TECTONIC EVOLUTION OF AN EARLY CRYOGENIAN LATE-MAGMATIC BASIN IN CENTRAL MADAGASCAR

Costa, R.L.[1], Schmitt, R.S. [1,2*], Collins, A.S. [3], Armistead, S.E. [3,4,5], Gomes, I.V.[1, 2], Archibald, D.B.[6], Razakamanana, T. [7]

[1]Programa de Pós-graduação em Geologia, Universidade Federal do Rio de Janeiro (PPGL/UFRJ), CEP 21941-916, Rio de Janeiro, Brazil – raisa.ric23@gmail.com

[2]Departamento de Geologia, Instituto de Geociências, Universidade Federal do Rio de Janeiro, Av. Athos da Silveira Ramos, 274/ bloco J, sala 022 - Cidade Universitária – Rio de Janeiro – RJ – CEP 21941-909 – Brazil – schmitt@geologia.ufrj.br

[3]Tectonics and Earth Systems (TES), Mawson Geoscience Centre, Department of Earth Sciences, The University of Adelaide, SA 5005, Australia – alan.collins@adelaide.edu.au (@geoAlanC)

[4]Geological Survey of Canada, 601 Booth Street, Ottawa, ON, K1A 0E9 Canada – sarmistead@laurentian.ca (@geoSheree)

[5]Mineral Exploration Research Centre, Harquail School of Earth Sciences, Laurentian University, Sudbury, ON, P3E 2C6 Canada

[6]Department of Earth Sciences, St. Francis Xavier University, Physical Sciences Complex, 5009 Chapel Square, Antigonish, NS, Canada, B2G 2W5 – darchiba@stfx.ca (@darchibald12)

[7]Département des Sciences de la Terre, Université de Toliara, Toliara, Madagascar – razakamananat@yahoo.fr

* Corresponding author at: Departamento de Geologia, Instituto de Geociências, Universidade Federal do Rio de Janeiro, Av. Athos da Silveira Ramos, 274/ bloco J, sala 022 - Cidade Universitária – Rio de Janeiro – RJ – CEP 21941-909 – Brazil . Tel.: +55 21 996388859 - E-mail: schmitt@geologia.ufrj.br

This manuscript is not published and is under review in the Journal of African Earth Sciences. Please note that subsequent versions of this manuscript will have slightly different content. If accepted, the final version of this manuscript will be available via the 'Peer-reviewed Publication DOI' link on the right-hand side of this webpage.

*corresponding author: Renata da Silva Schmitt – schmitt@geologia.ufrj.br

Highlights:

- Early Cryogenian basin with volcanoclastic contribution in Central Madagascar
- Three distinct tectonic models for Late Tonian-Early Cryogenian
- Extensional intracontinental magmatic setting related to an outboard subduction
- A transform continental setting resolves the tectonic Cryogenian paucity
- Continental collision dated with 550 Ma metamorphic rims on zircons

ABSTRACT

Central and southern Madagascar comprise a number of distinctive Archaean crustal blocks (the Antongil-Masora and Antananarivo domains) overlain by Proterozoic supracrustal sequences, preserved in the East African Orogen. Here, we present U–Pb and Lu–Hf isotopic data for two supracrustal units from detrital and metamorphic zircon grains. The lower sequence is comprised of quartzite and calc-silicate units with a major Palaeoproterozoic detrital zircon source and a minor Archaean contribution with a maximum depositional age of ca. 1780 Ma. This sequence reflects a stable shelf sedimentation within the Antananarivo Domain and is correlated with the Itremo Group. U–Pb and Hf data are equivocal in determining the direct sources for the Archaean and early Palaeoproterozoic detrital zircon grains. However, the abundant ca. 2.3-1.8 Ga detrital grains are correlative with the Congo-Tanzania-Bangweulu Block and as these are close to the inferred age of the Itremo Basin, these are interpreted to be single cycle detritus. This implies that the Congo-Tanzania-Bangweulu craton was close to central Madagascar at ca. 1.8-1.6 Ga and the lower sequence would correspond to an originally contiguous late Palaeoproterozoic to early Mesoproterozoic sedimentary basin across central

Madagascar. The upper metasedimentary unit has contrasting detrital sources and is represented mostly by biotite-plagioclase paragneiss, with an inferred psammitic protolith interleaved with volcanic/subvolcanic andesitic/rhyolitic dikes. The predominant Tonian-aged population (ca. 860-710 Ma) are igneous zircon grains with ε_{Hf}(t) values varying from -15.1 to -29.2 and T_{DM} Hf model ages between ca. 3.4 and 2.6 Ga. These grains were derived from the ca. 850-750 Ma Imorona-Itsindro magmatic suite. Their Neoarchaean-Palaeoproterozoic cores are interpreted as xenocrysts, reinforcing that the Imorona-Itsindro magmatism has a prominent continental reworking component. The probable tectonic setting for this Early Cryogenian sedimentary basin would represent a transition from an intra-arc to an intracontinental setting related to an outboard subduction, partially jammed at ca. 710 Ma due to the subduction of a ridge-transform system. The analogue would be the western US, where the Basin and Range region corresponds to a wide rift associated with a major mantle thermal anomaly. The absence of geological units and structures between ca. 720 and 635 Ma in central Madagascar corroborate with this model for a transition to a transform continental setting. The pre-Gondwana amalgamation convergence in the Ediacaran-Cambrian, that deformed and metamorphosed all units in central Madagascar units, is accounted for by ca. 550 Ma metamorphic rims on zircon grains from the quartzites in the Itremo Group.

Keywords

Imorona-Itsindro Suite, late-magmatic continental basin, Cryogenian, detrital zircon, Central Gondwana

1. Introduction

The East African Orogen (EAO) involves a collection of Neoproterozoic microcontinents and arc terranes lodged between older cratonic domains that coalesced during the final assembly of supercontinent Gondwana (Collins and Pisarevsky, 2005; Schmitt et al., 2018), extending from the Arabian Peninsula along eastern Africa, southern India, Sri Lanka and into Antarctica (Grantham et al., 2011; Jacobs and Thomas, 2004; Jacobs et al., 1998). This orogenic system resulted in the closure of the Mozambique Ocean between the eastern and western Gondwana blocks during the Neoproterozoic-Cambrian transition (Merdith et al., 2017; Tucker et al., 2014). The record of convergence between the Greater Dharwar Craton and the Congo/Tanzania/Bangweulu blocks (southern India and eastern Africa, respectively), is preserved within the Malagasy basement (Fig. 1a; Collins, 2006; Stern, 1994, 2002).

This basement in central and southern Madagascar comprise a number of distinctive Archaean crustal blocks (Antongil-Masora and Antananarivo domains) that are overlain by several Proterozoic sedimentary and volcano-sedimentary sequences (Fig. 1b), metamorphosed during the Neoproterozoic-Cambrian (De Waele et al., 2011; Tucker et al., 2007). The metasedimentary units include the Itremo, Ikalamavony, Ambatolampy, Maha, Manampotsy, Sahantaha, Andrarona and Molo groups (Bauer et al., 2011, Archibald et al., 2015; Cox et al., 2004a; De Waele et al., 2011). These supracrustal rocks record hundreds of millions of years of convergent tectonics, as pre-, syn-, late- and post-tectonic basins.

Although these metasedimentary sequences have U–Pb provenance studies (Cox et al., 1998, 2004a; De Waele et al., 2011; Boger et. al. 2014), many questions remain regarding the relationship of these sedimentary and volcanic rocks with two well described magmatic arcs: the Dabolava Suite that formed on a juvenile intra-oceanic arc at ca.1000 Ma (Archibald, et al., 2018; CGS, 2009a, 2009b; Tucker et al., 2007), and the Imorona-Itsindro Suite that formed in a continental arc setting at ca. 850-750 Ma (Archibald et al., 2016, 2017; Boger et al. 2014, 2015; Handke et al., 1999). In addition, the period between ca. 700 Ma, the age of the last preserved pre-collisional magmatic unit, and ca. 575 Ma, the age of the main metamorphic and deformational event related to the EAO within Madagascar, is still poorly understood in terms of the tectonic regime.

Here, we present new geological and geochronological data for two supracrustal units from the Ikalamavony region in the west-central part of Madagascar, including U–Pb geochronology and Lu-Hf isotopes on detrital and metamorphic zircon grains. We examine the existence of an early Cryogenian sedimentary basin with an important volcanoclastic contribution that developed coevally with Imorona-Itsindro suite magmatic activity. We also investigate a pre-Neoproterozoic sedimentary basin that is tectonically juxtaposed with the Early Cryogenian sequence.

2. Tectonic setting and geology

Precambrian tectonic domains make up the Malagasy basement (Fig. 1). The oldest blocks, the Antongil and Masora cratons, are located on the eastern coast and consist of Paleoarchaean to Mesoarchaean ortho- and paragneisses (ca. 3330-3140 Ma) with Neoarchaean granitic and metasedimentary rocks (ca. 2540-2500 Ma; BGS, 2008; Schofield et al., 2010; Tucker et al., 1999, 2011b). Nonetheless, two key differences between the cratons are acknowledged. The first is the ensuing magmatic activity recorded at ca. 2350 Ma and ca. 2150 Ma that occurs only in the Antongil Craton (Schofield et al., 2010; Bauer et al., 2011). The second is the Palaeoproterozoic supracrustal unit that overlies the Masora Craton, the so-called Maha Group (maximum depositional age ca. 1700 Ma), cropping out in the eastern portion of Madagascar (De Waele et al., 2008, 2011). Additionally, Bauer et al. (2011) reported a sequence of low-grade Proterozoic sediments overlying the Antongil Craton (Fig.1a), the Andrarona Group. The basal component of the Andrarona Group, the Ankavia Formation, has a maximum depositional age of 2355 ± 11 Ma (2σ), while the upper formation within this group – the Andratany Formation – has euhedral zircons of interpreted volcanic origin yielding an age of 1875 ± 8 Ma (2σ). According to Tucker et al. (1999, 2014), these rocks in the Antongil and Masora cratons are not considered to be affected by Neoproterozoic tectonothermal events and they correlate with the Greater Dharwar Craton, only separating during Gondwana break-up at ca. 85 Ma (Storey et al., 1995).

The central highlands of Madagascar comprise the Antananarivo Domain (Fig. 1b; Collins, 2006; Kröner et al., 2000). It consists of Neoarchaean stratified gneisses of the Sofia and Vondrozo groups, to the north and the south, represented by metasedimentary and metavolcanic rocks, intruded by the granitoids of the Siderian

Betsiboka Suite (ca. 2.52–2.49 Ga; BGS-USGS-GLW, 2008; Collins, 2006; Collins et al., 2003a; Kröner et al., 2000; Roig et al., 2012, Tucker et al., 2014).

The Antananarivo cover units Orthogneisses of the Antananarivo Domain are interleaved with a supracrustal package, the Ambatolampy Group (Figs. 1b and 2), consisting of graphite-bearing sillimanite ± mica schist/paragneiss with abundant quartzite beds (BGS-USGS-GLW, 2008). Limited U–Pb geochronology on detrital zircon was used to suggest a maximum depositional age of 1056 ± 54 Ma; with most detrital zircon dated between ca. 2740 and ca. 1800 Ma (Figs. 1b and 2; BGS-USGS-GLW, 2008). In contrast, Archibald et al. (2015) analysed multiple samples and obtained an age of 1836 ± 25 Ma for the youngest detrital zircon from the Ambatolampy Group. They proposed that the ca. 1.0 Ga sample, dated previously, was part of the Manampotsy Group and that the rest of the mapped Ambatolampy Group is very similar and related to the Itremo Group, forming a contiguous Palaeo- to early Mesoproterozoic basin (Fig. 1b and 2). The Maha and Sahantaha groups also overlie Archaean rocks and have comparable detrital zircon ages to the Itremo Group (Archibald et al., 2015; Cox et al., 2004a; De Waele et al., 2011).

The Manampotsy Group, between the Antongil-Masora cratons and the Antananarivo domain, is a package of supracrustal rocks with pods of mafic-ultramafic rocks with abundant intrusions of tonalitic to granitic bodies (Fig. 1b - BGS-USGS-GLW, 2008; Key et al., 2011). The younger population of U–Pb ages on detrital zircon grains ranges between ca. 840 Ma to ca. 790 Ma but abundant Archaean grains also occur (Collins et al. 2003a; Tucker et al. 2011b). This indicates that sedimentary protoliths were deposited from recently formed crustal sources as well as from the proximal basement. This group has been interpreted as representing volcanic-arc related sediments, deposited within an active forearc margin basin, representing the Betsimisaraka Suture (Collins et al., 2006; Fig. 1b) (BGS-USGS-GLW, 2008; Key et al., 2011; Collins et al. 2003b). This unit is also interpreted as deposited in an intracontinental basin in the Tonian produced by continental extension lasting more than 80 m.y (Tucker et al., 2011b).

The Itremo Group is a sedimentary sequence that overlies the Antananarivo Domain and is bound on its east by the Betsileo Shear Zone (Fig.1b - Collins et al. 2000). It also occurs as an NNW-SSE aligned belt separating the Antananarivo from the Ikalamavony Domain to the west (Figs. 1b and 2). The Itremo Group includes a series of metasedimentary and metabasic rocks, plus gneisses with late Archaean to Palaeoproterozoic sources (Cox et al. 1998; 2004a; Fernandez and Schreurs, 2003; Fitzsimons and Hulscher 2005; Armistead et al. in review). It consists of quartzite, metadolomite and metapelite. Detrital zircon grains from quartzites present prominent age peaks at ca. 1850 Ma and ca. 2500 Ma (Cox et al., 1998, 2004b, Fitzsimons and Hulscher, 2005) with a maximum depositional age of ca. 1700 Ma (Cox et al., 2004a; Fernandez and Schreurs, 2003). In addition, basic volcanic rocks, interlayered with metapelites and above the carbonate units (Cox et al., 1998), are intruded by ca. 850–750 Ma plutonic rocks of the Imorona-Itsindro Suite (BGS-USGS-GLW, 2008; CGS, 2009a, 2009b; Cox et al., 2004b; Fernandez and Schreurs, 2003; Collins et al. 2003c; Tucker et al., 2007). The Itremo Group was deformed into early recumbent folds, that either pre-date the Imorona-Itsindro Suite (Collins et al. 2003c) or formed synchronously with magmatism (Armistead et al., 2020), and refolded into late upright folds either progressively (Armistead et al., 2020) or much later during the Cryogenian–Cambrian (Collins et al. 2003c; Tucker et al. 2007).

2.2. The Ikalamavony Domain

The NNW-SSE Ikalamavony Domain contains metavolcanosedimentary rocks of the Ikalamavony Group intruded by, and derived from, the broadly contemporaneous ca. 1.0 Ga Dabolava suite (Figs. 1b and 2 - CGS, 2009a, 2009b; Cox et al., 2004b; Tucker et al., 2007, 2011a, 2011b). The Ikalamavony Group is mainly composed of metapelite, metapsammite and metavolcanic rocks, interpreted as a volcanosedimentary marginal sequence related to a magmatic arc environment with a Stenian-Tonian age (ca. 1080–980 Ma; Archibald et al., 2018; Tucker et al., 2011a). In addition, its detrital zircon age spectra are distinct from the Itremo Group (Tucker et al., 2011a). The Ikalamavony Group shows a prominent ca. 1.0 Ga source, linking it to the Dabolava Suite, with only a minor Palaeoproterozoic contribution (Archibald et al., 2018). Both the Dabolava Suite and the Ikalamavony Group are interpreted as forming in a subduction-related ca. 1000 Ma island arc setting in the Mozambique Ocean, outboard of the Archaean to early Palaeoproterozoic shield of Madagascar (Fig. 1b - Archibald et al., 2018). Then the Ikalamavony domain accreted to the Malagasy basement before the intrusion of the ca. 850–750 Ma Imorona-Itsindro Suite (Archibald et al., 2018; Fig. 1b).

Another Neoproterozoic metasedimentary unit, the Molo Group, was suggested by Cox et al. (2004a) based on detrital zircon age populations of three samples (quartzite, biotite gneiss, hornblende metapsammite). It has a late Meso- to Neoproterozoic detrital zircon signature, with three major age peaks at ca. 1090–1030 Ma, ca. 840–780 Ma and ca. 690–630 Ma, and with only minor grains that are older than ca. 1090 Ma. The depositional age of the Molo Group is constrained to between 613 ± 9 Ma (youngest detrital zircon), and 556 ± 10 Ma (average age of the metamorphic overgrowths; Cox et al. 2001, 2004a).

2.4. The SW terranes

The southwestern part of Madagascar there are four distinct tectonic domains (Boger et al., 2014, 2019): the Anosyen, the Androyen, the Graphite and the Vohibory (Fig. 1b). The Anosyen Domain is the most extensive and is comprised of two distinct groups of paragneissic rocks: (a) the Iakora Group—pelitic and calc-silicate gneisses with terrigenous or chemical sedimentary origin; and (b) the Horombe Group—extrusive volcanics, or pyroclastic/epiclastic volcanic sedimentary rocks (Boger et al., 2014). While the siliciclastic sedimentary rocks of the Iakora Group encompass mostly ca. 2400–1600 Ma aged detrital zircon with a modal age peak at approximately 1850 Ma, the Horombe Group has peraluminous (ca. 1800-700 Ma, Tôlanaro Formation) and metaluminous (ca. 820–760 Ma, Benato Formation) rocks with ages similar to the Imorona–Itsindro Suite (Kröner et al., 1999; Tucker et al., 2011a; Collins et al., 2012; Boger et al., 2014).

Published age data from the Androyen Domain are limited. Possible granitic protolith ages of ca. 2200–1800 Ma (Tucker et al., 2011a, 2014), considerable Palaeoproterozoic detritus (Collins et al. 2012) and a spread of ages between ca. 620 and ca. 520 Ma are interpreted to reflect two closely spaced (ca. 630–600 Ma and 580–520 Ma) periods of high-temperature metamorphic zircon growth (Tucker et al., 2011a; Boger et al., 2015). This domain also includes two groups of stratified units (Mangoky and Imaloto groups), which are intruded by ca. 930–910 Ma magmatic rocks from Ankiliabao Suite (GAF-BGR, 2008b).

The narrow and elongate Graphite Terrane (Boger et al., 2019) have gneisses that hosts the Ankiliabo Suite, therefore predating by several hundreds of millions of years the formation of the intermediate to felsic protoliths of the Vohibory Domain, to the west (Fig. 1b). In addition, felsic magmas in the Graphite Domain are originated from shallower crustal source rocks (Boger et al., 2019). Strongly positive initial εNd and relatively low age corrected 87Sr/86Sr(m) indicate that the gneisses of the Graphite Domain, similar to those of the Vohibory Domain, likely formed in an intra-oceanic environment (Boger et al., 2019).

The Vohibory Domain is represented by felsic and mafic orthgneisses, which are intercalated with paragneisses and marbles (Boger et al. 2019). Mafic orthogneisses are suggested to represent a blend of mid-ocean-ridge, back-arc and island arc basalt with inferred extrusion ages between ca. 850–700 Ma (Emmel et al., 2008; Jöns and Schenk, 2008), while younger felsic gneisses are dated between 670-630 Ma (Boger et al., 2015). In addition, the metasedimentary rocks yield a unimodal population of detrital zircon with an age range between ca. 900 Ma and ca. 750 Ma (de Wit et al., 2001; Emmel et al., 2008; Jöns and Schenk, 2008; Collins et al. 2012). The main metamorphic and deformational phase occurred at ca. 630–600 Ma, with a minor impacts of a 580-520 Ma orogenic event (Emmel et al., 2008, Jöns and Schenk, 2008, Collins et al., 2012; Boger et al., 2015).

