euhedral to subhedral shapes, oscillatory zoning (ES-1), and Th/U ratios between 0.3 and 1.0 (Table 1). Many grains also preserve core structures that are particularly bright relative to the surrounding rims (ES-1). However, the cores were not discernibly older or chemically distinct from the rims.

Seven analyses from grain cores (points 2, 4, 5, 7, 8, 9, 10, 11) define a linear array with an upper intercept at 3226 ± 9 Ma (MSWD = 1.07) (Fig. 7b). One bright, prismatic, homogeneous zircon forms an outlier to this main population, with younger $^{207}\text{Pb}/^{206}\text{Pb}$ ages and enrichment in Th and U (Table 1). The younger age recorded by this grain, together with extremely low Th/U ratios (<0.01) suggests that it grew during a subsequent metamorphic event, although the age is poorly constrained. We consider that 3226 ± 9 Ma is the crystallization age of the Mooihoek gneiss, in good agreement with ages of other trondhjemites emplaced around the margins of the greenstone belt (e.g., 3227 ± 1 Ma Kaap Valley; Kamo and Davis, 1994) as well as with the porphyritic rocks of the Usutu suite exposed in the eastern margin of the greenstone Belt (Schoene and Bowring, 2010). Notably, greenstone remnants around the Stolzberg pluton yielded a number of metamorphic zircons (with Th/U <0.2), which gave a similar age of 3227 ± 7 Ma for metamorphism (Dziggel et al., 2005), suggesting that the emplacement of the Mooihoek gneiss took place during a time regional deformation and amphibolite-facies metamorphism.

4.4. The Dalmein pluton (DL)

Sample DL is from a K-feldspar megacrystic quartz-monzonite from the central part of the pluton ($26^{\circ}06'15''\text{S}$ – $30^{\circ}55'05''\text{E}$) (Fig. 4a). The zircons vary in brightness and shape and contain cores and darker rims (ES-1). Both cores and rims group have the same age and compositions, although the cores have suffered less Pb loss. Thirty-two point analyses from the different zircon grains vary from concordant to strongly discordant. They plot around a line that intersects upper Concordia at 3199 ± 12 Ma (2σ ; MSWD = 8.1). Although several attempts to eliminate scatter around the Pb loss chord has been made, all points were considered to be representative of one $^{207}\text{Pb}/^{206}\text{Pb}$ age group – the large uncertainty to this age relates to six points (4, 6, 9, 8, 15, 24) that plot slightly above and below the regression line. A similar, albeit more precise, upper intercept age of 3203 ± 7 Ma (MSWD = 3.2) is indicated by 26 points that plot directly on the discordia line (Fig. 7c). This age is consistent with a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean of 3201 ± 12 Ma defined by the seven most concordant points (discordance <0.05) and is here interpreted as the best estimate for the crystallization age for the Dalmein pluton.

5. Discussion

The Southern Theespruit and Mooihoek gneisses (Fig. 4a) record a structural and magmatic evolution akin to most gneiss domes in the southern Barberton terrane (e.g., Kisters and Anhaeusser, 1995; Kisters et al., 2003; Diener et al., 2005; Lana et al., 2010). The Southern Theespruit gneiss crystallized at 3234 ± 12 Ma, contemporaneously with the 3236 ± 1 Ma (Schoene et al., 2008) Nelsoghtte TTG pluton in the northern terrane. This was followed shortly by a second TTG magmatic pulse that led to the emplacement of various

TTG bodies including the 3226 ± 9 Ma Mooihoek gneiss (Fig. 7b), the 3229.8 ± 0.6 Ma (Schoene et al., 2008) Vlakplaats granodiorite in the southern terrane, and the regionally extensive 3227 ± 1 Ma (Kamo and Davis, 1994) Kaap Valley pluton in the northern terrane.

The main fabric forming event in the high-grade gneisses of the southern Barberton terrane occurred at peak metamorphic conditions and was contemporaneous with the emplacement of the numerous granodiorite-granite sheets, which includes not only the 3226 ± 9 Ma Mooihoek gneiss to the south of the Dalmein (Fig. 5d and e), but also 3229 ± 5 Ma and 3212 ± 1 Ma syn kinematic dikes in the central Stolzberg complex (Dziggel et al., 2005; Schoene et al., 2008) and 3223 ± 2 Ma granodiorite dikes and syn kinematic granodiorites – eastern margin of the greenstone belt (Schoene et al., 2008; Schoene and Bowring, 2010). A mid-crustal depth of exhumation for the Mooihoek and Southern Theespruit bodies is indicated by the solid-state deformation and amphibolite-facies metamorphism, typified by the Stolzberg complex (Dziggel et al., 2005; Schoene et al., 2008), Theespruit complex (Diener et al., 2005) and Steynsdorp complex (Schoene et al., 2008; Lana et al., 2010).

