# 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 are linked to orogenesis of the East African Orogen. In this study, we integrate remote sensing, field data and thermochronology 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 an ~E–W shortening event (D3). Together these produced type 1 and type 2 fold interference patterns throughout the Itremo and Ikalamavony domains. We show that the Itremo and Ikalamavony domains were deformed together in the same orogenic system, which we interpret as the c. 630 Ma collision of Azania with Africa along the Vohibory Suture in southwestern Madagascar. In eastern Madagascar, deformation is syn- to post-550 Ma, which likely formed in response to final closure of the Mozambique Ocean along the Betsimisaraka Suture that amalgamated Madagascar with the Dharwar Craton of India. Apatite U–Pb and novel LA-QQQ-ICP-MS muscovite and biotite Rb–Sr thermochronology indicate that much of central Madagascar cooled through ~500°C at c. 500 Ma.

**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 (Figure 1a) (Collins, 2006; Collins and Windley, 2002; Tucker et al., 1999). Of particular interest and contention, is how and when the Archean 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 of the Neoproterozoic (e.g. Merdith et al., 2017).

Central Madagascar comprises the Archean Antananarivo Domain, the Proterozoic Itremo sub-domain and the Neoproterozoic Ikalamavony Domain. These terranes are bound by two postulated major sutures; the eastern Betsimisaraka Suture and the western Vohibory Suture (mapped as the Ampanihy shear zone in Figure 1b). These sutures resulted from at least two distinct orogenic events that amalgamated central Madagascar, the Dharwar Craton of India, and Africa within Gondwana (Figure

1a). 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 Archean, 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). An age of c. 750–650 Ma for this suture has alternatively been proposed (Fitzsimons and Hulscher, 2005). Several authors have proposed that the central Madagascar–Africa collision occurred at c. 650–630 Ma (Collins and Pisarevsky, 2005; Collins and Windley, 2002; Emmel et al., 2008; Horton et al., 2016; Jöns and Schenk, 2011). Other authors have suggested a c. 850–750 Ma age for a suture in western Madagascar (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 obscured the record of earlier events through high temperature resetting of key minerals used for thermochronology and metamorphism.

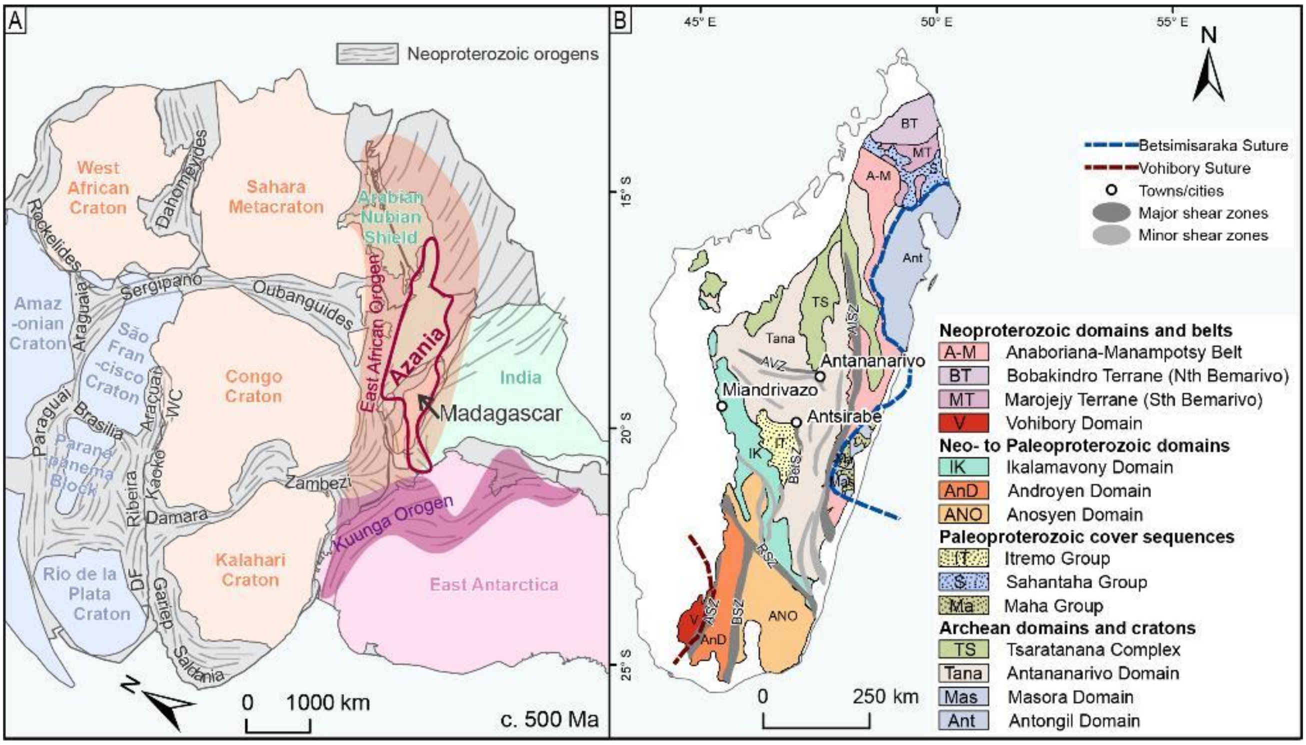


Figure 1 a) Tectonic map of Gondwana made using GPlates exported geometries from Merdith et al. (2017) in ArcGIS; projected in Hotine Oblique Mercator with Madagascar in the centre (reconstructed position, longitude=-75° and latitude=+40°). DF=Dom Feliciano Belt, WC=West Congo; b) Present day map of 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 geology

Madagascar is made up of several domains ranging in age from Archean 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 1b; Figure 2). 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 a maximum depositional age of c. 1600 Ma (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). 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).

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). Unique to the Ikalamavony Domain is the c. 1000 Ma 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). This 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. The relationship between the Ikalamavony Domain and the Itremo Group remains poorly understood, and is the focus of this study.

To the south of these metasedimentary terranes are the Proterozoic Anosyen, Androyen and Vohibory domains (Boger et al., 2014; de Wit et al., 2001; Emmel et al., 2008; Horton et al., 2016; Jöns and Schenk, 2008). In northern Madagascar is the c. 800–700 Ma Bemarivo Domain, which formed as an exotic juvenile arc terrane that amalgamated with Madagascar at c. 520 Ma, possibly in relation to the Betsimisaraka Suture (Armistead et al., 2019; Jöns et al., 2009; Thomas et al., 2009).

