Paleomagnetism of 1.79 Ga Pará de Minas mafic dykes: Testing a São Francisco/Congo-North China-Rio de la Plata connection in Columbia


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Contents lists available at ScienceDirect
Precambrian Research

journal homepage: www.elsevier.com/locate/precamres

ARTICLE INFO

Keywords:
Paleomagnetism
Pará de Minas dykes
Amazonian Craton
Columbia/Nuna

ABSTRACT

Paleogeographic connections between São Francisco/Congo, North China and Río de la Plata Cratons through the Paleo- to Mesoproterozoic have been proposed on the basis of geological and paleomagnetic data. We conducted new paleomagnetic analyses for mafic dykes of the 1790 Ma Pará de Minas swarm, occurring in the southern São Francisco Craton. The data revealed south/southwestern, low inclination magnetic directions after alternating field (AF) and thermal demagnetization, providing a paleomagnetic pole at 39.8°S, 196.8°E (A95 = 17.0°). The characteristic remanent magnetization is interpreted as a thermo-remanent magnetization carried by stable ‘pseudo-single domain’ magnetite grains, being acquired during the cooling of the dykes at 1790 Ma, as attested by a positive baked contact test performed on one of the dykes. The new pole in conjunction with 1790–1750 Ma paleomagnetic poles available for other blocks allowed an improved reconstruction of Columbia/Nuna supercontinent at this time frame. Laurentia, Baltica, Siberia, proto-Amazonia, West Africa and proto-Australia were positioned as in previously proposed models. The main difference from our reconstruction concerns the link of the São Francisco/Congo, North China and Río de la Plata cratons, implying that much of the Columbia supercontinent had already agglutinated around 1850–1800 Ma ago. This configuration is consistent with the onset of eventual taphrogenic episodes over most blocks, accompanied by the emplacement of mafic dyke swarms and granitic intrusions at 1790–1750 Ma.

1. Introduction

Since the first attempts to reconstruct its paleogeography almost two decades ago (Rogers and Santosh, 2002; Zhao et al., 2002; Meert, 2002), several models of the Paleo- to Mesoproterozoic Columbia (Nuna) supercontinent were proposed in the literature, mainly on the basis of paleomagnetic constraints and geological connections, including LIP barcode matches (e.g., Zhao et al., 2004; Pesonen et al., 2012; Johansson, 2009; Evans and Mitchell, 2011; Zhang et al., 2012; Bispo-Santos et al., 2014; Pisarevsky et al., 2014; Xu et al., 2014; Pehrsson et al., 2016; Salminen et al., 2016; 2017; Meert and Santosh, 2017, and references therein). However, no consensus has been achieved either on the paleogeography itself or on the time of continent’s maximum agglutination. For instance, several authors consider that Columbia assembled during the 1900–1800 Ma global peak in accretionary orogens around the Earth (e.g., Rogers and Santosh, 2002; Zhao et al., 2002, 2004; Evans and Mitchell, 2011; Yakubchuk, 2010), but others postulate that it had a much later assembly, around 1650–1580 Ma (e.g., Pisarevsky et al., 2014). The scarcity of reliable well-dated paleomagnetic poles from key units that formed Columbia and the limitations of the paleomagnetic method render this issue a particularly difficult task.

The São Francisco Craton (SFC) in eastern South America and its African counterpart, the Congo Craton, are an example where only few paleomagnetic data are available for the Paleo- to Mesoproterozoic interval (see review in D’Agrèlla-Filho and Cordan, 2017). The SFC and the Congo Craton share a common Paleo- to Mesoproterozoic evolution, indicated by the coherent evolution of their accretionary and collisional belts adjoining Archean blocks that eventually formed the São Francisco-Congo paleocontinent at ca. 2.0 Ga (Teixeira et al., 2017a and references therein). In the context of Columbia, the SFC-Congo paleocontinent was often considered as the ancient nucleus of the Atlantica continent (Rogers, 1996; Zhao et al., 2002; Rogers and Santosh, 2002; Yakubchuk, 2010). More recent reconstructions of the SFC-Congo in Paleo- to Mesoproterozoic times have tentatively linked this unit to North China and also to other blocks, including Siberia, Baltica and Río...
de la Plata (e.g., Ernst et al., 2013; Pisarevsky et al., 2014; Peng, 2015; Cederberg et al., 2016; Salminen et al., 2016; Girelli et al., 2018; Chaves et al., 2019). These reconstructions are based mainly on barcode matches of dyke swarms, roughly coeval tectonic basins and secondarily by paleomagnetic evidence.

Teixeira et al. (2017b) considered a connection between São Francisco-Congo paleocontinent and the North China Craton mainly based on the broadly correlative evolution among the ultimate stages of the 2.47–2.0 Ga Minas orogeny (southern SFC) and the Jiao-Liao-Ji belt, active from 2.20 to 1.85 Ga, along the eastern North China Craton, including the occurrences of khondalite rocks in both blocks. A different scenario for the SFC-Congo and North China was proposed by Peng et al. (2011) and Peng (2015) in which instead of the Minas orogen, the Itabuna-Salvador-Curacá orogen (northern SFC) was connected with the Jiao-Liao-Ji Orogen (North China Craton).

Alternatively the late Paleoproterozoic evolution of the São Francisco-Congo paleocontinent allows a correlation with the Trans-North Orogen of the North China Craton which is considered as the result of the collision between the Eastern and Western blocks by ~1.85–1.80 Ga ago, forming a unified continental lithosphere (e.g., Wang et al., 2010; Liu et al., 2012; Zhang et al., 2014; Peng et al., 2012, 2014; Wang et al., 2016; Liu et al., 2012; Guo et al., 2017). The khondalite belt/Fengzhen belt (~1.95–1.93 Ga) in the Western North China Craton is another key marker of such collisional dynamics. In a recent paper, Xu et al. (2017) used anisotropy of magnetic susceptibility (AMS) analysis of the 1.78 Ga Xiong’er volcanic rocks (LIP-like) to propose a connection of North China with SFC-Congo, Siberia and Rio de la Plata cratons at this time.

