U-Pb baddeleyite dating of intrusions in the south-easternmost Kaapvaal Craton (South Africa): revealing multiple events of dyke emplacement

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Dissertations in Geology at Lund University,
Master's thesis, no 457
(45 hp/ECTS credits)

Department of Geology
Lund University
2015

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Cover Picture: Overview of the 10 m wide ENE– trending dyke ENEDo8. Photo: Emilie Larsson.

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Larsson, E., 2015: U-Pb baddeleyite dating of intrusions in th south-easternmost Kaapvaal Craton (South Africa): revealing multiple events of dyke emplacement. *Dissertations in Geology at Lund University*, No. 457, 30 pp. 45 hp (45 ECTS credits).

Abstract: This study presents U-Pb baddeleyite ages for five dolerite dykes and one sill in the Mesoarchean granitic basement of south-easternmost Kaapvaal Craton in northern KwaZulu-Natal, South Africa. The oldest dyke is ENE-trending, and is dated to 2729 ± 5 Ma. One ENE- and two SE-trending dyke are dated to 2580 ± 3 Ma, 2574 ± 5 Ma and >2517 ± 5 Ma (minimum age for the latter given by the $^{207}Pb/^{206}Pb$ date for the "oldest" fraction), respectively. These are interpreted to belong to a common ca. 2577 Ma event of dyke emplacement. A dolerite sill was dated to 2423 ± 7 Ma and an ENE-trending dyke to 2168 ± 5 Ma. The ENE-trend is shared among all dated generations, and their trends are interpreted to largely be controlled by basement structures pre-dating the younger generations of dykes. The 2729 Ma dyke can be correlated to the 2724 Ma age of the Phokwane Formation and the 2733 Ma age of the Mohle Formation (Hartswater Group, Ventersdorp Supergroup), extending Ventersdorp volcanism into the southeastern part of the Kaapvaal Craton. The ca. 2577 Ma generation represents a new magmatic event on the Kaapvaal Craton. It bears a temporal similarity to the ca. 2575 Ma Great Dyke on the Zimbabwe Craton. However, different trends and a distance of over 800 km between these units argues against a Kaapvaal-Zimbabwe connection at 2575 Ma in their present-day positions. The ca. 2423 and 2168 Ma intrusions also represent new magmatic events in this part of the Kaapvaal Craton, although similar ages have been reported for the Westerberg Sill and Hekpoort volcanics. Based on these new temporal constraints, the data in this study supports a juxtaposition of Kaapvaal and Pilbara in Vaalbara, at 2729 Ma. The new 2168 Ma age in this study can be temporally linked to magmatic events in the Superior and Wyoming cratons, and is supportive of the Kaapvaal, Superior and Wyoming cratons being in close proximity at these times.

Keywords: U-Pb, baddeleyite, Kaapvaal Craton, mafic dykes, geochronology

Supervisor(s): Ulf Söderlund, Ashley Gumsley & Andreas Petersson

Subject: Bedrock Geology

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U-Pb baddeleyit datering av intrusioner i sydöstligaste Kaapvaal Kratonen

EMILIE LARSSON

Larsson, E., 2015: U-Pb baddeleyite dating of intrusions in the south-easternmost Kaapvaal Craton (South Africa): revealing multiple events dyke emplacement. *Examensarbeten i geologi vid Lunds universitet*, Nr. 457, 30 sid. 45 hp.

Sammanfattning: I den här studien presenteras U-Pb baddeleyit åldrar för fem diabasgångar och en sub-horisontell intrusion från KwaZulu-Natal provinsen i den sydöstra delen av Kaapvaal kratonen. Den äldsta diabasgånen är ONO- strykande och är daterad till 2729 ± 5 Ma. Två diabasgångar, en ONO-strykande och två med SO-strykande är daterade till 2580 ± 3 Ma, 2574 ± 5 Ma och >2517 ± 5 Ma (minimi ålder framtagen av ^{207}Pb/^{206}Pb den äldsta fraktionen). Dessa är tolkade till att tillhöra samma magmatiska event vid ca. 2577 Ma. Dessutom daterades en sub-horisontell intrusion till 2423 ± 7 Ma och en ONO-strykande diabasgång till 2168 ± 5 Ma. Då ONO-riktningen är representerad i alla generationerna tolkas denna strykning som en redan existerande strykning, åtminstone bland de yngre generationerna, och representerar eventuellt en svaghets-riktning i berggrunden. 2729 Ma generationen kan korreleras med den 2724 Ma gamla Phokwane Formationen och den 2733 Ma gamla Mohle Formationen (Hartswater Group, Ventersdorp Supergroup), vilket därmed utökar den kända utbredningen av Ventersdorp magmatismen sydöst om tidigare kända exponeringar. Den ca. 2577 Ma gamla generationen representerar ett nytt, tidigare okänt magmatiskt event på Kaapvaal kratonen. Denna ålder är jämnårig med den ca. 2580 Ma gamla Great Dyke på Zimbabwe Kratonen. Baserat olika trender och den stora distansen på över 800 km mellan Great Dyke och 2577 Ma generationen i denna studie anses en Kaapvaal-Zimbabwe kontakt vid 2575 Ma osannorlik med dagens placering av blocken. Intrusionerna med åldrarna 2423 Ma och 2168 Ma representerar också tidigare okänd magmatism på Kaapvaal-kratonen, även om liknande åldrar finns rapporterade från Westerberg Sill och Hekpoort volcanics. Baserat på dessa tidsramar stödjer resultaten i denna studie en länk mellan Kaapvaal och Pilbara (Vaalbara) vid 2729 Ma. Dessutom är 2168 Ma åldern likåldrig med magmatism på Superior- och Wyoming-kratonen, vilket innebär att dessa kan ha varit lokaliserade i närheten av varandra vid denna tidpunkt.

Nyckelord: U-Pb, baddeleyit, Kaapvaal, diabasgångar, geokronologi

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1 Introduction

The arrangement of cratonic blocks throughout the geological history has long been a challenging research topic (Bleeker 2003). There are several known reconstructions, where all or some of the crustal blocks have amalgamated into supercontinent configurations. The more recent assemblies are better known, whereas further back in geological history paleoreconstructions become highly uncertain. Postulated supercontinents include the Archaean Kenorland (e.g. Williams et al. 1991), Columbia/Nuna at 1.8–1.5 Ga, Rodina at 1.1–0.6 Ga and Pangea at 0.3–0.2 Ga, (e.g. Li et al. 2008; Bradley 2011; Zhang et al. 2012). Today, there are about 35 large cratonic fragments of Archaean age, dispersed around the globe. Kaapvaal is one of these cratonic fragments and dates back to between 3.6 and 3.1 Ga, making it among the oldest known cratons (Bleeker 2003). The presence of rifted and collisional margins near its borders, indicate that the Kaapvaal craton previously been part of a larger ancestral landmass. Even though pieces of the puzzle are missing, the most accepted hypothesis suggests juxtaposition to the Pilbara Craton of NW Australia. Kaapvaal and Pilbara are suggested to have formed the Vaalbara supercraton, which formation dates back to at least 3.1 Ga with subsequent breakup prior to 2.1 Ga (Cheney, 1996; Zegers et al. 1998); alternatively, Kapvaal and Pilbara were blocks of the Kenorland supercontinent. Vaalbara, together with Superia and Slave cratons are three of the better known crustal blocks (supercratons) that may have existed contemporaneously during the Archaean and prior to modern-style plate tectonics started to operate. This hypothesis is based on their stratigraphic and structural histories (e.g. Bleeker 2003).

Coeval short-lived mafic magmatic events on crustal blocks can indicate that they were part of an ancestral, coherent landmass, which has since rifted apart (Bleeker 2003; Bleeker & Ernst, 2006). The plumbing systems to these ancient magmatic events are, due to erosion of extrusive rocks, represented by dyke swarms and sill provinces. Crystallization ages of these magmatic events can be arranged in a magmatic "barcode" which subsequently can be compared with corresponding barcodes of other cratons. One or several matches of coeval events indicate that cratons were 'nearest neighbors' (Bleeker 2004). Once matches have been found, additional geological data such as paleomagnetic signatures and geochemistry can be used to confirm or dismiss a next-neighbor relation (Bleeker & Ernst 2006; Davies & Heaman 2014).

The Archean basement in the eastern part of the Kaapvaal Craton exposes a large number of dyke swarms with various trends and cross-cutting relations, indicating multiple tensional events associated with mafic magmatism. Recent U-Pb geochronology of dykes in the eastern part of the Kaapvaal Craton reveal the occurrence of three regional swarms of Archaean dykes at 2.70–2.66 Ga, all which radiate out from a centre in the Bushveld Complex. Apart from the 2.70–2.66 Ga dykes, the NE-trending dyke swarm is intermixed with 1.88–1.84 Ga dykes and the SE-trending dyke swarm is intermixed with 2.97 Ga dykes (Olsson et al. 2010; 2011; Olsson 2012). Further south, Rådman (2014) identified 2.66 Ga NE-trending dykes, identical in age to the dykes further north, possibly indicating a more extensive event. In the same area, a SE-trending dyke was dated to 2.87 Ga which is coeval and possibly related to the Hlagothi Complex (Gumsley et al. 2013).

In this study, age determinations of dykes with an ENE- and SE-trend in the KwaZulu-Natal province (SE Kaapvaal), originally assumed to represent two separate generations based on the two distinct trends and cross-cutting relationships, have been attempted using U-Pb baddelyite (ZrO_2) geochronology. Baddeleyite is an accessory mineral that is widely used for dating the crystallization of mafic and other silica-undersaturated rocks (Heaman & LeCheminant 1993). It has a high U content and low, or negligible Pb content, hence ideal for obtaining precise U-Pb ages. Baddeleyite is commonly formed in magmatic systems, as it readily transforms into zircon under metamorphism (Heaman & LeCheminant 1993; Söderlund et al. 2013). The results of this study will provide information to the barcode and assist to better constrain robust paleoreconstructions for the Kaapvaal Craton.

2 Regional Geology

The Kaapvaal Craton (Fig. 1) is bordered by the Proterozoic Namaqua-Natal Mobile Belt in the west and south. The Lebombo monocline forms the boundary in the east with the Mozambique Belt and the Zimbabwean Craton to the north, separated by the Limpopo Belt (e.g., de Wit et al. 1992). The craton was formed between 3.6 and 3.1 Ga, which makes it among the oldest pieces of known crust in the world, and also one of the most well-preserved (Bleeker 2003; Hunter et al. 2006). The Kaapvaal Craton is made up of the ~ 3.65 Ga Ancient Gneiss Complex (AGC) and the Barberton Greenstone Belt nucleus, upon which younger granitoid suites and greenstone belts were consolidated (de Wit et al., 1992; Eglington & Armstrong, 2004).

By 3.7–3.1 Ga the Kaapvaal Craton had stabilized, and sedimentary basins began to form on the craton (de Wit et al. 1992; Brandl et al. 2006). The first three successions were the Dominion Group, the Witwatersrand Supergroup (and the Pongola Supergroup equivalent in the east) and the Vendersdorp Supergroup. The Dominion Group, which is the oldest sequence consists of 3074 ± 6 Ma (U-Pb zircon) old volcanic rocks (Armstrong et al. 1991). The Witwatersrand Supergroup largely consists of clastic sediments and was formed during at least two sedimentary cycles (McCarthy 2006). The lower and more extensive West Rand Group consists of fine quartzites, conglomerates, shales and banded iron formations, whereas the Central Rand Group consists of coarser quartz-

