Metamorphic evolution of Pan-African granulite facies metapelites from Southern Madagascar

Gregor Markl a,*, Jürgen Bäuerle b, Djordje Grujic b

a Institut für Mineralogie, Petrologie und Geochemie, Universität Tübingen, Wilhelmstrasse 56, D-72074 Tübingen, Germany
b Geologisches Institut, Albert-Ludwigs-Universität, Albertstrasse 23 b, D-79104 Freiburg, Germany

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Abstract

Granulite-facies, metapelitic gneisses from five localities in Southern Madagascar exhibit very similar mineral assemblages and textures and reveal a very similar P-T evolution. They show an early assemblage of Grt + Sil + Qtz + Bio + Kfsp + Rt + Ilm ± Pl ± Spl which records P-T conditions of about 880 ± 60°C at 8 ± 1 kbar. In some samples, this assemblage is overprinted by the later assemblage Crd + Grt + Sil + Bio + Kfsp + Qtz + Mt ± Crn which equilibrated at about 690 ± 40°C and 4 ± 1 kbar. The main Crd-producing reaction Spl + Qtz = Crd was aided by the further reactions Sil + Bio + Qtz = Crd + Kfsp + H2O and Grt + Sil + Qtz = Crd. Crn formed from relictic Spl, probably during cordierite formation, by the oxidation reaction 3Hc + 0.5O2 = 3Crn + Mt. H2O activities calculated from phase equilibria for both metamorphic events range from 0.6 to 1 which is in agreement with the observation of partial melting during the HT stage. We interpret age data from the literature to suggest that the HT:HP event occurred at about 560–580 Ma, while the MT:MP event could be related to intrusive activity between 520 and 540 Ma, possibly during the same orogenic cycle. The similarities in textures and equilibration conditions among the Malagasy samples strongly imply — in contrast to the opinion of many previous workers — a common tectonometamorphic history of the various parts of Southern Madagascar between 600 and 500 Ma. The P-T-conditions recorded by the samples and the age data correspond closely to P-T-paths deduced for other Gondwana fragments like Southern India, Sri Lanka and parts of East Antarctica. This correspondence may help to constrain the location of the suture for the collision between East and West Gondwana in the late Proterozoic. © 2000 Elsevier Science B.V. All rights reserved.

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1. Introduction

The Panafircan tectonometamorphic event occurred between 950 and 450 Ma (Kröner, 1984) and traces of it are recorded on almost every continent, mostly in form of large scale deforma-
tion, intrusive activity and metamorphic overprints. A key to the understanding of this Panafri
can epoch is the fate of supercontinents—first the break-up of Rodinia, and later the assem-
bly of Gondwana. The supercontinent of
Gondwana was assembled around 500–550 Ma
and it involved — as far as we know today —
Africa, South America, East Antarctica, India,
Australia and most parts of Europe in addition to
Sri Lanka and Madagascar (Stern, 1994; and
references therein). The Panafri
can tectonic, mag-
matic and/or metamorphic imprint has recently
been investigated intensively in most of these ar-
areas in order to reconstruct the detailed history of
Gondwana assembly, break-up and, first of all,
the internal structure. Special emphasis has been
placed on South-west Africa (e.g. Windley, 1995;
Jacobs et al., 1997), Southern India (e.g. Santosh,
1986; Chacko et al., 1987; Braun et al., 1998), Sri
Lanka (e.g. Schumacher et al., 1990; Faulhaber
and Raith, 1994; Hiroi et al., 1994), East Antarc-
tica (Shiraishi et al., 1994; Jacobs et al., 1998;
Markl and Piazolo, 1998; Piazolo and Markl,
1999) and Madagascar (Nicollet, 1985, 1990; Win-
dley et al., 1994; Kröner et al., 1996; Martelat et
al., 1997; Nicollet et al., 1997; Paquette and
Nédélec, 1998; de Wit et al., in press). While each
area investigated has some specialities, most of
them share some common features including in-
trusive activity of similar tectonic affinity and
similar ages, similar metamorphic histories with
similar metamorphic ages and even a similar in-
ventory of rock types. Naturally, most of these
areas record most intensely or even exclusively the
late stages of Gondwana evolution, i.e. the late
stages of the Panafri
can event, depending on the
intensity of the overprint. This is also true for the
here investigated southern part of Madagascar.

With respect to its metamorphic evolution, the
southern part of Madagascar has been the subject
of some controversy within the last few years,
because earlier work seemed to indicate greatly
varying P-T-conditions in Southern Madagascar
(e.g. Moine et al., 1985; Nicollet, 1985; Acker-
mand et al., 1989; Nicollet, 1990; Kröner et al.,
1996; Martelat et al., 1997). These results were
interpreted to show an W-E decrease in pressure
(Nicollet, 1990), to reflect ‘elevator-tectonics’
along shear zones with vertical displacements of
up to 10 km (Martelat et al., 1997) or to reflect
various metamorphic events. As this confusion
about the tectonometamorphic history of the
southern part of Madagascar prohibits a detailed
comparison within the context of Gondwana evo-
lution and break-up, the present contribution is
aimed at clarifying the tectonometamorphic evo-
lution of Southern Madagascar during the late
stages of the Panafri
can event. We present new
petrological data derived from metapelitic rocks
which were collected in an area about 80 000 km²
large (Fig. 1). These new samples, investigated
with identical methods to enable a detailed com-
parison among them, place new constraints on the
Panafri
can effects in this area.

2. Geological setting and sample localities

2.1. Geology

The western third of Madagascar is composed

![Fig. 1. Location of the sample area in Madagascar.](image)
of late Paleozoic to recent sedimentary rocks, whereas the eastern two thirds of Madagascar consist mainly of Precambrian and early Paleozoic metamorphic and magmatic rocks. The Precambrian basement of southern Madagascar has been strongly deformed, commonly to upper amphibolite or granulite grade (Nicollet, 1990; Martelat et al., 1997), and intruded by numerous ‘stratoid’ granitoid intrusions during the Panafri
can orogenesis (Nédélec et al., 1994).