Here we report a new paleomagnetic pole for NW-trending Pará de Minas mafic dyke swarm that occurs in the southern SFC (Chaves, 2001; Chaves and Corrêa Neves 2005). This swarm was first paleomagnetically studied in the 1990’s (D’Agrella-Filho, 1992), the results disclosing a southwestern direction with low inclination for some of the analyzed samples. However, poor geochronological control and evidence of distinct generations of dykes in the region of Pará de Minas-Lavras (Chaves, 2013, Pines, 1997) hampered a robust age assignment of the pole. Recently, new U-Pb (on zircon) dating were performed on distinct dykes from the Pará de Minas region, which yielded similar crystallization ages at ca. 1795 Ma for three dykes, among other younger ages for some nearby dykes (Cederberg et al., 2016). In face of the new high-quality geochronological data, we resampled the Pará de Minas dykes that yielded ages at 1795 Ma in order to confirm the previous paleomagnetic results and test its robustness with a baked contact test. We complemented the new sampling with the analysis of previous collected blocks using more modern paleomagnetic techniques. Our combined dataset together with the available paleomagnetic poles for other cratonic units allowed new insights on the reconstruction of the Columbia supercontinent and the test of previous paleogeographic scenarios for 1.79 Ga.

2. Geological setting

The SFC-Congo Craton hosts two segments of an accretionary-collisional orogen (e.g., the Minas orogeny) that encompass the Archean continental blocks (Fig. 1). This polycyclic framework attained tectonic stability at ca. 2.0 Ga, giving rise to the São Francisco-Congo paleocontinent (Teixeira et al., 2017a,b; Alkmim and Teixeira, 2017). Shortly after, it underwent intraplate tectonics such as the Espinhaço rift system, as well as recurrent igneous episodes highlighted by mafic dike swarms (e.g., Silva et al., 1995; Chaves, 2013; Oliveira et al., 2013; Cederberg et al., 2016; Chaves et al., 2019). In the Neoproterozoic the Archean sialic remnants and Paleoproterozoic rocks akin to the proto-SFC were dispersed and later reassembled and reworked along the Brasiliano belts that surround the SFC. The occurrences of basement rocks in northern portion of the Tocantins province as well as those within the Neoproterozoic orogenic framework (Fuck et al., 2014; Cuadros et al., 2017; Teixeira et al., 2017a) suggest a much larger extent of the São Francisco-Congo paleocontinent at ca. 2.0–1.9 Ga than at the time of the assembly of the Western Gondwana.

In the central and northern SFC (Fig. 1B), Archean blocks (Gavião, Jequió and Serrinha) occur adjacent to the Paleoproterozoic Itabuna /Salvador/Curacá (ISC) belt, also termed Eastern Bahia orogenic domain, which is composed of tonalitic, trondhjemitic and charnockitic rocks and associated arc-type volcano-sedimentary sequences (e.g., Barbosa and Sabaté, 2002; Oliveira et al., 2004; Barbosa and Barbosa, 2017). The collisional orogeny developed during the transition between the Rhyacian and Orosirian periods, and eventually assembled at ca. 2.0 Ga through scarp tectonics of the intervening Archean blocks, including those in the African counterparts (Barbosa and Barbosa, 2017 and references therein). The evolution of the Eastern Bahia orogen is roughly coeval with the 2.47–2.00 Ga Minas Orogen (Fig. 1C) in the southern SFC (Alkmim and Teixeira, 2017 and references therein). The Minas orogeny produced a series of accretionary-collisional belts composed of plutonic rocks and volcanic-sedimentary rocks, namely the Mineiro, Mantiqueira and Juiz de Fora belts (e.g., Ávila et al., 2014; Teixeira et al., 2015; 2017a,b; Noce et al., 2007; Heilbronn et al., 2010; Barbosa et al., 2019). Crustal exhumation and final regional cooling of the gneissic-granitic complexes in the southern SFC occurred between ca. 2.1–1.9 Ga (Teixeira et al., 2000, 2017a).

Distinct generations of Proterozoic mafic dyke swarms transect the crystalline basement of the SFC, ranging in age from the Late Archean (e.g., Uauá dykes) to the Early Neoproterozoic (e.g., Ilhéus-Olivença and Diamantina dykes; Evans et al., 2016; Chaves et al., 2019). They witnessed extensional episodes through time and space (Girardi et al., 2017 and references therein), such as the Pará de Minas swarm. Some dykes crosscut the Espinhaço intracratonic rift system (Chemal, Jr. et al. 2012; Guadagnin and Chemal, 2015).

The NW-SE-trending Pará de Minas dykes (Chaves, 2001) is the most prominent dyke swarm in the southern part of the SFC, where other swarms with distinct ages and directions are also present (e.g., Girardi et al., 2017 and references therein). Most swarms transect the Archean granite-greenstone basement and the Early Paleoproterozoic Minas Supergroup rocks, occurring to the north of the Bom Sucesso-Jequiabá lineament (Fig. 1C, Neri et al., 2013; Moreno et al., 2017; Teixeira et al., 2017a). From a tectonic point of view the Pará de Minas dyke swarm is an anorogenic episode that occurred after the tectonic stability of the adjoining Mineiro belt. In a broader sense this mafic magmatism unravels a breakup attempt of the thick lithosphere of the São Francisco-Congo paleocontinent. This episode is also coeval with the intraplate basal volcanics in the Espinhaço Supergroup (e.g., Chemal, Jr. et al., 2012 and references therein).

The Pará de Minas dykes are composed of fine- to coarse grained diabases to gabbro of tholeiitic affinity (Girardi et al., 2017 and references therein). Most dykes are massive, but porphyritic types also occur, being characterized by plagioclase phenocrysts parallel to their dyke walls. Anisotropy of magnetic susceptibility (AMS) measurements on the Pará de Minas dykes indicated sub-horizontal magma fluxes from NW to SE, in accordance with the field observations (Raposo et al., 2004; Chaves, 2014). The dykes of the Pará de Minas swarm can be up to 100 m thick and up to 150 km long (Chaves, 2013; Cederberg et al., 2016). They show ophitic to sub-ophitic textures and usually crop out as pristine rocks. Petrographically these rocks are essentially composed of labradorite-andesine and augite, with minor amounts of interstitial K-feldspar, quartz, apatite, zircon or baddeleyite, and opaque minerals. In some outcrops the dykes are crosscut by narrow fractures at the margins of which fine-grained chlorite, quartz, clay minerals and sericite are the most common alteration phases. Sometimes augite is partially replaced by tremolite-actinolite (Chaves, 2013; Chaves and Corrêa Neves, 2005; Girardi et al., 2017).

