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Strain pattern and late Precambrian deformation history in southern Madagascar

Jean-Emmanuel Martelat [a,b,*], Jean-Marc Lardeaux [a], Christian Nicollet [b], Raymond Rakotondrazafy [c]

[a] *Laboratoire de Dynamique de la Lithosphère, Université Claude Bernard, UMR 5570, 43 Bd du 11 Novembre, 69622 Villeurbanne, France*

[b] *Département de Géologie, Université Blaise Pascal, UMR 6224, 5 Rue Kessler, 63038 Clermont Ferrand, France*

[c] *Laboratoire de Pétrologie, Université d'Antananarivo, BP 906, Antananarivo, 101, Madagascar*

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Abstract

This paper examines the thermo-mechanical evolution of the lower crust, in Precambrian times, with an example from southern Madagascar. The finite strain pattern is derived from the study of satellite images complemented by field structural analysis. The finite geometry reflects the superposition of two distinct finite strain patterns, D1 and D2. The geodynamic significance of the D1 event remains unclear. However, the D2 finite strain pattern is partitioned between a network of kilometric vertical shear zones and folded domains, and is consistent in the Mozambique belt with east–west shortening in a transpressive regime.The metamorphic conditions associated with the two finite strain fields show that rather uniform, high temperatures were ubiquitous in southern Madagascar. By contrast, regional pressure differences were controlled primarily by the major shear zones (D2). Chronological constraints obtained from monazite (U–Th–Pb electron microprobe dating) indicate ages of 590–530 Ma for the D1 event and 530–500 Ma for the D2 event. The two distinct strain fields D1 and D2 are continuous in time.The late pan-African transpressional tectonic regimes, contemporaneous with granulitic facies metamorphic conditions, resulted in strong strain partitioning between shear zones and folded domains. The D2 transpression was efficient and pervasive enough to exhume pieces of the lower crust. The amount of exhumation was controlled by the D2 strain gradient (amount of pure shear with respect to simple shear).

Keywords: Granulite; Pan-African; Shear zones; Strain partitioning; Transpression

1. Introduction

Knowledge of both structure and petrology of the continental crust is of fundamental importance in the understanding of Earth dynamics. Together with geophysical investigations and ex-

* Present address: Laboratoire Pétrologie et Tectonique, UFR Sciences Terre, Université Claude Bernard Lyon I, 43 Bd du 11 Novembre, 69622 Villeurbanne cedex, France. Fax: + 33-04-72-44-85-93.

E-mail address: jemartel@ens-lyon.fr (J.-E. Martelat)

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perimental studies, the study of deeply eroded Precambrian terrains offers a unique opportunity to investigate significant volumes of intermediate to low continental crust [see Harley (1992) for a general discussion]. The Precambrian belt of Madagascar represents a more-or-less continuous section in granulites and high-grade amphibolites with exceptionally diversified and preserved mineralogies (Moine et al., 1985; Nicollet, 1988, Nicollet, 1990). Consequently, southern Madagascar offers an excellent opportunity to establish the deformation history of a significant piece of lower crust (> 100 000 km^2). We have established the bulk finite strain pattern at the continental scale using satellite imaging complemented by field observations. In the following sections we discuss the finite strain pattern in relation to the kinematics, the P–T evolution and age distribution in southern Madagascar. This analysis allows us to examine the role of transpression tectonics for genesis and exhumation of high-grade metamorphic terrains.

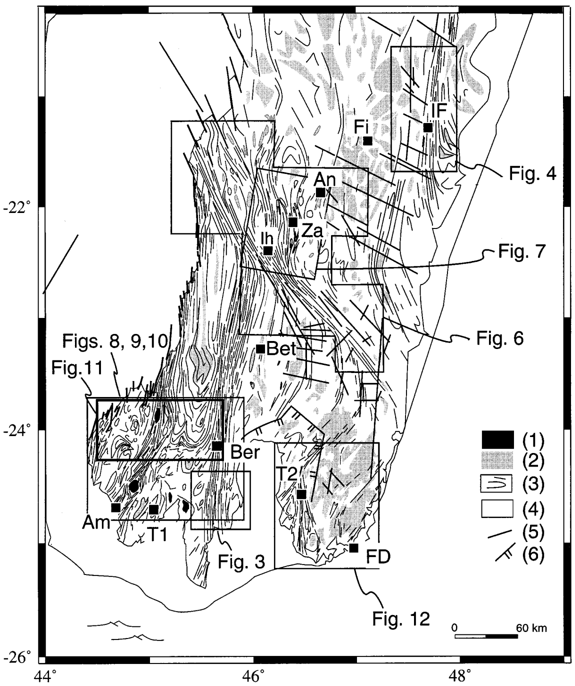

Fig. 1. Geological map in southern Madagascar with major tectonic features drawn from satellite images (Martelat et al., 1997; Martelat, 1998) and completed after Besairie (1970a,b). (1) Anorthositic bodies, (2) granitoids, (3) major foliation trends developed in migmatites and rocks metamorphosed under granulite and high amphibolite facies, (4) post-Cambrian sediments and volcanic rocks, (5) major faults, (6) post-Cambrian caldera. Black squares: towns of Ampanihy (Am), Ankaramena (An), Betroka (Bet), Beraketa (Ber), Fianarantsoa (Fi), Ifanadiana (IF), Ihosy (Ih), Tranoroa (T1), Tranomaro (T2), Zazafotsy (Za). The locations of Figs. 3, 4, and 6–12 are indicated.

2. Geological setting

It is well established, particularly with the paleomagnetic constraints (Norton and Sclater, 1979; Powell et al., 1980) that Madagascar is located in the middle Gondwana and represents the eastern front of the Mozambique belt (Pallister, 1971; Shackleton, 1986; Stern, 1994). This belt is regarded as a collision belt developed between two previously separated continental plates, namely east (India, Antarctica, Australia) and west Gondwana (Africa and South America). However the orogenic evolution of the Mozambique Belt remains poorly understood. Particularly discussed are the precise duration and the timing of the pan-African tectonic events between 950 and 550 Ma (Shackleton, 1986). It is generally admitted that an early period of rifting (750–700 Ma) is followed by continental collision and tectonic escape in a Tibetan style during 700–640 Ma (Key et al., 1989; Stern, 1994). The evolution of the belt after 640 Ma, and especially the significance of the widespread granulitic metamorphism at around 550 Ma, remains unclear. This late pan-African history, well developed in southern Madagascar, is the aim of this paper.

Various lithologies have been mapped in southern Madagascar (Fig. 1), i.e. orthogneisses, paragneisses, marbles, granitoids, migmatites, metabasites and anorthosites (Besairie, 1970a,b). The regional distribution of lithologies evolves from granitoids and k-feldspar gneisses in the east to plagioclase bearing gneisses, metabasites and amphibolites, with local anorthosite massifs (Ashwal et al., 1998), to the west. These metamorphic rocks are overlain on the west coast by rift-related late Phanerozoic through Cenozoic sediments. A Phanerozoic volcanic massif is also preserved between the villages of Tranomaro and Beraketa. Whereas Archean and/or early Proterozoic ages have been depicted in northern Madagascar (Caen-Vachette, 1979; Guerrot et al., 1991; Nicollet et al., 1997; Tucker et al., 1999), in the southern part of the island a generalized pan-African syngranulitic tectonic evolution is now well established (580–530 Ma, Andriamarofahatra et al., 1990; Paquette et al., 1994; Kröner et al., 1996; Montel et al., 1996). In all the lithologies, an extensive granulite facies metamorphism (800 ± 50°C, Nicollet, 1990; Windley et al., 1994; Martelat et al., 1997; Martelat, 1998) is widespread, especially between 20 and 25° of southern latitude. In the following we discuss the tectonic context associated with this granulite facies metamorphism.

3. The bulk strain pattern

3.1. Methods

The deformation history and global kinematic evolution of the area were deduced from the study of the bulk finite strain pattern (Choukroune, 1987; Gapais et al., 1987). The latter was obtained through the detailed mapping of foliation and lineation trajectories in addition to observation of meso- to micro-scopic criteria of coaxial or non-coaxial strain (e.g. symmetry or asymmetry of shear bands, tails around porphyroclasts, folds). The integration of all these data, from metric to kilometric scales, allowed us to constrain the three-dimensional (3D) finite strain pattern and the displacement field at the scale of the whole of southern Madagascar.

A part of the finite geometry was depicted by satellite imaging. We used 20 scenes of SPOT data from the south of Madagascar (Fig. 2) comprising three spectral bands (0.5–0.59 μm; 0.61–0.68 μm; 0.79–0.89 μm) or one spectral band (0.51–0.73 μm), with 10 m or 20 m ground resolution respectively, taken from vertical views of 60 × 60 km^2 surfaces measured between June and September. False color compositions and black and white pictures were interpreted from isolated images or mosaics at a scale of 1/100 000 in a Laborde-type field projection on a Hayford ellipsoid. Common digital processing was carried out: linear stretching, filtering for edge and edge directional enhancement, intensity hue saturation restitution and principal component analysis (Martelat, 1998). This was done especially over the key domains where complex structures interfere with one another.

