

Decompression cooling of the basement garnet-sillimanite paragneisses from Palmas (Tocantins, Brazil)

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Abstract: Protolithic rocks of the paragneisses were well sorted high aluminous pelitic sediments. Medium pressure regional metamorphism of slow heating, indicated by slow garnet nucleation rate (3.8×10^{-10} N/cm³s), attained the culmination conditions at 604 ± 30 °C and pressure of 3.4 ± 0.4 Kbar at the depth of ca. 12-14 km. Garnet-sillimanite assemblage without foliation had been formed, representing one dominant Precambrian metamorphic event. The rock cooling from the peak conditions was slow ($1.2\text{--}3.3$ °C/Ma), as calculated from garnet concentration profile and might have lasted ca. 30-80 Ma. However, the final process of the cooling and uplift was more rapid when closing P-T conditions (515 ± 15 °C and 2.5 ± 0.3 Kbar) were established. No retrograde fluid infiltration that could have changed the peak assemblage has been observed. Paragneisses are considered to represent a low pressure portion of the cratonic crystalline complex.

Key words: Precambrian metamorphism, thermobarometry, cooling, metapelites, Palmas, Brazil.

Introduction

Many uncertainties surround the metamorphic processes in the crust, their later cooling and uplift from the depth to the positions where they are currently exposed. Paragneisses are particularly frequent in Precambrian terranes and are generally thought to represent the exhumed roots of the old mountain belts complexes and the associated metamorphism is considered to be the result of burial during crustal thickening.

Detailed studies however point at more complex development of paragneisses than the simplified crustal thickening and exhumation. Many paragneisses develop under significant perturbation of "normal" continental geotherm with different intensity and extent. Lower pressure paragneisses may also be formed, in some cases, with andalusite and cordierite occurrence in rocks. Some Precambrian metamorphites show high pressure isobaric cooling generally accepted as the evidence of prolonged residence in the lower crust after the peak temperature mineral assemblage had been formed and deformation later followed. But many data suggest that in some Precambrian terranes deformation continued during cooling at constant or slightly increasing crustal thickness (Phillips & Wall, 1981; Warren, 1983; Clarke et al., 1987). The crustal stability that apparently postdates metamorphic processes in many metamorphic terranes contrasts with the isothermal, rapid uplift of metamorphic rocks from the roots of many Phanerozoic orogenic belts (see e. g. Selverstone et al., 1984), where the effects of extension, erosion, or both seem to terminate the metamorphic process (England & Richardson, 1977; England, 1987). The crustal stability after metamorphism and deformation is that the synmetamorphic deformation did not result in the

overthickened crust and the rocks in these terranes were probably uplifted to the Earth's surface by later events which were casually unrelated to thermal metamorphism. The general cause of high grade metamorphism in many terranes must be external to the rocks we observe, and most probably external to the crust (Vernon et al., 1990). Thermal perturbations required for the high grade facies formation are established with difficulty by just an overthickening of the crust (Thompson & England, 1984). In many events they would occur late in the evolution of a mountain belt during the isothermal decompression of deeply buried rocks. Although the remnants of an eroded mountain chain may cool isobarically from a late thermal peak of metamorphism (Thompson & England, 1984; Ellis, 1987). The thinning, deformation and peak metamorphism in many terranes preclude such a tectonic setting. The heat contribution to a terranes by magmas is often presented as a direct cause of some high grade metamorphism. However, the magmas commonly do not appear to be abundant enough or they are not synchronous with the metamorphic peaks (Wells, 1980).

Field occurrence

Metapelitic garnet gneisses occur near capital of Tocantins, Palmas, along the banks of the creek Corrente Mirindiba. Measured dip direction and dip was $156/45^\circ$. Rock occurred also as isolated blocks of different sizes, maximum 3-5 m in diameter, within the field area given by the coordinates: W- $47^\circ 50'$ S- $10^\circ 10'$; W- $48^\circ 00'$ S- $09^\circ 50'$; W- $48^\circ 20'$ S- $10^\circ 20'$ (see Fig. 1a,b).

Studied gneisses occurrence is located in the Proterozoic crystalline basement near the western edge of post orogenic Paleozoic-Mesozoic Parnaíba (Maranhão) sedi-