Until recently the age of this dyke swarm was constrained only by a 1740 ± 54 Ma Rb-Sr whole rock isochron date, obtained from samples of six nearby dykes in the study region (Chaves, 2014), as well as by the
1714 ± 5 Ma U-Pb age for the Ibirité Gabbro occurring in the Quadrilátero Ferrífero (Silva et al., 1995). New and more precise dates were obtained recently in six dykes using the U-Pb method in baddeleyite (Cederberg et al., 2016). Three dykes yielded identical (within error) U-Pb crystallization ages of 1798 ± 4 Ma, 1791 ± 7 Ma and 1793 ± 18 Ma, respectively, for the MG-3, MG-5 and MG-6 samples (see Figs. 3, 5, 6 of Cederberg et al., 2016), defining for the first time the precise time of emplacement of the Pará de Minas swarm. Other two

Fig. 1. (A) São Francisco and Congo cratons in the Gondwana configuration, and surrounding Neoproterozoic belts (South America in its present position); (B) Paleoproterozoic orogenic framework of the northern São Francisco Craton (SFC), showing the Itabuna-Salvador-Curaçá belt (also known as the Eastern Bahia orogenic domain); PA – Paramirim corridor. (C) – Paleoproterozoic orogenic framework of the southern SFC showing the Mineiro, Mantiqueira and Juiz de Fora belts (Modified after Teixeira et al., 2017b). Keys: RV = Neoarchean Rio das Velhas Supergroup; SG = Sabará Group (Early Paleoproterozoic Minas Supergroup; It = Itacolomi Group (< 1.96 Ga); CSZ = Cláudio Shear Zone; QF = Quadrilátero Ferrífero. Open circles indicate the studied sites: 1 to 11. * - U-Pb (baddeleyite) dated sites (Cederberg et al., 2016). Towns: CB (Campo Belo), IT (Itapecerica), CL (Cláudio), PE (Perdões), OL (Oliveira), PT (Passa Tempo), BS (Bom Sucesso), BH (Belo Horizonte), PM (Pará de Minas), OP (Ouro Preto), RC (Resende Costa), SJ (São João del Rei), CO (Conselheiro Lafaiete). See text for details.
samples (MG-4, MG-7) gave roughly similar crystallization ages of 1702 ± 13 Ma and 1717 ± 11 Ma, whereas the MG-8 sample yielded 766 ± 36 Ma. Collectively these precise U–Pb ages confirmed that more than one generation of mafic dykes occur in the southern part of the SFC. Only the oldest ages are considered here as representing the Pará de Minas swarm and used as a time constraint for our paleomagnetic investigation.

3. Sampling and analytical methods

Fourty-three oriented cylindrical cores were extracted from four dykes of the Pará de Minas swarm, using a gasoline-powered drill. Both, magnetic and sun compasses were used for orientation of the samples. Three of these dykes (sites 1, 3 and 4 in Table 1 and Fig. 1C) correspond, respectively, to the MG3, MG6 and MG5 samples dated at ca. 1790 Ma by Cederberg and colleagues (see above). The fourth sampled outcrop (site 2 in Table 1 and Fig. 1C) is from a ca. 20 m thick dyke (undated) cutting the Archean gneissic basement. Eight cylindrical cores were extracted from the dyke itself and from the country rock at distances of 0.6 m, 1.10 m, 10 m, and 15 m away from the dyke for a baked contact test. In this outcrop, the sharp contact between the dyke and country rocks was under water and then it could not be sampled. In addition, we reanalyzed samples from seven other dykes (Fig. 1C and Table 1) collected in the previous paleomagnetic study (D’Agrella-Filho, 1992). Normally, three to four hand samples were sampled from each dyke at that time, which were oriented by magnetic and solar compasses. Only at two sites (2 and 10), field observation permitted to infer that the dykes are sub-vertical.

The new sampled cores and the cylinders extracted from the ancient block samples were cut in cylindrical specimens of 2.2 × 2.5 cm, which were then submitted to stepwise alternating field (AF) and thermal treatments to isolate the characteristic remanent magnetization (ChRM). Steps of 2.0 mT, 2.5 mT, 5 mT or 10 mT were employed for AF demagnetization using a 2-axes tumbler AGICO AF demagnetizer. An ASC dual-chamber furnace model TD48 was used for thermal demagnetization in steps of 50 °C (from 100 °C to 500 °C) and 20 °C (above 500 °C). Remanent magnetization measurements were performed using an AGICO JR-6 spinner magnetometer. Some samples with low magnetic intensities were demagnetized in the same steps using an automated three-axis AF demagnetizer coupled to a 2G Enterprises model 755U cryogenic magnetometer. All equipments are housed in a magnetically shielded room with ambient field < 500 nT. Magnetic components were identified in orthogonal plots (Zijderveld, 1967) and calculated by principal component analysis (Kirshvink, 1980) using the Remasoft-3.0 program from AGICO (Chadima and Hrouda, 2006). At least four successive demagnetization steps were used to calculate vectors using least-squares fits, and an upper limit for mean angular deviation (MAD) of 12° was applied. Fisher’s (1953) statistics was used to calculate vector mean directions and the corresponding paleomagnetic pole.