In the field, the major tectonic structures are bedding-parallel foliation planes, which are recognizable at a metric or centimetric scale. Field strike measurements, corrected for local and regional magnetic deviation (which could reach 20°; Cattala, 1961), fit within ± 5°, with linear trends mapped from SPOT images (Martelat et al., 1995; Martelat, 1998). The data indicate that the textures mapped on satellite images correspond to foliation planes on the field. Consequently, in southern Madagascar, high-quality SPOT data and the specific geologic parameters of this area (relief, soil, vegetation, hydrographic framework, contrasted lithologies, high tectonic transposition), allow a textural foliation map to be drawn at a crustal scale (Martelat et al., 1995, 1997).

3.2. Crustal strain partitioning

At the crustal scale, high strain is widespread and strain partitioning results in the juxtaposition of domains with complex elliptical and folded structures, so-called ‘dome-and-basin domains’ (Windley and Bridgwater, 1971), bounded by ductile shear zones (SZs). The dome-and-basin domains comprise closed shapes with axial ratios from 1:3 to 1:5. These geometries essentially correspond to folded structures (interference pattern geometries), more rarely to diapiric structures (localized granites or migmatitic bodies). The transition from dome-and-basin domains to ductile SZs is frequently characterized by a progressive increase in the axial ratios of the closed shapes by up to 1:20. Such a spatial evolution underlines a regional strain gradient (Fig. 3). Consequently, on satellite images, we observe localization of strain within regularly oriented kilometric SZs and areas of lower strain corresponding to dome-and-basin domains. These SZs are developed throughout the whole of southern Madagascar, where they form an anastomosing network.

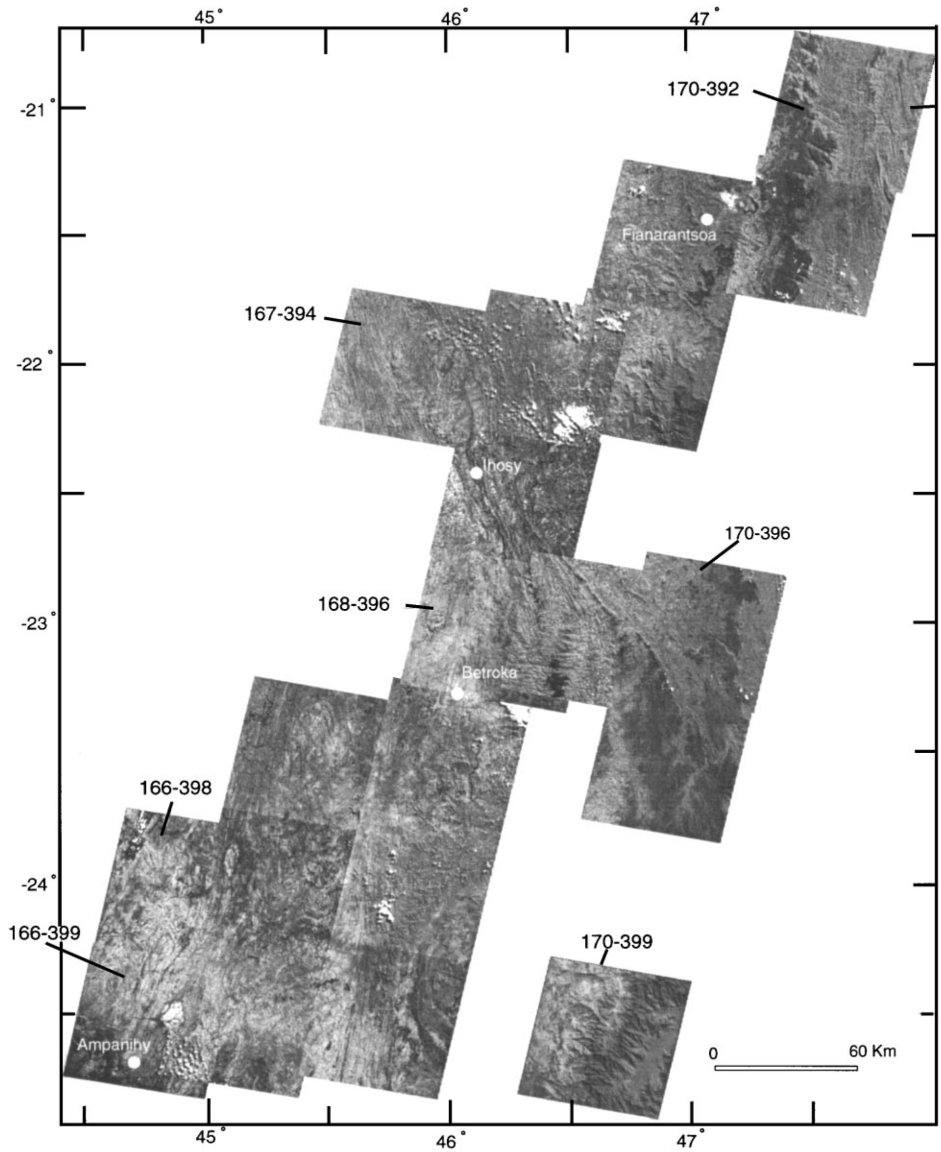

Fig. 2. Mosaic of 20 SPOT satellite images acquired on the southern part of Madagascar. Numbers correspond to SPOT classification (KJ). On the two images, KJ: 166-399 and 166-398, a regular structure oriented N 10 (Ampanihy SZ) is clearly visible and contains two anorthositic bodies (a rounded one, called Sariриaky, in white, is located just above the town of Ampanihy).

4. Strain pattern at different scales

In this section, we refer to four areas [(1) Ifanadiana–Fianarantsoa, (2) Ankaramena–Ihosy, (3) Ampanihy–Beraketa, (4) Tranomaro–Fort Dauphin], with good outcropping conditions and exposures, where geological mapping of finite strain elements, shear indicators and fold geometries reveal superimposed ductile deformations. In the following, we present the regional tectonic framework (SZ geometry) and the field data in the different sectors from north to south and south-east; the locations of all the figures are indicated in Fig. 1.

4.1. Sector 1: Ifanadiana–Fianarantsoa

4.1.1. SZ geometry

Eastwards of Fianarantsoa, a very regular zone, the Ifanadiana SZ, is oriented N 0 with a minimum length of 120 km (Figs. 2 and 4). Its width is poorly defined, as the vegetation and major brittle fault systems on the eastern coast are highly developed and disturb the satellite imaging of ductile structures. Reasonably, the width of this structure is between 10 and 20 km. The foliation trajectories, in agreement with the Besairie (1970b) map, allow us to observe another SZ to the south (Fig. 13) that corresponds to an anastomosed southern continuation of the Ifanadiana SZ.

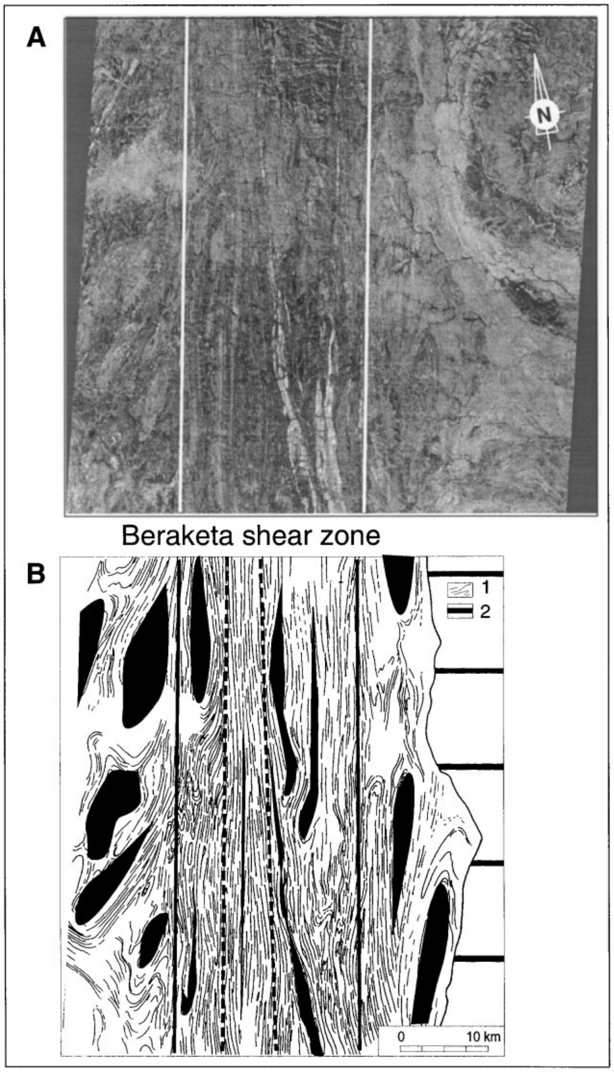

Fig. 3. Precise limits of a kilometric SZ depicted from the evolution of fold geometries (Beraketa SZ, KJ: 168–399, Fig. 2). (A) Satellite image; white lines correspond to the SZ limits. (B) Interpretive map obtained from (A): (1) transposed granulite; (2) post-Cambrian volcanic rocks. Black thin lines: foliation planes, thick black lines: SZ limits, dashed thick black lines: most transposed zone. The black areas underline the spatial evolution of the closed folded structures from preserved domains (dome-and-basin domains) to the most transposed ones.