The most striking tectonic features of the Malagasy basement are large shear zones (see Fig. 2). In zones marked by a higher degree of partial melting and high fluid flow (Pili et al., 1997a), Panafri
can deformation was localised in several major and a dozen minor shear zones (Pili et al., 1997b). These zones are characterised by N-S striking, sub-vertical foliation and a sub-horizon
tal to doubly plunging mineral lineation (Martelat et al., 1997). One major shear zone is the NW-SE oriented Bongolava-Ranotsara shear zone, which possibly consists of a en échelon arranged system of shear zones (de Wit et al., 1998). According to most authors this shear zone may represent a boundary between dominantly Archaean crust to the north and dominantly Proterozoic crust to the south (e.g. Ashwal et al., 1998). The other major shear zones are oriented approximately N-S among which the Ampanihy and the Beraketa (or Vorokafotra in Rolin, 1991) shear zone show the best outcrops. They strike about N 10–15° (Martelat et al., 1997), are 10–20 km in width and extend along strike for at least 350 km. North of the Bongolava-Ranotsara shear zone, the Axial Shear Zone of Windley et al. (1994) has the same N-S strike. Between the N-S striking shear zones there are areas which display done-and-basin and mushroom-fold interference structures (Martelat et al., 1997). The areas south of Bongolava-Ranotsara shear zone were subdivided into three (e.g. Besairie, 1964, 1967; Hottin, 1976) to six (Windley et al., 1994) north-south striking belts based on differences in lithostratigraphy and structure. In these belts, paragneisses with sillimanite and cordierite, graphite schists, marbles, calcisilicates and quartzites dominate. Some garnet-cordierite gneisses interpreted to represent former rhyolites have been called leptynites in the French literature (e.g. Hottin, 1976; Nicollet, 1990). The Ampanihy shear zone between two western belts contains two pre-deformational anorthosite massifs (Boulanger, 1959; Ashwal et al., 1998). The massifs behaved as effectively rigid inclusions during deformation resulting in mega-scale (~ 75 and ~ 100 km², respectively) mantled porphyroclasts (Grujic and Mancktelow, 1998). Granites and charnockites of late Proterozoic to Cambrian age range from outcrop scale leucosomes in the migmatites to kilometre scale intrusions. These are undeformed or slightly folded, rarely sheared. In the area of the Axial Shear Zone (Windley et al., 1994), the basement is still dominated by granulite and upper amphibolite facies gneisses, migmatites and many conformable granites.

According to some authors (e.g. Nicollet, 1990; Martelat et al., 1997; Pili et al., 1997b), the shear zones appear to separate blocks with progressively deeper crustal levels from east to west. Based on various geobarometric studies using various tech
iques, pressure was interpreted to increase from ~ 3–5 kbar in the east to ~ 8–11 kbar in the west. Temperature estimates range-depending on pressure — from 620 to 980°C (Ackermand et al., 1989; Gregoire and Nédélec, 1997). The various pressure and temperature estimates have been inter
preted to reflect the same tectonometamorphic event. Based on stable isotope geochemistry and gravimetric data, Pili et al. (1997a) inferred a mantle connection of the major shear zones. The typical spacing between the steep belts is compat
ible with the wavelength of lithospheric buckling (e.g. Martinod and Davy, 1992). From these ob
servations, Pili et al. (1997b) propose that the major shear zones initiated at instabilities in the mantle while the minor shear zones initiated at mid-crustal levels.

2.2. Sample localities

In the course of this study, five metapelitic gneisses from different localities were investigated (see Fig. 2). Sample 10 and 24 were collected in the Andringitra mountains about 7 km to the W (10) or NNW (24) of the village Ambalamandray which is situated about 50 km S of Ambalavao. Both samples are from the country rock in
the floor of a large syenite intrusion, within the Axial Shear Zone of Windley et al. (1994). Sample 37 comes from a hill about 3 km W of the road at the village of Ampandrandava which lies about 10 km N of Beraketa. The outcrop is situated in the Beraketa shear zone or the Betroka tectonic belt of Windley et al. (1994). Sample 99, from the eastern edge of the Ampanihy shear zone, was collected in a riverbed about 10 km E of Ampanihy, on the road to Tranoroa. Sample 163,
finally, was found at a quarry about 10 km W of Ihosy and 5 km W of Ankily. The quarry is situated close to the Bongolava-Ranotsara shear zone and belongs to the Betroka belt. In the outcrops, all samples are associated with textures clearly indicative of partial melting and migmatization.

3. Petrography and mineral chemistry

3.1. Petrography

The investigated samples are typical metapelites which all bear garnet, sillimanite, quartz, K-feldspar, biotite, ilmenite and rutile. Additionally, some samples contain plagioclase, spinel, magnetite, corundum or cordierite (see Table 1 for detailed listings of the assemblages). The mineral textures are very similar in all samples and therefore, the textural features described below were observed in all samples unless stated otherwise. Textures are documented in Fig. 3.

The samples are weakly to strongly foliated gneisses which macroscopically all show garnet as rounded grains in addition to biotite, feldspars and silky white needles of sillimanite. Sample 24 almost exclusively consists of sillimanite with only minor amounts of the other minerals (Fig. 3C).

