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The Imataca Complex, NW Amazonian Craton, Venezuela: Crustal evolution and integration of geochronological and petrological cooling histories

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SHRIMP U/Pb-zircon data and Nd mean crustal residence ages indicate that the Imataca Complex developed from an Archean (≥3.2 Ga) continental protolith which has undergone considerable isotopic disturbance plus and juvenile accretion during late-Archean (~ 2.8 Ga) times. Transamazonian granulites experienced peak metamorphic conditions of 750 – 800 °C, 6 – 8 kbar with associated transpressive thrusting and tectonic imbrication. Geochronology on zircon, pyroxene and garnet constrains the timing of peak metamorphism at 1.98 – 2.05 Ga. Diffusion modeling of Fe-Mg exchange between biotite inclusions and host garnet yields (near metamorphic peak) cooling rates of 50 – 100 °C/Ma, with petrological cooling rates being generally consistent with cooling rates determined from geochronology. Combining the retrograde P-T path with cooling rates suggests that after the metamorphic peak, large portions of the Imataca Complex were exhumed from 30 to 17 km at a rate of 7 – 2 km/Ma. After this, exhumation rates progressively decreased as the rocks approached the surface. Rapid overall uplift/erosion had ceased when the rocks passed below 600 – 550 °C at 2.01 – 1.96 Ga ago. Observed variations in mineral cooling ages are interpreted as to reflect episodic differential tectonic exhumation within major fault systems. Inferred (maximum) ages of fault re-activation generally coincide with major continental accretion events in the Amazonian Craton and reflect long-term thermal evolution of the Imataca terrane, as conditioned by variable response to continued continental convergence during the Proterozoic.

Introduction

High-grade (granulite facies) metamorphic rocks formed during ancient orogenic events yield information on early continental crust formation processes and therefore constitute important indicators of the long-term chemical and thermal evolution of the Earth (e.g., Fyfe, 1978; Ben Othman and Allégre, 1984; Richter, 1984). Moreover, in recent years it has become increasingly recognized that the pressure-temperature-time (P-T-t) paths deciphered from mineral assemblages in such rocks provide important constraints on geophysical and tectonic models of lower crust deformation, uplift and subsequent erosion (e.g., England and Richardson, 1980; England and Thompson, 1984; Bohlen, 1987). Application of tectonic models to the evolution of metamorphosed lower crust not only requires information on protolith (nature-age) characteristics (e.g., DePaolo et al., 1991), but also relies on a detailed understanding of the cooling rates of granulitic rocks. Indeed, the derived cooling patterns are essential to characterize the exhumation processes (uplift and/or erosion) that brought the rocks towards the surface (e.g. England and Molnar, 1990). Most cooling data on metamorphic terranes has been obtained through radiometric dating ("*geochronological cooling rates*"; Spear and Parrish, 1996) of different minerals with appropriate isotope closure temperatures (e.g., Cliff, 1985). Cooling rates can also be determined from analyses of diffusional zoning in metamorphic minerals ("*petrological cooling rates*"; Spear and Parrish, 1996). However, except for Spear and Parrish (1996) research on the ≤ 80 Ma "old" Valhalla Metamorphic Complex (British Columbia, Canada), few systematic studies have been published comparing petrological and geochronological cooling rates. As cautioned by Spear and Parrish (1996), such a comparison is of critical importance in order to better assess the internal consistency of both methods; furthermore, cooling rates determined by geochronological and petrological methods may provide complementary information and therefore it is useful to use both methods on thermochronological studies of metamorphic terranes.

In this study, results of sensitive high-resolution ion microprobe (SHRIMP) U/Pb-isotope measurements are presented for zircons from felsic to intermediate granulites from the San Felix-Upata area (Figure 1) in the La Encrucijada domain (Ascanio, 1985) of the Imataca Complex (Venezuela) provide a time framework for protolith formation and high-grade metamorphism. Petrological investigation of Imataca Complex samples is used to define the main fea-