2.4. The Imorona-Itsindro magmatic unit

The Antananarivo (including the Itremo Group) and Ikalamavony domains, plus the Masora Craton, are intruded by Tonian-aged, weakly peraluminous granitic and gabbroic rocks of the Imorona-Itsindro Suite (Fig. 1b and 2 - Archibald et al., 2016, 2017; Boger et al., 2015; Zhou et al., 2018). These rocks are interpreted as subduction related (Handke, et al., 1999; Archibald et al. 2016; 2017), with U–Pb zircon emplacement ages between ca. 850 Ma and ca. 750 Ma, but major magmatic activity from ca. 800 Ma to 780 Ma (Archibald et al., 2016; Handke et al., 1999;

McMillan et al., 2003; Tucker et al., 2011a; Kröner et al. 2000). Its origin is hypothesized as a result of the convergence between the Archaean nuclei of the Antongil-Masora cratons (Dharwar Craton – India; Fig 1a) and the Antananarivo (Malagasy basement) domain, that resulted in the generation of Andean-type magmatism due to subduction and the closure of the Neoproterozoic Mozambique Ocean (BGS-USGS-GLW, 2008; Collins, 2006; Collins and Pisarevsky, 2005; Collins and Windley, 2002; Kröner et al. 2000). The subduction polarity was originally proposed to be east-dipping, from a trench located to the west and subduction under the Antananarivo Domain (Handke et al. 1999). A number of authors then reinterpreted this to indicate west-dipping subduction from a trench located along the proposed Betsimiaraka Suture separating the Antananarivo Domain from the Antongil Domain (Fig. 1b - Kröner et al. 2000; Collins and Windley, 2002; Collins et al. 2003a; Collins and Pisarevsky, 2005; Fitzsimons and Hulscher 2004; Archibald et al. 2016; 2017; Armistead et al. 2018; 2020). In contrast to the subduction origin hypotheses, Zhou et al. (2015) analysed the chemistry of many fewer intrusions than Archibald et al. (2016; 2017), but suggested that instead, the suite represents melting of the so-called Greater Dharwar Craton during mantle plume induced rifting.

The southwestern tectonic domains have no record of the ca. 850–750 Ma Imorona-Itsindro Suite, but age equivalent volcanic or intrusive rocks are present in the Anosyen Domain (Tucker et al. 2011; Boger et al. 2014). Boger et al. (2014) interpreted the Imorona-Itsindro magmatic event as related to a subduction zone located to the west, supporting the original interpretation of Handke et al. (1999), but contrasting with interpretations of others (Collins, 2006; Collins and Pisarevsky, 2005; Collins and Windley, 2002; Kröner et al., 2000). Nevertheless, the ca. 550 Ma suture zone is placed between the Androyen and the Anosyen domains (Boger et al., 2015, 2019).

2.5. The Ediacaran–Cambrian tectonic overprint

The entire Antananarivo Domain was deformed and metamorphosed up to granulite facies conditions (e.g. Cenki-Tok et al., 2015) and intruded by Ediacaran-Cambrian syn- to late-tectonic granitic dykes and plutons of the Ambalavao and Maevarano plutonic suites (Fig. 1b and 2; Archibald et al., 2019; Paquette and Nédélec, 1998; Boger et al., 2014; Horton et al., 2016). This major tectonic event was related to the final Gondwana amalgamation, when the Malagasy shield was the site of east-directed translation of crystalline nappes, high-grade metamorphism and widespread magmatism (Tucker et al., 2011b; Collins et al. 2003c). The post-orogenic plutonic suites intruded between ca. 580 and 520 Ma (Archibald et al., 2019; Goodenough et al., 2010). Goodenough et al. (2010) interpreted Maevarano Suite as the final stages of an extensional collapse that followed asthenospheric upwelling.

Two events of deformation and high-temperature metamorphism occur in southwestern Madagascar: one at ca. 620–600 Ma (recorded only in the Androyen and Vohibory domains) and a second at ca. 580–520 Ma that is widespread in all domains (Ashwal et al., 1999; de Wit et al., 2001; Jöns and Schenk, 2011). The second event is accompanied by the emplacement of the Ambalavao Suite (Fig. 1b

- Ashwal et al., 1999; de Wit et al., 2001; Emmel et al., 2008; Jöns and Schenk, 2008, 2011; Tucker et al., 2007).

3. Method

3.1. Zircon separation and imaging

Eight samples of approximately 2.5 to 3 kg were collected for petrography and geochronology. The selected samples were crushed in a jaw crusher, disk mill grinder, and sieved to zircon fraction between 79–425 µm. After sample crushing, heavy minerals were separated by manual panning and then using magnet, heavy liquids technique (methylene iodide – density: 3.32g/cm^3) and Frantz Isodynamic Magnetic Separator. These steps were performed at the laboratories of University of Adelaide, Australia. Four samples yielded sufficient zircon grains and were hand-picked from the heavy mineral fractions. Zircon grains were mounted in a circular epoxy resin, polished and then imaged in reflected light (RL). Cathodoluminescence (CL) images were obtained to investigate the internal structure of the zircon grains, acquired by Philips XL40 Scanning Electron Microscope equipped using a tungsten filament electron source and a Gatan CL detector attached for high-resolution imaging at Adelaide Microscopy.

3.2. Zircon U-Pb geochronology

Zircon U-Pb geochronology was performed at Adelaide Microscopy, University of Adelaide, by Laser Ablation Inductively Coupled Mass Spectometry (LA-ICP-MS) using an 7500cs ICPMS unit coupled to a New Wave 213 nm Nd-YAG laser using a spot size of 30 µm and frequency of 5 Hz. U-Pb-Th isotope fractionation was corrected using GEMOC GJ-1 zircon ($^{207}Pb/^{206}Pb$ age = 607.7 ± 4.3 Ma, $^{206}Pb/^{238}U$ age = 600.7 ± 1.1 Ma and $^{207}Pb/^{235}U$ age = 602.0 ± 1.0 Ma; Jackson et al., 2004).

Data were processed using GLITTER software (Griffin et al., 2008). Concordia diagrams were produced using IsoplotR (Vermeesch, 2018) and Kernel Density Estimates (KDE) were produced in R using a bandwidth of 25 Ma. Data within 10% of concordance are included (Fig. 6). Furthermore, $^{207}Pb/^{206}Pb$ and $^{206}Pb/^{238}U$ ages were used for zircon grains older and younger than 1.3 Ga, respectively. Probability density plots from figure were constructed using ISOPLOT 4.15 software in Microsoft Excel (Ludwig, 2012) with discordance equal or < 10%.

3.3. Zircon Lu-Hf analysis

Seventy-nine Lu-Hf isotopes were analysed on the Thermo-Scientific Neptune Multi-Collector ICP-MS with an attached New Wave 193 Excimer laser ablation system at the University of Wollongong. A beam diameter of 50 µm was used. Typical ablation times were ~ 45 seconds using a 5 Hz repetition rate and an intensity of ~4.40 J/cm^2. Zircon grains were ablated in a helium atmosphere that was then mixed with argon upstream of the ablation cell.

Plešovice and Mudtank zircon standards were analysed before and during the analysis of unknowns to assess instrument stability and performance. Twenty Plešovice standard analyses were made and yield an average of 0.282471 ± 0.000042, which is within uncertainty of the published value of 0.282482 ± 0.000013 by (Sláma et al., 2008). Twenty Mudtank standard analyses were made and yield an average of 0.282505 ± 0.000047, which is within uncertainty of the published value of 0.282507 ± 0.000006 (Woodhead et al., 2004).

Zircon data were reduced using Iolite and normalised to $^{179}Hf/^{177}Hf$ = 0.7325. Values for $^{176}Hf/^{177}Hf_{CHUR(t)}$ were calculated using modern $^{176}Hf/^{177}Hf$ = 0.282785 (Bouvier et al., 2008), modern $^{176}Lu/^{177}Hf$ = 0.0336 (Bouvier et al., 2008), and ^{176}Lu

decay constant of 1.865×10^{-11} year^{-1} (Scherer et al., 2001). Values for crustal model ages (T_{DMC}) were calculated using a ^{176}Lu decay constant of 1.865×10^{-11} year^{-1} (Scherer et al., 2001), modern $^{176}Hf/^{177}Hf = 0.28325$, modern $^{176}Lu/^{177}Hf = 0.0384$ (Griffin et al., 2002), and a bulk crust value of $^{176}Lu/^{177}Hf = 0.015$ (Griffin et al., 2002). Uncertainties for $\varepsilon_{Hf}(t)$ are calculated as the $^{176}Hf/^{177}Hf_{Sample}$ uncertainty converted to epsilon notation (i.e. $[(^{176}Hf/^{177}Hf_{2\sigma})/0.282785) * 10{,}000]$ and are reported at the 2σ level).

4. Results

4.1. Geology of the study area

The studied area is located 22 km southeast of the village of Ikalamavony (Fig. 2) and includes the contact between the Antananarivo and the Ikalamavony domains (Fig. 1b and 2). Geological mapping was performed at 1:25,000 scale (Fig. 3) and the geological units are described below.

4.1.1. Metasedimentary units

We identified three metasedimentary units, from bottom to top: 1- quartzite with interleaved (garnet-sillimanite) muscovite-biotite schists; 2- calc-silicate rocks, and 3- fine-grained plagioclase-biotite paragneiss.

The first and structurally lowest sequence comprises aligned NNE-SSW mountain ranges (Fig. 3 and 4a). Thick layers, up to 5 meters, of pure quartzite predominate on the bottom (Fig. 4a – samples DA13-039 and DA13-045 – see item 4.1.3 and 4.1.2, respectively). Towards the top, packages of thin layered quartzite beds, varying from 5 to 10 cm thick (sample DA13-036 – item 4.1.1), are interlayered with (garnet-sillimanite) muscovite-biotite-schists (Fig. 4b). The thin-layered quartzites are more micaceous than the thicker beds, with biotite and muscovite (Fig.

5a). Accessory minerals such as tremolite occur on layers near the contact with the upper calc-silicate unit (Fig. 5a).

The interleaved (garnet-sillimanite) muscovite-biotite-schists are weathered, rarely preserving garnet, mostly replaced by oxides (Fig. 4b). The mineral assemblage indicates sillimanite zone upper amphibolite facies metamorphism (Fig.5b and 5c).There is a metamorphic foliation (S_n), represented by mica orientation and stretched quartz grains, parallel to the primary bedding (S_0). This foliation locally transposes crenulation surfaces, which represent an earlier tectonic foliation S_{n-1} (Fig. 5b and 5c), attesting a polyphase ductile deformation event.

The compositional variety of this sequence reflects changes of the sedimentary protolith, considered here as primary bedding - S_0 (Fig. 4b). Although there is indication of shearing and transposed foliation, parallel to bedding, the distribution of facies indicates a fining upward sequence.

The second unit is a calc-silicate package structurally above the quartzites (Fig. 3), composed of dark green rocks (~10 meters thick; Fig. 4c) with intermittent beds of white marbles, up to 2 meters thick (Fig. 4d), and also impure quartzite, up to 1 meter thick (Fig. 4e). It is comprised of mainly calcium-rich minerals (Fig. 4c), such as diopside, amphibole, epidote, and accessory minerals quartz, biotite, plagioclase and titanite. The unit displays mylonitic foliation and a transposed crenulation cleavage. At the bottom, a green calc-silicate diopside gneiss is dominant (Fig. 4c). Towards the top, the rocks show intercalation of coarse-grained white marble beds, constituted mostly by recrystallized calcite (Fig. 4d). One 2-meter-thick bed of impure foliated feldspar-bearing quartzite (Fig. 4e).

The third supracrustal unit is a fine-grained homogeneous plagioclase-biotite-paragneiss (Fig. 4f and 4g - sample DA13-030 – item 4.1.4). It comprises layers of up to 15 meters thick (Fig. 4g), constituted by quartz, plagioclase and biotite (Fig. 5d). Accessory minerals are white micas, zircon and Fe-oxides. The protolith is interpreted as a psammitic rock that might have been a fine-grained arkose with plagioclase. The mineral composition of this unit, plagioclase, quartz and mica, indicates that the sources were predominantly andesitic. The contact with the calc-silicate unit is concordant but the beds show high strain and stretching lineation that might indicate that it is a tectonic contact.

The three sedimentary-derived units present a continuous tectonic foliation dipping W-SW, parallel to compositional layering, with evidence of transposition (Fig. 5b and 5c) related to a down dip stretching lineation. Kinematic indicators suggest a top to E-NE movement (Fig. 2 and 3).

4.1.2. (Meta) igneous units

Three different (meta) igneous units were recognized in the studied area (Fig. 3): 1- metafelsite tabular bodies, 2- porphyritic orthogneiss, both correlated with the Imorona-Itsindro Suite; and 3- granite (correlated with Ambalavao Suite - Fig.1b).

The metafelsite bodies occur as dykes/sills, usually parallel and subparallel to the S_0/S_n, intruding the calc-silicate and the biotite-paragneiss units. They are up to 3 meters thick, and are shown on the geological map as dyke swarms and branches (Fig. 3 and 4h). They show fine-grained, occasionally porphyritic texture. The composition is rhyolitic, with quartz, microcline, biotite and muscovite. The dykes have quartz and K-feldspar as phenocrysts. The metafelsites present a tectonic foliation and some phenocrysts are also stretched (Fig. 4h). Archibald et al. (2016) dated one sample of this unit (DA13-029), interleaved with the calc-silicate unit, yielding a U-Pb zircon age of 828 ± 14 Ma that was interpreted as the emplacement age (see location on Fig. 3).

The porphyritic orthogneiss is recognized at the NE portion of the area (Fig. 3) and exhibits penetrative foliation marked by oriented biotite and stretched quartz. It is also correlated with the Imorona-Itsindro Suite according to the age obtained by Archibald et al. (2016) within the studied area (758 ± 10 Ma – sample DA13-020- also indicated on Fig. 3). The crosscutting relationships between the orthogneiss and the quartzitic unit are poorly constrained. Near the contact with the orthogneiss, mafic dykes with tectonic foliation crosscut the quartzite layers (Fig. 4i).

The third igneous unit is represented by a large pluton in the centre of the area (Fig. 3). These pinkish fine- to medium-grained granodioritic rocks are correlated with the Ambalavao Suite, and crosscut all other units (Fig. 4j). The rocks are commonly hololeucocratic and isotropic. Locally, some biotite foliation is recognized. The major body is dated with U-Pb zircon analysis from two samples, one at 550 ± 12 Ma and another at 539 ± 5.5 Ma (Archibald et al., 2019). A folded granitic dyke that cross cut the biotite-paragneiss was also dated at ca. 540 Ma (DA-13-031 – Figs. 3; Archibald, 2019) that is interpreted as the crystallization age (Fig. 4g). The composition and age of the deformed dyke is similar to the large pluton, therefore, we interpret this as the same magmatic event. The deformation pattern affected only the thin vein probably because of thickness and relation with the host rock (Fig. 4g). Although it is folded, it clearly cross-cuts the metamorphic foliation of the biotite-paragneiss.

4.2. U-Pb detrital zircon data

A total of 511 LA-ICP-MS U-Pb analyses of zircon cores and rims were randomly performed on 416 detrital zircon grains extracted from four rock samples (DA-13- 045; DA-13-039; DA-13-036; DA-13-030 – Fig. 3). Of all, 307 zircon grains yielded 294 concordant $^{207}Pb/^{206}Pb$ and/or $^{206}Pb/^{238}U$ ages, with 48 concordant ages collected from metamorphic rims (Supplementary table 1).

4.2.1. Sample DA13-045

This quartzite sample was collected on Route Nationale 7, near Ankaramena town in the Ambalavao District (Fig. 2), at the lower portion of the quartzite unit (item 2.2.1), near the contact with the Antananarivo Domain basement. It is a massive quartzite with locally more foliated layers with biotite and tourmaline. It grades towards the top, westwards, to calc-silicate layers. A quartzite in this outcrop was also sampled and dated by Tucker et al. (2011a; sample MJY-08-55). They obtained for 70% of the populations ages modes at 2.55-2.40 Ga, 2.7 Ga and 2.9 Ga. Thirty per cent of their sample provided age modes at 2.1-2.0 Ga and 1.8 Ga. Fifteen concordant rims yielded metamorphic ages between 550 and 450 Ma.

Our sample shows zircon grains with grain size length from ~120 to 250 µm with aspect ratios of 1:1 and 2:1 (length to width). The external grain morphology is similar to sample DA13-036, showing sub-rounded grains, with rounded/sub-rounded xenocrystic cores and homogeneous rims exhibiting a dark CL response (Fig. 7b) and most of the xenocryst cores show complex internal growth zoning.

Eighty analyses were performed on zircon cores and rims, of which 46 analyses yielded ages ≤10% discordant including 36 zircons cores and 10 metamorphic rims obtained from 44 zircon grains (Supplementary table 1).

$^{207}Pb/^{206}Pb$ ages vary between 3337 ± 37 to 1784 ± 26 Ma and the major population is within the interval ca. 2160–1820 Ma (15; 41% of the total concordant analyses) and interval ca. 2690–2400 Ma (13; 36% of the total concordant analyses) (Fig. 6b). The oldest zircon cores show $^{207}Pb/^{206}Pb$ Paleoarchaean ages (3337 ± 37 Ma and 3222 ± 21 Ma; Fig. 7b: #045-059 and #045-010) and 3 grains have Mesoarchaean ages (3073 ± 20 Ma, 3020 ± 23 Ma, and 2860 ± 22 Ma; Fig. 7b: #045-075, #045-009 and #045-006). The youngest detrital zircon grain has an age of 1784 ± 26 Ma.

4.2.2. Sample DA13-039

Sample DA13-039 was collected at the road from the town of Mangidy to Ikalamavony (Fig. 2). The sample is a coarse-grained quartzite from the lower sequence within the quartzite unit (described on item 2.2.1). Zircon grain lengths are between ~120 and 260 µm with aspect ratios of 1:1 and 2:1 (length to width). Sub-rounded grains, with sub-rounded xenocryst cores and homogeneous rims exhibiting a black CL response (Fig. 7c), characterize the zircon morphologies for this sample. Most of the xenocryst cores show complex growth zoning.

Ninety-three analyses of zircon cores and rims were performed, of which 71 analysis yielded ages with ≤ 10% discordance, including 60 zircon cores and 10 metamorphic rims from 65 zircon grains (Supplementary table 1). $^{207}Pb/^{206}Pb$ ages of concordant analysis vary from 3099 ± 52 to 2038 ± 23 Ma and the main contribution is ca. 2500 Ma. Three major intervals are observed: ca. 2750–2670 Ma (7 analyses; ~10% of the total), ca. 2500–2400 Ma (17 analyses; ~24% of the total) and ca. 2300–2030 Ma (28 analyses; 39% of the total concordant data- Fig. 6). Five concordant analyses on the oldest zircon grains present Mesoarchaean $^{207}Pb/^{206}Pb$ ages ranging from 3099 ± 52 Ma to 2940 ± 24, approximately 7% of all concordant data (Fig. 7c: #039-074, #039-042 and #039-015). The most significant population of detrital zircon have Siderian $^{207}Pb/^{206}Pb$ ages between ca. 2500 and 2330 Ma, ~25% of the data (18 concordant ages; Fig. 6). The youngest detrital zircon core yielded an age of 2038 ± 23 Ma (Fig. 7c: #039-007).

4.2.3. Sample DA13-036

Sample DA13-036 was collected at the road between the town of Mangidy and Ikalamavony, near the junction to Solila (Fig. 2). The sample is a quartzite from the structurally lowest sequence containing tremolite (described on item 2.2.1)**.** The detrital zircon grains have grain size length from ~ 110 to 230 µm with aspect ratios of 1:1 and 2:1 (length to width). Sub-rounded grains, with rounded/sub-rounded xenocryst cores and blackish homogeneous rims (Fig. 7a), characterize the external zircon morphology. Xenocryst cores show complex internal growth zoning.

One-hundred zircon cores and rims were analysed, of which 80 analyses yielded ages that are ≤ 10% discordant, including 57 zircon cores and 23 metamorphic rims from 65 zircon grains (Supplementary table 1). $^{207}Pb/^{206}Pb$ ages vary from 3358 ± 36 to 1814 ± 40 Ma and the most abundant ages are between ca. 2700 and 1900 Ma. The probability density plot presents two major age intervals at ca. 2250–1810 Ma (25 analyses, ~31% of the total concordant ages) and ca. 2680–2390 Ma (27 analyses, ~34% of the total) (Fig. 6). This latter group represents the most significant number of detrital zircon grain ages, with a peak of $^{207}Pb/^{206}Pb$ ages between ca. 2590–2390 Ma consisting of 26% of the total data (21 concordant ages).

There is one zircon core with a Palaeoarchaean $^{207}Pb/^{206}Pb$ age (3358 ± 36 Ma; Fig. 7a: #036-039) and another with Mesoarchaean age (2914 ± 21 Ma; Fig. 7a:

#036-094). The youngest detrital zircon core yielded an age of 1814 ± 40 Ma (Fig. 7a: #036-010). Only one detrital zircon grain has a Th/U ratio lower than 0.1 (0.03; Fig. 7a: #036-021), within a homogenous CL-domain presumably a metamorphic grain of 2.5 Ga.

4.2.4. Sample DA13-030

Sample DA13-030 was collected near Solila junction, south of the road from Mangidy to Ikalamavony (Fig. 2 and 3). The sample is a fine-grained homogeneous plagioclase-biotite paragneiss (described on item 2.2.1). The detrital zircon population shows grain size lengths varying from ~90 to 340 µm with aspect ratios of 1:1 and 2:1 (length to width). In terms of morphology, the detrital zircon grains are very distinct from the previous three samples. In general, external zircon morphology shows mostly poorly rounded grains and rarely rounded grains (Fig. 7d). Sometimes zircon grains have sub-rounded to rounded xenocryst cores showing complex growth zoning. However, the predominant internal structure is the typical igneous oscillatory zoning (Fig. 7d).