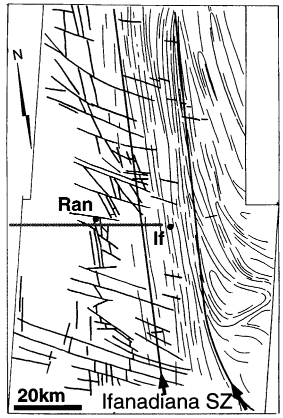

Fig. 4. Interpretative map obtained from satellite images (KJ: 170-392, 170-393, Fig. 2) showing detailed foliation trajectories (oriented N 0) and major brittle faults (oriented N 120, sector 1, Ifanadiana–Fianarantsoa in text). Arrows show the limits of the major SZ (N 0, Ifanadiana SZ). Location of east–west cross-section in Fig. 5 is indicated by the gray line from Ifanadiana (If) through Ranomafana (Ran).

4.1.2. Field data

The sector in Fig. 5 exhibits a very pronounced S1 compositional layering. The transposed layers comprise various lithologies with different thicknesses (1 mm to more than 20 m). The S1 foliation plane is dominantly N 0–N 20, shallowly dipping to the west; the dominant stretching lineation L1 is east–west (N 250–N 280) plunging to the west. These structural observations are compatible with the structural data derived from anisotropy of magnetic susceptibility from Grégoire and Nédélec (1997). The tectonites show L < S or S = L shape fabrics (Flinn, 1965). Sheetlike granites, with sharp contacts between surrounding para- or ortho-gneisses, are parallel to the S1 foliation plane. These syntectonic granites, 10–50 m or more thick by 50 km long, seem to be similar to the so-called 'stratoid granites' well known north of Antananarivo (Emberger, 1958; Nédélec et al., 1994, 1995). We observe, at the metric scale, recumbent folds (F1) with dominantly N 270 and more rarely N 0, subhorizontal to gently west-dipping hinge lines. Conjugate shear bands are developed in both *XZ* and *YZ* sections of the finite strain ellipsoid. These geometries indicate vertical shortening under mainly coaxial strain (D1). In some places, scarce asymmetric folds (F1) coherent with the orientations of intruding leucosomes in gneisses indicate a local non-coaxial strain with a systematic westwards normal displacement.

The S1 foliation can also be locally folded into kilometric open folds with subhorizontal axes (Fig. 5), yielding at map-scale the rounded shapes similar to the type I interference pattern observed in the dome-and-basin domains. These F2 folds with north–south axial planes are coherent with a global east–west shortening (D2). In some places, the first S1–L1 fabric is reworked by a new penetrative S2–L2 fabric. The new S2 foliation is vertical and L2 is subhorizontal, giving rise to the development of S = L tectonites (Flinn, 1965). These vertical domains reach 500 m wide by several kilometers long, but near Ifanadiana these structures are homogeneous at a kilometric scale (Figs. 4 and 5). This zone corresponds to the Ifanadiana SZ, depicted on the satellite images. Inside this SZ, some shear bands are conjugate at angles of 35–40° and numerous isoclinal folds show hinge lines varying from vertical to horizontal attitudes. These features (strong symmetry of the tectonic fabric and of strain indicators) are coherent with a strong east–west horizontal shortening (D2) under mainly coaxial strain.

4.2. Sector 2: Ankaramena–Ihosy

4.2.1. SZ geometry

Close to Ihosy village (Fig. 6) we find two main structural directions: one N 0 and the other N 140–150. The first is coherent with the geometry of the Beraketa SZ (northern termination), whereas the second is representative of the Bongolava–Ranotsara SZ (Bazot, 1976; Nicollet, 1988). The deflection of the foliation trajectories in the vicinity of the Bongolava–Ranotsara SZ looks like a crustal scale C/S structure (Berthé et al., 1979) indicating a sinistral sense of shear. Such a geometry is also well evidenced by the pattern of a small-scale SZ network (3–5 km wide, located northward from Ihosy, Fig. 5). The N 140-oriented Bongolava–Ranotsara SZ is badly defined to the east, where the density of N 140 faults increases, and to the west, where the structure interferes with smaller anastomosing SZs. Westwards, this structure disappears beneath the post-Cambrian sedimentary cover. From this limit, up to the Mozambique channel, numerous kilometric brittle faults are observed in the Cretaceous cover (Premoli, 1977) and are in the prolongation and parallel with ductile foliation plane

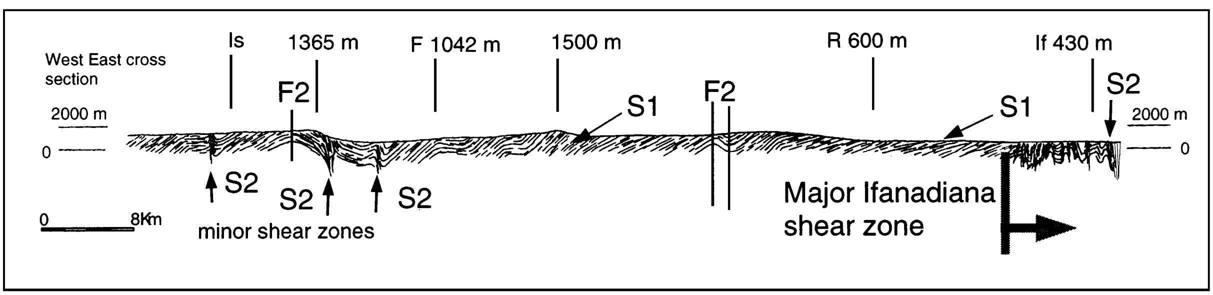

Fig. 5. East–west cross-section showing the two main foliation types (S1–S2) and open to upright F2 folds going inside the major or minor vertical SZs.

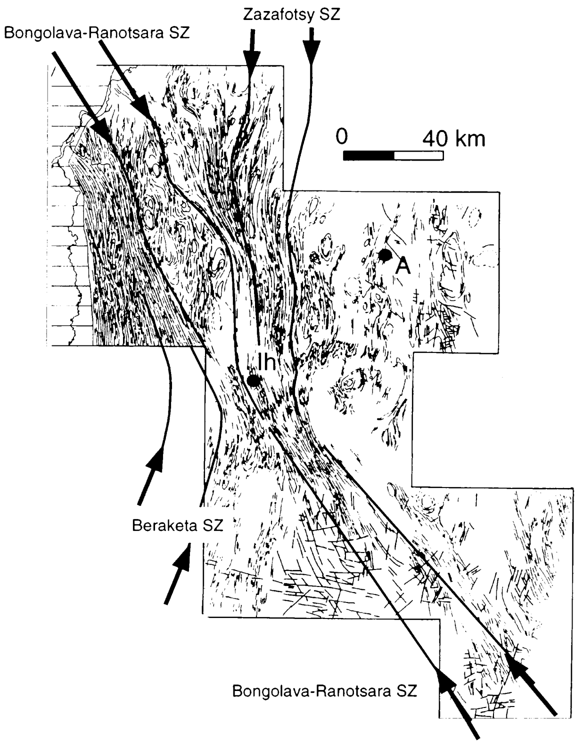

Fig. 6. Detailed foliation trajectories from Fig. 1 (sector 2, Ankaramena–Ihosy in text). Same legend as in Fig. 3. The heavy black lines define the limits of the Bongolava–Ranotsara SZ, Beraketa SZ, and Zazafotsy SZ. The density of brittle structures increases from north-east to south-east and disturb the interpretation of ductile structures.

(top left, Fig. 1). Brittle faults developed in this cover could be controlled by older tectonic structures at depth, such as Bongolava–Ranotsara SZ; we guess that the Bongolava–Ranotsara SZ reaches at least 350 km in length. Near Ihosy the width of this SZ is about 20 km, in other places the width of the Bongolava–Ranotsara SZ is more difficult to trace, because the SZ boundaries interfere with kilometric-scale normal faults and other SZs to the south and to the north.

As demonstrated in Fig. 6, north of Bongolava–Ranotsara, an N 0-trending SZ extends from Ihosy to at least 120 km northward. This structure is not as regular as the southwestern SZ, and its width is unclear (between 10 and 20 km), but this structure is clearly distinct from the Bongolava–Ranotsara SZ. We refer to this structure as the Zazafotsy SZ.

4.2.2. Field data

From Ankaramena to Zazafotsy (Fig. 7), we clearly follow the transposition from D1 geometries into D2 fabrics: 20 km west of Ankaramena, the regional subhorizontal S1 fabric is gently folded into kilometer-scale F2 open folds with subhorizontal hinge lines (synforms and antiforms). Near the boundary of the Zazafotsy SZ the folds become upright and unrooted (Fig. 7). The S1 plane (N 0–N 20) is transposed into a new penetrative N 170, subvertical S2 foliation plane, and the L1 east–west stretching lineation is progressively replaced by the L2 subhorizontal lineation.

Between the towns of Zazafotsy and Ihosy, vertical structures dominate. The F2 folds are upright and unrooted with folded, subhorizontal hinge lines. In this transect we cross two major SZs, the Bongolava–Ranotsara underlined by an N 140 vertical foliation plane (S2) and the Zazafotsy SZ underlined by N 170 vertical planes (S2). In these SZs the rocks are S–L tectonites (Flinn, 1965).

