Structural evolution and medium-temperature thermochronology of central Madagascar: implications for Gondwana amalgamation

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Abstract

Madagascar occupied an important place in the amalgamation of Gondwana, and preserves a record of several Neoproterozoic events that can be linked to orogenesis of the East African Orogen. We integrate remote sensing, thermochronology and field data to unravel complex deformation in the Ikalamavony and Itremo domains of central Madagascar. The deformation sequence comprises a gneissic foliation (S1), followed by south to south-west directed, tight to isoclinal, recumbent folding (D2). These are overprinted by north-trending upright folds that formed during a ~E–W shortening event. Together these produced type 1 and type 2 fold interference patterns throughout the Itremo and Ikalamavony domains. Apatite U–Pb and muscovite and biotite Rb–Sr thermochronometers indicate that much of central Madagascar was thermally reset to at least ~500°C at c. 500 Ma. Deformation in west-central Madagascar occurred between c. 750 Ma and c. 550 Ma, and we suggest this deformation formed as a response to the c. 650 Ma collision of Azania with Africa along the Vohibory Suture in southwestern Madagascar. In eastern Madagascar, deformation is syn- to post-550 Ma, which formed in response to the final closure of the Mozambique Ocean along the Betsimisaraka Suture that amalgamated Madagascar with the Dharwar Craton of India.

Keywords: thermochronology, Gondwana, remote sensing, GIS, supercontinents

Supplementary material:

Supplementary A: Detailed geological map of central Madagascar

Supplementary B: Detailed methodology and geo/thermochronology results

Supplementary C: Isotopic data for geo/thermochronology

Supplementary D: Landsat images and structural interpretation examples

1. Introduction

The amalgamation of central Gondwana occurred through convergence at several discrete subduction and collisional zones; collectively forming the East African Orogen. Madagascar was located in the centre of Gondwana and provides an ideal natural laboratory to study how this supercontinent coalesced (Collins, 2006; Collins and Windley, 2002; Tucker et al., 1999). Of particular interest and contention, is how and when the Archaean nucleus of Madagascar amalgamated with the Dharwar Craton of India to the east, and East Africa to the west, as well as smaller continental blocks of equivocal origin. Reconciling this tectonic history has major implications for global plate tectonic models during the Neoproterozoic (e.g. Merdith et al., 2017).

Madagascar is made up of several domains ranging in age from Archaean to Neoproterozoic (Figure 1b). The centre of Madagascar is made up of the Antananarivo Domain, which has a basement of c. 2500 Ma magmatic gneisses (Collins and Windley, 2002; Kröner et al., 2000; Tucker et al., 1999), known as the Betsiboka Suite (Roig et al., 2012), interleaved with the Ambatolampy Group granulite- and amphibolite-facies metasedimentary rocks (Archibald et al., 2015). To the east of the Antananarivo Domain is the Antongil-Masora Domain, which contains gneisses that are c. 3100 Ma and c. 2500 Ma, and are interpreted as a continuation of the Dharwar Craton of India (Armistead et al., 2017; Schofield et al., 2010; Tucker et al., 1999).

Overlying the Antananarivo Domain is the Itremo Group (Figure 1). Classified as a sub-domain of the Antananarivo Domain by Roig et al. (2012), the Itremo Group is comprised of quartzites, schists and marbles with maximum depositional age of c. 1600 Ma (Costa et al., Submitted; Cox et al., 1998; Fernandez et al., 2003). The Itremo Group is interpreted as a continental margin sequence that was deposited on the Antananarivo Domain basement (Cox et al., 1998; 2004). To the southwest, thrust over the Itremo Group, is the Ikalamavony Group within the Ikalamavony Domain, similarly made up of quartzites, schists and marbles but with a maximum depositional age of c. 1000 Ma (Archibald et al., 2017a; Tucker et al., 2014). In places the Ikalamavony Domain is in tectonic contact directly with the Antananarivo Domain basement, with no Itremo Group rocks separating them (Figure 1b; Figure 2). To the south of these metasedimentary terranes are the Proterozoic Anosyen, Androyen and Vohibory domains (Boger et al., 2014; Emmel et al., 2008; Jöns and Schenk, 2008). In northern Madagascar is the c. 800–700 Ma Bemarivo Belt, which formed as an exotic juvenile arc terrane that amalgamated with Madagascar at c. 520 Ma (Armistead et al., 2019; Jöns et al., 2009; Thomas et al., 2009).

The central Madagascan terranes discussed in this paper are bounded by two major sutures; the eastern Betsimisaraka Suture and the western Vohibory Suture (mapped as the Ampanihy shear zone in Figure 1). These resulted from at least two distinct major orogenic events that amalgamated central Madagascar, the Dharwar Craton of India and

Africa within Gondwana (Figure 1). However, the timing, location, and direction of subduction leading to these orogenic events remain contentious. Two end-member models are generally evaluated for the amalgamation of Madagascar; 1) that the Dharwar–central Madagascar collision (eastern suture) occurred in the late Archaean, and that central Madagascar and the Dharwar craton existed as the “Greater Dharwar Craton” through the entire Proterozoic eon (Tucker et al., 2011), and that widespread Neoproterozoic–Cambrian magmatism and metamorphism in Madagascar resulted from Madagascar–Africa collision (western suture); or 2) that the Dharwar Craton and central Madagascar were separate terranes that were sutured during a major Ediacaran–Cambrian East African orogenic event (the Malagasy Orogeny of Collins and Pisarevsky 2005), marked by the Betsimisaraka Suture in eastern Madagascar (Figure 1b). In this model, the central Madagascar–Africa collision occurred at c. 650–630 Ma (Collins and Pisarevsky, 2005; Collins and Windley, 2002; Emmel et al., 2008; Jöns and Schenk, 2011) as the East African Orogeny. Other authors have suggested a c. 750–650 Ma age for the eastern suture (Fitzsimons and Hulscher, 2005), or a c. 850–750 Ma age for the western suture (Moine et al., 2014). The proximity of these two suture zones makes it difficult to unravel the timing of events, as more recent events have the potential to overprint and obliterate the record of earlier events, such as through high temperature resetting of key minerals used for thermochronology and metamorphism.

The cross-cutting relationships and deformation history of the terranes that make up Madagascar can provide clues as to the timing of major orogenic events in Madagascar. Here we use structural geology to understand the deformation history of a part of central Madagascar, which lies between the two hypothesised suture zones. We have also collected U–Pb zircon, U–Pb apatite, Rb–Sr muscovite and Rb–Sr biotite thermochronology data to provide some absolute timing constraints on deformation in the region.

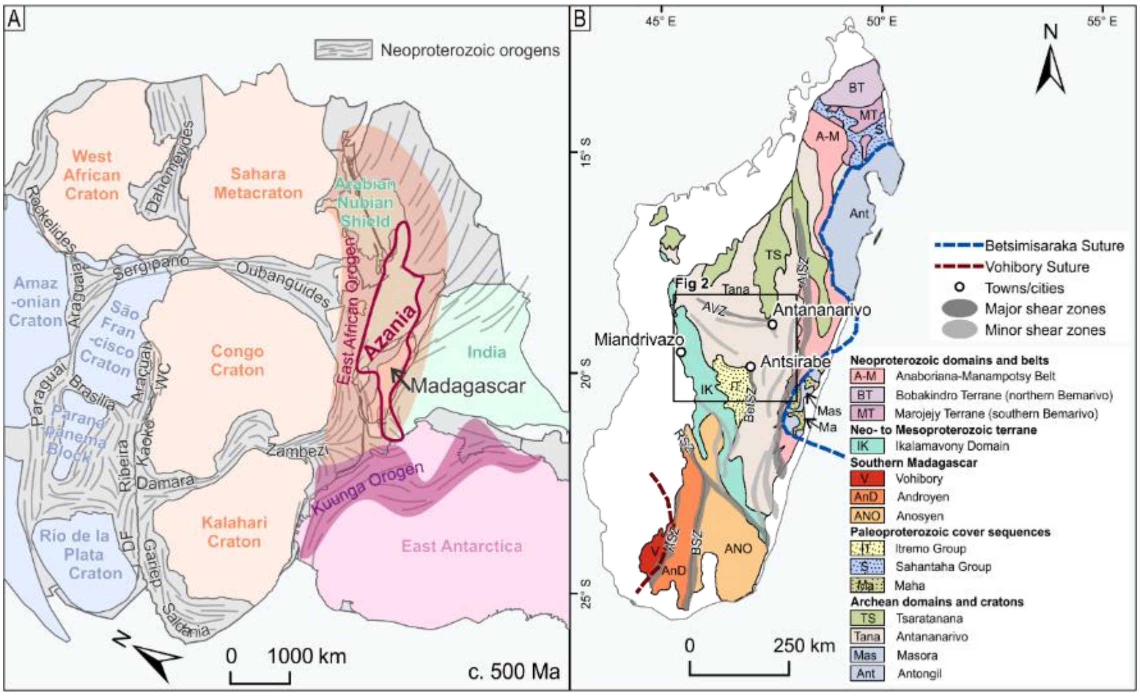

Figure 1 a) Tectonic map of Gondwana made using GPlates exported geometries from Merdith et al. (2017) in ArcGIS; projected in Hotine Oblique Mercator with Madagascar in the centre (reconstructed position, longitude=-75 and latitude=+40). DF=Dom Feliciano Belt, WC=West Congo; b) Present day map of the geological domains of Madagascar after (De Waele et al., 2011). AISZ=Angavo-Ifanadiana shear zone, AVZ=Antananarivo virgation zone, BetSZ=Betsileo shear zone, RSZ=Ranotsara shear zone, BSZ=Beraketa shear zone, ASZ=Ampanihy shear zone.

