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ORIGINAL PAPER

Kinematics and significance of a poly-deformed crustal-scale shear zone in central to south-eastern Madagascar: the Itremo–Ikalamavony thrust

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Abstract Across the crystalline basement of Madagascar, late Archaean rocks of the Antananarivo Block are tectonically overlain by Proterozoic, predominantly metasedimentary units of the Ikalamavony and Itremo Groups of the Southwest Madagascar Block. The generally west-dipping tectonic contact can be traced for more than 750 km from NW to SE and is referred to here as the Itremo–Ikalamavony thrust. The basal units of the SW Madagascar Block comprise metasedimentary quartzites with the potential to preserve a multistage deformation history in their microstructures. Previous studies suggest contrasting structural evolutions for this contact, including eastward thrusting, top-to-the-west directed extension and right-lateral strike-slip deformation during the late Neoproterozoic/Ediacaran. In this study, we integrate remote sensing analyses, structural and petrological fieldwork, as well as microstructural investigations of predominantly quartz mylonites from the central southern segment of the contact between Ankaramena and Maropaika. In this area, two major phases of ductile deformation under high-grade metamorphic conditions occurred in latest Neoproterozoic/early Phanerozoic times. A first (Andreaba) phase produces a penetrative foliation, which is parallel to the contact between the two blocks and contemporaneous with widespread magmatism. A second (Ihosy) phase of deformation folds Andreaba-related structures. The investigated (micro-)structures indicate that (a) juxtaposition of both blocks possibly already occurred prior to the Andreaba phase, (b) (re-)activation with top-to-the-east thrusting took place during the latest stages of the Andreaba phase, (c) the Ihosy phase resulted in regional-scale open folding of the tectonic contact and (d) reactivation of parts of the contact took place at distinctively lower temperatures post-dating the major ductile deformations.

Keywords Madagascar · Gondwana · Southwest Madagascar Block · Antananarivo Block · Thrust contact · East African Orogen · Itremo–Ikalamavony thrust

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Introduction

The crystalline basement of central and southern Madagascar is divided by a shallowly west-dipping shear zone into the late Archaean gneisses, migmatites and granitoids (e.g. Tucker et al. 1999, 2007, 2011a, 2014; Kröner et al. 2000; GAF-BGR 2008a) of the Antananarivo Block (Kröner et al. 2000; Collins 2006) to the east and Proterozoic, mainly metasedimentary rocks to the west (e.g. Itremo Group, Cox et al. 1998, 2004, Itremo or Ikalamavony–Itremo domain, GAF-BGR 2008b; Tucker et al. 2011a, b; Amborompotsy–Ikalamavony Group, Moine 1967; Müller 2000, or Ikalamavony domain, Tucker et al. 2014). The latter are grouped together with the Anosyen, Androyen and Vohibory domains as the Southwest (SW) Madagascar Block (Giese et al. 2011) (Fig. 1). This subdivision into Antananarivo and SW Madagascar Block was based on different deformation behaviours during latest stages of the East African Orogeny when the rheologically stiffer Antananarivo Block acted as an indenter deforming the rheologically softer, predominantly metasedimentary lithological assemblages of the SW Madagascar Block in a transpressive

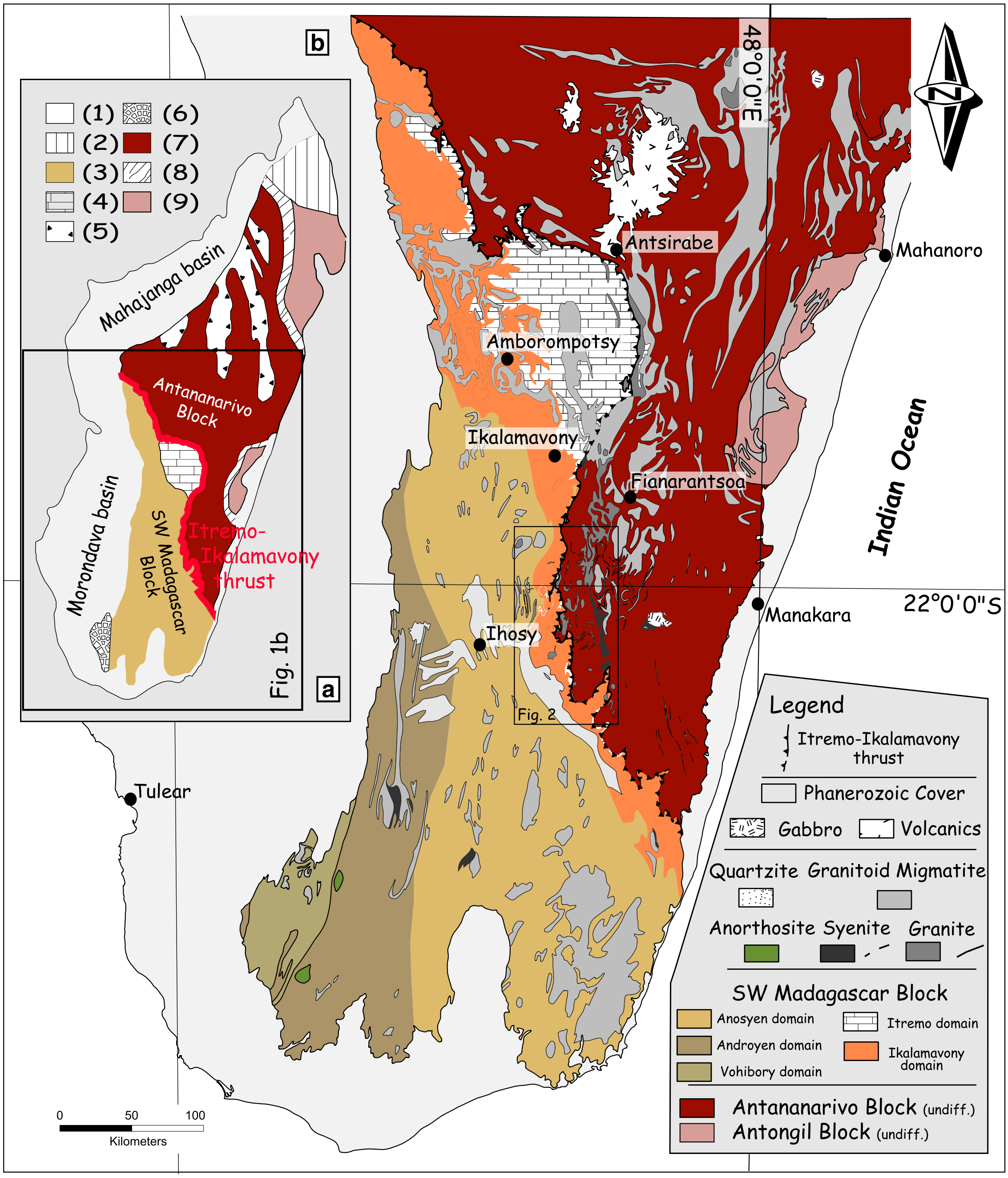


Fig. 1 **a** Geological overview map of Madagascar modified after Bésairie (1969/1970) and Collins and Windley (2002). Legend key: *1* Phanerozoic cover, *2* Bemarivo Belt, *3* Proterozoic supracrustal units of the SW Madagascar Block, *4* Itremo Group, *5* Tsaratanana Sheet, *6* Vohibory Complex, *7* Antananarivo Block, *8* Betsimisiraka Suture, *9* Antongil Block. **b** Simplified geological map of central and southern Madagascar, geology modified after Bésairie (1969/1970). Note the tectonic contact between the Antananarivo Block and the SW Madagascar Block can be traced across the whole crystalline basement of Madagascar from the NW to the SE for ~750 km. *Triangles* along the contact point towards the hanging wall. *Box* indicates map location of Fig. 2

regime (Schreurs et al. 2010; Giese et al. 2011). It does not refer to the palaeogeographic position of individual tectonic domains of the SW Madagascar Block.

While the rocks of the Antananarivo Block, underlying the relatively shallow, mostly westward-dipping contact, are metamorphosed at granulite facies conditions (e.g. Kröner et al. 2000), the metamorphic grade of the metasedimentary units of the overlying SW Madagascar Block varies from greenschist/amphibolite facies in the Itremo domain in central Madagascar (e.g. Fernandez et al. 2003), to upper amphibolite to granulite facies conditions north, south and west of the Itremo (e.g. Markl et al. 2000; Tucker et al. 2007).