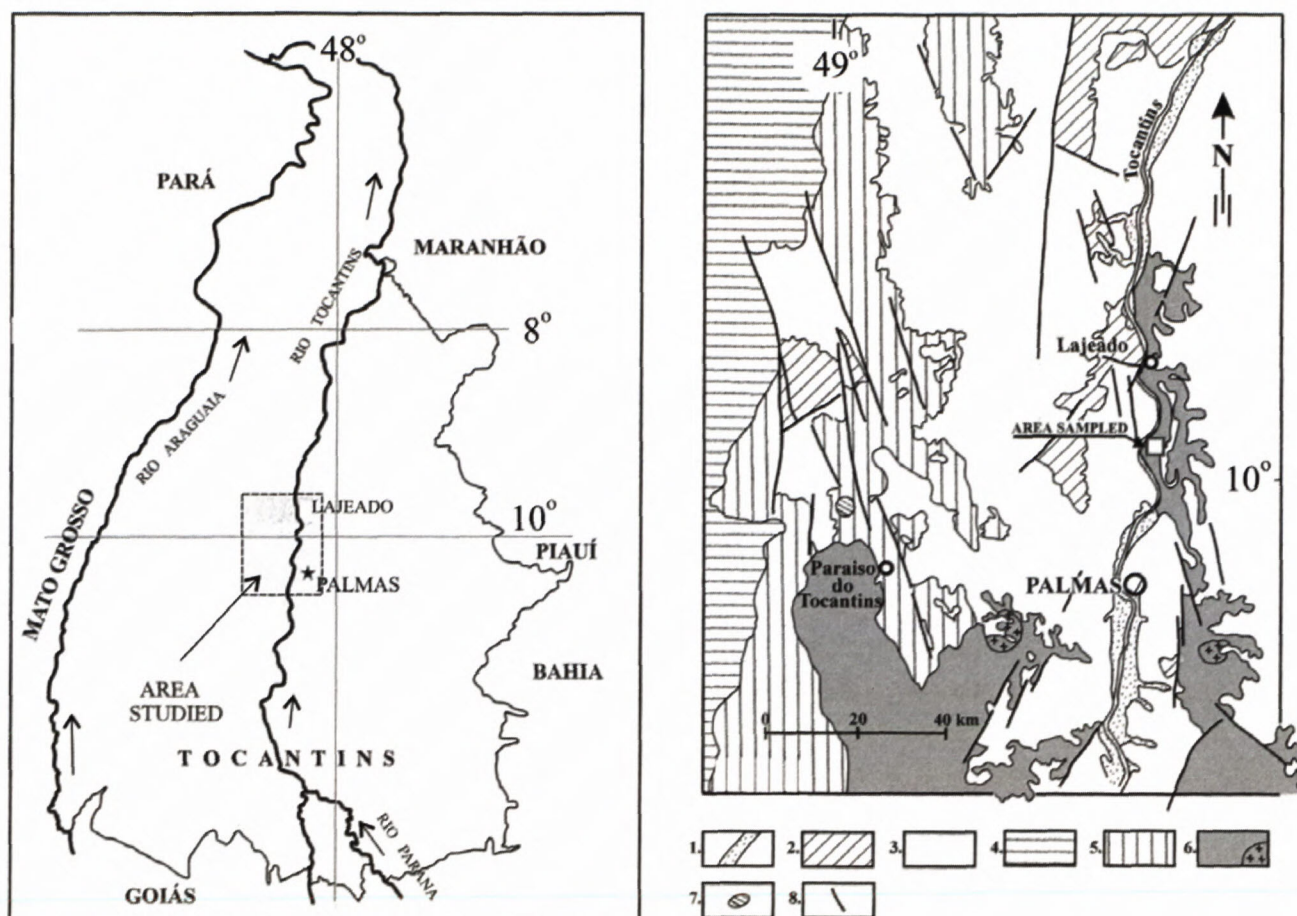


Fig. 1a,b. Schematic geological map of the studied area compiled from geological map *Projeto Radambrasil*, 1981. 1 - alluvial sediments of the Tocantins river (Quaternary), 2 - arenites, silts, locally conglomerates (Carboniferous), 3 - arenites, silts, conglomerates (Silurian-Devonian), 4 - metasediments of the Tocantins Group (Middle Precambrian), 5 - metamorphic rocks of the Araxa Group (Middle Precambrian), 6 - gneisses and granites of the Gioano Complex (Lower Precambrian), 7 - igneous alkaline rocks, 8 - airborne faults. Sampled area locations - banks of the Corrente Mirindiba.

mentary basin. Sediments of the inverted Parnaíba basin currently forming elevated plateau overlie São-Francisco crystalline shield, which is from the western border rimmed by the Neoproterozoic Tocantins mobile belt. The crystalline Goiania Complex (*Projeto Radambrasil*, 1981), is composed of metapelitic garnet gneisses, migmatites, amphibolites, quartzites, granites, pegmatites, granulites, marls, sporadically basic and ultrabasic intrusive bodies.

Garnet gneisses are mostly massive with no distinct foliation and some isolated blocks have a lensoid appearance. Some may be characterized by an inhomogeneous structure, reflecting probably the protolithic compositional changes.

Petrography and structure of the garnet-sillimanite paragneisses

The outcropping rocks show small differences in mineral modes. They are medium to coarse grained with semipelitic composition and exhibit granoblastic to lepidoblastic texture. No banded structures composed of leucocratic and melanocratic parallel bands are present.

The main minerals of the studied gneisses are: sillimanite (Sil, ~18 mod.%), garnet (Grt, ~8 mod.%), biotite (Bt, ~40 mod.%), plagioclase (Pl, ~10 mod.%) and quartz (Qtz, ~22 mod.%). Sil and Bt are dominant mineral phases. Rare postculmination minerals with quantitative amount less than 1 mod.% are andalusite (And) and cordierite (Crd). In accessory amount there are present: ilmenite (Ilm), zircon (Zrn), apatite (Ap) and chlorite (Chl).

Garnet porphyroblasts up to ~1 cm in diameter are regularly scattered in the rock matrix and are microscopically homogeneous with occasional poikilitic minute inclusions of Bt, Pl and Qtz. No significant inclusions were observed within garnet, that may serve as a strong reminiscence of prograde breakdown of e.g. chloritoid, staurolite, deciphering thus the progressive pre-culmination metamorphic stages with the Grt+Sil equilibrium assemblage. Garnet rims are in few cases slightly chloritized as the consequence of the restricted post-culmination retrograde reactions. At the edges and interstitions no signs of development of the symplectite textures have been observed.

On the basis of the mode of appearance, only one type of biotite is distinguishable forming the equilibrium cul-

mination matrix with the other felsic minerals. Biotite as a dominant phyllosilicate shows no sub-parallel alignment, flakes' orientation and do not represents distinctive rock schistosity. As the biotite is abundant and pyroxene is still missing, the transition of rocks from upper amphibolite to granulite facies conditions has not been attained.

The presence of Al_2SiO_5 minerals in rocks is important as their stabilities are controlled by temperatures and solid pressures (see e.g. Richardson et al., 1969; Holdaway, 1971). The presence of metastable Al_2SiO_5 relics is particularly important for inferring earlier metamorphic P-T conditions and assessing the metamorphic trajectory. No pseudomorphs after andalusite, staurolite or chloritoid reaction rests have been found, indicating the progressive passing into the sillimanite stability field.

The coexistence of Grt and Sil is important in order to estimate the peak metamorphic condition on the basis of their structural appearance. This assemblage is commonly observed in the region, without primary or secondary muscovite, except some vanishing retrograde modal amount. If K-feldspars would have been present that might have formed an argument that muscovite was not more stable in the presence of quartz. A few fine muscovite and chlorite grain flakes seldom develop at matrix mineral contacts and are considered to be produced during the retrogression with restricted hydration.

The rock structure, common occurrence of Grt+Sil+Qtz assemblage and decompression reaction $\text{Grt} + \text{Sil} + \text{Qtz} = \text{Crd}$ would be responsible for rare cordierite presence. Microscopic reaction domains, restricted garnet and sillimanite decomposition reaction features support this presumption. Some matrix domains favour tiny cordierite development at the grain contacts of Bt+Qtz and Bt+Sil.