In the Zazafotsy SZ, despite the high strain underlined by the intense tectonic transposition, we never observe clear asymmetry of the strain markers and the development of conjugate shear bands, and isoclinal folds is compatible with bulk coaxial strain due to an east–west shortening. On the contrary, in the Bongolava–Ranotsara SZ the deflection at a kilometric scale of the foliation trajectories indicates a non-coaxial deformation coherent with a sinistral strike-slip motion. Moreover, in the field, numerous shear bands or drag folds are compatible with sinistral strike-slip. However, subordinate conjugate meter-scale SZs, as well as symmetric boudinage, are also observed and are compatible with a global shortening perpendicular to the walls of the SZ.

4.3. Sector 3: Ampanihy–Beraketa

4.3.1. SZ geometry

In the southern and southwestern part of Madagascar we can recognize three major SZs (Fig. 8): Ejeda, Ampanihy and Beraketa [also called Vorokafotra SZ by Rolin (1991)] that correspond to regular zones of 20 km width, oriented between N 0 and N 15. The length of the Beraketa SZ is at least 250 km (Fig. 1); its northern limit becomes parallel to the N 140 trending Bongolava–Ranotsara SZ (Fig. 6). The length of the Ejeda and Ampanihy SZs are unknown because their northern and southern extents lie beneath the post-Cambrian sedimentary cover. We note a clear discontinuity between the three SZs and the adjacent dome-and-basin domains.

4.3.2. Field data

In this part of the island the vertical S2 foliation planes consist regionally of a transposed lithological layering. Inside the Beraketa SZ, the S2 foliation plane is commonly oriented N 0, 90°. The stretching lineation L2 is subhorizontal, and only locally plunges 5–30° to the north (Figs. 9–11). Towards the west, in the adjacent dome-and-basin domain, the foliation planes strongly

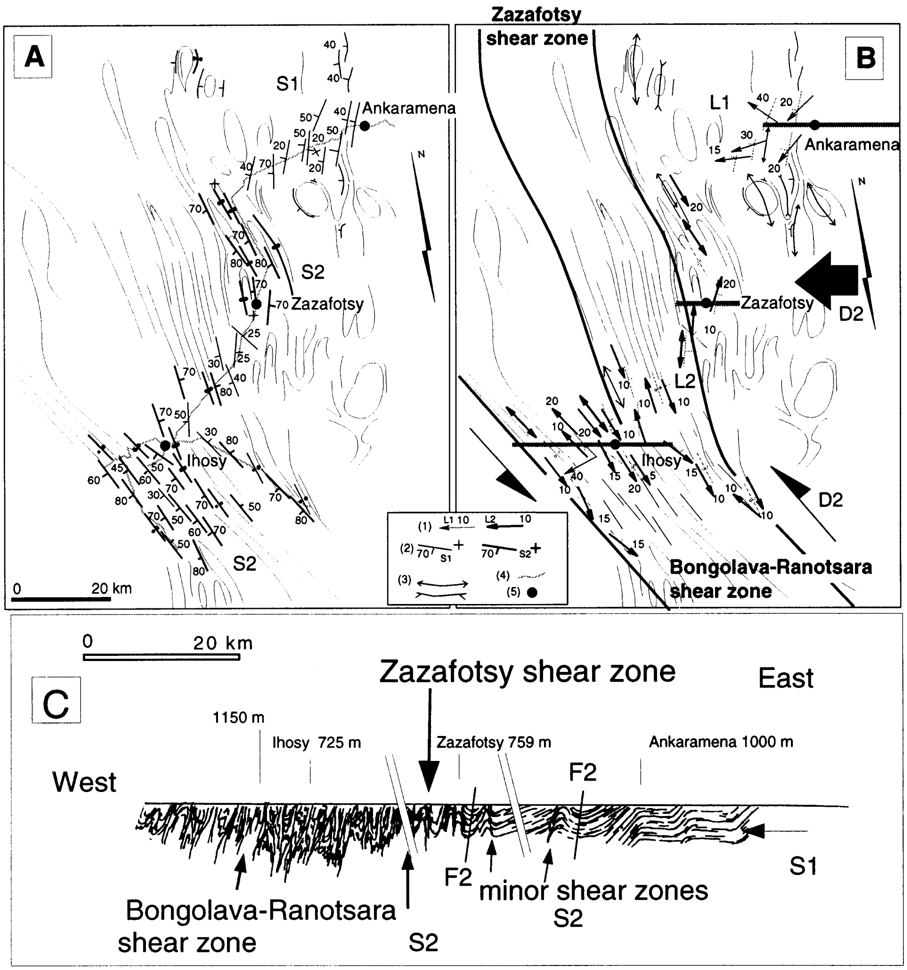

Fig. 7. Geometry from sector 2. (A) Foliation map and (B) lineation map: (1) stretching lineation L1 and L2, (2) foliation plane S1 and S2, (3) axial plane of synclines and anticlines, (4) roads, (5) villages. In this region two major SZs interfere: the Bongolava–Ranotsara SZ (N 140) and the Zazafotsy SZ (N 0). The limits of the SZ are indicated by heavy black lines. (C) Vertical east–west cross-section. The S1 foliation plane is transposed into the new S2 foliation plane. The location of the three cross-sections is given by the thick gray lines passing through the towns of Ankaramena, Zazafotsy and Ihosy.

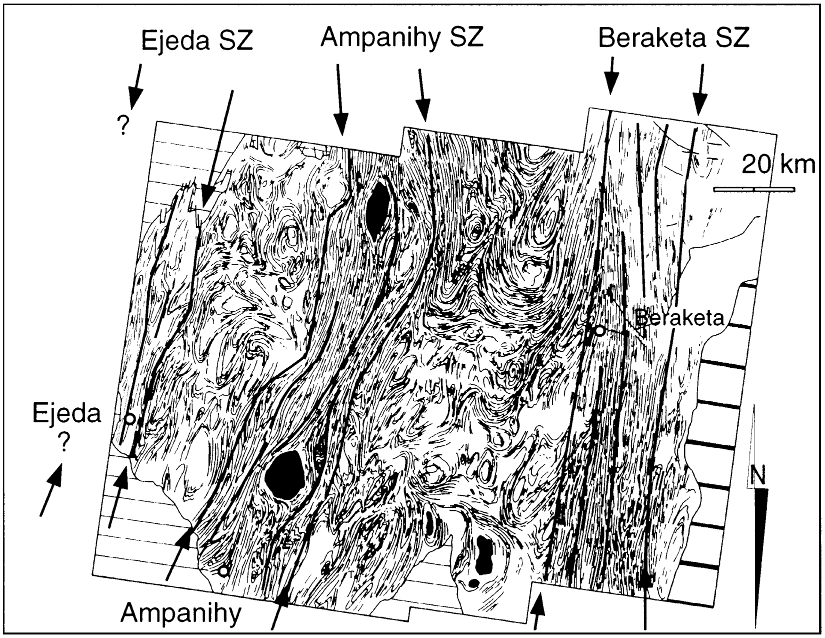

Fig. 8. Detailed foliation trajectories from Fig. 1 (sector 3, Ampanihy–Beraketa in text). Thinnest black lines are foliation planes. Arrows show the limits of the different SZs (Beraketa SZ, Ampanihy SZ, Ejeda SZ). Two large black, ovoid shapes in Ampanihy SZ are the anorthositic bodies; other symbols same as in Fig. 3.

dip. S1 planes are folded into kilometric F2 folds with vertical axial planes and subhorizontal folded hinge lines. The F2 folds may be associated with a new crenulation cleavage (vertical S2 plane), giving rise to an intersection lineation. On the map, the folds are ovoid with two dominant axial traces (N 10 and N 90, Fig. 9). Sometimes they form complex 'mushroom' shapes with the second fold axial trace oriented N 0 or N 90 (Fig. 9). In some places the S1 foliation trajectories underlined typical type III interference patterns (Fig. 11), with the last fold axial trace oriented N 0. The L1 stretching is globally oriented N 250, plunging towards the south or the south-west (20–60°). L1 stretching lineations are reoriented by the F2 folds (west from Beraketa and Tranoroa towns, Fig. 10).

In the Ampanihy SZ, S2 planes are mainly vertical. They are rotated at the vicinity of the two anorthositic bodies (Saririaky and Ankafotia), which behave as rigid particles in a ductile matrix (Martelat et al., 1997). On the north-east boundary of the Ampanihy SZ the S1 foliation planes dip towards the east (40–70°), whereas on the southwestern limit the planes are dipping west (40–70°). Some subhorizontal stretching lineations L2 can be observed, underlined by quartz rods. To the north-west, south-east and south-west of the SZ we clearly observed the transposition of steeply plunging L1 lineation, oriented N 220 or N 80, into a subhorizontal L2 lineation with an N 5 orientation.

Westwards from Ampanihy SZ the folds are more closed with highly folded hinge lines and the L1 lineations are more steeply dipping than in the other eastern dome-and-basin domains.

At the eastern boundary of the Ejeda SZ the S2 planes are vertical and oriented N 5, and they bear a dominant L2 (N 30, 50°S) stretching lineation.