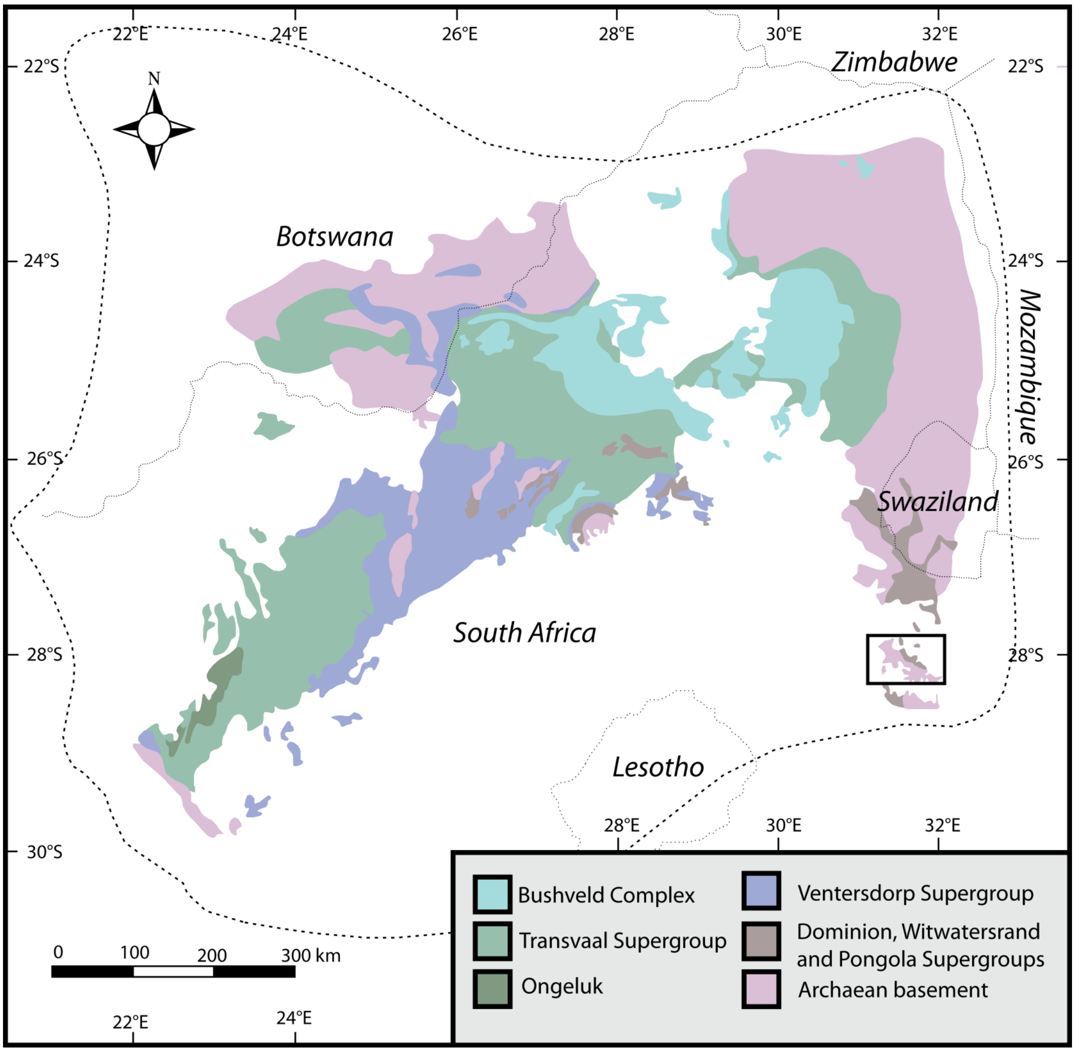

Fig. 1. A simplified map over the Kaapvaal Craton where the study area is marked with a rectangle frame. Modified from Frick (1997) and Gumsley (2015).

ites and conglomerates (Armstrong et al., 1991). A 2914 ± 8 Ma volcanic unit in the Central Rand Group represents the only dated unit in the Witwatersrand Supergroup (Armstrong et al. 1991). Coeval to the Witwatersrand Supergroup, the Pongola Supergroup was formed on the south-eastern part of the craton (Gold 2006; Wilson et al. 2013). The Pongola Supergroup consists of the lower, mainly volcanic Nsuze Group dated to between 2985 ± 1 Ma and 2966 ± 2 Ma (Mukasa et al. 2013; Hegner et al. 1984; 1994), which may correlate to the Dominion Group (cf. Cole 1994). The upper mainly sedimentary Mozaan Group can be correlated to the Witwatersrand Supergroup, directly based on lithostratigraphy (Beukes & Cairncross 1991; Gold 2006).

Overlying the Witwatersrand Supergroup is the volcanic-sedimentary succession of the Ventersdorp Supergroup, which has the largest amount of volcanic rocks preserved on the Kaapvaal Craton (van der Westhuizen & de Bruiyn 2006). The source of this extensive volcanism is not yet fully understood. According to Stanistreet & McCarthy (1991), it might be a result of extension related to the juxtaposition of Kaapvaal and Zimbabwe cratons, while Eriksson et al. (2002) stated it to be connected to a mantle plume event. Armstrong et al. (1991) obtained U-Pb zircon ages of 2714 ± 8 Ma for the Klipriviersberg Group, at the base of the supergroup, and 2708 ± 5 Ma for the Makwassie quartz porphyry, higher up in the Platberg Group. Wingate (1998), however, dated the Derdepoort Basalt, a correlative to the Klipriviersberg Group, to 2781 ± 5 Ma (U-Pb on zircon) and De Kock et al. (2012) dated the lower and upper Hartswater Group, a correlative to the Platsberg Group, to 2733 ± 3 Ma and 2724 ± 6 Ma (U-Pb on zircon). These older age determinations of correlatives to the Platberg Group and Klipriviersberg Group indicate that the ages by Armstrong et al. (1991) may be discrepant.

The Transvaal Supergroup predates the Vendersdorp Supergroup, and includes four major units. The Protobasinal rocks, the Black Reef formation, the carbonate and banded iron formation platforms of the Chunniespoort, Ghaap and Taupone groups and the chemical and clastic sediments of the Pretoria, Postmasburg and Segwagwa groups. These units have been preserved in the stratigraphically similar Transvaal basin, and Griqualand West basin which is subdivided into Ghaap Plateau sub-basin and Prieska sub-basin in South Africa and the Kanye Basin in Botswana (Eriksson et al., 2006).

Various units of sedimentary and volcanic origin with only limited lateral distribution form the Protobasinal rocks of the Transvaal Supergroup (Eriksson et al., 2006). Lavas in the upper part of the Buffelsfontein Formation of one of these Protobasinal fills have been dated to between 2657 Ma and 2659 Ma (Rb-Sr on whole rock; SACS 1993). The Protobasinal fill rocks are overlain by the Black Reef Formation in the Transvaal Basin and the Vryburg formation in the Griqualand West Basins. The Black Reef formation consists of mainly siliciclastic deposits in a fining upwards succession, while the Vryburg Formation consists of silisiclastic sediments, carbonates and volcanics (Eriksson et al. 2006). These volcanics have been dated to 2642 ± 3 Ma using Pb-Pb evaporation on zircon (Walraven and Martini 1995).

The Black Reef Formation in the Transvaal basin is overlain by the Chuniespoort Group. This group is composed of mainly carbonate platform deposits in the lower parts, and banded iron formations in the upper parts (Eriksson et al., 2006). At the base of this group a tuff layer from the Upper Oaktree Formation was dated to ca. 2585 Ma (Martin et al., 1998). In the Griqualand West Basin, the Vryburg Formation is succeeded by the Ghaap Group in the Prieska subbasin, which hosts the Nouga Formation. In the middle part of this formation a tuff layer was dated to 2588 ± 6 Ma (Altermann & Nelson 1998). The equivalent Ghaap Group Formation in the Ghaap Plateau subbasin hosts the Monteville Formation with a tuff layer in the upper part dated to 2555 ± 19 Ma (U-Pb on zircon; Altermann & Nelson 1998). At the top of the Ghaap Group in Griqualand West Basin, the Daniëlskuil Formation has been dated to 2432 ± 31 Ma (U-Pb zircon age; Trendall et al. 1990).

On top of the Ghaap Group in the Griqualand West sub-basins lies the Postmasburg Group comprised of sandstone, shale, diamictites and conglomerates at the base, and lavas from the Ongeluk Formation at the top (Eriksson et al., 2006). In the Ongeluk Formation, basaltic andesite lava yield a Pb-Pb whole rock age of 2222 ± 13 Ma (Cornell et al. 1996). In the Transvaal Basin, the Chuniespoort Group is overlain by the Pretoria Group, a 2.4–2.1 Ga succession of sandstone, conglomerates, diamictites and carbonate rocks interbedded with three larger volcanic units. The Bushy Bend Lava Member of the Timeball Hill Formation in the basal part, followed by the Hekpoort Formation and the Machadodorp volcanic Member in the Silverton Formation (Reczko et al. 1995; Eriksson et al. 2006). The Bushy Bend lavas yield a preliminary zircon age of ca 2350 Ma (Eriksson et al., 1995) whereas the Hekpoort Formation yielded a Rb-Sr isochron age of 2184 Ma (Cornell et al. 1996). The Machadodorp volcanic Member has not yet been dated. Hence, the ca. 2.2 Ga dates of Hekpoort and Ongeluk units can be correlated, and could be part of a common flood basalt event (Reczko et al., 1995; Eriksson et al., 2006).

The Transvaal Supergroup is covered and intruded by the Bushveld Complex layered intrusion, dated to 2056 ± 0.5 Ma (Pb-Pb on zircon; Zeh et al. 2015) and 2058 ± 2 Ma (U-Pb baddeleyite; Olsson et al. 2010).

The area of study is located in KwaZulu-Natal province in the south-eastern part of the craton, and is referred to as the southeasternmost window (Klausen et al. 2010; Lubnina et al. 2010). The area exposes Archaean granite and greenstone basement penetrated by mafic intrusions of various age and is surrounded by Jurassic sedimentary rocks from the Karoo Supergroup. In some areas of the window, the southernmost parts of the Pongola Supergroup are exposed on top of Archaean basement (Lubnina et al. 2010; Klausen et al. 2010). The Namaqua-Natal belt is located 50 km away from the study area. It is a tectono-stratigraphic sequence formed between ca. 1200 to 1000 Ma, during the Namaqua Orogeny, creating a continuous orogenic belt that stretches from the Northern Cape to the KwaZulu-Natal (Cornell et al. 2006). This event could have had a major impact on the area and might have altered the rocks studied.

2.1 Mafic dyke intrusions

Mafic dyke swarms in the eastern and northern basement part of the Kaapvaal Craton generally trend in a NW, E-W or NE direction. These trends originate from the Pongola and Vendersdorp rifting events, and east-west compression associated to the Bushveld Complex (Uken & Watkeys 1997). Because the trends follow large scale basement structures, the trend of individual dyke must not necessarily reflect the ambient stress regime, i.e. the least compressional stress direction being perpendicular to the dyke (Uken & Watkeys 1997; Jourdan et al., 2004; 2006).

The basement inliers exposed through the Phanerozoic cover successions in the northern KwaZulu-Natal area of South Africa form part of the southeasternmost window of the Kaapvaal Craton, revealing mafic dykes and sills of different generations and trends. According to Klausen et al. (2010) and Lubnina et al. (2010), dyke swarms with an E- to ENE-, NE- and some with a SE-trend dominate in that area. Additionally, Lubnina et al. (2010) states that the southeasternmost window is extensively cross-cut by dykes and sills of Jurassic age, penetrating the Karoo cover sequence.

Prior to this study, Petzer (2007) investigated chemical and petrological characteristics of dykes of two different trends (ENE and NE), as well as relative age relationships between dykes with four different trends (ENE, NE, SE and N-S). He suggested that the two suites with ENE- and NE-trends most likely represented two different generations. Through cross-cutting relations, the SE-trending dykes were recognized as the oldest, followed by ENE-trending dykes, NE-trending dykes and finally the youngest (Karoo) N -S-trending dykes (Petzer 2007). In the eastern Kaapvaal Craton, previous U-Pb baddeleyite datings on NE-, E- and SE- trending dolerite dykes radiating out from the Bushveld Complex, reveal ages between 2.70 and 2.66 Ga. The NE- and SE- trending dykes are intermixed with ~1.85 Ga dykes and ~2.97 Ga dykes, respectively (Olsson et al. 2010; 2011; Olsson 2012). Further south, in the southeast window, NE-trending feldspar-phyric dykes dated to ~2662 Ma form the ca. 90 km wide and 190 km long White Mfolozi Dyke Swarm (Rådman 2014). Rådman (2014) compared the White Mfolozi Dyke Swarm ages to the partly coeval NE-trending dykes further north in the eastern window, dated by Olsson et al., (2010; 2011), and suggests that the ca. 2.66 Ma dykes may belong to a common, larger swarm. Also in the southeast window, layered sills of the Hlagothi Complex and a SE-trending dyke are dated to 2866 ± 2 Ma and 2874 ± 2 Ma, respectively (Gumsley et al. 2013). In addition to the U-Pb ages, Lubnina et al. (2010) investigated the palaeomagnetism of Neoarchean-Paleoproterozoic dykes in the area, where the dykes were assigned to swarms with ages of 2.95, 2.65 and 1.90 Ga (ages determined by Olsson et al., 2010). In the south-easternmost window, Lubnina et al. (2010) sampled six dykes for paleomagnetic measurements. In the northern part two NE-trending dykes (NL-11 and NL-12) yield an age of 1.90 Ga. Further down the White Mfolozi River, an E-W trending dyke (NL-14) was investigated yielding an age of 2.65 Ga. In the White Mfolozi tributary river, the Mpembeni River, a SE-trending dyke (NL-13a) is cross-cut by an E-W-trending dyke (NL-13b) which in turn is cross-cut by a SE-trending dyke (NL-13c). The dyke NL-13a was estimated to have an age of 2.95 Ga, NL-13b to an age of 2.65 Ga and NL-13c to an age of 0.18 Ga (Lubnina el al. 2010). Klausen et al. (2010) also suggest an age of ca. 2.95 Ga for the ENE-trending dykes based on geochemical data and petrography, and an age of 2.65 Ga to the NE-rending dykes.

3 Field geology and petrography

The samples for this study were taken from dolerite dykes exposed along the Mhlatuze River and the White Mfolozi River, and along its tributary rivers, called the Mpembeni and the Ntinini, in the KwaZulu Natal province of South Africa (Fig. 2). The sampled dykes trend in either SE or ENE directions. Field observations suggest that the SE-trending dykes are older than the ENE-trending dykes (Klausen et al. 2010; Lubnina et al. 2010). Also, a NE-trending dyke, which may be part of the ca. 2.66 Ga NE- trending White Mfolozi Dyke Swarm (Rådman 2014), is cross-cut by an ENE-trending dyke, suggesting the ENE-trending dolerite dyke to be younger than ca. 2.66 Ga. The dykes are absent from the Phanerozoic Karoo and Natal cover rocks, indicating the dykes must be older than 0.18 Ga.