## 1.2. 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). The structural relationships between these domains has not yet been studied in detail. We use structural geology and various geochronological methods 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; Figure 1b; 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 Archean 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 shortening event occurred within a sinistral transpressive regime and was interpreted as being associated with the Ranotsara Shear Zone (Figure 1b) in southern Madagascar (Tucker et al., 2007). Although these models are similar in their sequence and style of deformation, they 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 (approximately the location of the Betsimisaraka Suture in Figure 1b) was studied from a structural perspective in Collins et al. (2003a); Martelat et al. (2000); Nédélec et al. (2000); Raharimahefa and Kusky (2006); Raharimahefa and Kusky (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). A syn-kinematic granite within the Angavo Shear Zone constrains deformation here to c. 550 Ma (Raharimahefa and Kusky, 2010).

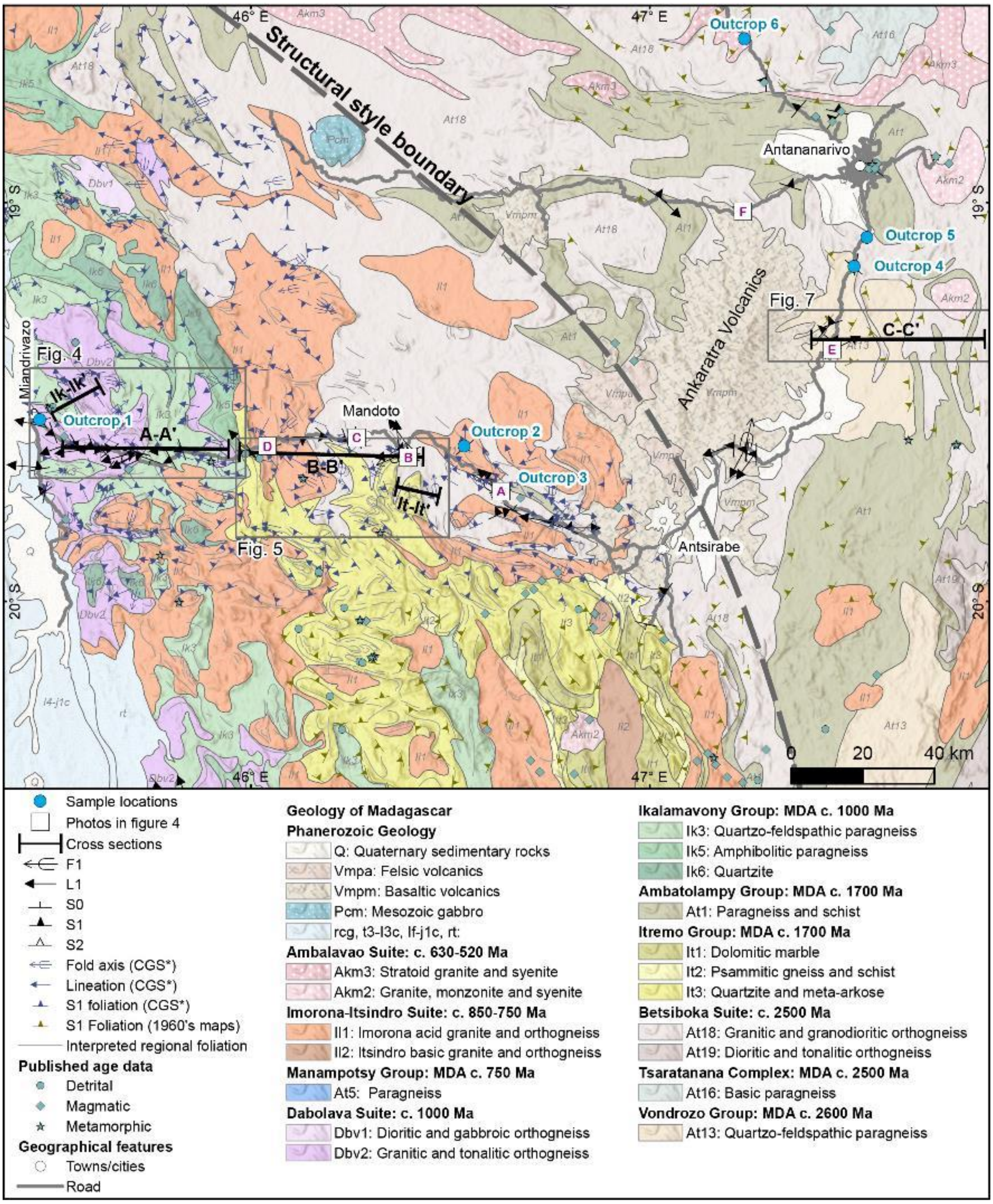


Figure 2 Geological map of central Madagascar (Roig et al., 2012) with sample locations, photo locations from Figure 6, our remote sensing interpretation, new structural measurements and structural measurements from Macey et al. (2009); Moine (1968); Service Géologique de Madagascar (1962); Service Géologique de Madagascar (1963a); Service Géologique de Madagascar (1963b), and published geochronology. Numbers associated with these points are provided in a detailed PDF copy of this map where layers can be turned on and off, and further information obtained from the model tree in a pdf viewer (Supplementary A).

Precise dating of deformation in Madagascar is difficult due to resetting from successive overlapping thermo-tectonic events. Latest metamorphism in the Anosyen and Androyen domains to the south of our study area is constrained to c. 580–520 Ma (Figure 3) (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). Horton et al. (2016); Jöns and Schenk (2011) demonstrated that in southern Madagascar, high-grade metamorphism yielded ages of c. 650–600 Ma west of the Vohibory Suture (Figure 3), but recorded ages of c. 560–530 Ma to the east of this suture. 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.

The cross-cutting relationships and deformation history of the rocks within the terranes that make up Madagascar can provide clues as to the timing of major orogenic events. Here we use structural geology to understand the deformation history of a poorly understood part of central Madagascar, which lies between the two hypothesised suture zones. 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 (Supplementary B; Supplementary C). Rb–Sr mica laser ablation QQQ-ICP-MS dating in particular is a novel technique and this research represents some of the first published ages using this technique. These isotopic systems span a wide range of closure temperatures from which we can reconstruct the temporal and thermal evolution of this region.