In contrast to the Bongolava–Ranotsara SZ, we have not observed clear deflection of the foliation trajectories at a kilometric scale in the Beraketa, Ampanihy and Ejeda SZs (Fig. 9). At the outcrop-scale we observe primarily conjugated shear bands, symmetrical boudinage, contradictory shear sense criteria and numerous metric upright folds with vertical to horizontal hinge lines in the Ampanihy and Ejeda SZs coherent with dominant coaxial strain and horizontal east–west shortening. In the Ampanihy SZ we proposed a sinistral displacement (Martelat et al., 1995, 1997). On the other hand, in the Beraketa SZ, asymmetric tails around minerals, drag folds and shear bands indicate a more important non-coaxial strain. These non-coaxial kinematic indicators are compatible with a dextral slip system (Rakotondrazafy, 1992).

The kilometric F2 folds observed in the two dome-and-basin domains delimited by the Ejeda, Ampanihy and Beraketa SZs result from the deformation of the S1 planes and give rise to kilometric-scale synforms and antiforms and associated minor folds. This folding event is compatible with a dominant east–west horizontal shortening. F2 folds show vertical axial planes and subhorizontal, gently folded hinge lines. They can interfere with the previous F1 folds described in sector 1 (horizontal axial planes and N 0 or N 90 horizontal hinge lines) and lead to the development of the type II and type III interference patterns (Ramsay, 1967; Thiessen and Means, 1980).

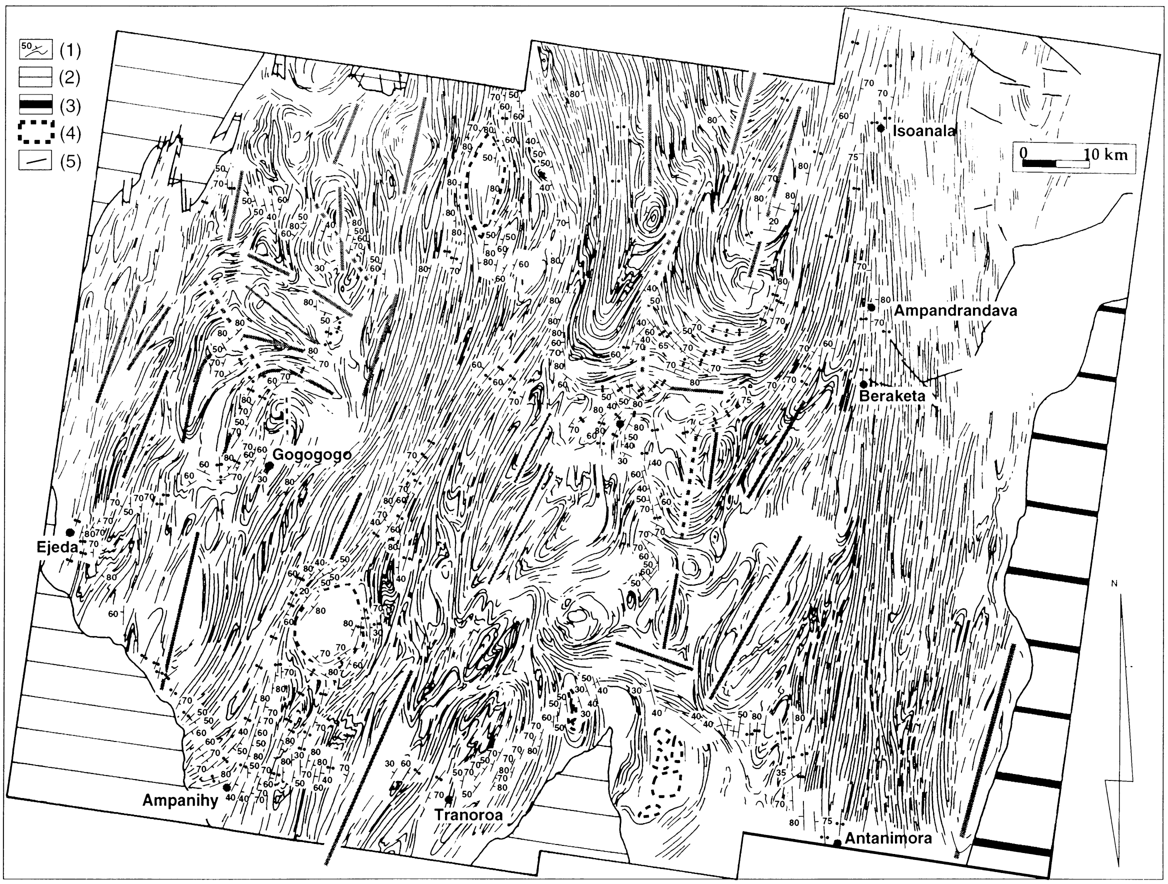

Fig. 9. Foliation trajectories (sector 3). (1) Granulite and foliation plane, (2) post-Cambrian sediments, (3) post-Cambrian volcanic rocks, (4) anorthosite bodies, (5) faults. Gray lines: axial traces of the major folds, dotted gray lines correspond to second fold axial trace.

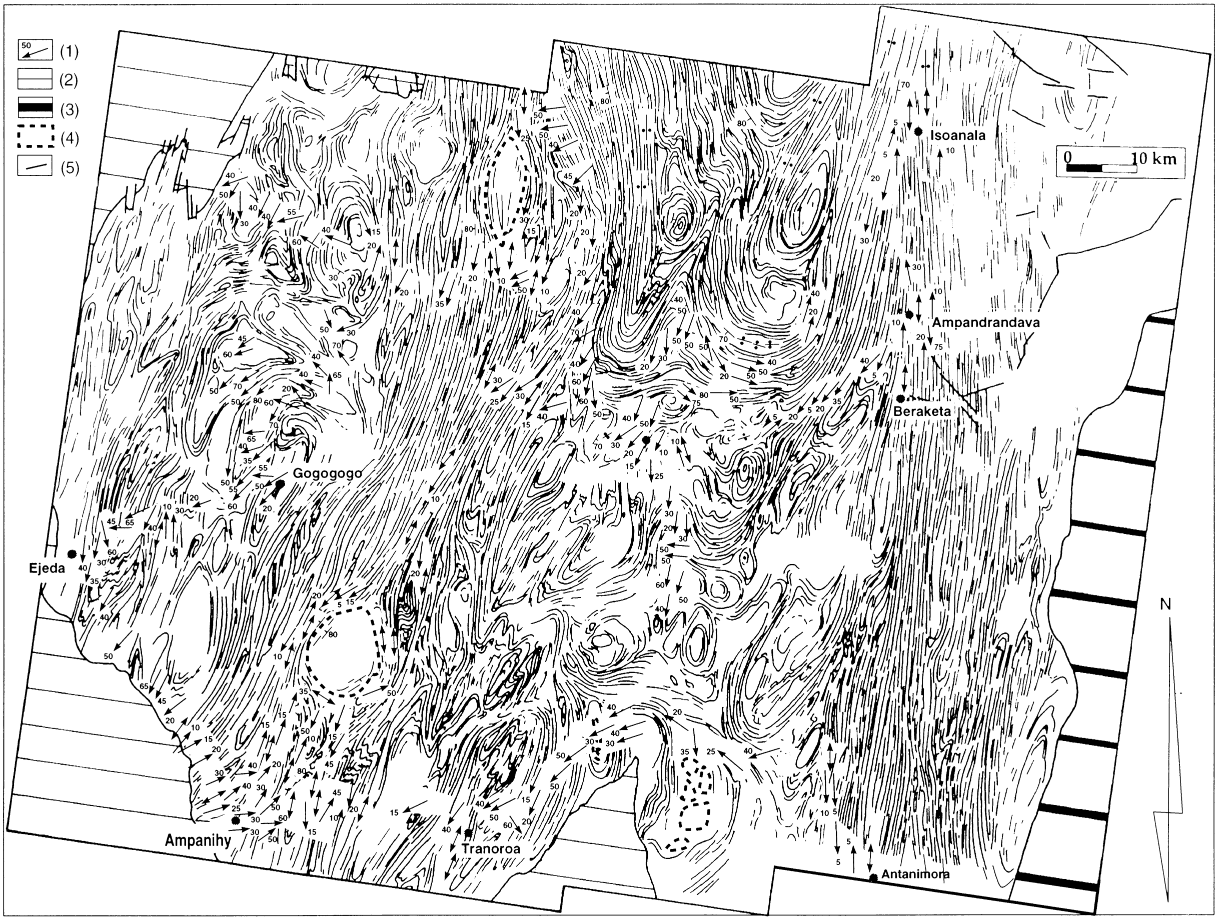

Fig. 10. Detailed stretching lineation trajectories from sector 3. (1) Granulite and stretching lineation; other symbols same as in Fig. 9.

4.4. Sector 4: Tranomaro–Fort Dauphin

4.4.1. SZ geometry

In the southeastern part of Madagascar highly strained domains are also observed from satellite images (170-399, Fig. 2) and geological maps (de la Roche and Marchal, 1956; Noizet et al., 1958; Bazot et al., 1978). These features tend to be less regular and smaller (3–5 km width by 100 km long) than the previously described zones. Among these smaller SZs we distinguish the Fort Dauphin SZ and the Tranomaro SZ systems (near the villages of Tranomaro and Fort Dauphin, Fig. 12).