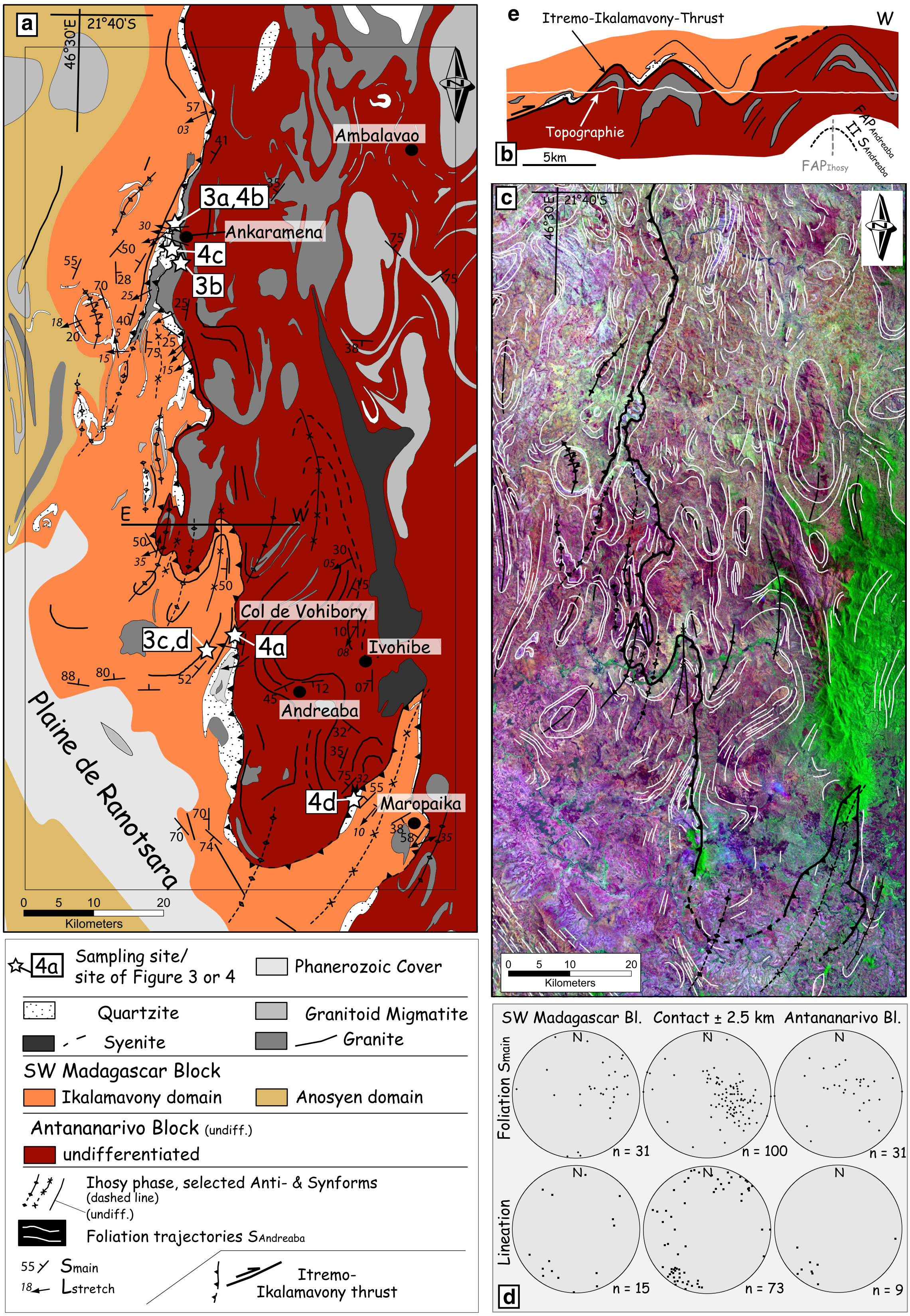
This contact, that we refer to as the "Itremo–Ikalamavony thrust", can be traced for more than 750 km from NW to SE Madagascar (Fig. 1). The identification and thorough characterization of such a major contact are of paramount importance for understanding the geodynamic evolution during and after amalgamation of Gondwana and for improving palaeogeographic reconstructions.

There is general agreement that the boundary separating the Antananarivo and the SW Madagascar Block represents a tectonic contact (e.g. Windley et al. 1994; Collins et al. 2000; Müller 2000; Tucker et al. 2011b). The nature, timing and significance, however, of associated deformation along and across this contact are controversial (e.g. Gregoire 1999; Collins et al. 2000; Müller 2000; Fernandez and Schreurs 2003; Tucker et al. 2007, 2011a, b; De Waele et al. 2011; Boger et al. 2014). Most of the detailed work carried out until now has focussed on the central part, where the low-grade Itremo Group metasediments are in contact with the Antananarivo Block (Fernandez and Schreurs 2003; Fernandez et al. 2003; Nédélec et al. 2003; Tucker et al. 2007). Only a few studies have investigated the prolongation of the contact to the north and to the south (e.g. Gregoire 1999; Collins et al. 2000; Müller 2000; Tucker et al. 2011b). Parts of this contact are also referred to as the Betsileo (extensional) shear zone (Collins et al. 2000). In this study, we integrate remote sensing analysis, structural and petrological fieldwork as well as microstructural and textural analysis of predominantly mylonitic quartzites to characterize the central southern part of the contact, where high-grade metasediments of the Ikalamavony Group directly overlie the rocks of the Antananarivo Block. We will demonstrate that this contact has a prolonged deformation history of reactivation both at high- and low-grade metamorphic conditions and with contrary kinematics (thrusting and normal faulting). These results and its geodynamic implications will then be discussed in the light of the whole contact. Investigation of major fault zones can also be economically important since perculating hydrothermal fluids can precipitate minerals such as gold (e.g. Morey et al. 2007), uranium, copper and many others (e.g. Pal et al. 2010). To date, such mineralizations have not been observed along the Itremo–Ikalamavony thrust, but both domains, the Ikalamavony domain in the hanging wall and the Antananarivo domain in the footwall, comprise a variety of mineral deposits (GAF-BGR 2008a, b).

Geology and structures

The Antananarivo Block is the largest tectonic unit of the crystalline basement of Madagascar (e.g. Collins 2006) and consists mainly of late Archaean granitoids, orthogneisses and migmatites (Kröner et al. 2000) with minor occurrences of Bt-gneiss, sillimanite-bearing quartzites, sillimanite-graphite schists, cordierite gneisses and amphibolites (Tucker et al. 1999; GAF-BGR 2008a). These rocks, together with parts of the SW Madagascar Block, were intruded by (a) large volumes of middle Neoproterozoic calc-alkaline intrusives between 800 and 720 Ma (e.g. Imorona–Itsindro suite, Handke et al. 1999; Tucker et al. 1999, 2007, 2014), (b) high-alkaline granites to syenites in the form of ~N–S elongate plutons (e.g. Andringitra syenite, Nédélec and Grujic 2001) and as foliation parallel sills (Nédélec et al. 1995; Paquette and Nédélec 1998; Gregoire 1999) in Late Neoproterzoic times and (c) voluminous granitic plutons and foliation parallel sills of the Ambalavao Suite (GAF-BGR 2008b; Tucker et al. 2014) in early Cambrian times (e.g. Vohimavo granite; Handke 2001, Laimanasika granite; Giese et al. 2011). In addition, exclusively the Ikalamavony domain hosts gabbroic to granitic calc-alkaline intrusions of the ~1 Ga old Dabolava suite (Tucker et al. 2014), suggesting that the Ikalamavony tectonic domain formed in an island (Boger et al. 2014) or magmatic arc setting (Tucker et al. 2014).

The Ikalamavony (e.g. Moine 1967) and Itremo domains (e.g. Fernandez et al. 2003; Cox et al. 2004) are the easternmost largely metasedimentary units of the SW Madagascar Block (see above, Giese et al. 2011). The Itremo Group mainly consists of quartzites, dolomitic marbles and metapelites, metamorphosed at greenschist to amphibolite facies conditions (e.g. Fernandez et al. 2003). They unconformably overlie amphibolites and gneisses, which are likely rocks of the Antananarivo Block (Collins 2006; Boger et al. 2014; Tucker et al. 2014). The Ikalamavony domain, north, south and west of the Itremo Group (Fig. 1) predominantly consist of quartzites, marbles, calc-silicates, amphibolites, meta-psammites and—pelites as well as feldspatic leucogneisses interpreted as metavolcanics (Rakotoarimanana 2001). Metamorphic grade within the Ikalamavony domain reached higher amphibolite facies conditions west and southwest of Itremo (Tucker et al. 2007; GAF-BGR 2008b), towards the southeast anatexis, and migmatization of Ikalamavony gneisses suggests slightly increasing temperatures exceeding 700 °C (GAF-BGR 2008b; Horton et al. 2016). Tucker et al. (2011a, b, 2014) point out the distinctly different detrital zircon age populations, intrusive suites and depositional settings between the Itremo and Ikalamavony Groups and argue


a
21°40'S
46°30'E
Ambalavao
3a,4b
Ankaramena
4c
3b
E
W
Col de Vohibory
3c,d
4a
Ivohibe
Andreaba
4d
Maropaika
Plaine de Ranotsara
0 5 10 20
Kilometers
e
Itremo-Ikalamavony-Thrust
W
Topographie
b
5km
FAP Andreaba
II S Andreaba
FAP Ihosy
c
21°40'S
46°30'E
0 5 10 20
Kilometers
Sampling site/
site of Figure 3 or 4
Phanerozoic Cover
Quartzite
Granitoid Migmatite
Syenite
Granite
SW Madagascar Block
Ikalamavony domain
Anosyen domain
Antananarivo Block (undiff.)
undifferentiated
Ihosy phase, selected Anti- & Synforms
(dashed line)
(undiff.)
Foliation trajectories SAndreaba
Smain
Lstretch
Itremo-
Ikalamavony thrust
SW Madagascar Bl.
Contact ± 2.5 km
Antananarivo Bl.
Foliation Smain
Lineation
n = 31
n = 100
n = 31
n = 15
n = 73
n = 9
d