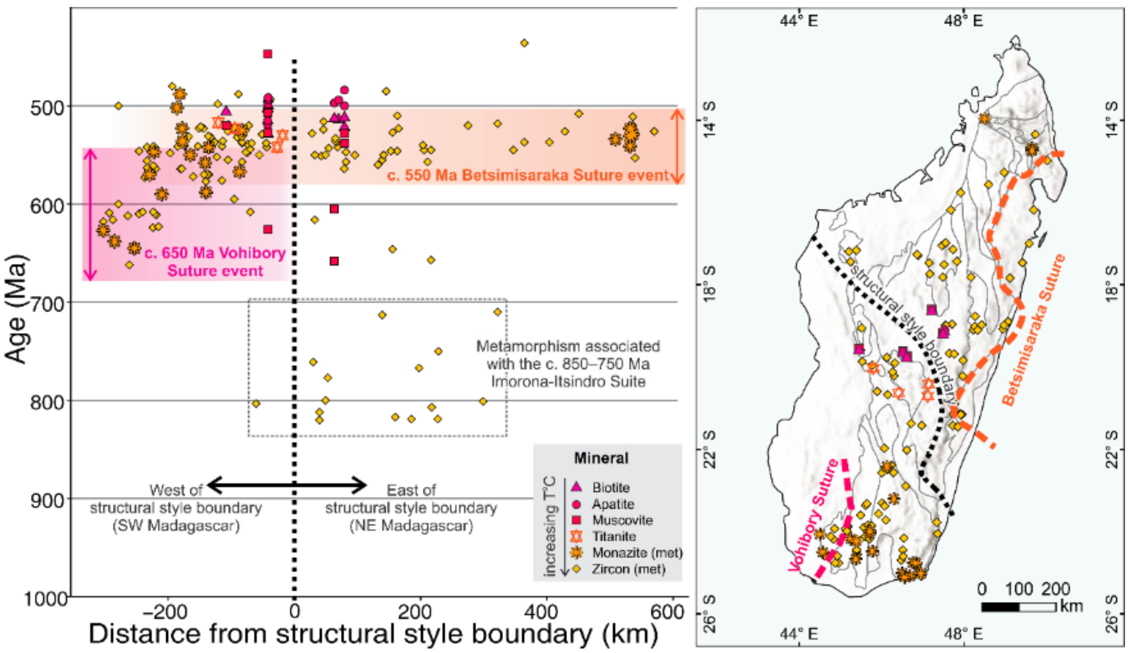


Figure 3 Summary of published metamorphic data for Madagascar and new data collected in this study. Biotite, apatite and muscovite are from this study. Metamorphic minerals zircon, monazite and titanite are from the compilation of Tucker et al. (2014). Locations of data points shown in the map to the right, terranes are the same as those from Figure 1b.

# 2. 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). Examples of Landsat images used for interpretation are presented in Supplementary D. East of Antsirabe, polydeformed folds are not observed and the structural style changes significantly. We have delineated this as a 'structural style boundary' in Figure 2. Here we further extend previous interpretations of the Itremo sub-domain (Collins et al., 2003b; Tucker et al., 2007) to the Ikalamavony Domain, where identification of lithostratigraphy from remotely sensed data is more difficult, and interpretation is less straightforward.

## 2.1. Ikalamavony Domain

The Ikalamavony Domain contains metasedimentary rocks of the Ikalamavony Group, which are dominated by paragneiss, schist, quartzite and amphibolite (Figure 4). We observe many of the gneisses with bands of mylonite, indicating a high strain environment. Based on remote sensing data, we interpret a thrust fault separating the Ikalamavony Domain and Itremo sub-domain (Figure 5). This fault is interpreted based on the sharp contrast in lithologies and the linear nature of the fault observed in remote sensing data. Due to the scarcity of fresh outcrop, we were unable to observe this fault in the field, however rocks were more strongly deformed in this area.

### *2.1.1. 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 sub-domain, are easy to recognise in remotely sensed data due to the large contrast in different Landsat bands (e.g. Supplementary D). In the Ikalamavony Domain the orientation of measured S1 foliations is dominantly northwest trending, and lineations and fold axes plunge moderately toward the west.

### *2.1.2. D2 Deformation*

D2 deformation is most easily identifiable from remotely sensed data due to the large scale (>1 km) wavelength folds. F2 antiforms and synforms are identifiable by the repetition of mapped geological units and constrained by structural measurements (Figure 4). 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 would have formed with axial traces striking ~east-west, however due to subsequent deformation during D3 and D4, they are now preserved with variable orientations.

### *2.1.3. 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. Type 1 and type 2 fold interference patterns are observed in remotely sensed data (Figure 4; Figure 5). Type 1 folds occur when an upright folding event is overprinted by an orthogonal upright folding event (Grasemann et al., 2004) and is expressed in Supplementary D(b). 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 5. We interpret D3 as the result of ~northeast-southwest shortening (present day orientation). Cross-section Ik–Ik' (Figure 4) 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.

### *2.1.4. 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. Ik-Ik' cross-section in Figure 4), and curve to become north- to northeast-trending in the centre of the map in Figure 4. We suggest this is caused by large wavelength (~30–50 km), F4 open folding with approximately east-west shortening.