4.4.2. Field data

In this sector, the S1 fabric dips shallowly (N 0–N 45, Fig. 12) and remains parallel to the syntectonic sheetlike granites, which also should be equivalent to north stratoid granites. This foliation plane is locally deformed and involved in F2 open folds (synclines and anticlines). As previously described, the L1 stretching lineation is globally oriented N 240, plunging 15–30° to the west or the east. The structures S1–L1 are, in some places, as seen near Tranomaro village, transposed into north–south vertical S2 planes bearing a subhorizontal L2 stretching lineation. In such cases, more upright and sometimes unrooted

F2 folds are well developed. This evolution in space clearly represents a finite strain gradient giving rise to an anastomosed framework of minor ductile SZs. In these SZs, asymmetric and symmetric shear criteria indicate, as in previously described SZs, a combination of: (1) pure shear associated with the regional east–west shortening and (2) simple shear associated with, in this case dextral, strike-slip shearing.

5. Deformation history

In southern Madagascar (from 20° of southern latitude) the deformation history clearly results from the superposition of two finite strain patterns D1 and D2 (Fig. 13). The D1 strain pattern is characterized by a subhorizontal foliation plane and an L1 mineral stretching lineation globally trending east–west (N 240–N 270). The S1 fabric corresponds to transposed lithologies giving rise to a compositional layering. Shear criteria are observed both in *XZ* and *YZ* sections of the finite strain ellipsoid and are coherent with vertical shortening with local westwards normal displacements. Numerous syntectonic sheetlike granites ('stratoid' granites) were emplaced in this tectonic context. Because of the intensity of the D2 tectonic event, and of the lack of clear kinematic shear criteria, the tectonic significance of the D1 strain pattern remains unclear.

The D1 structures were reworked during the D2 event, when gradual folding of the S1 planes led to complete transposition of these planes into the high-strain SZs. In the SZs, a new penetrative S2 vertical foliation and a subhorizontal lineation developed. Two kinds of SZ were identified: a major SZ (15–25 km wide by more than 100 km long) and a minor SZ (3–5kilometers wide by less than 100 km long). With the exception of the Bongolava–Ranotsara SZ, which is elongated N 140, all the SZs are globally parallel and north–south oriented.

In the folded domains (dome-and-basin domains), the geometry indicates an evolution from open to upright folds. The folds are more tight in the western part of the studied area (i.e. west of Ampanihy SZ) and, south of the Bongolava–Ranotsara SZ, the intensity of the folding process is compatible with a D2 strain gradient increasing from east to west.

The D2 finite strain pattern shows a strong strain partitioning between vertical SZs and the surrounding dome-and-basin domains that represent less deformed crustal blocks. The east–west D2 horizontal shortening is accommodated by

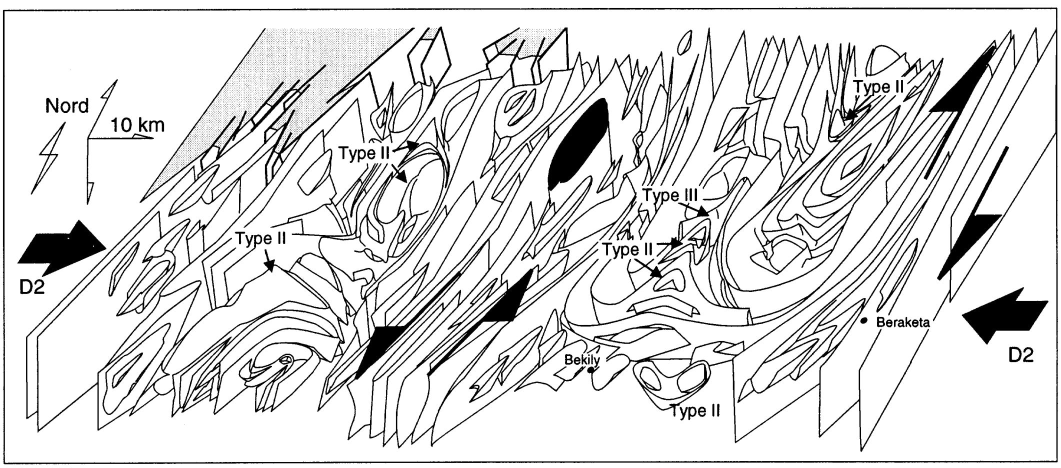

Fig. 11. Interpretative 3D block diagram from sector 3 (Figs. 9 and 10) showing the different interference pattern types (I, II, III; Ramsay, 1967). Gray shading indicates post-Cambrian sediments; the black surface corresponds to the northern anorthositic body (Ankafotia) located in the Ampanihy SZ.

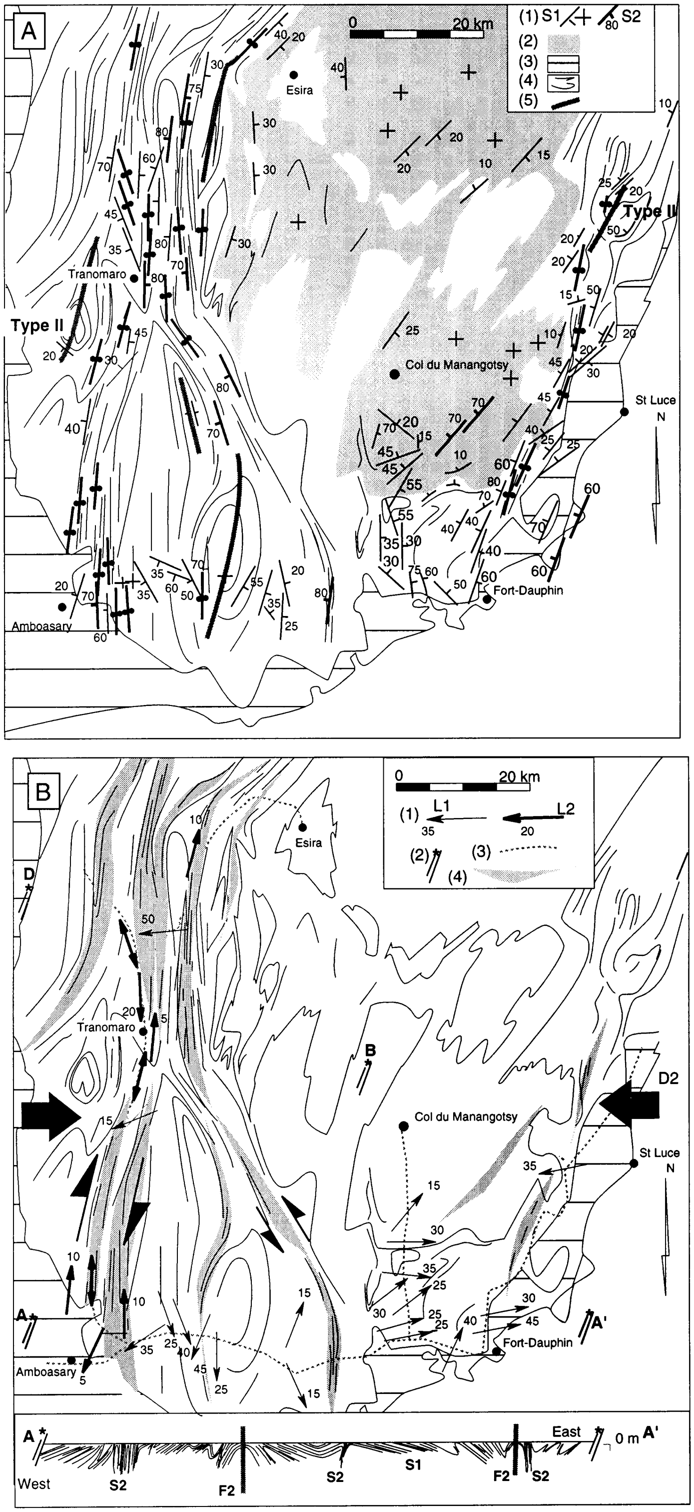

Fig. 12. 3D geometry in south-eastern Madagascar (sector 4, Tranomaro–Fort Dauphin in text) compiled from our observations and from previous data from Bazot et al. (1978), de la Roche and Marchal (1956) and Noizet et al. (1958). (A) Foliation trajectories: (1) foliation plane S1, S2, (2) granitoids, (3) post-Cambrian sediments, (4) granulite, (5) fold axial trace. (B) Stretching lineation map: (1) stretching lineation L1, L2, (2) location of east–west cross-sections, (3) principal roads, (4) minor SZs (3–10 km wide).