Microscopically, garnet shows inclusions of quartz, both feldspars, biotite, fibrolite (sample 163, only in the outer parts of the garnet crystals) and in sample 37 of spinel (Fig. 3D). In cordierite-bearing samples, garnet shows either straight, stable grain boundaries with cordierite or it appears to be replaced at its outermost rims by cordierite (e.g. in sample 24). With biotite, quartz and the feldspars, garnet forms straight grain boundaries indicative of equilibration (Fig. 3D). Apart from the few inclusions in sample 37, garnet is never in contact with spinel, but is always separated by a rim of cordierite from it. Based on the textures, though, garnet and spinel are assumed to have formed a stable assemblage before cordierite formation.

Sillimanite occurs as euhedral, large crystals in all samples. In most samples, though, an evidently later generation of fibrolitic sillimanite surrounds and overgrows the euhedral crystals (Fig. 3C). Additionally, the fibrolite observed in the outer parts of garnet crystals in sample 163 appears to represent an earlier generation formed prior to the euhedral sillimanite crystals, possibly during an earlier metamorphic event or during garnet growth on the prograde path. Euhedral sillimanite is stable with all minerals except for cordierite in some parts of sample 10 and 24.

Cordierite occurs in large aggregates in three of the samples. It surrounds and encloses spinel (Fig. 3F), quartz or garnet. Its relation to biotite is somewhat unclear but it appears to replace biotite in some places while they coexist with each other in others. Symplectites of cordierite with quartz and sillimanite occur between spinel and K-feldspar in sample 37 (Fig. 3B). In sample 10 and 24, sillimanite is also found as relics enclosed in cordierite grains.

Spinel, magnetite and corundum form conspicuous intergrowths in sample 10 and 37 (Fig. 3G), while spinel, magnetite and sillimanite are associated in sample 37 only. Spinel appears to have been in equilibrium with garnet, biotite and sillimanite, but it was obviously replaced by cordierite. Either prior to or during this reaction, spinel decomposed to form magnetite with corundum or sillimanite, either as extremely fine-grained intergrowths in the case of corundum (Fig. 3G) or as larger grains and mantles in the case of sillimanite (Fig. 3H).

Biotite forms well-shaped lath-like crystals of a distintively reddish brown colour coexisting with quartz, the feldspars, garnet and sillimanite and, in some cases, with cordierite. It also occurs as inclusions within K-feldspar, quartz, garnet,

<table>
<thead>
<tr>
<th>Sample</th>
<th>Pl</th>
<th>Spl</th>
<th>Mt</th>
<th>Crn</th>
<th>Crd</th>
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<td>x</td>
<td>x</td>
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<td>24</td>
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<td>37</td>
<td>x</td>
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<td>99</td>
<td>x</td>
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<td>163</td>
<td>x</td>
<td>x</td>
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</table>

* All samples contain garnet, ilmenite, rutile, biotite, sillimanite, K-feldspar and quartz.
Fig. 3. Microtextures in the samples of this study: A–D are photomicrographs; E–H are back-scattered electron images. The size of view is about 1.5 mm in A–D and about 1 mm in E–H.
spinel and cordierite. Characteristic biotite-quartz symplectites indicative of the former presence of orthopyroxene have been observed in sample 10 (Fig. 3E).

In most samples, two feldspars were present at peak metamorphic conditions. The alkali feldspar typically shows the fine lamellar texture of perthitic exsolutions. Plagioclase does not show such features. Ilmenite and rutile occur as rounded grains, at some places in contact with each other, in all samples. In sample 24 rutile forms conspicuous intergrowths with quartz in biotite.

3.2. Textural interpretation

The samples record two distinctly different metamorphic assemblages: an earlier one which reached the stability of the assemblage Grt + Sill + Qtz + Ilm + Rt ± Spl ( + Opx?) with K-feldspar, plagioclase, biotite and possibly magnetite as additional phases. This assemblage is a clear indicator of a high-temperature granulite-facies metamorphism with pressures in the rutile stability field. The equilibration of this assemblage appears to have erased any signs of earlier metamorphic assemblages the only possible exception being the fibrolite and biotite inclusions in some garnets — these may point to another high-grade metamorphism prior to the one mentioned here or they are relics of the prograde path.

The stable peak metamorphic assemblage was later partially overprinted by the cordierite-bearing assemblage Grt + Crd + Sill + Bio + Kfs + Pl + Qtz. This partial overprint suggests slow reaction kinetics as a result of small amounts of reaction-enhancing fluid in the rock. The main cordierite-producing reaction in formerly spinel-bearing rocks was

\[ 2\text{Spl} + 5\text{Qtz} = \text{Crd} \] (1)

which explains the mantle textures of cordierite around quartz and spinel. The Crd-Qtz-Sill-symplectite in sample 37 formed most probably by a reaction like

\[ 2\text{Spl} + 2\text{Kfs} + 2\text{H}^+ + 2\text{H}_2\text{O} + \text{Sill} + \text{Crd}. \] (2)

In sample 24, where no spinel was observed, the main cordierite-producing reaction was obviously

\[ 2\text{Grt} + 4\text{Sill} + 5\text{Qtz} = 3\text{Crd}. \] (3)

Some textures may also point to reactions involving the dehydration of biotite, but the detailed reaction is not obvious to formulate. The difference in textures — in some cases, cordierite replaces garnet and sillimanite, while in others, they stably coexist with each other — is hence related to the presence or absence of spinel and this in turn may depend on the bulk rock Fe-Mg ratio as a slightly more magnesian bulk rock composition may displace the FAS-invariant spinel + Qtz + cordierite + garnet + sillimanite point to higher temperatures and pressures (see, e.g. Bucher and Frey, 1994). Alternatively, spinel in this sample was completely consumed during the cordierite-forming event. In sample 163, cordierite did not replace spinel possibly because the spinel in this sample is extremely Zn-rich (see below).