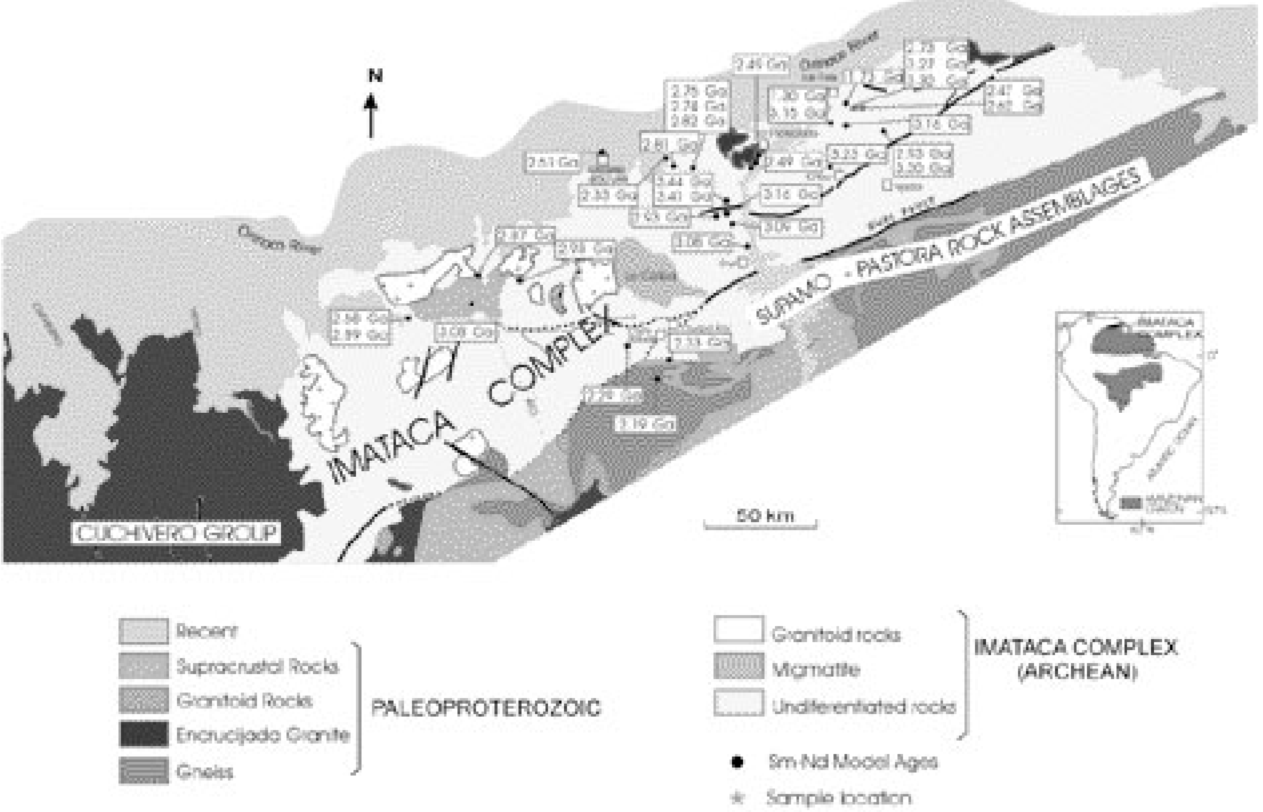

***Figure 1** Simplified geological map of the Imataca Complex area, northwestern Amazonian Craton, with sample locations (Compiled from Win et al., 1993 and Teixeira et al., 2002).*

tures of regional Transamazonian granulite metamorphism and the petrological cooling rates are inferred from diffusion modeling of Fe-Mg exchange between biotite inclusions and their garnet hosts. The data are compared with new Sm-Nd, Rb-Sr and previously reported ^{40}Ar/^{39}Ar (Onstott et al., 1988) mineral cooling ages in order to assess the thermochronology of the San Felix-Upata granulites of Imataca Complex, whereas Sm-Nd whole rock isotopic data are used to constraint the crustal residence ages and the composition of the protoliths in relationship with the mechanisms of crustal accretion. Finally, we discuss implications of our results for the geodynamic evolution of Imataca Complex.