◂**Fig. 2** **a** Detailed geological map around the central southern segment of the contact. Selected strike and dip direction of S_{main} and plunge direction, respectively, of associated stretching and/or mineral lineations as indicated. Traces of selected (partly assumed) Ihosy phase fold axial planes affecting the tectonic contact in the immediate vicinity of the contact are shown. *White boxes* indicate sample locations of outcrops and microstructures presented in Figs. 3, 4 and 5. Location of map is given in Fig. 1. **b** Schematic E–W cross section illustrating the folded Itremo–Ikalamavony thrust contact (not to scale). Trace of cross section is indicated in Fig. 2a; FAP, fold axial plane; II, parallel; Andreaba and Ihosy refer to deformation phases as described in the text. **c** Landsat ETM*7 false-colour composite image (bands 7, 4 and 1 = *red*, *green*, and *blue*) with interpretation of ductile foliation trajectories of S_{main} (=$S_{Andreaba}$) and fold axial traces of Ihosy phase upright folds. Image corresponds to same area as presented in Fig. 2a. Landsat ETM*7 images courtesy of U.S. Geological Survey (LP DAAC, NASA, USGS). **d** Stereographic equal area, lower hemisphere projection of S_{main} (pole projection) and associated stretching and/or mineral lineation, divided in measurements within rock of the SW Madagascar Block (*left*), within both blocks but in proximity to the contact (±2.5 km max, usually ±500 m; middle) and within the Antananarivo Block (*right*)

against older studies, suggesting that the protolith rocks of both groups represent lateral equivalents formed in a distal-to-proximal shelf environment (e.g. Cox et al. 2004; Collins et al. 2001a, b, 2003; Collins and Windley 2002; Tucker et al. 1999, 2007). Since the depositional age of the Itremo (meta-)sediments is not well constrained, the formation of these rocks either predates (e.g. Tucker et al. 2014) or postdates (Boger et al. 2014) formation of the Ikalamavony Group and intrusion of the Dabolava suite. Tucker et al. (2014) postulate sedimentation of the Itremo Group on the western shelf of the eastern Gondwanian craton between <1.78 and 1.1 Ga, followed by the formation of an active magmatic arc with deposition of the Ikalamavony Group and intrusion of the Dabolava suite. In contrast, Boger et al. (2014) account the Ikalamavony rocks to an island arc that was formed between ~1.07 and 1.01 Ga, subsequently accreted to the Antananarivo(-Dharwar)craton between 1 and 0.92 Ga and finally covered by an extensive blanket of Itremo sediments thereafter. The latter authors base their arguments on the exclusivity of the Dabolava intrusives to the Ikalamavony domain, but the common intrusive history during Imorona–Itsindro and Ambalavao magmatism. Further, they doubt the exceptionally young zircon ages of *c.* 950 and 840–750 Ma within the Itremo Group (Cox et al. 1998; De Waele et al. 2011) to represent metamorphic zircon growth but interpret them as detrital minerals. In either case, both Itremo and Ikalamavony domains rest on top of the Antananarivo Block.

Raoelison (1997) describes a west-directed thrusting of the Itremo group on top of gneisses of the Ikalamavony Group; Tucker et al. (2007, 2011a, b, 2014) suggest the latter Group to represent an eastward transported thrust nappe (their high-grade internal nappes), which was thrust upon the Itremo Group (their low-grade external nappes), together representing a nappe stack tectonically overlying the Antananarivo Block. Also, Nédélec et al. (2003) and Fernandez et al. (2003) identified mylonites at the base of the Itremo Group, interpreted as the result of ductile top-to-the-east thrusting. Nédélec et al. (2003) infer an age of ~645 Ma for the tectonic transport, predating the assumed extensional collapse at ~630 Ma (Paquette and Nédélec 1998; Collins et al. 2000), while Tucker et al. (2007) restrict emplacement of nappes to the period between 720 and 570 Ma based on crosscutting relationships of intrusive rocks. According to Collins et al. (2000), the northern contact between the Itremo Group and the Antananarivo Block preserves structures indicating top-to-the-east thrusting, while the southern contact [south of Antsirabe (Fig. 1)] features a major top-to-the-west extensional detachment of Neoproterozoic age, the Betsileo shear zone. As postulated by the latter authors, this extensional shear zone can be traced to the south of the low-grade Itremo Group, where high-grade rocks of the Ikalamavony Group directly overlie rocks of the Antananarivo Block. In this region of central southern Madagascar (south of Fianarantsoa, Figs. 1, 2), at least two major phases of ductile deformation between 550 and 520 Ma can be identified within the Antananarivo, Ikalamavony and Anosyen domains (Martelat et al. 2000; Giese et al. 2011) indicating antecedent juxtaposition of all three domains: a first phase of pervasive deformation (Andreaba phase of Giese et al. 2011; D1 of Martelat et al. 2000) occurred under HT/HP granulite facies conditions in the Anosyen domain (M1; 800–950 °C, 0.7–0.9 GPa; Nicollet 1990; Martelat et al. 1999; Markl et al. 2000) and at least higher amphibolite facies conditions within Ikalamavony and Antananarivo domains (690 °C, 0.65 GPa in Ikalamavony domain; GAF-BGR 2008b; Horton et al. 2016). Widespread granitic intrusions of the Ambalavao suite (~550–510 Ma; Giese et al. 2011), migmatization and anatexis (Kröner et al. 1996) occurred largely coeval with deformation. However, a variable strong solid-state overprint in migmatites and granites indicates that the majority of (partially) molten rocks crystallized prior to cessation of deformation (Giese et al. 2011). The resulting recumbent folds and penetrative foliation (Andreaba phase; Giese et al. 2011), which very often shows an E–W- to NE–SW-oriented mineral and/or stretching lineation (Martelat et al. 2000, this study), were subsequently affected by E–W transpression (Martelat et al. 2000) of the Ihosy phase producing ~N–S-trending upright folds (Fig. 2; Schreurs et al. 2010; D2 of Martelat et al. 2000). This deformation resulted in crustal-scale strain partitioning creating low-strain zones where folds gently folding the $S_{Andreaba}$ are separated by ~N–S-trending high-strain zones. The low-strain zones are lacking a newly developed axial planar foliation, while the high-strain zones are characterized by a near-vertical axial planar foliation and rootless folds (Martelat et al. 2000). Metamorphic conditions of M1 may pertain during this deformation phase, as indicated by anatexis, charnokitization and garnet crystal plastic behaviour associated with high-strain zone formation (Martelat et al. 2012). Gregoire (1999) proposes localization of such a high-strain zone with right-lateral displacement parallel to the contact between the Antananarivo and the SW Madagascar Block north of Ankaramena. Small, apparently undeformed discordant dykes and leucosomes in both the Antananarivo and the SW Madagascar Block, crosscutting the foliation(s) of both deformation phases, yield crystallization ages between ~530 and 515 Ma (Kröner et al. 1996; Müller 2000; Giese et al. 2011). These are interpreted to mark the termination of major ductile deformation in central southern Madagascar (Schreurs et al. 2010). Syn- to post-tectonic intrusions of the Ambalavao suite (GAF-BGR 2008b; Tucker et al. 2011b; Boger et al. 2014), dated between ~545 and 530 Ma, are present in different parts of Madagascar, e.g. the Vohimavo granite in the Itremo Group (Handke 2001) and the Carion granite in the Antananarivo Block in central Madagascar (Kröner et al. 2000; Meert et al. 2003).

Parts of central southern Madagascar have preserved evidence for a younger HT/MP (M2) granulite facies metamorphism, overprinting the HT/HP M1 event, characterized by cordierite-producing mineral assemblages (650–730 °C, 0.3–0.5 GPa; Gregoire 1999; Markl et al. 2000; Martelat et al. 2000) and partial melting in the middle crust (Berger et al. 2006). Boger et al. (2012) identifies cordierite production during both prograde and retrograde metamorphisms in the Anosyen domain with deformation of the prograde cordierite-bearing mineral assemblages during Ihosy phase upright folding and subsequent poikiloblastic cordierite overgrowing prograde mineral assemblages during retrogression. In the Ikalamavony domain, Giese et al. (2011) describe cordierite-bearing leucosomes overprinting and partly discordantly crosscutting Ihosy phase structures (their Fig. 4c, d) clearly indicating late to post-Ihosy phase timing of this metamorphism, which may be correlated with retrograde metamorphic path described by Boger et al. (2012). Markl et al. (2000) suggest that M2 metamorphism is related to intrusive activity between 540 and 520 Ma, whereas in contrast, Berger et al. (2006) interpret Cambrian/Ordovician and Ordovician/Silurian monazite ages between 500 and 490 Ma, and 460 and 420 Ma, respectively, as the timing of metamorphism. However, the metamorphic overprint is very often incomplete (e.g. Markl et al. 2000) and seems to be predominantly, but not exclusively, affecting rocks that are more intensely deformed by the Ihosy phase (Berger et al. 2006; Giese et al. 2011). Increased fluid flow in near-vertical shear zones might be responsible for this phenomenon (Pili et al. 1997, 1999).

Field area

The contact between the Ikalamavony Group and the Antananarivo Block has been investigated in greater detail in the area between southern latitude of 21°45′ and 22°44′. In the area around Ankaramena and the Col de Vohibory (Fig. 2), the base of the Ikalamavony Group is marked by thick sequences of metasedimentary quartzites, which can easily be traced on satellite images and in the field (Fig. 2c). Away from these areas, the quartzites become more impure and pass into partly migmatized biotite ± amphibole gneisses, one of the typical lithologies of this tectonic unit (e.g. Moine 1967; Müller 2000). In these areas, however, the discrimination between the two blocks becomes very difficult in the field as well as on remote sensing imagery, especially when tracing the contact further SE of Maropaika (Fig. 2c). We therefore concentrate our structural investigations on the areas featuring quartzite at the base and within the Ikalamavony Group.