The majority of the samples were taken by A. Gumsley during a field trip in October 2013. Complimentary sampling was made by the author during a second field trip in August 2014. The samples were collected from river outcrops, preferably in the center of and/or in the coarser portions of the dolerite dykes. Four samples were taken from SE-trending dykes, and eleven from dykes trending in an ENE direction. One of the SE-trending dykes, here called SED03, was previously investigated paleomagnetically by Lubnina et al. (2010). They referred to this intrusion as a dyke, NL-13c in their study. However, based on careful observations in the field this intrusion is interpreted to be a subhorizontal intrusion which hereinafter will be referred to as a sill. Ten of the samples yield baddeleyite, of these two SE-trending dykes, three ENE-trending dykes and one sill were chosen for dating. In the following sections, details are given about the dykes and their petrography (sample sites of these samples are shown in fig. 2 and coordinates are given in tab. 2). The petrography of the samples was studied in thin sections at the Department of Geology at Lund University. The samples display a variable degree of alterations in greenschist facies, but with no signs of deformation and all dykes are near-vertical unless stated otherwise (Fig. 7).

3.1 SE-trending dykes

3.1.1 SED01

The sample was taken from a fine- to medium-grained, massive dolerite dyke. The dyke is approximately 10 m wide and its strike is 150°. The dyke is exposed in a White Mfolozi River tributary, the Ntinini.

The sample is dominated by sericite (replacing plagioclase), amphibole (actinolite-tremolite) and chlorite. Alterations makes it difficult to distinguish the original texture and mineralogy and only remnants of plagioclase and pyroxene are preserved. Clinopyroxene is almost completely altered to amphibole and chlorite through uralitization and chloritization. The sample is rich in quartz as quartz-ribbons. Accessory minerals include biotite, opaque minerals and primary amphibole (Fig. 7a).

3.1.2 SED02

This sample is from a 50 m wide, medium-grained dolerite dyke. It is located in eastern part of the study area within the White Mfolozi River. It is visible from the bridge on the R33 road to Ulundi (Fig. 3). The contact with the country rock, an Archaean granodiorite, is clearly visible, and no signs of metamorphism could be observed in the field. The strike is approximately 150°.

The rock has lath-shaped plagioclase surrounded by partly enveloping to subequal clinopyroxene crystals, forming a subophitic bordering to intergranular texture with minor alterations. The degree of alteration is variable throughout the rock. The sample is

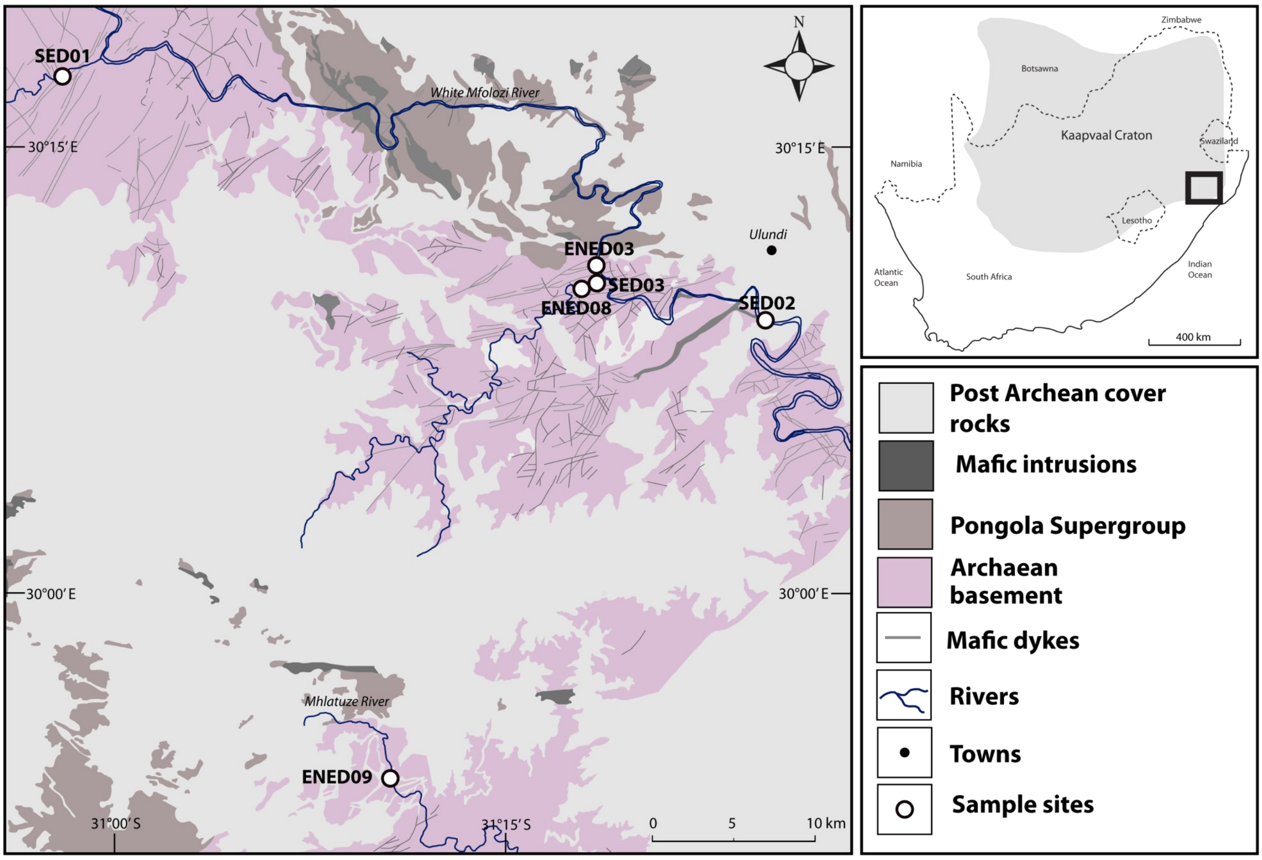

Fig. 2. Geological map over the sample site for all dated dykes and a regional overview map, with the field area marked for orientation showing the borders of South Africa, and the Kaapvaal Craton highlighted in grey. Modified from Wolmarans & Linström (1988).

made up of at least 50% plagioclase, which in some areas are altered through seritization. Approximately 20% of the sample is made up of clinopyroxene which in some regions are fresh and in others more altered. The clinopyroxene is replaced by amphibole and chlorite which together make up another ca. 20% of the sample. The final 10% of the sample is made up of opaque minerals and minor amounts of quartz (Fig. 7b).

Fig. 3. Dyke SED02 seen from the bridge on R33 towards Ulundi.

3.2 Sub-horizontal intrusion (sill)

3.2.1 SED03

SED03 is exposed in the river bank of the Mpembeni River. It was originally interpreted to be a SE-trending dolerite dyke, but was subsequently identified as a sub-horizontal intrusion, with a dip direction of ca. 40° towards NE, and a dip of ca. 15–30° (Fig. 4).

The dolerite is fresh, grey-black and unaltered. Based on paleomagnetic data, Lubnina et al. (2010) interpreted this sill to be an intrusion of the Karoo event (ca. 0.18 Ga). The SED03 sill cross-cuts an ENE -trending dyke (ENED08; NL-13b) dated to ca. 2.65 Ga in the same paleomagnetic study by Lubnina et al. (2010; Fig. 5).

This sample has a subophitic texture and moderate alterations. Abundant plagioclase (comprising at least 70% of the sample) is partly altered to sericite and enveloped by a groundmass of partly altered clinopyroxene and other unidentified phases. Additionally opaque minerals and primary amphibole occur sparsely (Fig. 7c).

Fig. 4. The subhorisontal intrusion SED03 with its 15–30° dip towards the NE marked.

3.3 ENE-trending dykes

3.3.1 ENED03

The ENED03 is from a 30 m wide dolerite dyke with 082° strike that crops out in the White Mfolozi River and has an observed contact with the granodiorite country rock. It is medium-grained, with feldspar phenocrysts and has a grey-green color. Lubnina et al. (2010) has previously assigned an age of ca. 2.65 Ga to this dyke (sample NL-14 in their study).

This sample displays extensive alteration, making the original texture and mineralogy hard to distinguish. However, in patches, plagioclase is better preserved. The sample has a groundmass dominated by seriticite, amphibole and chlorite. Plagioclase is often altered into sericite, and clinopyroxene is variably altered into amphibole, chlorite and an unidentified brown mineral. Patches of talk is intermixed with magnetite, indicative of olivine breakdown (Fig. 7d).

3.3.2 ENED08

This ca. 10 m wide fine-grained dolerite dyke trends approximately 070°. It is located in the Mpembeni River, a tributary river of the White Mfolozi River from where it is visible on both sides of a meandering bend. Lubnina et al. (2010) interpreted this dyke to be ca. 2.65 Ga in age since it cross-cuts a braided SE-trending dyke (NL-13a in their study; Fig. 5), which they assigned an age of 2.95 Ga based on its paleomagnetic directions. ENED08 is in turn cross-cut by the SED03 sill, which Lubnina et al. (2010; NL-13c in their study), interpreted to be Karoo in age (Fig. 5).

This dolerite is fine grained and has a subophitic to intergranular texture and has been altered under upper greenschist facies metamorphism. The primary mineralogy is discernible, but both plagioclase and clinopyroxene are altered to secondary minerals and amphibole, respectively. Amphibole is more abundant compared to pyroxene. Accessory phases include muscovite, biotite, quartz, opaque phases and baddelyite (Fig. 7e).

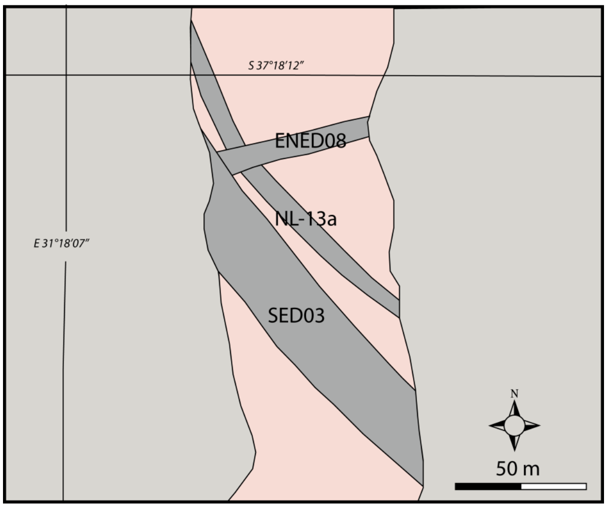

Fig. 5. A map illustrating the cross-cutting relationships between the SED03 sill, and the dykes ENED08 and NL-13a (Lubnina et al, 2010).

3.3.3 ENED09

ENED09 is located further south, in the Mhatatuze River. The dolerite dyke strikes 070°, is approximately 10 m wide and medium-grained. The contact to the country rock is exposed and the dyke cross-cuts a feldspar-phyric NNE-trending dyke (called JR03/ BCDX-27 in Rådman, 2014; Fig. 6). The NNE-trending dyke JR03 of Rådman (2014) has not been dated, but shows geochemical similarities to, and is likely part of the White Mfolozi Dyke Swarm, which has an precise age of 2662 Ma (Rådman, 2014).

This sample has minor alterations and elongated plagioclase crystals (ca 60 volume-%) are set in a matrix of large, partially enveloping pyroxene crystals forming a subophitic texture. Plagioclase is often virtually unaltered. Clinopyroxene often show simple twinning and is partly altered through uralitization and minor chloritization. Additionally, the sample contains minor amounts of biotite, opaque phases, quartz and baddeleyite.

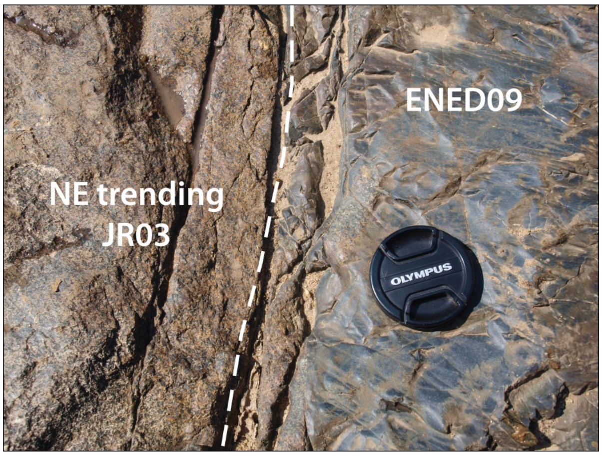

Fig. 6. The chilled margin of ENED09 at the contact to the cross-cutted NE trending dyke, JR09 (Rådman, 2014).