Geological setting

The Archean Imataca Complex (Figure 1; Chase, 1965; Kalliokoski, 1965; Hurley et al., 1976) comprises the Venezuelan NW corner of the Amazonian Craton (Cordani and Brito Neves, 1982) and forms an ENE-trending, fault-bounded (~500 km long) block, extending from below the northern Orinoco river floodplains to the Guri fault zone (Onstott and Hargraves, 1981), that separates the Complex from the Paleoproterozoic (2.2 – 1.95 Ga) Supamo-Pastora granite-greenstone belt (Maroni-Itacaíunas Province) to the south (Figure 1; Tassinari and Macambira, 1999; Tassinari et al., 2000). The Imataca Complex consists predominantly of felsic, quartzo-feldspathic, granulites and variably migmatized ortho-/paragneisses (with local anatectic granitoids) that display (subordinate) compositionally-intergradational variations into intermediate/mafic compositions (gneisses, granulites). The paragneisses include minor interlayered dolomitic marbles and BIFs that constitute huge iron deposits of Algoma type. The main protolith of the Imataca Complex has been interpreted as representative of a differentiated calc-alkaline magmatic series (Dougan, 1977). However, except for granitic orthogneisses of clear igneous derivation, the lack of geochemical coherence among immobile element contents in mafic/intermediate lithotypes suggests that some (if not most) of these rocks represent metamorphosed chemical/clastic sediments that could have been derived from older granitic-basaltic (greenstone) sequences.

Imataca Complex rocks have been intensely deformed, metamorphosed and intruded by several granitic (e.g., La Encrucijada) plutons during the Transamazonian orogeny (Montgomery et al., 1978; Gibbs and Barron, 1983; Ascanio, 1985; Onstott et al., 1989; Swapp and Onstott, 1989). Metamorphic grade was at granulite facies over most of the northern areas of the Imataca Complex but changed to amphibolite facies towards the south, on approaching the Guri fault zone. Previous petrographic studies (Dougan, 1977) and our own field observations indicate that granulites south of Dougan's "pyroxene limit" have been extensively retrograded and transformed to amphibolites during superimposed cataclastic recrystallization; these features have regional expression, being observed for several kilometers along the main Guri (> 1 km wide shear zone) and other (associated) fault zones (e.g., El Pao, Rio Claro; Ascanio, 1985), implying that the amphibolite/granulite transition in the Imataca Complex does not represent prograde metamorphism. All those major fault zones have been subsequently reactivated for several times (e.g., Gibbs and Barron, 1993) and they are of utmost significance on the tectonic framework of the Imataca Complex. Indeed, geological, palaeomagnetic and isotopic data (Onsttot and Hargraves, 1981; Tassinari and Macambira, 1999) indicate that the Guri fault zone corresponds to a major transamazonian continental suture, reflecting late-Paleoproterozoic accretion of the Imataca terrane into the proto-Amazonian Craton. During continental collision the Maroni-Itacaíunas terrane was thrust over the Imataca Complex (Ascanio, 1985), and the ensuing transpressive regime between these two terranes resulted in left lateral slip (Swapp and Onstott, 1989) with development of northward verging thrusts and complex imbrication within the Imataca Complex (Ascanio, 1985). The early stages of cooling and uplift of Imataca granulites may be related to thrusting along the major fault zones.

Analytical techniques

Chemical mineral analyses were obtained by electron microprobe (Jeol Superprobe 733, Centro de Geologia FCUL, Portugal) analyses of polished, carbon-coated thin sections, using a combination of natural and synthetic standards; typical errors are less than 2 % for major elements.

Rb-Sr and Sm-Nd isotopic analyses were carried out at the Centro de Pesquisas Geocronológicas (USP, Brazil). Standard analytical procedures were used for Rb-Sr and Sm-Nd analyses according to the methodology described by Tassinari et al. (2003) and Sato et al. (1995). Rb, Sr, Sm and Nd contents were measured by isotopic dilution techniques. Sr isotopic ratios were normalized to $^{86}Sr/^{88}Sr = 0.1194$ and replicate analyses of $^{87}Sr/^{86}Sr$ for the NBS987 standard gave a mean value of 0.71028 ± 0.00006 (2σ) with blank levels at 5 ng. Nd ratios were normalized to a $^{146}Nd/^{144}Nd = 0.72190$ and the average of $^{143}Nd/^{144}Nd$ for La Jolla and BCR-1 standards were 0.511847 ± 0.00005 (2σ) and 0.512662 ± 0.00005 (2σ) respectively; blanks levels were less than 0.03 ng during the period of analysis. Sr and Nd isotopic data were obtained on a multi-collector VG 354 Micromass mass spectrometer.