The very conspicuous spinel-corundum-magnetite intergrowths in sample 10 and 37 and the spinel-sillimanite-magnetite textures in sample 37 (Fig. 3H) are readily explained by simple oxidation reactions:

\[ 3\text{Hc} + 0.5\text{O}_2 = 3\text{Crn} + \text{Mt} \] (4)

\[ 3\text{Hc} + 0.5\text{O}_2 + 3\text{SiO}_2 = 3\text{Sill} + \text{Mt}. \] (5)

3.3. Analytical techniques

Electron Microprobe analyses were performed on a CAMECA SX100 at the Institute of Mineralogy, Petrology and Geochemistry at the University of Freiburg, Germany with internal PAP-correction (Pouchou and Pichoir, 1984, 1985). CAMECA-supplied natural and synthetic standards were used for most of the major and minor elements. Measuring times per element were 20 s with an emission current of 10 nA and an acceleration voltage of 15 kV.
3.4. Mineral chemistry

3.4.1. Garnet

Both Grs and SpS component are below 5% in all samples (Table 2). All garnets are almandine-dominated and show similar variations between a Mg-enriched core and an Fe-rich rim (Fig. 4). Most crystals show very flatly U-shaped zonation patterns in terms of $X_{Fe}$, some are, however, completely unzoned. Very minor late diffusive reequilibration is evidenced by tiny zoning in micrometer-sized parts of garnets bordering biotite grains. This is also taken as evidence that most of the garnet and biotite preserved peak metamorphic compositions. Andradite component based on the ideal stoichiometry formula calculation is below 5% in all analyses.

3.4.2. Biotite

Biotite has extremely Al- and especially Ti-rich compositions with up to 7.2 wt.% TiO$_2$ in sample 37 (see Fig. 5A and Table 3). Ti is positively correlated with $X_{Fe}$, which varies between 0.35 and 0.6 in sample 10 and 99, between 0.25 and 0.45 in sample 37, and between 0.5 and 0.6 in sample 24 and 163. $X_{Fe}$ in each sample is lowest in the grains which are enclosed in spinel and is highest in those which are enclosed in feldspar, quartz or cordierite. Those enclosed in garnet show intermediate values. Sample 37 exhibits significantly lower Al, but much higher halogen contents than the rest of the samples (see Fig. 5B and Table 3). $F$ ranges between about 0.6 afu (atoms per for-
Fig. 5. Biotite compositions in the samples of this study: $X_{Fe}$ is plotted against Ti (A); and Al (B) content per formula unit.

Table 3
Selected biotite analyses used for geothermobarometry

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<td>0.00</td>
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<td>0.21</td>
<td>0.64</td>
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<td>0.44</td>
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<tr>
<td>F</td>
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<td>95.22</td>
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<td>95.15</td>
<td>95.80</td>
<td>95.25</td>
<td>95.13</td>
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Formula based on ideal number of oxygens

| Si   | 2.60  | 2.63  | 2.63  | 2.60  | 2.75  | 2.72  | 2.61  | 2.67  | 2.61  | 2.71  |
| Al   | 1.51  | 1.57  | 1.53  | 1.69  | 1.22  | 1.24  | 1.56  | 1.51  | 1.82  | 1.53  |
| Ti   | 0.36  | 0.29  | 0.28  | 0.21  | 0.36  | 0.40  | 0.26  | 0.28  | 0.01  | 0.17  |
| Fe$^{3+}$| 1.30 | 1.26  | 1.15  | 0.90  | 1.02  | 1.03  | 1.32  | 1.30  | 1.01  | 1.04  |
| Mg   | 1.01  | 1.05  | 1.22  | 1.47  | 1.46  | 1.37  | 1.10  | 1.05  | 1.62  | 1.43  |
| Mn   | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.01  | 0.01  | 0.00  | 0.00  | 0.00  |
| Na   | 0.01  | 0.02  | 0.02  | 0.02  | 0.00  | 0.00  | 0.02  | 0.01  | 0.00  | 0.02  |
| K    | 0.94  | 0.93  | 0.94  | 0.91  | 0.93  | 0.94  | 0.95  | 0.94  | 0.78  | 0.90  |
| Cl   | 0.02  | 0.01  | 0.03  | 0.03  | 0.08  | 0.08  | 0.04  | 0.05  | 0.00  | 0.00  |
| F    | 0.11  | 0.10  | 0.13  | 0.10  | 0.58  | 0.53  | 0.18  | 0.17  | 0.03  | 0.02  |
| Sum  | 7.75  | 7.76  | 7.79  | 7.81  | 7.74  | 7.72  | 7.83  | 7.77  | 7.86  | 7.81  |
| $X_{Fe}$| 0.56 | 0.55  | 0.48  | 0.38  | 0.41  | 0.43  | 0.55  | 0.55  | 0.38  | 0.42  |
Fig. 6. Feldspar compositions in the samples of this study, plotted into the Or-An-Ab triangle. (A) alkali feldspar; and (B) plagioclase.

the absence of significant fluid species in the cordierite channels is inferred.

3.4.5. Spinel

Spinel composition is widely varying as a result of the later reactions mentioned above. Spinel coexisting with magnetite and corundum or sillimanite is the most Mg-rich in the respective samples and reached an $X_{Mg}$ of about 0.45, while else it has $X_{Mg}$ values between 0.2 and 0.35 (Fig. 7 and Table 4). $Fe^{3+}$ contents are always below 0.1 in all samples. In sample 163, which has only a few spinel grains, these are extremely rich in gahnite component with ZnO contents up to 13 wt.%, whereas spinel in the other two samples is almost devoid of Zn.

3.4.6. Other minerals

Corundum, rutile and sillimanite were checked to have almost perfect ideal composition. Ilmenite has no compositional specialities apart from up to 0.3 wt.% MgO and up to 0.8 wt.% MnO. Hematite component is below 0.03 in all analyses.