For this sample, 238 analyses on zircon cores and rims were performed, of which 147 analyses yielded ages that are ≤ 10% discordant, including 143 zircon cores and some igneous rims, plus four metamorphic dark rims. All data was obtained from 133 zircon grains. $^{207}Pb/^{206}Pb$ ages of concordant analyses vary from 2708 ± 40 Ma to 709 ± 11 Ma (Supplementary table 1). The most significant population presents Tonian–Cryogenian ages varying from ca. 860 to 710 Ma (Fig. 6d; 93 analyses; 65% of all concordant data). A minor contribution is represented by the interval ca. 2.7 to 1.8 Ga (Fig. 6d), with a Siderian peak of ca. 2.5–2.4 Ga (16 analyses; ~11% of all concordant data). The oldest detrital zircon is Neoarchaean in age (2708 ± 40 Ma; Fig. 7d: #030-163), that together with other 11 Neoarchaean zircon grains represent ~8% of the total concordant analyses. Some of these zircon grains have Th/U ratios lower than 0.1 (3 analyses; ~2% of all data) (Fig. 7d: #030-074, #030-219 and #030-220). These metamorphic detrital grains have ages of 1826 ± 39 Ma, 809 ± 11 Ma, and 753 ± 10 Ma respectively.

4.3. U–Pb metamorphic ages

Metamorphic rims were analysed in the detrital zircon population from all four samples, and 46 ages were obtained (Supplementary table 1). These concordant ages vary from 616 ± 8 Ma (Fig. 8c: #045-070 – oldest rim) to 504 ± 6 Ma (Fig. 8c: #036-009 – youngest rim). There is a significant distinction between the three samples from the quartzite unit (DA13-036, DA13-045 and DA13-039) and the sample from the paragneiss unit (DA13-030). The latter shows very thin dark metamorphic rims that surround the mostly prismatic detrital grains. These thin rims were difficult to date with the LA-ICP-MS technique (Fig. 7d and 8c). They have Th/U ratios between 0.03 and 0.08. Only three spots were analysed giving ages of: 614 ± 8 Ma, 598 ± 8 Ma and 581± 8 Ma (Fig. 8c).

The three samples from the quartzite unit show thick, dark metamorphic rims that truncate the internal morphology of the rounded to sub-rounded detrital zircon cores (Fig. 7a, 7b, 7c and 8c). We obtained 43 ages from the metamorphic rims of these samples constraining an interval between ca. 620 to 500 Ma, with weighted-average age of 550 ± 8 Ma (Fig. 8a and 8b). The Th/U ratios for these metamorphic rims are 0.18-0.09, for sample DA13-036; 0.01 and 0.04 for sample DA13-045, and 0.01–0.02 for sample DA13-039. Sample DA13-045 has one rim analysis with a Th/U ratio of 0.21 (Fig. 7b: 045-047).

4.4. Lu–Hf data

We performed 79 Lu–Hf in zircon analyses on the same four samples analysed for U–Pb (Supplementary table 2, Fig. 9). Most of the analyses were done on the detrital zircon grains and two analyses on metamorphic rims (Fig. 7a: #036-003 and #036-046). The spots were located on the same growth domain as the U-Pb analysis spot, characterized by the same internal growth pattern (Fig. 7).

The three quartzite samples have a similar Hf signature from their detrital populations (Fig. 9). The Palaeo- to Mesoarchaean populations (ca. 3.33–2.86 Ga), show juvenile signatures with ε_{Hf}(t) ranging from +6.8 to -2.1 and T_{DM} Hf model ages between 3.5 and 3.0 Ga. The Mid to Late Neoarchaean grains (ca. 2.75–2.60 Ga) show ε_{Hf}(t) values ranging from +7.4 to -7.0 and T_{DM} Hf model ages between ca. 3.6 and 2.7 Ga, suggesting both a juvenile and crustal contribution for the detrital zircons. In the transition between Archaean and Palaeoproterozoic through to the Siderian period (ca. 2.53 Ga – 2.34 Ga) the detrital population has ε_{Hf}(t) values ranging from +1.4 to -17.8 and T_{DM} Hf model ages between 2.9 and 3.9 Ga. This large variation from slightly juvenile to evolved Hf isotopic signatures are common in all three samples. The Rhyacian (2.3–2.05 Ga) to Orosirian (2.05–1.8 Ga) sources exhibit smaller variation, with U–Pb ages between ca. 2.29 to 1.78 Ga and ε_{Hf}(t) ranging from +2.0 to -12.9 with T_{DM} Hf model ages between 3.6 and 2.6 Ga. However, the Hf isotopic signature is more evolved, differing from the older Archaean grains.

Sample DA13-030 is the only sample that shows a Neoproterozoic detrital population. The grains analysed for Lu–Hf have U–Pb ages between ca. 1.01 Ga and 0.75 Ga. There is a strong variation on the nature of the sources with ε_{Hf}(t) ranging from -0.5 to -29.2 and T_{DM} Hf model ages between ca. 3.4 and 1.8 Ga. Two groups can be identified (Fig. 9). The older group with U-Pb ages from 1.01 Ga to 0.89 Ga is more juvenile. The second and youngest group of analysis, with U–Pb ages from 0.82 Ga to 0.75 Ga is much more evolved with high negative $\varepsilon_{Hf}(t)$ values. This sample also contains some older detrital grains. The Siderian sources (2.53–2.38 Ga) have $\varepsilon_{Hf}(t)$ values varying from +7.0 and -1.7 with T_{DM} Hf model ages between 3.1 Ga and 2.5 Ga, suggesting a more juvenile nature, similar to the quartzite zircons. The Mid to Late Palaeoproterozoic (2.05–1.83 Ga) detrital zircons present more crustal contribution with $\varepsilon_{Hf}(t)$ varying from -7.2 to -13.4 and T_{DM} Hf model ages between 2.9 Ga and 3.4 Ga, also showing similarities with the quartzite data.

5. Discussion

The detrital zircon data from the supracrustal rocks in this study define two different age groups: one group with predominant Palaeoproterozoic peaks and no Neoproterozoic contribution (DA13-036, DA13-039 and DA13-045); and one sample with a major early Neoproterozoic age population (Tonian; DA13-030).

5.1. Stratigraphic correlation

The three samples of the first group belong to the same stratigraphic quartzite unit that overlies the western limit of the Antananarivo domain (Figs. 1b and 2). Two samples (DA13-039 and DA13-045) were collected at the base, near the contact with the Antananarivo basement, and sample DA13-036 was obtained at the upper part of the sequence, where the quartzite is interlayered with (garnet-sillimanite) muscovite-biotite schists (Figs. 3 and 4b). This variation in rock type and, therefore, the sedimentary protolith, indicates a fining upward sequence. The U–Pb detrital zircon data show populations of ca. 2.5 Ga, ca. 2.1 Ga, with a minor contribution of ca. 3.3–2.9 Ga aged zircon and an absence of Neoproterozoic grains. This provenance pattern is similar to the (meta)sedimentary rocks from the Itremo Group (Fig. 10; Cox et al., 1998, 2004b; De Waele et al., 2011; Fitzsimons and Hulscher, 2005; Tucker et al., 2011a; Armistead et al. in review). Therefore, it indicates that the NNW-SSE-oriented belt of the Itremo Group may continue southwards following the main thrust that separates it from the crystalline rocks of the Antananarivo domain (Fig. 1b and 2).

Tucker et al. (2011a) dated a sample of the quartzite sequence from Ankaramena (Fig. 2), at the same outcrop where we sampled DA13-045 and obtained similar data. They also suggested that these quartzites may correlate with the Itremo Group and are separated from other feldspathic metasedimentary rocks of the Ikalamavony Domain by a significant disconformity. Alternatively, we could posit that these rocks are tectonic slices of the Itremo Group intercalated with the younger ca. 1.0 Ga paragneisses from the Ikalamavony Group during the late Neoproterozoic convergent tectonics. Other possibility is that these quartz-rich clastic protoliths are related to the Ikalamavony Group, and deposited in the Tonian, but derived exclusively from Archaean and Palaeoproterozoic source rocks. We note that these different interpretations have significant implications for whether the Ikalamavony Domain originated as an exotic volcanic island arc (Tucker et al. 2014; Archibald et al. 2018) or formed on the margin of the Antananarivo Domain.

In the studied area, a two-metre thick quartzite layer within a calc-silicate unit, above the quartzite unit, showed a detrital zircon pattern similar to the quartzite samples analysed here (Armistead et al., in review - Fig. 4e; sample MAD17-11-4A). This calc-silicate unit, composed of diopside gneiss interlayered with coarse-grained white marble layers, is here interpreted also as part of the Itremo Group (Fig. 3, 4c and 4d).

The depositional age window for the quartzite samples from the Itremo Group is wide, in between the youngest detrital zircon that is 1784 ± 26 Ma (#045-043; Supplementary table 1), the age of the metamorphic rims (550 ± 8 Ma; Fig. 8), and the age of the crosscutting Ambalavao granite, dated in the area by Archibald et al. (2019), at 549 ± 9 Ma and 544 ± 7 Ma (samples DA13-037 and DA13-038; Fig. 3). It is likely that this window can be narrowed by considering that these quartzites are crosscut by the ca. 850–750 Ma Imorona-Itsindro Suite (Archibald et al., 2016, 2017) elsewhere in the Itremo Group outcrop (Collins et al. 2003c). This large time gap allows various interpretations for the depositional history of the Itremo basin (Tucker et al., 2011a, Boger et al., 2014; Armistead et al., in review).

Our second group is represented by the plagioclase-biotite paragneiss (Figs. 3 and 4f - DA13-030), that shows a major detrital zircon population of Tonian-age (ca. 860–710 Ma), with minor peaks of Neoarchaean-Palaeoproterozoic ages and a single Mesoproterozoic grain (ca. 1500 Ma). This pattern is in contrast to the samples from the Itremo Group, with a major Palaeoproterozoic source (Fig. 6).

Based on the younger detrital zircon age and the few metamorphic rims dated in this sample, the deposition of this unit is constrained to the Cryogenian–early Ediacaran (ca. 710–610 Ma; Fig. 6 and 8). The plagioclase-biotite paragneiss is also crosscut by a folded granitic dyke dated at ca. 540 Ma (DA13-031; D.B. Archibald unpublished data; Figs. 3 and 4g). This age coincides with the metamorphic rims of the dated samples (Fig. 8).

Archibald et al. (2016) obtained a U–Pb zircon crystallization age of 828 ± 14 Ma for metafelsite bodies dykes/sills that intruded parallel and subparallel to the main foliation within the calc-silicate unit, included here in the Itremo Group (DA13-029 - Fig. 3 and 4h). These metafelsites are parallel to the plagioclase-biotite paragneiss and were not observed crosscutting them. To the NE of the area, Archibald et al. (2016) dated a porphyritic orthogneiss at 758 ± 10 Ma (sample DA13-020; Fig. 3) and correlated it with the Imorona-Itsindro Suite. This age overlaps with the youngest detrital zircon population that we dated in the paragneiss (Fig. 6).

One striking point is the consistent morphology of the detrital zircon grains from the plagioclase-biotite gneiss (DA13-030). They are predominantly subhedral with weakly rounded terminations, well-developed oscillatory zoning and with Th/U ratios higher than 0.1. The main age range is ca. 860–710 Ma (Fig. 11), same as the Imorona-Itsindro Suite. Sample DA-13-030 has ε_{Hf}(t) values ranging from -15.1 to -29.2 and T_{DM} Hf model ages between ca. 2.6 and 3.4 Ga (Fig. 9). The ε_{Hf}(t) for the

Imorona-Itsindro suite rocks (ca. 850–750 Ma) are between -4 to +14.0 in the Ikalamavony Domain and -7 to -37.3 in the Antananarivo Domain (Archibald et al., 2016). We therefore suggest that these detrital zircons were likely sourced from the Imorona-Itsindro Suite plutonic rocks or derived from coeval volcanic rocks erupted at the same time magmatic event.

In addition, some of the igneous zircons from the Tonian-aged detrital zircon population have cores with very distinct growth patterns and ages between ca. 2.70 to 1.82 Ga, with one analysis of ca. 1.54 Ga (Fig. 7d). Tonian-aged domains with magmatic oscillatory zoning mantling these zircon cores support as well the derivation from the Imorona-Itsindro magmatic event (Fig. 7d). Consequently, the Neoarchaean to the Palaeoproterozoic cores could be interpreted as xenocrysts within ca. 860–710 Ma igneous zircons. This reinforces the continental nature of the Imorona-Itsindro Suite magmatism, which is predominantly non-juvenile (Archibald et al., 2016; 2017).

Sample DA-13-030 also has a minor population of ca. 1.00–0.91 Ga igneous detrital zircons, with ε_{Hf}(t) varying from -0.5 to -5.1 and T_{DM} Hf model ages between 2.1 and 1.8 Ga, representing a slightly evolved source (Fig. 9). This group overlaps with the age and Hf isotope signature of the Dabolava magmatic suite (ca. 1.08–0.98 Ga; Fig. 11), which is considered to be the main source for the Ikalamavony Group (Archibald et al., 2018; Tucker et al., 2007). The Dabolava Suite has ε_{Hf}(t) varying from +15 to +5 (Archibald et al., 2018, Tucker et al., 2011a). Therefore, the zircon source for the paragneiss are slightly more isotopically evolved than the igneous rocks of the Dabolava Suite. Therefore, this paragneiss cannot be included in the Ikalamavony Group, to the west (Ikalamavony domain - Fig. 1b and 2), that shows a unimodal ca. 1.0 Ga contribution and is cross cut by a rhyolite layer dated at ca. 1015 Ma (CGS, 2009a).

According to published geological maps from Central Madagascar (e.g. Roig et al., 2012), this sample would be part of the Ikalamavony Group (Fig. 1b and 2). However, our data don't fully support this, because the major population of detrital zircon is ca. 800 Ma, whereas the Ikalamavony Group is dominated by ca. 1.0 Ga detritus. Therefore, the mapped 'Ikalamavony Group' that covers much of the Ikalamavony Domain may well contain more than one supracrustal unit of different age and depositional environment (as originally proposed in Cox et al. 2004a, Tucker et al., 2011, Boger et al., 2014). The Ediacaran-Cambrian tectonic event then intercalated these distinct lithostratigraphic groups, hindering the geological interpretation without more detailed mapping and geochronological controls.

Consequently, data from sample DA13-030 indicate that there is an Early Cryogenian supracrustal unit, not previously described in this region, interpreted here as volcaniclastic in origin, coeval to the major magmatic event of the Imorona-Itsindro Suite at ca. 850–750 Ma. The detrital age spectra is similar to the Manampotsy Group (to the east), but with an important discrepancy (Fig. 1b and 10). In the Manampotsy Group, the major detrital zircon population is Archaean-Palaeoproterozoic in age (BGS-USGS-GLW, 2008), which might be related to the basin being located in between Antongil/Masora cratons and Antananarivo domain (Fig. 1b).

Another supracrustal unit correlative would be the Late Neoproterozoic Molo Group that shows few Archaean and Palaeoproterozoic age modes, together with a more abundant Neoproterozoic detrital zircon population (Fig. 10; Cox et al., 2004a). The dated Molo sample was collected close to Ankaramena (Fig. 2), but differs from our unit since it contains younger detrital zircons (ca. 613 Ma) and age populations of ca. 1000 Ma, ca. 800 Ma and ca. 650 Ma (Fig. 10).

Within the southwestern Malagasy tectonic domains, two supracrustal units have similar geological characteristics and geochronological data to the sample DA-13-030. The Benato formation from the Anosyen Domain (Fig. 1b) is composed of quartz-feldspathic paragneisses with volcanic arc geochemical signature and volcanic zircon ages of ca. 796–626 Ma, interpreted to represent the extrusion/deposition of the Benato Formation (Boger et al., 2014). It differs from our sample, due to the lack of older zircons (e.g. 1.0 Ga, 2.2. Ga). The other unit is the Vohibory series, in the Vohibory Domain (Fig. 1b), that consists of metasedimentary rocks with marbles and juvenile amphibolite (Collins et al., 2012). Detrital zircon ages here range between ca. 900–750 Ma. This series is interpreted as deposited during the Cryogenian with Late Tonian to Cryogenian sources in an intra oceanic-arc setting that formed between the Antananarivo Domain and the Tanzania Craton (Collins et al., 2012; Fig. 1a). The difference from our sample is the lack of older zircon sources and all detritus in the Vohibory metasedimentary rocks can be correlated with local sources in the Vohibory Domain.

5.2. Potential Pre-Neoproterozoic sources

The Pre-Neoproterozoic zircons from all four samples are here correlated with the major Precambrian crustal blocks of central-east Gondwana (Fig. 1a and 10) including the Tanzania Craton (Africa), Dharwar Craton (India), Antongil-Masora blocks (Madagascar), and the Antananarivo Domain (Madagascar).

There are three Palaeoarchaean detrital zircons (samples DA13-036 and DA13-045) with U–Pb ages between ca. 3.35 and 3.22 Ga with $\varepsilon_{Hf}(t)$ varying from

+6.8 to +1.9 and T_{DM} Hf model ages between 3.5 Ga and 3.3 Ga. Similar juvenile detrital zircons are found in metasedimentary rocks throughout the Dharwar Craton with $\varepsilon_{Hf}(t)$ varying from +8 to -4 (Fig. 1a - Armistead et al., 2018; Lancaster et al., 2015; Maibam et al., 2016; Sarma et al., 2012). However, isotopically juvenile igneous rocks of this age in the Dharwar Craton are not recognized (Collins et al., 2015; Glorie et al., 2014; Mohan et al., 2014; Praveen et al., 2014; Yang et al., 2015). Similar rare Palaeoarchaean grains are found in the Southern Irumide Belt of Zambia (along the southern margin of the Congo-Tanzania-Bangweulu Block; Alessio et al., 2019). Supracrustal rocks with similar detrital zircon ages and Hf isotopic values are found in the Tanzania craton (Thomas et al. 2016).

The Mesoarchaean population (ca. 3.2–2.8 Ga), from samples DA13-036, DA13-039 and DA13-045, has $\varepsilon_{Hf}(t)$ values varying between +5.3 to -2.1 and T_{DM} Hf model ages between 3.5 Ga and 3.1 Ga. The Antongil and Masora domains, in eastern Madagascar (Fig. 1b), contain ca. 3200–3000 Ma igneous rocks that are interpreted to have been part of the Dharwar Craton at the time (Armistead et al., 2018; Schofield et al., 2010; Tucker et al., 1999, 2014). These transitional juvenile to crustal sources occur throughout the Dharwar Craton with $\varepsilon_{Hf}(t)$ varying from ~ +7 to -12 (Armistead et al., 2018; Collins et al., 2015; Glorie et al., 2014; Lancaster et al., 2015; Maibam et al., 2016; Mohan et al., 2014). Alternatively, these zircon grains could have been derived from the Tanzania Craton, which hosts a range of late Mesoarchaean aged igneous zircons (ca. 2820–2610 Ma) with $\varepsilon_{Hf}(t)$ varying from ~ +4.0 to -5.0 (Thomas et al., 2016).

The Mid-Neoarchaean to Mid-Siderian population (ca. 2.7 to 2.3 Ga) is more abundant in samples DA13-036, DA13-039 and DA13-045 with $\varepsilon_{Hf}(t)$ varying from +7.4 to -17.8 (but predominantly between +2 and -8), and T_{DM} Hf model ages between 3.9 Ga and 2.5 Ga (Figs. 6 and 9). In addition, sample DA13-030 also contains a minor population in this interval between 2.70 Ga and 2.34 Ga with $\varepsilon_{Hf}(t)$ varying from +7.0 to -1.7 and T_{DM} Hf model ages between 3.1 Ga and 2.5 Ga (Fig. 6 and 9). The best source candidate for these zircon grains is the Betsiboka Suite of the Antananarivo Domain (Figs. 1b and 10). Similar ages and Hf values occur in the Dharwar Craton, which show $\varepsilon_{Hf}(t)$ values varying from +8 to -12 (Glorie et al., 2014; Praveen et al., 2014; Yang et al., 2015). In contrast, there is no record in the Tanzania craton of igneous rocks with ages between 2.6–2.3 Ga (Fig. 10; Thomas et al., 2016).