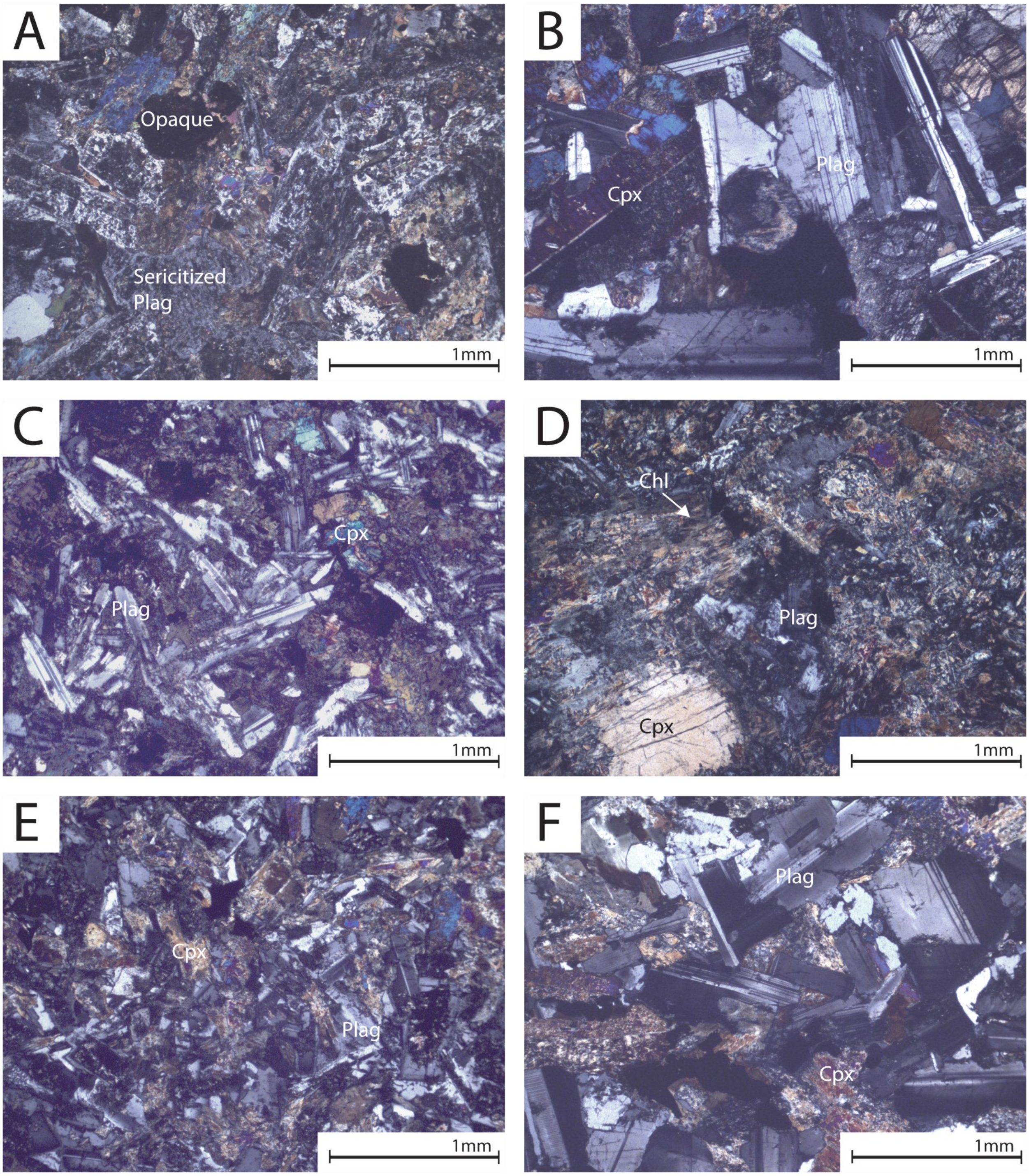

Fig. 7. Overview of the thin sections in crossed polars for a general idea of grainsize and degree of alteration; A) SED01: One of the more altered thin sections. Plagioclase is almost completely altered to sericite and pyroxene is almost completely altered to a fine mass of chlorite and amphibole. B) SED02: One of the better preserved and coarse grained samples. Only minor alterations on plagioclase, pyroxene is altered to amphibole and cholorite C) SED03: Plagioclase is unevenly altered to sericite in a groundmass of partly altered clinopyroxene D) ENED03: One of the more altered samples. A larger clinopyroxene is altered to amphibole and chlorite, starting from the rims. Groundmass of sericite, amphibole and chlorite E) ENED08: Relatively fine grained sample F) ENED09: One of the less altered samples. Plagioclase is relative unaltered but clinopyroxene is partly altered to secondary mineral.

4 Analytical protocol

Samples for this study were collected in the KwaZulu Natal area in October 2013 and August 2014. The samples were prepared for geochronological analysis and studied in thin section at the Department of Geology at Lund University.

4.1 U / Pb dating of baddeleyite

The samples were crushed using a sledgehammer and grinded into finer fractions using a swing-mill. The samples were suspended in water with a few drops of detergent to reduce adhesive forces and keep all grains and particles in suspension. A portion of the sample (ca. 30 grams) was then placed on a Wilfley shaking table following the procedures of Söderlund & Johansson (2002). After a few minutes only the finest and densest material remain on the deck and was collected in a large plastic beaker. The collected material, ideally made up of baddeleyite, apatite and opaque phases, is transferred to a petri dish. Magnetic minerals were removed using a pencil hand magnet. Finally the baddeleyite crystals were hand picked under an binocular microscope and the best quality grains were selected for isotopic analysis.

The baddeleyite grains were transferred into Teflon capsules and cleaned in multiple steps using 3.5 N nitric acid (HNO_3) including a ca. 30 minutes step in 3.5 N nitric acid on hotplate. Finally one drop of a ^{205}Pb-$^{236-233}U$ isotopic tracer was added and the sample was dissolved in 10 drops of hydrofluoric acid (HF) in an oven (190°C) over 2—3 days. This ensures complete dissolution and full homogenization of U and Pb from the sample and isotopic tracer.

At the Swedish Museum of Natural History in Stockholm the dissolved samples were put on hotplate and dried. 10 drops of 6 N HCl and 1 drop of 0.25 N phosphoric acid (H_3PO_4) were added and the mixture was then evaporated, leaving only a tiny drop. The sample drop was then taken up with 2 µl Si-gel and placed on an outgassed Rhenium (Re) filament. By letting a weak current through the filament the sample was dried. The filament temperature was then gradually increased until the phosphoric was burnt off and the sample turned white.

Mass spectrometry analysis was performed on a TIMS Finnigan Triton mass spectrometer. The intensities of ^{204}Pb, ^{205}Pb, ^{206}Pb, ^{207}Pb and ^{208}Pb were measured at a filament temperature of ca. 1200°C to 1250° C. Depending on signal intensities and the recorded $^{206}Pb/^{205}Pb$ ratio, the measurements were made in either static mode with Faraday Cups (for stronger intensities) or in peak-jumping mode using a Secondary Electron Multiplier (SEM, for weaker intensities). The U-isotopes ^{233}U, ^{236}U and ^{238}U were measured in peak-jumping mode at a temperature of approximately 1350°C. Data reduction was made in Microsoft Office Excel in an "in-house" program made by Per-Olof Persson (Swedish Museum of Natural History, Stockholm) based on the algorithms of Ludwig (2003).

5 Results

5.1 Geochronology

A total of six samples were selected for dating comprising three ENE-trending dykes and three dykes with a SE-trending direction. Between three and four baddeleyite fractions were prepared for each sample. U-Pb TIMS data is found in Table 1 and concordia diagrams in Fig. 8. A summary of the age results, with sample coordinates and trends is shown in Tab. 2.

5.1.1 SED01

Three fractions with 4–6 small grains each were analysed. The baddeleyite grains were variably frosty, indicating recrystallization of baddeleyite to zircon. The three fractions scatter and preclude a linear regression (Fig. 8a). Therefore only the calculation of a minimum age for this sample was possible, given by $^{207}Pb/^{206}Pb$ date of 2517 ± 3 Ma (fraction C).

5.1.2 SED02

A regression comprising three fractions of four small and light colored baddeleyite grains yield an upper intercept age of 2568 ± 4 Ma and a lower intercept age of 616 ± 290 Ma, with a mean square weighted deviation (MSWD) value of 2.5. One point plots 2 % discordant while two points overlap and plots slightly above the concordia (1 and 3.5 % discordant). A weighted mean $^{207}Pb/^{206}Pb$ age of 2574 ± 5 Ma (MSWD = 0.82) was calculated using fractions B and C, and is considered the best age estimate for this sample (Fig. 8b).

5.1.3 SED03

Four fractions were analysed, each with 5–8 moderately brown and minute (<40 µm in length) baddeleyite crystals. One fraction plots concordant within error, whereas the other fractions plot 2.5–9.5 % discordant. The upper and lower intercept ages are 2423 ± 7 Ma and 510 ± 180 Ma, respectively (MSWD = 0.27). The age 2423 ± 7 Ma is interpreted to represent the crystallization age for this sample (Fig. 8c).

5.1.4 ENED03

Three fractions of 3–8 small and frosty baddeleyite grains were analysed. The analyses are 3 to 11 % discordant and yield upper and lower intercept ages of 2168 ± 7 and 73 ± 210 Ma, respectively (MSWD = 0.75). The upper intercept age at 2168 ± 7 Ma is interpreted to be the age of crystallization (Fig. 8d).

5.1.5 ENED08

Baddeleyite grains were frosty and scarce in this sample making selection of high-quality crystals difficult, explaining the overall discordancy of analyses (5.5–16 %). Three fractions of 3–4 small medium-brown grains were analysed yielding upper and lower intercept ages of 2729 ± 5 and 228 ± 57, respectively (MSWD < 0.1). The upper intercept interpreted to rep-

Tab. 1. U-Pb data for the analysed fractions.

Table 1. U-Pb TIMS data

Analysis no. (number of grains)	U/ Th	Pbc/ Pbtot[1)]	$^{206}Pb/^{204}Pb$ raw[2)]	$^{207}Pb/^{235}U$ [corr][3)]	± 2s % err	$^{206}Pb/^{238}U$	± 2s % err	$^{207}Pb/^{235}U$ [age, Ma]	$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$	± 2s	Concord-ance
SED01												
Bd-1 (4 grains)	4.8	0.055	1107.0	9.7121	0.86	0.42827	0.86	2407.9	2297.9	2502.2	3.8	0.918
Bd-2 (5 grains)	10.5	0.127	466.0	9.7799	1.47	0.43866	1.46	2414.3	2344.6	2473.5	7.0	0.948
Bd-3 (6 grains)	6.9	0.025	2478.8	10.2304	0.44	0.44715	0.42	2455.8	2382.6	2517.1	2.5	0.947
SED02												
Bd-1 (4 grains)	19.5	0.059	969.9	11.2030	0.54	0.47701	0.53	2540.2	2514.2	2560.9	2.7	0.982
Bd-2 (4 grains)	13.7	0.205	240.5	11.7553	1.02	0.49646	0.97	2585.1	2598.6	2574.6	5.8	1.009
Bd-3 (4 grains)	6.0	0.464	93.6	12.0308	2.85	0.51026	2.81	2606.8	2657.7	2567.5	14.6	1.035
SED03												
Bd-1 (5 grains)	14.2	0.140	448.5	8.4561	1.88	0.39858	1.87	2281.2	2162.5	2389.4	8.5	0.905
Bd-2 (5 grains)	14.5	0.174	362.7	9.7646	2.39	0.45182	2.38	2412.8	2403.3	2420.8	10.7	0.993
Bd-3 (5 grains)	13.3	0.052	1180.0	9.2915	1.10	0.43323	1.00	2367.2	2320.2	2407.9	8.3	0.964
Bd-4 (8 grains)	15.1	0.055	1156.1	9.4967	0.96	0.44075	0.96	2387.2	2354.0	2415.7	4.0	0.974
ENED03												
Bd-1 (8 grains)	6.9	0.096	601.3	7.2066	0.82	0.38675	0.79	2137.3	2107.7	2165.9	4.9	0.973
Bd-2 (5 grains)	7.7	0.211	250.2	6.8964	1.32	0.36960	1.14	2098.2	2027.5	2168.3	12.0	0.935
Bd-3 (3 grains)	6.7	0.144	451.0	6.4607	2.18	0.34775	2.16	2040.6	1923.8	2160.7	10.6	0.890
ENED08												
Bd-1 (3 grains)	8.5	0.032	1925.2	10.8364	0.54	0.42255	0.53	2509.2	2272.1	2707.1	2.8	0.839
Bd-2 (3 grains)	7.9	0.054	1138.9	12.7082	0.86	0.49095	0.85	2658.3	2574.8	2722.4	3.5	0.946
Bd-3 (3 grains)	5.06	0.081	739.6	12.6727	1.30	0.48957	1.29	2655.6	2568.8	2722.5	5.4	0.944
ENED09												
Bd-1 (2 grains)	0.5	0.031	1359.8	11.7180	0.58	0.49326	0.58	2582.1	2584.8	2580.1	2.6	1.002
Bd-2 (4 grains)	17.7	0.034	1854.4	11.0780	0.57	0.46926	0.54	2529.7	2480.3	2569.6	3.6	0.965
Bd-3 (4 grains)	15.1	0.019	3280.1	11.0507	0.33	0.46761	0.31	2527.4	2473.1	2571.3	2.1	0.962

[1)] Pbc = common Pb; Pbtot = total Pb (radiogenic + blank + initial).

[2)] measured ratio, corrected for fractionation and spike.

[3)] isotopic ratios corrected for fractionation (0.1% per amu for Pb), spike contribution, blank (0.45-1.0 pg Pb and 0.045-0.1 pg U), and initial common Pb. Initial common Pb corrected with isotopic compositions from the model of Stacey and Kramers (1975) at the age of the sample.

resent the crystallization age (Fig. 8e).