The magnetic mineralogy was investigated by curves of isothermal remanent magnetization (IRM). Low-field magnetic susceptibility versus temperature curves were performed from −194 °C to 700 °C using a CS-4 apparatus coupled to a KLY-4S Kappabridge instrument (AGICO). Hysteresis and IRM acquisition curves were performed with a MicroMag vibrating sample magnetometer model 3900 (Princeton Measurements Corporation). Bulk coercive force (Hc), saturation magnetization (Ms), and saturation remanent magnetization (Mr,50, 500) were obtained after subtraction of the paramagnetic contribution from the high field portion of hysteresis curves. The coercivity of remanence (Hr) was obtained by applying increasing reverse fields after the sample was submitted to a field of 1 T.

4. Paleomagnetic results

Natural remanent magnetization (NRM) of the new samples varies from 13.8 to 26.3 nT/m to up to 333 A/m. The extremely high intensities were all found in site MG4. This site had the greatest differences in orientation from solar and magnetic compasses, presented a fast decay of the natural remanent magnetization during demagnetization and showed random paleomagnetic directions. All this suggests this outcrop was affected by lightning strikes (e.g. Salminen et al., 2013). NRM of block samples vary between 6.9 and 9.5 A/m, being therefore consistent with those of the new samples.

The new sites 1, 2 and 3 yielded S/SW directions with low inclinations (Fig. 2a–c) after elimination of a low-coercivity or low-unblocking temperature secondary component during AF and thermal demagnetization, respectively. Similar directions were disclosed for the block samples (Fig. 2d–f).

Site mean directions are shown in Fig. 3a with a cluster around the mean of D = 222.4°, I = −13.1° (N = 10, α95 = 18.8°, k = 7.6),
Fig. 2. Examples of alternating field (a, b, d) and thermal (c, e, f) demagnetization. For each sample, the stereographic projection, orthogonal (Zijderveld) projection and the normalized intensity curve (M/Max versus alternating field (H) or temperature) are shown.
Fig. 3. (a) site mean directions for the Pará de Minas dykes (the mean (in red) with its confidence circle (α95) is also shown); (b) specimen magnetization directions isolated for the gneissic basement rocks (the mean (in red) with its confidence circle (α95) is also shown). Symbols ⧫ and ⧫ represent the present geomagnetic and dipolar fields, respectively. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Fig. 4. Baked contact test (Site 2): examples of AF and thermal demagnetization for samples from the dyke (a and b); samples at 0.6 m from the contact (c and d); and sample at 15 m from the contact (d). For each sample, the stereographic projection, the normalized intensity curve (M/Max versus alternating field (H) or temperature), and orthogonal (Zijderveld) projection are shown.
Fig. 5. Examples of low- (left) and high-temperature (right) thermomagnetic curves (magnetic susceptibility versus temperature). Heating in red and cooling in blue. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
Fig. 6. Examples of Isothermal remanent magnetization (IRM) acquisition curves (induced magnetic moment versus magnetic field).
which is here named as component A. Based on this component we calculated a paleomagnetic pole for the Pará de Minas dykes (PM pole) at 39.8°S, 196.8°E (N = 10, A95 = 17.0°, K = 9.0).

4.1. Baked contact test

The baked contact test was performed for site 2 where the gneissic host rock was cored at 0.6 m, 1.10 m, 10 m, and 15 m away from the contact of a 20 m-thick dyke. Samples from the dyke itself revealed the characteristic direction of magnetization with low-inclination and SW trend (Fig. 4a, b). Two farthest samples, at 10 m (DMG2K) and 15 m (DMG2L) from the contact, respectively, yielded after AF treatment a distinct but coherent northeast direction with an upward inclination (named here as component B) (Fig. 4e). The same component B is shown in Fig. 3b for other samples from the crystalline basement (DMG2K and DMG2L; Table 1).

In Fig. 4, two samples were chosen to represent the contact zone (at 0.6 m) in the host rock. A representative example of AF demagnetization is shown in Fig. 4c (sample DMG2I1A). At low coercivity (0–4 mT), a SW-trending component was isolated that matches the characteristic remanent magnetization extracted from the dyke. At higher coercivities, a southeast, upward inclination direction was isolated, which is located between components A and B. The thermal treatment of the sister sample DMG2I3 (Fig. 4d) also provided a south-southwest direction with a low positive inclination (component A) carried by magnetic grains with high unblocking temperatures (560–600 °C) and an intermediate component between A and B. Sample DMG2J (not shown), located at 1.1 m from the contact, yielded inconsistent results.

4.2. Magnetic mineralogy

Magnetic properties were characterized by hysteresis, IRM and thermomagnetic curves. In thermomagnetic curves, most samples show the Verwey transition at −153°C, which is typical of magnetite (Fig. 5a, b). Reversible and irreversible high-temperature thermomagnetic curves were obtained at high-temperature and are exemplified in Fig. 5. Samples are characterized by a pronounced Hopkinson peak just before the decay at around 580 °C (Fig. 5a, b), which is also characteristic of fine-grained magnetite (Dunlop and Özdemir, 1997). In contrast, samples from site 9 (block samples DMG75, DMG76 and DMG77) show thermomagnetic curves with a strong paramagnetic signal characterized by a well-defined hyperbola at lower temperatures, superimposed to the susceptibility decay typical of magnetite grains at around 580 °C (Fig. 5c).

IRM curves are also typical of low-coercivity magnetite (Fig. 6). Most samples saturated at fields below 100 mT. The exception was again the sample from site 9 (sample DMG75E in Fig. 6), for which saturation occurred at higher fields, around 400 mT and may thus have some contribution of higher coercivity minerals such as hematite or goethite. This information was complemented by hysteresis loops performed for one sample from each dyke. Typical examples are shown in Fig. 7. Coercivities are comprised between 4.4 mT and 31.1 mT, which are also typical of magnetite. $M_r/M_s$ versus $H_r/H_c$ ratios were plotted in a Day’s plot (Fig. 8). Most samples fall along a trend parallel to the theoretical single domain/domain (SD/MD) mixing curves of Dunlop (2002) and fall in the pseudo-single domain field of the Day diagram (Day et al., 1977). Two samples (DMG2L and DMG2K), however, plotted in the multidomain (MD) field. These samples correspond to the country rock samples collected the farthest from the dyke in the baked contact test.