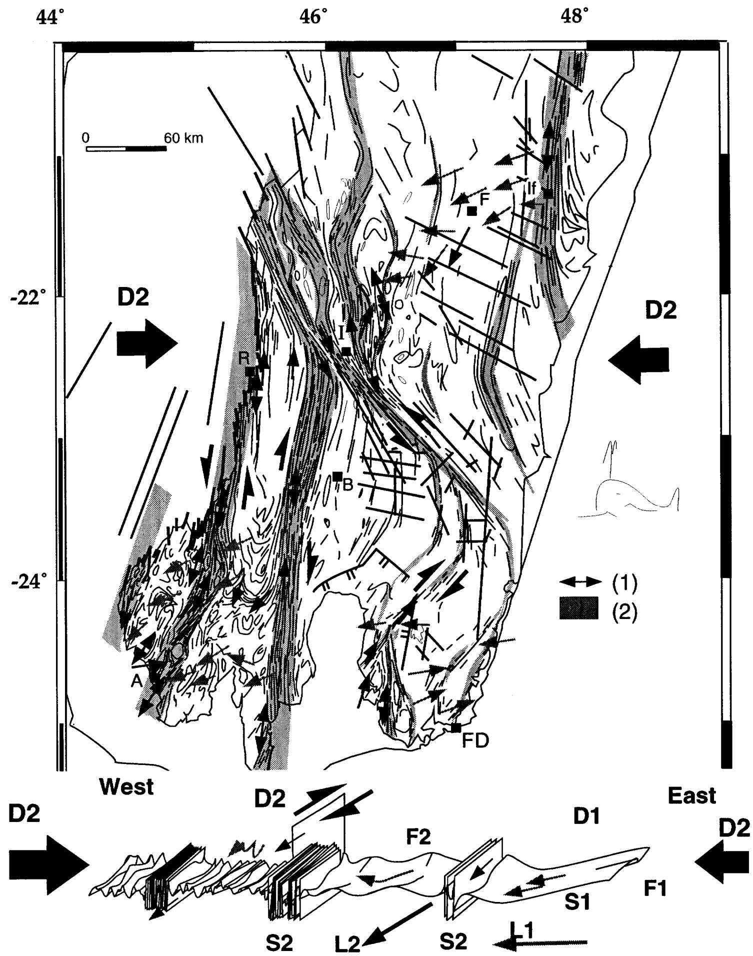

Fig. 13. 3D finite strain pattern of southern Madagascar result showing the superposition of the two events D1 and D2. (1) Stretching lineation: black arrow = L2, gray arrow = L1, (2) SZs, other symbols same as in Fig. 1. Bottom portion of figure is a schematic east–west cross-section. The kinematics of the last D2 event are indicated by the thick black arrows.

folding in the blocks (pure shear-dominated domain), yielding interference pattern geometries with previous F1 folds (D1 structures) and strike-slip displacement (simple shear dominated domain) in the SZs. Moreover within the SZs, combined pure shear and simple shear accommodates east–west shortening. Thus, this D2 finite strain pattern reflects a transpressive tectonic regime, as defined by Robin and Cruden (1994) and Sanderson and Marchini (1984), with different ratios of simple shear versus pure shear (Thompson et al., 1997) inside the SZs as well as between the SZs (i.e. in the folded domains).

6. Metamorphism and deformation

Both D1 and D2 finite strain patterns are coeval with high-temperature metamorphism, as demonstrated in the field by the syntectonic character of the migmatization: melt emplacements are observed in asymmetric or symmetric inter-boudins, shear bands or hinges of F1 or F2 folds. Some amphibole-bearing leucosomes are developed in biotitic gneisses (tonalite to granodiorite) with biotite melting at high-temperature (Kenah and Hollister, 1983). The high-temperature metamorphism is also shown by granulite facies mineral assemblages that underlined both D1 and D2 tectonic fabrics. Typical assemblages (De Waard, 1965; Vielzeuf, 1983) include two pyroxenes–plagioclase–quartz assemblages with rare garnet in metabasites and spinel–quartz associations in metasediments. The precise metamorphic conditions associated with the two finite strain fields have been measured by applying standard methods (e.g. exchange thermobarometry; Spear, 1993) and multiequilibrium thermobarometry (TWEEQ software; Berman, 1988, 1990) on mafic and silico-aluminous lithologies (Fig. 14; Nicollet, 1985, 1988, 1990; Rakotondrazafy, 1992; Martelat, 1998). Our results are compatible with other estimations (Ackermand et al., 1989, 1991; Nédélec et al., 1992; Grégoire and Nédélec, 1997) and demonstrate that uniform high-temperature conditions (750–850°C) existed throughout southern Madagascar. The calculations also reveal regional pressure differences between the different crustal blocks separated by the SZs. Maximum pressure conditions ($P = 12$ kbar) occur in the southwestern part of the area studied, whereas minimum pressures ($P = 5$ kbar) were recorded in the eastern block and northern block (Fig. 14). The calculated P–T paths indicate decompression under high-temperature conditions. The highest pressure and decompres- sion under high temperature were recorded westwards. Eastwards rocks recorded smaller pressure and a weak decompression observable on the field with numerous destabilization in paragneisses of garnet and sillimanite into cordierite. On the northern blocks we only observe the cooling of rocks at the end of the metamorphic phase. Since the D2 SZs juxtapose blocks recording different amounts of decompression, we suggest that the D2 transpressive regime was responsible for the main part of the exhumation of Madagascar granulites.

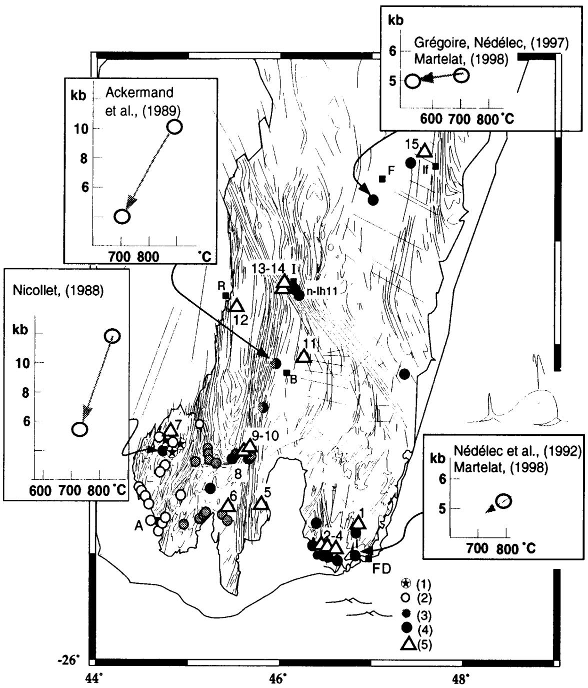

Fig. 14. P–T–t data and finite strain pattern. Foreground: the different P–T–t paths from Nicollet (1988), Ackermand et al. (1989), Nédélec et al. (1992), Grégoire and Nédélec (1997) and Martelat (1998). Background: the different circles correspond to our pressure estimates at the same fixed temperature: 800°C (Martelat et al., 1997; Martelat, 1998). White circle with star: $P > 10$ kbar, white circle $P = 8$–10 kbar, gray circle $P = 6$–8 kbar, black circle $P < 6$ kbar. Triangles correspond to samples used for monazite probe dating (see Table 1 and text for explanation).

7. Age of deformation

Chronologic estimates using the E-probe dating monazite method (Th–U–Pb, Montel et al., 1996) have been undertaken to obtain a better constraint on the age of deformation of these rocks. The age determination is based on the simultaneous measurement, directly on thin sections, of Th–U–Pb concentrations in monazite with an analytical protocol after Montel et al. (1994, 1996). On a single thin section numerous ages can be obtained. A statistical method developed by Montel et al. (1996) is used to deal with the large number of data and to characterize the population of results (unimodal, bimodal, etc.). Results obtained by both methods (E-probe dating and conventional isotopic methods) show a good degree of reliability. Thus E-probe dating monazite is relevant to date magmatism or metamorphism events in a huge variety of tectonic regimes (Montel et al., 1994, 1996). This in situ dating method allows the different measures in the monazite and its relations with other minerals and with the

Table 1
Ages obtained from in situ measurement, on thin sections, of Th–U–Pb concentrations in monazite (Montel et al., 1994, 1996). The first column corresponds to sample relationships with the different SZs: inside mSZ (minor SZ), inside MSZ (major SZ) or outside SZ (in dome-and-basin domains or crustal blocks). The different blocks are individualized by the SZs (B–A: block between Beraketa and Ampanihy SZs; W–A: block westwards from Ampanihy SZ; F–T: block between Fort Dauphin and Tranomaro SZ systems; north-east: block westwards from Ifanadiana SZ). The precise location of samples is indicated in Fig. 14. The ages are obtained by statistical calculations; for some samples there are two or three (sample 13) populations of results

Location	Rock type	Mineral paragenesis	Age (Ma)	Fig. 14: no
Block F–T (south-east)	Charnockite	Qtz-Fk-Pl-Opx-Bt-Grt	558 ± 12; 646 ± 19	1
Block F–T (south-east)	Mineral of monazite		550 ± 20	4
Block T–B (center)	Leucocratic gneiss	Qtz-Fk-Grt-Opx-Bt	590 ± 32	5
Block T–B (center)	Paragneiss	Bt-Crd-Sap-Krn	567 ± 18; 1003 ± 53	11
Block B–A (west)	Paragneiss	Bt-Grt-Crd	529 ± 21	6
Block B–A (west)	Paragneiss	Qtz-Spl-Fk-Pl-Bt-Grt-Crd	539 ± 20; 1389 ± 85	8
Block B–A (west)	Leucocratic gneiss	Qtz-Fk-Grt-Bt	534 ± 23	12
Block W–A (south-west)	Paragneiss	Qtz-Fk-Pl-Bt-Grt-Crd-Sill	529 ± 23	7
Block north-east	Leucocratic gneiss	Qtz-Fk-Grt-Bt	515 ± 20; 848 ± 45	15
mSZ south-east	Charnockite	Qtz-Fk-Pl-Opx-Bt-Grt	536 ± 7	2
mSZ south-east	Charnockite	Qtz-Fk-Pl-Opx-Bt-Grt	523 ± 8	3
MSZ Beraketa	Paragneiss	Qtz-Spl-Fk-Bt-Grt-Crd	502 ± 32	9
MSZ Beraketa	Paragneiss	Qtz-Fk-Pl-Bt-Grt-Sill-Opx-Sap-Crd	488 ± 24; 736 ± 34	10
MSZ Bongolava-R	Paragneiss	Qtz-Fk-Pl-Bt-Grt-Sill-Crd	375 ± 28; 481 ± 41; 1752 ± 103	13
MSZ Bongolava-R	Charnockite	Qtz-Fk-Pl-Opx	430 ± 32	14

tectonic structures to be controlled. Finally, this method is one order of magnitude less precise than conventional isotopic methods. In the southern Madagascar island, the method was performed on various leucocratic lithologies, such as gneisses or charnockites, and one giant monazite crystal in a leucosome (15 samples, results are presented in Table 1 with locations from Fig. 14).

