Reviewing the puzzling intracontinental termination of the Araçuaí-West Congo orogenic belt and its implications for orogenic development

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Abstract

Palinspastic reconstructions suggest that the late Proterozoic–Cambrian Brasiliano/Pan-African orogenic belt in southeast Brazil and west Congo terminated northwards into an embayment within the São Francisco-Congo cratonic unit. The orogenic shortening that created the Araçuaí-West Congo orogen in this embayment has been explained by tightening of the horseshoe-shaped São Francisco-Congo craton in a fashion referred to as "nutcracker tectonics". We show that this model is incompatible with the general orogenic evolution proposed in recent literature, which involves (1) ~50 m.y. of subduction of oceanic crust and associated arc formation, followed by (2) collisional orogeny and crustal thickening. Quantitative considerations show that the original nutcracker model is too rigid

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to explain even the second, crustal thickening part, let alone any long pre-collisional history. To soften the model, we suggest that the so-called São Francisco – Congo bridge was broken by a ~150 km wide orogenic corridor along the current African Atlantic margin. This corridor adds sufficient mobility to the system to explain the orogenic thickening of the crust to 60-65 km. However, even with this additional softening the confined nature of this orogen is incompatible with prolonged arc development. We therefore suggest that oceanic crust was nonexistent or very limited in the Macaúbas basin, and reject the widely published model involving ~50 m.y. of subduction of oceanic crust and related arc development. Instead, we find strong support for a hot intracontinental orogen model in the currently available P-T, geochronologic, petrographic and structural data. In this model, extensive melting and flow of the middle crust is likely to have caused spreading of the upper crust in an orogenic setting that was created by collisions along the N, W and S margins of the São Francisco craton from ~630 Ma.

Keywords: Hot orogen; Confined orogen; Partial melting; Brasiliano/Pan-African belt

1. Introduction

Mountain belts tend to form connected systems that cross entire continents or supercontinents, such as the extensive Alpine-Himalayan orogenic system running from Asia through the Mediterranean region, the Paleozoic Caledonide-Appalachian system and the mostly Neoproterozoic Brasiliano-Pan–African system. Within these systems, individual orogenic elements form a connected network in which they change character between orthogonal, oblique and strike-slip, but rarely terminate without transfer of displacement to other plate-tectonic elements. And where they do, they tend to do so gradually.

The Araçuaí-West Congo orogenic belt developed by shortening of a pre-orogenic rift basin with or without oceanic crust (see discussion below), and the orogen is generally regarded to terminate abruptly into a rigid cratonic continental environment largely unaffected by the Brasiliano/Pan-African deformation (Figs. 1 and 2). The kinematics of the orogen is orthogonal shortening, mostly E-W but radial in the northern part. As discussed below, the Brasiliano/Pan-African Araçuaí-West Congo belt involves substantial crustal thickening and horizontal shortening even close to its northern termination, and thus appears to represent a rather odd example of an orogen that abruptly vanishes into a continental cratonic environment. More specifically, it is surrounded by Archean and Paleoproterozoic continents to the east (Congo craton), north and west (São Fancisco craton), and throughout its late Proterozoic orogenic evolution the Araçuaí-West Congo orogenic belt has therefore been classified as a confined (Pedrosa-Soares et al., 2001) or partially confined (Alkmim et al., 2006) orogen.

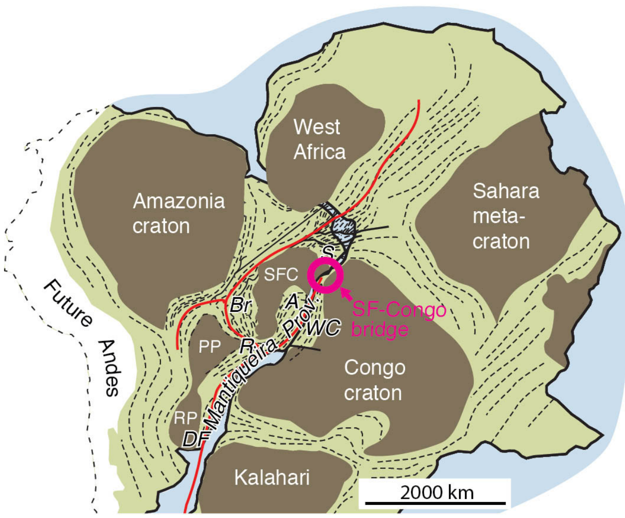

Figure 1. *Brasiliano-Pan-African orogenic belts of West Gondwana prior to the formation of the South Atlantic ocean. A= Araçuaí; Br=Brasilia orogen; DF=Dom Feliciano belt; PP= Paranapanema Craton; RP=Rio de La Plata craton; SFC=Sao Francisco craton; WC=West Congo.*

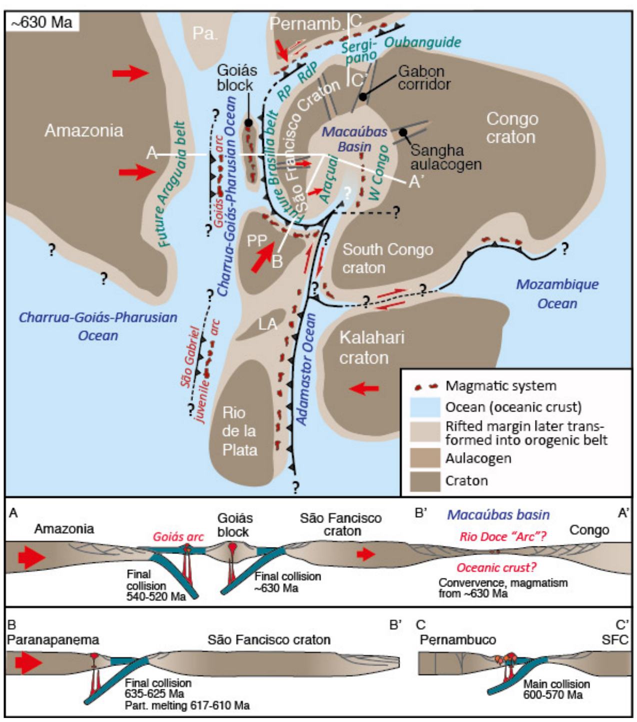

Figure 2. *Schematic tectonic setting immediately prior to the main collisional events between the São Francisco craton and surrounding cratonic and magmatic elements.*

Cross-sections through different parts of the margin are shown. Note that most authors since the late 1990s consider the Macaúbas Basin, which develops into the Araçuaí-W Congo orogen, to have hosted an ocean that started to subduct at this time (630 Ma) and until ~580 Ma (e.g., Pedrosa-Soares et al., 1998). LA=Luis Alves, PP=Paranapanema, RP= Rio Preto belt; RdP=Riacho do Pontal belt. Based in part on Meira et al. (2016).

The concept of a confined or "dead end" orogen is special, and its boundary conditions impose important constraints on the kinematic, strain and tectonic evolution of such orogenic systems. These conditions have not been sufficiently taken into consideration in the case of the Araçuaí-West Congo orogenic belt, and we here critically discuss the so-called confined model in the light of these boundary conditions. We conclude that both the existing confined orogenic model ("nutcracker tectonics"; Alkmim et al., 2006) and the widely published tectonic model for the orogenic evolution (e.g., Pedrosa-Soares et al., 1998; 2001; Gradim et al, 2014; Peixoto et al., 2015; Gonçalves et al., 2016; Richter et al., 2016; Tedeschi et al., 2016; Alkmim et al., 2017; Degler et al., 2017) need fundamental modifications, and argue that the orogenic evolution is better understood in terms of a hot orogen model (Vanderhaeghe, 2009; Jamieson and Beaumont, 2013) without prolonged oceanic subduction and magmatic arc development. We also point at data needed to better understand the evolution of this intriguing branch of the Brasiliano/Pan-African orogenic system.

2. General setting of the orogenic system

Reconstruction of the West Gondwana paleocontinent (Fig. 1) shows the Araçuaí-West Congo orogenic belt as a part of the Neoproterozoic Brasiliano/Pan-African orogenic system (Torsvik and Cocks, 2013). This orogenic system is defined by a network of orogenic belts formed by amalgamation of a plethora of larger and smaller cratonic continents into West Gondwana in the Neoproterozoic-Cambrian, following extensive Neoproterozoic rifting (Trompette, 1994, 2000). The Araçuaí-West Congo orogen is the northern part of one of these belts, known as the Mantiqueira province. This province stretches from Uruguay and northwards along the southeast coast of Brazil, and formed during convergent movements between the São Francisco, Congo, Kalahari, Rio de la Plata, and Paranapanema cratons (Fig. 1). The Ribeira belt is the central section of the Mantiqueira province, and connects the Dom Feliciano and Araçuaí-West Congo belt belts (Fig. 1). The restored width of the Araçuaí-West Congo orogenic belt is ~650-700 km, about twice that of the transpressional Ribeira belt to the south (Fig. 3). In the following we will describe the São Francisco-Congo craton and the Araçuaí-West Congo and associated orogenic belts, before discussing the problems associated with the current model and suggesting an alternative evolutionary model for the Araçuaí-West Congo orogen.