Zircon U/Pb isotopic data were obtained from the Australian National University SHRIMP I instrument, using a ~30 μm diameter spot. Stern (1998) and Williams (1998) describe calibration methods and analytical procedures. $^{206}Pb/^{238}U$ ratios have an error component (typically 1.5 to 2.0%) from calibration measurements using the standard zircons. U abundance was calibrated against 238 ppm U (< ± 10%) fragments of the single crystal SL 13 standard and Pb/U was calibrated against the multi-crystal standard AS57 (1100 Ma; Paces and Miller, 1993). All errors take into account non-linear fluctuations in ion counting rates beyond that expected from counting statistics (e.g. Stern, 1998). Age calculations were performed using the Ludwig (1998) ISOPLOT/Ex program.

Sample descriptions

Studied samples were collected from road cuttings and from an abandoned quarry near the village of San Félix (on the highway to Upata; see Figure 1). Within the collection area, the general strike of metamorphic foliation is close to E-W (steeply dipping to S) and predates the development of NE-SW shear zones that represent local expressions of the main (e.g., Guri, El Pao) regional faults. Typical bluish-quartz, garnet-orthopyroxene, and two-pyroxene bearing granulites are exposed in the sampling area, of which samples V1 to V8 are representative. Sample V9 is a felsic (quartzo-feldspathic) segregation in garnet-granulites (selected for zircon geochronology). Finally, sample V10 is a quartz + plagioclase + K-feldspar + biotite ± muscovite "blastomylonite" collected from a shear zone. It has a planar fabric defined by aligned biotite and quartz plus feldspars that have experienced deformation and grain size reduction.

The metamorphic history of granulites is best reflected in samples V-6, V-7 and V-8; a summary description of the petrography, mineral-chemistry and thermobarometric investigation of these samples is given below.

Metamorphism

Granulites of the Imataca Complex were previously described by Dougan (1974) and Swapp and Onstott (1989).

Of the three samples selected for this petrologic study, two (V6, V7) are garnet + orthopyroxene + plagioclase + K-feldspar + biotite + quartz bearing granulites and V8 is a two-pyroxene + plagioclase + K-feldspar + hornblende + biotite + quartz granulite.

Granulite V8 has typical granuloblastic texture and relatively homogeneous mineral compositions. Pyroxenes have very small amounts of non-quadrilateral components and display slight core to rim Fe/Mg zoning (cpx: 0.66 → 0.64; opx: 1.23 → 1.27). Other FeMg minerals do not display detectable zoning; orthopyroxene (opx), clinopyroxene (cpx), amphibole (hb) and biotite (bio) have (average) Fe/Mg values ordered as follows: $opx^{Fe/Mg} > hb^{Fe/Mg = 1.19} > bio^{Fe/Mg = 0.90} > cpx^{Fe/Mg}$. Amphiboles display extensive tschermak substitution (Si ~ 6.4, Al ~ 2.0 a.p.f.u.) and significant A-site occupancy ($K^A > 0.32$), indicating crystallization under high-temperature conditions. X_{an} in plagioclase ranges from 0.42 to 0.44, whereas X_{ab} in K-feldspar is ~ 0.1.

Garnet granulites have granuloblastic texture, particularly well equilibrated in sample V6. Sample V7 has higher modal amounts of garnet and displays gneissic structure due to alternating garnet+ biotite- and feldspar-richer layers. Orthopyroxene and (particularly) garnet contain abundant biotite inclusions (especially, in sample V7) and larger orthopyroxene crystals surround (earlier) garnet, separating it from plagioclase. These features, coupled with complex plagioclase + K-feldspar intergrowths, suggest that the decompression P-T path followed by these rocks may have reached conditions that were close to those of biotite dehydration melting ,

biotite + quartz → orthpyroxene + K-feldspar + liquid (1)