5.1.6 ENED09

Three fractions of 2–4 small and light-brown baddeleyite crystals were analysed. One of the fractions plots concordant within error, whereas the two other plot discordant by 3–4 %. A regression yields an upper intercept at 2580 ± 3 Ma and a lower intercept at 396 ± 130 Ma (MSWD = 1.2).Crystallization is interpreted to have occurred at 2580 ± 3 Ma (Fig. 8f).

6 Discussion

6.1 Interpretation of baddeleyite U-Pb results

As previously stated, baddeleyite is ideal for dating the crystallization age of mafic igneous rocks. It is sensitive to metamorphism and alteration at which baddeleyite reacts with silica to form polycrystalline zircon (Heaman & LeCheminant 1993; Söderlund et al., 2013). The ease by which baddeleyite is transformed to zircon explains why baddeleyite rarely plot perfectly concordant, especially for rocks that underwent alteration. The data for the dolerite samples investigated in this study are variably discordant, indicating that many grains are rimmed by secondary zircon.

The lower intercept ages of SED02 (616 ± 290 Ma), SED03 (510 ± 180 Ma) and ENED09 (396 ± 130 Ma) presumably indicate baddeleyite-to-zircon reactions linked to the Pan African Orogeny at ca. 500 Ma (Kröner & Stern 2005). Similarly, the lower intercepts of ENED03 (73 ± 210 Ma) and ENED08 (228 ± 57 Ma) might be influenced by the widespread Karoo magmatic event, at ca. 180 Ma (Svensen et al. 2012). All samples are likely to have been affected by some degree of metamorphism at one or several occasions. It is, however, hard to deduce the exact cause for the difference in lower intercept ages despite the close proximity of the samples. In thin section, it is obvious that the degree of alteration is variable between the different samples, where ENED03, ENED08 and SED01 are the most altered whereas SED03, ENED09 and SED02 are the least altered. There is no obvious explanation for this variation. Locally, protected areas could have sustained alteration better or Si-rich fluids could have affected rocks locally. A thin-section might not be representative for an entire dyke. Nevertheless, the degree of alteration seen in the thin-sections can to some extent be correlated to the discordance of the

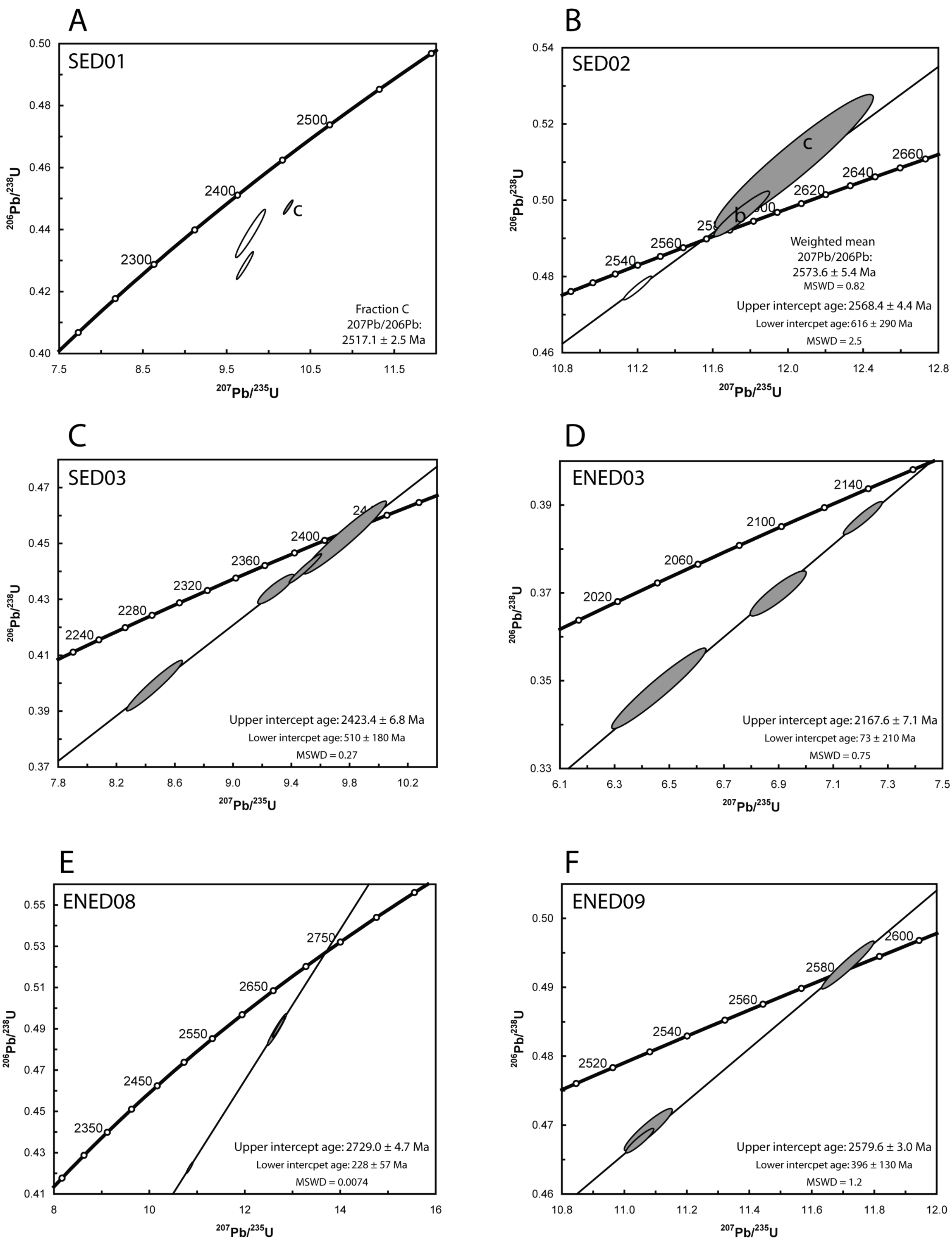

Fig. 8. Concordia diagrams from U-Pb TIMS analyses. The ^{207}Pb/^{206}Pb date for fraction C (grey ellipse) provides a minimum age of this sample. A free regression with all fractions and a weighted mean age is given by fraction B and C in SED02. The remaining, SED03, ENED03, ENED08 and ENED09 are plotted and calculated using free regressions.

Tab. 2. Sample coordinates and ages.

SAMPLE	COORDINATES	AGE (MA, 2σ)
SED01	28.22007°S, 30.95752°E	>2517 ± 3
SED02	28.34967°S, 31.40493°E	2574 ± 5
SED03	28.332632°S, 31.302685°E	2423 ± 7
ENED03	28.32339°S, 31.30208°E	2168 ± 7
ENED08	28.333260°S, 31.299697°E	2729 ± 5
ENED09	28.60901°S, 31.17285°E	2580 ± 3

data. The two least altered samples SED02 and ENED09 display the least discordant analyses while samples with more altered thin sections generally plot more discordant.

6.2 Two dyke trends with three dyke generations

Fig. 9 shows the general trends of the dykes investigated in this study. The dykes ENED08 (2729 Ma), ENE09 (2580 Ma) and dyke ENED03 (2168 Ma) follow an ENE trend, implying that the ENE trend is represented by at least three generations of dykes. Dyke swarms with multiple dyke generations are suggested to be controlled by pre-existing structures in the basement rather than representing the extensional stress regimes during intrusion. This should affect the younger generations as the oldest generation might be connected to the primary basement structures (Uken & Watkeys 1997; Jourdan et al. 2004; 2006). Dyke swarms with multiple generations can, for instance, be seen e.g. in the Pará de Minas dyke swarm on the São Francisco Craton, with at least three dyke generations (Cederberg 2013). The dykes SED01 (> 2517 Ma) and SED02 (2574 Ma) follow an SE trend. Three of the dated dykes, SE-trending SED01 (≥ 2517 Ma) and SED02 (2574 Ma) and ENE-trending ENED09 (2580 Ma) are interpreted to belong to a common, approximately 2577 Ma event. This implies that the 2577 Ma generation represents both SE- and ENE-trending dykes, however, the SE trend is only represented in the 2577 Ma generation (Figure 9).

The ENE trend may be an ”inherited” dyke direction for the younger 2577 Ma (ENED09) and 2168 Ma (ENED03) generations. However, it might represent the true basement control stress pattern for the oldest, ENED08 dyke and ENE-SWS basement lineaments controlling the paleostress in northern KwaZulu-Natal. The SE trend is however only represented in the 2577 Ma generation and could therefore represent basement control stress pattern for that generation of dykes.

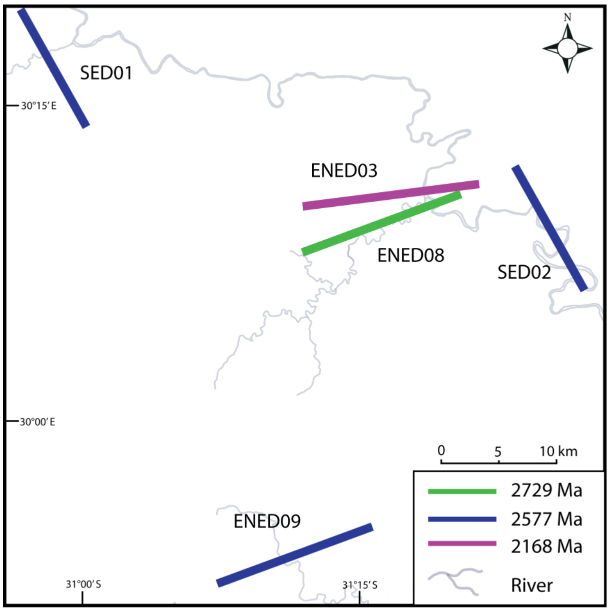

Fig. 9. Principled map over the study are showing approximate age and trend of the dykes.

6.3 Magmatism on Kaapvaal and other cratons

The results of this study reveal four generations of dykes and sills, spanning over 500 million years. Because of this time span, these generations will herein be discussed separately.

6.3.1 The 2729 Ma generation

The Ventersdorp Supergroup is a volcano-sedimentary sequence formed by multiple episodes of volcanic activity, exposed primarily in the central and western part of the Kaapvaal Craton (van der Westhuizen & de Bruiyn 2006). The ‘traditionally’ accepted ages for the lower Klipriviersberg and upper Makwassie volcanics within the Ventersdorp Supergroup are 2714 ± 8 Ma to 2708 ± 5 Ma (Armstrong et al. 1991). However, Wingate (1998) dated the Derdepoort volcanics, in a separate graben, to 2780 ± 5 Ma, which are thought to be synchronous with the lower Klipriviersberg volcanics. Further, the Modipe gabbro, coeval with the Derdepoort volcanics, was recently dated to 2784 ± 1 Ma (U -Pb on baddeleyite; Denyszyn et al. 2013). The age of 2780 Ma questions the correlation between the Derdepoort and Klipriviersberg, unless the U-Pb dates of the Klipriveirsberg volcanics are incorrect.

The oldest dated sample in this study is the 2729 ± 5 Ma old ENED08 dyke, which is 15-20 Myr older than the younger ages for the Klipriviersberg and Makwassie volcanics, presented by Armstrong et al. (1991). However, it is possible that the ENE-trending dyke ENED08, despite its more south-eastern location on the Kaapvaal Craton, is connected to Ventersdorp magmatism. In addition, ENED08 can be correlated to a pyroclastic surge and ash fall deposits, recently dated to 2733 ± 3 Ma (Mohle Formation), and a quartz-feldspar porphyry to 2724 ± 6 Ma (upper Phokwane Formation) in the Hartswater Group (de Kock et al.

2012). The Hartswater Group is a volcano-sedimentary succession in the western part of the Kaapvaal Craton, which has been stratigraphically correlated to the Platberg Group (includes the dated volcanics of the Makwassie Formation) of the Vendersdorp Supergroup (De Kock et al. 2012). Additionally, ENED08 can be correlated to a felsic porphyry described as part of the Makwassie Formation near Amalia, which has been dated to 2729 ± 3 Ma (U-Pb zircon; Poujol et al. 2005).

The Ventersdorp Supergroup magmatism has been studied primarily in the northern and western parts of the craton. Until now, there are no known exposures of Venterdorp magmatism in the KwaZulu-Natal province (van der Westhuizen et al. 1991). This study thus, extends the Ventersdorp magmatic event into the KwaZulu-Natal area, approximately 300—500 km east of the previously known exposures (see Fig. 1 and Fig. 10 for orientation).

On the Pilbara Craton, Arndt et al. (1991) published U-Pb zircon dates of volcanics from six different stratigraphic sites in the lower part of the Fortescue Group, which gave a mean age of 2765 Ma. Higher in the stratigraphy of the Fortescue Group, the Pillingini Tuff was dated at 2715 ± 6 Ma. Arndt et al. (1991) describes the 2765 Ma mean age as the onset of volcanism and the 2715 Ma age as the age at which volcanism ceased in the Fortescue. Blake et al. (2004) dated a mafic and a felsic tuff layer in the Forescue Group at 2741 ± 3 and 2724 ± 5 Ma respectively using U-Pb on zircon. The latter is a match to the ENED08 dyke in this study (within error) and the other tuffs are within a similar time interval.