Structural data

The contact between the Antananarivo Block and the SW Madagascar Block is oriented parallel to the regional tectonic foliation $S_{Andreaba}$, which dominates also the rocks in the footwall and is parallel to the lithological stratification that may still represent original bedding (S_0) of the metasedimentary sequence in the hanging wall (Figs. 1, 2). Usually, the $S_{Andreaba}$ and the contact strike ~N–S and dip gently towards the west (Fig. 2d). There is no significant change in orientation of the main foliation $S_{Andreaba}$ in close vicinity to the contact (i.e. dip measurements within about ±2.5 km east and west of the contact) and farther away (distance of about ±20 km; Fig. 2d). However, south of Ankaramena and as far as Maropaika the contact is, together with $S_{Andreaba}$, folded by ~N–S-trending Ihosy phase folds, resulting in a complex trace because of the rugged topography and changes in dip direction (Fig. 2a). A newly developed axial plane foliation related to the Ihosy phase is not recognized along this part of the contact. The rocks in both the hanging and footwall are affected by open to tight upright Ihosy phase folds (Fig. 2b). About 10 km SW of Ankaramena, quartzite outlines an east vergent fold, with a shallow-dipping western limb and a steeply westward-dipping eastern limb. In the core of this circular antiform, high-grade migmatites are present, probably belonging to the Antananarivo Block. This interpretation, however, remains speculative because of the deep weathering of the rocks and their sparse exposure. South of Ankaramena, the sequence of quartzites, is duplicated, in between which, high-grade migmatized amphibolite and orthogneisses, probably also belonging to the Antananarivo Block, can be found. In both, the rocks of the hanging wall and the footwall W to SW and NE plunging stretching and/or mineral lineations (Qtz, Fsp or Bt, mineral abbreviations after Siivola and Schmidt 2007) are abundant (Figs. 2d, 3a, b, c). Much less frequent, lineations are parallel to the strike of the contact. In particular, the SW and NE plunging lineations seem to predate the folding associated with the Ihosy phase, representing a former unique population, most likely related to the Andreaba phase, which was folded around the ~N–S-trending Ihosy phase folds (Fig. 2d). Retrodeforming the upright Ihosy phase folds would reorient the SW and NE plunging lineations into a subhorizontal orientation, likely representing the major tectonic transport direction of the hanging wall towards the NE. While the stretching lineation is best developed with NE and SW plunging maxima within close vicinity to the contact (±2.5 km), there is a fairly large scatter in different trends within the Ikalamavony domain in the hanging wall (Fig. 2d). Stretching lineations within the Antananarivo Block farther away from the contact (>2.5 km) are scarce but uniformly plunging to the SW suggesting less reorientation by Ihosy phase upright folding (Fig. 2d). West of the Col de Vohibory, a spectacular L-tectonite (Flinn 1965) features a SW plunging lineation composed of rodded mylonite (sample JGI 110; Fig. 3c, d). Within the quartzitic sequence, the structures visible in outcrop vary significantly from virtually undeformed, coarse- and sugar-grained quartzite beds to extremely elongate, mostly fine-grained quartz bands parallel to S_{Andreaba}. Occasionally, oblique foliations in pure quartz mylonites with shape-preferred orientation (SPO) in uniform-sized mm-scale grains indicate predominantly a ~top-to-the-east sense of shear (Fig. 3a). Impure quartzites and gneissic rocks, mainly featuring mylonitic microstructures like S–C-type fabrics (Lister and Snoke 1984) point towards the same kinematics. In this case, an intersection lineation, trending parallel to the contact, is visible.

Microstructures, shape-preferred orientation (SPO) and crystallographic preferred orientation (CPO)

Oriented samples of pure quartzites/quartz mylonites from (a) different locations along strike of the contact and (b) at variable distance from the base of the metasedimentary sequence were analysed in order to examine the kinematics and conditions of deformation (Figs. 2, 3a, 4). Quartz mylonites have the potential to record temperature (e.g. Kruhl 1998; Stipp et al. 2002a) and sense of shear (e.g. Schmid and Casey 1986; Passchier and Trouw 2005) during deformation. The temperature can be estimated either based on the opening angle of the resulting c-axis fabric (Kruhl 1998) or on the inferred deformation mechanism of a dynamic recrystallization microstructure (Stipp et al. 2002a). In both cases, however, temperature must be the primary controlling factor since dynamic recrystallization microstructures are also sensitive to strain rate and hydraulic weakening (Law 2014 and references therein). Shear sense can be derived from either the development of a crystallographic preferred orientation (CPO), when different intracrystalline slip systems are activated, resulting in unique pole figures for c- (and a-) axes orientation (e.g. Schmid and Casey 1986; Passchier and Trouw 2005), or the development of a shape-preferred orientation (SPO) with elongate grain shapes forming oblique to the orientation of the shear zone boundaries (Herwegh and Handy 1996). SPOs are easily reset and start forming from the beginning of every deformational increment, while the formation of a CPO requires steady-state conditions (e.g. Herwegh and Handy 1996). In particular, CPOs can thus retain information about earlier deformation phases/increments when only partially overprinted by subsequent deformation (Herwegh and Handy 1996). Quartz CPOs from polymineralic rocks may not be representative for regional shear sense but rather represent the local scale of the mineral aggregate (Kilian et al. 2011), and thermal overprint can alter, amplify or erase older CPOs (Augenstein and Burg 2011). Thin sections of our samples all indicate the preservation of more than one increment of deformation at different conditions; the resulting c-axis patterns (pole figures) thus represent the bulk fabric of finite strain—and not steady-state deformation (see below). However, different recrystallization mechanisms can clearly be distinguished. We thus regard the opening angle (Kruhl 1998) as not representative to extract temperature but apply the thermometry after Stipp et al. (2002a) who calibrated the temperatures of naturally deformed quartz veins to syn-tectonically grown mineral assemblages and compared the natural microstructure to experimental ones (Stipp et al. 2002a, b).

Although the orientation of most of the samples did not vary significantly with regard to the external reference system given by the S_{Andreaba} and L_{Andreaba} (all thin sections are cut parallel to the lineation and perpendicular to the foliation), the results show a large variety of microstructures, indicating that different deformation mechanisms were involved at significantly varying temperatures and/or strain rates (Fig. 4). Furthermore, microstructures indicate that tectonic movement was not restricted to a unique narrow shear zone, but was distributed within a several hundreds of metres thick metasedimentary pile by (re)-activating shear planes predominantly oriented parallel to the S_{Andreaba}.

The most common high-temperature (HT) microstructures include:

