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Paleomagnetic study of NeoArchean–Paleoproterozoic dykes in the Kaapvaal Craton

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ABSTRACT

431 oriented samples were collected from 27 dolerite dykes at 17 sites, belonging to 2.95, 2.65, and 1.90 Ga swarms, that trend SE, E and NE, respectively from the Bushveld Igneous Complex into the eastern Kaapvaal Craton (ages determined by Olsson et al., 2010; Olsson in Söderlund et al., this volume). Samples were analyzed for paleomagnetism and also anisotropy of magnetic susceptibility (AMS). For the 2.95 Ga SE-trending dykes high temperature/coercivity 'P' component has unblocking temperatures up to 590 °C and coercivity 40–90 mT and demonstrate SSW declination and intermediate positive inclination. Based on positive contact and conglomerate tests we argue for a primary origin of this component. The paleopole (BAD), calculated from '*P*' component, does not correspond to any of the previously obtained Archean–Paleoproterozoic paleopoles for the Kaapvaal Craton, and represents a new key pole for 2.95 Ga. The high-coercivity '*H*' component for the 2.65 Ga-old E-trending dykes has a SSW declination and steep positive inclination. Paleomagnetic pole (RYK), recalculated from this component, is close to the paleopoles, obtained by Wingate (1998) and Strik et al. (2007) for 2.78 Ga Ventersdorp volcanics. The third group, NE-trending dykes of the 1.90 Ga Black Hill swarm demonstrate an '*M*' component with dual polarity high-coercivity component with SSE-declination and negative intermediate inclination. The paleopole (BHD), calculated from this component is close to the 1.87 Ga pole of the Kaapvaal Craton obtained by Hanson et al. (2004). Overprint directions include a very well developed thermo-chemical overprint (Dec = 329° Inc = −36°), which is believed to be associated with a ∼0.18 Ga regional 'Karoo' thermal event.

1. Introduction

Many greenstone belts terminate at a high angle to the current margins of the Kaapvaal Craton, indicating that this craton is truncated and may have belonged to a larger supercraton that, possibly, fragmented at different stages during the Paleoproterozoic, prior to the amalgamation of the Limpopo, Kheis, Namaqua-Natal and Mozambique mobile belts (e.g., Jacobs et al., 2008). Various Neoarchean supercratonic arrangements have been proposed (e.g., Rogers, 1996; Aspler and Chiarenzelli, 1998), but there is no consensus regarding nearest neighbors for the Kaapvaal Craton, except for a proposed linkage ("Vaalbara") with the Pilbara craton (Cheney, 1996; Wingate, 1998; Nelson et al., 1999; Eriksson et al., 2009). Other than this link, the identification of other blocks formerly adjacent to other sides of the Kaapvaal in the NeoArchean–Paleoproterozoic is unknown.

Nevertheless, the robust identification of former nearest neighbors to the Kaapvaal Craton in the latest Archean to early Proterozoic is of significance for allowing the correlation of geological belts and boundaries, sedimentary basins, lithospheric roots, etc., between the Kaapvaal Craton and its former neighbors. In addition, there are potential economic implications to be gained from the tracing of metallogenic provinces between blocks. For instance, the economically important ca. 2060 Ma Bushveld event consists of the well-known Bushveld Igneous Complex as well as additional smaller intrusions and extrusions which are widespread in the Kaapvaal Craton (e.g., Mapeo et al., 2006). Additional units of the Bushveld event should be present on the former nearest neighbors to the Kaapvaal Craton.