1.1. Regional structural and geochronological framework for central Madagascar

This study focuses on central Madagascar including parts of the Ikalamavony, and Antananarivo domains and the Itremo sub-domain (Figure 1b, Figure 2). Structural geology and various geochronological methods are used to define and distinguish deformation events in central Madagascar. Collins et al. (2003b) and Tucker et al. (2007) undertook comprehensive studies of the structure of the Itremo Group in central Madagascar. This area contains spectacularly folded sequences visible from satellite imagery. Collins et al. (2003b) interpreted a D1 event that produced 10 km scale recumbent, isoclinal folding predating c. 800–780 Ma intrusive rocks of the Imorona-Itsindro Suite. D2 was interpreted as a local deformation event that occurred synchronously with c. 800–780 Ma intrusions. D3 was interpreted as an east-west shortening event with thrusting and at least two phases of upright folding. D4 is expressed as post-550 Ma normal shearing and locally marks the boundary between the least metamorphosed parts of the Itremo Group and the granulite-facies Betsiboka Suite and Ambatolampy Group of the Antananarivo Domain (Betsileo Shear Zone; Collins et al. 2000). Tucker et al. (2007) interpreted a similar history for the Itremo Group with km-scale fold and thrust nappes, and east-directed vergence. This resulted in inversion and repetition of the Archaean Antananarivo Domain gneisses and the Proterozoic Itremo

Group, with high-grade (old) rocks being thrust over low-grade (young) rocks. The inversion was followed by east-west shortening that resulted in upright folding of nappes to produce km-scale fold interference patterns. This event occurred within a sinistral transpressive regime and was interpreted as being associated with the Ranotsara Shear Zone in southern Madagascar (Tucker et al., 2007). These two models for the Itremo Group differ in that Tucker et al. (2007) interpreted the timing of deformation as occurring after c. 720 Ma, whereas Collins et al. (2003b) interpreted the early nappes as forming before 800–780 Ma, and the upright folding having occurred after the c. 780 Ma intrusive rocks.

The region between the eastern-most part of our study area and the east coast of Madagascar was studied from a structural perspective in Collins et al. (2003a); Martelat et al. (2000); Nédélec et al. (2000); Raharimahefa and Kusky (2006, 2009); Raharimahefa et al. (2013). Interpretations of this region generally include a D1 event characterised by N-S striking foliations that dip to the west, with a top to the east sense of movement (Collins et al., 2003a; Nédélec et al., 2000). These rocks are reworked by D2 shear zones such as the Angavo Shear Zone and the Antananarivo virgation zone (Figure 1b) that underwent low-pressure, granulite conditions (Nédélec et al., 2000; Paquette and Nédélec, 1998). D3 is characterised by >20 km wide mylonitic high-strain zones and smaller discrete shear zones (Collins et al., 2003a). These dip gently to the west, with a top to the east sense of movement. D4 is characterised by poorly preserved late stage folding (Collins et al., 2003a). Raharimahefa and Kusky (2006, 2009) interpreted three folding events associated with ~N-S striking shear zones along the northeastern and southeastern margins of the Betsimisaraka Suture. A syn-kinematic granite within the Angavo Shear Zone constrains deformation here to c. 550 Ma (Raharimahefa and Kusky, 2010).

Precise dating of deformation in Madagascar is difficult due to resetting from successive overlapping tectono-thermal events. Latest metamorphism in the Anosyen and Androyen domains to the south of our study area is constrained to c. 580–520 Ma (Collins et al., 2012; de Wit et al., 2001; Martelat et al., 2000; Paquette et al., 1994) and attributed to high-strain shearing along the Ampanihy and Beraketa shear zones (Figure 1b) (Boger et al., 2015; Boger et al., 2014). Jöns and Schenk (2011) demonstrated that in southern Madagascar, high-grade metamorphism yielded ages of c. 650–600 Ma in the west, but recorded ages of c. 560–530 Ma in the east. In central Madagascar, U–Pb dating of zircon rims and titanite have been used to constrain latest metamorphism in the Itremo Group to c. 550–500 Ma (Tucker et al., 2007). Further east, between the eastern-most part of the study area and the east coast of Madagascar, metamorphism has been dated to c. 560–520 Ma (BGS-USGS-GLW, 2008; Collins et al., 2003c; Kröner et al., 2000). From this, it is clear that whatever event was taking place at c. 580–520 Ma, its effects were widespread and resulted in metamorphism throughout most of Madagascar, sparing, perhaps, the far southwest.

In this study we attempt to link up previous structural studies and further extend these interpretations to cover the entire central Madagascar region. We have used remotely sensed data such as satellite imagery and Landsat to interpret the structural framework of central Madagascar, and integrated existing geochronological and structural data (Supplementary A). We have ground-truthed this interpretation by collecting structural data and key rock samples for U–Pb zircon, U–Pb apatite, Rb–Sr muscovite and Rb–Sr biotite analysis. These systems span a wide range of closure temperatures from which we can reconstruct the temporal and thermal evolution of this region.

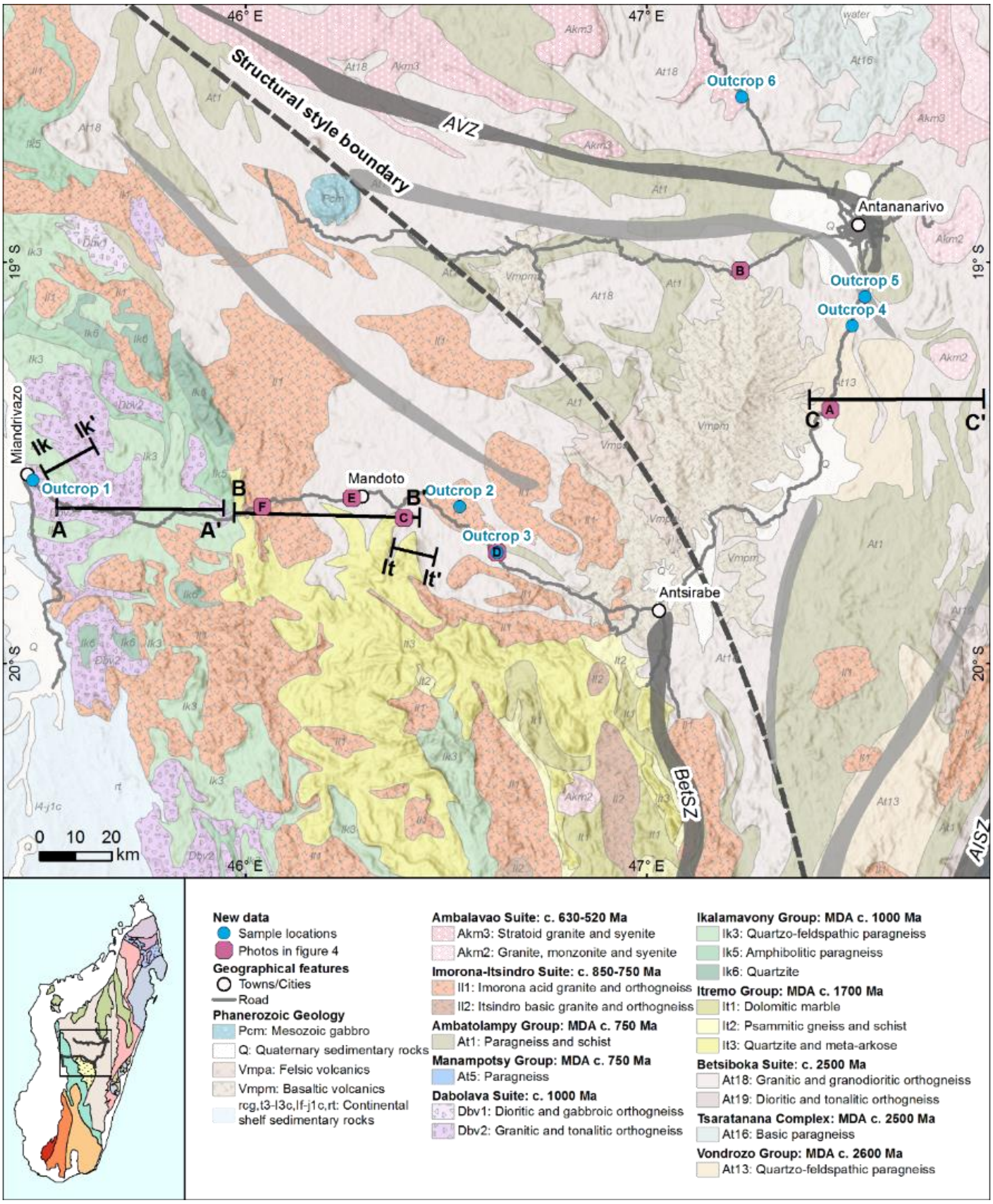

Figure 2 Geological map of central Madagascar (Roig et al., 2012) with sample locations, photo locations from Figure 4. Our remote sensing interpretation, new structural measurements and structural measurements from Macey et al. (2009); Moine (1968); Service Géologique de Madagascar (1962, 1963a, 1963b), and published geochronology are provided in a detailed PDF copy of this map where layers can be turned on and off in a pdf viewer (Supplementary A).

2. Thermochronology

A range of magmatic and orthogneiss samples were collected with the aim of having a representative sample set of the major magmatic suites of central Madagascar. This is important for determining overprinting relationships of key structural events, and determining relative and absolute timing constraints on these events. We used four geochronology/thermochronology techniques: zircon U–Pb (closure temperature ~900–1000°C), apatite U–Pb (closure temperature ~350–550°C), muscovite Rb–Sr (closure temperature ~500–600°C) and biotite Rb–Sr (closure temperature ~300–400°C). Detailed methodologies for these techniques are provided in Supplementary B. Due to the abundance of samples (41 in total), detailed results for each sample and outcrop are also provided in Supplementary B. Sample descriptions, location and age data are summarised in Table 1. Isotopic data are given in Supplementary C.