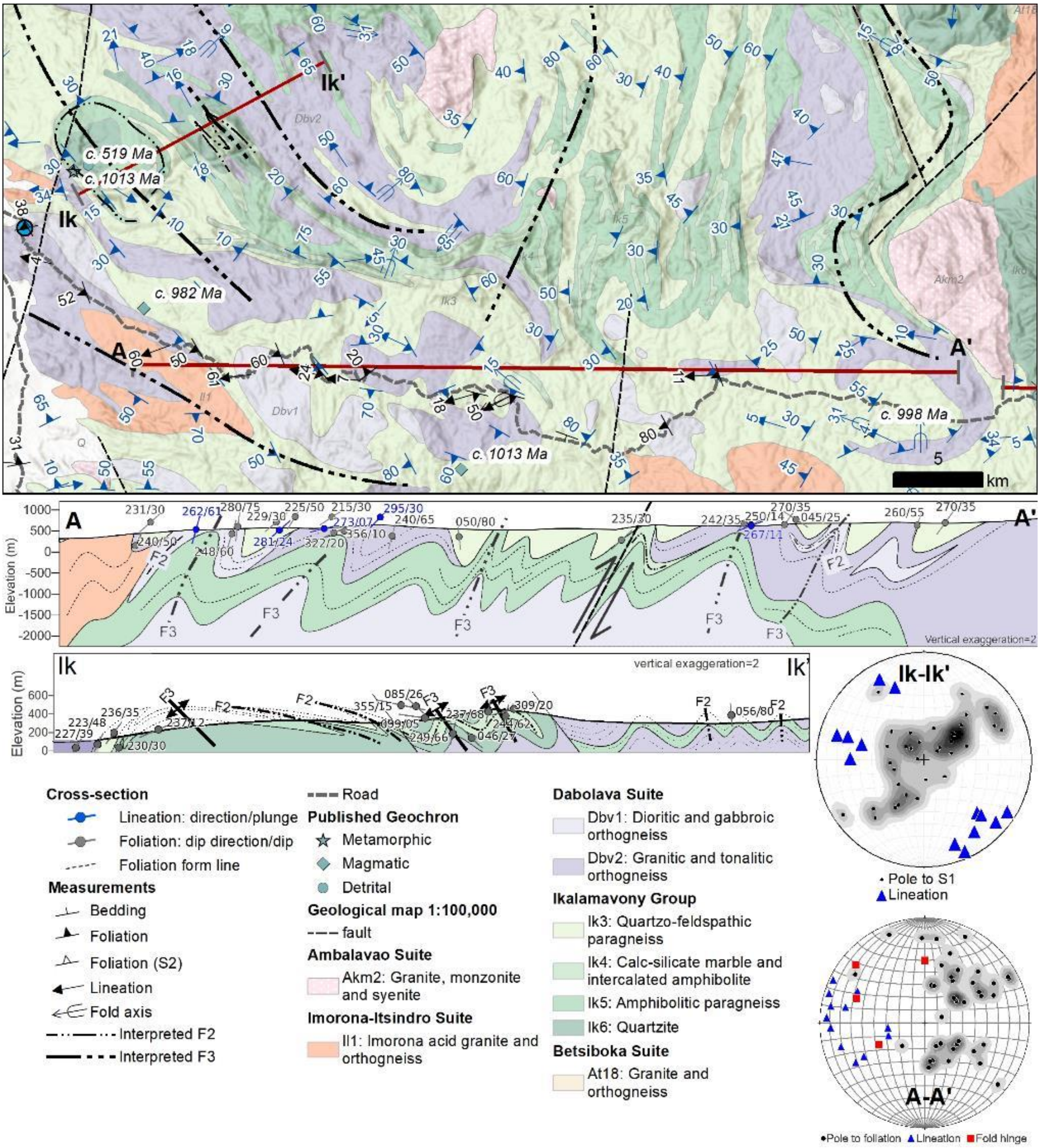


Figure 4 Geological map, structural data and cross-sections through the Ikalamavony Domain using both our new data (black) and previously published data (blue) (Macey et al., 2009). Fold axis measurements are generally interpreted to be F2 folds since we do not observe any overprinting fold generations in the field. A-A' shows the overall trend of structures in the Ikalamavony Domain. Ik–Ik': an example of type 2 fold interference patterns with north-west trending third generation upright folds. 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). Geological polygons from Council for Geosciences 1:100000 mapsheets (Macey et al., 2010).

## 2.2. 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 5) 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-Antananarivo Domain 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.

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 (documented in Table 1).

### *2.2.1. D1 Deformation*

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 strike dominantly north-northwest, with lineations and fold axes plunging moderately toward west-northwest. 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 (Figure 6a). 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 6b). Here we observe S0 as the interbedded pebble layers, and S1 as the flattening of pebbles.

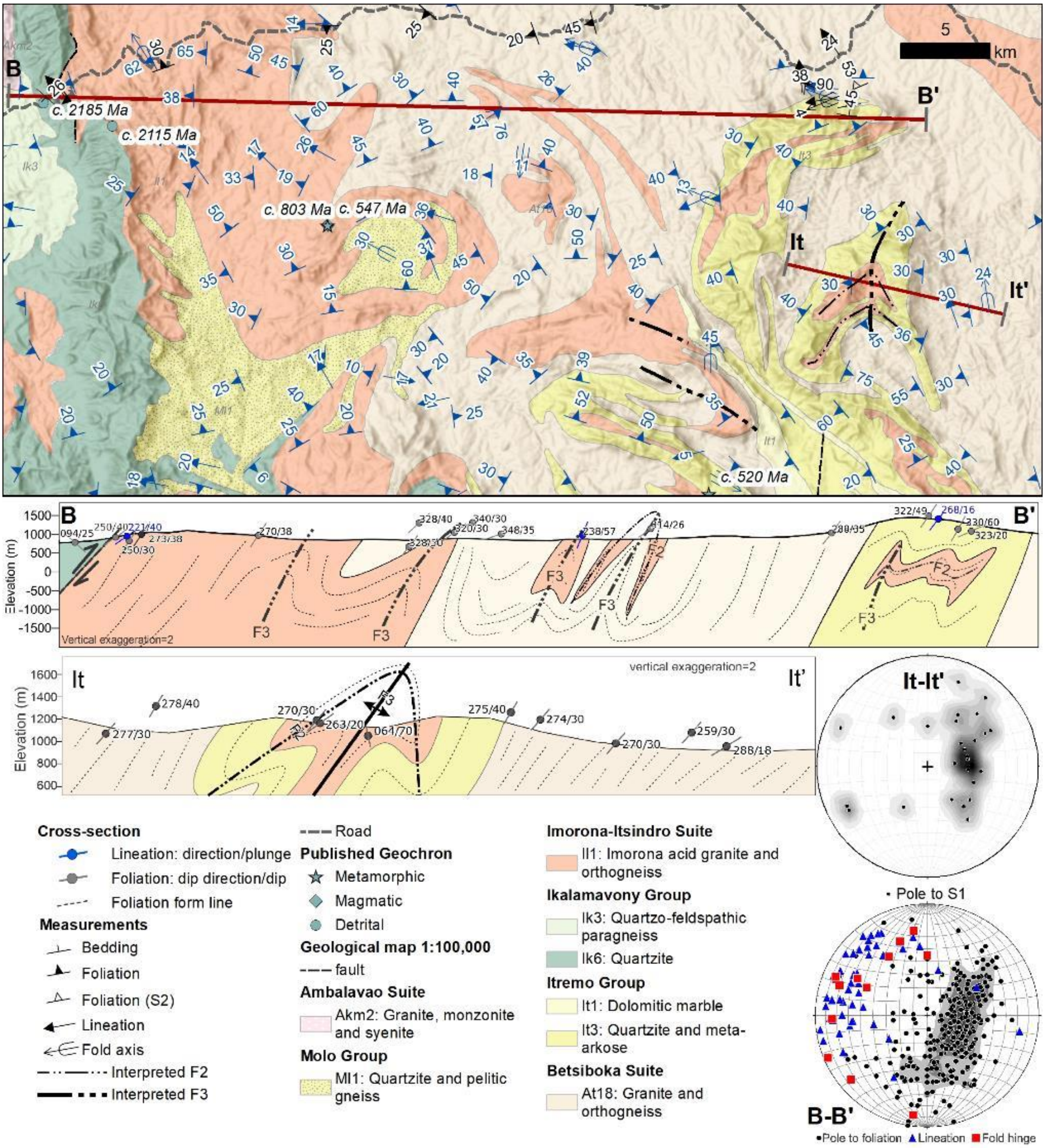


Figure 5 Geological map, structural data and cross-section through the Itremo-Antananarivo Domain using both our new data and previously published data (Macey et al., 2009). Measured fold axes are generally interpreted to be F2 folds since we do not observe any overprinting fold generations in the field. Similar to the Ikalamavony transect, the Itremo transect contains moderately to steeply west-dipping foliations, west-plunging lineations and west to northwest-plunging 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).