The Mid-Rhyacian to Mid-Orosirian population (ca. 2.2 to 1.8 Ga) is also significant (Figs. 6 and 9). Samples DA13-036, DA13-039 and DA13-045 have U–Pb ages between 2.29 Ga and 1.78 Ga with $\varepsilon_{Hf}(t)$ varying from +2.0 to -12.9 and T_{DM} Hf model ages between 3.3 Ga and 2.6 Ga. Sample DA13-030 has detrital zircons with U–Pb ages between 2.26 Ga and 1.82 Ga with $\varepsilon_{Hf}(t)$ varying from -7.2 to -13.4 and T_{DM} Hf model ages between 3.4 Ga and 2.9 Ga. The Hf isotopic data show a trend from more juvenile, older sources to more evolved, younger zircon sources (Fig. 9). In addition, sample DA13-030 has two age groups. One group has ca. 2.0 Ga U–Pb ages with lower $\varepsilon_{Hf}(t)$ values (-11.6 and -13.4) and T_{DM} Hf model ages between 3.4 Ga and 3.3 Ga. Another group has U–Pb ages of ca. 1.8 Ga with slightly higher $\varepsilon_{Hf}(t)$ values (-7.2 and -7.3) and T_{DM} Hf model ages between 3.0 Ga and 2.9 Ga. This indicates two distinct crustal sources. Sample DA13-039 has only U–Pb detrital zircon grains dated between ca. 2.29–2.03 Ga with negative $\varepsilon_{Hf}(t)$ values and T_{DM} Hf model ages between ca. 3.6 Ga and 2.8 Ga. A possible source for these zircons is the igneous rocks from the Usagaran Belt that marks the eastern Tanzania Craton with ca. 2.0–1.8 Ga ages and $\varepsilon_{Hf}(t)$ values between 0 and -6 (Figs. 1b and 10; Reddy et al. 2003; Thomas et al., 2016).

For the Mesoproterozoic Era, a single grain in sample DA13-030 has U–Pb age of 1.54 Ga but no Hf data (#030-221 – Supplementary table 1). The Dharwar Craton has no record of magmatism or detrital zircon grains of this age. Thomas et al. (2016) describe magmatic rocks of ca. 1.5 Ga in the Tanzania Craton.

We can here conclude that the sources for the three quartzite samples, correlated with the Itremo Group, could be derived either from the eastern cratons (Dharwar, Antongil, Masora) and/or the western craton (Tanzania; Fig. 1a). The Paleoarchaean and Mesoarchaean detrital zircon populations could have come from both the Dharwar and the Congo-Tanzania-Bangweulu cratonic sources (Fig. 1a). The Neoarchaean sources (ca. 2.8 to 2.5 Ga), are present today in India and eastern Madagascar blocks, including the Antananarivo Domain (e.g. Betsiboka Suite - BGS-USGS-GLW, 2008; Kabete et al., 2006; Kroner et al., 2000; Tucker et al., 1999, 2007). The younger Palaeoproterozoic sources (ca. 2.2 to 1.8 Ga) are abundant only in the Tanzania Craton (Thomas et al., 2016). Cox et al. (1998, 2004a, 2004b) and Fitzsimons and Hulscher (2005) proposed correlation between the Antananarivo Domain and the Tanzania Craton. Collins and Pisarevsky (2005) described a Neoarchaean to early-Palaeoproterozoic igneous province comprising present-day Saudi Arabia, Madagascar and southern India and named this continental block Azania, composed primarily of ca. 2.45–1.90 Ga rocks (Collins et al., 2001; Collins and Pisarevsky, 2005; Kroner et al., 1999, 2000; Küster et al., 1990; Lenoir et al., 1994; Paquette and Nédélec, 1998; Teklay et al., 1998; Whitehouse et al., 2001). Our data are inconclusive but indicate that the sources for the Itremo-correlated quartzite do not easily fit with exclusively “Dhawar” or “Tanzanian” sources. However, the abundant late Palaeoproterozoic detritus that define the maximum depositional age, and may be close to the age of deposition (Cox et al., 1998), are most strongly correlated with similar sources for the Irumide and Southern Irumide belts of the Congo-Tanzania-Bangweulu Block (Alessio et al. 2019; Armistead et al. in review). It seems apparent that by ca. 1.7 Ga, a stable basin was developed that was predominantly sourcing African Palaeoproterozoic sources, with perhaps some contribution from recycled, older detritus.

The maximum depositional age for our three quartzite Itremo Group samples is marked by the youngest detrital zircon grain of 1784 ± 26 Ma. This is coherent with the depositional age proposed for the Maha Group, a package of Palaeoproterozoic metasedimentary rocks that overlies (perhaps tectonically) the Archaean rocks of the Masora Domain (Fig. 1b; BGS-USGS-GLW, 2008; Tucker, et al., 2011b). The Maha Group has ca. 2770 Ma to ca. 1800 Ma detrital zircon populations and a maximum depositional age of 1797 ± 18 Ma (Fig. 10; De Waele et al., 2011). We also follow Archibald et al. (2015) in correlating the Ambatolampy Group as a part of this supracrustal metasedimentary package including the Itremo and Maha groups (max. dep age 1.8 Ga; Archibald et al., 2015).

The similar pattern of detrital zircons sources, plus the maximum depositional age of ca. 1.7 Ga, for these three groups—Itremo (Antananarivo Domain), Ambatolampy (Antananarivo Domain) and Maha (Masora Craton)—suggest that they could be related to the same sedimentary basin (Figs. 1b and 10). In addition, the basin can be extended to include the Iakora Group (Anosyen Domain) in southern Madagascar (Boger et al., 2014) and the Sambirano–Sahantaha Group from the southern Bemarivo Domain (De Waele et al., 2011; Boger et al., 2014; Armistead et al., 2019), which is thrusted onto the Antongil Craton, to the north (Fig.1b; Schofield et al., 2010; Thomas et al., 2009; Armistead et al., 2019). These metasedimentary packages have similar late Palaeoproterozoic maximum depositional ages (Fig. 1b; Collins et al., 2012; Boger et al., 2014; De Waele et al., 2011). Taken together, they appear to represent an originally contiguous Palaeoproterozoic to early Mesoproterozoic sedimentary basin across and/or on the margins of Madagascar.

The minimum depositional age of this basin or basins is ca. 0.85 Ga as constrained by the emplacement of the ca. 850–750 Ma Imorona-Itsindro magmatic suite. The Dabolava Suite (ca. 1080–980 Ma) does not crosscut any of these formations. This is part of the logic to suggest that the Ikalamavony Domain and, thus, the Dabolava Suite formed exotic to the rest of Madagascar (Tucker et al. 2014; Archibald et al., 2016, 2018). Furthermore, the predominantly quartzite nature of the arenites within the Itremo Group and equivalents, differ lithologically from the Tonian-aged sedimentary rocks that formed coeval with the Dabolava Suite (Ikalamavony Group). The latter usually contain appreciable feldspathic and lithic grains and are interbedded with volcanic rocks that are distinct from the Itremo, Ambatolampy and Maha groups. The proposed sedimentary environment for the Itremo Group is a stable shelf with quartz arenites, pelites and carbonate rocks—possibly a passive margin succession. This suggests a tectonically quiescent period from ca. 1700 to ca. 850 Ma in the Antananarivo Domain (Tucker et al., 2014).

Alternatively, it could be argued that the quartzites mapped by us as the Itremo Group (Fig. 2) are part of the same sequence as the structurally overlying plagioclase-biotite paragneiss (mapped as the Ikalamavony Group in Fig. 2) and both were part of a Cryogenian sedimentary basin. The layers are tectonically parallel , however, there are important discrepancies between these two groups here studied, including: (i) the Itremo Group quartzites represent a more mature sedimentary protolith, also reflected on the round detrital zircons, while the plagioclase-biotite paragneiss protoliths represent feldspar-rich, immature sediments with sub- to euhedral detrital zircon grains. (ii) The detrital zircon data implies a major change of source. (iii) The absence of metamorphic zircon rims on the plagioclase-biotite paragneiss suggests that it either came from distinct crustal levels and was juxtaposed during collision or the distinct original composition favoured the zircon growth.

5.3. Tectonic evolution of the Cryogenian continental sedimentary basin

The sample here presented deposited in a Cryogenian (<710 Ma) continental basin that developed coeval with the late stages of the Imorona-Itsindro Suite, its major source (population 860–750 Ma from sample DA13-030 – Figs. 6 and 11). We propose here that this Cryogenian late-magmatic unit can be either correlated with the Manampotsy group (in the contact between Masora and Antananarivo domains– Fig.1b) or the Benato formation (in the Anoysen domain – Fig 1b), implying that this basin late-Imorona-Itsindro magmatic suite could have been much wider. In addition, based on mineral composition, well-preserved plagioclase grains, and the absence of K-feldspar we propose that the source was andesitic in composition. The Archaean/Palaeoproterozoic zircon cores, within the Tonian-aged igneous crystals (Fig. 6 and 7), indicate that the source for the Imorona-Itsindro Suite was most likely the basement of the Antananarivo Domain. The Hf isotopic evidence also supports this correlation. A minor detrital population of ca. 1.0 to 0.9 Ga zircon in sample DA13-030 (Fig. 11) could be derived from the Dabolava Arc rocks that crop out to the west, although the Hf detrital data indicate that this population is less juvenile (Fig. 1b and 2). These crystals are not xenocrysts within the ca. 860–710 Ma igneous zircons, but detrital zircon with a single age. The predominant late-Tonian-aged detrital zircon, the spatial relation with Imorona-Itsindro Suite and the ca. 2.8–1.7 Ga xenocrysts support that this sample was deposited in a basin that developed coevally to the Imorona-Itsindro magmatic event, with some contribution from the Ikalamavony domain, attesting that both Antananarivo and Ikalamavony domains were attached by ca. 850 Ma (Figs. 11 and 12).

We present three models of Late Tonian-Early Cryogenian tectonic setting for the sedimentary basin recorded by sample DA-13-030 (Fig. 12). These tectonic models are based on previous work and the interpretation of our new data, in order to constraint the tectonic setting for the deposition of this Cryogenian basin. The two first alternatives encompass subduction zones which magmatic arc is represented by the Imorona-Itsindro suite.

The first model postulates that the Imorona-Itsindro Suite formed in a continental arc related to a west-dipping subduction zone as proposed by several authors (BGS-USGS-GLW, 2008; Collins, 2006; Collins and Pisarevsky, 2005; Collins and Windley, 2002; Kröner et al., 2000). In this case, the sources for pre-Neoproterozoic detrital zircon grains would be the Antananarivo Domain (including the Itremo Group), since the “Indian” sources (Masora and Antongil) were distal and separated by an ocean (Fig. 12a).

The second scenario would be the eastward subduction of oceanic crust (Handke et al., 1999; Boger et al., 2014), and emplacement of the Imorona-Itsindro Suite also as a continental magmatic arc. In such a circumstance, the Antananarivo and the Antongil-Masora Domains would be connected to the Dhawar Craton corroborating to the suggestion that the major source for the older detrital zircons would be the “Indian” blocks (Fig. 12b). This is supported by the similarities between the Manampotsy metasedimentary sequence and the unit represented here by sample DA-13-030.

The third model envisaged is modified from Tucker et al. (2011b) proposal of an intracontinental setting with a period of crustal extension. According to these authors, the Imorona-Itsindro Suite and its related sedimentary rocks are the product of continental extension (Fig. 12c). The pre-Neoproterozoic sources for the detrital zircon would mostly be the “Indian” and “Malagasy” blocks, which is a similar situation to the model of the east-dipping subduction (Fig. 12b). Nevertheless, there is no evidence of rift sequences related to this tectonic environment (Collins et al. 2003a). The Cryogenian biotite-plagioclase paragneiss is interpreted as a psammitic rock with well-preserved plagioclase clasts indicates a fast sedimentation rate, more likely represented by an orogenic setting.

An intriguing chapter in the Neoproterozoic tectonic evolution of central Madagascar is the paucity of geological units, structures and ages throughout the Cryogenian period (ca. 720–635 Ma). If we consider the subduction models, there should be an explanation for the long period of tectonic quiescence and paucity of magmatic arc activity (Imorona-Itsindro Suite) before the well-documented continental collision at ca. 550 Ma. The two subduction model alternatives (Fig. 12a and 12b) do not explain the tectonic quiescence of the Cryogenian (ca. 720–635 Ma), between the Tonian widespread magmatic activity and the major Ediacaran-Cambrian continental collision, registered in the samples of this study and in the overall literature of the Antananarivo and Ikalamavony domains.

The two first tectonic scenarios pointed out would need to stall for ca. 100 m.y. (Fig. 12). That is why we suggest here that the third alternative does not exclude a prior subduction. In this case, this scenario is proposed for after 710 Ma and could be a consequence for the subduction stall of either tectonic settings (a) and (b). The tectonic setting for a Cryogenian basin would be an extensional intracontinental context late-Imorona Itsindro Suite. The corresponding rift related sequences are absent from the record, which is expected if it is considered that a continental collisional event took place later at 550 Ma. Major uplift and crustal thickness related to the Gondwana amalgamation would contribute to the erosion of pre-orogenic upper crustal units.

The tectonic environment of the Great Basin of the western US, for example, configures a middle to late Cenozoic extension (more than 60 m.y.) of an Archaean and Palaeoproterozoic continental crust, which involved an initial intra-arc to back arc deformation and later a transtensional torsion of the continental block inland from the evolving San Andreas transform system (Dickinson, 2006). This wide rift system is a product of the interaction between the subduction of ridge-transform systems that affects the thermal gradients beneath the upper lithospheric mantle (Dickinson, 2006). The continental extension hypothesis for the Tonian tectonic evolution in central Madagascar, proposed by Tucker et al. (2011b), partially shown in our model 3 (Fig. 12c), could be compared to this recent environment, if it is considered that there was a previous subduction.

We envisage that the Tonian-Cryogenian magmatism and its associated sedimentary basins might be related to a continental extension due to an outboard subduction system, partially jammed due to the subduction of a ridge-transform system. This would be coherent also with the subsequent tectonic quiescence of ca. 100 m.y that preceded the final continental collision. This alternative could be related to subduction stall between Androyen and Anosyen domains or among the Masora craton and the Antananarivo domain (Fig.12c). Therefore, either alternative (a) or (b) from figure 12 could evolve to alternative (c). This would differ from the original purpose of Tucker et al. (2011b), that presumes no subduction at this time.

If we consider the western US setting as an analogue, the development of a transform continental setting would resolve the tectonic Cryogenian paucity. Other Gondwana-forming belts do have magmatic provinces that are either related to subduction or intracontinental rifting in the period between ca. 850-750 Ma, such as the Kaoko belt (Goscombe and Gray 2007, Konopasek et al., 2017), the Ribeira belt (Meira et al., 2015; Schmitt et al., 2016) or the Arabian Nubian Shield (Johnson et al. 2011). Other orogens have well-documented magmatism related with sedimentary rift sequences in the same period, like the Damara belt (Miller et al., 2009; Nascimento et al., 2016).

The advance of the converging pre-Gondwana blocks finally collide at ca. 560–520 Ma deforming most central Madagascar lithostratigraphic units. This would be triggered by an outboard subduction setting, probably related to the SW terranes of Madagascar, especially the Vohibory Domain (Fig. 1b), and/or the juvenile arc terranes of eastern Africa including the Tanzanian Ntaka Terrane (Mole et al. 2018) and the Cabo Delgado Terrane of Mozambique (Bingen et al. 2009). These late Ediacaran to early Cambrian pervasive multi-phase ductile structures and high-grade metamorphism collisional event affected most of the Malagasy terranes (Collins, 2006; Tucker et al., 2011b; Armistead et al., 2020). The tectonic pile, preserved in the studied area, intercalated units of distinct tectonic pre-collisional settings (e.g. Itremo and Ikalamavony Groups; Fig. 3). In addition, the detrital zircon grains from the three quartzite samples show thick metamorphic rims of ca. 550 Ma (Fig. 8). The event coincided with widespread magmatism at ca. 540 Ma (Ambalavao Suite), which marked the terminal Gondwana amalgamation (Fig. 1a; Schmitt et al., 2018).

6 - Conclusions

The data presented here reinforce a hypothesis that the tectonic pile of metasedimentary units from the Itremo-Ikalamavony Domain in central Madagascar records two distinct basins in time and environment. The lower sequence comprises quartzite and calc-silicate units with a major Palaeoproterozoic source and a minor Archaean contribution with a maximum depositional age of ca. 1.78 Ga, based on U–Pb detrital zircon data. This sequence reflects stable shelf sedimentation on the basement and or the margins of the Antananarivo Domain and is correlated with the Itremo Group. The sources are varied and detritus could be derived from either the eastern cratons (Dharwar, Antongil, Masora) or the western craton (Tanzania). The abundant near-depositional-age Palaeoproterozoic detritus were likely sourced from eastern Africa. Our data support an originally contiguous late Palaeoproterozoic to early Mesoproterozoic sedimentary basin that blanketed central Madagascar, and possibly included eastern Africa.

The upper sequence is a metasedimentary unit represented mostly by biotite-plagioclase paragneiss, with psammitic protolith, interleaved with volcanic/subvolcanic andesitic/rhyolitic dykes/sills/flows. The predominant Tonian-Early Cryogenian (ca. 860-710 Ma) detrital zircon population is probably derived from the Imorona-Itsindro Suite magmatic rocks, with comparable $\varepsilon_{Hf}(t)$ values and T_{DM} Hf model ages. Some Tonian-aged detrital zircon grains preserve Neoarchaean to Palaeoproterozoic cores interpreted as xenocrysts, reinforcing the continental nature of the Imorona-Itsindro magmatism. This is supported by Hf isotopic data that indicate an evolved, crustal signature.

Our preferred tectonic scenario for this Early Cryogenian sedimentary basin would be a continental back arc setting associated with an outboard subduction system, which became locked-up by the subduction of a ridge-transform system. We suggest that Holocene North America is a modern analogue. We envisage that extension was triggered by distal subduction to the west, represented by the Vohibory Domain of SW Madagascar and corollaries in eastern Africa. The apparent period of tectonic quiescence in central Madagascar between ca. 720 and 635 Ma is explained in this model by the conversion of this subduction margin into a transform margin. This still enables the oblique approach of Neoproterozoic India with Madagascar and Africa to collide in the Ediacaran-Cambrian (Collins and Pisarevsky 2005; Merdith et al. 2017) deforming central Madagascar and documented here by ca. 550 Ma zircon metamorphic rims in the Itremo Group quartzites.

Acknowledgments

Authors acknowledge the careful review of Steven Boger and an anonymous reviewer. R.S. Schmitt acknowledges the CNPq Grant 311748/2018-0. ASC acknowledges Australian Research Council grant FT120100340. We would like to thank the undergraduate geology student Luan Dias for helping draft the figures. The Razafinjoelina family, in particular Auguste and Berthieu are thanked for providing transportation and assistance in the field. The fieldwork was supported by Petrobras/CENPES through the Gondwana cooperation project with UFRJ. This paper is a contribution to IGCP-628 ‘‘The Gondwana geological Map and the tectonic evolution of Gondwana” and IGCP 648 “Supercontinent Cycles and Global Geodynamics”.

REFERENCES

Alessio, B.L., Collins, A.S., Siegfried, P.R., Glorie, S., De Waele, B., Payne, J., Archibald, D.B., 2019. Neoproterozoic tectonic geography of the south-east Congo Craton in Zambia as deduced from the age and composition of detrital zircons. Geoscience Frontiers, 10, 2045-2061.

Archibald, D.B., Collins, A.S., Foden, J.D., Payne, J.L., Taylor, R., Holden, P., Razakamanana, T., Clark, C., 2015. Towards unravelling the Mozambique Ocean conundrum using a triumvirate of zircon isotopic proxies on the Ambatolampy Group, central Madagascar. Tectonophysics 662, 167-182.

Archibald, D.B., Collins, A.S., Foden, J.D., Payne, J.L., Holden, P., Razakamanana, T., De Waele, B., Thomas, R.J., Pitfield, P.E.J., 2016. Genesis of the Tonian Imorona–Itsindro magmatic Suite in central Madagascar: Insights from U–Pb, oxygen and hafnium isotopes in zircon. Precambrian Research 281, 312-337.

Archibald, D.B., Collins, A.S., Foden, J.D., Razakamanana, T., 2017. Tonian arc magmatism in central Madagascar: the petrogenesis of the Imorona-Itsindro Suite. The Journal of Geology 125(3), 271-297.

Archibald, D.B., Collins, A.S., Foden, J.D., Payne, J.L., Macey, P.H., Holden, P., Razakamanana, T., 2018. Stenian-Tonian arc magmatism in west central Madagascar: The Genesis of the Dabolava Suite. Journal of the Geological Society 175, 111-129.