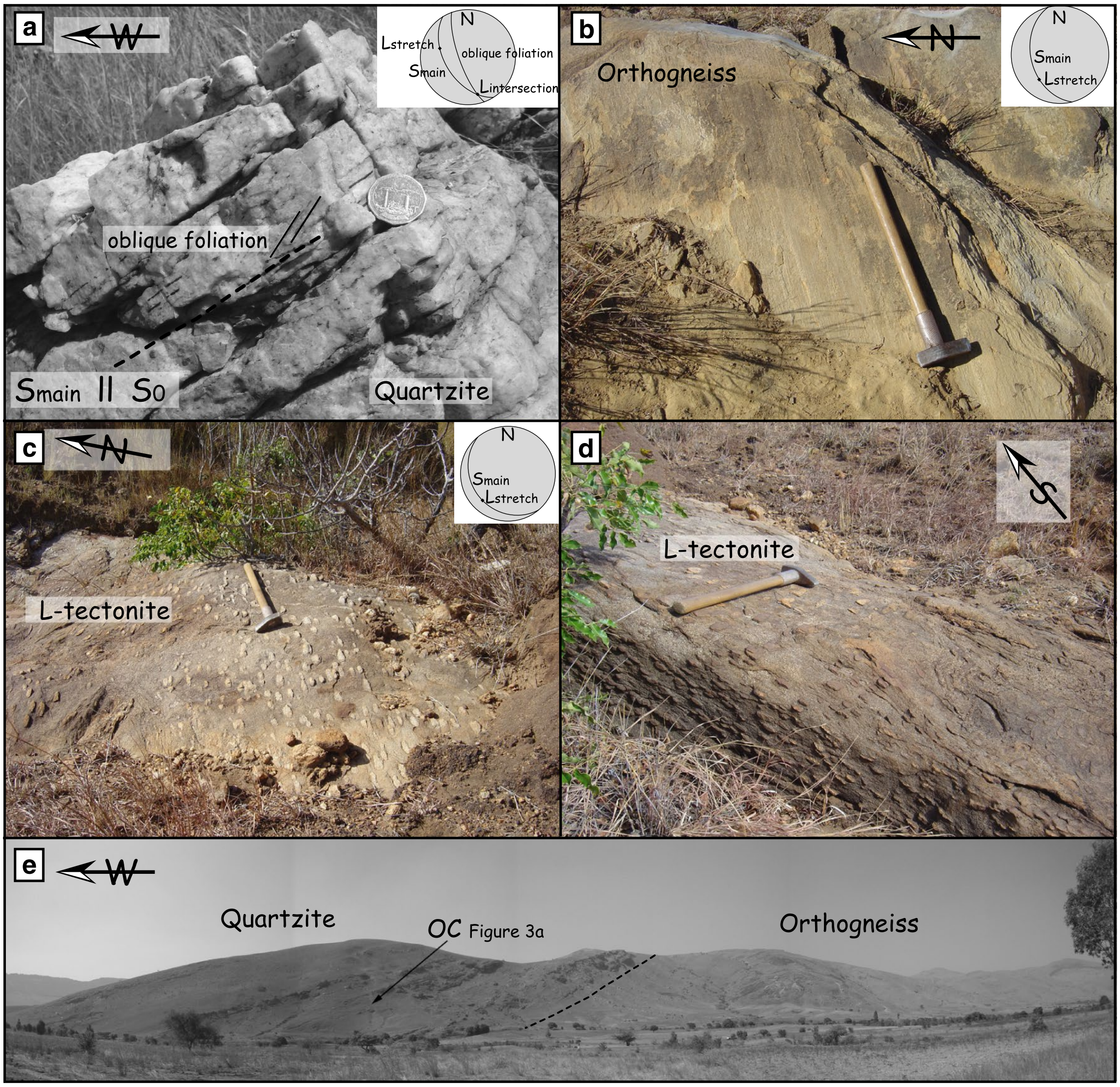


Fig. 3 **a** Quartzite of the Ikalamavony Group north of Ankaramena in the hanging wall of the contact. Outcrop features a mylonitic, oblique foliation to the composite foliation, indicating a top-to-the-ENE tectonic transport direction, which is roughly consistent with the west plunging stretching lineation. Structural data presented in stereographic projection, equal area and lower hemisphere. Outcrop site is indicated in Figs. 2a, 3e. **b** Strongly foliated orthogneiss from the Antananarivo Block in the footwall of the contact. Feldspar aggregates define a stretching lineation, which gently plunges to the SW (parallel to the hammer). **c**, **d** L-tectonite (mylonitic gneiss) in the Ikalamavony Group. Mylonite rods are strongly elongated and moderately flattened, microstructures refer to Fig. 5. Note the identical orientation of foliation and lineation with the orthogneiss in Fig. 3b, illustrated by the stereographic projections (although both sites are ~60 km apart along strike of the contact). **e** Overview picture showing quartzites of the Ikalamavony Group in the hanging wall, and orthogneisses of the Antananarivo Block in the footwall of the contact (shown by *dashed line*). The *arrow* indicates site of outcrop (OC) of Fig. 3a. Image is taken west of Ankaramena viewing towards the north

(a) fast grain boundary migration (GBM) recrystallization in pure quartz aggregates, which is assumed to be related to temperatures in excess of ~500 °C (e.g. Stipp et al. 2002a), resulting in grain sizes of several mm (Fig. 4a);

(b) dominant [c] direction of grain-internal slip, resulting in CPO with c-axes maxima close to the stretching lineation (Mainprice et al. 1986; Fig. 4a); temperatures assumed for the activation of [c] directed slip are ~650–750 °C (Mainprice et al. 1986).

(c) Occurrence of chessboard patterns indicating prism- and basal-plane subgrain boundaries in quartz, which is regarded as a geothermobarometer (Kruhl 1996). P–T conditions for coeval activation of both slip systems are restricted to the quartz α–β transition (Kruhl 1996; Fig. 4b), e.g. at pressures of ~M1 (0.8 GPa) minimum temperatures are ~770 °C, at pressures of ~M2 (0.4 GPa) minimum temperatures are ~660 °C;

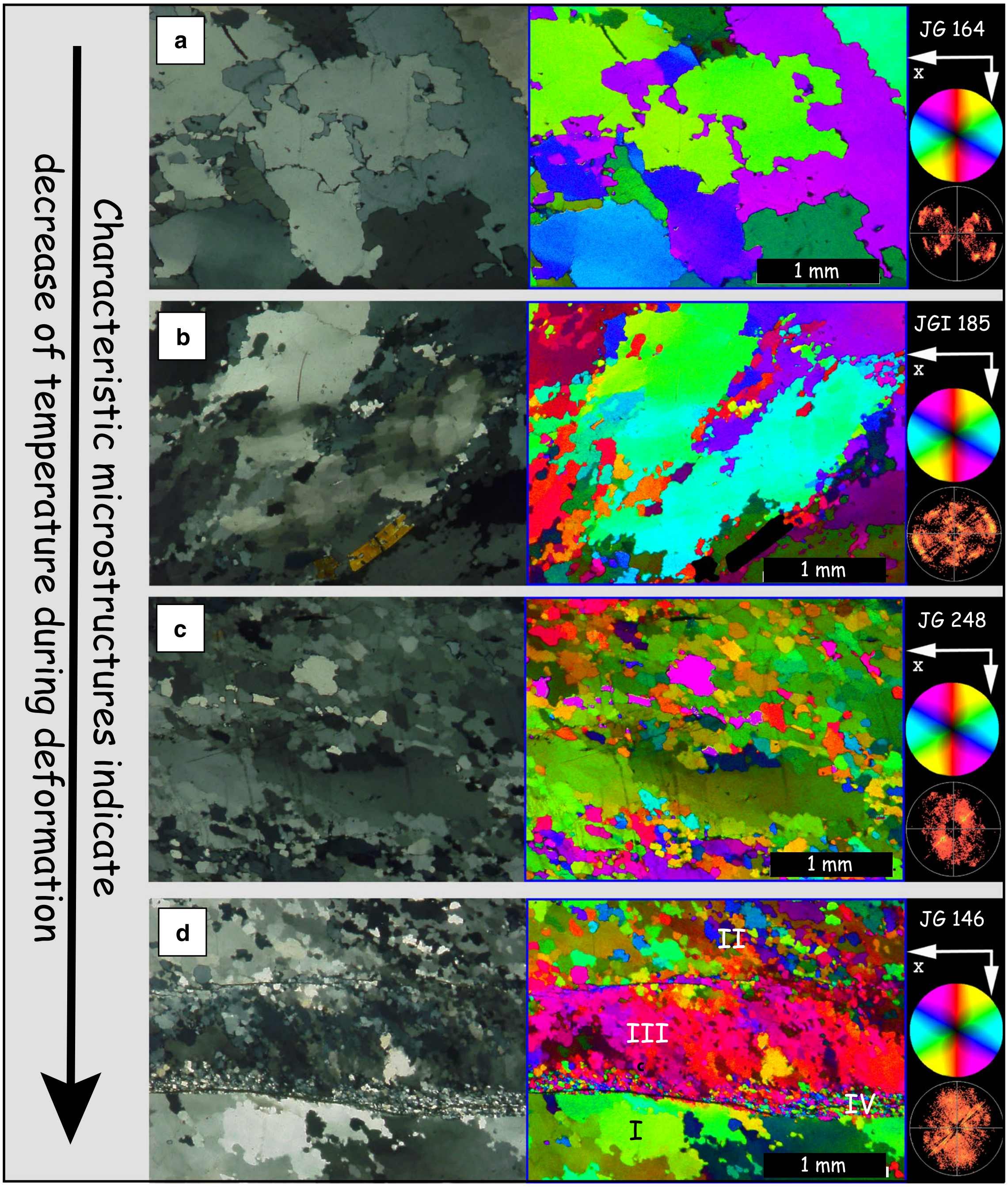
(d) stability of sillimanite syn- and post-dating shearing of the L-tectonite (sample JGI 110) as indicated by late to post-deformation dilatant microcracks perpendicular to the stretching direction (Fig. 5) that are healed with fine-grained sillimanite.

Low-temperature (LT) microstructures are mainly characterized by ribbon grains and dynamically recrystallized quartz aggregates showing subgrain rotation (SGR) recrystallization (~400–500 °C; Stipp et al. 2002a) and grain boundary bulging (BLG, ~280–400 °C; Stipp et al. 2002a), resulting in distinctively smaller grain sizes (µm to mm) than HT (>500 °C) fabrics (e.g. Stipp et al. 2002a) and localization of strain (Fig. 4c, d).

In the following, microstructures associated with HT conditions during deformation will be considered to be generated at temperatures in excess of ~500 °C. Temperatures might have reached much higher conditions, e.g. 770 °C and 0.8 GPa for the quartz α–β transition (which would be consistent with the regional M1 metamorphic event). In any case, 500 °C is a minimum estimate required for the HT microstructures. Consequently, microstructures related to temperatures less than ~500 °C will be considered as LT microstructures.