#### 2.2.2. 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 6c,d). 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 (Figure 5). The original orientation of F2 folds would have had ~east-west striking axial traces, but have been subsequently deformed during D3 and D4. Further south where structures are more north-south trending, Tucker et al. (2007) interpreted east or south-east directed vergence from these fold trends.

#### 2.2.3. 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 an ~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 sub-domain (Figure 5).

Adding to the complexity of the structure in It–It' is the juxtaposition of older units (the Archean 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 Archean and Proterozoic rocks. This interpretation accounts for why the It–It' section contains older units that appear structurally above younger units.

#### 2.2.4. D4 Deformation

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 4), F3 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 structural style boundary between the western and eastern transects delineated in Figure 3. This regional variation may relate to D4 deformation or may relate to orogenic bending as orogenesis progressed.

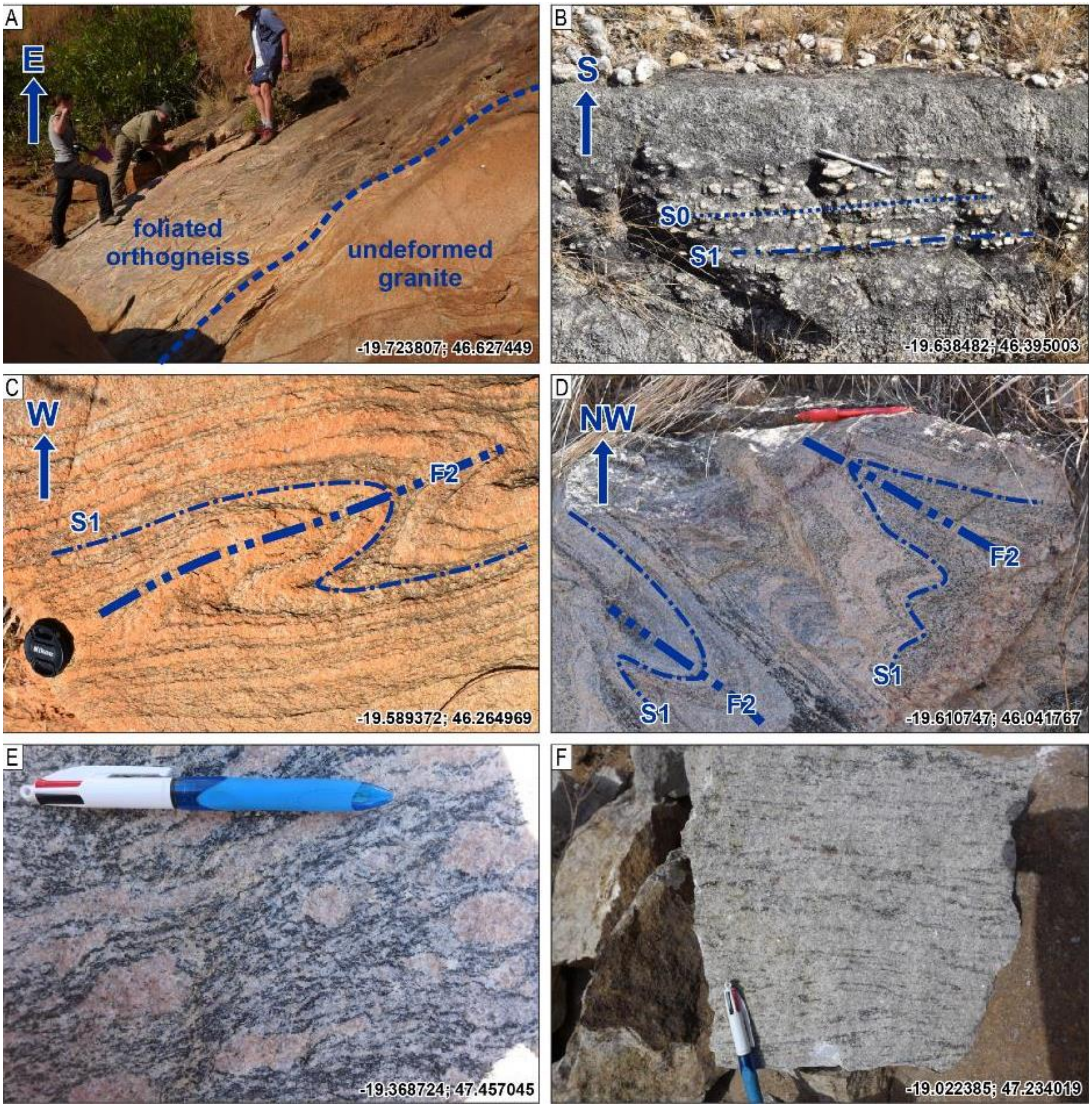


Figure 6 Examples of samples and field outcrops; latitude and longitude given in lower right corner and shown in Figure 2; , a) outcrop of foliated gneiss (left) intruded by undeformed granite (right); b) flattened conglomerate where S1 is parallel to S0 in the Itremo Group of the Itremo sub-domain; c) S1 foliation folded around an F2 fold in the Itremo sub-domain; d) S1 foliation folding around F2 folds in the Itremo sub-domain; e) S1 foliation in a sample of augen gneiss of the Betsiboka Suite in the Antananarivo Domain; and f) S1 foliation in a sample of gneiss from the Antananarivo Domain, west of Antananarivo.

## 2.3. Antananarivo Domain

Precambrian outcrops between Antsirabe and Antananarivo are scarce due to the widespread coverage of the Ankaratra Volcanics (Figure 2). Generally, deformation in this area is much less intense than the Ikalamavony Domain and Itremo sub-domain, and we observe fewer deformation events (Figure 7). To the east of the Antananarivo–Antsirabe road (Figure 2; Figure 7), there is a distinct change in structural trend. In the Ikalamavony and Itremo domains, structures dominantly trend northwest. North of Antananarivo, structures trend ~west (approximately the location of Antananarivo virgation zone in Figure 2), and between Antananarivo and Antsirabe (Figure 7) structures trend ~north. This region was studied in detail in Collins et al. (2003a); Nédélec et al. (2000); Raharimahefa and Kusky (2006); Raharimahefa and Kusky (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.

#### *2.3.1. 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 (Figure 6e,f). The foliation is commonly preserved by the alignment of biotite, feldspar and quartz in orthogneisses. Structural measurements indicate that S1 foliations between Antsirabe and Antananarivo dominantly strike ~north-northeast, with dips moderately to the west (Figure 7). North of Antananarivo, S1 is more variable, and folded following the Antananarivo virgation zone (e.g. Nédélec et al., 2000). We observe a well-defined gneissic foliation within the Archean Betsiboka Suite, which may have originally formed prior to Neoproterozoic deformation. If this is the case, then subsequent Neoproterozoic deformation was approximately the same orientation, as we do not observe cross-cutting fabrics.