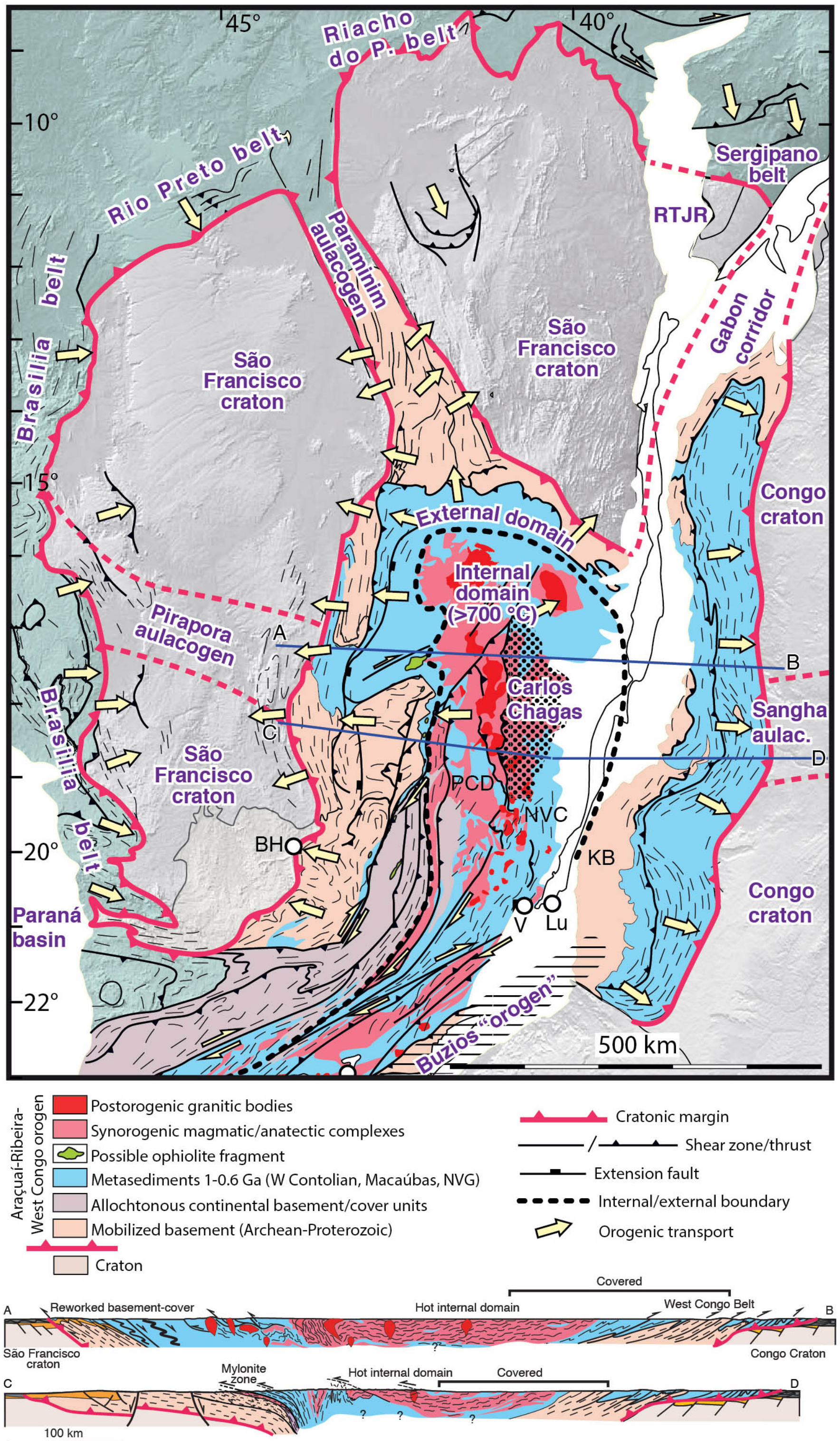

Figure 3. *Simplified geologic map of the Araçuaí-West Congo and northern Ribeira orogen, with Congo restored to its pre-Atlantic rifting situation with respect to South*

America. Yellow arrows represent kinematics during the main/late stages of orogeny, and are in part from Alkmim et al. (2006). Based on maps from the Geological Survey of Brazil (CPRM) and Tack et al. (2001). Cross-sections are based on Tack et al. (2001), Alkmim et al. (2006), and Vauchez et al. (2007). Metasediments1-0.6 Ga (blue) range from very low grade in the foreland to high-T paragneisses in the hot internal zone of the orogen. Dotted ornament indicate the Carlos Chagas anatectic domain. BH=Belo Horizonte; KB=Kimezian basement (reworked); Lu=Luanda; NVC=Nova Venécia Complex; PCD=Plutonic Central Domain of Mondou (2012); RJ=Rio de Janeiro; RTJR=Recôncavo-Tucano-Jatoba rift; SP=Sao Paulo; V=Vitoria. Geographic coordinates refer to current Brazil.

2.1 The São Francisco-Congo craton and its rift arms

The São Francisco-Congo craton consists of Archean and Paleoproterozoic rock complexes older than ~1.8 Ga, covered by a variety of supracrustal rocks of late Paleoproterozoic to late Mesoproterozoic age (Espinhaço Supergroup), followed by the rift and continental margin deposits of the Neoproterozoic Macaúbas Group (Alkmim et al., 2017). Deposition of the Macaúbas Group was related to rifting following the formation of Rhodinia at ~1.0 Ga. Most likely the São Francisco-Congo craton was not part of Rhodinia (Evans et al., 2009; 2016), but this uncertainty does not affect the late Neoproterozoic orogenic development discussed here.

The shape of the craton in the study area mimics that of a southward-opening horseshoe (Fig. 2). This shape is broken by several rift arms that were variously reactivated during the Araçuaí-West Congo orogeny. The Paramirim (Cruz and Alkmim, 2017) and Pirapora aulacogens dissect the craton into a southern, northern and northeastern part, and third rift arm, the Sangha aulacogen (Alvarez, 1995), extends into the Congo craton (Fig. 3). In addition, a 150-200 km wide north-trending orogenic corridor, informally named the Gabon corridor (Fossen et al., 2017), occurs along the African side of the South Atlantic margin, where the orogenic front continues for several hundred kilometers beyond the termination on the Brazilian side before getting buried under younger deposits (NE part of the map in Fig. 3). The Gabon corridor, which has received little attention in the previous literature, may well be a pre-orogenic rift segment similar to the better-exposed Paramirim aulacogen to the west, but subjected to more intense Pan-African reactivation. If so, it is an important tectonic element that breaks the São Francisco-Congo "bridge" and provided increased flexibility during Neoproterozoic rifting and the Brasiliano orogeny.

All of these rift arms radiate from a center located in the northern Araçuaí-West Congo orogen, hinting that a plume may have been located in this location during rifting. The largest rift was trending southwards from this rift center along what is now the Araçuaí-West Congo orogen. As a whole, this rift system accommodated the opening of the pre-orogenic Macaúbas basin in the cratonic embayment, as well as the orogenic shortening across the Araçuaí-West Congo orogen.

2.2 The São Francisco-Congo cratonic bridge

A key point in the following discussion is the widely accepted idea that the Congo and São Francisco cratons were physically connected from the Paleoproterozoic until the Cretaceous opening of the Atlantic ocean by what has been referred to as the São Francisco-Congo cratonic bridge (Porada, 1989; Pedrosa-Soares et al., 2001; Alkmim et al., 2006; Barbosa and Barbosa, 2017; Degler et al., 2018). This "cratonic bridge" has been discussed in detail by Alkmim et al. (2006), who presented the following main arguments in favor of a connection between the São Francisco and Congo cratons: 1) lack of Neoproterozoic orogenic deformation along the coast of Bahia and Gabon (a point discussed in Section 5), 2) paleomagnetic poles roughly coinciding for the two sides of the bridge (McWilliams, 1981; D'Agrella Filho et al., 1990, 2004; Renne et al., 1990), and 3) the width of the Atlantic margin being narrow, which they consider to be characteristic of rifted cratonic crust. This cratonic bridge represents a key element in a poorly understood geometric situation that puts important restrictions on the kinematic evolution of the northern part of the Araçuaí-Ribeira-West Congo orogenic system.

In spite of the general acceptance of the cratonic bridge, the prevailing tectonic model for the confined orogen south of the cratonic bridge is that of eastward subduction of oceanic crust under the West Congo rifted margin (Fig. 2), and subsequent collision between the West Congo margin and the eastern margin of the São Francisco craton (Pedrosa-Soares et al., 1998; Alkmim et al., 2006; Vauchez et al., 2007). Trompette (1994, 1997), on the other hand, considered the Araçuaí-West Congo belt as "partly or totally intracratonic" (Trompette, 2000), with the Adamastor ocean "ending northwards in a complex and wide continental rift system identified in the Araçuaí-Ribeira-West Congo belt" (Trompette, 1994), largely similar to our interpretation shown in Fig. 2.

2.3 Brasiliano orogenesis

A protracted Neoproterozoic orogenic history created the Brasiliano orogenic system, which around the São Francisco craton includes the Araçuaí-West Congo-Ribeira orogenic belt along its eastern margin, the Brasilia belt along its southern and western margins, and the Rio Preto, Riacho do Pontal and Sergipano belts to the north (Fig. 2). All of these orogenic belts are connected as parts of the Brasiliano orogenic system, which developed as a result of convergent to oblique interaction between different cratonic and arc elements, in the Brasilia belt from as early as ~900 Ma (Pimentel, 2016). These interactions culminated around 630-600 Ma to form West Gondwana, although orogenic pulses and events locally occurred as late as the early Cambrian (Schmitt, 2004). The São Francisco craton was affected by all of the major collisions, and each of its surrounding belts is briefly summarized in the following.

2.3.1 The northern Brasilia belt

The northern Brasilia orogenic belt is the result of collision of the Archean-Proterozoic Goiás microcontinent, the 670-639 Ma Goiás arc, and possibly other magmatic arc systems with the western passive margin of the São Francisco craton. The major continental collisional stage of this convergent history occurred at ~630 Ma (Pimentel, 2016; Fuck et al., 2017) or 640-610 Ma (Valeriano et al., 2008). The Amazon craton itself, also moving eastward relative to the São Francisco continent, collided later, probably between 540-520 Ma (Valeriano et al., 2008) or around 550 Ma (Moura et al., 2008) to form the Araguaia belt (Fig. 2). This late concluding event may explain thin-skinned thrusting of the Bambuí Group cover of the São Francisco craton (Reis and Alkmim, 2017). The Bambuí Group has recently been suggested to be as young as 550-542 Ma (Warren et al., 2014). If correct, the Brasilia orogeny seems to have lasted until the dawn of the Cambrian.