Inferring sense of shear from thin section

The extraction of shear sense will be based on crystallographic preferred orientation (CPO) and shape-preferred orientation (SPO). CPO from quartz grains deformed at the presumably highest temperatures (i.e. sample JG 164, S_{Andreaba} 325/20, L_{Andreaba} 038/02) forms a slightly skewed type I c-axis cross girdle (e.g. Passchier and Trouw 2005) with an opening angle of ~100° pointing towards top-to-the-southwest thrusting (Fig. 4a). Because of the large grain size, however, statistics to confirm this CPO are not well constrained, and in our opinion this sense of shear has be treated with caution. However, both the high opening angle and the microstructure indicating fast grain boundary migration recrystallization (Stipp et al. 2002a) support high temperatures during deformation (Fig. 4a). A second sample featuring both HT and LT microstructures (sample JGI 185, S_{Andreaba} 238/50, L_{Andreaba} 280/30) features an SPO of large (several mm) grains indicating a top-to-the-east thrusting sense of shear (Fig. 4b). The texture indicated by the *c*-axis pole figure corresponds to the bulk c-axis orientation of finite deformation, thus mingling HT and LT CPOs. However, c-axis colour coding of the large grains (green colour) indicates c-axis maxima at the periphery close to the L_{Andreaba} and thus corroborates top-to-the-east thrusting at high temperatures (Fig. 4b). Red and pink colours of smaller recrystallized grain sizes (0.1–0.5 mm) surrounding the large grains point to an opposite sense of shear (i.e. top-to-the-west extension) re-activating the same shear plane (S_{Andreaba}) at significantly lower temperatures as indicated by the dominant subgrain rotation recrystallization mechanism (e.g. Stipp et al. 2002a; Fig. 4b). A similar kinematic evolution can be deduced from sample JG 248 (S_{Andreaba} 250/32, L_{Andreaba} 220/15) where not fully overprinted domains of large ribbon grains indicate an inherited strong CPO by the green colour coding (Fig. 4c). Again, the c-axis pole figure mingles HT and LT quartz fabrics, but the green colour points towards c-axis maxima at a low angle to the L_{Andreaba}, thus suggesting again top-to-the-northeast thrusting followed by reactivation of the same shear plane S_{Andreaba} at significantly lower temperatures as indicated by the subgrain rotation recrystallization dominated microstructure of the newly recrystallized grains (Fig. 4c). Those newly crystallized grains with diameters between 0.1 and 0.4 mm feature an asymmetrically inclined SPO indicating top-to-the-westsouthwest extension. The inherited HT CPO has not been completely overprinted but pink, red and orange colours of newly formed grains suggest the onset of the formation of a new CPO that was "frozen in" before steady-state deformation was reached (Fig. 4c). Also, sample JG 146 (S_{Andreaba} 340/50, L_{Andreaba} 022/32) shows very similar evolution when compared to samples JGI 185 and JG 248. However, in this thin section four different domains indicating different conditions, mechanisms and kinematics during deformation can be clearly distinguished: (I) large quartz grains (~1 mm in diameter) feature a strong CPO (green colours) suggesting a c-axis maxima again close to the L_{Andreaba} pointing towards top-to-the-southwest thrusting (see orientation of the S_{Andreaba} and L_{Andreaba}). (II) Intermediate grain sizes (0.1–0.4 mm) of quartz featuring subgrain rotation recrystallization microstructures develop—if at all—a very weak SPO suggesting top-to-the-northeast extensional sense of shear. A CPO has not been commonly established, but the HT CPO of domain (I) has been clearly altered (as visible by the green, blue, pink and orange colours in the colour-coded Fig. 4d). (III) In between domains (I) and (II), a shear zone of about one mm thickness has developed parallel to the S_{Andreaba} featuring dominantly subgrain rotation and grain boundary bulging recrystallization (Stipp et al. 2002a). Both a strong CPO (indicated by red and pink colours in colour-coded image) and SPO point towards top-to-the-northeast extensional sense of shear. While the upper boundary to the intermediate-sized grains of domain (II) is slightly gradual, the lower boundary to the large grains of domain (I) is sharp and truncates HT quartz grains (Fig. 4c). Domain (IV) corresponds to very small-sized quartz grains (<50 µm) that are aligned along the lower boundary of the shear zone of domain (III) and include thin strongly aligned biotite crystals (Fig. 4c).



No SPO or CPO is developed within quartz grains in this domain, which may indicate a change from predominantly dislocation creep to viscous granular flow (Fig. 4c). The absolute directions of shear in this sample have to be treated with caution since they may have been rotated as this sample derived from the western limb of a Ihosy phase open fold (Fig. 2a).

A more comprehensive kinematic evolution could be derived by integrating more diverse mylonitic lithologies (rather than basically concentrating on quartz mylonites) that have different potential to record a more widespread inventory of shear fabrics such as mica fish, rotated clasts, intrafolial folds or flanking structures (e.g. Wennberg 1996; ten Grotenhuis et al. 2002, 2003; Mukherjee and Koyi 2009, 2010; Mukherjee 2011, 2013, 2014; Mukherjee et al. 2015).

Summary of results from microstructures

Most of the microstructures do not point to a common kinematic evolution: very often HT microstructures or relicts of HT deformation (e.g. core and mantle structures around

◂**Fig. 4** Quartz microstructures from different locations (Fig. 2) along the central southern segment of the contact between the Antananarivo Block and the SW Madagascar Block. *Left* and *right columns* show the same thin section area as (*left*) microphotographs with crossed polarizers, and (*right*) colour-coded Qtz-c-axis orientation images (AVA, "Achsenverteilungsanalyse" calculated using GeoVision software; Fueten and Goodchild 2001). Colour coding for *C*-axis orientation is given by the stereographic look-up table, scale of *left-hand pictures* equals the *scale bar* present in the AVA image. Pole figs. of quartz *c*-axis. All thin sections are cut perpendicular to the foliation (*short arrow*) and parallel to the stretching lineation (indicated by *x* on the *long arrow*). Sample numbers are provided in the *upper right corner*. From **a–d**, microstructures preserved are indicative of decreasing temperature during deformation. **a** Large grain sizes of several millimetres and fast grain boundary migration recrystallization (GBM) as the dominant dynamic recrystallization mechanism require high temperatures (>500 °C, according to Stipp et al. 2002a). Although no obvious shape-preferred orientation (SPO) is recognized, the grains are characterized by internally homogeneous crystallographic orientation (hardly any subgrains or undulatory extinction) and a considerable crystallographic preferred orientation (CPO). (b) Large quartz clasts, featuring chessboard patterns (temperatures and pressures around quartz α–β transition, Kruhl 1996), are forming an oblique foliation to S_{main}. These grains are preserved as core and mantle structures with small subgrains forming preferentially along grain boundaries. These small grains (<100 µm), evolved by subgrain rotation recrystallization, probably also by bulging recrystallization (BLG) suggest deformational overprint of the HT fabrics at intermediate to low temperatures (~400–500 °C and lower, according to Stipp et al. 2002a). Their crystallographic orientation is different to that preserved in the HT clasts. **c** Large ribbon grains are rarely preserved, more or less equigranular quartz grains which formed by subgrain rotation recrystallization dominate the fabric. The orientation of the well-developed SPO is contrary to the preserved CPO. **d** In between two discrete µm scale shear zones a dominant shear zone (SZ) of ~1 mm thickness discordantly cuts HT fabrics preserved below and intermediate temperature fabrics preserved above the SZ. Second-phase minerals (biotite) are present within the bordering SZs. Inside the SZ the orientation of a strong SPO indicates a similar shearing direction as the strong CPO, but again contrary oriented to the CPO preserved within the HT fabrics. There is no strong CPO developed within the smallest (<50 µm) grains of the lower boundary of the SZ. This might be an expression of a change from dominantly dislocation creep to viscous granular flow. *I*, *II*, *III* and *IV* correspond to different microstructural domains, see text for explanation

large quartz clasts with internal chessboard patterns; Fig. 4b) feature (partial) overprint at lower temperatures and/or higher strain rates (Fig. 4c, d). Usually the CPO preserved in quartz, associated with HT microstructures, is distinctly different compared to CPOs of newly recrystallized quartz overprinting the fabric (see above, Fig. 4). Although a shape-preferred orientation (SPO) is rarely observed in samples preserving HT microstructures, a top-to-the-east movement could be inferred from an SPO in one HT quartz mylonite sample north of Ankaramena (i.e. sample JGI 185, Figs. 3a, 4b).

On the other hand, most microstructures related to a LT overprint develop SPOs or localized mm-scale shear zones, which very often indicate different kinematics than those indicated by preserved crystallographic preferred orientations (CPO) of HT fabrics (Fig. 4c, d). Moreover, several samples feature partial static annealing camouflaging deformation-related microstructures and others show variable influence of second-phase effects, comparable to those observed in carbonate or harzburgitic mylonites, such as pinning of grain boundaries and thus influencing the effective grain size (e.g. Ebert et al. 2007; Herwegh et al. 2011; Linckens et al. 2011, 2015). Substantial strain localization from outcrop to grain scale can be observed in the L-tectonite (sample JGI 110) as a function of the mineral assemblage: highly stretched sillimanite domains localize the strain, whereas Qtz-Fsp-Bt domains are remarkably less deformed (Figs. 3c, d, 4).

In summary, the microstructures along this part of the contact indicate that (a) deformation is inhomogeneously distributed across a relatively wide sheared sequence including metasediments, gneisses and granitoids, (b) some deformation is coeval with high temperatures, which is also recorded in the surrounding metamorphic rocks (Figs. 4a, b, 5), (c) part of the high-temperature fabrics indicate top-to-the-east/northeast sense of shear (Figs. 3a–c, 4b), (d) re-activation and incomplete overprinting of HT fabrics occurred at lower temperatures (or, although unlikely, at dramatically higher strain rates), and (e) deformation is partly associated with a change in kinematics indicated by a different sense of shear (e.g. Fig. 4d). Therefore, we regard the interpretation of regional kinematics for this contact and identification of sense of shear only on a microstructural basis as challenging, since the spatial and temporal context of the microstructures remains speculative, both between and within individual samples. However, there is no doubt that deformation occurred along the contact during both high- and low-temperature conditions.

Discussion

The data presented above indicate that deformation along the investigated part of the contact occurred repeatedly under significantly varying conditions and kinematics, including HT and LT fabrics. Since LT fabrics certainly re-activate the same shear plane that already accommodated HT shearing, the distinction of LT and HT fabrics in the field is challenging—macroscopic observations are entirely in line with contractional tectonics, i.e. thrusting. Nevertheless, by integrating structural and microstructural observations with geochronological data and metamorphic conditions associated with deformation from the literature, several constraints regarding the significance, timing and kinematics of deformation along this contact can be deduced.