Garnet ($Alm_{75-78}Py_{9-13}Gr_9Sps_{3-4}$) and orthopyroxene ($En_{36-38}Fs_{61-63}Wo_{5-6}$) compositional variations are dominated by almandine-pyrope and enstatite-ferrosilite solid solutions. Almandine/pyrope and enstatite/ferrosilite are zoned such that Fe/Mg increases in garnet (V6: 6.25 → 7.37; V7: 5.90 → 7.51) and decreases in orthopyroxene (V6: Fe/Mg = 1.86 → 1.67; V7: 1.91 → 1.83) from core to rim. Garnet zoning may be symmetrical about the rim, but it is also observed in garnet touching biotite and/or orthopyroxene (V6: $Fe/Mg^{garnet} = 6.25 \rightarrow 8.68$), resulting from diffusion controlled Fe-Mg exchange on cooling (Spear and Florence, 1992). Detailed analyses of biotite inclusions in garnet also reveal that there is a correlation between biotite Fe/Mg and biotite size (smaller biotite inclusions have lower Fe/Mg: V7: 60 µm – Fe/Mg = 0.89; 250 µm – Fe/Mg = 1.35). Biotite Fe/Mg is also a function of the location in the sample, such that matrix biotites (away from garnet) generally have higher Fe/Mg values than those of biotites included in garnet (V7: $Fe/Mg_{matrix\ bio} = 1.46 - 1.55$). Some late stage biotites may have formed by retrograde operation (from left to right) of reaction (1), but the small K [illegible] (see also, Spear and Markusen, 1997) ensures that the early matrix biotites still provide reliable indicators of peak metamorphic conditions. All these features are consistent with the interpretation that the Fe-Mg zoning in garnet, and biotite Fe/Mg variations are products of diffusion in response to gradients caused by Fe-Mg exchange between biotite+orthopyroxene and adjacent garnet (Spear and Parrish, 1996; biotite inclusion data will be used below to infer petrological cooling rates). Plagioclase is homogeneous within each sample ($X_{an} = 0.23$ in V6, $X_{an} = 0.25$ in V7), whereas X_{ab} in K-feldspar ranges from 0.10 to 0.17.

Thermobarometry: Peak metamorphic conditions and P-T path

Previous estimates of Transamazonian peak metamorphic conditions in the Imataca Complex have been reviewed by Swapp and Onstott (1989), who suggested T and P in the range of 750 – 800 °C and 8 – 8.5 kbar for granulites at El Pao mine. Compositions of garnet, orthopyroxene, biotite and plagioclase coexisting with quartz in samples V6 and V7 provide useful constraints on metamorphic conditions at San Felix. Calculations based on the TWQ approach (Berman, 1988; 1991; Berman and Aranovich, 1996) yield peak (core) temperature and pressure estimates at 740 ± 20 °C, 6.7 ± 0.4 kbar for sample V7. TWQ geothermobarometric results agree with those obtained from several other methods ($T_{garnet\text{-}matrix\ biotite} = 724 - 774$ °C: Hodges and Spear, 1982; $T_{garnet\text{-}orthopyroxene} = 747 -$

787°C: Lee and Ganguly, 1988; T = 786 ± 93 °C, P = 6.7 ± 1.3 kbar: Holland and Powell, 1998). These are also consistent with two-pyroxene (767 ± 20 °C: Andersen et al., 1993) and hornblende-plagioclase (770 ± 40 °C: Holland and Blundy, 1994) geothermometry on sample V8 (two-pyroxene granulite) from a nearby outcrop. Swapp and Onstott (1989) mineral data for garnet granulite sample IM115, recalculated according to the TWQ geothermobarometric method, yield 814 ± 20 °C and 7.7 ± 0.4 kbar. This suggests either that peak T-P metamorphic conditions were slightly higher at El Pao mine, or that San Felix granulites were re-equilibrated at lower than peak conditions. Geothermobarometric data from sample V6 support the re-equilibration hypothesis; estimated core to rim T and P conditions decrease from 660 ± 40 °C to 570 ± 20 °C and 5.9 ± 0.8 kbar to 4.2 ± 0.4, indicating that extensive recrystallization proceeded with decreasing temperature and pressure. Thus, overall data suggest that San Felix granulites reached peak metamorphic conditions at 750 – 800 °C, 6 – 8 kbar, followed by decompression and cooling.