2.1. Zircon U–Pb data

Thirteen magmatic samples from six key localities were analysed for U–Pb zircon geochronology. Concordia plots are shown for each locality in Supplementary B; Figure A.2 and age data is summarised in Table 1 of the manuscript. Samples were generally very discordant and difficult to interpret magmatic crystallisation ages. This is not surprising given all samples have been thermally reset during a younger event, as revealed by both the apatite U–Pb and mica Rb–Sr results. If samples contained sufficient concordant data, we used a weighted average to calculate the magmatic crystallisation age (e.g. M16-45). However, as most data were discordant we generally calculated upper intercept ages of discordia lines of those data that appeared to form a linear trend away from the concordia curve. Data points that were very discordant and did not plot along a linear trend with other data ellipses, were generally excluded from age calculations. Many of these points may represent inherited zircons that have become discordant, or zircons that have undergone a complex history of multi-stage lead-loss. However, without several analyses forming a trend, it is impossible to interpret these 'outliers' with confidence.

2.2. Apatite U–Pb data

Apatite has a U–Pb closure temperature of ~450–550°C (Chamberlain and Bowring, 2001; Schoene and Bowring, 2007), which makes it a potentially useful system for understanding the thermal evolution of orogenic events. Coupled with other minerals with different closure temperatures, we can reconstruct the thermal and tectonic evolution of central Madagascar.

Of the thirteen samples separated for heavy mineral analysis, ten yielded sufficient apatite for U–Pb analysis. Supplementary B; Figure A.3 shows Concordia plots for all samples analysed, and grouped by locality. Discordia lines were calculated for each sample. Some analyses in each sample plotted significantly off the discordia lines, and were excluded from the calculated intercept ages (higher transparency ellipses in Supplementary B; Figure A.3). Some of these analyses appear to form a distinct trend, and so we calculated an intercept age based on these analyses for each locality, as indicated by the red discordia lines and red outlined ellipses in Supplementary B; Figure A.3.

2.3. Biotite and muscovite Rb–Sr data

Muscovite and biotite have closure temperatures of ~500–600°C (Armstrong et al., 1966) and ~300–400°C (Del Moro et al., 1982; Jenkin et al., 2001; Verschure et al., 1980) respectively, which makes them useful for understanding medium-temperature geological events. Supplementary B; Figure A.4 and Figure A.5 show isochron plots for all samples that produced reasonable age calculations, and are grouped by locality. Isochron lines were calculated for each sample. Some analyses in each sample plotted significantly off the isochron lines, and were excluded from the calculated intercept ages (higher transparency ellipses in Supplementary B; Figure A.4 and Figure A.5).

Table 1 Summary of sample descriptions, outcrop and cross-cutting relationships, and age data. Letters given for each outcrop are the interpreted order of formation/intrusion, based on cross-cutting relationships. All zircon ages are interpreted as magmatic crystallisation ages except for metamorphic ages indicated by (*) and lower intercept ages indicated by (#).

Sample ID	Transect	Outcrop	Sample description	Magmatic Suite	Latitude	Longitude	Elevation (m)	Zircon U–Pb age (Ma)	Apatite U–Pb age (Ma)	Muscovite Rb–Sr age (Ma)	Biotite Rb–Sr age (Ma)
M16-24	West	1	K-spar granite	Ambalavao	-19.5443	45.47028	182	576 ± 24	–	519 ± 69	505 ± 59
M16-32	West	2/A	Coarse-grained gneiss	Betsiboka	-19.6107	46.53399	989	2553 ± 24	519 ± 11	446 ± 161	502 ± 20
M16-33	West	2/D	Microgranodioritic dyke-undeformed	Imorona-Itsindro	-19.6107	46.53399	989	798 ± 24	–	–	–
M16-34	West	2/C	Thin dyke intruding M16-32	Betsiboka	-19.6107	46.53399	989	2511 ± 14	515 ± 7	–	–
M16-35	West	2/B	k-spar rich deformed dyke	Betsiboka	-19.6107	46.53399	989	2583 ± 26 2494 ± 14 (*)	502 ± 6	–	513 ± 18
M16-15	West	3/A	Orthogneiss	Betsiboka	-19.7239	46.62736	1067	2456 ± 17	492 ± 5	624 ± 152	528 ± 18
M16-16	West	3/B	Granite	Imorona-Itsindro	-19.7239	46.62736	1067	795 ± 24	498 ± 7	506 ± 82	499 ± 68
M16-17	West	3/C	Pegmatite veins, k-spar rich	Imorona-Itsindro	-19.7239	46.62736	1067	–	494 ± 7	526 ± 39	492 ± 51
M16-46	East	4/A	Orthogneiss	Betsiboka	-19.1599	47.51211	1351	2522 ± 8 543 ± 27 (#)	497 ± 15	604 ± 211	512 ± 24
M16-47	East	4/B	Undeformed cross-cutting granite	Imorona-Itsindro	-19.1599	47.51211	1351	798 ± 48 532 ± 44 (#)	–	657 ± 98	–
M16-45	East	5	Fine-grained granite	Ambalavao	-19.0869	47.54429	1312	543 ± 18	507 ± 35	–	512 ± 16
M16-52	East	6/A	Grey granite, very weakly foliated	Ambalavao	-18.589	47.23721	1359	568 ± 16	484 ± 14	527 ± 51	511 ± 16
M16-53	East	6/B	Cross-cutting pink granite	Ambalavao	-18.589	47.23721	1359	c. 568	500 ± 10	537 ± 35	521 ± 18

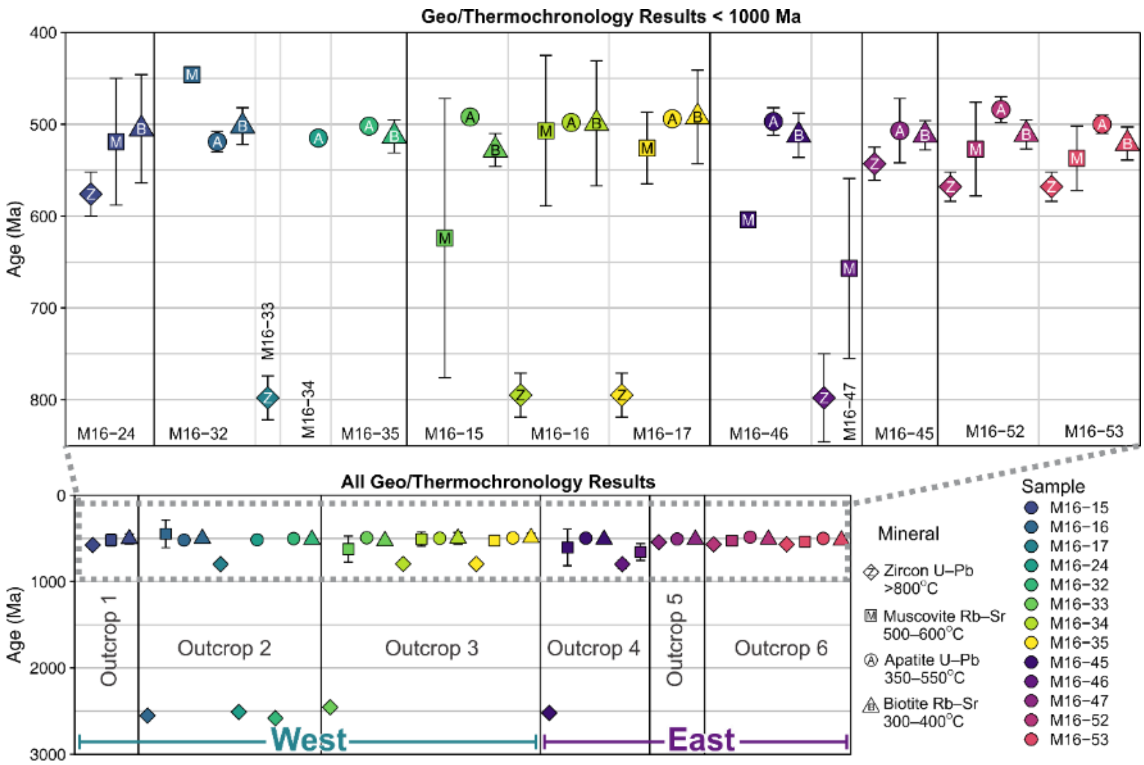

Figure 3 Summary of geo/thermochronology data for each sample and outcrop collected in this study. Error bars are 2σ. Sample locations shown in Figure 2. The bottom figure shows all of the sample data collected, and the top image zooms in on all data that are younger than 1000 Ma.

3. Structural geology methods

3.1. Remote sensing methods

We used high resolution aerial imagery and Landsat 8 data to define the structural framework for the study area (Supplementary D). Structural trends and lithological boundaries were delineated from the ESRI world imagery basemap and Landsat 8 data in ArcGIS. Examples of Landsat images and bands used as well as our structural interpretation are documented in Supplementary D. Structures in the Itremo Group are easily identifiable due to relatively low vegetation cover and a strong contrast between quartzites and other rock types. Faults were defined by small offsets in lithologies or as large linear features. Lithologies were identified by similar signals in Landsat data and aerial imagery and boundaries were determined accordingly. Following the identification of major rock packages, lithological trends (S1, S2 etc.) and faults, we were able to identify fold interference patterns and interpret the major deformation events responsible for producing these poly-deformed folds.