Consequently, this cordierite microscopic appearance and its crossing of the biotite culmination grains forms an argument for its post-peak development during decompression processes. In few cases, minute chlorite grains

participate on the Grt decomposition. Only some muscovite flakes are associated with plagioclase at their grains interstitions. At some garnet edges, tiny aggregates of muscovite and chlorite indicate the restricted complex retrograde reactions with a minimum extent. Thus, the thermal culmination mineral assemblage was almost untouched by the later retrograde reactions.

The garnet grains have no significant mineral relics useful for reasonably interpretation of early stages of recrystallization, what makes difficult to infer the preculmination metamorphic trajectory of these rocks on the basis of microscopic mineral appearance. The post-culmination retrograde processes producing tiny amounts of andalusite, cordierite, chlorite and muscovite have left the peak mineral assemblage unhydrated.

The paragenetic and structural relationships among garnet, biotite and sillimanite in the studied rock domains suggest, that the peak temperature and pressure formed the well equilibrated mineral assemblage.

Geothermobarometry, diffusion and garnet crystal size data

Chemical analyses of minerals were carried out on CAMECA SX 100 electron microprobe analyser with an accelerating voltage of 15 kV, 20 nA beam at GÜDŠ (Dr. P. Konečný, Bratislava, Slovakia). The analyses of garnets, biotites and plagioclases are given in Tab.1.

Chemical composition of the garnet porphyroblasts (Tab.1) was determined in five rock samples, where mineral equilibrium domains were confirmed. The garnet grains are unzoned, almost homogeneous and the compositional uniformity suggests that diffusion homogenization was realized during metamorphic thermal culmination of this terrane. The compositional changes at the garnet edges and mineral grain contacts in the analysed domains are ascribed to the post-culmination processes

Tab. 1. Chemical analyses of coexisting garnets, biotites and plagioclases from Palmas paragneisses.

GARNET ANALYSES

N°	1.	2.	3.	4.	5.	6.	7.	8.	9.	10.	11.	12.
Sample	I.1c	I.1r	I.3c	I.3c'	I.3r	II.2c	II.2r	II.3c	II.3r	II.4c	II.4c'	II.4r
SiO ₂	36.66	37.52	37.18	37.14	36.56	35.65	36.71	36.91	36.78	36.78	36.71	36.67
TiO ₂	0.00	0.02	0.00	0.01	0.03	0.00	0.03	0.01	0.05	0.08	0.07	0.00
Al ₂ O ₃	20.66	20.88	20.32	20.31	20.18	19.85	20.19	20.41	20.23	20.35	20.38	20.38
Cr ₂ O ₃	0.00	0.00	0.00	0.03	0.02	0.04	0.07	0.01	0.00	0.05	0.03	0.00
FeO	37.60	37.67	37.39	37.57	37.55	37.24	37.54	38.08	37.30	37.00	37.76	38.09
MnO	1.65	1.35	1.45	1.49	1.56	1.70	1.99	1.56	1.65	2.30	1.22	2.08
MgO	2.26	2.04	2.11	2.32	1.91	2.27	1.61	2.58	1.90	1.45	2.63	1.57
CaO	1.12	1.57	1.74	1.25	1.71	1.15	1.62	0.92	1.68	1.25	0.89	1.61
Total	99.95	101.05	100.19	100.12	99.52	97.90	99.76	100.48	99.59	99.66	99.73	100.40
Cations on 12 oxygen basis												
Si	2.986	3.014	3.018	3.016	2.998	2.977	3.007	2.993	3.009	3.020	2.994	2.991
Ti	0.000	0.001	0.000	0.000	0.002	0.000	0.002	0.001	0.003	0.004	0.004	0.000
Al	1.984	1.977	1.944	1.944	1.951	1.954	1.949	1.951	1.951	1.969	1.959	1.960
Cr	0.000	0.000	0.000	0.002	0.001	0.002	0.004	0.001	0.000	0.003	0.002	0.000
Fe	2.562	2.530	2.538	2.551	2.575	2.601	2.572	2.582	2.551	2.541	2.576	2.599
Mn	0.114	0.092	0.099	0.103	0.109	0.120	0.138	0.107	0.115	0.159	0.084	0.144
Mg	0.275	0.244	0.256	0.281	0.233	0.283	0.197	0.312	0.232	0.177	0.320	0.191
Ca	0.098	0.135	0.151	0.109	0.150	0.103	0.142	0.080	0.147	0.109	0.077	0.141

Tab. 1 (continued)

PLAGIOCLASE ANALYSES

N°	1.	2.	3.	4.	5.	6.	7.	8.	9.	10.
Sample	I.1m	I.1r	I.3m	I.3r	II.2m	II.2r	II.3m	II.3r	II.4m	II.4r
SiO ₂	61.25	60.37	61.20	60.50	60.44	59.08	60.40	59.63	59.71	59.81
Al ₂ O ₃	24.32	25.04	24.31	25.00	24.89	25.71	25.60	25.33	25.36	25.35
FeO	0.59	0.02	0.59	0.02	0.03	0.42	0.05	0.03	0.03	0.05
CaO	5.36	6.36	5.32	6.36	6.48	7.37	4.48	6.82	6.77	6.81
Na ₂ O	8.30	8.00	8.32	8.00	7.86	7.44	8.91	7.79	7.84	7.86
K ₂ O	0.12	0.05	0.03	0.03	0.04	0.03	0.04	0.04	0.03	0.04
Total	99.94	99.86	99.77	99.91	99.74	100.05	99.44	99.67	99.74	99.92
Cations on 8 oxygen basis										
Si	2.724	2.688	2.725	2.691	2.693	2.638	2.666	2.665	2.666	2.666
Al	1.274	1.314	1.276	1.311	1.307	1.353	1.332	1.334	1.334	1.332
Fe	0.022	0.000	0.021	0.000	0.001	0.015	0.002	0.001	0.001	0.002
Ca	0.255	0.303	0.253	0.303	0.309	0.353	0.212	0.326	0.323	0.325
Na	0.716	0.690	0.719	0.690	0.679	0.644	0.763	0.675	0.678	0.679
K	0.007	0.002	0.001	0.002	0.002	0.002	0.002	0.002	0.0011	0.002