#### *2.3.2. D2 Deformation*

We do not observe D2 structures at the outcrop scale in this section. However, the repetition of mapped Ambatolampy Group within the Betsiboka Suite (Figure 7), indicates that the two elongated Ambatolampy Group bodies in Figure 7 represent tight to isoclinal F2 folds.

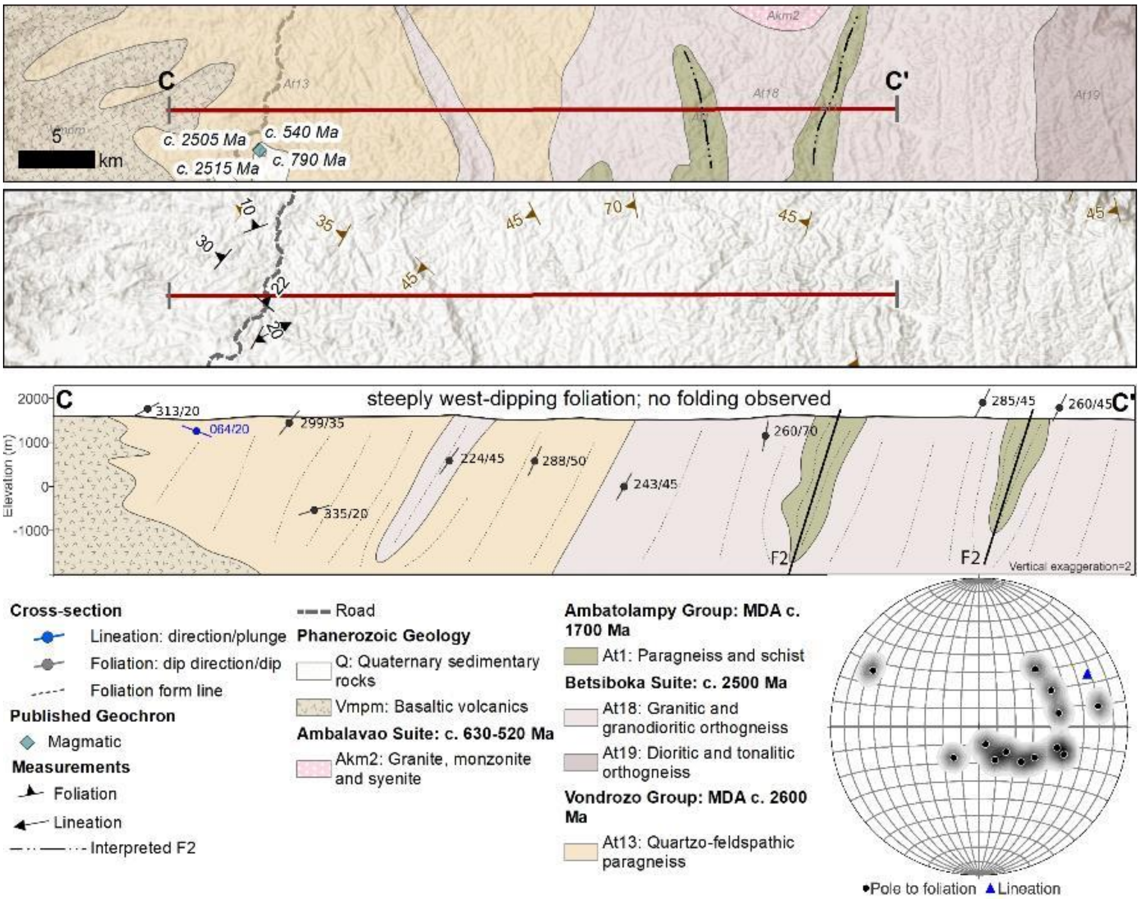


Figure 7 Geological map, structural data and cross-section through the Antananarivo Domain. This transect does not contain the complex folding that we observe in the Ikalamavony and Itremo domains. Here, foliations are steeply west-dipping. Geological polygons from Roig et al. (2012). Sections generated 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).

# 3. 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). The results are summarised in Figure 8 and documented more extensively in Supplementary B. Detailed methodologies for these techniques are also provided in Supplementary B. Rb–Sr mica dating using LA-QQQ-ICP-MS is a novel technique, and this data presents some of the first published data of its kind. The good agreement of Rb–Sr mica ages with U–Pb apatite ages, which have similar closure temperatures, demonstrates that Rb–Sr mica dating is a useful tool for dating medium-temperature events. Due to the abundance of samples (41 in total), detailed results for each sample and outcrop, including plots of isotopic data, are provided in Supplementary B. Sample descriptions, location and age data are summarised in Table 1. Isotopic data are given in Supplementary C.

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 | Undeformed K-feldspar rich granite. Intrudes deformed gabbro interpreted as the Dabolava Suite | Ambalavao | -19.5443 | 45.47028 | 182 | 576 ± 24 | – | 519 ± 69 | 505 ± 59 |
| **M16-32** | West | 2/A | Coarse-grained gneiss with 1–2 cm biotite phenocrysts | Betsiboka | -19.6107 | 46.53399 | 989 | 2553 ± 24 | 519 ± 11 | 446 ± 161 | 502 ± 20 |
| **M16-33** | West | 2/D | Undeformed fine-grained granodioritic dyke, intrudes M16-32 | 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-feldspar rich deformed dyke | Betsiboka | -19.6107 | 46.53399 | 989 | 2583 ± 26<br>2494 ± 14 (*) | 502 ± 6 | – | 513 ± 18 |
| **M16-15** | West | 3/A | K-feldspar and biotite rich, foliated orthogneiss | Betsiboka | -19.7239 | 46.62736 | 1067 | 2456 ± 17 | 492 ± 5 | 624 ± 152 | 528 ± 18 |
| **M16-16** | West | 3/B | Undeformed granite, occasionally very weakly foliated | 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 | c. 795 | 494 ± 7 | 526 ± 39 | 492 ± 51 |
| **M16-46** | East | 4/A | Foliated orthogneiss, medium-grained, K-felsdpar and biotite rich | Betsiboka | -19.1599 | 47.51211 | 1351 | 2522 ± 8<br>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<br>532 ± 44 (#) | – | 657 ± 98 | – |
| **M16-45** | East | 5 | Medium-grained granite, undeformed | Ambalavao | -19.0869 | 47.54429 | 1312 | 543 ± 18 | 507 ± 35 | – | 512 ± 16 |
| **M16-52** | East | 6/A | 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 K-feldspar rich granite | Ambalavao | -18.589 | 47.23721 | 1359 | c. 568 | 500 ± 10 | 537 ± 35 | 521 ± 18 |