2.3.2 The southern Brasilia belt

The southern Brasilia belt was formed by the northward motion of the Paranapanema/Rio de Plata continent, causing accretion of a significant orogenic wedge of allochthonous units onto the southern São Francisco margin. In general, outboard (arc-related) terranes tectonically overlie high-grade units with anatectic domains and retrogressed eclogite, again overlying low-grade units of reworked São Francisco margin affinity (Campos Neto et al., 2011; Valeriano et al., 2008). This pile of thrust nappes developed diachronously with the age of deformation younging toward the São Francisco craton (Campos Neto et al., 2011). Anatectic melting is dated at 617-610 Ma (Martins et al., 2009), and has tentatively been associated with channel flow of the middle crust after crustal thickening (Fig. 4) (Campos Neto et al., 2011). Given the fact that extensive partial melting requires something like 20 Ma of continent-continent collision (Jamieson et al., 2011; Vanderhaeghe, 2009), the main collisional event must have initiated around or before 637 Ma, which corresponds well with the 650-630 Ma age suggested by Valeriano (2017). Hence, the main collisional event of the southern Brasilia belt appears to be broadly synchronous with the main event in the northern Brasilia belt.

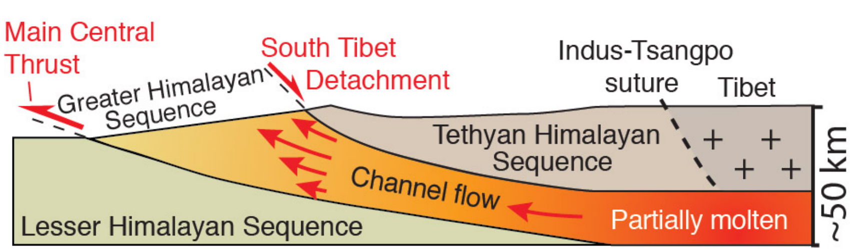

Figure 4. *The concept of channel flow (e.g., Nelson et al., 1996) in the context of the Himalayan orogen. Hot and partially molten rocks flow within a channel from the lower or*

middle crust toward the foreland under the weight of an overlying orogenic edifice (plateau). Modified from Webb et al (2011). Our suggestion is that the hot internal part of the Araçuaí-West Congo belt represents an erosional section through the partially molten crust.

2.3.3 The Sergipano belt

The Sergipano belt (Fig. 2) is the south-verging orogenic belt located immediately north-northeast of the São Francisco craton. It consists largely of low-grade shelf sediments thrust southward onto the São Francisco craton and intruded by granitic magma. This occurred in response to southward movement of the Pernambuco block to the north from ca. 630 Ma (Oliveira et al., 2006), i.e. contemporaneous with major orogenic activity in the southern and western Brasilia belts. Convergent movements appear to have continued at least until 570 Ma, with muscovite defining the pervasive D2 foliation dated at 591±4 Ma (40Ar/39Ar) (Oliveira et al, 2010). Araujo et al. (2013) suggest that this collision happened around 590-580 Ma, contemporaneous with extensive transcurrent shearing in the Pernambuco block (Archanjo et al., 2013). On a larger scale, it connects with the Oubanguide orogen in NW Africa (Trompette, 2000) and the Riacho do Pontal and Rio Preto belt to the west (Fig. 2).

2.3.4 The Rio Preto and Riacho do Pontal belts

This 600 km long part of the Brasiliano system borders the São Francisco craton to the north and northwest, and connects with the North Brasilia and Sergipano belts (Fig. 2). The Rio Preto and Riacho do Pontal orogenesis involved a pre-Brasiliano (~900 Ma and younger) rift system and passive margin, through a combination of N-S shortening and lateral escape (i.e., partitioned transpression). Collision tectonics is believed to have initiated at around 620 Ma, after a period of northward subduction of oceanic crust and related arc development (Caxito et al., 2017).

2.3.5 *Timing of collisions around the São Francisco craton*

The São Francisco craton was affected by collisions from all of the aforementioned orogenic belts, notably the prolonged collisional history to the west (the northern Brasilia orogeny) and collisions in response to north or northeastward motion of the Paranapanema craton to the south. Most of these belts appear to record main collisional events at roughly 630 Ma, which seem to have started the shortening of the confined Macaúbas basin and the crustal thickening that lead to the formation of the Araçuaí-West Congo orogen described below. In particular, the prolonged collisional history to the west (the northern Brasilia orogeny) and collisions in response to north or northeastward motion of the Paranapanema craton to the south were important for the development of the Araçuaí-West Congo orogen.

3. The Araçuaí-West Congo orogen

The Araçuaí-West Congo orogen consists of an external fold-and-thrust belt and a wide and hot internal domain characterized by high temperatures and extensive partial melting and magmatism (Pedrosa-Soares et al., 2001; Vauchez et al., 2007; Cavalcante et al., 2013, 2014, 2016; Alkmim et al., 2017) (Fig. 3). These two parts are separated by a 5 km thick high-T/low-P mylonitic thrust zone with top-to-foreland sense of shear, according to Vauchez et al. (2007). The pre-orogenic basin, here referred to as the Macaúbas basin, consists mainly of the up to 10 km thick Macaúbas Group in Brazil and much (~4 km) of the West Congolian Group in Congo, and records mostly the pre-orogenic rift basin history. Initial rifting is constrained by dating of magmatic activity and detrital zircons to around 850 Ma in the Macaúbas Group on the São Francisco craton (Alkmim and Martins-Neto, 2012 and references therein) with evidence of somewhat earlier rift initiation elsewhere in the Araçuaí-West Congo orogen (Tack et al., 2001; Pedrosa-Soares et al., 2008), and both groups include poorly constrained glacial deposits. Formation of oceanic crust in this basin has been suggested, based on limited occurrences of rather poorly dated (816±72 Ma, Sm-Nd whole-rock isochron) amphibolite and ultramafic rocks on the Brazilian side of the orogen (Pedrosa-Soares et al., 1998).

Metasediments and metavolcanics also occur as migmatites and migmatitic granulites in the central part of the orogen. Some of these are probably highly altered sediments of the Macaúbas Group, while other parts (Nova Venécia Complex) have been interpreted as syn-orogenic (back-arc) deposits whose depositional age is bracketed by their youngest detrital zircon age of 606±3 Ma and intrusions dated at 593±8 Ma (Richter et al., 2016).

3.1 The external fold-and-thrust belt

The external belt of the orogen has a narrow unmetamorphosed to low-grade thin-skinned foreland part that involves the sedimentary succession covering the São Francisco and West Congo continents (Bambuí and West Congolian groups; Tack et al., 2001; Reis and Alkmim, 2015). However, basement involvement is seen relatively close to the orogenic front, locally triggered by reactivation of pre-orogenic rift faults (Alkmim et al., 2017). The metamorphic grade increases into the orogenic belt, where allochthonous basement soon exhibits ductile fabrics of greenschist to amphibolite facies. Mylonitic basement rocks of high-temperature (~750 °C; Vauchez et al., 2007) amphibolite facies occur in the mylonite zone that marks the base of the external domain, with kinematic indicators consistent with thrusting toward the west foreland (e.g., Vauchez et al., 2007). In the southern Araçuaí and into the Ribeira belt, there are also large elongated units of variously sheared magmatic and gneissic rocks interpreted as terranes of both continental margin and arc affinity (Heilbron et al., 2008).

The kinematics of this entire external belt is everywhere top-to-the-craton (Figs. 3 and 5a), with some local evidence of late extensional reactivation (Marshak et al., 2006). The metamorphic conditions increase progressively from very-low grade along the cratonic margin to amphibolite and granulite facies close to the border of the internal domain (Pedrosa-Soares et al., 2001).

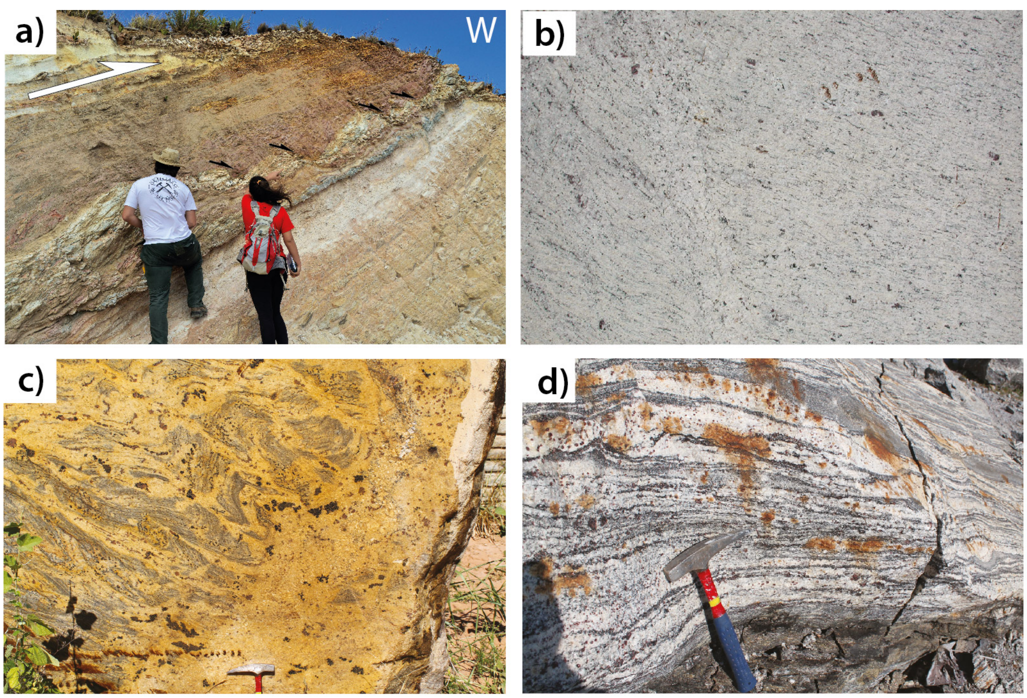

Figure 5. *Field aspects of the Araçuaí belt. (a) Asymmetric boudinage in low-grade metasediments showing top-to-foreland (W) thrusting near the thrust front; (b) diatexite*

with magmatic foliation marked by aligned biotite and feldspar; c) and d) metatexites exhibiting a migmatitic foliation associated with leucosome rich in garnet, forming networks of interconnected melt, which suggest high volume (>40%) of magma during deformation (i.e., magmatic state deformation).