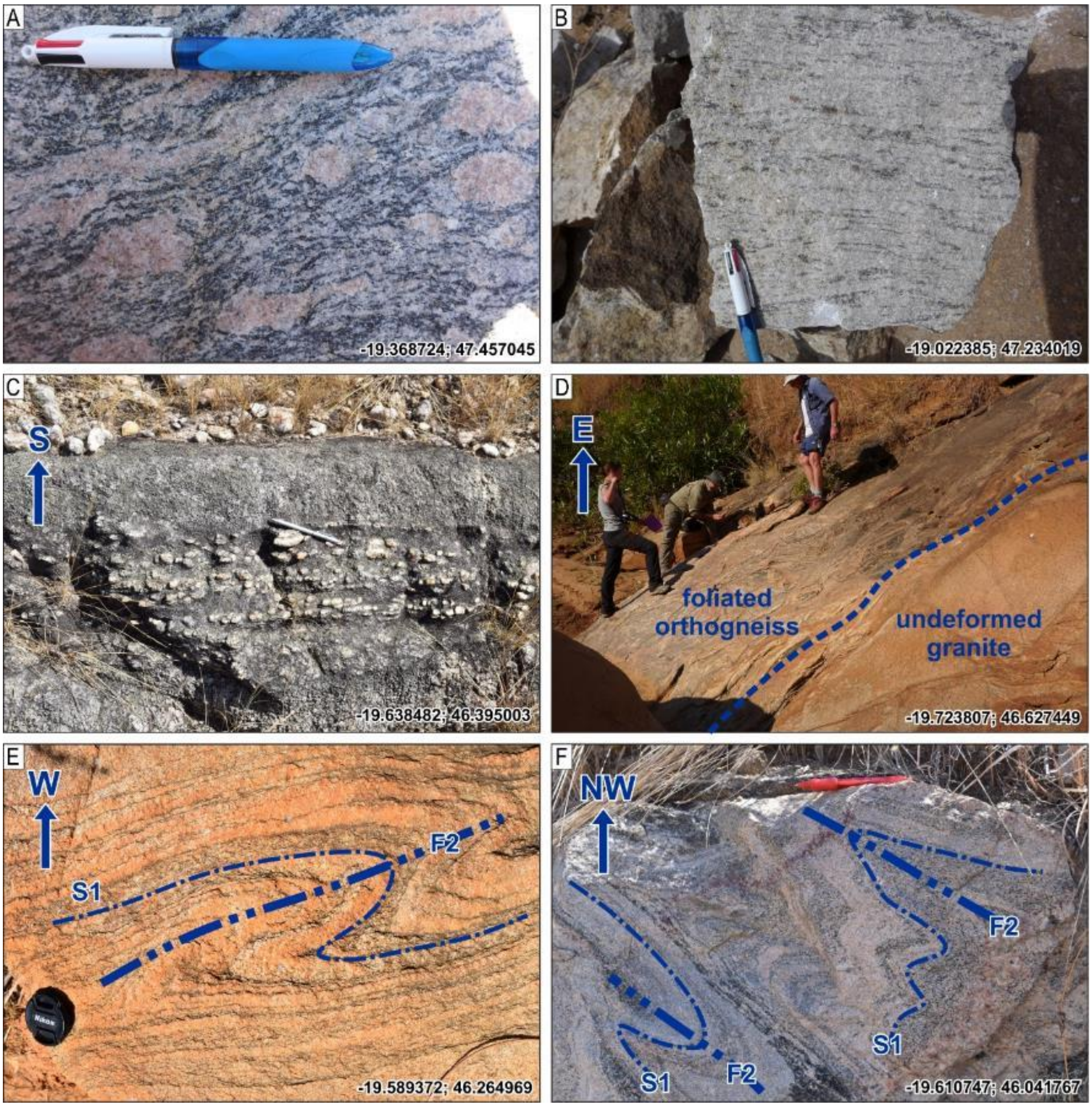

Figure 4 Examples of samples and field outcrops; latitude and longitude given in lower right corner and shown in Figure 2; a) S1 foliation in a sample of augen gneiss of the Betsiboka Suite in the Antananarivo Domain, b) S1 foliation in a sample of gneiss from the Antananarivo Domain, west of Antananarivo, c) flattened conglomerate where S1 is parallel to S0 in the Itremo Group of the Itremo Domain, d) outcrop of foliated gneiss (left) intruded by undeformed granite (right), e) S1 foliation folded around an F2 fold in the Itremo Domain, and f) S1 foliation folding around F2 folds in the Itremo Domain. Locations of field photos shown in Figure 2.

3.2. Field methods

Several hundred structural measurements were taken from over 70 localities. Data collected by the Council for Geoscience during the World Bank project in Madagascar were also used, which contains measurements for bedding and foliation (Macey et al., 2009). We additionally georeferenced geological maps (Moine, 1968; Service Géologique de Madagascar, 1962, 1963a, b) and extracted structural readings. Based on broad lithological and structural styles across the region, we have divided the study area into three sections. The Ikalamavony transect was conducted along the ~east–west road between Miandrivazo and the boundary of the Ikalamavony Domain (Figure 5), the Itremo section was conducted along the same road from the Ikalamavony-Itremo boundary toward the east approximately 50 km. The strike of S1 fabrics interpreted from remotely sensed data very closely match those measured at outcrops, we can therefore be confident that our interpretation of S1 structures from remotely sensed data is reliable.

4. Structure of central Madagascar

Large-scale structures, fold interference patterns, faults and shear zones are recognisable in remotely sensed data in the region west of Antsirabe (Figure 2). East of Antsirabe, polydeformed folds are no longer observed and the structural style changes significantly. We have delineated this as a 'structural style boundary' in Figure 2. The Itremo Nappes in the Itremo Domain have been investigated extensively due to their prominence in remotely sensed data and availability of outcrops (Collins et al., 2003b; Tucker et al., 2007). Here we further extend these interpretations to the Ikalamavony Domain, where identification of lithostratigraphy from remotely sensed data is more difficult, and interpretation is less straightforward. We have integrated our new interpretation from remotely sensed data with new 1:100,000 mapping and available structural data (Macey et al., 2009), to interpret the deformation history of the Ikalamavony, Itremo and Antananarivo (sub)domains.

We have constructed several cross-sections of the three central Madagascan domains and key type 2 fold interference patterns (Figure 5, Figure 6, Figure 7, Figure 8). We used the QGIS qProf plugin to construct cross-sections. Structural data within ~2 km of the section were included and projected onto the profile. The Africa Digital Elevation Model (30 m resolution) was used to construct the topographic profile.

4.1. Ikalamavony Domain

The Ikalamavony Domain contains metasedimentary rocks of the Ikalamavony Group, which are dominated by paragneiss, schist, quartzite and amphibolite. Generally these are finer grained than the Itremo Group. We observe many of the gneisses with bands of mylonite, indicating a high strain environment. Unique to the Ikalamavony Domain is the Dabolava Suite, which is composed of granitic to gabbroic orthogneiss (Archibald et al., 2017a). The Dabolava Suite and the age-equivalent Ikalamavony Group have been interpreted as an oceanic arc terrane (Archibald et al., 2017a). The terrane must have accreted prior to the intrusion of the c. 850–750 Ma Imorona-Itsindro Suite, which intrudes the Ikalamavony, Itremo and Antananarivo domains—placing a minimum age on the juxtaposition of the three central Madagascan domains.

4.1.1. D1 Deformation

The first recognisable deformation event at the outcrop scale is defined by a pervasive foliation observed in orthogneisses, paragneisses and metasedimentary rocks. In orthogneisses and paragneisses, the foliation is typically defined by the elongation and alignment of biotite, feldspar and quartz. In metasedimentary rocks such as schist and paragneiss, the foliation is commonly defined by the orientation of biotite crystals and biotite rich layers. Primary sedimentary features such as bedding were difficult to recognise due to significant metamorphism and recrystallisation.

In remotely sensed data, linear or curvilinear trends such as ridges, are interpreted as being representative of the S1 foliation. Quartzite units in particular, which are less common than in the Itremo Domain, are easy to recognise in remotely sensed data due to the large contrast in different Landsat bands (e.g. Figure 7). In the Ikalamavony Domain the orientation of measured S1 foliations is dominantly northwest trending, and lineations and fold axes plunge moderately toward the west.

4.1.2. D2 Deformation

D2 deformation is most easily identifiable from remotely sensed data due to the large scale (>1 km) wavelength of folds. F2 antiforms and synforms are identifiable by the repetition of mapped geological units and constrained by structural measurements. D2 is defined by tight to isoclinal folds with axial traces approximately parallel to S1 in fold limbs. At the outcrop scale we observe these as decimetre- to metre-scale asymmetric, tight to isoclinal folds. F2 folds are similar-type folds, with thickened hinge zones and thinned limbs. An axial planar foliation is difficult to recognise in outcrops, but sometimes occurs as the alignment of biotite in hinge zones. Due to the isoclinal nature of folding, F2 axial traces are approximately parallel to S1 at the regional scale. F2 folds have been subsequently deformed during D3 and D4, however their original orientation would have been preserved with ~east-west striking axial traces.

4.1.3. D3 Deformation and associated fold interference patterns

We do not observe evidence for a third generation deformation event at the outcrop scale, however D3 folds are recognisable in remotely sensed data. The folding of F2 folds during D3 has produced a series of fold interference patterns similar to those in the Itremo Domain. Type 1 and type 2 fold interference patterns are observed in remotely sensed data (Figure 7). Type 1 folds occur when an upright folding event is overprinted by an orthogonal upright folding event (Grasemann et al., 2004). Type 2 fold interference patterns occur when a recumbent folding event is orthogonally overprinted by an upright folding event (Grasemann et al., 2004) and is expressed in Figure 7. We therefore interpret D3 as the result of ~northeast-southwest shortening (present day orientation). Cross-section Ik–Ik’ (Figure 7) has F2 folds that are very tight to isoclinal, with axial traces approximately parallel to F3 in F3 fold limbs. This formed by ~southeast-directed recumbent folding that was overprinted by a north to northwest trending F3 fold.

4.1.4. D4 Deformation

The axial traces of F3 folds vary across the Ikalamavony Domain, indicating a fourth generation of deformation. For example the F3 fold axes vary from northwest-trending in the west near Miandrivazo (e.g. the cross-section in Figure 7c), and curve to become north- to northeast-trending in the eastern side of Figure 7. We suggest this is caused by large wavelength (~30–50 km), F4 open folding with approximately east-west shortening. This event may have occurred in the late stages of folding and thrusting of the Itremo sub-domain and Ikalamavony Domain, or may be related to far-field deformation associated with orogenesis in eastern Madagascar (Collins et al., 2003a).