5. Discussion

5.1. Significance of the baked contact test

The dyke itself yielded a S-SW low-inclination characteristic remanence direction similar to that isolated for the other studied dykes. A different E-NE magnetic direction (component B) with moderate negative inclination was found for the two samples collected the farthest from the contact (at 10 and 15 m), which is different from the recent geomagnetic field and from the characteristic component of the dykes (Fig. 3). Sample from host rock collected close to the dyke (~0.6 m away from the contact) present a multi-component behavior. A direction similar to the dyke (component A) was isolated using either AF or thermal treatment (Fig. 4c and d). However, the component B characteristic of the host rocks was not completely isolated in these samples, with the AF treatment showing an evolution along a great circle that encompasses the component B. This behavior occurs when two magnetic components present coercivity spectra overlap (Halls, 1978). One possible explanation is that as a result of the thermal effect of the dyke, new magnetic minerals grew in the sample at 0.6 m whose
Most samples fall in the Pseudo-single domain (PSD) proposed by Dunlop (2002). Percentages of MD grains in the mixture are also shown. Gneissic samples DMG2L and DMG2K collected at 10 m and 15 m from the contact test are positive, that is, component A disclosed for the Pará de Minas swarm. Therefor, the PM pole can be used to infer the paleogeographic position of the SFC around 1790 Ma ago. Therefore, the new 1790 Ma PM pole determined here is an excellent opportunity to test these reconstructions, since the 1790–1780 Ma North China dykes were paleomagnetically studied by several previous studies (e.g., Peng et al., 2015; Cederberg et al., 2016; Teixeira et al., 2017b; Xu et al., 2017; Girelli et al., 2018). However, different relative positions between these two blocks in the context of Columbia have been proposed, some of them including the Rio de la Plata Craton as a nearest block (Peng, 2015; Cederberg et al., 2016; Xu et al., 2017; Girelli et al., 2018).

In summary, although the PM pole satisfies only four out of the seven Van der Voo’s criteria, it fulfills the two most important criteria: (i) the PM pole is well dated at 1794 Ma (Cederberg et al., 2016), and (iv) a positive baked contact test attests to the primary nature of component A.

5.3. Sào Francisco-Congo and North China connection

Teixeira et al. (2017b) documented the geologic-tectonic matches between the Minas orogeny and the Jiao-Liao-Ji Orogen located in North China Craton. The progressive closure of an intervening Paleo-proterozoic ocean eventually built the connection between the SFC-Congo and the North China Craton at ca. 2.0 Ga. In addition, widespread intraplate magmatic activity within the São Francisco-Congo paleocontinent at ca. 1.79–1.70 Ga match the ca. 1.78 Ga Xiong’er LIP which is located in the southern edge of the North China Craton (e.g., Peng et al., 2011; Peng, 2015; Cederberg et al., 2016; Xu et al., 2017; Girelli et al., 2018). However, different relative positions between these two blocks in the context of China have been proposed, some of them including the Rio de la Plata Craton as a nearest block (Peng, 2015; Cederberg et al., 2016; Teixeira et al., 2017b; Xu et al., 2017; Girelli et al., 2018).

Therefore, the new 1790 Ma PM pole determined here is an excellent opportunity to test these reconstructions, since the 1790–1780 Ma North China dykes were paleomagnetically studied by collection of their remanence carriers and testing the possible paleocontinental reconstructions. The new PM pole (39.8°S, 196.8°E) shows a Fisher’s statistical parameters of χ95 = 17° and κ = 9.0. These parameters did not satisfy the second criterion of Van der Voo (1990), which establishes χ95 < 16° and κ > 10.0. Also, the precision parameter (κ = 9.0) of the PM pole implies in an angular dispersion (σ) of 27.0° for the paleolatitude (λ) of 6.6°, calculated using the Pará de Minas mean magnetic inclination I = −13.1°. A much lower value for σ, of ca. 11°, for the paleolatitude of ~7° is predicted from Proterozoic models of secular variation (Smirnov et al., 2011; Veikolkainen and Pesonen, 2014). An increase in the dispersion of directions could be related to the low geomagnetic dipolar field at 1800 Ma when relatively enhanced non-dipole fields could prevail (Smirnov, 2017; Biggin et al., 2015; Kirschker et al., 2019).

(iii) Least-squares fit from orthogonal diagrams (Principal Component Analysis; Kirshvink, 1980) were used to calculate ChRM components after AF and thermal demagnetization;

(iv) A positive baked contact test was obtained for site 2, attesting the primary nature of component A;

(v) The studied dykes are not metamorphosed or deformed and preserve original textures (Chaves, 2013; Chaves and Corrêa Neves, 2005; Cederberg et al., 2016; Girardi et al., 2017). We note that one dyke in the southern SFC yielded a Neoproterozoic age, which may arise some speculation about tectonic movements in the studied area after the main Pará de Minas dyke swarm was intruded (e.g., Teixeira et al., 2000). But the coherence of dyke’s trends, their vertical position across the whole region and the coherence of magnetic directions between the different studied dykes suggest that if these movements did occur, they were minor and do not affect significantly the paleomagnetic results;

(vi) The analyzed sites disclosed only one polarity. However, the dispersion of site mean directions suggests they average out the secular variation of the geomagnetic field;

(vii) The PM pole is close to recent poles (50–0 Ma) for South America (Torsvik et al., 2012). However, the positive baked contact test obtained for site 2 proves the Pará de Minas component A is of primary origin.

In summary, although the PM pole satisfies only four out of the seven Van der Voo’s criteria, it fulfills the two most important criteria: (i) the PM pole is well dated at 1794 Ma (Cederberg et al., 2016), and (iv) a positive baked contact test attests to the primary nature of the characteristic remanence carried by the Pará de Minas dykes. Therefore, the PM pole can be used to infer the paleogeographic position of the SFC around 1790 Ma ago.