BIOTITE ANALYSES

N°	1.	2.	3.	4.	5.	6.	7.	8.	9.	10.
Sample	I.1m	I.1r	I.3m	I.3r	II.2m	II.2r	II.3m	II.3r	II.4m	II.4r
SiO ₂	33.63	35.08	34.18	34.69	33.21	33.72	33.86	34.67	33.70	33.85
TiO ₂	2.98	1.92	2.49	2.50	2.22	2.92	1.92	2.39	2.82	2.08
Al ₂ O ₃	18.13	20.04	19.23	19.36	19.13	18.59	19.19	18.87	19.32	19.26
Cr ₂ O ₃	0.09	0.05	0.12	0.14	0.13	0.24	0.08	0.09	0.09	0.18
FeO	23.02	21.31	21.62	21.14	22.53	22.75	22.30	21.55	21.90	23.85
MnO	0.08	0.05	0.02	0.00	0.02	0.06	0.04	0.05	0.01	0.07
MgO	7.92	8.27	7.63	7.65	7.00	7.19	7.26	7.78	6.93	7.03
CaO	0.01	0.03	0.00	0.01	0.03	0.03	0.01	0.03	0.05	0.01
Na ₂ O	0.19	0.33	0.37	0.36	0.36	0.34	0.34	0.33	0.29	0.28
K ₂ O	9.01	8.61	8.63	8.57	8.63	8.42	8.42	8.43	8.52	7.99
F	0.00	0.04	0.13	0.13	0.15	0.25	0.18	0.00	0.00	0.65
Cl	0.11	0.10	0.11	0.13	0.10	0.09	0.09	0.11	0.10	0.09
Total	95.21	95.83	94.53	94.31	93.34	94.81	93.69	94.30	93.73	95.34
Cations on anhydrous 22 oxygen basis										
Si	5.240	5.334	5.297	5.319	5.251	5.254	5.310	5.352	5.279	5.240
Ti	0.349	0.220	0.291	0.291	0.264	0.342	0.227	0.277	0.332	0.242
Al	3.330	3.591	3.512	3.532	3.564	3.415	3.548	3.433	3.566	3.513
Cr	0.011	0.006	0.015	0.018	0.016	0.030	0.010	0.012	0.011	0.022
Fe	3.000	2.710	2.803	2.736	2.979	2.965	2.924	2.912	2.869	3.087
Mn	0.011	0.006	0.003	0.000	0.003	0.008	0.006	0.007	0.002	0.009
Mg	1.840	1.875	1.764	1.765	1.649	1.671	1.698	1.751	1.619	1.622
Ca	0.002	0.005	0.000	0.002	0.005	0.005	0.001	0.005	0.008	0.002
Na	0.059	0.098	0.112	0.110	0.111	0.104	0.105	0.100	0.088	0.085
K	1.792	1.670	1.708	1.692	1.699	1.716	1.684	1.660	1.702	1.579

associated with decompression uplift and cooling of the whole metamorphic complex. The garnet porphyroblasts remained with idioblastic morphology and restricted re-sorption.

Almandine component in garnets is ranging considerably high for metapelitic paragneisses. Almandine \approx 83.6–84.5 mol.%, pyrope \approx 6.0–10.4 mol.%, spessartine 3.0–4.6 mol.% and grossular \approx 3.2–5.1 mol.%. Postculmination diffusion rims of most garnet grains are also slightly richer in almandine and spessartine component than the cores. The grossular component is almost constant (Fig. 2). At the garnet edges no symplectitic features and garnet overgrowth were developed.

Matrix biotite has a stable chemical composition within the sample and thus documents one dominant equilibrium recrystallization event during culmination of metamorphic process.

Plagioclases appear to be microscopically homogeneous. Lamellar twinning is not frequent. The microprobe analyses of discrete plagioclase grains (Tab. 1) in the rock domains reveal a weak compositional zoning. The composition of matrix plagioclase and at the plagioclase grain edge against garnet has been used for pressure calculations of the culmination and closing retrograde processes.

Fig. 2. Biotite-garnet grain edge. The composition profile documents the relative homogeneity of the garnet and the continuous slow cooling process. The concentration change of Mg and Fe in garnet when normalized as the function of garnet size depends on type of cooling and time. The schematic approach presupposes that Mg and Fe redistribution process has been formed by effective diffusion in the post-culmination cooling stage and is independent on the given Mg and Fe concentration. Rim diffusion penetration distance is strictly dependent on temperature and time.

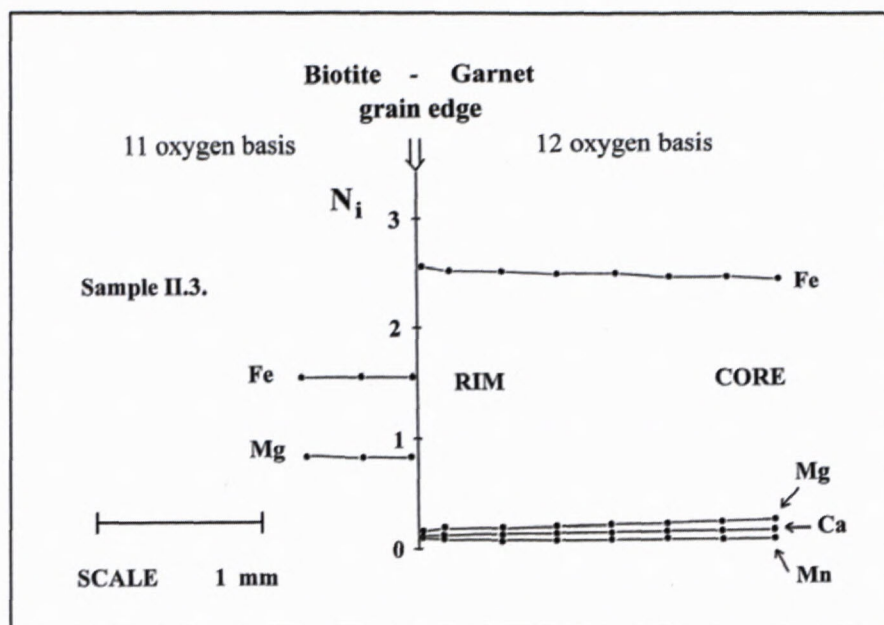
In order to calculate the metamorphic culmination recrystallization temperatures, the reaction Almandine + Phlogopite = Pyrope + Annite with calibrations of different authors was applied and the obtained temperature data are listed in Tab. 2. Consequently, the prograde culmination, as well as the closing retrograde temperatures are presented with respect to their scattering. Culmination temperature data are consistent with the mineral assemblages and sillimanite presence in the samples. The P-T calculations used different calibrations and THERMOCALC software of Powell & Holland (1988, 1998) and Powell et al. (1998). See Tabs. 2 and 3 for details.