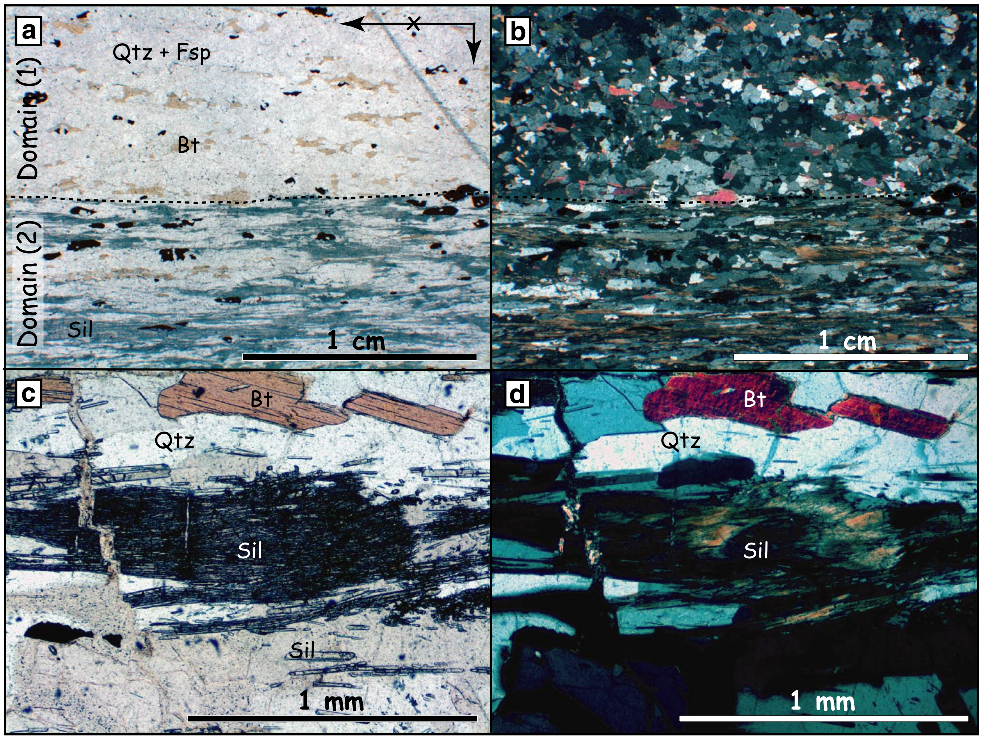


Fig. 5 Microphotographs of L-tectonite (sample JGI 110, Fig. 2). Overview image under **a** plane and **b** crossed polarized light. Domain (*1*) consists of moderately flattened quartz, feldspar and biotite, which traces the foliation and is distinctly less deformed than domain (*2*). The latter domain consists mainly of fibrous and blocky sillimanite defining the foliation, less quartz, feldspar, biotite and an opaque phase. Domain (*2*) corresponds to the highly stretched and moderately flattened mylonite rods (Fig. 3c, d). **c**, **d** Show a close-up into domain (*2*) with fibrous sillimanite (fibrolite) and prismatic blocky sillimanite crystals parallel to the foliation. Dilatant microcracks perpendicular to the foliation pull the sillimanite apart; cracks are healed and partly filled by finest grained sillimanite needles. Thin section is cut perpendicular to the foliation (*short arrow*) and parallel to the stretching lineation (indicated by *x* on the *long arrow*)

Timing of deformation and relative displacement along the Itremo–Ikalamavony thrust

The region is dominated by an open folded, usually shallow W-dipping main foliation, which was produced by deformation associated with the Andreaba phase, confined to the period between ~550 and 520 Ma (Giese et al. 2011). This main foliation (S_{Andreaba}) is parallel to the contact/thrust between the SW Madagascar and Antananarivo Blocks and occurs concordantly within the rocks of the Ikalamavony Group in the hanging wall and the Antananarivo domain in the footwall. The sedimentary protoliths of the rocks from the Ikalamavony Group were either deposited on the basement of the Antananarivo Block and were only slightly displaced thereafter (hence representing a parautochthonous cover as proposed by De Waele et al. 2011 for the prolongation of this thrust north of the Itremo domain), or the Ikalamavony Group formed elsewhere (e.g. in an island arc setting as proposed by Boger et al. 2014) and collided and thrusted onto the Antananarivo domain before about 800 Ma predating the common intrusion of the Imorona–Itsindro magmatic suite (e.g. Tucker et al. 2014). Because of the exclusivity of the about 1 Ga old Dabolava intrusive suite to the Ikalamavony domain, the latter scenario appears more likely (Boger et al. 2014). So when did the (micro-)fabrics described in this study develop? Before, during or after the Andreaba phase?

Although an accretion of the Ikalamavony island arc to a continental margin must have resulted in severe deformation, no (micro-)structures can unequivocally be linked to this event due to the subsequent intense tectonic and metamorphic overprint with the pervasive structures developed during the Andreaba phase.

The Andreaba deformation and related foliations, lineations, folds and microstructures (see discussion below) could be the result of substantial crustal thickening, which might be due to nappe stacking of the western units of the SW Madagascar Block onto the Ikalamavony Group (Müller 2000), and the latter directly thrusting onto the Antananarivo Block in northern and southern Madagascar. In central Madagascar, the Ikalamavony Group overthrusts the Itremo Group and both overthrust the Antananarivo Block (Tucker et al. 2007). In this case, thrusting is coeval with magmatism and high-grade metamorphism, involving partial melting. The emplacement of high-alkaline (stratoid) granites in Neoproterozoic times is reported for central northern Madagascar (e.g. Nédélec et al. 1994, 1995, Paquette and Nédélec 1998). Although slightly older in central northern Madagascar, stratoid granites as well as high-alkaline granites and syenites also intrude the SW Madagascar Block and the Antananarivo Block in central southern Madagascar (e.g. Delbos 1959; Bésairie 1969/1970; Giese et al. 2011). The production of high-alkaline rocks and their emplacement as foliation parallel sills are suggested to be related to and coeval with extensional tectonics in Madagascar during the late Neoproterozoic (e.g. Nédélec et al. 1994, 1995; Paquette and Nédélec 1998). This inferred extension is suggested by Collins et al. (2000) and Collins (2006) to be contemporaneous with the extensional Betsileo shear zone with a top-to-the-west sense of shear, parallel to $S_{Andreaba}$. This suggestion, however, raises the question, why the Betsileo shear zone itself is not intruded by high-alkaline sills. Actively deformed rocks, e.g. in shear zones, are often assumed to have higher permeabilities and hence would provide preferred pathways for fluids and melts (e.g. Handy et al. 2001).

Relative age and timing of microstructures and CPOs within the Itremo–Ikalamavony thrust

HT microstructures

The Andreaba phase foliation dominates large regions in southern Madagascar (Martelat et al. 2000; their S1) and very often features a lineation, which is similarly oriented to the majority of those lineations associated with deformation along and across the contact between both blocks (Fig. 2). Deformation occurred under HT/HP granulite facies conditions (e.g. Martelat et al. 1999, 2000). This observation agrees with the stability of sillimanite during deformation (Figs. 3c, d, 5). However, regional deformation was restricted to a period of abundant granitic intrusions, migmatization and anatexis in the gneissic rocks of both the hanging wall and the footwall and ceased before the end of magmatic activity (Giese et al. 2011). Since large volumes of partially molten rock, characterized by a distinct weaker rheology (Handy et al. 2001) than the surrounded meta-quartzitic sequence, exist, there is no reason why deformation should localize within the quartzitic sequence at that time. The preservation of hardly annealed HT dynamic recrystallization microstructures in quartz suggests that development of the latter structures postdates the major thermal and magmatic event. This is supported by solid-state deformation of syn-tectonic magmatic rocks (e.g. similarly oriented lineations; Giese et al. 2011). One possible scenario could involve solid-state deformation being restricted to the very short time and temperature window, when the majority of the melt in rocks underlying the meta-quartzitic sequence crystallizes and excess fluid is released (Berger et al. 2008). The crystallized rocks immediately increase in strength and the rheological contrast to the quartzite inverts. The released fluid could diffuse into the metasedimentary sequence, probably preferentially along more permeable layers, and thus lowering the differential stresses resulting in localized deformation re-activating the mechanically contrasting contact. This hypothesis requires further integration of more detailed process-oriented studies.

Re-activation and later incomplete overprint of HT microstructures (Fig. 4) impede the extraction of reliable information about kinematics and conditions of deformation from the microstructures, since it is doubtful that steady-state fabrics (e.g. Herwegh and Handy 1996, Herwegh et al. 1997), which are essential for this application, are preserved. In any case, (micro-) structures associated with HT shear deformation producing the ~W to ~SW and ~NE plunging mineral and/or stretching lineation during the Andreaba phase predate later folding by N–S-trending Ihosy phase folds since they are themselves folded in several places (Fig. 2). However, some of the Ihosy phase folds are overturned, and the east vergence coupled with duplication of the sequence suggests again eastward transport of the Ikalamavony Group.