Paleomagnetism offers a robust approach for producing well-constrained reconstructions. There have been several previous paleomagnetic studies of NeoArchean and Paleoproterozoic units of the Kaapvaal (e.g., Pisarevsky, 2005). However, the general prob-

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lem with obtaining Precambrian paleomagnetic poles, not only in the Kaapvaal Craton, but globally, has been the absence of precise dates on the units studied paleomagnetically. This problem has been noted by Buchan et al. (2000), who surveyed the available 'key paleomagnetic' poles for Laurentia and Baltica, i.e., those for which there was a robust paleomagnetic direction that averaged secular variation, was demonstrated primary on the basis of a field test, and had also a precise age date. Of the hundreds of poles published for Precambrian units, only a handful were considered 'key poles', and of these, the majority were in the Canadian shield.

A reliable Proterozoic APWP (Apparent Polar Wander Path) for the Kaapvaal Craton does not exist. Toward this goal we initiated a paleomagnetic study of units that have been precisely dated by the U–Pb baddeleyite method in companion studies (Olsson et al., 2010; Olsson in Söderlund et al., 2010). As in these dating studies, we focus on the dyke swarms that have trends which appear to radiate from the eastern side of the Bushveld Complex. We aim to obtain new paleomagnetic results on the two swarms precisely dated in the above mentioned study by Olsson et al. (2010): a SW-trending swarm of age 2965 Ma (referred to as 2.95 Ga), an E–W trending swarm with an age range of 2685–2660 Ma (referred to as 2.65 Ga). We also present paleomagnetic results for a NE-trending dyke swarm with a preliminary age of ca. 1.90 Ga (Olsson in Söderlund et al., 2010; Klausen et al., 2010). We propose new paleomagnetic poles for the Kaapvaal Craton for these times. We also use AMS (anisotropy of magnetic susceptibility) data in order to determine magma flow direction (e.g., Ernst and Baragar, 1992; Tarling and Hrouda, 1993; Polteau et al., 2008).

2. Regional geology

We refrain from any extensive review of South Africa's Neoarchean to Paleoproterozoic geology and refer the reader to both Johnson et al. (2006) and regional geology sections in companion papers in this volume (Olsson et al., 2010; Klausen et al., 2010). Instead, we focus our brief summary on (1) host rock units within the studied part of the Kaapvaal Craton (Fig. 1a), (2) surrounding cover rocks that are younger than 2.7 Ga, and (3) relevant aspects of the complex array of mafic dyke swarms that cross-cut various parts of the eastern Kaapvaal Craton (Fig. 1b–e).

2.1. Eastern Kaapvaal Craton basement and Pongola Supergroup

Fig. 1(a) shows the distribution of samples which were collected from dykes hosted in the oldest (Paleo- to Mesoarchean), eastern part of the Kaapvaal Craton (e.g., Eglington and Armstrong, 2004), including two sites (NL11 and NL12) within a ca 3.0–2.9 Ga lava-sediment cover sequence (Pongola Supergroup). In general, tonalite–trondhjemite–granodiorite (TTG) gneisses predominate within the Paleoarchean, while a granodiorite–monzonite–syenite (GMS) suite of intrusions make up most of the Mesoarchean (Robb et al., 2006). Neoarchean granitoids are not directly relevant to our study, except for the fact that most SE-trending mafic dykes do not cut post-Pongola granitoids, in agreement with this swarm's syn-Pongola age (2.95 Ga; Olsson et al., 2010). The eastern Kaapvaal Craton (Fig. 1a) also includes some of the oldest (3.5–3.2 Ga) and best preserved greenstone belts in the world (Brandl et al., 2006), but these are generally too dark to provide any good contrast to the same cross-cutting mafic dyke swarms that are more readily mapped in equally old granitoids.

2.2. Dyke swarms across eastern Kaapvaal Craton

There were few precise, absolute ages on the eastern Kaapvaal Craton dykes prior to the precise U–Pb dating by Olsson et al. (2010) and Olsson in Söderlund et al. (2010) which identified at least three separate swarms: (1) the 2.95 Ga SE-trending Badplaas swarm, (2) the 2.65 Ga E–W trending fanning Rykoppies swarm, and (3) the 1.90 Ga NE-trending Black Ridge swarm.