Out of the major SZs, in the region dominated by the flat-lying structures (D1 event), the ages lies between 590 and 515 Ma. These data are consistent with the results of Paquette et al. (1994) obtained on granites and charnockites by single U–Pb on monazites and zircons or Sm–Nd isotopic studies on minerals and whole rocks. Their results come from metagranite showing a gently dipping foliation associated with the D1 structures. They depicted a synchronous peak of metamorphism in all the south-east between 580 and 565 Ma (Paquette et al., 1994). These results are also coherent with the age of 565 Ma obtained with U–Pb on zircon in a pyroxenite by Andriamarofahatra et al. (1990) located in the south-east outside of the vertical SZs. Consequently, we propose that the D1 tectonic is efficient in the period of 590–530 Ma. Considering the results older than 600 Ma (Table 1), and because magmatic ages (800 to 630 Ma) exist in the northern part of Madagascar (Paquette and Nédélec 1998; Tucker et al., 1999), we suggest that they are compatible with inherited ages kept in the cores of the monazite crystals corresponding to protolith ages or remnant tectonics of older tectonic events.

In the minor SZs (D2 structures), we also obtained pan-African ages between 540 and 520 Ma. Paquette et al. (1994) date the formation of zircon-bearing calcite veins inside the Tranomaro minor SZ at around 520 Ma. In the major SZs (samples 9, 10, 13, 14), where the transposed rocks show a vertical foliation plane, we systematically identify younger ages between 500 and 375 Ma: 500–490 Ma in the Beraketa SZ; ages between 480 and 430 Ma and one population around 375 Ma in the Bongolava–Ranotsara SZ. Because the ages below 500 Ma are without tectonothermal significance in the Mozambique belt (Stern, 1994), we suggest that the syngranulitic D2 strain pattern, associated with crustal fluid (CO_2 rich) flow (Pili et al., 1997a), finished between 530 Ma and 500 Ma. As discussed previously, kilometric brittle faults are controlled by previous main ductile tectonic structures (especially Bongolava–Ranotsara SZ). Thus the youngest ages (< 500 Ma) should be coherent with a resetting (Pb loss with late fluid circulation in the upper crust) during the development of brittle structures, possibly during the opening of the Mozambique channel (from Jurassic to Cretaceous).

8. Discussion and conclusions

A regional study that integrates multi-scale tectonic analysis, metamorphic petrology and monazite E-probe dating has been realized in southern Madagascar. The granulitic Precambrian rocks and structures are attributable to late pan-African time (590–500 Ma). In the southernmost part of the area studied (south of Fianarantsoa), two major pan-African deformation events have been recognized. The first one (D1) is represented by a flat-lying granulitic foliation and stratoid granites coeval with isoclinal folds and bearing E–W-stretching lineations. The second one (D2) corresponds to the development of a vertical SZ network (the major ones being rooted in the mantle; Pili et al., 1997b) and to steep refolding of D1 fabrics. In these SZs, which are mainly trending between N 0 and N 20 in the actual position of the island, a poorly deepening stretching lineation is well developed. Only the Bongolava–Ranotsara, the latest SZ cross-cutting other major SZs, is oriented N 140. Thus, tectonic history results from the superposition of the two late pan-African syngranulitic finite strain patterns D1 and D2 continuously developed through time. The D1 finite strain pattern was developed at around 590–530 Ma, whereas the D2 event was effective until 530–500 Ma (D1 and D2 seems to represent a whole progressive deformation event).

Concerning the previous tectonic structures, the D1 event should be either compatible with thickening processes or with post-collisional extension. In both tectonic contexts, high-temperature low pressure metamorphism conditions and huge amounts of granite should be produced (England and Thompson, 1984; de Yoreo et al., 1989, Malavielle et al., 1990; Gardien et al., 1997). For example, in northern Madagascar extensional tectonics was proposed to explain the emplacement of stratoid granites (Nédélec et al., 1994, 1995). Regarding the intensity of D2 reworking, in southern Madagascar a kinematic history of D1 event is questionable. As in other Precambrian fields, the geodynamic significance of such horizontal tectonic structures remains unclear (Harley, 1992).

The D2 bulk strain pattern is clearly related to a transpressional regime during bulk horizontal shortening of heated crust. This transpression, contemporaneous with granulite facies conditions, resulted in a strain partitioning between the anastomosing SZ network and folded domains. The dome-and-basin domains are folded under a pure shear-dominated strain regime, whereas the SZs show a component of the simple shear regime. Inside the SZs the strain also varies, from simple to pure shear, depending on their positions in space.

The vertical SZs separate blocks recording different pressure conditions (from high pressure, 12 kbar, to low pressure, 5 kbar). These regional pressure differences, controlled by kilometric tectonic structures, are compatible with granulites exhumation (at least 20 km of vertical motion) during compression under a transpressive regime (D2 strain pattern). The amount of uplift that juxtaposes contrasted metamorphic domains was controlled by the D2 strain gradient. The uplift reached a maximum in domains affected by intense bulk horizontal shortening and where the amount of pure shear, with respect to simple shear, was the highest (Fig. 15). This hypothesis is confirmed by the strain gradient increases in blocks from east to west and also by the changes of lithologies from acid to more basic from east to west where deeper crustal levels outcrop.

Thus, Madagascar represents an opportunity of studying the evolution and exhumation of deep-seated rocks during an obliquely convergent (transpressive) orogen. The cratons convergence (Tanzanian craton westwards and Dharwar craton eastwards, Fig. 15) promotes the shortening of a hotter and softer piece of continental lithosphere. We suggest that such mechanical heterogeneites (i.e. cratons) in the soft heated lithosphere could control strain partitioning at the scale of the whole of Gondwana. We think that the SZs dominated by pure shear are located at the front of the cratons with respect to the east–west convergence (oriented N 0 Ampanihy or Zazafotsy SZs). The SZs dominated by simple shear are located to the sides of the cratons (oriented N 140 SZs Bongolava–Ranotsara SZ and its prolongation in Kenya).

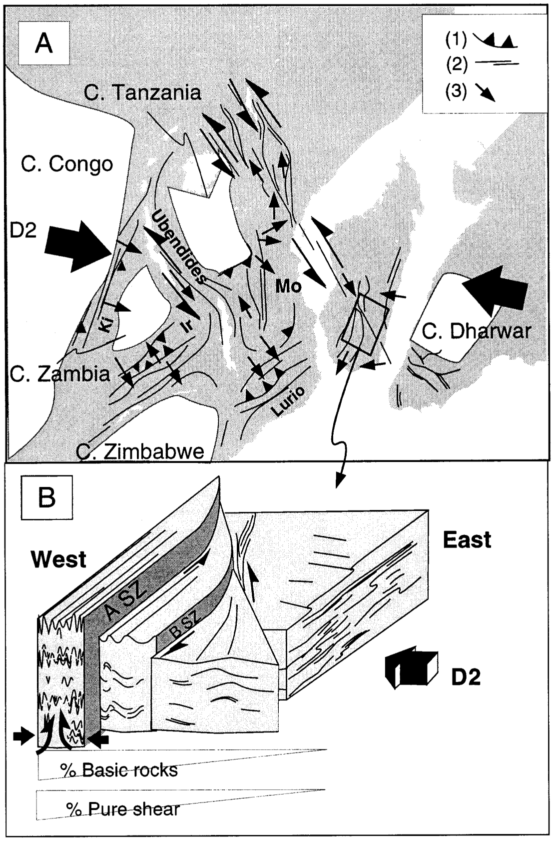

Fig. 15. (A) Madagascar in the framework of late pan-African convergence of cratons. (1) Major horizontal SZs, (2) major vertical SZs, (3) dominant direction of stretching. C., Mo, Ir and Ki correspond respectively to craton, belts of Mozambique, Irumides and Kibarides. This tectonic reconstruction was modified after Norton and Sclater (1979), Powell et al. (1980), Klerkx et al. (1987), Daly (1986), Schackleton (1986) and Martelat (1998). (B) Interpretative block diagram showing the D2 dynamic event in the southern part of Madagascar. The two vertical planes in heavy gray represent vertical major SZs of Ampanihy and Beraketa, and the oblique SZ corresponds to the Bongolava–Ranotsara structure and the northern structure to the Zazafotsy SZ. The zone where the exhumation of the lower crust is more efficient is located in the west (dominated by a pure shear strain regime). Mafic lithologies increase from east to west as the deepest part of the crust outcrops. These two draws underlined the control of cratons on the strain partitioning and subsequent vertical motion of rocks at the scale of the Gondwana.

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