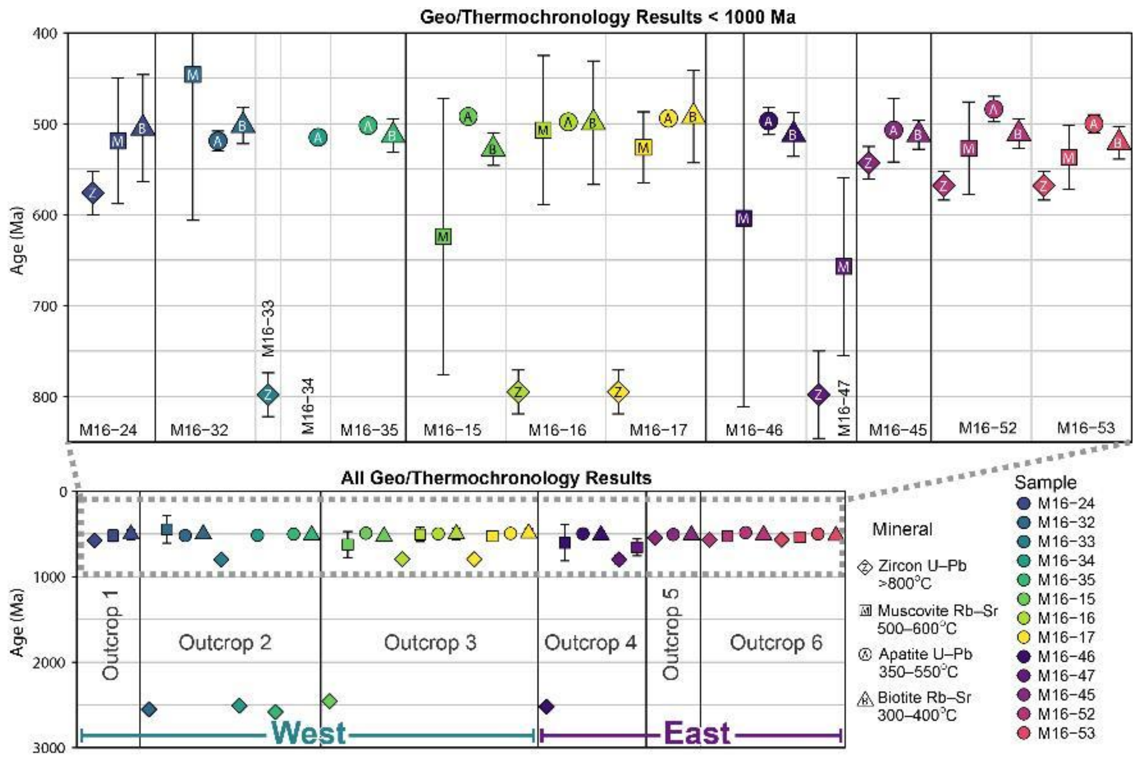


Figure 8 Summary of geo/thermochronology data for each sample and outcrop collected in this study. Error bars are 2σ. Sample locations shown in Figure 2. The bottom figure shows all of the sample data collected, and the top image zooms in on all data that are younger than 1000 Ma. Plots of isotopic data associated with these calculated ages are presented in Supplementary B.

# 4. Discussion

## 4.1. Structural evolution of central Madagascar

The structural style of the Ikalamavony and Itremo domains are indistinguishable and we suggest that D1–D3 (and possibly D4) in both domains were the result of 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, commonly 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 D1–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).

A structure with very similar geometry and orientation to the type 2 interference pattern highlighted in Figure 5, was modelled by Armistead et al. (2018). They 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).

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, developed as a result of ~northeast-southwest shortening (present day orientation). We propose that continental collision southwest of the Ikalamavony Domain and Itremo sub-domain was responsible for deformation in central Madagascar. Subduction prior to continental collision was ~southwest dipping.

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).

The weakly defined, open folding associated with D4 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).

In the eastern part of the study area, east of the 'structural style boundary' in Figure 2 and Figure 3, the deformation style changes in orientation and intensity. Transect C–C' is less deformed than the Ikalamavony and Itremo transects (Figure 9), and we do not observe any complex fold interference patterns 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 550 ± 4 Ma (Raharimahefa and Kusky, 2010). Collins et al. (2003a) constructed a cross-section from Antananarivo eastwards to Brickaville along the east coast of Madagascar (Figure 9). 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.

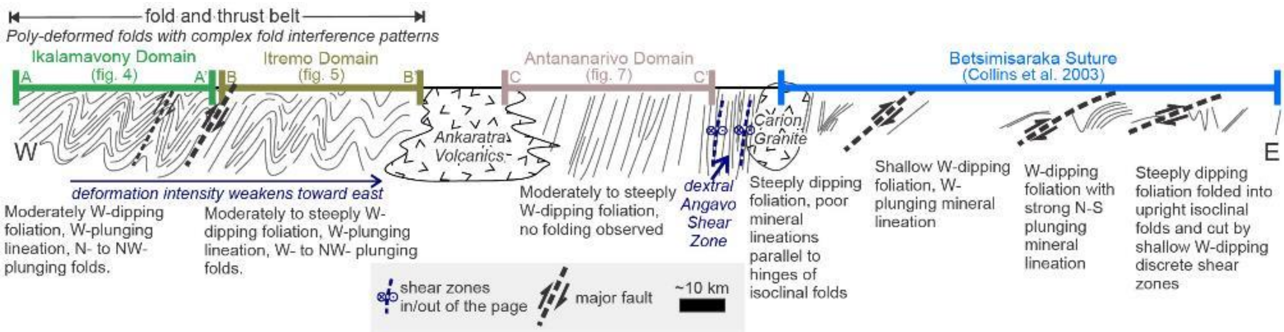


Figure 9 Schematic cross-section through central Madagascar from Miandrivazo to Brickaville, combining the cross-sections of Figure 4, Figure 5, Figure 7 and that of Collins et al. (2003a). The Ikalamavony and Itremo domains preserve the same structural styles and kinematics. A change in structural style 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.

## 4.2. Temporal constraints on deformation

### 4.2.1. *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); Raharimahefa and Kusky (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 1b; Figure 9). This indicates 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).

### 4.2.2. *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 8). 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 within a narrow age range from 519 ± 11 Ma to 484 ± 14 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 within a narrow range between 657 ± 98 Ma and 446 ± 161 Ma for muscovite (albeit, large uncertainties) and between 528 ± 18 Ma and 492 ± 51 Ma for biotite. 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 either reset to at least ~500°C or cooled through ~500°C at c. 500 Ma.

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. 520–480 Ma thermo-tectonic event affected the entire island such that it cooled synchronously through ~500–300°C at c. 500 Ma. The c. 520–480 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 regions distal from the collision zone and in rocks that are not in contact with the Ambalavao Suite — 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.