5.2. The Pará de Minas pole

The new Pará de Minas (PM) pole (196.8°E, 38.8°S, χ95 = 17°, κ = 9.0) satisfies four out of the seven quality criteria proposed by Van der Voo (1990):

(i) The documented U-Pb ages for the Pará de Minas swarm yielded very consistent ages of 1798 ± 4 Ma, 1791 ± 7 Ma and 1793 ± 18 Ma (see above), which likely represents the age of acquisition of the remanent magnetization;

(ii) Component A was determined using 104 specimens from 10 sites. The results yielded the paleomagnetic pole PM (39.8°S, 196.8°E) which shows the Fisher’s statistical parameters of χ95 = 17° and κ = 9.0. These parameters did not satisfy the second criterion of Van der Voo (1990), which establishes χ95 < 16° and κ > 10.0. Also, the precision parameter (κ = 9.0) of the PM pole implies in an angular dispersion (σ) of 27.0° for the paleolatitude (λ) of 6.6°, calculated using the Pará de Minas mean magnetic inclination I = −13.1°. A much lower value for σ, of ca. 11°, for the paleolatitude of ~7° is predicted from Proterozoic models of secular variation (Smirnov et al., 2011; Veikolkainen and Pesonen, 2014). An increase in the dispersion of directions could be related to the low geomagnetic dipolar field at 1800 Ma when relatively enhanced non-dipole fields could prevail (Smirnov, 2017; Biggin et al., 2015; Kirschker et al., 2019).

(iii) Least-squares fit from orthogonal diagrams (Principal Component Analysis; Kirshvink, 1980) were used to calculate ChRM components after AF and thermal demagnetization;

(iv) A positive baked contact test was obtained for site 2, attesting the primary nature of component A;

(v) The studied dykes are not metamorphosed or deformed and preserve original textures (Chaves, 2013; Chaves and Corrêa Neves, 2005; Cederberg et al., 2016; Girardi et al., 2017). We note that one dyke in the southern SFC yielded a Neoproterozoic age, which may arise some speculation about tectonic movements in the studied area after the main Pará de Minas dyke swarm was intruded (e.g., Teixeira et al., 2000). But the coherence of dyke’s trends, their vertical position across the whole region and the coherence of magnetic directions between the different studied dykes suggest that if these movements did occur, they were minor and do not affect significantly the paleomagnetic results;

(vi) The analyzed sites disclosed only one polarity. However, the dispersion of site mean directions suggests they average out the secular variation of the geomagnetic field;

(vii) The PM pole is close to recent poles (50–0 Ma) for South America (Torsvik et al., 2012). However, the positive baked contact test obtained for site 2 proves the Pará de Minas component A is of primary origin.

In summary, although the PM pole satisfies only four out of the seven Van der Voo’s criteria, it fulfills the two most important criteria: (i) the PM pole is well dated at 1794 Ma (Cederberg et al., 2016), and (iv) a positive baked contact test attests to the primary nature of the characteristic remanence carried by the Pará de Minas dykes. Therefore, the PM pole can be used to infer the paleogeographic position of the SFC around 1790 Ma ago.
several authors (e.g., Halls et al., 2000; Piper et al., 2011; Zhang et al., 2012; Xu et al., 2014). For this purpose, we present two alternative paleogeographies for the SFC-Congo and North China, based on the 1790–1780 Ma paleomagnetic poles.

Fig. 9A presents a first model assuming a connection between North China (NC) and SFC-Congo at 1790 Ma, where the Mineiro and Jiao-Liao-Ji belts face each other (slightly modified from Teixeira et al., 2017b). The Xiong’er plume center (red star), the 1790–1780 Ma Taihang-Yinshan dyke swarm in North China and the Pará de Minas dyke swarm in SFC are also shown. Note that the Pará de Minas dykes strike almost perpendicular to the Xiong’er magma source in this model (Fig. 9A).

Fig. 9B shows an alternative configuration (preferred) for the SFC-Congo and North China, considering the 1790 Ma anti-pole of North China as having the same polarity of the Pará de Minas (PM) pole. In this model the Xiong’er LIP event can be considered as the plume source of the Pará de Minas dykes, as already suggested by Peng (2015). Also, the position of the Xiong’er magma source to the W/NW of the Pará de Minas dykes in Fig. 9B is reinforced by the AMS measurements and field observations that indicate these dykes were mainly fed by sub-horizonal magma fluxes from NW to SE (in their present position) (Raposo et al., 2004; Chaves, 2014).

Although other configurations of SFC-Congo and North China may also explain the 1790 Ma Xiong’er plume in a radiating geometry of the roughly coeval dyke swarms in these two blocks (e.g. Peng, 2015; Cederberg et al., 2016; Xu et al., 2017), our preferred reconstruction (Fig. 9B) is the only one that is also consistent with paleomagnetic constraints. In the following topic we will expand our alternative (and paleomagnetically consistent) scenario to include Rio de la Plata and India as the other blocks containing the radiating dyke swarms that could be genetically related with the Xiong’er plume center.

5.4. Supercontinent Columbia at ca. 1790–1750 Ma

Paleogeographic configurations of Columbia have been tested by means of paleomagnetic poles and geological evidence by many authors (e.g., Evans and Mitchell, 2011; Zhang et al., 2012; Xu et al., 2014; Pisarevsky et al., 2014). Fig. 10 shows a possible Columbia reconstruction with the respective 1790–1750 Ma poles. The Euler rotation poles for each block (and respective paleopoles) are presented in Table 2 and in the legend of Fig. 10.

Laurentia, Baltica, Siberia, proto-Amazonia, West Africa and proto-Australia were placed according to the reconstructions of Evans and Mitchell (2011); Zhang et al. (2012); Bispo-Santos et al. (2014) and Xu et al. (2014), among others. Baltica-Amazonia link follows the SAMBA model of Johansson (2009) whereas the position of proto-Australia is from the model of Xu et al. (2014). The relative position of North China and SFC-Congo is the same as in Fig. 9B. India is positioned to the south of the block formed by North China and SFC-Congo, considering the available geological and paleomagnetic evidence (Zhao et al., 2002, 2003, 2004; Zhang et al., 2012; Xu et al., 2014; Shankar et al., 2018). According to these works, the Paleoproterozoic Trans-North China...
(TNC) belt would be correlative to the Central Indian Tectonic Zone (CITZ) (Fig. 10). Recently, geophysical data suggested that North China and India formed a single block during the Paleoaneropaleozoic (Cao et al., 2019). In Fig. 10 the (present) northern limit of the Rio de la Plata Craton is connected to the (present) south-western limit of North China, whereas the (present) northwest portion of north Australia is linked to the Rio de la Plata and the São Francisco-Congo paleocontinent.