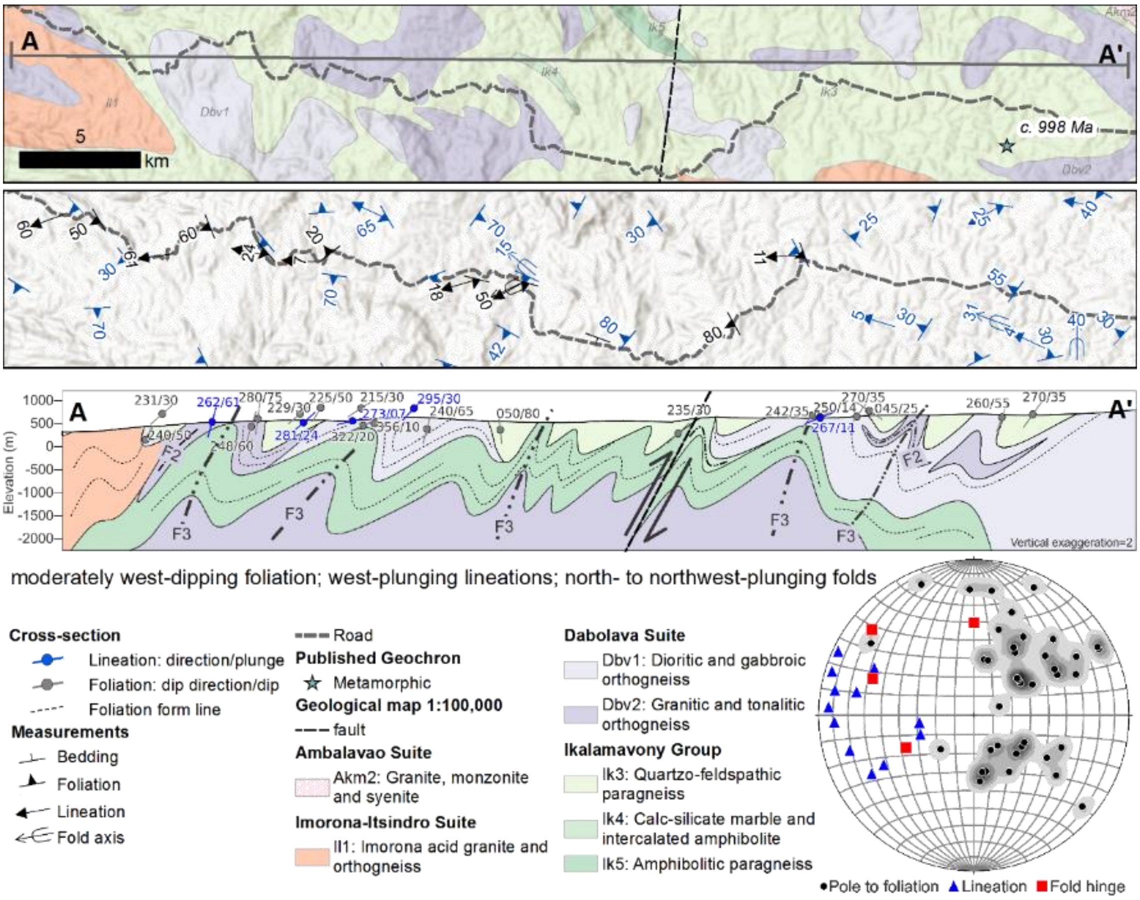

Figure 5 Geological map, structural data and cross-section through the Ikalamavony Domain. In this transect foliations are moderately west-dipping; lineations are west-plunging, and folds plunge toward north to northwest.

4.2. Itremo-Antananarivo Domain

The Itremo Group is a continental marginal sequence deposited on basement rocks of the Antananarivo Domain (e.g. Cox et al., 1998; 2004). Therefore, we consider these 'domains' together. Transect B–B' (Figure 6) contains metasedimentary rocks of the Itremo Group, which are dominantly quartzites, marbles and schists, with minor conglomerates. The majority of quartzites that we observe are strongly recrystallised and it is often difficult to recognise primary sedimentary features. The Itremo Group overlies the Archean Betsiboka Suite and was intruded by the c. 850–750 Ma Imorona-Itsindro Suite, after early deformation (Collins et al., 2003b). Together, these suites of rocks underwent a complex deformation history that must post-date the intrusion of the Imorona-Itsindro Suite.

4.2.1. D1 Deformation

In remotely sensed data, D1 structures are prominent features. Quartzite units in particular are easy to recognise in remotely sensed data due to the large contrast in different Landsat bands (e.g. Figure 7). The orientation of S1 is variable in the Itremo sub-domain due to the abundance of poly-deformed folds. Similar to the Ikalamavony Domain, foliations trend dominantly north-northwest, with lineations and fold axes plunging moderately toward the west.

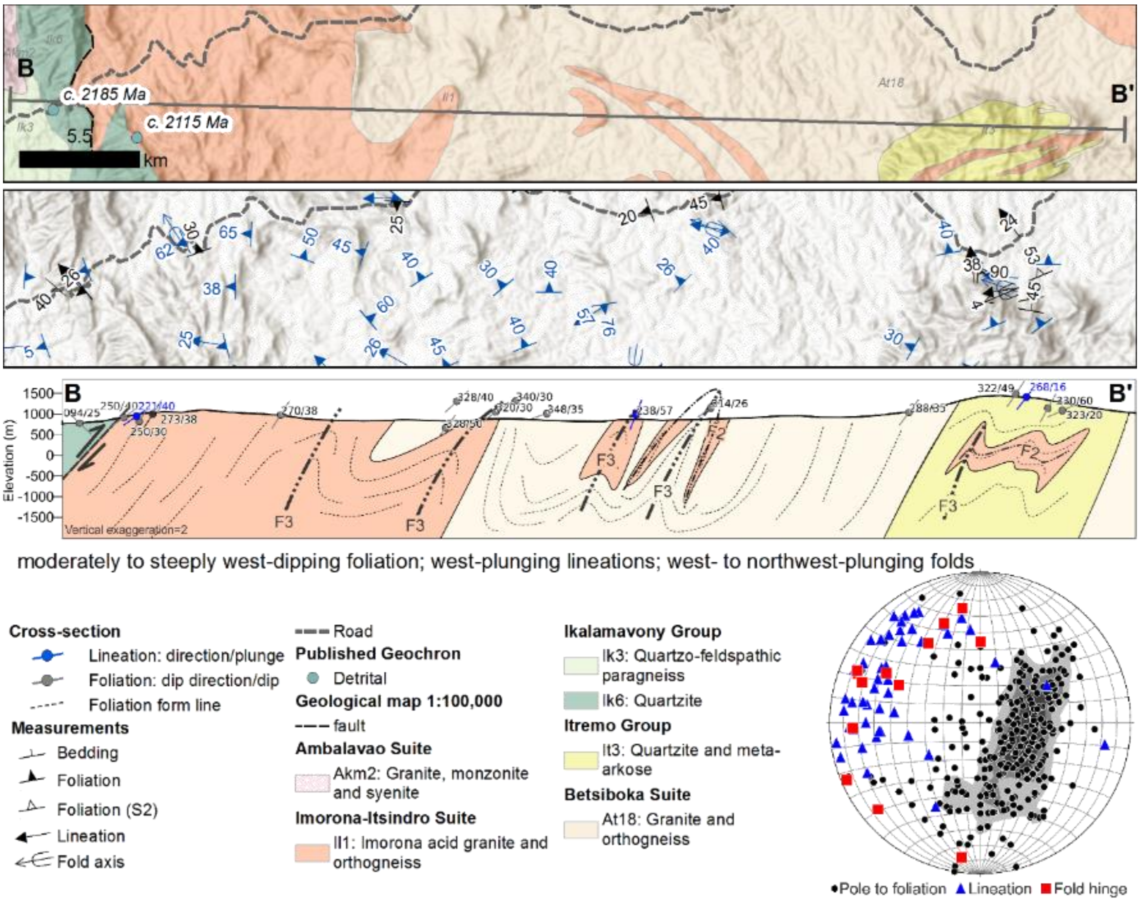

Figure 6 Geological map, structural data and cross-section through the Itremo-Antananarivo domain. Similar to the Ikalamavony transect, the Itremo transect contains moderately to steeply west-dipping foliations, west-plunging lineations and west to northwest-plunging folds.

Like the Ikalamavony Domain, the first generation foliation at the outcrop scale is typically defined by the elongation and alignment of biotite, feldspar and quartz in orthogneisses and paragneisses. In metasedimentary rocks such as quartzites and marbles, the foliation is sometimes defined by the orientation of biotite crystals and biotite rich layers, but is often difficult to recognise due to significant recrystallisation of quartz and a lack of other minerals. Primary sedimentary features such as bedding were difficult to recognise in quartzites due to significant recrystallisation. Within the quartzite packages, there are several conglomerate units with large (up to ~5 cm) pebbles (Figure 4c). Here we observe S0 as the interbedded pebble layers, and S1 as the flattening of pebbles.

4.2.2. D2 Deformation

Identical to the Ikalamavony Domain, D2 is defined by tight to isoclinal folds with axial traces approximately parallel to S1 in fold limbs. At the outcrop scale we observe these as decimetre- to metre-scale asymmetric, tight to isoclinal folds (Figure 4e,f). F2 folds are similar-type folds, with thickened hinge zones and thinned limbs. F2 axial traces are approximately parallel to S1 at the regional scale. F2 folds are recognisable in remotely sensed data as ~500–1000 m wavelength, tight to isoclinal folds.

F2 folds have been subsequently deformed during D3 and D4, however their original orientation would have been preserved with ~east-west striking axial traces. Further south where structures are more north-south trending, Tucker et al. (2007) interpreted east or south-east directed vergence from these fold trends. We suggest both the east to south-east vergence outlined in Tucker et al. (2007), as well as the ~south to southwest-directed vergence identified in this study, developed as a result of ~northeast-southwest shortening (present day orientation).