6.3.2 The 2577 Ma generation

The SE-trending dyke SED02 and ENE-trending dyke ENED09, dated at 2574 ± 5 and 2578 ± 3 Ma, are located ca. 35 km apart, and are interpreted to belong to a common ca. 2577 Ma magmatic event. Also the SED01 dyke is interpreted to belong to this event of dyke intrusions, although spurious data preclude a date to be calculated.

The ~2577 Ma ages manifest the discovery of a new event of dyke intrusions, not yet identified on the Kaapvaal Craton. However, tuff layers with similar ages occur in the Transvaal Supergroup, more specifically in the Nauga (Ghaap Group in the Griqualand West area), and the Upper Oak Tree Formation (Chunniespoort Group in the Transvaal area), which are stratigraphically correlated. These tuffs are located ca. 800 km west and 400 km northwest relative to the mafic dykes investigated herein (Fig. 10). The tuff layers have been found in two separate exposures of the Upper Oaktree Formation. These tuff layers have been dated using U-Pb on zircon to 2583 ± 5 Ma and 2588 ± 7 Ma (Martin et al. 1998). Tuff layers in the middle and upper part of Nauga Formation are dated to 2588 ± 6 Ma and 2549 ± 7 Ma, respectively (Altermann & Nelson 1998). Rudolph et al. (2006) presented an age of 2580 ± 16 Ma for concordant zircon from a tuff layer in the Nauga Formation and interpreted this to be the time of deposition of this layer.

Tuff layers identified on the Pilbara Craton are similar in age to the 2577 Ma generation. At the base of the Hamersley Group, a tuff layer in the Marra Mamba Iron Formation was dated to 2597 ± 5 Ma, whereas another tuff in the overlying Wittenoom Formation yield an age of 2561 ± 8 Ma using U-Pb on zircon (Trendall et al. 1998).

The only known mafic intrusions with matching ages belong to the 2580 ± 10 Ma Caraíba Complex on the São Francisco Craton in Brazil (Oliveria et al. 2004), and the Great Dyke on the Zimbabwe Craton in Zimbabwe. The Great Dyke and its satellite intrusions yield U-Pb zircon and baddeleyite ages in the 2571 to 2579 Ma age interval (Armstrong & Wilson 2000; Wingate 2000; Söderlund et al. 2010).

6.3.3 The 2423 Ma generation

The 2423 ± 7 Ma age of the SED03 sill overlaps within error to the U-Pb baddeleyite ages of ca. 2.43 Ga, determined for the Ongeluk sills (Gumsley, pers. com. 2015) in the Griqualand West basin, located ca. 800 km to west of SED03 (Fig. 10). The Ongeluk lava has previously been dated to a significantly younger age of 2222 ± 13 Ma (Pb-Pb whole-rock, Cornell et al. 1996). The lava is situated on top of the glacially derived diamictites of the Makgayene Formation that is regarded to be part of a global glaciation event that coincided with the transition from anoxic to aerobic conditions at 2.4 Ga (e.g. Kirschvink et al. 2000; Polteau et al. 2006; Hoffman 2013). Assuming the true age of the Ongeluk lava is ca. 2.4 rather than 2.2 Ga, then there is a distinct temporal link between the herein dated sill and the Ongeluk lavas on the other side of the craton. Additionally, the Westerberg Sill (potentially a feeder to the Ongeluk volcanics) intruding the Kuruman Banded Iron Formation, in the Griqualand West Basin, can potentially be linked to the 2.4 Ga event (Fig. 10). It yields U-Pb baddeleyite ages of ca. 2441 Ma and ca. 2426 Ma (Kampmann et al. submitted).

Globally, the 2423 Ma age is similar to ages of dykes on numerous Archaean cratons, although perfect age matches are rare. On the Superior Craton in Canada, the du Chef Dyke Swarm has been dated to 2408 ± 3 Ma (Krogh, 1994). Numerous dykes of the Scourie swarm in Scotland on the North Atlantic Craton yield ages within the 2418–2375 Ma interval (Heaman & Tarney 1989; Davies & Heaman 2014). On the Kola-Karelia Craton in Norway the Ringvassøy mafic dykes yield an age of 2403 ± 3 Ma (U-Pb on zircon and baddeleyite; Kullerud et al. 2006). In the Zimbabwe Craton, the 2408 ± 2 Sebanga Poort Dyke of the Sebanga Dyke Swarm is another similar age (baddeleyite U-Pb age; Söderlund et al. 2010). The regional-scale Widgiemooltha dyke swarm on Yilgarn craton in western Australia was dated to 2411 ± 4 Ma and 2410 ± 2 Ma (Dohler & Heaman, 1998), and the Binneringie Dyke of the same swarm was dated to 2418 ± 3 Ma (Nemchin & Pidgeon 1998). The Binneringie Dyke is

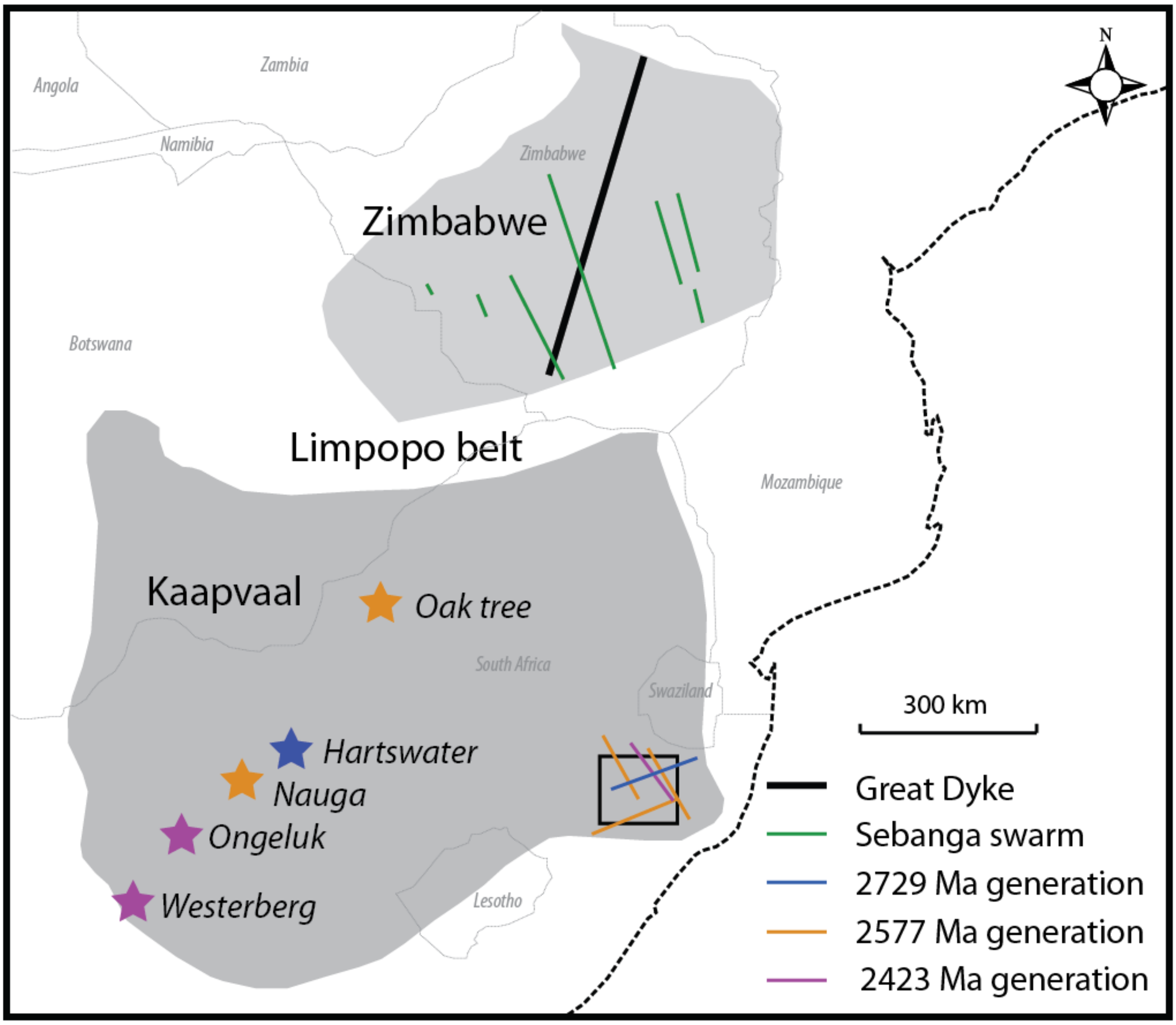

Fig. 10. Schematic map over Kaapval and Zimbabwe craton in current day positions. The dykes and sill of the 2729 Ma, 2577 Ma and 2423 Ma generations are magnified and displayed in different colours. Stars coloured after the different generations display the locations of the discussed units with ages corresponding to the different generations (modified from Söderlund et al. 2010 and McCourt et al. 2013).

coeval (within error) to the 2423 Ma generation in this study. Extensive mafic igneous activity between 2450 Ma and 2440 Ma is seen on several Archaean cratons. Examples include the above mentioned and the 2.45–2.48 Ga Matachewan Dyke Swarm on Superior Craton (Halls, 1991; Heaman, 1997; Ernst & Bleeker, 2010), and the 2.45–2.51 Ga Baltic Igneous Province on the Karelia Craton (Ernst & Buchan, 2001). This igneous activity led Heaman (1997) to speculate about a global mafic magmatic event around this time. According to Heaman (1997), the investigated Matachewan igneous event is amongst the earliest to preserve magmatism associated with flood-basalt volcanism, layered mafic intrusions and radiating dyke swarms that was to recur repeatedly during the Proterozoic and Mesozoic, forming large igneous provinces (LeCheminant & Heaman, 1989; Heaman 1997; Bleeker & Ernst, 2006). This type of mafic magmatism differs from the Archean type, typically associated with greenstone belts and komatiites (Heaman 1997). The transition in magmatic style could be associated with changes in heat transportation in the mantle, and may have influenced the break-up of the supercontinent Kenorland, during this time-period (Heaman 1997).

Although mafic units in Kapvaal Craton dated to between 2441 Ma and 2423 Ma (including SED-03) are not perfectly coeval with regional dyke swarms elsewhere, they may be part of a global ca. 2.4 Ga magmatic event linked to break up of the first supercontinent.

6.3.4 The 2168 Ma generation

The 2168 Ma ENE- trending dyke ENED03 is the youngest dyke dated in this study. There are no known magmatic correlatives of this age on the Kaapvaal Craton. The closest is the 2184 Ma date of the Hekpoort volcanic rocks from the Pretoria group (Transvaal Supergroup, located ca. 500 km northwest of the study area), and refers to a Rb-Sr isochron age (Cornell, et al. 1996). The Hekpoort volcanics are by tradition correlated to the Ongeluk volcanics, ca. 600 km southwest of Hekpoort, in the Postmasberg group. Recent age determinations demonstrate that the Ongeluk volcanics are ca. 2.43 Ga instead of 2.22 Ga (Gumsley pers. com. 2015), which render correlation between the two invalid (see also Bau et al. 1999; Moore et al. 2001, 2012; Hoffman, 2013). The radiometric dates used for the correlation between Ongeluk and Hekpoort lavas are presumably biased due to metamorphic/metasomatic alteration (Moore et al. 2001). Moore et al. (2012) suggest, partly based on U-Pb ages of detrital zircon from the Postmasburg Group and Pretoria Group (Dorland 2004), that the Ongeluk volcanics are considerably older than the Hekpoort volcanics, in agreement with the new age constraints for the Ongeluk volcanics (Gumsley pers. com. 2015). The detrital zircon obtained by Dorland (2004) originates from reworked proximal and volcanic sources from immediately underlying layers, at different stratigraphic levels in the Pretoria Group. These zircon provid maximum ages for formations overlying the Hekpoort volcanics. The Daspoort and Magaliesberg Formations, stratigraphically above the Hekpoort volcanics, have younger detrital zircon ages at 2220 Ma and 2205 Ma, respectively (Dorland 2004). With no known ~2220–2184 Ma ages on the craton, the 2168 Ma generation in this study may match the age of the Hekpoort volcanics.

There are several mafic dyke swarms worldwide that potentially could match the 2168 ± 7 Ma generation. The extensive Biscotasing dyke swarm, covering at least 300 000 km^2 on the Superior Craton in southern Canada is of the same age as ENED03. Reported ages for Biscotasing dykes are 2167 ± 1 Ma (U-Pb on baddeleyite and zircon, Buchan et al. 1993), 2168 ± 2 Ma and 2172 ± 1 Ma (U-Pb on baddeleyite, Halls & Davis 2004). In the Ungava region of northern Superior Craton, additional matches are the 2.17 Ga Payne River Dyke Swarm (S. Pehrsson 2000, cited in Ernst & Buchan, 2004) and the 2169 ± 2 Ma Cramolet Lake gabbro sill (U-Pb zircon age, Rohon et al. 1993). The Biscotasing and Payne Lake dyke swarms are possibly related as part of a common magmatic centre off the eastern tip of the Superior Craton, based on perfectly overlapping ages and areal distribution with a radiating pattern (French & Heaman 2010). To the west on the Superior Craton, the Margot Lake dyke, seemingly unrelated to the radiating dyke swarm mentioned above, provides an age of 2175 Ma (Hamilton & Stott, 2008; French & Heaman, 2010). A mafic dolerite dyke near the Wind River Mountains on the Wyoming Craton yield a U-Pb baddeleyite age of 2170 ± 8 Ma (Harlan et al. 2003). This dyke is suggested to be linked to the Biscotasing Dyke Swarm based on paleomagnetic constraint, supporting a Superior – Wyoming link (Roscoe & Card 1993; Ernst & Bleeker 2010). More distant intrusions of this generation can be found in the northern Dharwar Craton in India. The Bandepalem and Dandeli dykes, ca. 550 km apart, yield U-Pb baddeleyite and zircon ages of 2177 ± 4 Ma and 2181 ± 1 Ma, respectively (French & Heaman 2010). French & Heaman (2010) suggest these dykes to be part of a large radiating dyke swarm that includes the 2181 ± 2 Ma Duck Lake sill and the 2188 Ma Dogrib sill (Bleeker & Kamo 2003) on the Slave Craton.