Archibald, D.B., Collins, A.S., Foden, J.D., Payne, J.L., Holden, P. and Razakamanana, T., 2019. Late syn- to post-collisional magmatism in Madagascar: The genesis of the Ambalavao and Maevarano Suites. Geoscience Frontiers, 10, 2063-2084.

Armistead, S.E., Collins, A.S., Payne, J.L., Foden, J.D., De Waele, B., Shaji, E., Santosh, M. 2018. A re-evaluation of the Kumta Suture in western peninsular India and its extension into Madagascar. Journal of Asian Earth Sciences 157, 317-328.

Armistead, S.E., Collins, A.S., Payne, J.L., Cox, G.M., Merdith, A.S., Foden, J.D., Razakamanana, T., De Waele, B. 2019. Evolving marginal terranes during the Neoproterozoic supercontinent reorganisation: constraints from the Bemarivo Belt in northern Madagascar. Tectonics 38, 2019–2035.

Armistead, S.E., Collins, A.S., Redaa, A., Gilbert, S., Jepson, G., Gillespie, J., Blades, M.L., Foden, J.D., Razakamanana, T., 2020. Structural Evolution and Medium-Temperature Thermochronology of Central Madagascar: Implications for Gondwana Amalgamation. Journal of the Geological Society, London 177, 784–798.

Armistead, S.E., Collins, A.S., Schmitt, R.S., Costa, R.L., De Waele, B., Razakamanana, T., Payne, J.L., Foden, J.D. in review. Proterozoic basin evolution and tectonic geography of Madagascar during the Nuna/Columbia Supercontinent. Tectonics.

Ashwal, L.D., Tucker, R.D., Zinner, E.K., 1999. Slow cooling of deep crustal granulites and Pb-loss in zircon. Geochemica et Cosmochimica Acta 63, 2839-2851.

Bauer, W., Walsh, G.J., de Waele, B., Thomas, R.J., Horstwood, M.S.A., Bracciali, L., Schofield, D.I., Wollenberg, U., Lidke, D.J., Rasaona, I.T. and Rabarimanana, M.H., 2011. Cover sequences at the northern margin of the Antongil Craton, NE Madagascar. Precambrian Research 189(3-4), 292-312.

BGS-USGS-GLW, 2008. Revision de la cartographie geologique etminiere des zones Nord, Centre, et Centre Est de Madagascar. BGS Report CR/08/078 (Keyworth England).

Bingen, B., Jacobs, J., Viola, G., Henderson, H.C., Skåra, Ø, Boyd, R., Thomas, R.J., Solli, A., Key, R.M., Daud, E.X.F., 2009. Geochronology of the Precambrian crust in the Mozambique belt in NE Mozambique, and implications for Gondwana assembly. Precambrian Research 170, 231-255.

Boger, S.D., Hirdes, W., Ferreira, C.A.M., Schulte, B., Jenett, T., Fanning, C.M., 2014. From passive margin to volcano-sedimentary forearc: the Tonian to Cryogenian evolution of the Anosyen Domain of southeastern Madagascar. Precambrian Research 247, 159-186.

Boger, S.D., Hirdes, W., Ferreira, C.A.M., Jenett, T., Dallwig, R., Fanning, C.M., 2015. The 580–520 Ma Gondwana suture of Madagascar and its continuation into Antarctica and Africa. Gondwana Research 28(3), 1048-1060.

Boger, S.D., Maas, R., Pastuhov, M., Macey, P.H., Hirdes, W., Schulte, B., Fanning, C.M., Ferreira, C.A.M., Jenett, T., Dallwig, R., 2019. The tectonic domains of southern and western Madagascar. Precambrian Research, 327: 144-175.

Bouvier, A., Vervoort, J.D., Patchett, P.J., 2008. The Lu–Hf and Sm–Nd isotopic composition of CHUR: Constraints from unequilibrated chondrites and implications for the bulk composition of terrestrial planets. Earth and Planetary Science Letters 273(1-2), 48-57.

Cenki-Tok, B., Berger, A. and Gueydan, F., 2015. Formation and preservation of biotite-rich microdomains in high-temperature rocks from the Antananarivo Block, Madagascar. International Journal of Earth Sciences 105(5), 1471-1483.

CGS, 2009a. Map Explanation of 1:100 000 scale (Zone E) Sheets I46 – Ambararata, J46 – Beopoaka, I47 – Itondy, J47 – Belobaka, K47 – Ambatofotsy, I48 – Miandrivazo, J48 – Betrondro, K48 – Ambatondradama, I49 – Ankotrofotsy, J49 – Dabolava, K49 – Anjoma-Ramartina, L49 – Vasiana, M49 – Ankazomiriotra, N49 – Antsirabe. P.H. Macey, Miller, J.A., Armstrong, R.A., Bisnath, A., Yibas, B., Frost-Killian, S., Chevallier, L., Mukosi, N.C., Cole, J., le Roux, P. and Haddon, I.G. Ministère de L'Energie et des Mines - Project de Gouvernance des Ressources Minérales, Antananarivo, Madagascar and Council for Geoscience, Pretoria, South Africa.

CGS, 2009b. Map Explanation of 1:100 000 scale sheets (Zone F) G41 – Ambohipaky, H41 – Bevary, G42 – Mangoboky, H42 – Bekodoka, G43 – Andolamasa, H43 – Andrafialava and parts of G40 – Ankasakasa, F40 – Saint-Andre, F41 – Betsalampy, H40 – Maroboaly-Sud, I40 – Soalala-Sud, I41 – Andranomavo, F42 – Marovoay Kely, I42 – Mahabe, F43 – Bebao, F44 – Antranogoaika, G44 – Morafeno, I43 – Ampoza, H44- Bemolanga and I44 Makaraingo. P.H. Macey, Miller, J.A., Armstrong, R.A., Ingram, B.A., Bisnath, A., Yibas, B., Frost-Killian, S., Chevallier, L., Finkelstein, J. and Haddon, I.G. Ministère de L'Energie et des Mines – Project de Gouvernance des Ressources Minérales, Antananarivo, Madagascar and Council for Geoscience, Pretoria, South Africa.

Collins, A.S., 2006. Madagascar and the amalgamation of Central Gondwana. Gondwana Research 9, 3-16.

Collins, A.S., Pisarevsky, S.A., 2005. Amalgamating eastern Gondwana: The evolution of the Circum-Indian Orogens. Earth Science 71, 229-270.

Collins, A.S., Windley, B.F., 2002. The tectonic evolution of central and northern Madagascar and its place in the final assembly of Gondwana. The Journal of geology 110, 325-339.

Collins, A.S., Razakamanana, T., Windley, B.F., 2000. Neoproterozoic extensional detachment in central Madagascar: Implications for the collapse of the East African Orogen. Geological Magazine 137(1), 39-51.

Collins, A.S, Windley, B.F., Kröner, A., Fitzsimons, I., Hulscher, B., 2001. The Tectonic Architecture of Central Madagascar: Implication on the Evolution of the East African Orogeny. Gondwana Research 4, 152-153.

Collins, A.S., Fitzsimons, I.C.W., Hulscher, B. and Razakamanana, T., 2003a. Structure of the eastern margin of the East African Orogen in central Madagascar. Precambrian Research 123, 111-133.

Collins, A.S., Kröner, A., Fitzsimons, I.C.W. and Razakamanana, T. 2003b. Detrital Footprint of the Mozambique Ocean: U/Pb SHRIMP and Pb Evaporation Zircon Geochronology of Metasedimentary Gneisses in Eastern Madagascar. Tectonophysics. 375, 77-99.

Collins, A.S., Johnson, S., Fitzsimons, I.C.W., Powell, C.McA., Hulscher, B., Abello, J. and Razakamanana, T. 2003c. Neoproterozoic deformation in central Madagascar: a structural section through part of the East African Orogen. In: Proterozoic East Gondwana: Supercontinent Assembly and Breakup (eds. Yoshida, M., Windley, B.F. and Dasgupta S.). Geological Society, London, Special Publication, 206, 363-379.

Collins, A.S., Clark, C., Santosh, M., Sajeev, K., Kropinski, L., McKenzie, S., Hand, P.M., Kinny, P.D., 2006. The Mozambique ocean suture in southern India: Age and significance of granulites in the Palghat-Cauvery shear zone system. ASEG Extended Abstracts 1, 1-3.

Collins, A.S., Kinny, P.D., Razakamanana, T., 2012. Depositional age, provenance and metamorphic age of metasedimentary rocks from southern Madagascar. Gondwana Research 21(2-3), 353-361.

Collins, A.S., Patranabis-Deb, S., Alexander, E., Bertram, C.N., Falster, G.M., Gore, R.J., Mackintosh, J., Dhang, P.C., Saha, D., Payne, J.L., Jourdan, F., Backé, G., Halverson, G.P., Wade, B.P., 2015. Detrital mineral age, radiogenic isotopic stratigraphy and tectonic significance of the Cuddapah Basin, India. Gondwana Research 28(4), 1294-1309.

Cox, R., Armstrong, R.A., Ashwal, L.D., 1998. Sedimentology, geochronology and provenance of the Proterozoic Itremo Group, central Madagascar, and implications for pre-Gondwana palaeogeography. Journal of the Geological Society 155(6), 1009-1024.

Cox, R., Coleman, D.S., Wooden, J.L., DeOreo, S.B., 2001. A newly recognised Late Neoproterozoic metasedimentary sequence in Central Madagascar suggests terrane juxtaposition at 560±7 Ma during Gondwana assembly, in: Geological Society of America Abstracts with Programs 33, A436.

Cox, R., Coleman, D.S., Chokel, C.B., DeOreo, S.B., Wooden, J.L., Collins, A.S., De Waele, B., Kröner, A., 2004a. Proterozoic tectonostratigraphy and paleogeography of Central Madagascar derived from detrital zircon U-Pb age populations. The Journal of Geology 112, 379-399.

Cox, R., Fernandez, A., Schreurs, G., 2004b. Discussion on tectonic evolution of the Proterozoic Itremo Group metasediments in Central Madagascar. Special

Publication 206, 2003, 381-399. Journal of the Geological Society 161(3), 539-541.

De Waele, B., Horstwood, M.S.A., Bauer, W., Key, R.M., Pitfield, P.E.J., Potter, C.J., Rabarimana, M., Rafahatelo, J.-M., Ralison, V., Randriamananjara, T., Smith, R.A., Thomas, R.J. & Tucker, R.D., 2008. U-Pb detrital zircon geochronological provenance patterns of supracrustal successions in central and northern Madagascar. Colloquium of African Geology, Hammamat, Tunisia, pp. 3.

De Waele, B., Fitzsimons, I.C.W., Wingate, M.T.D., Tembo, F., Mapani, B. and Belousova, E.A., 2009. The geochronological framework of the Irumide Belt: A prolonged crustal history along the margin of the Bangweulu Craton. American Journal of Science, 309(2): 132-187.

De Waele, B., Thomas, R.J., Macey, P.H., Horstwood, M.S.A., Tucker, R.D., Pitfield, P.E.J., Schofield, D.I., Goodenough, K.M., Bauer, W., Key, R.M., Potter, C.J., Armstrong, R.A., Miller, J.A., Randriamananjara, T., Ralison, V., Rafahatelo, J.M., Rabarimanana, M., Bejoma, M., 2011. Provenance and tectonic significance of the Palaeoproterozoic metasedimentary successions of central and northern Madagascar. Precambrian Research 189, 18-42.

de Wit, M.J., Bowring, S.A., Ashwal, L.D., Randrianasolo, L.G., Morel, V.P.I., Rambeloson, R.A., 2001. Age and tectonic evolution of Neoproterozoic ductile shearzones in southwestern Madagascar, with implications for Gondwana studies. Tectonics 20, 1-45.

Dhuime, B., Hawkesworth, C. and Cawood, P., 2011. When continents formed. Science, 331(6014): 154-155.

Dickinson, W.R., 2006. Geotectonic evolution of the Great Basin. Geosphere 2(7), 353-368.

Emmel, B., Jöns, N., Kröner, A., Jacobs, J., Wartho, J.-A., Schenk, V., Razakamanana,T., Austegard, A., 2008. From closure of the Mozambique Ocean to Gondwana break-up: new evidence from geochronological data of the Vohibory terrane, southwest Madagascar. The Journal of Geology 116, 21-38.

Fernandez, A., Schreurs, G., 2003. Tectonic evolution of the Proterozoic Itremo Group metasediments in central Madagascar. Geological Society, London, Special Publications 206(1), 381-399.

Fitzsimons, I.C.W., Hulscher, B., 2005. Out of Africa: detrital zircon provenance of central Madagascar and Neoproterozoic terrane transfer across the Mozambique Ocean. Terra Nova 17(3), 224-235.

GAF-BGR, 2008a. Final report. Explanatory notes for the Anosyen Domain southeast Madagascar. Réalisation des travaux de cartographie géologique de Madagascar, révision approfondie de la cartographie géologique et minière aux échelles 1/100,000 et 1/500,000 zone Sud. Republique de Madagascar, Ministère de L’energie et des Mines (MEM/SG/DG/UCP/PGRM), 93 pp.

GAF-BGR, 2008b. Final report. Explanatory notes for the Androyen Domain southern Madagascar. Réalisation des travaux de cartographie géologique de Madagascar, révision approfondie de la cartographie géologique et minière aux échelles 1/100,000 et 1/500,000 zone Sud. Republique de Madagascar, Ministère de L’energie et des Mines (MEM/SG/DG/UCP/PGRM), 81 pp.

GAF-BGR, 2008c. Final report. Explanatory notes for the Vohibory Domain southwest Madagascar. Réalisation des travaux de cartographie géologique de Madagascar, révision approfondie de la cartographie géologique et minière aux échelles 1/100,000 et 1/500,000 zone Sud. Republique de Madagascar, Ministère de L'energie et des Mines (MEM/SG/DG/UCP/PGRM), 85 pp.

Glorie, S., De Grave, J., Singh, T., Payne, J.L., Collins, A.S., 2014. Crustal root of the Eastern Dharwar Craton: zircon U–Pb age and Lu–Hf isotopic evolution of the East Salem Block, southeast India. Precambrian Research 249, 229-246.

Goncalves, P., Nicollet, C., Montel, J.M., 2004. Petrology and *in situ* U–Th–Pb Monazite Geochronology of Ultrahigh-Temperature Metamorphism from the Andriamena Mafic Unit, North–Central Madagascar. Significance of a Petrographical *P*–*T* Path in a Polymetamorphic Context, Journal of Petrology, 45, 1923–1957.

Goodenough, K.M., Thomas, R.J., De Waele, B., Key, R.M., Schofield, D.I., Bauer, W., Tucker, R.D., Rafahatelo, J.-M., Rabarimanana, M., Ralison, A.V, Randriamananjara, T., 2010. Post-collisional magmatism in the central East African Orogen: The Maevarano Suite of north Madagascar. Lithos 116, 18-34.

Goscombe, B., Gray, D.R., 2007. The Coastal Terrane of the Kaoko Belt, Namibia: Outboard arc-terrane and tectonic significance. Precambrian Research 155(1-2), 139-158.

Grantham, G.H., Manhiça, A.D., Armstrong, R. A., Kruger,F.J., Loubser, M. 2011. New SHRIMP, Rb/Sr and Sm/Nd isotope and whole rock chemical data from central Mozambique and western Dronning Maud Land, Antarctica: Implications for the nature of the eastern margin of the Kalahari Craton and the amalgamation of Gondwana. Journal of African Earth Sciences 59, 74-100.

Griffin, W.L., Wang, X., Jackson, S.E., Pearson, N.J., O'Reilly, S.Y., Xu, X., Zhou, X., 2002. Zircon chemistry and magma mixing, SE China: In-situ analysis of Hf isotopes, Tonglu and Pingtan igneous complexes. Lithos 61, 237–269.

Griffin, W L; Powell, W.J.; O'Reilly, S.Y., 2008. GLITTER : Data reduction software for laser ablation ICP-MS. Laser Ablation ICP-MS in the Earth Sciences: Current practices and outstanding issues, 308-311.

Handke, M.J., Tucker, R.D., Ashwal, L.D., 1999. Neoproterozoic continental arc magmatism in west-central Madagascar. Geology 27, 351-354.

Horton, F., Hacker, B., Kylander-Clark, A., Holder, R., Jöns, N., 2016. Focused radiogenic heating of middle crust caused ultrahigh temperatures in southern Madagascar. Tectonics 35, 293-314.

Jackson, S.E., Pearson, N.J., Griffin, W.L., Belousova, E.A., 2004. The application of laser ablation-inductively coupled plasma-mass spectrometry to in situ U–Pb zircon geochronology. Chemical Geology 211(1-2), 47-69.

Jacobs, J., Fanning, C.M., Henjes-Kunst, F., Olesch, M., Paech, H., 1998. Continuation of the Mozambique Belt into East Antarctica: Grenville-Age Metamorphism and Polyphase Pan-African High-Grade Events in Central Dronning Maud Land. Journal of Geology 106, 385-406.

Jacobs, J., Thomas, R. J. 2004. Himalayan-type indenter-escape tectonics model for the southern part of the late Neoproterozoic–early Paleozoic East African–Antarctic orogen. Geology *32*, 721-724.

Johnson, P.R., Andresen, A., Collins, A.S., Fowler, A.R., Fritz, H., Ghebreab, W., Kusky, T., Stern, R.J., 2011. Late Cryogenian-Ediacaran history of the Arabian-Nubian Shield: A review of depositional, plutonic, structural, and tectonic events in the closing stages of the northern East African Orogen. Journal of African Earth Sciences. 61, 167-232.

Jöns, N., Schenk, V., 2008. Relics of the Mozambique Ocean in the central East African Orogen: evidence from the Vohibory Block of southern Madagascar. Journal of Metamorphic Geology 26, 17-28.

Jöns, N., Schenk, V., 2011. The ultrahigh temperature granulites of southern Madagascar in a polymetamorphic context: implications for the amalgamation of the Gondwana supercontinent. European Journal of Mineralogy 23, 127-156.

Kabete, J., Groves, D., McNaughton, N., Dunphy, J., 2006. The geology, SHRIMP U–Pb geochronology and metallogenic significance of the Ankisatra-Besakay District, Andriamena belt, northern Madagascar. Journal of African Earth Sciences 45(1), 87-122.

Key, R.M., Pitfield, P.E.J., Thomas, R.J., Goodenough, K.M., De Waele, B., Schofield, D.I., Bauer, W., Horstwood, M.S.A., Styles, M.T., Conrad, J., Encarnacion, J., Lidke, D.J., O'Connor, E.A., Potter, C., Smith, R.A., Walsh, G.J., Ralison, A.V., Randriamananjara, T., Rafahatelo, J.-M., and Rabarimanana, M., 2011. Polyphase Neoproterozoic orogenesis within the East Africa–Antarctica Orogenic Belt in central and northern Madagascar. Geological Society, London, Special Publications 357(1), 49-68.

Konopásek, J., Hoffmann, K-H., Sláma, J., Kosler, J., 2017. The onset of flysch sedimentation in the Kaoko Belt (NW Namibia) – Implications for the pre-collisional evolution of the Kaoko-Dom Feliciano-Gariep orogen. Precambrian Research 298, 220-234.

Kröner, A., Windley, B.F., Jaeckel, P., Brewer, T.S., Razakamanana, T., 1999. Precambrian granites, gneisses and granulites from Madagascar: new zircon ages and regional significance for the evolution of the Pan-African orogen.

Journal of the Geological Society (London) 156, 1125-1135.

Kröner, A., Hegner, E., Collins, A.S., Windley, B.F., Brewer, T.S., Razakamanana, T., Pidgeon, R.T., 2000. Age and magmatic history of the Antananarivo Block, central Madagascar, as derived from zircon geochronology and Nd isotopic systematics. American Journal of Science 300(4), 251-288.

Küster, D., Utke, A., Leupolt, L., Lenoir, J.L., Haider, A., 1990. Pan-African granitoid magmatism in northeastern and southern Somalia. Berliner Geowissenschaftliche Abhandlungen 120, 519-536.

Lancaster, P.J., Dey, S., Storey, C.D., Mitra, A., Bhunia, R.K., 2015. Contrasting crustal evolution processes in the Dharwar craton: insights from detrital zircon U–Pb and Hf isotopes. Gondwana Research 28(4), 1361-1372.

Lenoir, J.-L., Küster, D., Liégeois, J.-P., Utke, A., Haider, A., Matheis, G., 1994. Origin and regional significance of late Precambrian and early Palaeozoic granitoids in the Pan-African belt of Somalia. Geologische Rundschau 83, 624-641.

Ludwig, K.R., 2012. User's Manual for Isoplot 3.75. A Geochronological Toolkit for Microsoft Excel.