By the time of intrusion of the Dalmein pluton, tectonomagmatic activity in the southern parts of the Barberton belt had largely ceased (Fig. 4). The Dalmein pluton was emplaced at higher crustal levels, since the pluton itself is undeformed and sharply truncates the low-grade structures in the greenstone belt (e.g., Kromberg structure). This field observation is at odds with the previously reported age of 3216 ± 2 Ma for the Dalmein pluton, based on analysis of two fractions of titanite and one discordant analysis of zircon (Kamo and Davis, 1994), taken from a sample collected near the contact with the Theespruit Formation. However, four lines of evidence suggest that the 3216 ± 2 Ma age for the Dalmein is artificially old, and reflects the presence of xenocrystic, inherited material. First, the contact between the younger Dalmein intrusion sharply crosscuts the Mooihoek Gneiss and establishes the possibility of inheritance. Second, the Dalmein pluton is characterized by millimeter- to meter-sized xenoliths of country rock (Fig. 6a–c), consistent with the presence of xenocrysts. Third, a number of rocks surrounding the pluton contain zircons and titanites ranging in age from 3230 to 3200 Ma, including the Doornhoek pluton (Kamo and Davis, 1994), the Theespruit Formation amphibolites (Kisters et al., 2003; Diener et al., 2005), the Stolzberg pluton (Kamo and Davis, 1994; Schoene et al., 2008), and the Mooihoek and Southern Theespruit gneisses (this study). Finally, though the use of titanite as a thermochronometer in metamorphic rocks is well-established, there are numerous examples of inherited titanite in granitic magmas (Zhang and Schärer, 1996; Pidgeon et al., 1996; Aleinikoff et al., 1995; Corfu, 1996; Corfu and Stone, 1998; Aleinikoff et al., 2002; Jung and Hellebrand, 2007). The blocking temperature of titanite for

slowly-cooled metamorphic provinces is considered to be between 600°C (Mezger et al., 1991) and 700°C (Scott and St-Onge, 1995; Bingen and van Breemen, 1998), but the actual blocking temperature for individual grains will reflect the cooling rate (Dodson, 1973), as well as the grain size of the analysed material (Cherniak, 1993; Frost et al., 2000). Given the fast-cooling expected for the high crustal level emplacement of the Dalmein pluton, and that the >0.180 mg titanite fractions analysed by Kamo and Davis (1994) were fragments of much larger grains, we consider that titanite in Dalmein pluton has a high probability of preserving inherited ages.

We note that the older age assignment underpins recent tectonic models that emphasize an exceedingly fast transition (15 Ma) from collisional to extensional tectonics (Kisters et al., 2003; Diener et al., 2005). The younger age we report here suggests a more moderate pace of tectonic events with orogenic collapse now constrained to have occurred ca. 30 Ma after crustal thickening. This lengthened timescale demonstrates that orogenic processes operative during the Phanerozoic (e.g., Caby et al., 2001; Whitney et al., 2003; Bonamici and Duebendorfer, 2009) is obtained also for early-middle Archean times.

5.1. Timing constraints on the dome-and-keel structure

Structural studies in the Barberton granite-greenstone belt (Williams and Furnell, 1979; De Wit, 1983; Jackson and Robertson, 1983; De Ronde and De Wit, 1994; Lowe et al., 1999; Westraat et al., 2005; Belcher and Kisters, 2006) emphasize the role of NW–SE-directed crustal shortening that followed the D2 collisional event at 3230 Ma (e.g., Fig. 3). Shortening affected all stratigraphic units in the greenstone belt, from the upper Onverwacht volcanic sequences to the Moodies Group sediments, refolding the strata along NE–SW trending folds (Fig. 2). Within the same time frame, the Barberton belt witnessed the exhumation of high-grade rocks in the southern granitoid-gneiss terrane (e.g., the Theespruit Formation and the underlying Steynsdorp, Theespruit and Stolzberg domes; Fig. 1) from crustal depths of 25 and 35 km (Fig. 8a and b). The domes themselves have been shown to be part of a high-grade metamorphic core-complex zone (Kisters et al., 2003; Diener et al., 2005; Lana et al., 2010), which also include the Southern Theespruit Gneiss.

Exhumation of these rocks generated structural elements that are co-planar/co-linear with the NE-trending structural grain in the main greenstone belt (Fig. 2). For instance, strain in the Steynsdorp area (Figs. 2 and 4a) caused folding of the granitoid-greenstone contact into a NE-plunging conical fold, which is co-planar with the Steynsdorp and Kromberg fold structures in the main greenstone belt (e.g., Kisters and Anhaeusser, 1995; Lana et al., 2010). Small-scale folds in the Theespruit Formation (Kisters et al., 2003) are

Fig. 8. Extensional detachment and formation of core-complexes in the Barberton region. (a) Cartoon illustrating the structural uplift of the Steynsdorp, Theespruit, and Stolzberg domes underneath a shallowly NE dipping detachment. Much of the strain related to the exhumation was accommodated by the Theespruit Formation at the base of the greenstone belt. (b) *P*–*T* diagram summarising peak metamorphic pressures recorded in different areas of the southern terrane of the Barberton belt; (1) Northern margin of the Stolzberg pluton (Kisters et al., 2003); (2) Northern margin of the Theespruit pluton (Diener et al., 2005); (3) Northern margin of the Steynsdorp pluton (Lana et al., 2010); (4) Southern margin of the Stolzberg pluton (Dziggel et al., 2002); (5) Onverwacht rocks to the north of the Theespruit Pluton (Cloete, 1991).