4. Thermobarometry and estimation of fluid parameters

4.1. Methods

Pressure-temperature calculations were performed using conventional thermobarometers calibrated — if possible — for mineral compositions and under P-T conditions similar to the ones we found in our samples. For the first, higher pressure granulite-facies event we used the following calibrations:

- garnet-biotite thermometry after Patiño-Douce et al. (1993) because of the extremely high Ti and Al content of the biotite in our samples;
- garnet-ilmenite thermometry after Pownceby et al. (1987);
- garnet-sillimanite-plagioclase-quartz barometry after Koziol (1989) with adjustments for the sillimanite-kyanite transition after Holdaway (1971);
Table 4
Selected cordierite and spinel microprobe analyses used for geothermobarometry

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Formula based on ideal number of oxygens and, for spinel, cations

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Fig. 8. Results of the P-T-calculations for two representative samples. See text for discussion of the geothermobarometers used.
Fig. 9. Results of the P-T calculations for all samples of this study. The solid lines labelled ‘dry’, ‘0.1’, ‘0.2’, and ‘0.3’ are solidus curves for granitic melts at different water activities after Johannes and Holtz (1996).

- garnet-rutile-ilmenite-plagioclase-quartz barometry after Bohlen and Liotta (1986);
- garnet-rutile-ilmenite-sillimanite barometry after Bohlen and Liotta (1986);

The earlier assemblage — in the following referred to as high-temperature/high-pressure assemblage (HT/HP) — equilibrated in the range 820–950°C at 6.5–9.5 kbar. This is the range spanned by all samples, every single sample, however, shows a much smaller range of equilibration conditions (see Fig. 9). The very high temperatures of these granulites are indicated by all geothermometers and by the combination of all geothermobarometers used. The pressures are independently constrained by the appearance of rutile (Itaya et al., 1985) to be higher than about 6 kbar and by the absence of kyanite to be lower than about 10–12 kbar. The various samples appear to represent slightly different depths where metamorphism occurred, but obviously, they all have been subjected to the same geotherm during equilibration. The geotherm spanned by the five samples corresponds to very roughly 100°C/kbar (see Fig. 9).

For these calculations, garnet and biotite core compositions were used to constrain the higher pressure event, while intermediate garnet compositions (not rim, these were thought to have possibly equilibrated later as a result of lower temperature diffusion) were used with average cordierite compositions in the calculations for the lower pressure event. Garnet-biotite thermometry was not used to constrain the cordierite-event as a result of the difficulties in deciding whether biotite was still stable with cordierite or not. Preliminary tests revealed, however, temperatures in the same range as those recorded by the garnet-cordierite assemblage.

Further calculations concerning the H₂O activity, K⁺/H⁺ ratio, oxygen fugacity or silica activity in the metamorphic fluids were performed using the GEOCALC software of Lieberman and Petrakakis (1991) with the thermodynamic database of Berman (1988).

4.2. Results of P-T calculations

P-T calculations using the various geothermobarometers show two distinctly different sets of metamorphic conditions which are in agreement with additional constraints on minimum and maximum pressures and with petrogenetic grids for the observed assemblages (e.g. Bucher and Frey, 1994; p. 209). Fig. 8 shows the representative calculations for the two samples 24 and 37 and the relatively narrow overlap of the various geothermobarometers.

The earlier assemblage — in the following referred to as high-temperature/high-pressure assemblage (HT/HP) — equilibrated in the range 820–950°C at 6.5–9.5 kbar. This is the range spanned by all samples, every single sample, however, shows a much smaller range of equilibration conditions (see Fig. 9). The very high temperatures of these granulites are indicated by all geothermometers and by the combination of all geothermobarometers used. The pressures are independently constrained by the appearance of rutile (Itaya et al., 1985) to be higher than about 6 kbar and by the absence of kyanite to be lower than about 10–12 kbar. The various samples appear to represent slightly different depths where metamorphism occurred, but obviously, they all have been subjected to the same geotherm during equilibration. The geotherm spanned by the five samples corresponds to very roughly 100°C/kbar (see Fig. 9).

The later, cordierite-bearing assemblage — the medium-temperature/medium-pressure assemblage (MT/MP) — equilibrated between 650 and 730°C and between 3 and 5 kbar (Fig. 9). The range spanned by all samples is significantly smaller than the one indicated by the earlier HT/HP assemblage, but only three samples show the later overprint at all.
4.3. Constraints on H$_2$O activity and the composition of the fluid

4.3.1. H$_2$O activity and K$^+$/H$^+$ ratios

As the samples were subjected to partial melting at temperatures in the range 850–900°C, the H$_2$O activity must have been at or above 0.3 according to Johannes and Holtz (1996). Apart from this estimate, the H$_2$O activity was calculated independently from equilibria like

$$\text{Ann} + \text{Sill} + 2\text{Qtz} = \text{Kfsp} + \text{Alm} + \text{H}_2\text{O} \quad (6)$$

for both assemblages or, if it is inferred that cordierite in the spinel-bearing samples was still in equilibrium with biotite, from the reaction

$$2\text{Ann} + 6\text{Sill} + 9\text{Qtz} = 2\text{Kfsp} + 3\text{Fe-Crd} + 2\text{H}_2\text{O} \quad (7)$$

for the MT/MP assemblage. Calculations used the same garnet, cordierite and biotite compositions as for the respective P-T estimations (with a slightly more Fe-rich biotite for reaction 7). Almandine activities were calculated after Berman (1990), biotite, K-feldspar and cordierite activities with an ideal site mixing formulation.