Maibam, B., Gerdes, A., Goswami, J. N., 2016. U–Pb and Hf isotope records in detrital and magmatic zircon from eastern and western Dharwar craton, southern India: Evidence for coeval Archaean crustal evolution. Precambrian Research 275, 496-512.

McMillan, A., Harris, N.B.W., Holness, M., Ashwal, L., Kelley, S., Rambeloson, R., 2003. A granite-gabbro complex from Madagascar: constraints on melting of the lower crust. Contributions to Mineralogy and Petrology 145(5), 585-599.

Meira, V.T., Garcia-Casco, A., Juliani, C., Almeida, R.P., Schorscher, J.H.D., 2015. The role of intracontinental deformation in supercontinent assembly: insights from the Ribeira Belt, Southeastern Brazil (Neoproterozoic West Gondwana). Terra Nova 27(3), 206-217.

Merdith, A. S., Collins, A. S., Williams, S. E., Pisarevsky, S., Foden, J. D., Archibald, D. B., Blades, M.L., Alessio, B., Armstead, S., Plavsa, D., Clark, C., Müller, R.D., 2017. A full-plate global reconstruction of the Neoproterozoic. Gondwana Research 50, 84-134.

Miller, R. M. G., Frimmel, H. E., Halverson, G. P., 2009. Passive continental margin evolution. Neoproterozoic to Early Palaeozoic evolution of Southwestern Africa. In: Gaucher, C., Sial, A.N., Halverson, G.P., Frimmel, H.E. (Eds.), Neoproterozoic-Cambrian Tectonics, Global Change and Evolution: A Focus on Southwestern Gondwana. Developments in Precambrian Geology 16. Elsevier, pp. 161-181.

Mohan, M.R., Sarma, D.S., McNaughton, N.J., Fletcher, I.R., Wilde, S.A., Siddiqui, M.A., Rasmussen, B., Krapez, B., Gregory, C.J., Kamo, S.L., 2014. SHRIMP zircon and titanite U-Pb ages, Lu-Hf isotope signatures and geochemical constraints for ~2.56 Ga granitic magmatism in Western Dharwar Craton, Southern India: Evidence for short-lived Neoarchean episodic crustal growth?. Precambrian Research, 243, 197-220.

Mole, D.R., Barnes, S.J., Taylor, R.J.M., Kinny, P.D., Fritz, H., 2018. A relic of the Mozambique Ocean in south-east Tanzania. Precambrian Research, 305, 386-426.

Nascimento, D.B., Ribeiro, A., Trouw, R.A.J., Schmitt, R.S., Passchier, C.W., 2016.

Stratigraphy of the Neoproterozoic Damara Sequence in northwest Namibia: Slope to basin sub-marine mass-transport deposits and olistolith fields. Precambrian Research 278, 108-125.

Paquette, J.-L., Nédélec, A., 1998. A new insight into Pan-African tectonics in the East–West Gondwana collision zone by U–Pb zircon dating of granites from central Madagascar. Earth and Planetary Science Letters 155(1-2), 45-56.

Praveen, M.N., Santosh, M., Yang, Q.Y., Zhang, Z.C., Huang, H., Singanenjam, S., Sajinkumar, K.S., 2014. Zircon U–Pb geochronology and Hf isotope of felsic volcanics from Attappadi, southern India: implications for Neoarchean convergent margin tectonics. Gondwana Research 26(3-4), 907-924.

Reddy, S.M., Collins, A.S. and Mruma, A. 2003. Complex High-Strain Deformation in the Usagaran Orogen, Tanzania: Structural Setting of Palaeoproterozoic Eclogites. Tectonophysics. 375, 101-123.

Roig, J.Y.; Tucker, R.D.; Delor, C.; Peters, S.G. and Théveniaut, H., 2012. Carte Géologique de la République de Madagascar à 1/1,000,000. Ministère des Mines, PGRM, Antananarivo, République Madagascar 1.

Sarma, D.S., McNaughton, N.J., Belusova, E., Mohan, M.R., Fletcher, I.R., 2012. Detrital zircon U–Pb ages and Hf-isotope systematics from the Gadag Greenstone Belt: Archean crustal growth in the western Dharwar Craton, India. Gondwana Research 22, 843–854.

Scherer, E.; Munker, C.; Mezger, K., 2001. Calibration of the Lutetium-Hafnium Clock. Science 293(5536), 1766-1766.

Schmitt, R.S., Fragoso, R.A., Collins, A.S., 2018. Suturing Gondwana in the Cambrian: The Orogenic Events of the Final Amalgamation. In: Siegesmund, S., Basei, M.A.S., Oyhantçabal, P., Oriolo, S. (Eds.), Geology of Southwest Gondwana. Springer International Publishing, pp. 411-432.

Schmitt, R.S., Trouw, R.A.J., Schmus, W.R.V., Armstrong, R., Stanton, N.S.G., 2016. The tectonic significance of the Cabo Frio Tectonic Domain in the SE Brazilian margin: a Paleoproterozoic through Cretaceous saga of a reworked continental margin. Brazilian Journal of Geology 46, 37-66.

Schofield, D.I., Thomas, R.J., Goodenough, K.M., De Waele, B., Pitfield, P.E.J., Key, R.M., Bauer, W., Walsh, G.J., Lidke, D.J., Ralison, A.V., Rabarimanana, M., Rafahatelo, J.M., Randriamananjara, T., 2010. Geological evolution of the Antongil craton, NE Madagascar. Precambrian Research 182(3), 187-203.

Sláma, J., Košler, J., Condon, D.J., Crowley, J.L., Gerdes, A., Hanchar, J.M., Horstwood, M.S.A., Morris, G.A., Nasdala, L., Norberg, N., Schaltegger, U., Schoene, B., Tubrett, M.N., Whitehouse, M.J., 2008. Plešovice zircon - A new natural reference material for U–Pb and Hf isotopic microanalysis. Chemical Geology 249(1-2), 1-35.

Stern, R.J., 1994. Arc assembly and continental collision in the Neoproterozoic East African Orogen: Implications for the consolidation of Gondwanaland. Annual Review of Earth and Planetary Sciences 22, 319-351.

Stern, R.J., 2002. Crustal evolution in the East African Orogen: A neodymium isotopic perspective. Journal of African Earth Sciences 34, 109-117.

Storey, M., Mahoney, J.J., Saunders, A.D., Duncan, R.A., Kelley, S.P., Coffin, M.F., 1995. Timing of hot spot-related volcanism and the breakup of Madagascar and India. Science 267(5199), 852-855.

Teklay, M., Kroner, A., Mezger, K., Oberhansli, R., 1998. Geochemistry, Pb–Pb single zircon ages and Nd–Sr isotope composition of Precambrian rocks from southern and eastern Ethiopia: implications for crustal evolution in East Africa. Journal of African Earth Sciences 26, 207- 227.

Thomas, R.J., De Waele, B., Schofield, D.I., Goodenough, K.M., Horstwood, M., Tucker, R., Bauer, W., Annells, R., Howard, K., Walsh, G., Rabarimanana, M., Rafahatelo, J.M., Ralison, A.V. and Randriamananjara, T., 2009. Geological evolution of the Neoproterozoic Bemarivo Belt, northern Madagascar. Precambrian Research, 172(3-4): 279-300.

Thomas, R.J., Spencer, C., Bushi, A.M., Baglow, N., Boniface, N., de Kock, G., Horstwood, M.S.A., Hollick, L., Jacobs, J., Kajara, S., Kamihanda, G., Key, R.M., Maganga, Z., Mbawala, F., McCourt, W., Momburi, P., Moses, F., Mruma, A., Myambilwa, Y., Roberts, N.M.W., Saidi, H., Nyanda, P., Nyoka, K., Millar, I., 2016. Geochronology of the central Tanzania Craton and its southern and eastern orogenic margins. Precambrian Research 277, 47-67

Tucker, R.D., Ashwal, L.D., Handke, M.J., Hamilton, M.A., Le Grange, M., Rambeloson, R.A., 1999. U-Pb geochronology and isotope geochemistry of the Archean and Proterozoic rocks of north-central Madagascar. The Journal of Geology 107(2), 135-153.

Tucker, R.D., Kusky, T.M., Buchwaldt, R., Handke, M.J., 2007. Neoproterozoic nappes and superposed folding of the Itremo Group, west-central Madagascar. Gondwana Research 12(4), 356-379.

Tucker, R.D., Roig, J.Y., Macey, P.H., Delor, C., Amelin, Y., Armstrong, R.A., Rabarimanana, M.H., Ralison, A. V., 2011a. A new geological framework for south-central Madagascar, and its relevance to the “out-of-Africa” hypothesis.

Precambrian Research 185, 109-130.

Tucker, R.D., Roig, J.Y., Delor, C., Amelin, Y., Goncalves, P., Rabarimanana, M.H., R., A.V., Belcher, R.W., 2011b. Neoproterozoic extension in the Greater Dharwar Craton: a reevaluation of the “Betsimisaraka suture” in Madagascar. Canadian Journal of Earth Sciences 48(2), 389-417.

Tucker, R.D., Roig, J.Y., Moine, B., Delor, C., Peters, S.G., 2014. A geological synthesis of the Precambrian shield in Madagascar. Journal of African Earth Sciences 94, 9-30.

Vermeesch, P., 2018. IsoplotR: A free and open toolbox for geochronology. Geoscience Frontiers, 9(5), 1479-1493.

Whitehouse, M.J., Windley, B.F., Stoeser, D.B., Al-Khirbash, S., Ba-Bttat, M.A.O., Haider, A., 2001. Precambrian basement character of Yemen and correlations with Saudi Arabia and Somalia. Precambrian Research 105, 357-369.

Woodhead, J., Hergt, J., Shelley, M., Eggins, S., Kemp, R., 2004. Zircon Hf-isotope analysis with an excimer laser, depth profiling, ablation of complex geometries, and concomitant age estimation. Chemical Geology 209(1-2), 121-135.

Yang, Q. Y., Santosh, M., 2015. Zircon U–Pb geochronology and Lu–Hf isotopes from the Kolar greenstone belt, Dharwar Craton, India: Implications for crustal evolution in an ocean-trench-continent transect. Journal of Asian Earth Sciences 113, 797-811.

Zhou, J.-L., Li, X.-H., Tang, G.-Q., Liu, Y. and Tucker, R.D., 2018. New evidence for a continental rift tectonic setting of the Neoproterozoic Imorona-Itsindro Suite (central Madagascar). Precambrian Research, 306: 94-111.

Zhou, J.-L., Shao, S., Luo, Z.-H., Shao, J.-B., Wu, D.-T. and Rasoamalala, V., 2015.

Geochronology and geochemistry of Cryogenian gabbros from the Ambatondrazaka area, east-central Madagascar: Implications for Madagascar-India correlation and Rodinia paleogeography. Precambrian Research, 256(0): 256-270.

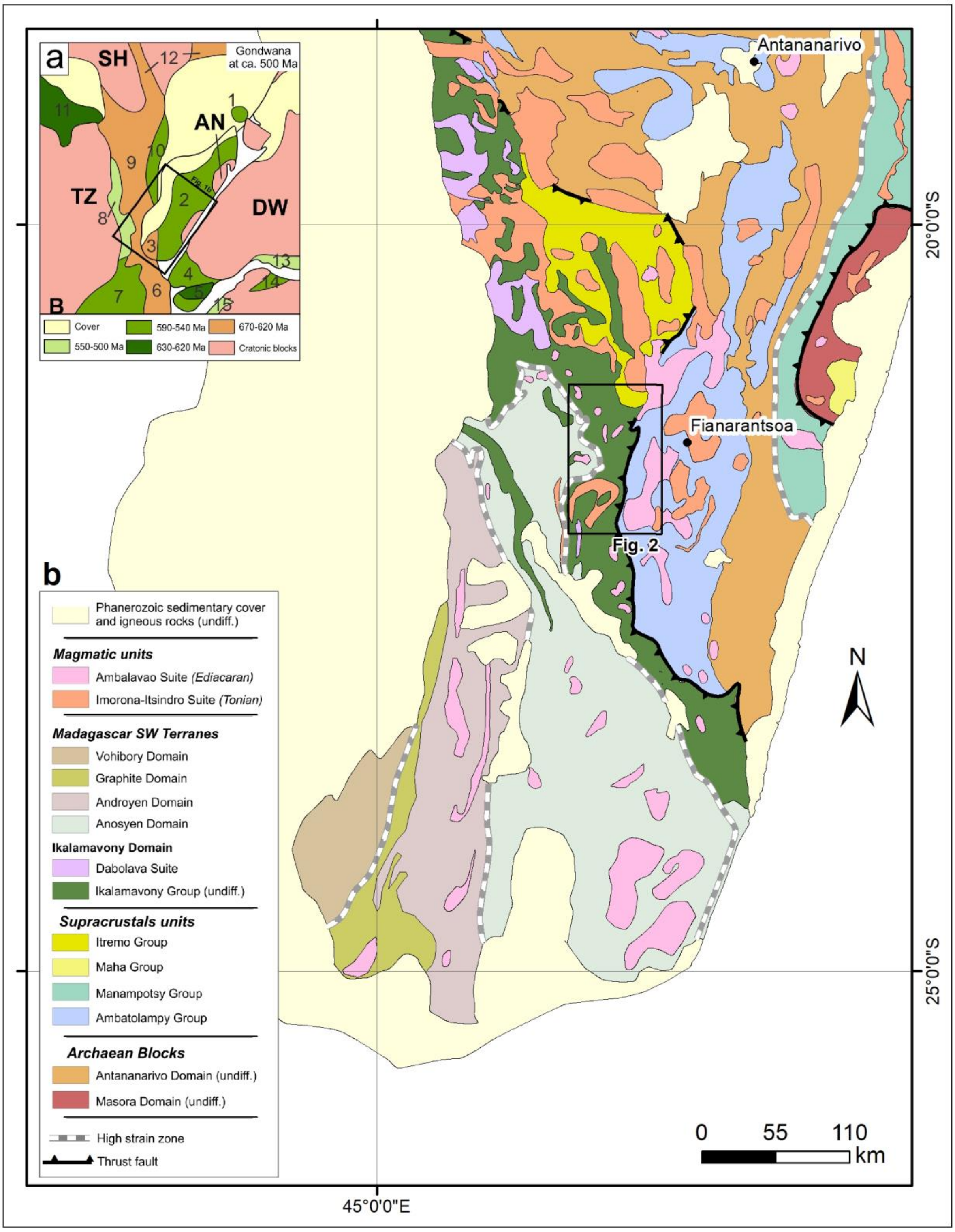


Figure 1: a) Gondwana configuration at ca. 500 Ma with pre-Gondwana cratons sutured by Neoproterozoic-Cambrian mobile belts (modified from Schmitt et al., 2018), highlighting central and south Madagascar. Pink polygons show Precambrian cratons, orange and green polygons show different orogenic events according to age intervals. The letters and number on the map represent Gondwana cratons and mobile belts, respectively. Abbreviations: EA=East Antarctica; B=Bangweulu Block; AN=Antogil Domain; DW=Dharwar; SH=Sahara and TZ=Tanzania; Legend key: 1-Seychelles; 2-Madagascar (Bemarivo, Antananarivo, Itremo-Ikalamavony, Androyen and Anosyen); 3-Madagascar (Vohibory); 4-Southern Granulites; 5-Sri Lanka; 6-Eastern Granulite; 7-Zambesi; 8-Western Granulite; 9-Arabian/Nubian Shield (South); 10-Galana (Azania); 11-Oubanguides; 12-Arabian/Nubian Shield (North); 13-Eastern Ghats; 14-Reworked border of the Napier Complex; 15-Prince Olaf Coast/Kemp Land; **b)** Simplified geological map of Central and Southern Madagascar showing the distribution of (volcano)metasedimentary units from Antananarivo and Itremo-Ikalamavony domains, Maha and

Manampotsy groups and undifferentiated Masora, Anosyen, Androyen and Vohibory domains, Imorona-Itsindro and Ambalavao suites and Phanerozoic undifferentiated rocks (modified from Roig et al., 2012).

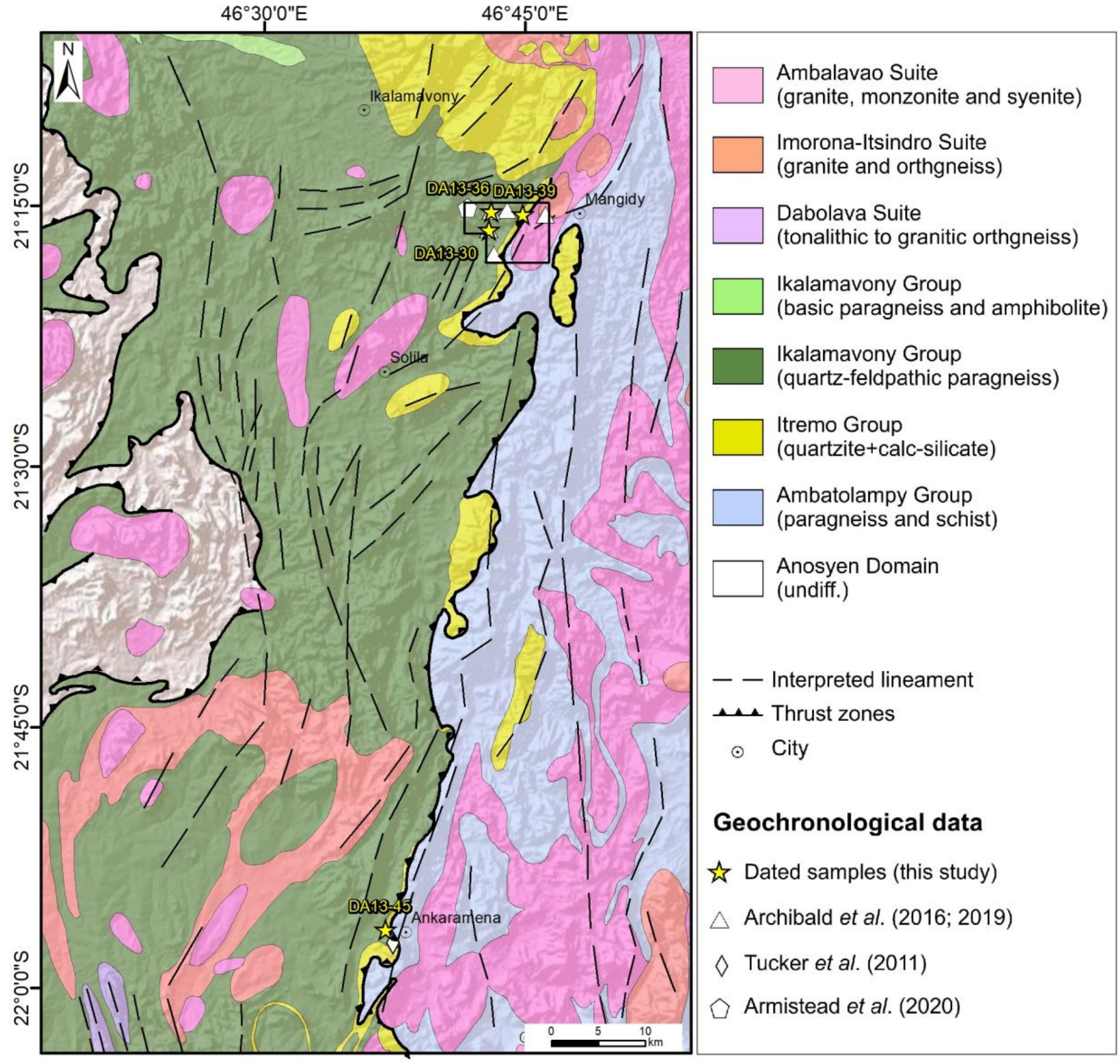


Figure 2: Geological map of western region of central Madagascar with location of dated samples, shown in figure 1 (modified from Roig et al., 2012). The mapped area is shown in the black polygon.