3.2 The internal domain (hinterland)

The internal hinterland of the Araçuaí-West Congo orogen defines the up to 250 km wide high-temperature core of the orogen, and is made up of high-grade metamorphic rocks and vast amounts of granites and migmatitic rocks that range in crystallization age from 630 to 480 Ma. This includes the Plutonic Central Domain of Mondou et al. (2012) and the Carlos Chagas anatectic domain (Figs. 3 and 5 b-d), which is a 100 by 300 km large area dominated by anatectic rocks formed by partial melting of the middle crust and deformed predominantly at the magmatic state (Cavalcante et al., 2013). The Plutonic Central Domain consists of tonalitic and granodioritic bodies (the "Galiléia" and "São Vitor") emplaced during a magmatic event at ~580 Ma and deformed at the magmatic state, coherently with their metasedimentary country rocks (Mondou et al. 2012). These bodies have calc-alkaline composition interpreted as representative of a magmatic arc, which would span from 630 to 585 or 580 Ma (e.g., Tedeschi et al. 2016) and imply the consumption of oceanic crust for 45-50 million years. Paragneisses showing evidence of partial melting are widespread in the internal domain (blue unit 3 in Fig. 3), and are at least in part considered to represent partially molten Macaúbas Group (e.g., Dias et al., 2016), but also synorogenic metasediments (Nova Venécia Group; Richter et al., 2016). Temperature estimates from different techniques consistently indicate peak metamorphic temperatures of 750-850 °C for this internal core of the orogen (Cavalcante et al., 2014;

Moraes et al., 2015; Dias et al., 2016) and pressures around 6-7 kbar (Munhá et al., 2005; Moraes et al., 2015). The abundant granitoid rocks in the hot internal orogenic domain have been separated into (super)suites representing “pre-collisional” arc magmatism (630-580), a “syn-collisional” (585-530 Ma) and a “post-collisional” (530-480 Ma) suite by Pedrosa-Soares et al. (2001, 2011; Gonçalves et al. 2016). The “pre-collisional” suite (also called G1) is dominated by I-type, metaluminous to slightly peraluminous expanded calc-alkaline granites, while the “syn-collisional” (also called G2) granites mostly consists of S-type, peraluminous, sub- to calc-alkaline granites (Gonçalves et al., 2014; Tedeschi et al., 2016). Acceptance of this classification and evolutionary model poses important implications for the tectonic and kinematic evolution of the Araçuaí-West Congo orogen, as will be discussed in more detail below.

4. How much shortening across the Araçuaí–West Congo orogen?

In general, to estimate the amount of shortening between two converging continents we need to consider both the contribution from oceanic consumption (subduction) and the horizontal shortening and corresponding vertical thickening of the continental margin during what is referred to as the continent-continent collision phase. We will start with the shortening involved in the crustal thickening process, generally considered as the result of continent-continent collision between the east margin of the São Francisco craton and the west margin of the Congo craton.

4.1 Shortening associated with the crustal thickening (“collision”)

The relationship between crustal thickening and convergence is not always straightforward at convergent plate boundaries. Vanderhaeghe and Duchêne (2010) have shown how the pattern of thickening relates to slab advancement or retreat, and to the degree of coupling between the mantle lithosphere and the overlying continental crust. However, in a confined orogen such as the Araçuaí–West Congo, where there is no evidence of deep subduction of continental crust and where continental accretion due to any slab rollback would be balanced by upper-plate stretching, is a simpler case. In this case the thickening of the continental crust is proportional to the horizontal shortening and convergence between the southern São Francisco and West Congo parts of the Congo craton during what is referred to as the collisional stage in the recent literature on this orogen.

Estimating the amount of shortening of continental crust across orogens commonly involves palinspastic reconstructions or section restorations. Such restorations are difficult to perform for the Araçuaí–West Congo orogen because of the lack of restorable allochthonous units (thrust nappes) and marker horizons, and poor depth control due to low topographic relief and little relevant geophysical data. Furthermore, deformation in the internal hot part of the orogen was disseminated and absorbed by partially molten rocks with little memory of strain and displacement (Vauchez et al., 2007). Hence, the best approach is to consider the transformation of thin, rifted crust to an overthickened orogenic continental crust. This involves assumptions regarding the preorogenic basin and the geometry and thickness of the resulting orogenic belt.

The proximal margins of the pre-orogenic Macaúbas basin are located under the foreland fold-and-thrust belt on both the São Francisco and West Congo sides of the orogen

(Tack et al., 2001; Pedrosa-Soares et al., 2008) (Fig. 3). This implies that the orogenic foreland closely coincides with the limits of the pre-orogenic volcano-sedimentary basin, whose attenuated crust was subsequently shortened, metamorphosed and incorporated into an orogenic crust that was thick and hot enough for extensive melting to occur.

The crustal thickness that was achieved during the Araçuaí–West Congo orogeny is revealed by metamorphic pressure estimates. Pressure associated with the metamorphic peak (ca. 580 Ma) have been calculated by several authors from different sections of the internal part of the orogen, and most of the data indicate pressures of 0.6-0.7 GPa (Munhá et al., 2005; Belém, 2006; Petitgard, 2009; Uhlein et al., 2009; Gradim et al., 2014; Cavalcante et al., 2014; Moraes et al., 2015; Dias et al., 2016; Gonçalves et al., 2016), with slightly higher pressures (~0.8 GPa) reported from the southernmost part of the Araçuaí belt (Bentos dos Santos et al., 2011) (Fig. 6). These data indicate that the present erosion level was located at depths of around 20-25 km during the metamorphic peak, and that the crustal thickness in the internal part of the orogen was fairly constant, as expected for a plateau-type orogen (Vanderhaeghe and Teyssier, 2001). With a uniform current crustal thickness of around 40 km (Assumpçao et al., 2017), this implies that the crust was fairly flat-based with a total thickness of 60-65 km across the hot internal part of the orogen at the time of peak Araçuaí-West Congo metamorphism (e.g., Cavalcante et al., 2014). Deep crustal subduction not only produces roots, but also channels of vertical extrusion along which (ultra)high pressure rocks are exhumed (Liou et al., 2004; Butler et al., 2013), and we find no trace of such extrusion in the Araçuaí-West Congo orogen.

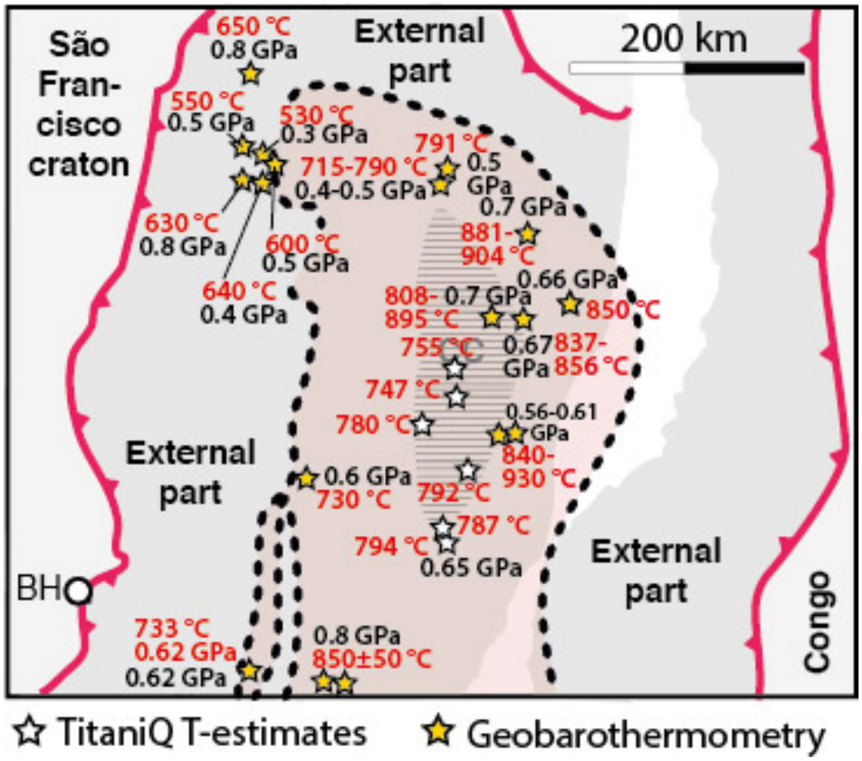

Figure 6. *Peak pressure and temperature estimates from the internal hot part of the orogen. TitaniQ temperatures represent minimum crystallization temperatures of quartz from Cavalcante et al. (2014 and 2018). Geobarothermometric data from Garcia et al. (2003), Schmitt et al. (2004), Munhá et al. (2005), Belém (2006), Petitgirard et al. (2009), Uhlein et al. (2009), Bento dos Santos et al. (2011), Gradim et al. (2014), Moraes et al. (2015), Degler et al. (2018).*

To estimate the pre-orogenic crustal thickness, we consider the current 40 km thickness of the São Francisco craton (Assumpção et al., 2017) to have been the cratonic thickness also in the Neoproterozoic. Furthermore, the pre-orogenic basin must have been wider than the ~600 km wide orogenic belt, and this continental crust was a rift or rifted margins with thinned crust. The crustal thickness in wide continental margin or rift settings is variable, but is usually around or slightly less than half of its original thickness (e.g., Faleide et al., 2008; Reston 2010; Huismans and Beaumont, 2014). Hence an average reduction from 40 to 20 km seems like a reasonable estimate. Basin sediments that were later metamorphosed during the orogeny and thus contributed to the Araçuaí–West Congo crust are included in this estimate, but any oceanic crust that might have existed is assumed to have been subducted and is therefore not considered.