6.4 The timing of formation of the Kalahari Craton

The Zimbabwe Craton and the Kaapvaal Craton form part of the greater Kalahari Craton, which formed by collision along the Limpopo belt during either the Neoarchean or Paleoproterozoic. Attempts to date the high-grade metamorphism indicate a collision at 2.7–2.6 Ga or at 2.0 Ga (e.g. Kamber et al., 1995; Holzer et al., 1999 Kreissig et al., 2001; Van Reenen et al., 2008). Söderlund et al. (2010) compared emplacement ages of regional dyke swarms and sill complexes from the Kaapvaal and Zimbabwe cratons. No age matches were found prior to ca. 2.0 Ga, whereas the oldest event common on both cratons is the 1.88–1.83 Ga mafic magmatic event. This includes the Mashonaland sills in Zimbabwe and the post-Waterberg sills, and Black Hills Dyke Swarm in Kaapvaal (Figure 10). Söderlund et al. (2010) therefore argued that collision did not occur until ca. 2.0 Ga.

The 2577 Ma generation in this study is coeval (within error) with the 2575 ± 2 Ma Great Dyke and its satellite intrusions (Fig. 11, Söderlund et al. 2010) in Zimbabwe, hence challenge the interpretation by Söderlund et al. (2010). However, the here dated intrusions in Kaapvaal Craton are located roughly 800 km from the Great Dyke in Zimbabwe. The Great Dyke and its sub-parallel satellites stretch along the entire Zimbabwe Craton from NNE to SSW, for approximately 550 km, but ends at the northern edge of the Limpopo Belt. This imprint is not traceable over the Zimbabwe – Kaapvaal Craton terrane boundary, and no coeval magmatism has previously been discovered. Furthermore, the 2577 Ma dykes in Kaapvaal have trends (SE- and ENE-trends) completely different from the Great Dyke which argue against that the Kaapvaal and Zimbabwe blocks were connected at 2575 Ma in their present-day positions (Fig. 10).

The sill SED03, dated to 2423 ± 7 Ma is similar in age to the younger 2408 ± 2 Ma Sebanga Poort dyke on the Zimbabwe Craton (Fig. 11, Söderlund et al. 2010). As with the Great Dyke, the Sebanga Poort dyke is located ca. 800 km northwest to the study area. The NNW-trending Sebanga Poort dyke is part of the Sebanga Dyke Swarm that intrudes the entire Zimbabwe Craton for 350 km, with a vast number of parallel dykes that sprawls over at least 500 km (Fig. 10). No magmatic correlatives are known in the area between the 2423 Ma SED03 dyke and the 2408 Ma Sebanga Poort dyke, and the ca. 15 Ma years disparity between them are substantial.

Ultimately, the differences between the intrusions on the two cratons make a genetic link between the events unlikely, and the 2577 Ma and 2423 Ma generations in this study do not indicate a amalgamation of Kaapvaal and Zimbabwe prior to ca. 2.0 Ga. This due to the difference in trend and different trendlines. However, a genetic link cannot be ruled out, and a common event before the amalgamation could be responsible for emplacement of the dykes.

6.5 Potential barcode reconstructions

Based on existing geological knowledge of Archean cratonic fragments worldwide, several crustal configurations have been proposed in the Archaean and early Proterozoic. Among these are the more accepted Superia, Sclavia and Vaalbara cratons (e.g. Bleeker 2003). Formation of the supercontinent Kenorland, is believed to have taken place at ca. 2.7 Ga, with subsequent break up between ca. 2.5–2.1 Ga (Aspler & Chiarenzelli 1998). The supercontinent is suggested to have joined mainly the cratons of present day North America, the Siberian and the Baltic cratons (Williams et al. 1991; Aspler & Chiarenzelli 1998). Based on slightly different geological records for crustal blocks today localized in the southern hemisphere, these are suggested to have formed another, coexisting continent. Different configurations of such a continent have been suggested, including the Vaalbara supercraton, made up by coherent Kaapvaal and Pilbara cratons (e.g. Trendall 1968; Nelson et al. 1992; Cheney 1996; de Kock et al. 2009; 2012), and Zimvaalbara, which places Kaapvaal together with Pilbara, Zimbabwe, São Francisco and possibly many Indian cratonic blocks (Stanistreet 1993; Aspler & Chiarenzelli 1998; Smirnov et al. 2013). Common for these previously proposed configurations are the accepted amalgamation of Kaapvaal and Pilbara cratons.

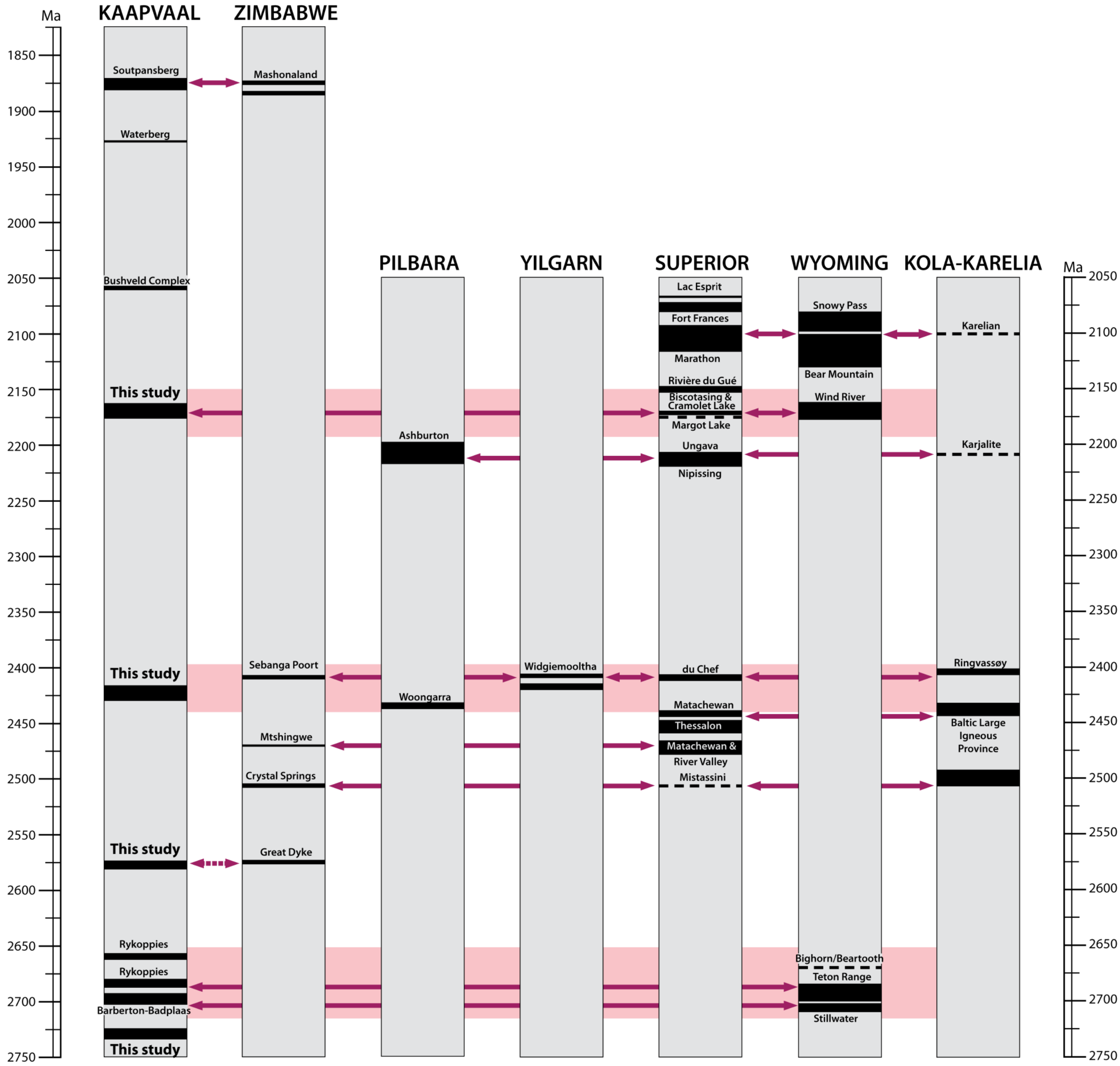

Fig. 11. Potential barcode reconstruction figure. The primarily discussed parts of the barcode reconstruction are marked in pink. The thickness of the bars shows the uncertainty in the ages and dashed lines indicate unpublished ages. The detail concerning the different barcodes can be seen in tab. 3.

Tab. 3. The barcode details for Figure 8, showing the age, error (if known), name, provenance and type of magmatism with the reference.

Age	Name	Craton	Type	Reference
1879 ± 1 & 1872 ± 2	Soutpansberg/Waterberg	Kaapvaal	dolorite sills	Hanson et al. 2004
1886 ± 2 & 1875 ± 2	Mashonaland	Zimbabwe	dolorite sills	Söderlund et al. 2010
1927 ± 1	Waterberg	Kaapvaal	dolorite sills	Hanson et al. 2004
2058 ± 2	Bushveld Complex	Kaapvaal	Layered mafic intrusion	Olsson et al. 2010
2069 ± 1	Lac Esprit	Superior	dyke swarm	Buchan et al. 2007
2076 +5/-4	Fort Frances	Superior	dyke swarm	Buchan et al. 1996
2092 ± 9	Snowy Pass	Wyoming	gabbro	Premo & van Schmus 1989
ca. 2100-2110	Karelian	Karelia	dykes	Vuollo & Huhma 2005 (cited in Ernst et al. 2013); Vuollo et al. 1995
2113 ± 15	Bear Mountain	Wyoming	mafic dykes	Bowers & Chamberlain 2006
2126 ± 1 - 2106 ± 2	Marathon	Superior	mafic dyke swarm	Halls et al. 2008
2149 ± 3	Rivière du Gué	Superior	dyke swarm	Maurice et al. 2009
2168 ± 7	2168 Ma generation	Kaapvaal	mafic dyke	This study
2167 ± 1, 2168 ± 2 & 2172 ± 1	Biscotasing	Superior	dyke swarm	Buchan et al. 1993; Halls & Davies 2004
2169 ± 2	Cramolet Lake	Superior	gabbro sill	Rohon et al. 1993
2170 ± 8	Wind River	Wyoming	mafic dyke	Harlan et al. 2003
ca. 2175	Margot Lake	Superior	dyke	Hamilton & Stott 2008 (cited in French & Heaman 2010)
ca. 2200-2220	Karjalite	Karelia	layered mafic sills	Vuollo & Phrainen 1992; Vuollo et al. 1995
2208 ± 10	Ashburton	Pilbara	mafic sills	Müller et al. 2005
2216 +8/ -4 & 2209 ± 1	Ungava	Superior	mafic dyke swarm	Buchan et al. 1998
2217 ± 4 & 2210 ± 4	Nipissing	Superior	diabase intrusions	Noble & Lightfoot 1992
2403 ± 3	Ringvassøy	Kola-Karelia	mafic dykes	Kullerud et al. 2006
2408 ± 3	du Chef	Superior	dyke swarm	Krogh 1994
2408 ± 2	Sebanga poort dyke	Zimbabwe	mafic dyke	Söderlund et al. 2010
2418 ± 3 & 2410 ± 2	Widgiemooltha	Yilgarn	mafic dyke swarm	Nemchin & Pidgeon 1998; Dohler & Heaman 1998
2423 ± 7	2423 Ma generation	Kaapvaal	mafic sill	This study
ca. 2450 - 2510 (2445-2433, 2505-2490)	Baltic large igneous province	Karelia	dykes & layered intrusions	Ernst & Buchan 2001 (and references therein)
2449 ± 3	Woongarra	Pilbara	large igneous province	Barley et al. 1997
2453 ± 6	Thessalon	Superior	basalt & rhyolite	Ketchum et al. 2013
2470 ± 1	Mtshingwe	Zimbabwe	mafic dyke	Söderlund et al. 2010
2473 +16/-9 & 2446 ± 3	Matachewan	Superior	dyke swarm	Heaman 1997
2475 ± 2	River Valley	Superior	intrusion	L. Heaman 2009 (cited in Söderlund et al. 2010)
ca. 2505	Mistassini	Superior	dyke swarm	Bleeker et al. 2008 (cited in Ernst & Bleeker 2010; Södelund et al. 2010)
2512 ± 2	Crystal Springs	Zimbabwe	mafic dyke	Söderlund et al. 2010
2571 ± 9, 2574 ± 2, 2575 ± 2 & 2579 ± 3	Great Dyke and satellites	Zimbabwe	mafic intrusion/dykes	Armstrong & Wilson 2000; Wingate 2000; Söderlund et al. 2010
2577	2577 Ma generation	Kaapvaal	mafic dyke	This study
ca. 2670	Bighorn/Beartooth	Wyoming	mafic dykes	Chamberlain 2014
2683 ± 2, 2686 ± 6 & 2659 ± 3, 2662 ± 3	Rykoppies	Kaapvaal	mafic dyke swarm	Olsson et al. 2010
2693 ± 8	Teton Range/Rendevous metagabbro	Wyoming	metagabbro	Chamberlain et al. 1999
2692 ± 1, 2698 ± 4	Barberton-Badplaas	Kaapvaal	mafic dykes	Olsson et al. 2011
2705 ± 4	Stillwater layered mafic complex	Wyoming	layered mafic complex	Premo et al. 1990
2729 ± 5	2729 Ma generation	Kaapvaal	mafic dyke	This study