The oldest 2.95 Ga-old SE-trending Badplaas dyke swarm probably acted as feeders to the Pongola volcanics (e.g., Hunter and Halls, 1992), because these dykes are generally cut by post-Pongola (NeoArchean in Fig. 1a) granitoids, are orientated sub-parallel to the possible orientation of a Pongola rift, and have roughly similar compositions (Klausen et al., 2010). Consistent 2.65 Ga ages negate previous inferences suggesting that E-trending Rykoppies dykes were feeding the neighboring E–W elongated Bushveld Complex (Uken and Watkeys, 1997; Anhaeusser, 2006). Instead, this swarm appears to be radiating (Olsson et al., 2010) and probably acted as feeders to the upper (Allanridge) lava formation within the Ventersdorp Supergroup (Klausen et al., 2010). Finally, a ca. 1.90 Ga age for NE-trending Black Ridge dykes argues for a widespread magmatic event which is correlated to sills in the Waterberg Group (Hanson et al., 2004), lavas within the Soutpansberg Group (Barker et al., 2006), and even dykes and the Mashonaland sills within the juxtaposed Zimbabwe Craton (Söderlund et al., 2010; Klausen et al., 2010).

In addition to the three dated swarms that have been studied herein there are also dykes present in each area that have other trends. These potentially represent additional magmatic events (e.g., the 0.18 Ga Karoo, 1.10 Ga Umkondo, 2.05 Ga Bushveld, 2.20 Ga Ongeluk, or other events), and will be targeted in the future for precise U–Pb dating and associated petrogenetic and paleomagnetic studies.

2.3. <2.65 Ga cover rocks and tectonic/metamorphic overprint

The eastern Kaapvaal Craton is disconformably bounded by two well-preserved supergroups: (1) the 2.65–2.05 Ga (Vaalian Epoch) Transvaal Supergroup towards the west (Fig. 1a; e.g., Eriksson et al., 2006), and (2) the 325–175 Ma Karoo Supergroup across its southern and eastern extent (e.g., Johnson et al., 2006). Both supergroups are comprised of relatively thick and varied sedimentary sequences, which are capped by rapidly emplaced and extraordinarily voluminous igneous deposits; i.e., the Bushveld Complex (Cawthorn et al., 2006) and the Karoo Igneous Province (Duncan and Marsh, 2006).

It is debated, whether the Bushveld Complex was formed (1) during a northerly collision with the Zimbabwe Craton, along the Limpopo Belt, (2) in combination with more westerly subduction prior to a collision with the Congo Craton, along the Ubendian Belt and/or (3) as a plume-induced Large Igneous Province. It is, at least, superseded by the emplacement of coeval (ca. 1.90 Ga) Soutpansberg lavas (Barker et al., 2006), Waterberg-hosted sills (Hanson et al., 2004) and mafic dykes and sills across northern Kaapvaal and southern Zimbabwe Craton (Olsson, unpublished results: Söderlund et al., 2010; Klausen et al., 2010), closely followed by the ca. 1.8 Ga Eburnean orogeny. The timing of collision of the Zimbabwe and Kaapvaal Cratons along the Limpopo Belt is constrained by this magmatic record (Söderlund et al., 2010). Specifically, the apparent absence of 2.06 Ga Bushveld magmatism in the Zimbabwe Craton and the corresponding absence of 2.58 Ga Great Dyke of Zimbabwe magmatism in the Kaapvaal Craton, suggest that the Zimbabwe and Kaapvaal Cratons were still separate at these times. Furthermore, since ca. 1.90 Ga magmatism is present in both Zimbabwe and Kaapvaal Cratons, then collision and welding of these two cratons must have occurred between 1.90 Ga and 2.06 Ga.

The ca. 0.18 Ga Karoo Large Igneous Province is related to Gondwana break-up, where a vast band across South America, southern Africa and Antarctica was affected by voluminous magmatism.