Using this assumption, the hot internal part of the orogen increased in thickness from 20 to 60 km over its current width of ~250 km (Fig. 7a) during the orogeny. Restoring to a pre-orogenic basin with 20 km crustal thickness gives a 750 km wide basin and 500 km of orogenic shortening, as illustrated schematically in Fig. 7b. For comparison, a 750 km wide continental basin is similar to the rifted South Atlantic margin across the Santos basin (Fig. 7c) (e.g., Szatmari and Milani, 2016), and the width of the Basin and Range basin in the western USA is around 800 km.

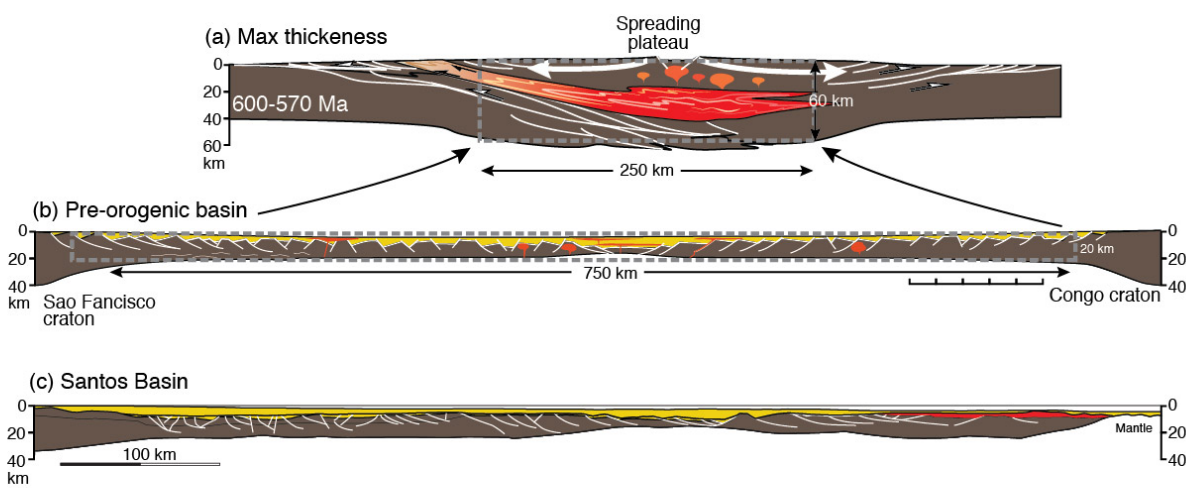

Figure 7. *a) Schematic cross-section through the Araçuaí orogen at the time of partial melting of the middle crust. b) Illustration of what the pre-orogenic Macaúbas basin may have looked like. An average Moho depth of 20 km is chosen. Stippled rectangle in a) reflects the average crustal thickness in the thick-skinned part of the orogen, and the corresponding pre-orogenic shape of this area is presented by the rectangle in b). c) The*

800 km wide rifted South Atlantic margin across the Santos basin, for comparison (from Magnavita, 2014 and Szatmari et al., 2016).

In addition, some foreland shortening outside the rectangular area in Fig. 7 occurred, but at least some of this foreland thrusting/thickening was driven by gravitational spreading of the internal part of the orogen after peak metamorphism, and this part should not be included. More dating of deformation in the foreland is necessary to distinguish between these two components. On the contrary, any material added or subtracted to the section by northward flow from the pinching point at the southern termination of the São Francisco craton would affect the amount of shortening to some extent. Similarly, introduction of intrusive rocks from the mantle during the orogeny would overestimate the amount of thickening. However, most of the magmatic rocks originated by partial melting of the crust (Gonçalves et al., 2017) and would therefore not affect the mass balance. Hence, our 500 km estimate of orogenic shortening is considered to be a reasonable first-order estimate.

According to most recent workers the Araçuaí–West Congo orogeny lasted for 50-55 m.y. (585-530 Ma) (Pedrosa-Soares, 2001, 2011; Gradim et al., 2014; Tedeschi et al., 2016; Alkmim, 2017). The convergence rate during continent collision is usually considerably lower than those typical for oceanic subduction, because of the gravitational resistance of continental crust to subduction. For instance, the convergence rate of the Himalayan system slowed down from >10 cm/y to 4.5 cm/y (Klootwijk et al., 1992). For a confined situation like the Araçuaí–West Congo orogen, the convergence may have been even slower. For example, a low average convergence rate of 1 cm/y would, over 50 m.y., produce 500 km of shortening across the Araçuaí-West Congo orogen, i.e. the same order of magnitude estimated above.

4.2 Implications of any "pre-collisional" subduction

Any subduction of oceanic crust prior to what is referred to as "collision" in the recent literature would imply convergence prior to the continental shortening discussed above. The prevailing model regarding oceanic crust in the Macaúbas basin and its consumption involves extensive arc magmatism and prolonged subduction of oceanic crust (Pedrosa-Soares et al., 2011; Gonçalves et al., 2014), as presented or assumed in a large number of recent contributions (e.g., Pedrosa-Soares et al., 1998, 2001, 2008; Alkmim et al., 2006; Gradim et al., 2014; Kuchenbecker et al., 2015; Moraes et al., 2015; Peixoto et al., 2015; Dias et al., 2016; Gonçalves et al., 2016, 2017; Richter et al., 2016; Tedeschi et al., 2016; Alkmim et al., 2017; Degler et al., 2017; Melo et al., 2017a, b). This model is based on geochemical and geochronologic data from magmatic rocks in the orogen, and argues for ~50 m.y. of arc magmatism and related subduction of oceanic crust. Subduction rates generally vary from 2-10 cm/y, for example the fast subduction of the oceanic part of the Indian plate under Asia at >10 cm/y (prior to the Himalayan collision) versus the slow subduction at ~2 cm/y for the Lesser Antilles system (Stein, 1983). Picking a slow subduction rate of 2 cm/y implies ~1000 km of shortening across the Macaúbas basin prior to continent collision. This most likely represents about ~1000 km of eastward displacement of the São Francisco craton relative to the Congo craton. Arguably, slab rollback could absorb a limited amount of these 1000 km by further stretching of an already thinned continental margin. At some point slab rollback would create an oceanic back-arc basin that would produce new oceanic crust. Regardless, the unsolvable problem of putting a 1000 km wide ocean into the confined environment of the Araçuaí-West Congo remains. An ocean close to this size (750 km) was schematically indicated by Richter et al. (2016) (Fig. 8c). However, by adding such an ocean to this embayment leaves far too little continental margin to even thicken the crust to normal thickness, let alone to build a 60-65 km thick orogenic crust, as shown in Fig. 7.

Other references to the size of this ocean have been made by Pedrosa-Soares et al. (1998), who state that "the extensive occurrence of syntectonic to late tectonic calc-alkalic granitoids along the internal domain of the Araçuaí belt implies that a reasonably large amount of ocean crust was consumed". As discussed above, "reasonable" implies something in the order of 1000 km or more. However, in another publication Pedrosa-Soares et al. (2001) state that "only a narrow oceanic lithosphere was generated, and it was subducted afterwards". Such self-contradictory statements illustrate the need for quantitative evaluations when considering tectonic models for the Araçuaí-West Congo orogen.