Since the Grt-Bt geothermometer is based on the Fe-Mg exchange reaction, Ferry & Spear (1978) give the thermometrical limits for additional components as Ca, Mn in garnets $(Ca+Mn)/(Ca+Mn+Mg+Fe)$ up to ≈ 0.2 and coexisting biotites $(Al^{VI}+Ti)/(Al^{VI}+Ti+Mg+Fe)$ up to ≈ 0.15 . The studied garnets satisfy well this composition requirements with limiting values of 0.06-0.09. However, the high aluminous biotites with compositional value 0.37-0.41 overstep this limit and affect the calculated temperature values given in Tab. 2.

Differences in calculated P-T data are mainly due to the calibration used and non-ideality of the mineral solid solutions. The high Ti and Al^{VI} contents of biotites may be responsible for the higher calculated temperatures obtained with $\ln K_D^{MgFe}$ v. T relation calibrated by Ferry & Spear (1978) and confirms the Thompson's (1976) calibration suitability applied to high grade metamorphic rocks (Essene, 1982; Indares & Martignole, 1985).

For geobarometric purposes the reaction Plagioclase = Grossular + Sillimanite + Quartz has been used and different calibrations adopted. The pressure calculation results are summarized in Tabs. 3 and 4. The peak temperature and the corresponding retrograde closure temperature for particular rocks (Tab. 2) are assumed to be the calculated average temperatures given by used calibrations.

The garnets edges are modified by penetration diffusion and the central parts are almost chemically homogeneous (see Fig. 2). Thus, the calculated pressures at thermal culmination and at retrograde thermal minimum



are presumed to be reasonable. The consequent construction of metamorphic post-culmination trajectory, presented graphically in Fig. 4, is based on calculated peak progressive culmination and closing retrograde P-T data.

The garnet post-culmination diffusion penetration rims (see Fig. 2) may serve for calculation of an approximate time span, that was needed for the diffusion penetration at the given retrograde P-T conditions.

Crank (1975) defines the relation among diffusion penetration distance X , diffusion coefficient $D_{(T,P)}$ and time (t) by the expression

$$X = \sqrt{4Dt} \quad (1)$$

The diffusion coefficient for Fe-Mg interdiffusion in garnet was calculated for specific P and T conditions according to the expression:

$$D_{(T,P)} = D_0 e^{[(\Delta E^* + (P-1) \cdot \Delta V^*)/RT]} \quad (2)$$

where D_0 is the preexponential factor, ΔE^* is the activation energy for Fe-Mg interdiffusion and ΔV^* is the activation volume.

Using the retrograde P-T data, 2.5 Kbar at 515-525 °C and measured penetration distance in garnet $\sim 14 \pm 2 \mu m$, the time required for penetration has been calculated in the considerably wide range ca. 9-70 Ma (see Tab. 5). Maximum diffusion penetration into a garnet will be given by time during which a rock has been exposed to higher temperatures. Speedy cooling of rock complex (say 80-120 °C/Ma) may restrict the diffusion significantly which will be limited by the small penetration distances. Slow cooling of $\sim 0.05-1$ °C/Ma, from higher culmination temperatures, causes the longer diffusion penetration distance in crystal. Temperature higher than ~ 750 °C causes rapid garnet homogenization. Thus the development of garnet rim composition is strictly temperature-time controlled phenomena.

The activation energy for diffusion plays the crucial role in penetration distance calculation and in the cooling rate approximate calculation. Using the obtained P-T data, garnet size, normalized concentration profile and

Tab. 2. Metamorphic recrystallization temperatures of basement paragneisses from Palmas. *

		lnK _D	T	F&S	N&H	G&S	H&L	Tn
I. 1.)	c	1.742	560	573	586	590	575	579
	r	1.967	505	497	514	521	520	523
3.)	c	1.732	561	570	585	586	572	579
	r	1.961	506	498	517	528	522	524
2.)	c	1.626	589	608	621	625	597	599
	r	1.995	499	489	507	529	515	517
II. 3.)	c	1.569	605	629	640	635	611	612
	r	1.955	508	500	518	530	523	539
II. 4.)	c	1.511	621	652	662	653	626	622
	r	1.965	505	497	515	540	521	537

*Based on the reaction: Almandine + Phlogopite = Pyrope + Annite

using equilibrium calibrations of: T - Thompson (1976); F&S - Ferry and Spear (1978); N&H - Newton and Haselton (1981); G&S - Ganguly and Saxena (1984); Holdaway and Lee (1977); Thoenen (1989). *c* - culmination temperatures of metamorphic recrystallization, *r* - retrograde closure temperatures.