Magnetic events (most of them dyke swarms) dated between 1790 and 1750 Ma are also shown in Fig. 10 (denoted by numbers 1 to 6), assuming the Xiong'er LIP as the magnetic center (in red). The geometry of the coeval dyke swarms of Pará de Minas (1) and Taingh-Yinshan (2) is consistent with such a hypothesis. Fig. 10 also shows the position of the 1790 Ma Florida dykes (3) of the Rio de la Plata Craton, the Hart dolerite (4) of proto-Australia, and the Pipilia (5) and Pebbair (6) dyke swarms in India (ca. 1765 Ma). The geometry of the latter two dyke swarms allows a relationship with the Xiong'er plume (Samal et al., 2019; Söderlund et al., 2019). For proto-Australia, the Hart dolerite (4) strikes almost perpendicular to the radial dyke geometry (Fig. 10), but their anisotropy of magnetic susceptibility (AMS) indicate a magma flux coming from present northwest (Kirscher et al., 2019), that is, in agreement with the proposed reconstruction. The Florida dykes do not strike in the direction of the Xiong'er magma source, but the available geological evidence is permissive with the paleogeography proposed in Fig. 10 (see below).

In defining the position of proto-Australia relative to Laurentia, Kirscher et al. (2019) used the 1790 Ma Hart AU1 pole from North Australia and the 1766 Ma Deschambault Pegmatites LA4 pole from Laurentia. We note that these poles are close together in our reconstruction (Fig. 10 and Table 2). However, no field tests were performed for the Deschambault Pegmatites, and its pole is very close to the present geomagnetic field. Besides, it falls far from the

Footnotes: Plate (Paleolatitude); Plong (Paleolongitude); Euler poles (used for each cratonic block); Ap (confidence circle - Fisher’s statistic parameters); Rlat (rotated latitude); Rlong (rotated longitude); Q (Quality factor, Van der Voo, 1990). Ref.: 1 - Bispo-Santos et al. (2014); 2 - Fedotova et al. (1999); 3 - Lubina et al. (2012); 4 - Pisarevsky and Sokolov (2001); 5 - Elming et al. (2009); 6 - Pisarevsky and Bylund (2010); 7 - Park et al. (1973); 8 - Gala et al. (1995); 9 - Irving et al. (2004); 10 - Symons et al. (2000); 11 - Shankar et al. (2018); 12 - Kirscher et al. (2019); 13 - Teixeira et al. (2013); 14 - This work; 15 - Xu et al. (2014); 16 - Halb et al. (2000); 17 - Piper et al. (2011); 18 - Zhang et al. (2012),

Table 2

Selected 1790–1750 Ma paleomagnetic poles from cratonic blocks that formed Columbia supercontinent.
Silvânia domains) in the Tocantins Province, probably represent an extension of terranes (Natividade, Almas-Conceição dos Tocantins, Cavalcante-Arraias domains), as well as with roughly coeval rocks occurring in the Silvânia region to the south within the Neoproterozoic Brasilia belt (e.g., Fuck et al., 2014; Sousa et al., 2016; Cuadros et al., 2017 and references therein). Altogether these basement rocks are relics of a much larger São Francisco-Congo paleocontinent that existed before the assembly of West Gondwana. Fig. 11 also shows the Mantiqueira and Juiz de Fora belts that are similarly considered to be relevant components of this paleocontinent, though deeply reworked in the Neoproterozoic.

The Paleoproterozoic orogenic framework that characterizes the southeast portion of the South America continent, such as the SFC, the Rio de la Plata craton (including the Florida dykes) and the Tocantins crystalline substratum is a result of a polycyclic history, arc-type that included successive juvenile growth combined with crustal reworking of the intervening ancient block. The main metamorphic phase peaked at ca. 2.1 Ga and eventual cratization occurred at ca. 1.90 (e.g., Oyhantçabal et al., 2018; Fuck et al., 2014; Sousa et al., 2016; Cuadros et al., 2017; Alkmim and Teixeira, 2017). Such an orogenic scenario, from a broader perspective, allows a suitable correlation with the Khondalite belt of North China, as outlined in Fig. 11. This particular Khondalite belt resulted from a long-lived Paleoproterozoic arc-continent accretion along the southern margin of the Yinshan Block culminating with the continent-continent collision of the Yinshan and Ordos blocks at ca. 1.95–1.90 Ga (Liu et al., 2017) at the time of the suturing of the Jiao-Liao-Ji belt along the Eastern North China block (Teixeira et al., 2017b and references therein).

We note that the cratonic rocks in South America and North China were crosscut at the 1790–1780 Ma by the Florida (Halls et al., 2001), Pará de Minas (Cederberg et al., 2016) and Taihang-Yinshan dyke swarms (fed by the Xiong’er rift system; Peng, 2015), respectively, representing the major intraplate features. Roughly contemporary intraplate magmatic events at 1790–1770 Ma, mostly of them represented by dyke swarms, are similarly documented in the Amazonian craton (AV, Avanavero LIP - Reis et al., 2013), Baltica (H, Hotting Gabbro, S, Småland dykes, T, Tomashgorod dykes - Elming et al., 2009; Pisarevsky and Bylund, 2010; Bogdanova et al., 2013, respectively), North Australia (Hart dolerite - Kirsch et al., 2019) and India (Pipilia and Pebbair dyke swarms - Samal et al., 2019; Söderlund et al., 2019; Srivastava et al., 2019) (see Fig. 10). Given the barcode matches, this fact indicates a worldwide scale phenomenon that succeeded tectonic stability of Paleoproterozoic continental lithosphere, implying that all these blocks were close neighbors at the time. Moreover, a connection around the Siberian rift system (black symbol in Fig. 10) which would be the central magma source of 1750 Ma radiating swarms seems to explain better the available data for Siberia, Laurentia and West Africa (Youbi et al., 2013), and possibly also for Baltica and proto-Amazonia. Based on the barcode match of four dyke swarms (at 1.87, 1.75, 1.35 and 0.72 Ga), Ernst et al. (2016) suggested a long-lived connection between the present southern Siberia and the present northern Laurentia since ca. 1.9 Ga (similar to that of Fig. 10).