These equilibria provide, as a result of the difficulty in discerning the accurate biotite composition with which the garnet under high-pressure and the cordierite under low-pressure conditions were in equilibrium, and as a result of uncertainties in the mixing properties of biotite, only rough estimates. As shown on Fig. 10A, H$_2$O activities during both the HT/HP and the MT/MP event are constrained between 0.6 and 1, if the whole range of possible pressures, temperatures, biotite and garnet compositions are taken into account. In detail, sample 24, 99 and 163 indicate slightly higher H$_2$O activities than sample 10 and 37 in the HT/HP calculations, but all samples with cordierite indicate H$_2$O activities of more than about 0.75 and possibly close to 1 (Fig. 10B). These estimates are in accordance with the relatively low halogen and especially Cl contents of the biotite which indicate a fluid of low salinity.

The formation of the cordierite-quartz symplectite by reaction (2) is dependent on pressure, temperature, H$_2$O activity and the K$^+$/H$^+$ activity ratio. As pressure and temperature are relatively well constrained, the K$^+$/H$^+$ activity ratio can be readily calculated for a fixed H$_2$O activity. In addition to Eq. (2), the following equilibria also allow the calculation of K$^+$/H$^+$ activity ratios in various samples during the MT/MP equilibration:

\[ \text{Ann} + \text{Kfsp} + \text{H}_2\text{O} = \text{Alm} + \text{Sill} + 2\text{Qtz} \]

\[ 2\text{Ann} + 6\text{Sill} + 9\text{Qtz} = 3\text{Fe-Crd} + 2\text{Kfsp} + 2\text{H}_2\text{O} \]

\[ 2\text{Ann} + 6\text{Sill} + 9\text{Qtz} = 3\text{Fe-Crd} + 2\text{Kfsp} + 2\text{H}_2\text{O} \]

Fig. 10. Diagrams relating the activity of annite component in biotite to the activity of water at 7–9 kbar, 850–950°C (A); and at 4 kbar, 650°C (B). The range of $a_{\text{H}_2\text{O}}$ relevant for the different samples is constrained by the intersection of the sample box with the respective reaction curve. See text for discussion.
Fig. 11. Diagrams relating oxygen fugacity to temperature (A); and silica activity (B). These diagrams were constructed to explain the corundum- and sillimanite-magnetite-spinel textures found in sample 10 and 37. Corundum-bearing assemblages are evidently stable under low silica activities only, irrespective of the temperature. See text for discussion.

the rock during cordierite formation could promote redox reactions. Hence, spinel breakdown was investigated as a function of temperature, oxygen fugacity and silica activity, as these are the parameters of most importance for equilibria (4) and (5). In Fig. 11A, equilibria among corundum, hercynite, sillimanite and quartz are calculated at 5 kbar and with an ideal site mixing model for the average spinel analysis from samples 10 and 37. For comparison, also the fayalite-magnetite-quartz buffer is shown. In Fig. 11B, these equilibria are calculated at 5 kbar, 650°C and with the same hercynite activity. Silica activity is defined such that pure quartz at standard conditions has an activity of 1. Interestingly, although the rocks are all quartz saturated, differences in silica activity explain the different textures more readily (Fig. 11B) than temperature differences. This indicates that the coexisting fluid was not in equilibrium with quartz in the corundum-producing samples, but had silica activities below 0.8. Differences in temperature would always favour the formation of sillimanite in the presence of a quartz-saturated fluid (Fig. 11A). Involvement of non-equilibrium growth of corundum is another possibility, but in the presence of a fluid at 650°C, this is considered unlikely. Furthermore, the presence of the sillimanite in one sample argues against this possibility. Oxygen fugacities in the fluid as indicated by these reaction textures are at least 2–3 orders of magnitude above the FMQ buffer at the same temperature (see Fig. 11A) — independent on the actual temperature.

Oxygen fugacity estimates of the high-temperature assemblage which involves ilmenite and rutile can be performed using the oxygen barometer of Zhao et al. (1999). Estimates based on this calibration using ilmenite compositions as reported in Bäuerle (1999) indicate oxygen fugacities of QFM ± 1.

5. Discussion and conclusions

5.1. Timing of Panafican magmatism, metamorphism and deformation on Madagascar

The style and exact timing of the Panafican
events in Madagascar is still a matter of debate. The oldest events are dated between 791 and 780 Ma, and are related to the emplacement of granitoids in Proterozoic metasediments of the Itremo region in central-western Madagascar (Ashwal and Tucker, 1997; Handtke et al., 1997). The sillimanite-cordierite-garnet metasediments in the Ihosy area have a depositional age of less than 720 Ma (Kro¨ner et al., 1996) based on detrital zircon ages. The igneous crystallization age of the anorthosites in southern Madagascar (located within the Ampanihy shear zone) is estimated at 660 ± 60 Ma by combined whole-rock and mineral Sm-Nd isochrons (Ashwal et al., 1998). The same authors suggest an emplacement event at or before the onset of the high grade metamorphism (559 ± 50 Ma). Recently, Paquette and Nédelec (1998) reported ca. 630 Ma U-Pb zircon ages on syntectonic alkaline stratoid granites of central-northern Madagascar (north of Itremo region). These granitoids are interpreted as being emplaced at mid-crustal levels during post-collisional extensional collapse (Nédelec et al., 1994; Paquette and Nédelec, 1998). From this, Paquette and Nédelec (1998) infer that collision between East- and West-Gondwana in this area started at about 700–650 Ma. However, de Wit et al. (1998) attribute an age of 660–620 Ma (U/Pb ages on zircon and monazite) to thrusting at mid-crustal levels associated with peak granulate-facies metamorphism. A later group of undeformed Pan-African alkaline granitoids in the Itremo region which clearly crosscut all the previous structures show ages between 580 and 520 Ma (Ashwal and Tucker, 1997; Handtke et al., 1997). Similarly, the post-tectonic Ambatolampy biotite-granite in central Madagascar has an age of ca. 540 ± 3 Ma (Tucker et al., 1999).