Tab. 3. Metamorphic recrystallization pressures of basement paragneisses from Palmas. *

aAn	aGr	G	N&H	P&H	P&H ^P		
I. 1.)	c	0.4440	0.0394	3.67	2.98	3.93	3.52
	r	0.5950	0.0500	2.84	1.96	3.13	2.71
I. 3.)	c	0.4387	0.0396	3.65	3.12	4.05	3.55
	r	0.6021	0.0552	3.09	2.19	3.38	2.87
II. 2.)	c	0.5065	0.0364	3.13	2.64	3.45	3.21
	r	0.6907	0.0494	2.34	1.31	2.49	2.23
II. 3.)	c	0.3334	0.0257	3.46	2.96	3.76	3.44
	r	0.6255	0.0532	2.98	2.09	3.24	2.87
II. 4.)	c	0.5129	0.0391	3.70	3.13	3.87	3.69
	r	0.6222	0.0495	2.75	1.84	2.98	2.66

* Based on the reaction: 3 Anorthite = Grossular + 2 Sillimanite + Quartz

using equilibrium pressure calibrations of: G - Ghent et al. (1979), N&H - Newton & Haselton (1981), P&H - Powell & Holland (1988), P&H^P-THERMOCALC, Powell & Holland (1988, 1998). Activity of anorthite (*aAn*) and grossular (*aGr*) is based on formulation of Newton & Haselton (1981). *c* - pressures of metamorphic recrystallization at thermal maximum. *r* - pressures at retrograde temperature minimum.

Tab. 4. Composition and pressure characteristics of basement paragneisses from Palmas using the calibration of Hoisch (1990, 991) and Koziol & Newton (1988). *

lnK _D R1	lnK _D R2	P _{R1}	P _{R2}	ΔV(J/bar)	K&N*		
I. 1.)	c	6.667	4.965	3.20	3.23	-0.2973	2.91
	r	6.189	4.285	2.61	2.63	-0.2491	2.39
I. 2.)	c	6.065	4.368	3.96	3.84	-0.2866	3.14
	r	6.112	4.219	2.91	2.91	-0.2285	2.73
II. 2.)	c	6.441	4.854	3.12	3.10	-0.2945	2.71
	r	6.891	4.961	2.09	2.15	-0.2439	1.87
II. 3.)	c	5.873	4.333	4.13	4.08	-0.3075	3.78
	r	6.347	4.458	2.69	2.73	-0.2321	2.52
4.)	c	6.658	4.613	2.82	3.81	-0.2883	3.36
	r	6.630	4.727	2.40	2.48	-0.2491	2.26

*R1: 1/3 Pyrope + 2/3 Grossular + Eastonite + 2 Quartz = 2 Anorthite + Phlogopite

R2: 1/3 Almandine + 2/3 Grossular + Siderophyllite + 2 Quartz = 2 Anorthite + Annite

ΔV - partial molar volume change; K&N* - calibration according to Koziol & Newton (1988).

Tab. 5. Diffusion penetration distance calculations for garnets from Palmas paragneisses.

	ΔE*[KJmol ⁻¹]	P[Kbar]	T [°C]	X [μm]	D ⁵¹⁵ [cm ² /s]	D ⁵²⁵ [cm ² /s]	t[Ma]	t[Ma]
I*	293.1	2.5	515 - 525	14	1.16x10 ⁻²¹	2.03x10 ⁻²¹	13	9
II*	273.8	2.5	515 - 525	14	2.21x10 ⁻²²	3.74x10 ⁻²²	70	41

I* according to activation energy [ΔE*] of Lasaga et al. (1977) and II* of Chakraborty & Ganguly (1992).

diffusion data, the petrological cooling rate estimates have been assessed (Tab. 6).

Tab. 6. Calculated petrological cooling rate estimates for basement paragneisses from Palmas *.

Sample:	II.3.
T [°C]:	624 ± 15
P [Kbar]:	3.65 ± 0.40
a [mm]:	3.41
	Diffusion coefficient
D _[T,P] :	7.26 × 10 ⁻¹⁹
D _[T,P] :	2.45 × 10 ⁻¹⁹
γ':	170
	Cooling rate estimates [°C / Ma]
S (E):	3.3
S (L):	1.2

* calculated for metamorphic culmination P-T conditions using the activation energy for diffusion (ΔE^*) according to E - Elphic et al., (1985) - $\Delta E^* = 224.4$ [KJmol⁻¹] and L - Lasaga et al. (1977) - $\Delta E^* = 293.1$ [KJmol⁻¹]. Cooling rate [S] has been calculated according to formula of Lasaga (1983), $S = RT^2 \gamma' / \Delta E^* a^2$, where γ' is given by the shape of the normalized measured composition profile and a is the garnet size.

If the calculated cooling rate estimates (~ 1.2 and 3.3 °C/Ma) have been accepted, then the linear cooling period from culmination to retrograde temperature conditions would have lasted approximately from ~ 27 to 80 Ma. This is relatively slow cooling if considering e.g. Spanish Betic metamorphic terrane with cooling rates of 200 °C/Ma (Zeck et al., 1992) or 150-350 °C (Monié et al., 1994) and corresponding exhumation rate of 5-10 km/Ma. Exhumation and cooling rates of Dora Maira terrane are 22 km/Ma and ca. 90 °C/Ma (Gebauer, 1996). Thermobarometric data and normalized concentration garnet zonality of some Variscan basement paragneisses of the Western Carpathians indicate cooling rate from ~ 4 up to 90 °C/Ma (Dyda, 2002).

Detailed analyses of crystal size distribution (CSD) data in chemical and technological processes (see e.g. Avrami, 1939; Randolph & Larson, 1971) attributed their wide application to different petrological studies of igneous and metamorphic rocks (see e.g. Kretz, 1973, 1974; Baronnet, 1984; Cashman, 1988, 1990; Marsh, 1988, 1998).

The CSD data for garnets in the unretrograded assemblage (Fig. 3) may have an importance for evaluation of the particular garnet growth crystallization process in this assemblage.

Plotting the population density $\ln(n)$ versus garnet crystal size (L) is usually a line with the slope $k/G\tau$ and intercept $\ln n_0$, where G is the average growth rate of a given mineral, τ is the growth residence time and k is the constant. As the average growth rate is known, the growth residence time may be calculated directly. The average nucleation rate (J) may be calculated from linear equation parameters given by the slope of the line - $\ln(n)$ versus size (L).

Crystal size distribution data of garnets from the assemblage II.3. (Fig. 3), give the calculated line, $\ln(n)$ versus size (L), determined by the slope: $b = 9.53$ [cm⁻¹] and intercept $\ln n_0 = 6.65$ [n/cm⁴]. Consequently, the calculated average nucleation rate is ~ 4 × 10⁻¹⁰ N/cm³s.