6. Summary and conclusions

A paleomagnetic pole (PM – 196.8°E, 38.8’S, A95 = 17.0°, K = 9.0) was determined for the well-dated 1790 Ma Pará de Minas mafic dyke swarm in the southern SFC, for which a positive baked contact test attests to the primary nature of the magnetic direction of the dykes. Paleomagnetic poles for the time span 1790–1750 Ma are documented for many cratonic blocks in the world which permitted to reconstruct the paleogeography of the Columbia supercontinent at the time (Fig. 10). The working model considers that the position of Tandilia terranes (Oyhantçabal et al., 2018) and rocks from boreholes cored at latitude 31°S (Rapela et al., 2007) considered as the tentative border of the RPC, though extensively overlain by Phanerozoic cover, allow a plausible connection with the late Paleoproterozoic rocks of the Tocantins Province (Natividade, Almas-Conceição dos Tocantins, Cavalcante-Arraias domains), as well as with roughly coeval rocks occurring in the Silvânia region to the south within the Neoproterozoic Brasilia belt (e.g., Fuck et al., 2014; Sousa et al., 2016; Cuadros et al., 2017 and references therein). Altogether these basement rocks are relics of a much larger São Francisco-Congo paleocontinent that existed before the assembly of West Gondwana. Fig. 11 also shows the Mantiqueira and Juiz de Fora belts that are similarly considered to be relevant components of this paleocontinent, though deeply reworked in the Neoproterozoic.

Fig. 11. Zoom of Fig. 10 focusing 1790–1780 Ma magmatic events (mafic dyke swarms in black – described in the legend of Fig. 11) and related granitoid bodies (in red)) and Paleoproterozoic events (2.4–1.8 Ga) inside São Francisco Craton (SF): NAC – Natividade/Almas-Conceição dos Tocantins-Cavalcante-Arraias Terranes; S – Silvânia; M – Mineiro Belt; MA - Mantiqueira belt; JF – Juiz de Fora belt; CG – Correntina-Guamambê Terrane; IB – Itacambira-Barrocaí Terrane; ISC – Itabuna-Salvador-Curaçá belt. Congo Craton (C): K – Kamizian belt. North China Craton (NC) – Archean Blocks: OB – Ordos Block; YB – Yinshan Block; EB – East Block. Paleoproterozoic belts: TNC – Trans-North China belt; JL – Jiu-Lu-Liu-Ji belt; KB – Khondalite belt. Rio de la Plata Craton (LP): Ta – Tandilla Terrane, PA – Piedra Alta Terrane, BH – Boreholes (Rapela et al., 2007). India (IN, after Sahota and Mazumder, 2012); CITZ – Central Indian Tectonic Zone. North Australia (NAU, after Sheppard et al., 2012); LB – Lambro Province; HC – Halls Creek orogen. The dotted line comprising Paleoproterozoic terranes (Natividade, Almas-Conceição dos Tocantins, Cavalcante-Arraias and Silvânia domains) in the Tocantins Province, probably represent an extension of the São Francisco Craton at Paleoproterozoic times, according to Sousa et al. (2016 and references therein) (see details in the text). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

1790–1770 Ma poles from Baltica (BA1, BA2, BA3, BA4) in Fig. 10, which would imply in some distance of Baltica relative to Laurentia at ca. 1780 Ma. However, the remaining 1790–1750 Ma poles from Laurentia (LA1, LA2, LA3) are close to the 1790–1770 Ma poles for Baltica (BA1, BA2, BA3, BA4) (see Table 2 and Fig. 10). In any case, if the Hart dolerite (UA1) pole has an age of 1790 Ma, as argued by Kirsch et al. (2019), either the Laurentia and proto-Australia link (Payne et al., 2009; Evans and Mitchell, 2011; Zhang et al., 2012; Xu et al., 2014) or the Baltica-Laurentia link (e.g., Salminen and Pesonen, 2007; Pisarevsky and Bylund, 2010; Pesonen et al., 2012; Pisarevsky et al., 2014; this paper) must be revised.

The paleomagnetic constraints for the Rio de la Plata craton (RPC) in our reconstruction rely on the paleopole of Florida dykes (Teixeira et al., 2013) and the geological similarities including barcode matches between this craton and the SFC-Congo and North China (see Fig. 11), as envisaged by Peng (2015) and Girelli et al. (2018) among others (e.g., Rapalini et al., 2015).

In our reconstruction (Fig. 11) the Paleoproterozoic Piedra Alta and
Laurentia, Baltica, Siberia, Proto-Amazonia, West Africa and proto-Australia is similar to the Columbia configuration presented by other authors (e.g., Evans and Mitchell; 2011; Zhang et al., 2012; Bispo-Santos et al., 2014; Xu et al., 2014).

The novelty in our reconstruction is the connection of the Rio de la Plata, São Francisco Craton, North China and India, constrained by the 1970–1780 Ma paleomagnetic data in the mafic dykes, as well as by a coherent geometry of the late Paleoproterozoic orogenic belts. Overall, the evolution of these belts in these cratons is consistent with the amalgamation of their Archean blocks, as exemplified by the final collision between the East and West blocks in North China along the Trans-North China belt at ca. 1850 Ma. After that an intraplate event produced 1790–1750 Ma mafic dyke swarms that crop out in most building blocks of Colombia, probably fed by LP centers such as the Xiong’er rift system (Figs. 10 and 11).

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Acknowledgements

We thank CNPq research fellowship to M. S. D’Agrella-Filho (303130/2014-8). We thank Augusto Rapalini and an anonymous reviewer for their comments that greatly improved the manuscript.

Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.precamres.2019.105584.

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