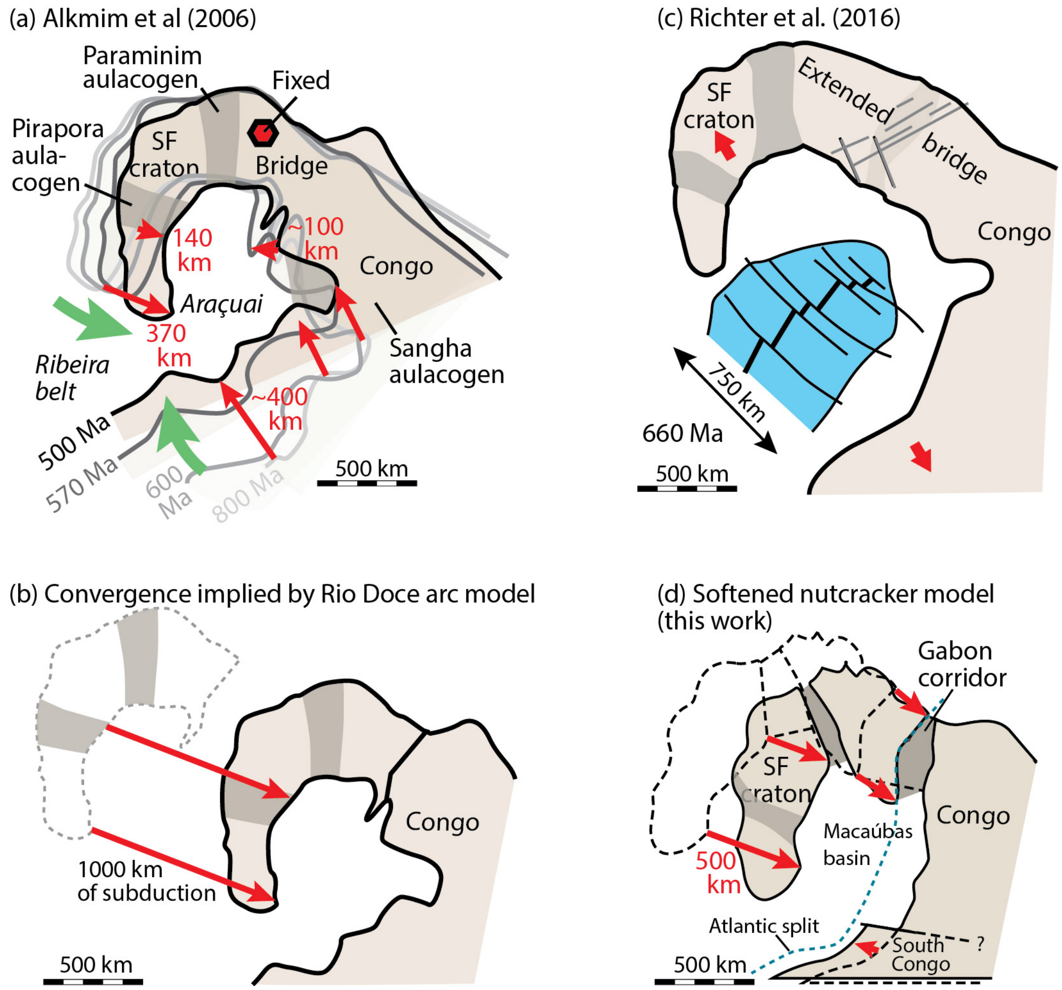

Figure 8. *a) Evolution of the confined Araçuaí orogen as interpreted by Alkmim et al. (2006, their fig. 15), showing progressive closing of the pre-orogenic Macaúbas basin at 800, 600, 570, and 500 Ma (successively less transparent). b) Restoration of the São Francisco craton to allow for a 1000 km wide ocean between the São Francisco and Congo cratons that was subducted during convergence. Note that the crustal thickening ("collisional phase") implies additional 500 km of displacement. c) Model presented by Richter et al. (2016), where an ocean has been allowed by extending the continental bridge considerably. Richter et al. provide no explanation for how this bridge shortened during the orogeny. d) Modified model, taking into account the Gabon corridor, here considered to accommodate ~200 km of orthogonal shortening.*

5. Kinematic models for the Araçuaí-West Congo orogen

The above discussion suggests that ~500 km of convergence is needed to form the Araçuaí-West Congo orogen in a intracontinental setting, and that additional convergence on the order of 1000 km is needed to explain the widely published model of extensive arc magmatism. In the following we will discuss the implications of these numbers in terms of the constraints imposed by the confined orogenic setting.

The kinematic evolution of the Araçuaí-West Congo orogen has been discussed qualitatively by Alkmim et al. (2006), who presented the "nutcracker" model where the Macaúbas basin and its underlying crust were shortened by anticlockwise rotation of the São Francisco craton relative to Congo (indicated by green arrows in Fig. 8). Figure 8a shows successive stages of Alkmim et al.'s model, from 800 Ma (pre-convergence) until the post-orogenic stage (500 Ma). This presentation shows that their model, based on their own illustration, produces only ~300 km of total convergence across the central Araçuaí part of the orogen, decreasing to the north and increasing to ~750 km at the southern termination of the São Francisco craton and into the Ribeira belt. Hence, this model has difficulties accounting for the crustal thickening associated with the Araçuaí-West Congo orogen, and has no room for any oceanic crust at all.

To illustrate the problems involved in incorporating the >1000 km of estimated subduction-related shortening, the São Francisco craton was moved 1000 km to the west in Fig. 8b. This amount of convergence would require the São Francisco craton to have moved completely independent of the Congo craton, which is incompatible with the nutcracker model and the idea of a cratonic bridge, as discussed above. In an apparent attempt to get around this problem, Richter et al. (2016) in their reconstruction (their figure 2) extended the bridge connecting the São Francisco and Congo cratons. Richter et al. (2016) provide no explanation as to how their shortening of this part of the craton was accommodated. However, we have already pointed out that the Gabon corridor (Fig. 8d) could accommodate such strain, but how much? This corridor is 150–200 km wide (depending on the interpretation and restoration of the passive margins in the area), and based on the global relationship between width and shortening across orogenic belts (Fig. 9) it seems unlikely to represent much more than 200 km of orthogonal shortening. Further, there is not room for any significant additional strike-slip deformation along the Gabon corridor, as such motions would be hampered by the transverse Sergipano belt to the north. A reconstruction similar to the one by Alkmim et al. (2006), but with the Gabon corridor added, allows for a total of ~500 km of shortening across the Araçuaí-West Congo orogen. This may be sufficient to explain the crustal thickening reflected by the geobarometric data, but it leaves no space for the subduction-related consumption of oceanic crust called for in most recent publications from this orogen (e.g., Pedrosa-Soares et al., 1998). In other words, the combined strain associated with the Gabon corridor, Paraminim aulacogen and other structures that split the craton into subunits cannot allow for much more than 500 km of convergence across the Araçuaí-West Congo orogen.

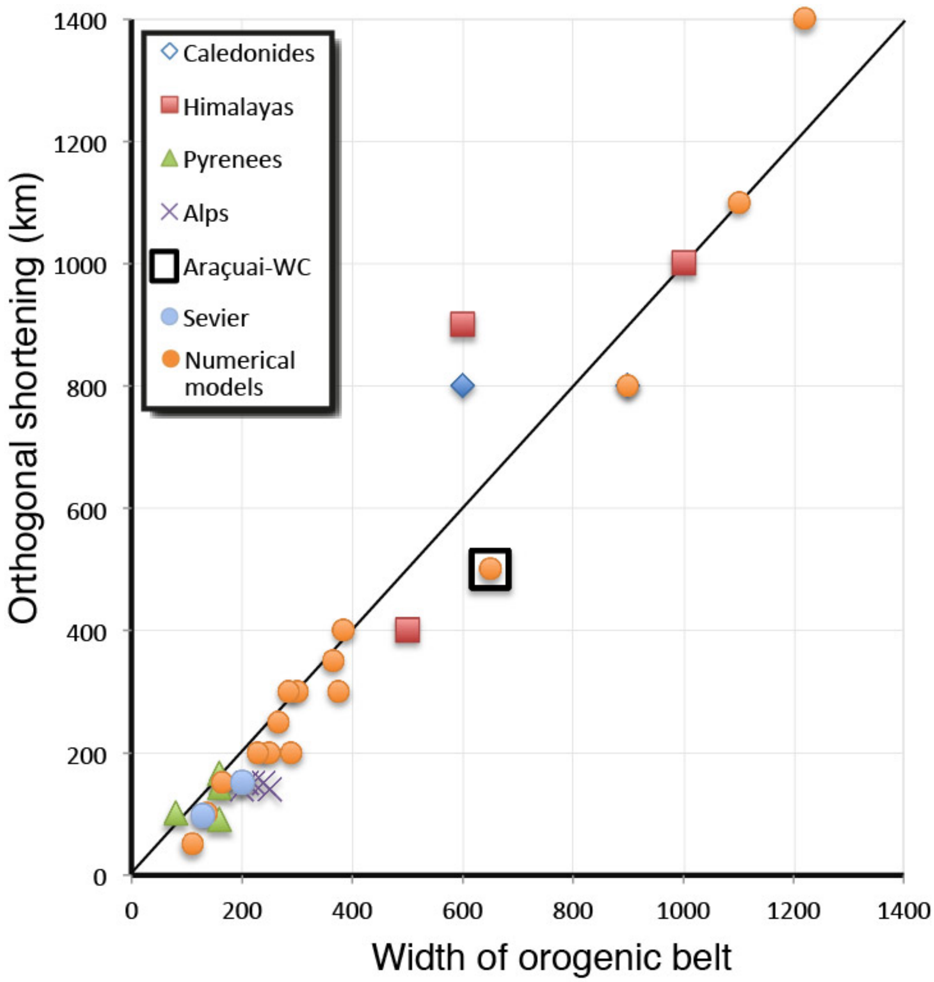

Figure 9*. Width of orogenic belts plotted against their orthogonal shortening, based on data from Armstrong, 1968; Beaumont et al., 2000, 2010; DeCelles and DeCelles 2001, 2002; Erdös et al., 2014; Long et al., 2011; Li et al, 2015; Mouthereau et al., 2014; Robinson and Delores, 2008; Rosenberg and Berger, 2009; Schmid and Kissling, 2000; Weil and Yonkee, 2012. The reference line indicates a linear 1:1 relationship between orthogonal shortening and orogenic width.*

We conclude that qualitatively, the "nutcracker" model by Alkmim et al. (2006) is a viable model, but that it needs some additional flexibility to accommodate the strain associated with the mountain building. We suggest that this flexibility can be added along the orogenic zone extending northward across the bridge along the current Atlantic margin. Furthermore, it is clear that the arc development presented in the literature is at odds with even the softened nutcracker model (Fig. 8d) or any other kinematic model proposed for the region (see Alkmim et al., 2006 for a review), and we raise the question if there was ever any significant amount of oceanic crust in this northern part of the Mantiqueira orogenic province.