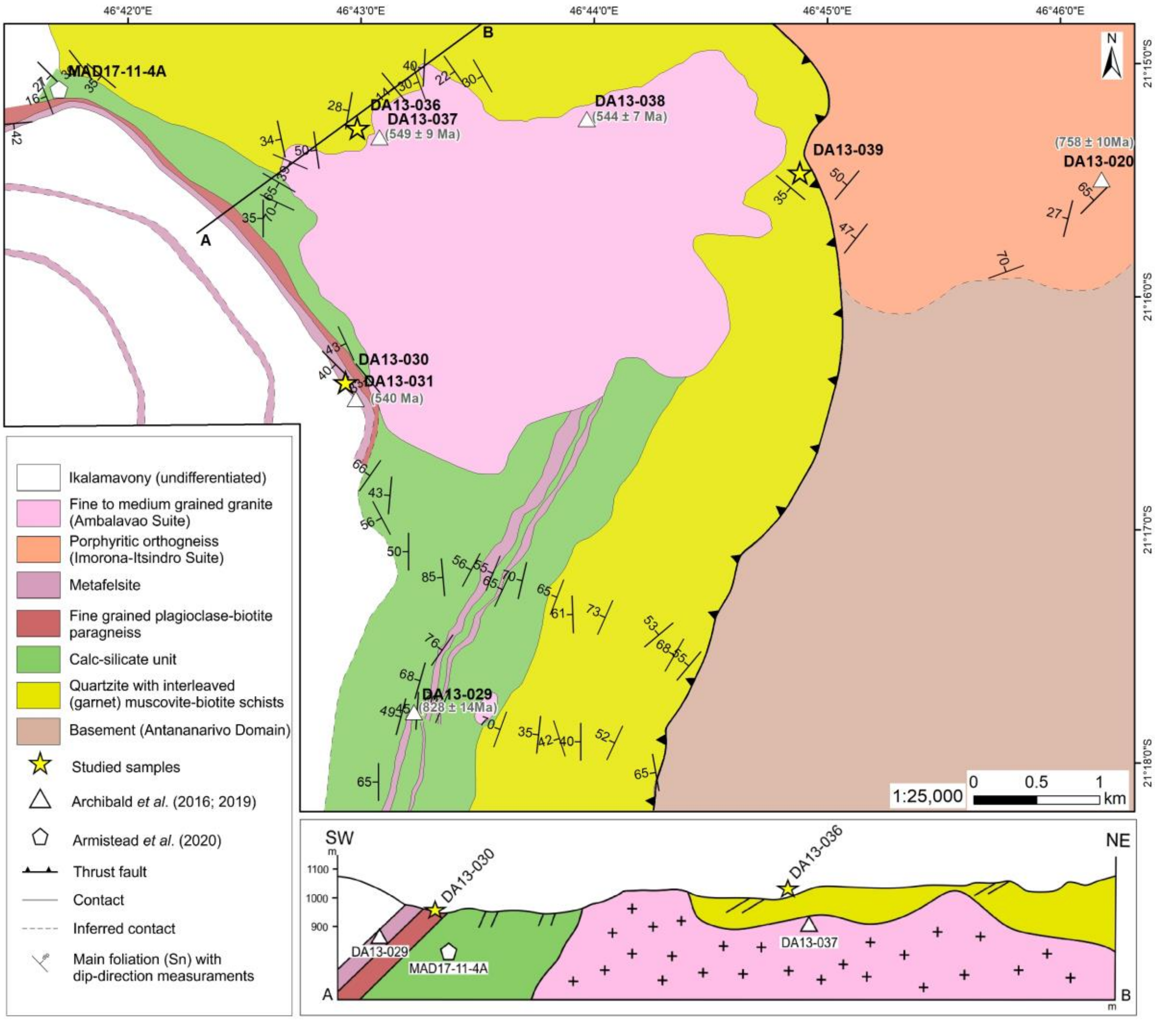


Figure 3: Detailed geological map of the area with measurements of main foliation, sample locations of this study (except DA13-045), and sample location of Archibald et al. (2016, 2019), and other sample, not analyzed. In addition, a SW-NE cross section, not scaled, is presented and the location of the area according to the Figure 2.

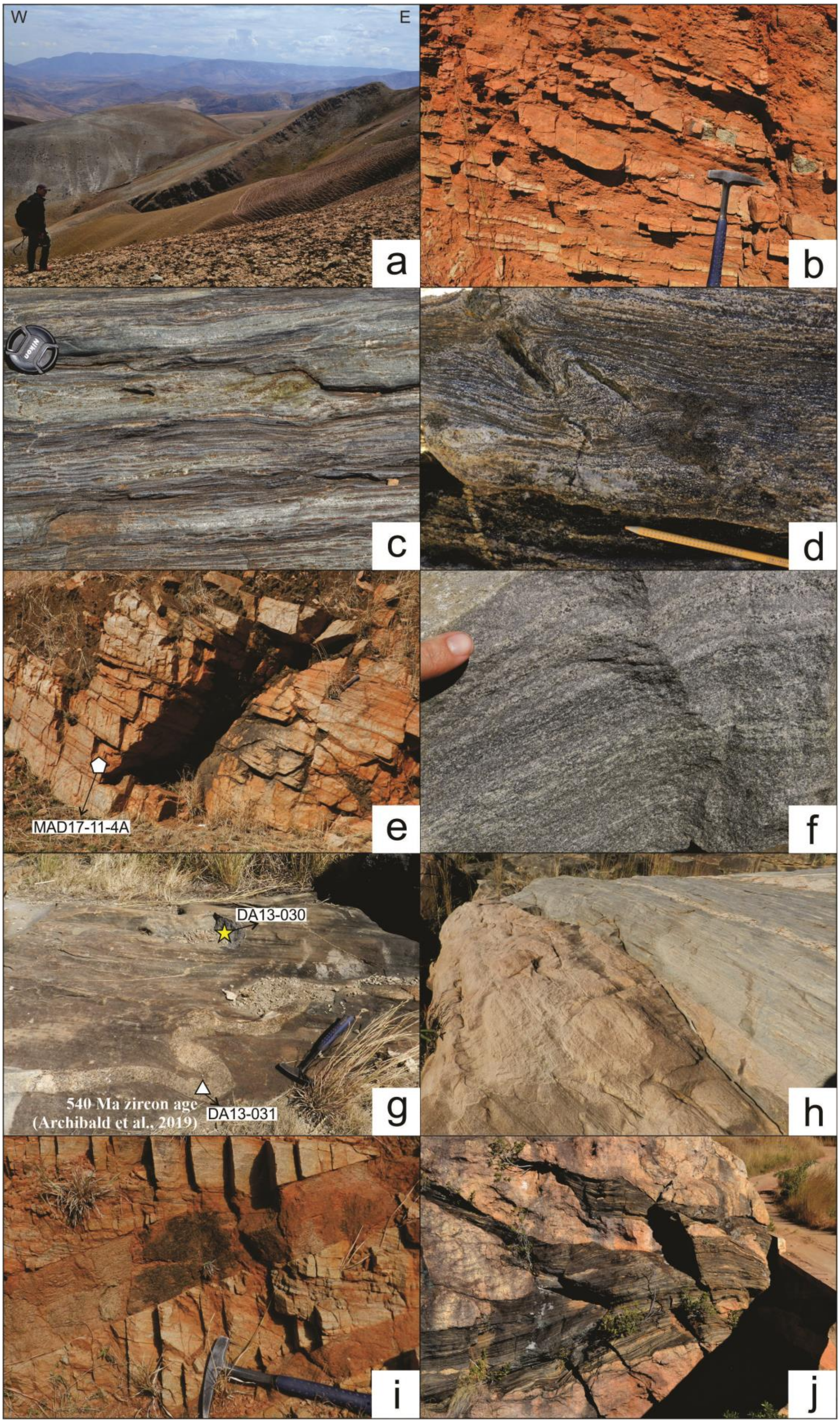

W
E
a
b
c
d
MAD17-11-4A
e
f
DA13-030
540 Ma zircon age
(Archibald et al., 2019)
DA13-031
g
h
i
j

Figure 4: a) Mountain ranges of the quartzite unit with interlayered (garnet- sillimanite) muscovite-biotite schists. The sedimentary layering is parallel to the tectonic foliation and both dip to west; **b)** Intercalation of quartzite and schist on the upper package of the quartzite unit; **c)** Diopside-gneiss of the bottom of the calc-silicate unit showing mylonitic foliation; **d)** Asymmetric fold in laminated white marble; **e)** Quartzite layer within the calc-silicate unit – note sample MAD17-11-4A – dated by Armstead et al. (in review); **f)** Fine grained homogeneous plagioclase-biotite paragneiss; **g)** Folded granitic dyke cross-cutting the plagioclase-biotite-paragneiss – note samples DA13-031 and DA13-030; dated by D.B. Archibald unpublished and our study, respectively; **h)** Contact between the plagioclase-biotite paragneiss (grayish on top) and the metafelsite (pinkish on bottom). Note concordant tectonic foliation; **i)** Foliated mafic dykes (granodioritic in composition) that cross-cut the quartzite; **j)** Ambalavao Suite pink granite intruding the calc-silicate sequence.

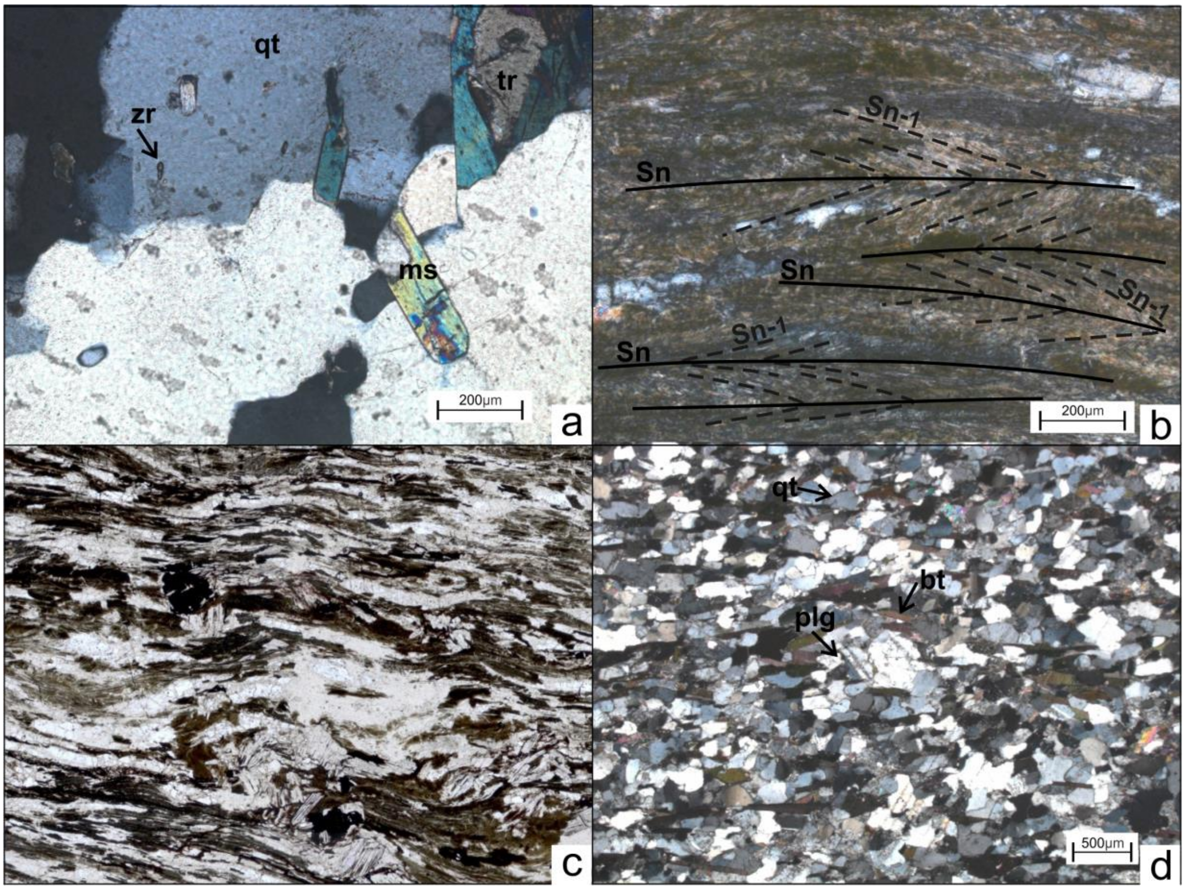


Figure 5: a) Photomicrograph of thin layered quartzite near the contact with the calc-silicate unit, showing muscovite, tremolite and zircon, in cross-polarized view; **b)** Cross-polarized view of a (Garnet-sillimanite) muscovite-biotite schist from the quartzite unit with tight chevron folds, showing that the Sn tectonic foliation (axial plane of the micro-folds), measured in the field, corresponds to a second phase of ductile deformation. The folded metamorphic minerals represent an older foliation (Sn-1); **c)** Plane-polarized view of the sillimanite-garnet schist, showing asymmetric micro-folds east-vergent; **d)** Fine grained plagioclase-biotite-paragneiss in thin section. This is sample DA-13-030, dated here.

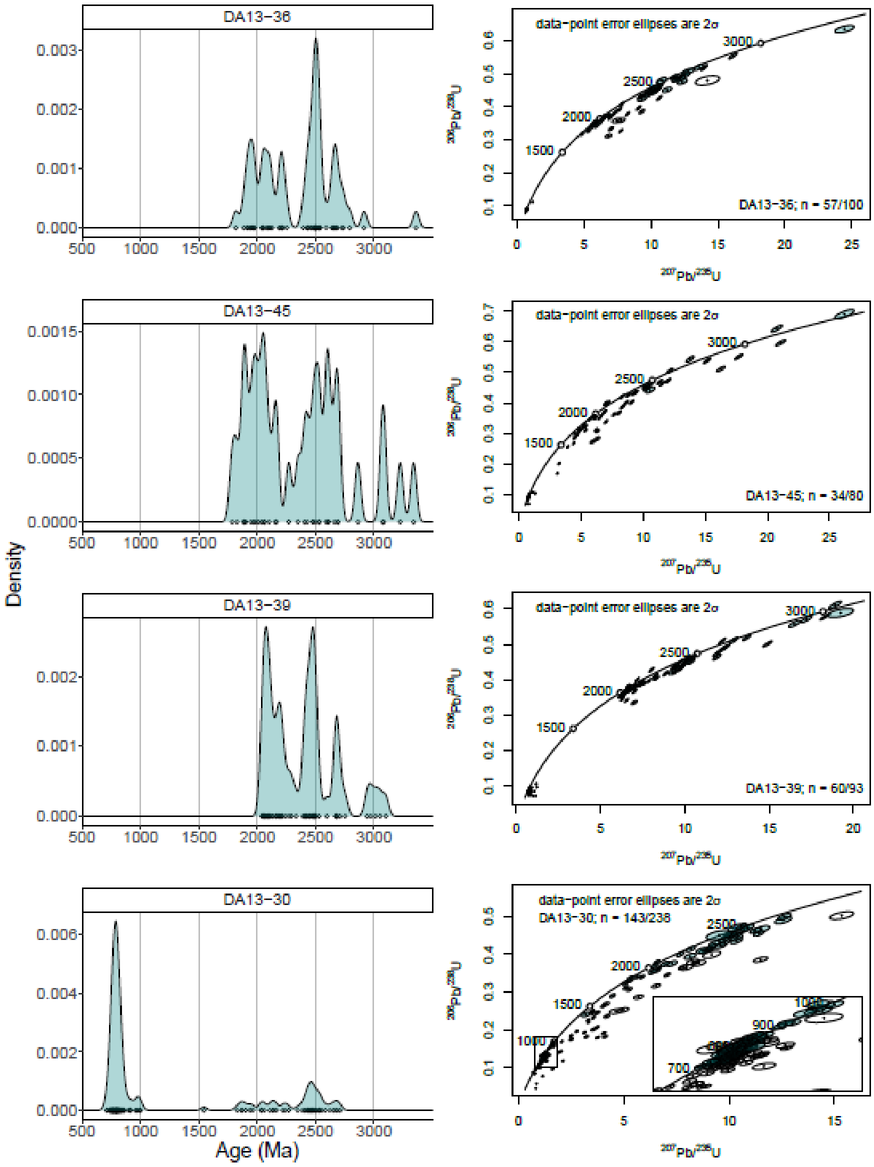


Figure 6: Probability density plots (left) and Concordia (right) diagrams for analyzed detrital zircon grains from samples DA13-036 (a), DA13-045 (b), DA13-039 (c) and DA13-030 (d). Within the Concordia diagram the analysis within 10% of concordance are coloured, and all other analyses are white.

Figure 7: Selected CL images of zircon grains with U-Pb and Lu-Hf analysis spots and Th/U ratios from samples: a) DA13-036 b) DA13-045 c) DA13-039 d) DA13-030.

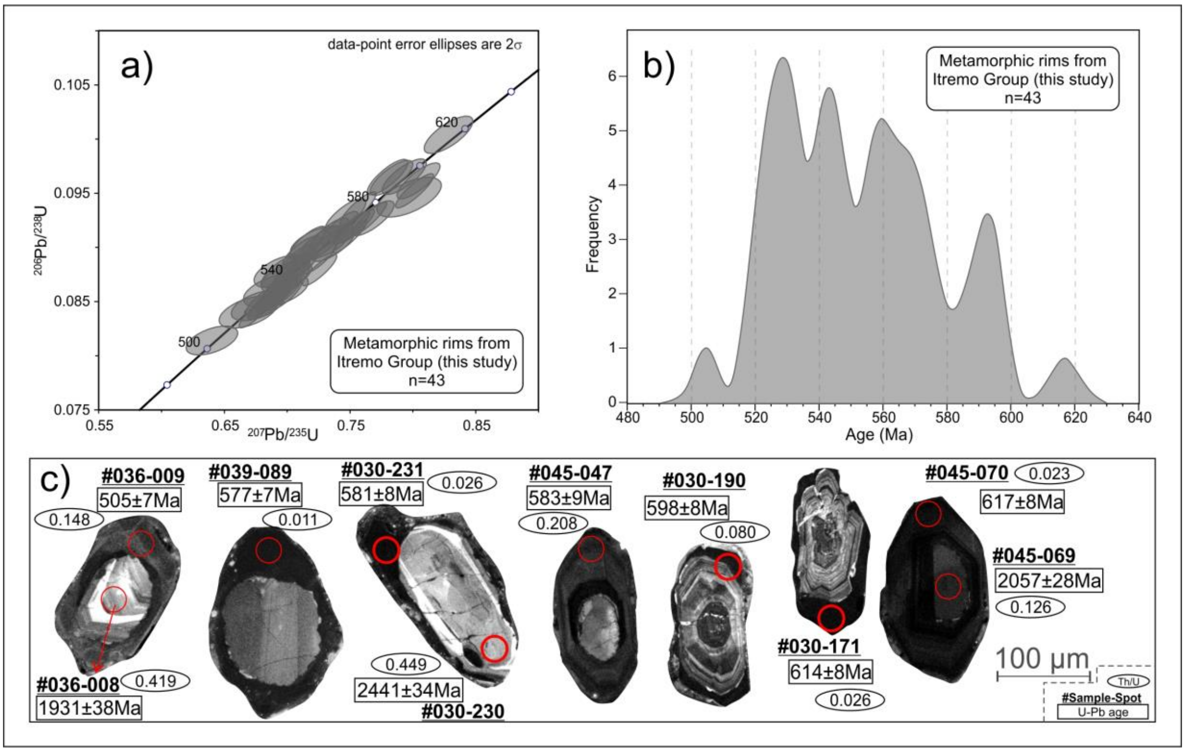


Figure 8: Metamorphic rim data from three quartzite samples. **a)** U-Pb concordia of metamorphic rim data; **b)** histogram diagram with the same data as in (a) and **c)** selected CL images from metamorphic zircon domains showing metamorphic rims with U-Pb ages and Th/U ratios. Data from the sample DA-030 are shown only in Fig. 8c.

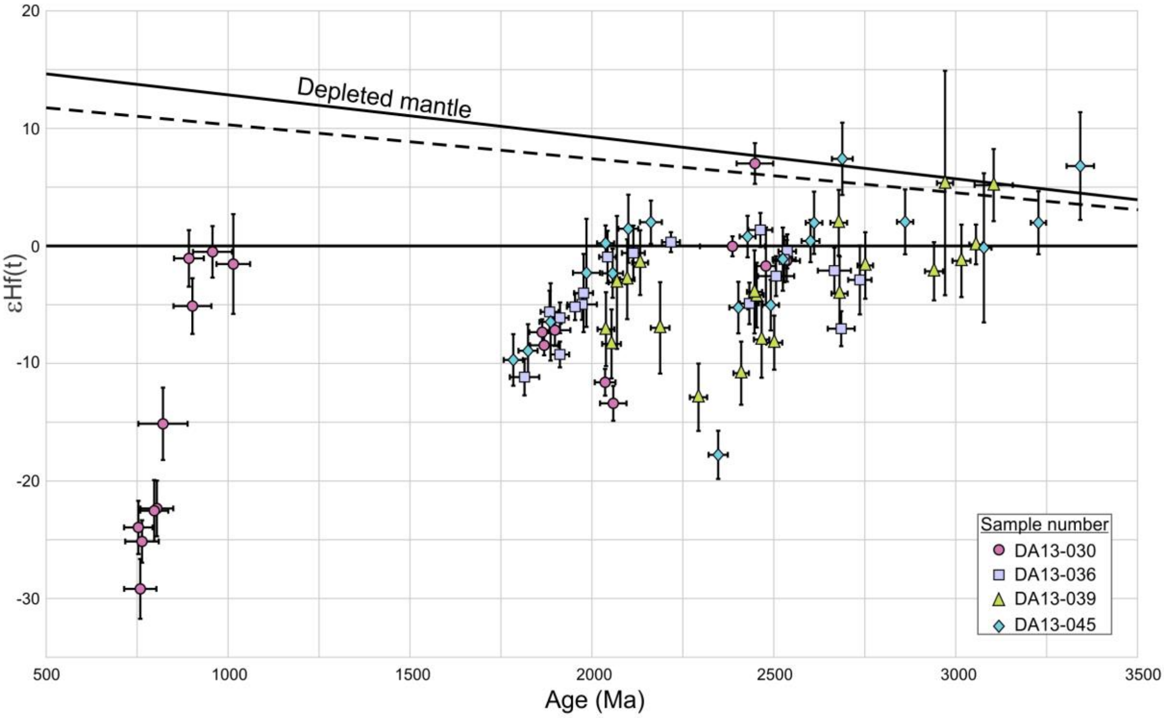


Figure 9: Hf isotopic evolution diagram with data obtained on detrital zircons from the four samples of this study. Error bars are in 2 sigma.