Nucleation rate (J) changes with metamorphic regimes. Contact metamorphism is characterized by the highest nucleation rates (e.g. $J \approx 5 \times 10^{-3}$) and high grade regional metamorphism with lower J values (ca. 3×10^{-8} - 4×10^{-6}). In regional metamorphic rocks of sillimanite zone from Waterville Formation (south-central Maine, USA) Cashman & Ferry (1988) give the average nucleation rate for garnets $J = 1.98 \times 10^{-6}$ and some Variscan regionally metamorphosed basement metapelites from the Western Carpathians have the garnet nucleation rate $J = 3 \times 10^{-8}$ - 1×10^{-7} N/cm³s (Dyda, 1997).

Thus, the obtained garnet crystal size data of the metapelitic gneisses from Palmas form an argument for the slow regional metamorphic pre-culmination heating rate, documented by the slow nucleation rate of the garnet.

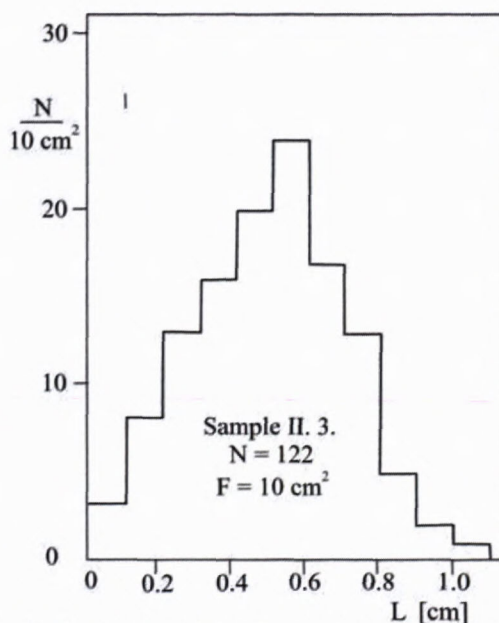


Fig. 3. Garnet crystal size frequency histogram in metapelitic gneisses from the Palmas area. The distribution data are used for garnet nucleation rate and growth estimates.

Results and discussion

The assessment of the thermal and pressure changes may form a basic frame for further collecting of petrological data. More observations of sequence and extent of metamorphic reactions on a larger field scale are needed for a better understanding of metamorphism in the region.

Metamorphic trajectory can be derived using different pieces of petrological evidence mainly based on the sequence of mineral assemblages, as the rock passes through the mineral stability fields during prograde metamorphism to the culmination P-T conditions. Subsequently, retrograde processes may occur, accompanied by

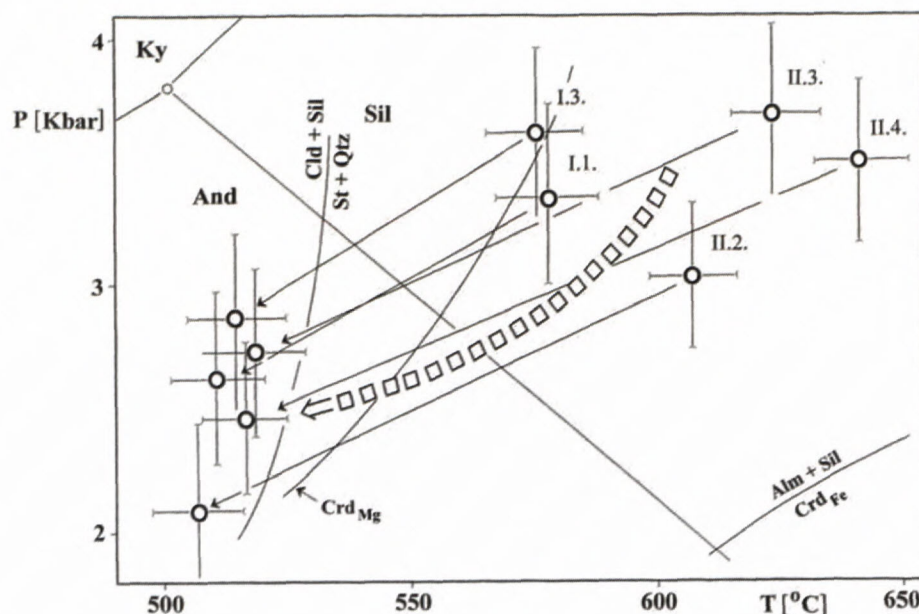


Fig. 4. The calculated metamorphic culmination and closing retrograde P-T characteristics of metamorphic recrystallization calculated from Grt-Bt-Plg-Sil-Qtz mineral equilibria. Temperature data were obtained on the basis of Grt-Bt calibrated equilibria (Ferry & Spear, 1978; Thompson, 1976; Newton & Haselton, 1981; Ganguly & Saxena, 1984; Hodges & Crowley, 1985). Pressure was calculated from calibrated equilibria Grt-Plg-Sil-Qtz (Ghent et al., 1979; Powell & Holland, 1988; Newton & Haselton, 1981) and equilibria including Grt-Bt-Ms-Plg-Qtz (Hirsch, 1990). The staurolite (St) and cordierite (Crd_{Mg} , Crd_{Fe}) appearance limits are taken from Bucher & Frey (1994).

changes in mineral assemblage. Heat adding intrusions often form early, followed by uplift that was caused by the rise of the crystalline domes (Schumacher et al., 1990).

The potential complexity of P-T paths makes generalization about their form difficult. Chamberlain & Karabinos (1987) have shown, that in structural terranes dominated by folding and thrusting, details of P-T paths can vary dramatically within a single metamorphic belt. This is due to the heat redistribution caused by folding of the isotherms and by differential cooling rates that are dependent on the rock's position in the structure following deformation. However, many models of geological processes and examples from metamorphic terranes (e.g. England & Thompson, 1984; Thompson & England, 1984) document, that P-T paths can be simplified and broadly classified as curved clockwise or curved counter-clockwise trajectories in P-T coordinates.

The field occurrence of garnet-sillimanite paragneisses from Palmas area, their recrystallization history inferred from microscopic studies and the P-T data suggest, that their petrological character can be useful for the tectono-metamorphic reconstruction of this terrane.