Age data for the formation of the shear zones have been interpreted differently by different authors. While Martelat et al. (1997) argued that all the major shear zones in southern Madagascar developed coevally in a transpressional regime, de Wit et al. (1998) interpreted the Bongolava-Ranotsara shear zone as post-dating the N-S trending shear zones. Consequently, Martelat et al. (1997) attributed a 580–550 Ma age to the formation of all shear zones, while de Wit et al. (1998) interpreted the N-S trending shear zones as having formed by ductile pure shear between 610 and 608 Ma and the sinistral Bongolava-Ranotsara shear zone during the youngest parts of a thermal event between 550 and 520 Ma.

High-grade metamorphism and anatexis in southern Madagascar (Ihosy area) are documented by monazite and zircon ages (Pb/Pb evaporation) between 561 ± 12 and 526 ± 34 (Andriamarofahatra et al., 1990; Kro¨ner et al., 1996). These ages are very similar to those obtained in southernmost Madagascar (Forth Dauphin-Tôlanaro) by Paquette et al. (1994). They interpret the zircon ages between 580 and 560 Ma to reflect the peak of high-grade metamorphism and synchronous intrusion of stratoid granitoids (the Anosyan massif). Applying the electron microprobe monazite dating method for the Th-U-Pb system, Nicollet et al. (1997) inferred that the high-temperature metamorphism in southern Madagascar occurred at 560–550 Ma. Subsequent cooling lasted until ca. 500 Ma. Younger ages from samples from within the shear zones (450–500 Ma in the Beraketa and 375–480 Ma in the Bongolava-Ranotsara shear zone) were attributed to hydrothermal processes during exhumation (Nicollet et al., 1997). However, these large difference in ages of peak metamorphic events could also be as a result of different age dating techniques. Nevertheless, the age of peak metamorphism at about 560 Ma is quite well constrained.

5.2. Implications for the P-T-t evolution of South Madagascar

In the present study, we investigated samples of similar bulk composition from a large area. While the sample density is somewhat low, the remarkable similarity in textures and calculated P-T conditions indicates a common tectonometamorphic history of Southern Madagascar during the Panafirican. The variations in P-T estimates for the early HT/HP event among the samples could in principle be caused by calculation uncertainties. However, the good agreement for the various geothermobarometers applied to every single sam-
ple argues against this. It is more likely that the relatively small variations of 2 kbar and about 100°C between the samples reflect slightly varying depths where these samples underwent the early granulite-facies metamorphism. If the calculated differences are real, the samples were obviously subjected to minor vertical depth changes before attaining a common crustal level where they experienced the following MT/MP metamorphism. With respect to the temperatures reached, it is most likely that the rocks suffered partial melting before they equilibrated as the granulites we see today. This equilibration may have obliterated any signs of earlier melting. Taking the published geochronological data detailed above into account, it appears most reasonable to identify the HT/HP granulite-facies metamorphism with the granulite-facies event of Andriamarofahatra et al. (1990), Paquette et al. (1994) and Nicollet et al. (1997) at about 550–580 Ma.

The later MT/MP event is thermobarometrically well constrained. At conditions of 4–5 kbar and at about 650–700°C, fluids of relatively high water activity entered the rocks and caused a partial reequilibration. As pointed out by Bucher and Ohta (1993) for cordierite-gneisses from Dronning Maud Land in Antarctica, these conditions strikingly correspond to the water-saturated granite solidus at these conditions. Bucher and Ohta (1993) inferred that fluid released from crystallising granitic plutons may be the source of the retrogressing fluids. This was further corroborated in the same area by Markl and Piazolo (1998) based on the halogen geochemistry of minerals from the granitoids and the surrounding metamorphic rocks. We propose that the same mechanism operated in Southern Madagascar. The reason, why some samples show a later MT/MP overprint while others do not is then related to the access of such granitoid-related fluids to the rocks. This access may vary strongly even on an outcrop scale. Age data of Kröner et al. (1996) and de Wit et al. (1998) indicate that this metamorphic event may be related to the magmatic activity between 540 and 520 Ma, where large amounts of granitic rocks intruded in central and southern Madagascar. The relatively small range in pressures within this sample group argues for the fact that after this metamorphism, the samples shared a common uplift and erosion history until today.

5.3. Comparison to previous results

Earlier work on granulites from Southern Madagascar gave highly variable and confusing results. P-T estimates which were in most cases ascribed to the same granulite-facies metamorphic event ranged from 9.5 kbar/980°C (Ackermand et al., 1989) to 4.5 kbar/620°C (Gregoire and Nédélec, 1997) with many values in between (e.g. Moine et al., 1985; Nicollet, 1990; Kröner et al., 1996; Martelat et al., 1997). Most of the previous studies were also performed on metapelitic rocks, so the phase assemblages used for thermobarometry were broadly similar to each other and to those of the present study. However, most of the temperature estimates from the literature are based on garnet-biotite thermometry using calibrations which are not really valid for the extremely Al- and Ti-rich, granulite-facies biotites encountered here. The same is true for TWEEQU calculations without a suitable mixing model for these extreme biotite compositions. It is therefore most probable that erroneous temperatures combined with T-dependent barometers (e.g. garnet-sillimanite-plagioclase-quartz) lead to erroneous P-T estimates which were then interpreted to reflect strongly variable P-T histories for samples from various localities in Southern Madagascar. Additionally, former workers did not always clearly distinguish the HT from the later MT metamorphic assemblage — the resulting mixture of data may add to the confusion.

For our samples, we could show that the use of other garnet-biotite thermometry calibrations which do not take the high Ti- and Al-contents into account, resulted in hugely varying and in most cases geologically unrealistic temperature and, consequently, pressure estimates. The coherence and similarity among our samples which span the whole area investigated by the various earlier authors imply that differences in the calibrations used may have created an artificially complicated picture of the Southern Madagascar tectonometamorphic evolution. This framework

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needs to be further filled with additional thermo-
barometric and geochronological data.