4.2.3. D3 Deformation and associated fold interference patterns

Similar to the Ikalamavony Domain, we do not observe evidence for a third generation deformation event at the outcrop scale, however D3 folds are recognisable in remotely sensed data. The majority of F3 fold axial traces are ~north-south striking, and orthogonally overprint F2 folds. We therefore interpret D3 as a ~east-west shortening event. The folding of F2 folds during D3 has produced a series of fold interference patterns. Type 2 fold interference patterns are observed in remotely sensed data in the Itremo Domain (Figure 7). Type 2 fold interference patterns occur when a recumbent folding event is orthogonally overprinted by an upright folding event (Grasemann et al., 2004) and is expressed in both examples in Figure 7.

Cross-section It–It' in Figure 7 represents a type 2 fold interference pattern. Armistead et al. (2018) modelled a structure of very similar geometry and orientation, and showed that this type of feature formed from south-directed, tight, recumbent folding that was orthogonally overprinted by third generation upright folding. In our example from the Itremo Group, the F2 recumbent folding formed during south to slightly south-west directed folding that locally formed by ~north-south shortening. The overprinting F3 upright fold formed during ~east-west shortening that resulted in a north to north-east trending axial trace. These interpreted kinematics are consistent with previous interpretations for deformation in the Itremo Group (Collins et al., 2003b; Tucker et al., 2007).

Adding to the complexity of the structure in It–It' is the juxtaposition of older units (the Archaean Betsiboka Suite) structurally above younger units (Paleoproterozoic Itremo Group). Tucker et al. (2007) observed that the km-scale fold and thrust nappes (our interpreted D2), resulted in the inversion and repetition of Archaean and Proterozoic rocks. This interpretation accounts for why the It–It' section contains older units that appear structurally above younger units.

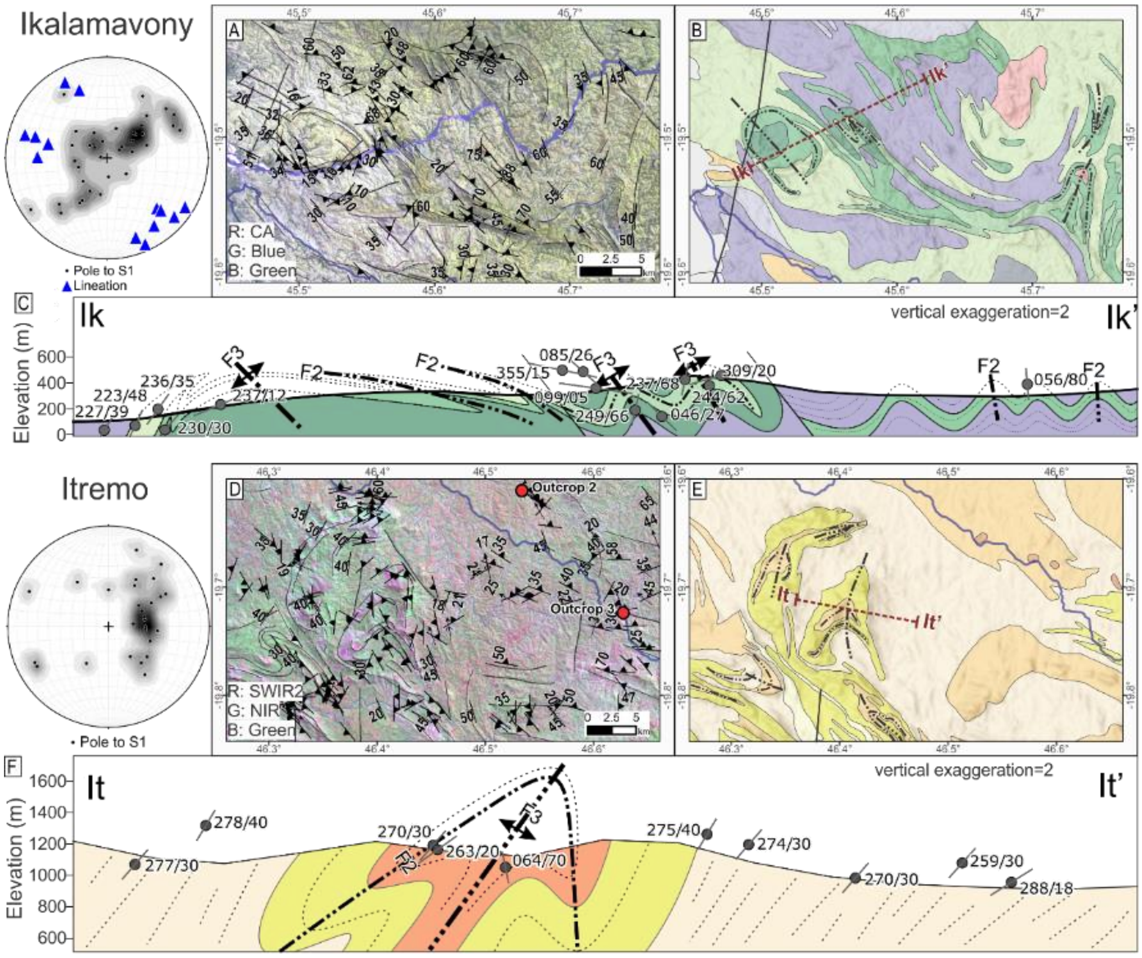

Figure 7 Examples of poly-deformed folds in the Ikalamavony and Itremo domains. Ik–Ik': an example of type 2 fold interference patterns with north-west trending third generation upright folds. It–It': an example of a type 2 fold interference pattern with D2 south-directed recumbent folding overprinted by a F3 north to north-east trending upright fold. Geological polygons from Council for Geosciences 1:100000 mapsheets (Macey et al., 2010). Sections generated using QProf plugin in QGIS. Structural measurements (dip direction/dip) within ~2km of the section are projected along the profile. Topographic profile derived from 30 arc-second DEM of Africa (USGS). Legend for geological units is the same as Figure 5 and Figure 6.

4.2.4. D4 Deformation

The axial traces of F3 folds vary across the Ikalamavony, Itremo and Antananarivo regions, as we have shown in the examples of It–It' and Ik–Ik' above. The trend of structures vary from the northwest of central Madagascar near Miandrivazo, to the southeast of the study area along the eastern margin of the Itremo Group (Figure 2). Near Miandrivazo (e.g.

Figure 5), D3 axial traces generally trend northwest-southeast. In the Itremo Group and further to the south, these structures are generally north-south trending. This trend broadly follows the curve of our transect boundary line between the western and eastern transects delineated in Figure 2. This regional variation may relate to D4 deformation or may relate to orogenic bending as orogenesis progressed.

4.3. Antananarivo Domain

Precambrian outcrop near Antsirabe is scarce due to the widespread coverage of the Ankaratra Volcanics. Generally, deformation in this area is much less intense than the Ikalamavony Domain and Itremo sub-domain, and we observe fewer deformation events. Here we only identify a primary foliation (S1), with no later deformation events recognised. Deformation becomes more intense further east, along the road west of Antananarivo where Collins et al. (2003a) have described at least four phases of deformation.

4.3.1. D1 Deformation

Much like the western transect, at the outcrop scale we observe a pervasive foliation within the Betsiboka Suite, which we interpret as an S1 foliation. The foliation is commonly preserved by the alignment of biotite, feldspar and quartz in orthogneisses. Structural measurements from Service Géologique de Madagascar (1963a) and Service Géologique de Madagascar (1963b) indicate that S1 foliations between Antsirabe and Antananarivo dominantly strike ~north-northeast, with dips moderately to the west (Figure 8). North of Antananarivo, S1 is more variable, and folded following the Antananarivo virgation zone (e.g. Nédélec et al., 2000).

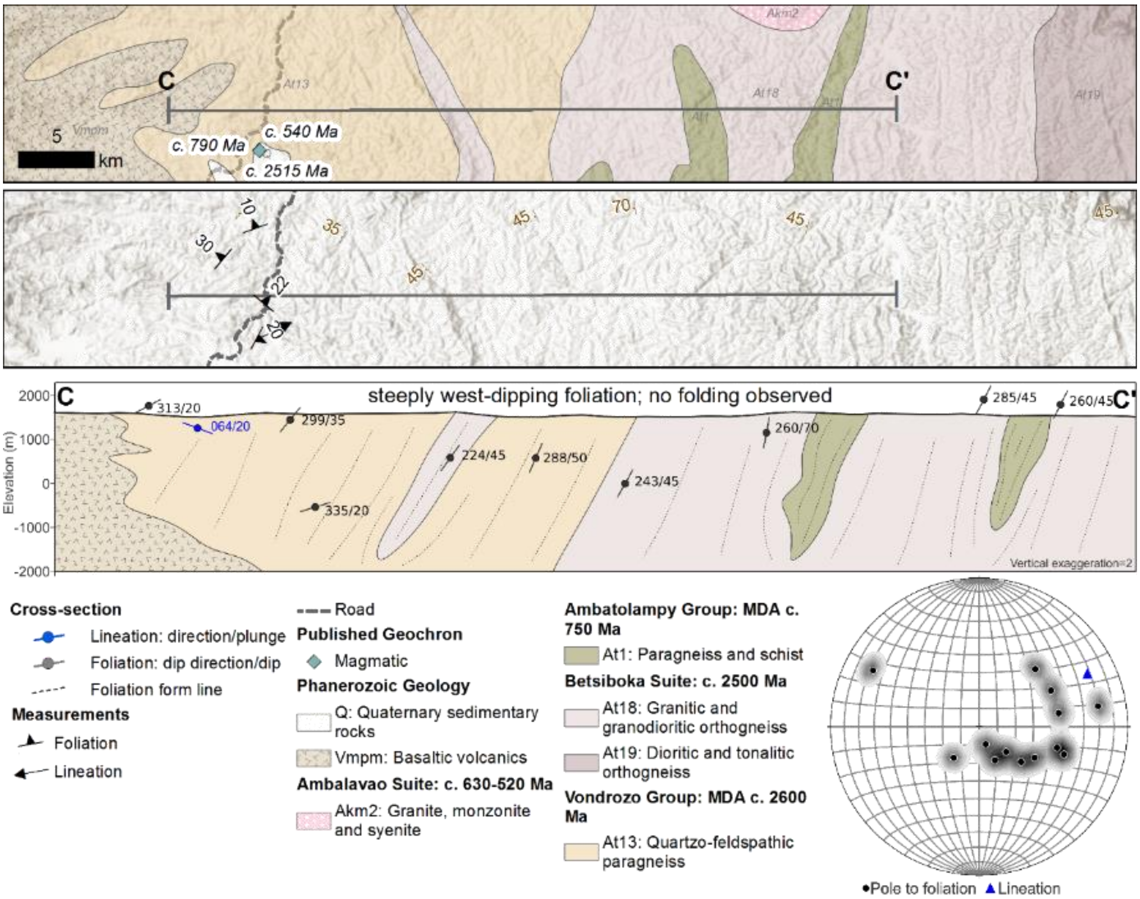

Figure 8 Geological map, structural data and cross-section through the Antananarivo Domain. This transect does not contain the complex folding that we observe in the Ikalamavony and Itremo domains. Here, foliations are steeply west-dipping.