6. Was there an ocean at all?

6.1 Possible ophiolite fragments

The possibility that the pre-orogenic Macaúbas basin rested on extended continental crust with no significant amount of oceanic crust has been advocated by Trompette (1994; 2000) and later by Meira et al. (2015), who suggested the entire Ribeira–Araçuaí orogen to be intracratonic. Contrary to this interpretation, restricted occurrences of amphibolites (interpreted as metabasalt) and metamorphosed ultramafic rocks have been presented as evidence of oceanic crust from the pre-orogenic basin (Pedrosa-Soares et al., 1998). In general, orogens that involve tens of millions of years of subduction and island arc activity, such as the Caledonides, Appalachians and the Himalayas all contain abundant evidence of oceanic crust in the form of well preserved ophiolite complexes, in addition to island-arc magmatism, even if we consider a high grade of chemical weathering. It is also worth noting that several ophiolites in the Alpine orogenic system have been suggested to be rift- or breakup-related, and not actual oceanic crust (Koglin et al., 2009; Dilek and Furnes, 2014). Hence, ophiolite fragments are not evidence for a former oceanic basin.

In the Araçuaí-West Congo orogen, the possible ophiolite fragments are small, strongly altered, do not display any characteristic ophiolite pseudo-stratigraphy, bear no information about the extent of oceanic crust, except that they are claimed to be almost 200 m.y. older than the early magmatism that is interpreted as arc magmatism and subduction initiation (Pedrosa-Soares et al., 1998). This is a very long time span (for comparison, nearly all current oceanic crust is younger than 200 m.y.) and would imply a several thousand kilometers wide pre-orogenic ocean, comparable in width to that of the Atlantic Ocean. Subducting such a wide ocean over 50 m.y. (630-580 Ma) is another challenge in a confined system such as the Araçuaí-West Congo. Furthermore, it is unusual in any orogen to preserve such old and dense oceanic crust. Instead, most orogenic ophiolites represent buoyant oceanic crust from small and young oceanic forearcs or backarcs basins (Stern, 2004).

There are several examples of orogens that only involved very small oceanic basins or no oceanic crust at all. The Alps is a well-known example, and it still contains ophiolitic rocks (e.g., Chenin et al., 2017). The Pyreneean orogen is another example where oceanic crust may not have been involved at all (Beaumont et al., 2000). Instead, a domain of hyperextend continental crust and extended subcontinental depleted mantle appears to have existed. Whether this was the case in the Macaúbas rift basin is unknown, but should be kept in mind.

6.2 Orogenic magmatism

The interpretation of large paleo-oceans in convergent settings typically relates to long-lived arc magmatism, identified by tectonic context and geochemical and isotopic signature. While a review of the vast amount of published geochemical data from the Araçuaí-West Congo orogen is outside of the scope of the present contribution, we note that magmatic rocks considered to represent a pre-collisional arc (Rio Doce arc; Tedeschi et al., 2016) was built upon Paleoproterozoic continental crust considered to represent the western margin of the Congo craton (Gonçalves et al., 2017). Even though some these early magmatic rocks share geochemical similarities with rocks from more modern continental arcs, for instance the Sierra Nevada arc and the Andean belt (Gonçalves et al., 2014, 2016), distinguishing between arc-generated magmatic rocks and magmatic rocks formed during hot continental orogenesis is not straight-forward (e.g., Barbarin, 1999). More specifically, the calc-alkaline composition of magmatic rocks from the central domain of the Araçuaí belt (Galiléia and São Vitor bodies) is not unequivocal evidence for subduction-related magmatic arcs, as suggested by Tedeschi et al. (2016). Such a composition can also be found in extensional settings, such as the Basin and Range province (Western USA) and the Gulf of California (e.g., Sheth et al. 2002) and in continental collision settings (e.g., Barbarin 1999). It also seems relevant in this context to point out that the basement to the Ediacaran Rio Doce magmatic rocks was already a juvenile Early Proterozoic magmatic arc, based on its geochemical signature, ε Nd values and the absence of inherited zircon grains (Noce et al. 2007). Hence the origin and tectonic implications of the 630-575 Ma magmatism in the Araçuaí belt should be critically reassessed, as already suggested by Meira et al. (2015).

7. A revised orogenic model

Concluding from the above that any pre-orogenic ocean must have been very small or absent, we would expect orogenic thickening between the two cratonic margins to have happened at a much earlier time than that postulated by most authors (585-580 Ma; Pedrosa-Soares, 2001, 2011; Gradim et al., 2014; Tedeschi et al., 2016; Alkmim, 2017). New radiometric and thermal data show that crystallization of the anatectic core of the orogen (Carlos Chagas anatectic domain) was going on already around 600 Ma, and that the middle crust at this point was already heated to more than 750 °C in a large (150,000 km^2) area (Fig. 6) (Cavalcante et al., 2018). The achievement of such high temperatures and associated widespread partial melting together with the transformation of a thinned crust to an overthickened orogenic crust requires time (~20 m.y.; e.g., Horton et al., 2016) (see below). Hence, thickening of the continental crust could well have started at 630-620 Ma. This eliminates the model involving prolonged subduction of a vast amount of oceanic crust. Hence these two lines of arguments (little or no oceanic subduction, and crustal thickening starting at 630-620 Ma) go very well together, and form the basis for an alternative, hot orogen model for the Araçuaí-West Congo orogen.

Below we outline a hot orogen model for the Araçuaí–West Congo orogen that conforms to the following conditions:

1) The total amount of convergence across the orogen was on the order of 500 km;
2) Only limited or no oceanic crust existed;

3) Much of the melt in the hot internal part of the orogen formed by partial melting of the middle crust, probably in response to orogenic thickening and radioactive decay;
4) Crustal thickening initiated earlier than 600 Ma, and probably before 620 Ma.

The first two points relate to the space problem involved in putting an ocean of any significant size into the confined Araçuaí–West Congo orogenic system (Fig. 8) and are already discussed above. The third point relates to evidence in favor of orogenic extrusion or channel flow in a hot overthickened crust, as presented by Cavalcante et al. (2013, 2014, 2016), and the fourth point is based on recent dating of the crystallization of mid-crustal melt in the central anatectic part (Carlos Chagas anatexite) of the orogen (Cavalcante et al., 2018).

7.1 The hot orogen model

A hot orogen model for the Araçuaí–West Congo orogen has developed over the last decade through the work by Vauchez et al. (2007), Petitgirard et al. (2007), Mondou et al. (2012) and Cavalcante et al. (2013, 2014, 2018), and has implications for the temporal and rheologic evolution of the Araçuaí–West Congo orogen that fits the kinematic constraints of this orogen quite well (Fossen et al., 2017). Elevated temperatures in the central part of the orogen may have several causes. One may be the high thermal gradient that can be expected for the pre-orogenic rift. The radial arrangement of rift arms centered at the location of the Araçuaí–West Congo orogen may indicate high heat flow from the mantle during the pre-orogenic rifting. Hence the crust may have been hot already at the onset of crustal thickening. The extensive magmatism governing the Araçuaí–West Congo hinterland (Pedrosa-Soares et al., 2011) also suggests anomalously high temperatures in this region both prior to, during and after the orogeny.

Furthermore, several lines of evidence suggest that elevated temperatures in this orogen also relate to heat production by radiogenic decay of fertile sediments buried (subducted) during the orogenic crustal thickening. One is the fact that melting occurred in the middle crust at 20-25 km depth, which favors a mid-crustal heat source. Another is the large volumes of potentially fertile sedimentary rocks in the Macaúbas basin, and the observation that large amounts of peraluminous melt was produced from metasedimentary midcrustal rocks, for instance in the Carlos Chagas anatectic domain (Cavalcante et al., 2013). In general, heat production by radiogenic decay of buried sedimentary rocks in collision zones leads to high temperatures (≥ 700 °C) in substantial volumes of the middle crust after ~20 m.y. of collision (England and Thompson, 1986; Jamieson et al., 1998; Sandiford and McLaren, 2002; Faccenda et al., 2008), generating a hot orogen with associated profound crustal weakening (Vanderhaeghe, 2009; Jamieson and Beaumont, 2013). This is the scenario suggested for the crust underlying the Tibetan plateau (Nelson et al., 1996; Vanderhaeghe and Teyssier, 2001; Zhang et al., 2004), where the present crustal thickness is at least doubled, where middle to lower crustal temperatures are well above 700 °C (Klemperer, 2006), and where melting started some 20-25 m.y. after the collision and is still ongoing 30 million years later (Jamieson et al., 2011). A schematic illustration of such an orogenic evolution is shown in Figure 10. Transferred to the Neoproterozoic Araçuaí–West Congo orogen where melting was ongoing at around 600 Ma (Cavalcante et al., 2018), this implies that orogenic thickening started before 620 Ma. Recent dating of recrystallization of detrital zircon at ~630-625 Ma in metasedimentary rocks in the Macaúbas basin, interpreted as a regional metamorphic event in a convergent regime (Schannor et al., 2018) is consistent with such a model.