Central southern Madagascar experienced renewed HT/MP granulite facies metamorphism (M2) with partial melting in the middle crust and cordierite-producing mineral assemblages (e.g. Kröner et al. 1996; Martelat et al. 1999; Berger et al. 2006) probably post-dating the Ihosy phase (Giese et al. 2011). If, however, HT microstructures are related to the c. 550–520 Ma Andreaba phase, then it is strange that the dynamic recrystallization microstructures of quartz are preserved and did not completely anneal since the sequence remained under HT granulite facies conditions of M2—lasting until at least 450 Ma—thus post-dating cessation of the Andreaba phase until at least 450 Ma, the proposed cessation of the M2 HT/MP granulite facies metamorphism (Giese et al. 2011). One reason for this lack of overprint could be related to the observation that retrograde cordierite-bearing mineral assemblages (M2 overprint) seems to be predominantly affecting intensely re-activated and/or newly developed structures directly associated with the Ihosy phase (high-strain zones, axial planar foliations, discrete shear zones, e.g. Ihosy quarry, Berger et al. 2006; Ikalamavony gneiss west of Ankaramena, Giese et al. 2011). Along and across the tectonic contact, the present-day structures are dominantly preserving those associated with the Andreaba phase, only being openly folded by the Ihosy phase but without development of an axial planar foliation. To conclude, the HT microstructures are confined to represent deformation associated with a late stage of the

Andreaba phase between 550 and 520 Ma based on overprinting and intrusive relationships. This observation contradicts the proposed right-lateral re-activation of the contact related to this event (Gregoire 1999).

LT microstructures

The (micro-)structures related to LT deformation post-date the Ihosy phase (550–520 Ma), since metamorphic conditions associated with the latter phase are still at high temperatures (Giese et al. 2011). According to Fitzsimons (2016), a change from contractional to extensional tectonics occurred at ~530 Ma when thickened orogenic crust forming a Tibetan-style plateau collapsed. This age is roughly in agreement with the proposed extensional movement post-dating 550 Ma intrusive rocks along the Betsileo shear zone of Collins et al. (2000) forming part of the Itremo–Ikalamavony thrust. The post-orogenic extensional collapse can also be found in many other parts of the former Pan-African orogenic belts and prevailed until 490 Ma in Madagascar (DeWit et al. 2001). Berger et al. (2006), however, identified cordierite-producing events in central southern Madagascar (Ihosy quarry) as young as ~440 Ma. This indicates that metamorphic conditions remained high grade during a prolonged period in the Palaeozoic, even after major ductile deformation ceased. The latter authors suggested minor, localized re-activation of Ihosy phase shear zones to produce the renewed cordierite (and monazite) growths. In this context, also the boundary between the SW Madagascar Block and the Antananarivo Block could have been re-activated. This renewed, more localized deformation is likely to document early stages of the final exhumation of the presently exposed basement to the surface—that may have been initiated by post-orogenic extensional collapse, since in this area titanite fission-track ages of 430–330 Ma (Emmel et al. 2004, 2006) and apatite fission-track ages of 380–170 Ma (Emmel et al. 2004, 2006; Seward et al. 2004) provide a tight timeframe for (rapid) cooling from amphibolite/granulite facies (M2; 730–650 °C, e.g. Markl et al. 2000) to fission-track retention temperatures (titanite partial annealing zone (PAZ) 310–265 °C, Coyle and Wagner 1998; apatite PAZ ~120–60 °C, Laslett et al. 1987; Corrigan 1993). Exhumation was previously proposed to be assisted by movement along near-vertical shear zones (Martelat et al. 2000; Emmel et al. 2008) and is possibly also accompanied by extensional re-activation of shallowly dipping segments of the contact between the SW Madagascar Block and the Antananarivo Block in Phanerozoic times.

Summary of structural evolution of the Itremo–Ikalamavony thrust

Based on the discussion above and the data presented, we suggest the following evolution of the central southern part of the contact between the SW Madagascar Block and the Antananarivo Block:

- The Ikalamavony Group and the Antananarivo Block were juxtaposed between ~900 and 800 Ma based on intrusion ages of Imorona–Itsindro and Dabolava magmatic suites, respectively (Tucker et al. 2014). Due to the strong structural and metamorphic overprint during subsequent deformation phases, structures related to this event cannot unequivocally be identified (see also discussion by Giese et al. 2011).
- Major (re-)activation of the contact occurred during late stages of the Andreaba phase (550–520 Ma), but post-dating the main crystallization of the magmatic intrusions and migmatites. The HT microstructures observed in quartz mylonites, as well as the stretching lineation produced in the PT field of sillimanite stability, are likely to be associated with this (re-)activation. The sense of shear is not totally resolved; however, the kinematic indicators observed suggest predominantly top-to-the-east/northeast movement. This inference is in agreement with the eastward thrusting of the nappe stack in central Madagascar as proposed by Tucker et al. (2007, 2011b, 2014). We interpret the contact between the Antananarivo Block and the SW Madagascar Block in central southern Madagascar to represent the southern continuation. Since the Itremo Group is replaced by the Ikalamavony Group along strike to the northwest and south, the contact with the Antananarivo Block is a composite and we refer to it as the Itremo–Ikalamavony thrust.
- ~E–W shortening associated with the Ihosy phase resulted in folding of the contact together with S_{Andreaba} by predominantly N–S-trending upright folds of regional scale. However, a major re-activation of contact-parallel shear zones during folding is not likely. The end of deformation of the Ihosy phase is indicated by late-stage leucosomes, crosscutting the Ihosy phase foliation in central southern Madagascar at ~530–515 Ma (Giese et al. 2011; Schreurs et al. 2010).
- LT extensional re-activation affects parts of the contact, post-dating the HT evolution of the crustal section (post-dating M2). This LT overprint is likely to be associated with Phanerozoic exhumation of this mid-crustal section that may have been initiated by post-orogenic extensional collapse (DeWit et al. 2001; Fitzsimons 2016).

Implications for Gondwana correlations

The understanding of the kinematic evolution of this contact and its geological significance are important since this contact can be traced all the way to the east coast and therefore might be a suitable structure for tight-fit Gondwana correlations between Madagascar and southern India. In southern India, the Karur–Kamban–Painavu–Trichur shear zone (KKPT, Ghosh et al. 2004) separates predominantly magmatic and metamagmatic rocks to the north of the shear zone from a paragneiss-dominated region to the south (Ghosh et al. 2004). According to Ghosh et al. (2004), the shape of the KKPT shear zone reflects a late refolding. Collins and Pisarevsky (2005) point to similarities of the Archaean rocks of the Madurai Block north of the KKPT shear zone, with those of the Antananarivo Block as part of the proposed former micro-continent Azania (Collins and Pisarevsky 2005), and Boger et al. (2014 and references therein) correlate the rocks of the Ikalamavony domain with the rocks in India south of the KKPT based on the unique Dabolava suite age constraints. Since in both cases structurally overlying metasediments and paragneisses of Proterozoic age are separated from the Archaean rocks by a shallow dipping, refolded ductile shear zone, we suggest a link between the Itremo–Ikalamavony thrust in Madagascar with the KKPT shear zone in southern India. This tentative link, however, needs further confirmations; Tucker et al. (2014), for example, trace the contact between the Ikalamavony and Antananarivo domains into the Palghat-Cauvery (vertical) high-strain zone in southern India based on similarities in structural evolution.

Conclusions

To unravel the geological evolution of shear zones requires a thorough integration of different investigation methods, concerning the various aspects that contribute to the modification of the shear zone in space and time. Therefore, a solid structural investigation of the shear zone and the surrounding geology is essential and serves as a basis for any interpretation. Integrating microstructures serves as a useful tool providing detailed information, which when integrated into a larger framework, contribute to a better understanding of the full geological evolution of the shear zone. However, interpretations regarding the significance of individual observed microstructures with regard to crustal-scale processes and the timing of those are crucial. Under- or overestimating the significance of the microstructures preserved might result in considerably different interpretations.

In this study, we were able to integrate and constrain the structural evolution of a complex tectonic contact in central southern Madagascar into the framework of distinct geodynamic scenarios (including thrusting and crustal thickening, regional metamorphism, magmatism, crustal shortening, exhumation and extensional tectonics) under consideration of changes in rheological properties in space and time and the preservation potential of structures and microstructures. Although quartz mylonites are very well suited to determine kinematics within sheared rock, to decipher a detailed polyphase evolution of a shear zone, different mylonitic lithologies should be sampled for more diverse shear sense indicators (e.g. mica fish, rotated clasts, intrafolial folds or flanking structures). This would be a good future addition to further unravel the history of this contact and should be critically considered for comparable studies of different shear zones.

We propose that during final amalgamation of this crustal segment of Gondwana the units of the SW Madagascar Block were tectonically displaced towards the E to NE on the Antananarivo Block prior to the Andreaba phase. Continuous ~E–W shortening during the Andreaba phase reactivated the Itremo–Ikalamavony thrust, and subsequently both blocks and the contact were affected by ~N–S-trending large-scale Ihosy phase folds. Both Andreaba and Ihosy deformation phases were continuous in time and were restricted to the narrow time window between ~550 and 520 Ma and coeval with amphibolite/granulite facies metamorphism and widespread granitic magmatism (Giese et al. 2011). The contact was repeatedly reactivated post-dating major ductile deformation probably related to an initial extensional collapse and phanerozoic exhumation of the crustal segment. This major contact, which can be traced over 750 km across Madagascar, may tentatively be linked to the KKPT SZ in southern India during Gondwanian times.

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