## 4.3. Tectonic model for the evolution of central Madagascar

The boundaries between the major domains in southern Madagascar represent possible suture zones responsible for deformation in the Ikalamavony Domain and Itremo sub-domain. 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 (Figure 10a) (Cox et al., 1998; 2004) and that the Ikalamavony Domain evolved as an exotic island arc terrane (Figure 10b) (Archibald et al., 2017a). The presence of the Imorona-Itsindro Suite in the Ikalamavony Domain indicates that it must have accreted to central Madagascar before c. 850 Ma (Figure 10c). A large west-dipping thrust fault separating the Ikalamavony Domain from the Itremo sub-domain (Figure 5; Figure 9), possibly represents this suture zone (schematic thrust in Figure 10c). 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 continental collision. Increasing deformation intensity in the Ikalamavony Domain and the orientation of structures imply that the collision zone must have lain southwest of these domains.

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. This interpretation was based on c. 630–600 Ma metamorphism restricted to the west of this high-strain zone, and widespread c. 580–520 Ma magmatism and metamorphism on both sides of the high-strain zone. In this model, the subduction zone was east-dipping (present day direction), and resulted in the syn- to post-tectonic Ambalavao granites throughout Madagascar (Figure 10). However, the structures we have described and those described by Tucker et al. (2007) require a ~west-dipping (top to the east; present day direction) sense of movement, making an east-dipping subduction zone beneath the Antananarivo Craton at this time unlikely.

These authors interpreted that another west-dipping subduction zone was active beneath the Vohibory Domain until c. 650 Ma (Figure 10e), and that collision between the Vohibory Domain and Androyen Domain occurred at c. 630–610 Ma (Figure 10f), outboard from the Antananarivo Craton (marked by the Ampanihy high-strain zone/Vohibory Suture in Figure 1b). In this model, subduction was west-dipping (Figure 10e,f). Horton et al. (2016) also interpreted that the Androyen and Anosyen domains collided with the Vohibory Domain at c. 630 Ma, based on monazite and zircon geochronology. In our view, this collision zone is most consistent with evidence we see from a structural perspective. We therefore propose that the Vohibory suture was responsible for high intensity deformation and polydeformed folds in the Ikalamavony and Itremo domains. It's likely that deformation associated with this event is present in the Androyen and Anosyen domains, however structures are much more steeply dipping and more highly strained here, meaning that recognition of distinct folding events is difficult (Figure 10e). As interpreted by Boger et al. (2014); Boger et al. (2019), the Vohibory Suture would have been west-dipping (present day direction), resulting in the emplacement of the c. 630 Ma Marasavoa Suite in the Vohibory Domain. Metamorphism of this age is also recorded in the Androyen Domain.

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 10g). Although we did not look at this region in great detail, the changes across central Madagascar from west to east, combined with extensive structural studies published on eastern Madagascar (Collins et al., 2003a; Martelat et al., 2000; Nédélec et al., 2000; Raharimahefa and Kusky, 2006; Raharimahefa and Kusky, 2009; Raharimahefa et al., 2013), indicate that a west-dipping subduction zone was active at c. 550 Ma, somewhere in the region of the Betsimisaraka Suture.

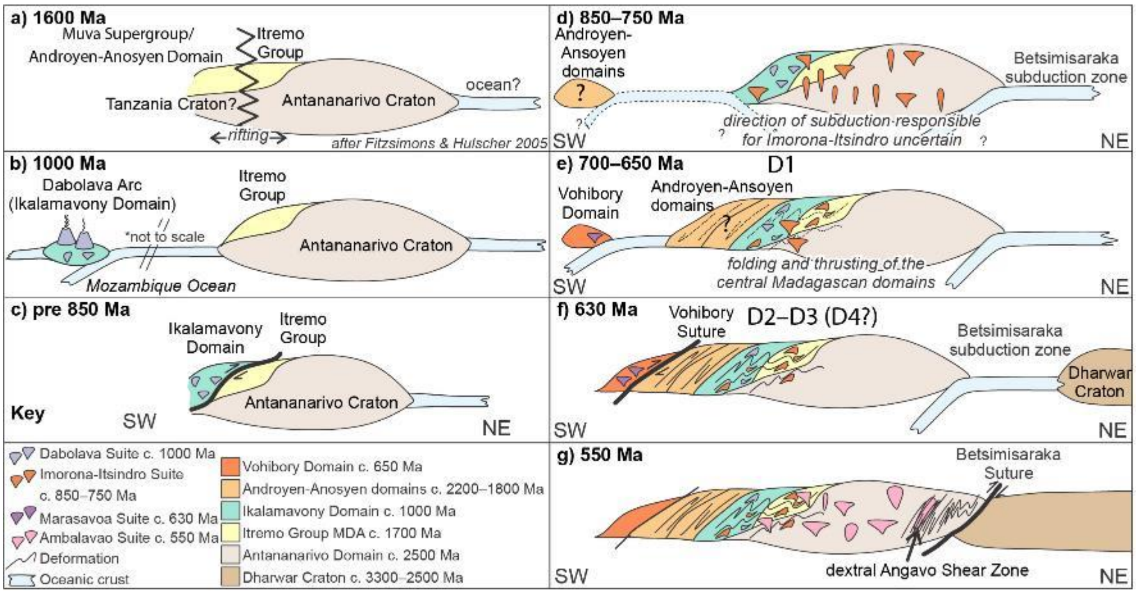


Figure 10 Schematic diagram showing our interpretation of the evolution of central Madagascar. a) Sometime after the deposition of the Itremo Group/Muva Supergroup onto the Antananarivo Craton/Tanzania Craton, these regions begin to rift (Cox et al., 2004; Fitzsimons and Hulscher, 2005); b) at c. 1000 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, with the beginning of deformation (D1); f) closure of the Mozambique ocean along the Vohibory Suture resulting in complex deformation that we have interpreted in central Madagascar (D2–D3) and possibly D4; and g) final amalgamation of central Gondwana along the c. 550 Ma Betsimisaraka Suture.

# 5. Conclusions

We have integrated remote sensing, field data and thermochronology to unravel complex deformation in the Ikalamavony and Itremo domains of central Madagascar. We have recognised four generations (D1–D4) of deformation that resulted in complex fold interference patterns in the Ikalamavony and Itremo domains of central Madagascar. We interpret deformation here as the result of c. 630 Ma collision of Azania with Africa along the Vohibory Suture in southwestern Madagascar. In eastern Madagascar, deformation is syn- to post-550 Ma, which likely formed in response to the final closure of the Mozambique Ocean along the Betsimisaraka Suture that amalgamated Madagascar with the Dharwar Craton of India. Apatite U–Pb and novel LA-QQQ-ICP-MS muscovite and biotite Rb–Sr thermochronology indicate that much of central Madagascar cooled through ~500°C at c. 500 Ma. We have shown the importance of using medium-temperature thermochronometers to date the cooling stages after orogenesis, and the potential for Rb–Sr mica dating to provide useful thermochronological constraints.

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