The paragneisses represent the culmination temperature of 604 ± 30 °C at pressure of 3.4 ± 0.4 Kbar with ca. 12-14 km of the overlying pile at one dominant metamorphic event. No data have been collected to classify and estimate the metamorphic intensity gradient with respect to the territorial scale.

The prograde P-T data form an important clue to the definition of the culmination of metamorphic conditions. However, the pre-culmination trajectory is not yet well constrained. The absence of inclusions in garnet and the coexistence of matrix phases did not allow to assess the prograde whole rock reactions. Neither muscovite nor

K-feldspars have been revealed as possible dehydration reaction participants and temperature indicators. The post-peak decompression and retrograde closing P-T conditions have been determined in rather narrow temperature range, from 604 ± 30 °C for culmination tempera-

tures, to 515 ± 15 °C for retrograde closing temperatures (Fig. 4). Symplectitic features, simple corona textures, resorbed or embayed garnets are absent. These textural features, together with the garnet rim zonality and limited retrograde hydration, indicate the slow cooling in depth of the gneissic complex from Palmas. Cooling generally attended decompression on the retrograde path and calculated P-T data involve decrease in pressure from 3.4 ± 0.4 to 2.5 ± 0.3 Kbar, indicating thus slow cooling and uplift. However, typical "isobaric cooling" gradients have rather flat dP/dT slopes ca. 0.3-0.5 Kbar per 100 °C (Harley & Hensen, 1990). The cooling retrograde processes led to the closing temperature and pressure conditions of 515 ± 15 °C and 2.5 ± 0.3 Kbar respectively. The post-peak mineral inter-granular equilibrium adjustment is considered to be a diffusion process of one dominant terminating metamorphic event.

In the studied paragneisses, sillimanite is the common Al-silicate mineral. Andalusite, or pseudomorphs after andalusite have not been found. Thus, it is presumed, that these paragneisses passed first through the kyanite stability field during the prograde metamorphism. Subsequent increasing temperature drove these rocks into sillimanite stability field. Avoiding probably the andalusite field, they are indicated by the clockwise curved P-T trajectory. The geothermobarometrical data also support a clockwise trajectory and imply terms of crustal thickening, compression prior to the reaching of the thermal and pressure maximum and subsequent slow cooling and decompression.

On the other hand, some cratonic metamorphic terranes characterized by isobaric cooling (e.g. Adirondacs, Namaqualand, Broken Hill) have demonstrated that evolved their thermal maximum along a counter-clockwise P-T path (see e.g. Bohlen, 1987; Waters, 1986). Such counter-clockwise P-T-t paths would be consistent with magmatic and/or extensional settings (Sandiford & Powell, 1986).

The studied garnet-sillimanite gneisses are located at the westernmost margin of the São Francisco craton, rimmed by the Neo-Proterozoic mobile belt.

Mobile belts of Brazil evolved during juxtaposition and amalgamation of single continental fragments of Gondwana supercontinent in Brasiliano tectonic cycle 950–600 Ma ago (Brito-Neves & Cordani, 1991; Brito-Neves et al., 1999). Tectonic style of mobile belts is typical for accretionary wedge structure developing due to closing of Goiás ocean (Pimentel & Fuck, 1992). Tocantins fold and thrust belt comprise strongly folded and sheared 1 200–600 Ma old rocks. Ensimatic units, relics of oceanic crust (Pimentel & Fuck, 1992; Viana & Pimentel, 1994, ex Strieder & Suita, 1999) and granulite metamorphism of the mafic-ultramafic complexes (Suita et al., 1994) are related to this event. After final closing of Goiás ocean, the continental collision of São Francisco and Amazonian shield started, probably 800 Ma ago (Strieder & Suita, 1999). CC collision 650–550 Ma ago imprinted to Tocantins mobile belt tectonic style of accretionary wedge. Contemporaneously, I-type to A-type granitoids intruded into the Neo-Proterozoic gneisses (Almeida et al., 1973, ex Nairn & Stehli, 1973). The Proterozoic Goiania Complex, where metapelitic garnet-sillimanite gneisses come from, tectonically belongs to the Goiás Median massif (sensu Strieder & Suita, 1999), triangular zone resulting from CC collision of Amazonian with São Francisco plate during Brasiliano orogeny. The last tectonic deformation phases took place 500 Ma ago (Strieder & Suita, 1999). The Pre-Cambrian crystalline complex has not been significantly disturbed since Silurian, but it is affected by younger deep seated warping and faulting (Fairbridge, 1975). During the Lower Tertiary, important block faulting in many areas took place, influencing thus geomorphology of the recent surface.

The gneissic complex from Palmas was isostatically and thermally well equilibrated. These results imply, that the paragneisses, presently outcropped, were buried at ca. 12–14 km depth throughout the long Precambrian time interval. Thus the presumed crustal residence is inferred for this gneissic terrane as a “true craton”, at least until its local and marginal remobilization and partial exhumation in late Proterozoic and younger tectonic events. The petrological features and P-T data may lead us to the conclusion that the metapelitic gneisses from Palmas are a low pressure portion of the cratonic crystalline complex. This is consistent with the data of DaCruz & Kuyumjian (1998) who consider this terrane to be subjected to medium pressure and variable temperatures up to amphibolite facies.

The rocks do not possess the synkinematic crystal growth (e.g. rotational garnets, etc.) or deformation features. They represent one dominant regional pre-Cambrian metamorphic thermal event. They cooled slowly isostatically in depth and in the final stages they were, presumably, rapidly uplifted. They represent the fragmented rigid rests of the western marginal part of the São Francisco craton.

The uplift movements were associated with the activity of the Late-Proterozoic mobile belt and fault tectonic activities between Amazonian and São Francisco cratons.

There is an absence of the deformational and/or further metamorphic events related to any tectonic activities within the studied rocks. Thus we presume, that reactiva-

tion connected with uplift of the Goiás Median complex was not significantly affected by the dynamo-metamorphic events, at least in the studied region.

Thus the gneissic complex remained unaffected by the post-metamorphic geological activities since Precambrian. This mechanism of stability should be one of the basic geological problems for the future studies as well as for geochronological investigation when considering the tectonic and metamorphic history of the gneissic complex from Palmas.

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