coaxial with NE-plunging folds in the Moodies Group and Fig Tree Group (Jackson and Robertson, 1983), and correspond to the unidirectional NE plunge of the L1 lineation in the high-grade gneisses (Fig. 2). Similarly, the isoclinal symmetrical folds, NE-plunging (L1) mineral lineation, and layer-parallel boudins observed in the Southern Theespruit and Mooihoek gneisses (Fig. 4a) display the same geometry to those folds and boudins in the Steynsdorp and Stolzberg plutons (Fig. 2) (e.g., Diener et al., 2005; Lana et al., 2010).

These regional constraints and our present structural/geochronological data suggest that the Southern Theespruit and Mooihoek gneisses were part of the younger TTG rocks emplaced between 3236 and 3227 Ma, and subsequently exhumed relative to the main greenstone belt as a consequence of late-D2 orogenic collapse. Peak metamorphism at ca. 3230 Ma (Dziggel et al., 2005) coincided with the onset of exhumation of the high-grade gneisses. Thermochronological data available for the kilometer-scale TTG plutons (see Schoene et al., 2008) are consistent with a long period of extensional unroofing and slow cooling of the gneiss domes around the greenstone belt in the wake of the ca. 3230 Ma metamorphism. This is particularly true for several granitoid plutons in the northern and southern terranes (e.g., Vlakplaats intrusive, Nelsooghte and Kaap Valley plutons, Fig. 2) that record hornblende and apatite ages consistent with cooling below 400 °C between 3225 and 3180 Ma (Layer et al., 1992, 1998; Schoene et al., 2008).

Much of the crust should have been exhumed before the emplacement of the Dalmein pluton at ca 3203 Ma. The Dalmein is a post-tectonic pluton that crosscuts not only the Kromberg structure, but also the limbs of the Steynsdorp and Onverwacht anticlines (Fig. 2) (Anhaeusser et al., 1981; Robb et al., 1983; De Wit, 1983). It therefore post-dates all the kilometer-scale folds in the greenstone belt, including the deformed Moodies Group, which is folded near the Steynsdorp dome and in the Eureka syncline (e.g., Jackson and Robertson, 1983; Lowe et al., 1999; Lowe and Byerly, 1999). Further south, the Dalmein truncates the NE-plunging tight folds in the Southern Theespruit and Mooihoek gneisses. It is thus reasonable that the main phase of kilometer-scale folding in the southern parts of the Barberton granite-greenstone belt, and the extension + cooling of the Meso-Archean crust, should be bracketed between the peak metamorphism at 3230 Ma and the emplacement of the Dalmein Pluton at 3203 ± 7 Ma. The relatively low metamorphic pressures recorded in the greenstone belt to the north of the Dalmein contact zone (Fig. 8b) (260 MPa, Cloete, 1991) indicate emplacement at relatively shallow levels (<11 km). Thus, the high-grade southern Barberton gneisses, which record low geothermal gradients (~ 20 °C/km; Fig. 8b), must have been cooler than 300 °C at ca. 3203 Ma, the time of the Dalmein intrusion. The Ar/Ar hornblende and apatite age data available for the Kaap Valley and Nelshoogte and Steynsdorp domes (Layer et al., 1992, 1998; Schoene et al., 2008), and the structural and metamorphic features in the high-grade gneisses do not support a regional high temperature (>500 °C) thermal overprint during the D3 event. Much of the deformation and thermal overprint related to D3 was largely confined to the margins of the kilometer-scale potassic batholiths such as the Heereveen, Mpuluzi and Pigg's peak Batholiths in the south (Belcher and Kisters, 2006; Schoene et al., 2008).

We thus propose that much of the present-day dome-and-keel structure of the Barberton greenstone belt resulted from a ca. 30 million year period of NE–SW extension of relatively hot (>500 °C) crust mid- to lower-crustal gneisses and concomitant TTG plutonism. The orthogonal strain partitioning that we infer, i.e., NE-extension in the lower to mid-crust and NW–SE compression in the upper crustal rocks, lends support to the crustal-scale chocolate-tablet boudinage model proposed for other Archean provinces. These models link extension/exhumation in the lower crust with contemporaneous shortening in the upper crust, commonly envi-

sioned as infolding of supracrustal rocks between rising domes of hot, mid-crustal basement (Marshak et al., 1992, 1997; Alkmin and Marshak, 1998; Marshak, 1999; Tohver et al., 2006, 2007). The thermal component necessary to weaken the lower- and mid-crust for this “regional boudinage” may be directly related to the high-grade metamorphism and voluminous TTG magmatism, since both processes are spatially and temporally linked to the D2 event (e.g., Kisters et al., 2003).

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.precamres.2009.12.001.

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