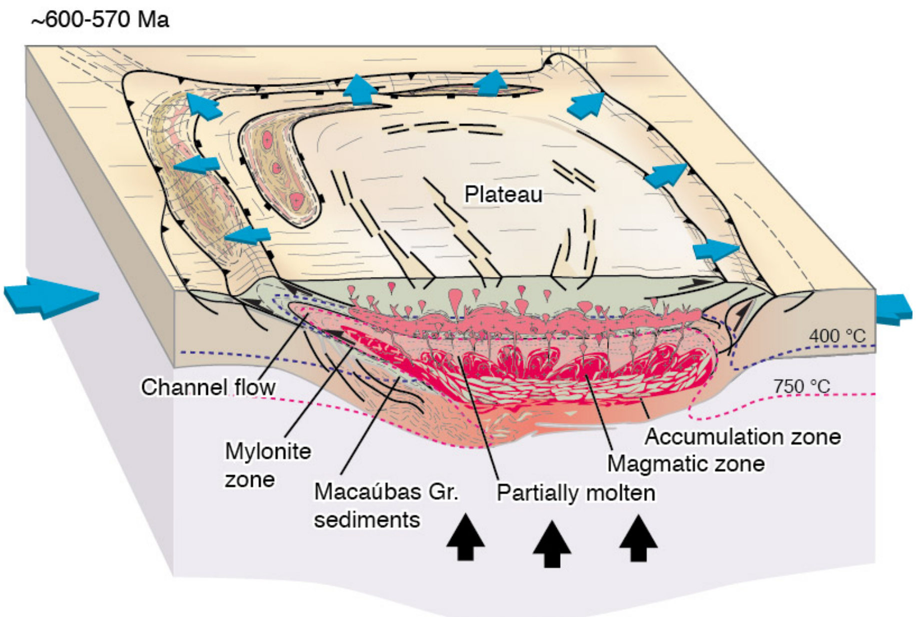

Figure 10. *Idealized illustration of the formation of a hot orogen at the time of maximum crustal thickness with extensive partial melting, where the molten crust is starting to flow toward the foreland (channel flow) which again leads to plateau uplift and exhumation of the lower crust (black arrows). Modified from Vanderhaeghe (2009).*

The Variscan and Grenville orogens represent ancient examples explained by similar processes (Vanderhaeghe et al., 1999; Turlin et al., 2018), and both the Variscan, Grenville and Himalayan orogens show evidence of lateral channel flow or extrusion (Fig. 4) of mid-crustal material under a stronger upper crust that stretched during slow gravity-driven orogenic spreading. This model fits the Araçuaí–West Congo orogen and its high temperatures and extensive partial melting, and estimates of temperature and magma viscosity from the anatectic part of the Araçuaí orogen (Cavalcante et al., 2014). Important evidence of flow of partially molten middle crust is based on structural mapping based on outcrop observations and magnetic fabrics determined by the AMS (Anisotropy Magnetic Susceptibility) method in the anatectic core of the orogen (Carlos Chagas anatectic domain; Cavalcante et al., 2013). This work shows a magmatic-state middle crustal flow pattern consistent with gravitational collapse of the upper and middle crust and the more or less radial top-to-foreland kinematics shown in Fig. 3. In this scenario gneiss domes tend to form, as seen in eroded hot orogenic belts as well as in numerical models of hot orogenic settings (e.g., Vanderhaeghe and Teyssier, 2001; Rey et al., 2009, 2010). The fluctuating structural pattern of mostly low-angle fabrics in the anatectic parts of the Araçuaí orogen presented by Cavalcante et al. (2013) is consistent with deep sections through the roots of such core complexes (e.g., Vanderhaeghe, 2009). Channel flow would also cause northward flow of high-grade rocks toward the São Francisco foreland (the bridge region), explaining the relatively short distance between the hot core of the orogen and its northern termination (Figs. 3 and 6). We do not see evidence for vertical extrusion of light lower crustal hot and partially molten material, such as suggested for the Variscan orogen by Schulmann et al. (2008, 2014). Such overturning produces steep foliation and lineation patterns overprinted by low-angle fabrics related to subsequent lateral extrusion, while the hot central part of the Araçuaí-West Congo belt exhibits low-angle fabrics and in-situ middle-crustal melting. Some melt is likely to have come from the lower crust and the mantle, but the majority of melt at the present mid-crustal erosion level appears to have formed within the middle crust, since we do not observe any intrusive relationships (Cavalcante et al., 2013).

Plateau collapse related to lateral flow of the middle crust from ~600 Ma implies the existence of an extending upper crustal orogenic lid that is now removed by erosion. Some of the upper crust is preserved in the Ribeira and Dom Feliciano belts to the south, where a system of middle Ediacaran (~600 Ma) to early Cambrian rifts have been mapped (Almeida et al., 2010, 2012). A collapsing hot orogen model provides a viable explanation for the formation of these upper crustal rift basins, and this situation is consistent with the constant crustal thickness reflected by P-T data (Fig. 6).

8. Final discussion and concluding remarks

The partly confined geometry of the Araçuaí–West Congo orogen limits the amount of shortening across the orogen to maximum ~500 km, excluding models that call for ~50 m.y. of arc magmatism and associated subduction of oceanic crust. Instead, a hot orogenic model is favored, where heating during crustal thickening contributed to extensive partial melting from ~600 Ma, implying initiation of crustal thickening before 620 Ma. This is close to the time (~630 Ma) of collisions along the north, west and south margins of the São Francisco craton (Fig. 2), which probably caused the shortening of the Macaúbas basin that lead to the formation of the Araçuaí–West Congo orogen. The hot orogeny model outlined above involves internal heat production by radiogenic decay of buried rocks and sediments. Crustal heat production from radioactive elements (U, Th, K) can be a sufficient source of heat for partial melting of thickened continental crust (Jamieson et al, 1998; Sandiford and McLaren, 2002; Faccenda et al., 2008), combined with heating of the system during the pre-orogenic stretching of the crust across the Macaúbas rift basin. A syn-orogenic mantle heat source has also been suggested by several authors (Gradim et al., 2014; Bento dos Santos, 2015; Tedeschi et al., 2016). Regardless, explaining extensive magmatism and partial melting of crustal material by prolonged island arc development is incompatible with the confined nature of this orogen. In contrast, the hot orogen model can explain much of this melt generation within a framework of a softened nutcracker model. Such a model has been successfully used to explain other orogens, including the Himalayas, Grenville and Variscan orogens (Beaumont et al., 2010; Jamieson et al., 2011).

The ~500 km of shortening estimated across the orogen occurred during the first and main part of the orogenic history, until extensive partial melting was established at ~600 Ma (Cavalcante et al., 2018). Once extensive melting of the middle crust was established, slow collapse of the central parts of the orogen may have driven thrusting toward the foreland, at least until 570 Ma, which is the youngest age of melt crystallization reported from the anatectic core (Cavalcante et al., 2018). However, evidence of younger orogenic activity has been reported. For example, the Três Marias Formation, which contains 558 Ma old detrital zircons (Kuchenbecker et al., 2015; Alkmim et al., 2017), has been involved in thrusting. Furthermore, the age of the youngest syn-kinematic intrusive body dated in the Araçuaí belt is ~530 Ma (the Ibituruna syenite; Petitgard et al., 2009), whereas the oldest post-kinematic granite is ~520 Ma (Noce et al., 2000). These observations may indicate a young pulse of orogeny, possibly related to the ~540 Ma Cabo Frio orogeny reported from the eastern Ribeira belt and the southernmost West Congo belt (Schmitt et al., 2016; Monié et al., 2012). Continuous orogenic convergence for ~100 m.y. is considered unlikely, as it would accumulate too much shortening. We suggest that late-orogenic thrusting driven by a collapsing hot central part of the orogen should be further considered as more data accumulate.

To understand this unusual termination of a large orogenic belt in a confined cratonic environment requires dedicated and high-quality dating of melt crystallization in a wider part of the hinterland. Such data should be compared with results from direct dating of thrusting in the low-temperature foreland fold-and-thrust belt, for example by Ar/Ar dating of micas grown below the retention temperature (Oriolo et al., 2018 and references therein). Furthermore, better mapping of the thermal structure of the orogen and numerical modeling of both the thermal and kinematic aspects discussed in this paper would be beneficial, and orogenic strain should be estimated across the reactivated Paraminim aulacogen. A better separation of the pre-orogenic rift sequence and the syn-orogenic deposits would also enhance our understanding of the orogenic evolution. Finally, the geochemical aspects of the various melts and magmatic rocks in the orogen require closer attention to explore alternatives to a conventional magmatic arc interpretation, since there is an overlap between the crystallization ages of the Galiléia and São Vitor bodies (Rio Doce magmatic arc; 580 Ma) and the Carlos Chagas anatexite (~600-570 Ma). Several fundamental implications of the kinematic constraints of this orogen are pointed out here, but there is a need for critical evaluation of both data and models, and there is also a general lack of quantitative structural considerations in the existing literature. Future work with this in mind will undoubtedly reveal new details about this intriguing part of the Brasiliano/Pan-African orogenic system. The main conclusions that we have been able to draw from the currently available data are as follows:

-A non-rigid cratonic model along the lines presented by Alkmim et al. (2006) is qualitatively viable only if loosened up by an orogenic corridor that breaks the “cratonic bridge” between the São Francisco and Congo cratons.

-This softened nutcracker model is at odds with the widely published idea of 50 m.y. of subduction-related arc magmatism in the Araçuaí–West Congo orogen, which should be reconsidered.

-Our "softened nutcracker model" can only accommodate ~500 km of orogenic shortening, which is required to form the 60-65 km thick orogenic crust.

-This hot orogen involves extensive partial melting of the middle crust, explainable by radiogenic decay of fertile sediments and crustal heating during pre-orogenic lithospheric thinning.

-Its extensively molten middle crust is likely to have produced foreland-directed gravity-driven flow (spreading) that influenced foreland deformation. Hence, late foreland thrusting does not necessarily directly reflect convergence but also relates to plateau collapse.

-There is a need to better constrain the timing of deformation in the orogen by geochronologic methods, particularly the low-temperature foreland deformation.

Acknowledgments

This work was supported by FAPESP projects 2015/23572-5, 2014/10146-5 2013/19061-0 and 2010/03537-7, and by the strong incentive for Brazilian research and the public universities by former presidents Luiz Inácio Lula da Silva and Dilma Rousseff. Additional support was provided by the Meltzer Research Fund (University of Bergen). CC greatly appreciates discussions with Alain Vauchez, which introduced her into the hot orogen model during her PhD research. Reviews by Olivier Vanderhaeghe and an anonymous reviewer are greatly appreciated.

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