4.4. Regional variation in deformation

Complex fold interference patterns are widespread throughout the Ikalamavony Domain and Itremo sub-domain (incorporating parts of the Betsiboka Suite). Tucker et al. (2007) proposed that complex folds in the Itremo sub-domain can be broadly considered in two groups; “high-grade, internal nappes” and “low-grade, external nappes.” These were considered to be separated by a west-dipping thrust fault, although the exact location of this boundary is ambiguous from the highly schematic diagrams presented in that study. We broadly agree that metamorphic grade and deformation intensity appears to increase toward the west, however a sharp tectonic boundary hasn’t been observed in this study within the Itremo Group. We do however see a major tectonic boundary between the Ikalamavony Domain and the Itremo/Antananarivo Domain. This boundary may more accurately reflect the boundary between the internal and external nappes proposed by Tucker et al. (2007).

Deformation intensity appears to weaken toward the east of the Itremo sub-domain, with an absence of complex fold interference patterns between Antsirabe and Antananarivo. The Imorona-Itsindro Suite in particular becomes progressively less deformed to the east. In the west, the Imorona-Itsindro Suite is folded into fold interference patterns, whereas in the east it only appears to be folded into weakly-defined F3 folds. This is consistent with our sampling of c. 850–750 Ma rocks along this weakly deformed margin (along the main road in Figure 2), where rock samples appear undeformed or very weakly deformed. Other authors have interpreted these rocks as strongly deformed further to the south (e.g. Archibald et al., 2016), where we and others have interpreted poly-deformed folds. Together with the observation that deformation and metamorphic grade increases toward the west in the Ikalamavony Domain, this implies that the tectonic event responsible for complex deformation in central Madagascar is focussed west of the study area, and its effects become less intense toward the east.

Constraining the timing of deformation within the Imorona-Itsindro Suite is particularly important due to its age (c. 850–750 Ma). Due to an intrusive contact between the Imorona-Itsindro Suite and surrounding rocks, it is difficult to determine whether the Imorona-Itsindro Suite underwent all phases of deformation, or whether it was intruded after the earliest D1 phase.

To the east of the Antananarivo–Antsirabe road (Figure 2; Figure 8), there is a distinct change in structural trend. In the Ikalamavony and Itremo Domains, structures dominantly trend northwest. North of Antananarivo, structures trend ~west (near samples M16-52 and M16-53), and between Antananarivo and Antsirabe (Figure 8) structures trend north or north-east. These structures were studied in detail in Collins et al. (2003a); Nédélec et al. (2000); Raharimahefa and Kusky (2006, 2009); Raharimahefa et al. (2013). The intensity of these structures increases toward the east, with at least four phases of deformation recognised resulting from the Betsimisaraka Suture in eastern Madagascar.

5. Discussion

5.1. Structural evolution of central Madagascar

The structural style of the Ikalamavony and Itremo domains are indistinguishable, and we suggest that they were deformed together in the same orogenic system. Type 1 and Type 2 fold interference patterns are common in fold-and-thrust belts, and more commonly form during progressive deformation rather than discrete deformation events. A myriad of complex processes ranging from rheological contrasts to progressive rotation during deformation, can cause fold structures with trends that are oblique to the transport direction of the overall fold-and-thrust belt (e.g. Poblet and Lisle, 2011). Therefore, we suggest that D2 and D3 in the Ikalamavony Domain and Itremo sub-domain formed during the same orogenic event, through progressive deformation, consistent with the interpretation based on metamorphism in this region by Moine et al. (2014).

As pointed out in Tucker et al. (2007), nappes in the southern Itremo Domain are east-verging and were likely produced within a zone of west-dipping subduction (present day direction). We interpret structures in our study areas of the Ikalamavony and Itremo Domains to be dominantly north-west trending, with northeast directed vergence. We propose that a subduction zone southwest of the Ikalamavony Domain and Itremo sub-domain was ~southwest-dipping, and was responsible for deformation in central Madagascar. Direct evidence for a suture in this region (e.g. obducted ophiolite sequences) has not been observed, however this zone separates terranes that have distinct geological histories and so a suture is a reasonable interpretation.

The boundaries between the major domains in southern Madagascar represent possible suture zones responsible for deformation in the Ikalamavony Domain and Itremo sub-domain. Boger et al. (2014); (2019) suggested that the Beraketa high strain zone that separates the Anosyen and Androyen domains represents a c. 580–520 Ma suture. The west-dipping c. 650 Ma Vohibory Suture (Figure 11), which separates the Vohibory Domain from the Androyen Domain, is also a possible suture responsible for deformation. Based on our observation that the c. 580–520 Ma Ambalavao Suite is undeformed in central Madagascar, the orogenic event must pre-date these magmatic rocks—therefore we suggest that the c. 650 Ma Vohibory Suture was responsible for complex deformation in the Ikalamavony and Itremo domains.

In the eastern part of the study area, east of the 'structural style boundary' in Figure 2, the deformation style changes in orientation and intensity. Our transect C–C' is less deformed than the Ikalamavony and Itremo transects, and we do not observe any complex folding here. The orientation of structures also change, and become more north to north-east trending. Further east of our C–C' transect is the dextral Angavo Shear Zone, which was active at c. 550 Ma (Raharimahefa and Kusky, 2010). Collins et al. (2003a) constructed a cross-section from Antananarivo eastwards to Brickaville along the east coast of Madagascar. This transect region contains a deformation sequence distinct from the Ikalamavony and Itremo domains and was therefore caused by a different tectonic event. Although controversial, there is significant metamorphic and structural evidence that the sequence of deformation described by Collins et al. (2003a) can be attributed to the c. 550 Ma Betsimisaraka Suture that amalgamated Madagascar with the Dharwar Craton of India.

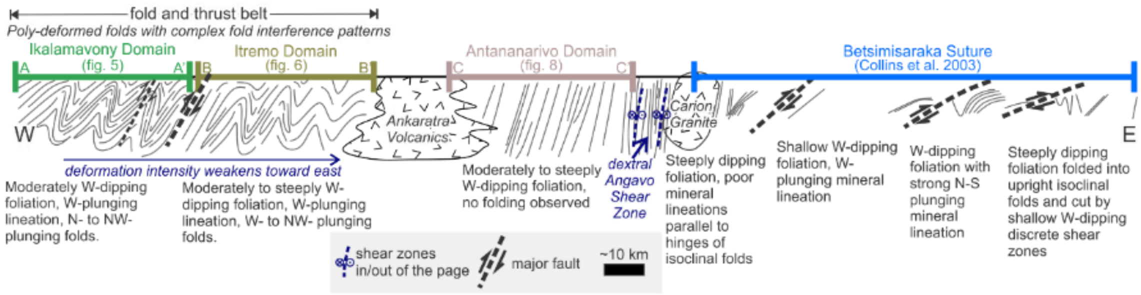

Figure 9 Schematic cross-section through central Madagascar from Miandrivazo to Brickaville, combining our new structural study with that of Collins et al. (2003a). The Ikalamavony and Itremo domains preserve the same structural styles and kinematics. A change in structural style occurs east of these sections, with the Antananarivo Domain section containing no complex fold interference patterns. Further east, Collins et al. (2003a) interpret intense deformation associated with the Neoproterozoic–Cambrian Betsimisaraka Suture.

5.2. Temporal constraints on deformation

5.2.1. Relative timing of deformation

Understanding the ages of geological units that are deformed and undeformed can help constrain the timing of deformation. At the regional scale in the western transect, the c. 850–750 Ma Imorona-Itsindro Suite is poly-deformed, and therefore was intruded prior to the onset of at least D2 and D3. In the eastern transect and in the region studied by Collins et al. (2003a); Nédélec et al. (2000); Raharimahefa and Kusky (2006, 2009), the Imorona-Itsindro Suite is not poly-deformed into complex fold interference patterns but instead is elongated along the length of the c. 550 Ma Angavo shear zone (Figure 2). This suggests that different structural regimes are responsible for deformation in the west and east of Madagascar.

In the Ikalamavony Domain and Itremo sub-domain, the c. 550 Ma Ambalavao Suite is undeformed and, therefore, provides a minimum age constraint on deformation here. In the east, the Ambalavao Suite is represented by both deformed and undeformed rocks. We therefore suggest that deformation in the Ikalamavony Domain and Itremo sub-domain occurred between c. 750 and c. 550 Ma, which is consistent with interpretations by Tucker et al. (2007). Deformation in eastern Madagascar likely occurred later at c. 550 Ma, which is consistent with age determinations for the Angavo Shear Zone and Antananarivo virgation zone (Meert et al., 2003; Nédélec et al., 2000; Paquette and Nédélec, 1998; Raharimahefa and Kusky, 2010).