Figure 2 summarizes the P-T path constraints based on pertinent reaction equilibria. Petrographic evidence to constrain the prograde P-T evolution of San Felix granulites has been mostly erased by subsequent reactions. However, our data are consistent with that of Swapp and Onstott (1989) and both suggest a clockwise P-T path involving decompression and heating to peak conditions, with the general absence of early kyanite in the Imataca rocks (Dougan, 1974; Swapp and Onstott, 1989) limiting the amount of decompression to < 2 kbar. A constraint on the retrograde P-T path comes from noting that 3qz + gr + 2alm = 6fs + 3an equilibrium (Berman and Aranovich, 1996) has an almost constant slope for the investigated compositions. Therefore, the retrograde path (Figure 2) was tentatively draw to follow that line down to about 600 °C, at ~ 13 bars/degree °C. Assuming a linear pressure gradient with depth and a constant rock column density of 2.7 g.cm-^{3}, the retrograde path corresponds to a temperature-depth gradient of 20 °C/km. Extrapolation of this path to the surface results in unreasonable hot surfaces temperatures (> 200 °C), indicating that geothermal gradients must have increased (to > 30 – 40 °C/km) as the rocks were exhumed to the surface (e.g., England and Thompson, 1984). Regardless of the actual meaning of the estimated retrograde P-T path, it is worth noting that the San Felix granulites must have remained at relatively high temperatures for long enough to allow the observed retrograde re-equilibration.

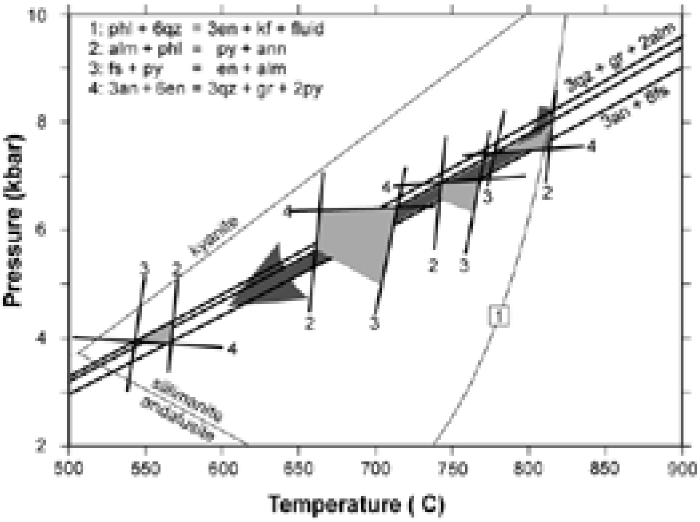

Figure 2 Pressure and temperature constraints for the Imataca Complex granulites. Displayed reaction equilibria calculated according to Berman (1988) and Berman and Aranovich (1996) thermodynamic data. "Arrow" illustrates the retrograde path as discussed in text.

Petrological cooling rates

The theory and methods that use chemical zoning in minerals to infer cooling rates have been discussed at length by a large number of workers (e.g. Dodson, 1973, 1986; Lasaga, 1983; Wilson and Smith, 1984; Spear, 1991; Spear and Parrish, 1996); therefore, only a brief summary is be provided here. The method follows the technique developed by Spear and Parrish (1996). Their approach provides a simple (but rigorous) characterization of the reaction framework that governs compositional boundary conditions, restricting the diffusion model analyses to Fe-Mg exchange between host garnet and biotite inclusions in order to assess cooling rates. It is assumed that Fe-Mg inter-diffusion is induced by compositional variations at garnet-biotite interfaces in response to changing *KD(Mg/Fe)*$^{gt\text{-}bio}$ values during the cooling process. As temperature decreases garnet becomes enriched in Fe/Mg and biotite becomes depleted, until, at sufficiently low temperature (T_c = closure temperature), the process effectively ceases. Considering mass balance requirements (the diffusive flux out of garnet must be matched by the diffusive flux into biotite), and noting that the diffusion process is rate limited by diffusion in garnet $D^{bio}_{Fe-Mg} >> D^{gt}_{Fe-Mg}$; Spear, 1991; Spear and Parrish, 1996), the Fe/Mg variations in biotite will be a function of the size of the biotite inclusions (smaller inclusions will change composition more than larger ones). The composition of each biotite inclusion can therefore be transformed into its respective closure temperature (by using the garnet core composition). Thus in each case, $T_c = f$ *(biotite size)* reflects the total diffusive flux out of garnet (Spear and Parrish, 1996), reflecting the thermochronological history (see, Dodson, 1973). The corresponding cooling rates are then obtained by comparison of biotite inclusion Tc data with the results of (computer) model diffusion calculations performed under known conditions.