5.4. Comparison to P-T-t-paths from Southern
India, Sri Lanka and East Antarctica

The results of the present study show remark-
able similarities to those from other continental
fragments of Gondwana today dispersed around
the Indian Ocean. As shown in Fig. 12, the south-
ern part of Madagascar was correlated by Wind-
ley et al. (1994) with the southern part of India
and further on with Sri Lanka and Lützow-Holm
Bay in East Antarctica. This correlation was
mainly based on structural arguments and the
most important correlation marks are the huge
shear zones found in South Madagascar and
South India, here especially the Anchankovil shear
zone which was interpreted to represent the con-
tinuation of the Bongolava-Ranotsara shear zone
in Madagascar.

Fig. 12B compares P-T paths from different
areas shown in Fig. 12A. It is obvious that the
P-T-paths of Sri Lanka, South India, South
Madagascar, Lützow-Holm Bay and Dronning
Maud Land as reported by our study, Ackermand
et al. (1989), Santosh and Yoshida (1992), Bucher
and Ohta (1993), Hiroi et al. (1994) and Piazolo
and Markl (1999) show remarkable similarities:
an early high-pressure granulite-facies event
which is dated to have occurred between 550 and 600
Ma in all of these areas is followed by a lower
pressure retrogression which typically occurs
around 650°C at 3 ± 1 kbar and which is, where it
is dated, related to charnockitic and A-type grani-
toid magmatism between 540 and 500 Ma (e.g.
Jacobs et al., 1998; Markl and Piazolo, 1998).
Additional similarities include, e.g. the occurrence
of massif-type anorthosites with intrusion ages
between 600 and 700 Ma in both Southern Mada-
gascar and Central Dronning Maud Land in
Antarctica (Ashwal et al., 1998; Jacobs et al.,
1998), while in Southern India, an early alkali
granitoid magmatism was dated to fall in this age
range (Rajesh and Santosh, 1996). The results of
Schumacher and Faulhaber (1994) indicate princi-
pally a similar conditions of metamorphism for
the Highland Complex (HC) of Sri Lanka com-
pared to Madagascar, however, in the HC an
oblique section through the middle and lower
crust appears to be exposed. Schumacher and
Faulhaber (1994) suggest that the observed steep P-T array in the HC is related to heating of the middle crust by pre- to synmetamorphic intrusions similar to Madagascar.

While these observations do neither provide compelling evidence nor refusal of any plate-tectonic reconstruction like the one of Windley et al. (1994) or Kröner et al. (1996), they nevertheless argue for common tectonometamorphic events in these areas. It is noteworthy that this is also true for central Dronning Maud Land which was according to Windley’s et al. (1994) reconstruction situated about 1200–1500 km to the south of Madagascar. In the light of the extremely close correspondence between the southern Indian Kerala Khondalite belt with South Madagascar and with central Dronning Maud Land in ages and type of magmatism including the occurrence of the rare late-Proterozoic massif-type anorthosites in Madagascar and central Dronning Maud Land, in the detailed metamorphic evolution and in the rock inventory — e.g. the striking and common occurrence of marbles bearing humite-group minerals associated with cordierite-sillimanite-garnet gneisses (see Piazolo and Markl, 1999, and references therein) — strongly argues for a reconstruction in which between 700 and 500 Ma South Madagascar, South India, Sri Lanka and central Dronning Maud Land were situated at least close to each other.

Another question concerns the mechanism by which a huge area like this underwent so extremely similar episodes and kinds of magmatism and metamorphism. If one takes the age data from Dronning Maud Land as the most detailed and reliable ones (these are SHRIMP data on zircons, Jacobs et al., 1998), the history of anorthosite formation, granulite-facies metamorphism, intrusion of A-type granitoids and concomitant upper amphibolite-facies retrogression took place over a period of about 100 Ma between 610 and 500 Ma. We argue that a period of basaltic underplating, hence the ponding of mafic magmas at the base of the crust, resulted in the formation of the anorthosites. The peak metamorphic granulite-facies metamorphism was probably related to the collision between East and West Gondwana, as this event most likely resulted in substantial crustal thickening. In the course of this process, the metasediments investigated in this study and those in the other comparable areas were buried to depths in excess of 35 km. Consequently, the areas compared above with their surprisingly similar history would help to constrain the suture zone of East and West Gondwana at about 560–580 Ma. This could be performed by affiliating similar P-T-t-domains to the same tectonic plate because lower and upper plate during collision should be expected to have different tectonometamorphic histories. The occurrence of the peculiar Panafican anorthosites may additionally indicate close proximity of southern Madagascar with central Dronning Maud Land before this collisional event. Later, during a syn- or post-collisional extensional collapse of the thickened crust at about 500–530 Ma, the lower crustal rocks were successively uplifted to higher levels. This may have involved low angle normal faults (not yet recognised in the field or already eroded) and vertical movements along the shear zones as well as erosion or active uplift as, e.g. shown by Thompson et al. (1997). Concomitantly, partial melting in the lower crust lead to renewed intrusion of granitoids into the middle crust, which at this time were made up by the rocks of this study. These intrusions triggered the formation of cordierite in the surrounding country rocks. Combining petrological, geochronological and structural informations and interpretations, we argue that mapping areas of closely similar tectonometamorphic histories and drawing the boundary between different P-Tt domains may finally result in the reconstruction of the large suture between both parts of Gondwana and in understanding the detailed processes during this supercontinent amalgamation. Granulites recording similar tectonometamorphic histories in the Alps (Droop and Bucher-Nurminen, 1984) and in the Himalayas (Lombardo et al., 1999) may furthermore indicate that these types of tectonics recorded by high-pressure granulites with later cordierite are typical for continent-continent collisions.
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References