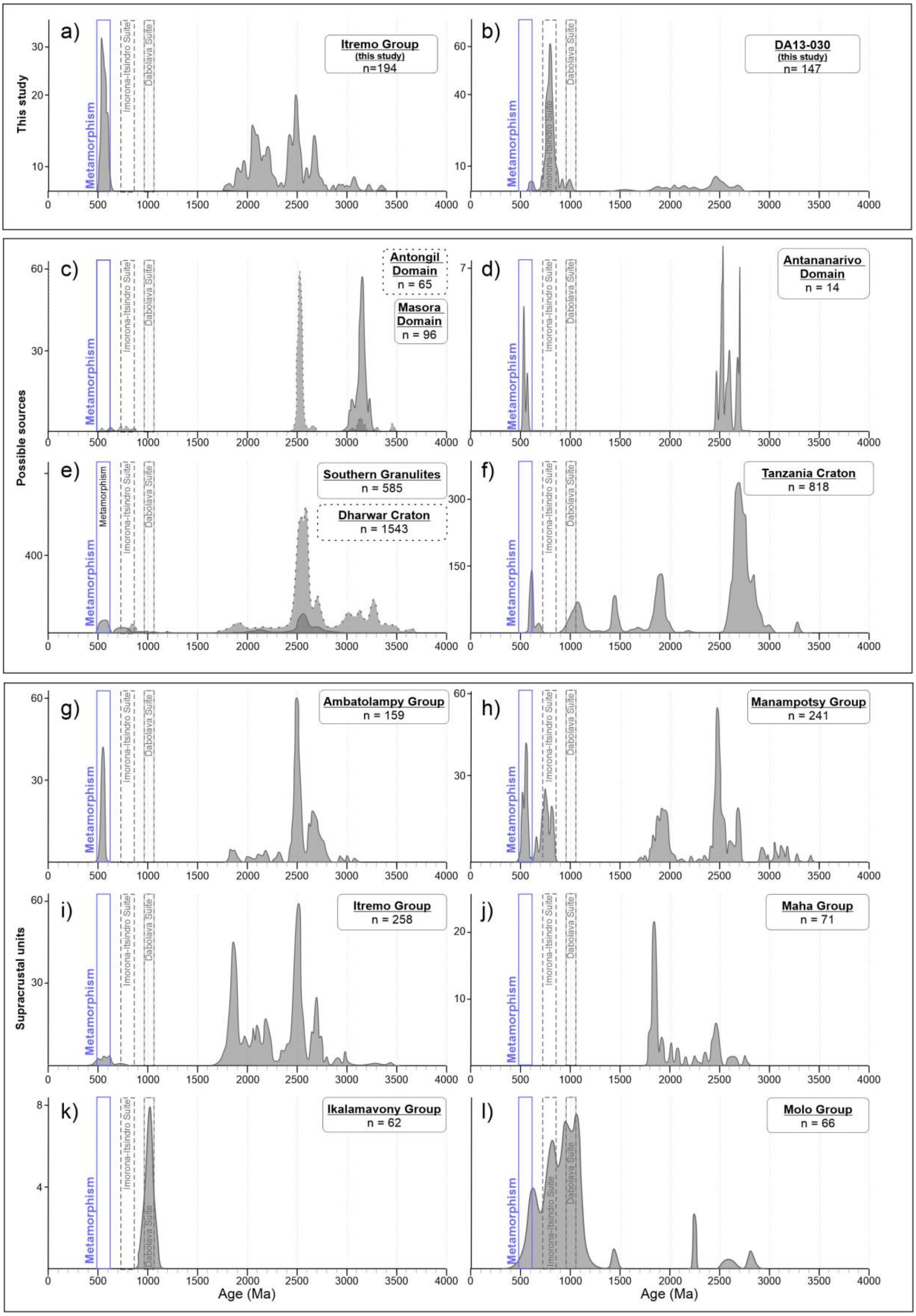


Figure 10: Comparative U–Pb age distribution of detrital zircon for supracrustal units in central Madagascar and their possible sources.. **a)** DA13-036, DA13-039 and DA13-045 (this study); **b**) DA13-030 (this study); **c)** Antongil and Masora domains (BGS-USGS-GLW, 2008; Collins et al. 2003; Schofield et al. 2010); **d)** Antananarivo Domain – Betsiboka Suite (BGS-USGS-GLW, 2008); **e)** Dharwar Craton (Collins et al. 2015;

Glorie et al. 2014; Ishwar-Kumar et al. 2013; Jayananda et al. 2013; Lancaster et al. 2015; Maibam et al. 2016); **f)** Tanzania Craton (Thomas et al. 2016); **g)** Ambatolampy Group (Archibald et al. 2015); **h)** Manampotsy Group (BGS-USGS-GLW, 2008; Tucker et al. 2011a); **i)** Itremo Group (Cox et al. 1998, 2004; De Waele et al. 2011; Fitzimons and Hulscher 2005; Tucker et al. 2011); **j)** Maha Group (De Waele et al. 2011); **k)** Ikalamavony Group (Tucker et al. 2011) and **l)** Molo Group (Cox et al., 2004).

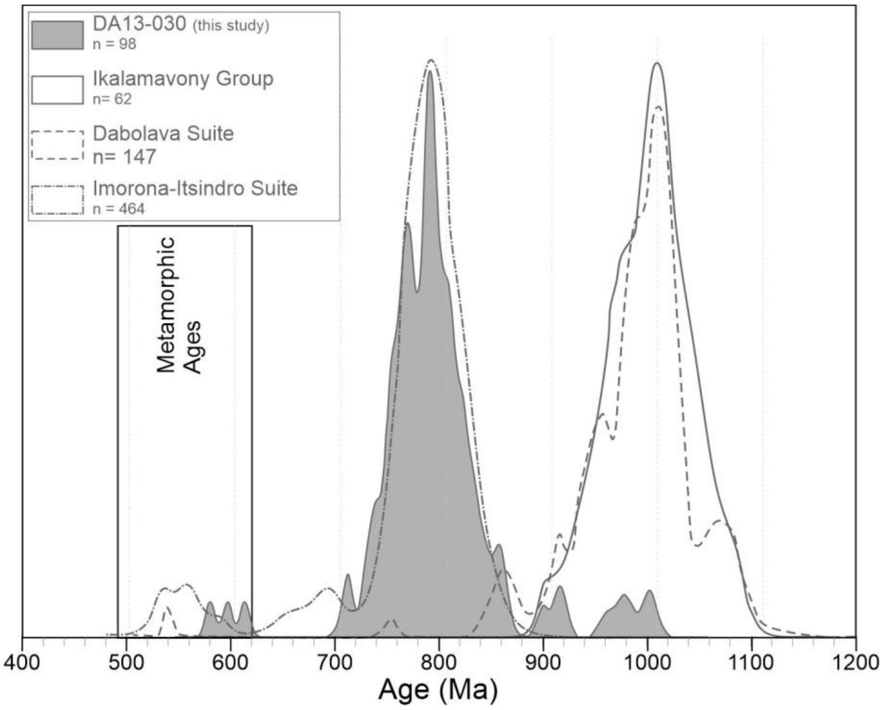
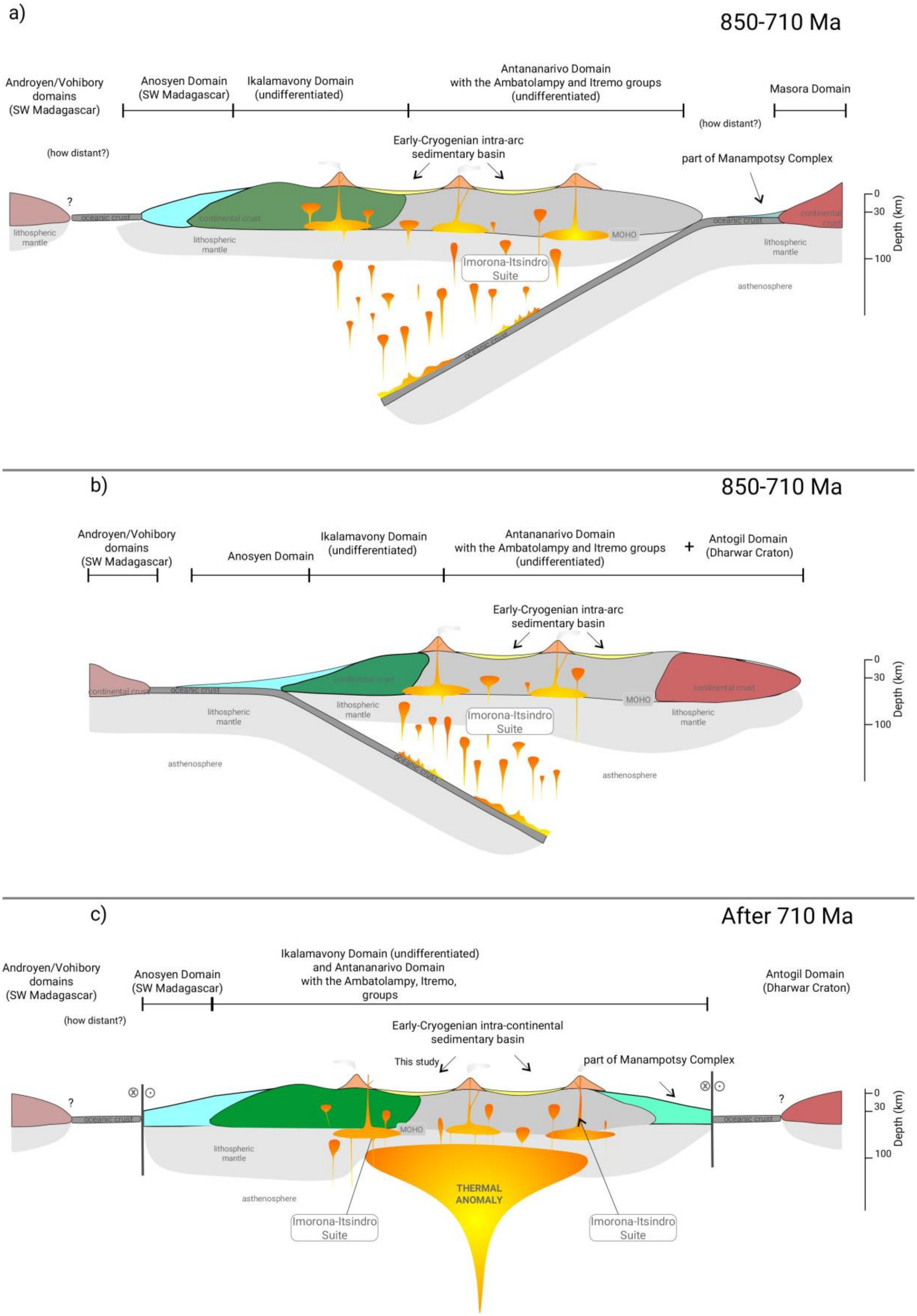


Figure 11: Comparative probability and density diagram to identify the possible main source for the population younger than 1.1 Ga, also more abundant, of sample DA13-030. U-Pb zircon data from Ikalamavony Group, Dabolava Suite and Imorona-Itsindro Suite is compiled from Tucker et al. (2011), Archibald et al. (2016, 2018) and BGS-USGS-GLW (2008).



1525
1526

1527 **Figure 12:** Different tectonic settings for Late Tonian Early Cryogenian interval (850-710 Ma) modified from
1528 diverse authors, to envisage the depositional scenario for the basin represented by sample DA13-030 from this
1529 study. **a)** West-dipping subduction related to a convergent continental margin between Antananarivo Domain
1530 (Madagascar) and Dharwar Craton (India) interpreted by Collins (2006), Key et al. (2011) and Kröner et al.

(2000); **b)** East-dipping subduction and the deposition of supracrustal units from the Anosyen Domain above Antananarivo Domain by Boger et al. (2014) and **c)** After 710 Ma an intracontinental setting would prevail late-Imorona-Itsindro magmatic activity due to subduction stall (this alternative is a variation of the model byTucker et al., 2011b, 2014).

FIGURE CAPTIONS

Figure 1: a) Gondwana configuration at ca. 500 Ma with pre-Gondwana cratons sutured by Neoproterozoic-Cambrian mobile belts (modified from Schmitt et al., 2018), highlighting central and south Madagascar. Pink polygons show Precambrian cratons, orange and green polygons show different orogenic events according to age intervals. The letters and number on the map represent Gondwana cratons and mobile belts, respectively. Abbreviations: EA=East Antarctica; B=Bangweulu Block; DW=Dharwar; SH=Sahara and TZ=Tanzania; Legend key: 1-Seychelles; 2-Madagascar (Bemarivo, Antananarivo, Itremo-Ikalamavony, Androyen and Anosyen); 3-Madagascar (Vohibory); 4-Southern Granulites; 5-Sri Lanka; 6-Eastern Granulite; 7-Zambesi; 8-Western Granulite; 9-Arabian/Nubian Shield (South); 10-Galana (Azania); 11-Oubanguides; 12-Arabian/Nubian Shield (North); 13-Eastern Ghats; 14-Reworked border of the Napier Complex; 15-Prince Olaf Coast/Kemp Land; b) Simplified geological map of central and southern Madagascar showing the distribution of (volcano)metasedimentary units from Antananarivo and Itremo-Ikalamavony domains, Maha and Manampotsy groups and undifferentiated Masora, Anosyen, Androyen and Vohibory domains, Imorona-Itsindro and Ambalavao suites and Phanerozoic undifferentiated rocks (modified from Roig et al., 2012).

Figure 2: Geological map of western region of central Madagascar with location of dated samples, shown in figure 1 (modified from Roig et al., 2012). The area mapped in detail is shown in the black polygon.

Figure 3: Detailed geological map of the area with measurements of main foliation, sample locations of this study (except DA13-045), and sample locations of Archibald et al. (2016; 2019 and unpublished data), and other samples that were not analysed. In addition, a SW-NE cross section, not scaled, is presented and the location of the area according to the Figure 2.

Figure 4: a) Mountain ranges of the quartzite unit with interlayered (garnet-sillimanite) muscovite-biotite schists. The sedimentary layering is parallel to the tectonic foliation and both dip to west; b) Intercalation of quartzite and schist on the upper package of the quartzite unit; c) Diopside-gneiss from the bottom of the calc-silicate unit showing mylonitic foliation; d) Asymmetric fold in laminated white marble; e) Quartzite layer within the calc-silicate unit – note sample MAD17-11-4A – was dated by Armstead et al. (unpublished); f) Fine-grained homogeneous plagioclase-biotite paragneiss; g) Folded granitic dyke cross-cutting the plagioclase-biotite-paragneiss – note samples DA13-031 and DA13-030; dated by D.B. Archibald unpublished and our study, respectively; h) Contact between the plagioclase-biotite paragneiss (grayish on top) and the metafelsite (pinkish on bottom). Note concordant tectonic foliation; i) Foliated mafic dykes that cross-cut the quartzite; j) Ambalavao Suite pink granite intruding the calc-silicate sequence.

Figure 5: a) Photomicrograph of thin layered quartzite near the contact with the calc-silicate unit, showing quartz (qt), muscovite (ms), tremolite (tr) and zircon (zr), in cross-polarized view; b) Cross-polarized view of a (Garnet-sillimanite) muscovite-biotite schist from the quartzitic unit with tight chevron folds, showing that

the Sn tectonic foliation (axial plane of the micro-folds), measured in the field, corresponds to a second phase of ductile deformation. The folded metamorphic minerals represent an older foliation (Sn-1); c) Plane-polarized view of the sillimanite-garnet schist, showing asymmetric micro-folds east-vergent; d) Photomicrograph of fine-grained plagioclase-biotite-paragneiss in thin section, showing quartz (qt), plagioclase (plg) and biotite (bt). This is sample DA-13-030, dated here.

Figure 6: Probability density plots (left) and Concordia (right) diagrams for analysed detrital zircon grains for samples DA13-036 (a), DA13-045 (b), DA13-039 (c) and DA13-030 (d).

Figure 7: Selected CL images of zircon grains with U-Pb and Lu-Hf analysis spots and Th/U ratios from samples: a) DA13-036 b) DA13-045 c) DA13-039 d) DA13-030.

Figure 8: Metamorphic rim data from zircon in the three quartzite samples. a) U-Pb concordia diagram; b) probability density diagram with the same data as in (a) and c) selected CL images from metamorphic zircon domains showing metamorphic rims with U-Pb ages and Th/U ratios. Data from the sample DA-030 is shown only in fig. 8c.

Figure 9: Hf isotopic evolution diagram with data obtained on detrital zircons from the four samples of this study. Error bars are in 2 sigmas. In addition to CHUR, the evolution of depleted-mantle (Griffin et al., 2002) and the evolution curve for new crust derived from the upper-mantle (Dhuime et al., 2011) are shown.

Figure 10: Comparative U-Pb age distribution of detrital zircon for supracrustal units in central Madagascar and their possible sources. Yellow bars show the main intervals – described in item 4.1. a) DA13-036, DA13-039 and DA13-045 (this study);

b) DA13-030 (this study); c) Antongil and Masora domains (BGS-USGS-GLW, 2008; Collins et al. 2003; Schofield et al. 2010); d) Antananarivo Domain – Betsiboka Suite (BGS-USGS-GLW, 2008; Tucker et al., 1999; Kröner et al., 2000); e) Dharwar Craton (Collins et al. 2015; Glorie et al. 2014; Ishwar-Kumar et al. 2013; Jayananda et al. 2013; Lancaster et al. 2015; Maibam et al. 2016); f) Tanzania Craton (Thomas et al. 2016); g) Ambatolampy Group (Archibald et al. 2015); h) Manampotsy Group (BGS-USGS-GLW, 2008; Tucker et al. 2011a); i) Itremo Group (Cox et al. 1998, 2004; De Waele et al. 2011; Fitzimons and Hulscher 2005; Tucker et al. 2011); j) Maha Group (De Waele et al. 2011); k) Ikalamavony Group (Tucker et al. 2011) and l) Molo Group (Cox et al., 2004).

Figure 11: Comparative probability density diagram to identify the possible main source for sample DA13-030. U-Pb zircon data from Ikalamavony Group, Dabolava Suite and Imorona-Itsindro Suite are compiled from Tucker et al. (2011), Archibald et al. (2016, 2018) and BGS-USGS-GLW (2008).

Figure 12: Different tectonic settings for Late Tonian to Early Cryogenian interval (ca. 850-710 Ma) modified from diverse authors, to envisage the depositional scenario for the basin represented by sample DA13-030 from this study. a) West-dipping subduction related to a convergent continental margin between Antananarivo Domain (Madagascar) and Dharwar Craton (India) interpreted by Collins (2006), Key et al. (2011) and Kröner et al. (2000); b) East-dipping subduction and the deposition of supracrustal units from the Anosyen Domain above the Antananarivo Domain as suggested by Boger et al. (2014) and c) After 710 Ma an intracontinental setting would prevail late to the Imorona-Itsindro magmatic activity due to subduction stall (this alternative is a variation of the model by Tucker et al., 2011b, 2014).

LIST OF SUPPLEMENTARY TABLES

Supplementary table 1: U-Pb data of detrital zircons from samples DA13-036, DA13-045, DA13-039 and DA13-030 (LA-ICP-MS)

Supplementary table 2: Lu-Hf data of detrital zircons from samples DA13-030, DA13-036, DA13-039 and DA13-045 (LA-ICP-MS)