5.2.2. Thermochronology

We have used minerals that record a range of temperatures in an attempt to capture different stages of the tectonic evolution of Madagascar. Our sampling included the major magmatic suites of Madagascar, including the c. 2500 Ma Betsiboka Suite, the c. 850–750 Ma Imorona-Itsindro Suite, and the c. 550 Ma Ambalavao Suite (Figure 3). Interestingly, apatite U–Pb ages—which record the age the minerals were cooled through ~350–550°C (Chamberlain and Bowring, 2001; Schoene and Bowring, 2007)—are all c. 500 Ma regardless of their magmatic crystallisation age. Muscovite and biotite, which have Rb–Sr closure temperatures of ~500–600°C (Armstrong et al., 1966) and ~300–400°C (Del Moro et al., 1982; Jenkin et al., 2001; Verschure et al., 1980), respectively, also record ages of c. 500 Ma. This implies that the final stages of orogenesis in Madagascar, regardless of whether this was in the west or east, affected the entire central region of the island, where rocks were heated to at least ~500°C.

Multiple thermochronometers have provided insights into the medium-temperature thermo-tectonic evolution across the western and eastern part of Madagascar. As we have shown here, the more recent c. 550–500 Ma thermo-tectonic event affected the entire island such that it cooled synchronously through ~500–300°C at c. 500 Ma. The c. 550–500 Ma regional thermal perturbation would have overprinted prior events, obscuring any evidence of a pre-existing thermo-tectonic evolution. Using thermochronometers that record temperatures higher than ~600°C (e.g. monazite U–Pb) in future research may be able to provide further constraints on the timing of orogenesis, particularly in the distal regions, where temperatures during the c. 550 Ma event may not have been hot enough to cause complete reset. Without direct dating of the structures observed, we need to look further afield for evidence of subduction and collision that resulted in deformation of central Madagascar.

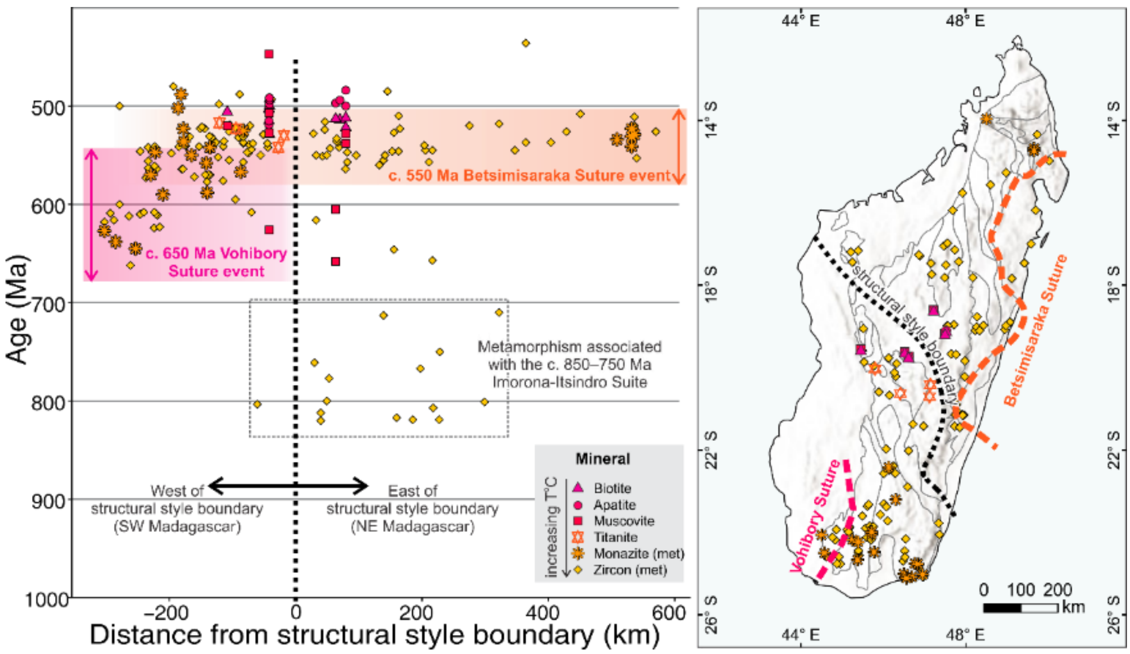

Figure 10 Summary of new thermochronology data and published metamorphic data. Biotite, apatite and muscovite are from this study. Metamorphic minerals zircon, monazite and titanite are from the compilation of Tucker et al. (2014). Locations of data points shown in the map to the right, terranes are the same as those from Figure 1b.

5.3. Tectonic model for the evolution of central Madagascar

Prior to the juxtaposition of the Itremo sub-domain and Ikalamavony Domain, we agree with previous interpretations that the Itremo Group was deposited on the Antananarivo

Domain basement (Cox et al., 1998; 2004) and that the Ikalamavony Domain evolved as an exotic island arc terrane (Archibald et al., 2017a). The presence of the Imorona-Itsindro Suite in the Ikalamavony Domain suggests that it must have accreted to central Madagascar before c. 850 Ma (Figure 11c). A large west-dipping thrust fault separating the Ikalamavony Domain from the Itremo sub-domain (Figure 6), possibly represents this suture zone (schematic thrust in Figure 11c). This implies west-dipping subduction, which is consistent with previous models for the accretion of the Ikalamavony Domain to central Madagascar (e.g. Boger et al., 2019).

Based on the interpreted kinematics and overprinting relationships, deformation of the Ikalamavony Domain and Itremo sub-domain was the result of west-dipping subduction. Increasing deformation intensity in the Ikalamavony Domain and the orientation of structures imply that this subduction zone must have lain southwest of these domains. Based on cross-cutting relationships, the timing of deformation is constrained between c. 750 and c. 550 Ma. We suggest this was in response to the Vohibory Suture (Figure 11e).

A change in deformation style and kinematics toward the east of Madagascar and younger geochronological constraints, indicate that complex folding in eastern Madagascar formed in response to a different event than that in the west (Figure 11g). Based on structural kinematics, this event was west-dipping, and was likely related to the c. 550 Ma Betsimisaraka Suture (Collins et al., 2003a).

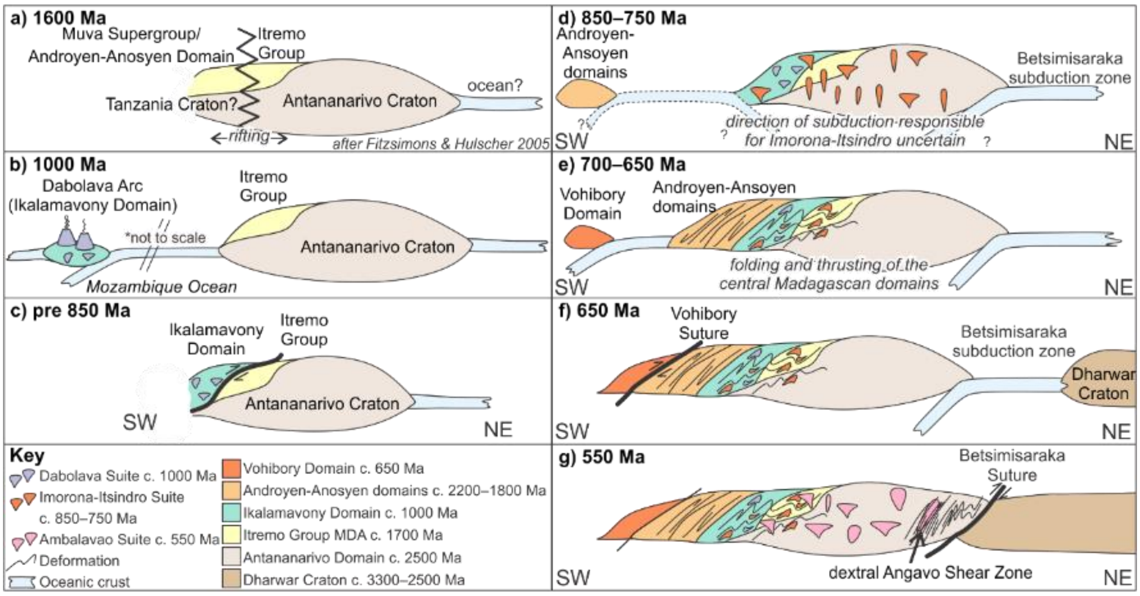

Figure 11 Schematic diagram showing our interpretation of the evolution of central Madagascar. a) Sometime after the deposition of the Itremo Group/Muva Supergroup onto the Antananarivo Craton/Tanzania Craton, these regions begin to rift (Cox et al., 2004; Fitzsimons and Hulscher, 2005); b) at c. 1000 Ma the Dabolava Arc forms as an oceanic island arc outboard from the Antananarivo Domain (Archibald et al., 2017a); c) prior to the intrusion of the c. 850–750 Ma Imorona-Itsindro Suite, the Ikalamavony Domain is thrust over the Antananarivo/Itremo Domain; d) the intrusion of the Imorona-Itsindro Suite resulting from Andean-type subduction, with polarity uncertain (Archibald et al., 2017b); e) west-dipping subduction beneath the Vohibory Domain, resulting in complex deformation that we have interpreted in central Madagascar; f) closure of the Mozambique ocean along the Vohibory Suture; and g) final amalgamation of central Gondwana along the c. 550 Ma Betsimisaraka Suture.

6. Conclusions

We have integrated structural analysis with medium-temperature thermochronology to constrain the timing of orogenic events in Madagascar. We observe complex fold interference patterns in both the Ikalamavony Domain and Itremo sub-domain. We suggest these formed in response to ~southwest-dipping subduction beneath the Vohibory Domain at c. 650 Ma. North-south trending structures in eastern Madagascar formed in response to the ~west-dipping c. 550 Ma Betsimisaraka Suture, which resulted in the emplacement of widespread granitic rocks and the resetting of medium-temperature minerals throughout much of central Madagascar. We have shown the importance of using medium-temperature thermochronometers such as U–Pb apatite to date the cooling stages post orogenesis, as well as the potential for Rb–Sr mica dating to provide useful thermochronological constraints.

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