6.5.1 Vaalbara

The link between Kaapvaal and Pilbara craton (northwestern Australia) has been suggested by several authors based on stratigraphic correlations and other similar geological features (e.g. Trendall 1968; Nelson et al. 1992; Cheney 1996; de Kock et al. 2009; 2012). The Ventersdorp Supergroup on Kaapvaal correlates to the volcano-sedimentary Fortescue Group on Pilbara Craton in Western Australia. This correlation is based on geochronology, lithological, geochemical and paleomagnetic similarities (e.g. Nelson et al. 1992; de Kock et al. 2009; 2012. In this study, the 2729 Ma age of the ENED08 dyke overlaps within error of the 2724 Ma age of a tuff layer in the Fortescue Group (Blake et al. 2004). The 2577 Ma generation is close in age to tuff layers dated on the Pilbara Craton. Additionally, similar ca. 2577 Ma tuff layers can be found on the Kaapvaal Craton. Despite the lack of exact age matches, the results from this study, revealing magmatism in the same age range as magmatism on the Pilbara, can be interpreted to further support a link between the Kaapvaal and Pilbara cratons, at least at 2729 Ma.

6.5.2 Kaapvaal - Superior - Wyoming

The absence of a ca. 2577 Ma magmatic event in Superior and Slave, as well as associated cratons, might indicate that the Kaapvaal Craton never was part of neither Sclavia or Superia, but that Vaalbara rather was a separate supercraton during this time. This corresponds to the previous ideas that Kaapvaal, Superior and Slave cratons have their ancestry from separate supercratons, based on differences in age of cratonization and cover sequence (Bleeker 2003). However, as can be seen in fig. 11, with references shown in tab. 3, the 2423 Ma sill in this study is close in age to the 2408 Ma du Chef Dyke Swarm (Krogh, 1994). Similarlily, the 2167 Ma dyke, dated in this study, is a possible match with the 2167–2172 Ma Biscotasing dyke swarm (Buchan et al. 1993; Halls & Davis 2004), the 2169 Ma Cramolet Lake gabbro sill (Rohon et al. 1993) and the 2.17 Ga Payne River Dyke Swarm (S. Pehrsson 2000, cited in Ernst & Buchan, 2004) on the

Superior Craton. The apparent polar wander paths of poles from the Kaapvaal and Superior cratons show that a juxtaposition of the cratons between 1110 Ma and 1880 Ma, is impossible. They could, however, have belonged to a common block from 1880 to 2220 Ma, though with a 60° distance between them. Between 2220 and 2680 Ma, the apparent polar wander path tests of the different poles attests to a possibility of a coherent landmass of many cratons (Evans & Pisarevsky 2008). A coherent landmass during the formation of the 2167 Ma generation makes a link between the 2167 Ma generation in this study, and the Superior dyke swarms possible. Comparative paleomagnetic studies of these dykes could provide further constraints to this possible link.

Another suggested nearest neighbour to the Superior Craton is the Wyoming Craton, based on sedimentological and stratigraphic similarities between the Huronian Supergroup on Superior Craton and the Snowy Pass Supergroup on Wyoming Craton, and widely distributed events at ca. 2100 Ma on both cratons (Roscoe & Card 1993; Heaman 1997; Ernst & Bleeker 2006). Ernst & Bleeker (2010) link the cratons based on geochronology and paleomagnetic constraints of the Biscotasing Dyke Swarm on Superior, and a dyke near Wind River Mountains on the Wyoming Craton. As can be seen in fig. 11, the 2167 Ma generation in this study can be correlated to both the Biscotasing dyke swarm, other 2.17 Ga dykes on the Superior Craton, as well as the 2170 ± 8 Ma dyke near Wind River on the Wyoming Craton (Harlan et al. 2003). Recently, Chamberlain (2014) suggested a Kaapvaal – Wyoming link at 2.7 Ga based on the similar ages of mafic dykes exposed in the Wyoming Craton and Kaapvaal Craton, an argument strengthened by 2.7 Ga metamorphism (12 kbar) on both cratons. Dated dykes in Bighorn and Beartooth Mountains on the Wyoming Craton reveal magmatic pulses at approximately 2.71, 2.69 and 2.67 Ga. The orientation of these and other dykes in the area suggest at least two mantle plume heads. One mantle plume at 2.71 Ga has a center in southern Montana (2705 ± 4 Ma Stillwater layered mafic complex; Premo et al. 1990) and the one at ca. 2.69 Ga with a center in Teton Range (near the 2693 ± 8 Ma Rendevous metagabbro; Chamberlain et al. 1999), in western Wyoming. Similar age-groups have been reported from the Kaapvaal Craton at 2.71 Ga and 2.69 Ga (Chamberlain 2014), e.g. the 2966 ± 1 and 2698 ± 1 Ma aged Barberton-Badplaas swarm, the older 2683–2685 Ma and younger 2659–2662 Ma dykes of the Rykoppies dyke swarm (Olsson et al. 2010; 2011) and the ca. 2710–2780 Ma Ventersdorp volcanics (e.g. Armstrong et al. 1991; de Kock et al. 2012; Denyszyn et al. 2013). One radiating dyke swarm in the northeastern part of Kaapvaal Craton have been dated to ca. 2700–2660 Ma, originating in the eastern parts of the Bushveld Complex (Olsson et al. 2010; 2011). According to Chamberlain (2014), these dykes may have fed the 2.67 Ga generation of mafic dykes found in the Wyoming Craton. The 2167 Ma dyke in this study provides a possible match to the 2170 ± 8 Ma dolerite dyke near Wind River Mountains on the Wyoming Craton (Fig. 11, Harlan et al. 2003). Preliminary paleomagnetic data from the dykes in Bighorn and Beartooth Mountains indicate similar paleolatitudes between Wyoming and Kaapvaal at 2.7 Ga. However, the paleolatitudes of the cratons diverge at 2.2 Ga, indicating rifting before the formation of the 2167 Ma and 2170 Ma dykes on Kaapvaal and Wyoming cratons, between 2.7 Ga and 2.2 Ga, under the assumption that the Ongeluk volcanics are of this age (Chamberlain 2014).

Fig. 12 shows a tentative reconstruction with a possible location of the Kaapvaal craton relative to Superior, Wyoming, Kola-Karelia, Zimbabwe and Yilgarn. Placing Kaapvaal to the left of Zimbabwe, Superior and Yilgarn, allows for a structural match between the ca. 2170 Ma dyke in this study and the ca. 2170 Ma Biscotasing dyke swarm on Superior (Halls & Davies 2004), which also has a temporal connected to the ca. 2170 Ma Wind River dyke on Wyoming (Harlan et al. 2003). This reconstruction also opens for a potential connection between the ca. 2420 Ma sub-horizontal sill and the ca. 2420–2410 Ma Widgiemooltha dyke swarm (Nemchin & Pidgeon 1998). Note that the assumed trend of ca. 2577 Ma generation becomes parallel with the ca. 2575 Ma Great Dyke on Zimbabwe Craton (Söderlund et al. 2010; for more information, see Fig. 10 and Tab. 3), but would imply a huge amount of distance between dykes when extrapolated along strike in this configuration. Also for reasons discussed a genetic link between these intrusions remains uncertain. The fact that magmatism of ages close to the 2423 Ma and 2167 Ma generations are traceable between several cratons that are suggested to be nearest neighbours at the time, provides further support for a Superior-Wyoming-Kaapvaal link during this time period.

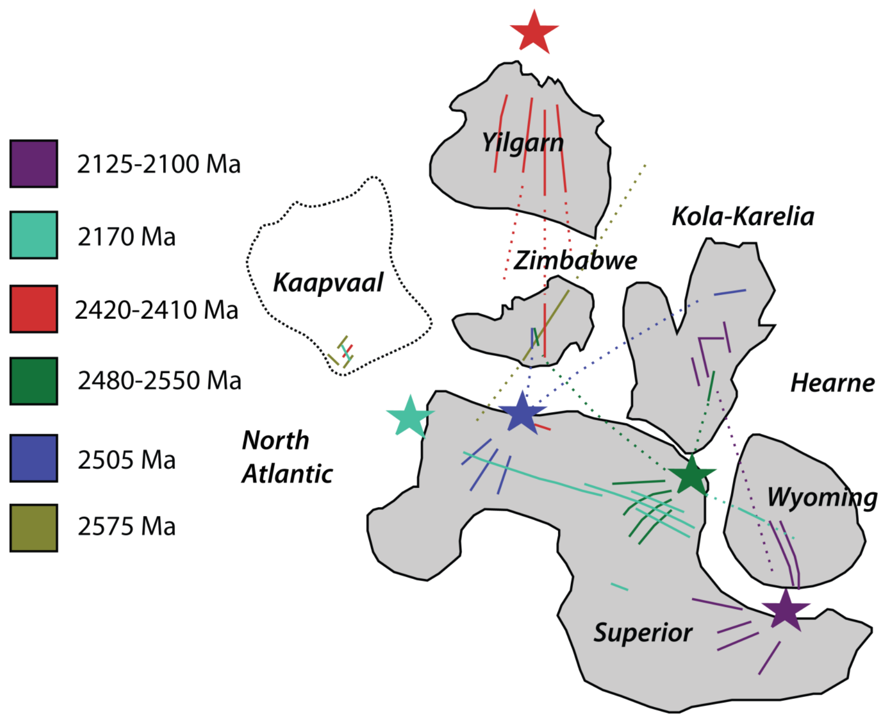

Fig. 12. Principled paleogeographic reconstruction based on dyke swarm patterns, involving Superior, Wyoming, Kola-Karelia, Zimbabwe, Yilgarn and Kaapvaal Cratons. The position of Kaapvaal is solely based on dyke patterns. The stars represent suggested mantle plume head positions. Paleomagnetic data are not accounted for. Modified after Pisarevsky et al. (2015) with position of Wyoming after Ernst & Bleeker (2006) and position of Zimbabwe and Yilgarn after Söderlund et al. (2010).

7 Conclusions

- Four generations of mafic intrusions in the KwaZulu-Natal area in the south-eastern most Kaapvaal Craton have been dated using U-Pb on baddeleyite to 2729 ± 5 Ma, 2577 Ma (2574 ± 5 Ma, 2580 ± 3 Ma), 2423 ± 7 Ma and 2168 ± 5 Ma.
- The 2729 Ma, 2577 Ma and 2168 Ma generations have an ENE-trend. The direction of dyke emplacement is potentially controlled by pre-existing structures in the basement, at least among the younger generations.
- The 2577 Ma generation have an SE-trend which might represent the true basement control pattern on that generation.
- The 2729 Ma generation is coeval with the 2724 Ma Phokwane Formation and 2733 Ma Mohle Formation of the Hartswater Group, correlated to the Platsberg Group of the Ventersdorp Supergroup. If the 2729 Ma generation in this study in fact is related to the Ventersdorp magmatism it extends its distribution ca. 300 km south-east of previously known exposures. Additionally, broadly coeval tuff layers in the Fortescue Group on Pilbara can strengthen a possible Kaapvaal-Pilbara link.
- The 2577 Ma dyke generation represents a new age on the Kaapvaal. Tuff layers in the Nauga and Oak Tree Formation of the Transvaal Supergroup are of similar age. Coeval ages can be found in the Caraíba Complex on the São Francisco Craton in Brazil and the Great Dyke and its satellites on the Zimbabwe Craton. Based on differences in trend, character and the large distance separating the 2577 Ma generation in this study and the Great Dyke, these do not suggest an amalgamation of Kaapvaal and Zimbabwe Craton prior to 2.0 Ga.
- The 2423 Ma aged sub-horizontal intrusion is similar to new ages of the Westerberg Sill. Similar ages on several cratons indicate a much larger magmatic event worldwide from 2450–2440 Ma.
- The 2168 Ma dyke age is a previously unknown magmatic age on Kaapvaal Craton, however it is coeval to other ages on the Superior Craton and on the Wyoming Craton. Additional coeval ages between the Kaapvaal and Wyoming Craton around 2700 Ma provides an additional link between the two cratons which data support a Kaapvaal-Superior-Wyoming amalgamation at these times.

8 Acknowledgements

First of all I would like to thank to my supervisor Ulf Söderlund and co-supervisors Ashley Gumsley and Andreas Petersson for all their help and support during this work. Lastly I would like to thank my friends and family for their support and encouragement, and Ludvig for the love and understanding.

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