In this study, initial conditions for the diffusion algorithm assume homogeneous biotite and garnet (of appropriate sizes and compositions) at the estimated peak T-P metamorphic conditions (T_0 = 800 °C, P = 7 kbar). As cooling proceeds, garnet-biotite interface compositions will change (as prescribed by the simulated thermal history, $T°C=g(t–Ma)$ in accordance to $K^{gt-bio}_{D(Mg/Fe)} = h(T(t))$ (Ferry and Spear, 1978) and solution of the diffusion equation

$$\frac{\partial C}{\partial T} = D^{gt}_{Fe-Mg} \times \frac{\partial^2 C}{\partial x^2} \qquad (2)$$

(Crank, 1975), is approximated by a finite difference model (diffusion in spherical→garnet and cylindrical→biotite geometries), using Fe-Mg interdiffusion coefficients $\left(D^{gt}_{Fe-Mg} = D_0 \times \exp\left(-Q/_{RT}\right)\right)$ calculated according to Lasaga (1979) from Chakraborty and Ganguly (1992; see also Ganguly et al., 1998) experimental data.

Figure 3 displays a comparison between observed T_c – log (biotite diameter) relations for San Felix garnet-granulite samples and the results obtained from the computational simulation of garnet-biotite Fe-Mg diffusion exchange. As it should be expected, there is a broad correlation between inclusion biotite size (30 – 250 μm) and garnet-biotite closure temperatures (550 – 720 °C). It can be seen that the San Felix T_c – log (biotite diameter) variation trend has a considerably higher slope than that of any line depicting modeled constant cooling rates in Figure 3; thus, a straightforward comparison with those model results is not justified. A feasible explanation for the data is that the retrograde P-T path of San Felix granulites proceeded under decreasing cooling rates.

Because of initial fast cooling, rapidly decreasing diffusive fluxes out of garnet will soon become unable to cope with compositional variations at the interfaces with the largest biotite crystals and these inclusions will partially close to exchange early in the cooling process. During subsequent cooling, larger biotites will behave like

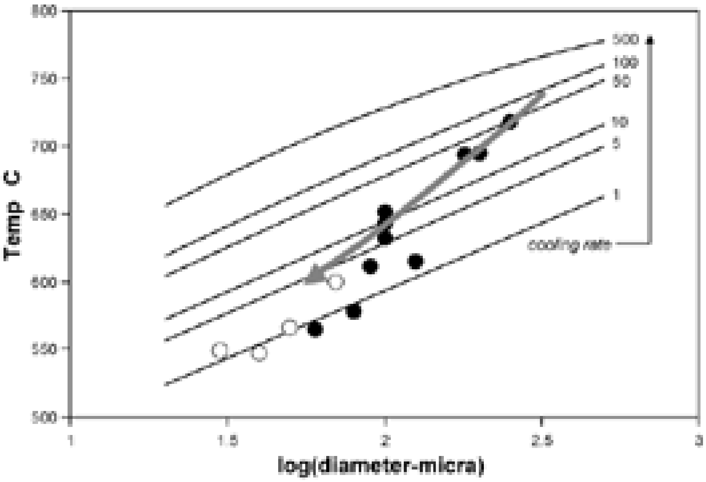

Figure 3** **Plot of closure temperatures, calculated from garnet (core)-biotite (inclusion) thermometry (Hodges and Spear, 1982), vs. biotite size (log diameter-(m) for San Felix granulites (circumferences: V6; filled circles: V7). Continuous thin lines are model results for constant cooling rates and thick arrow is model result for decreasing cooling rates, as discussed in text.

minerals with limited extent of diffusion (Ganguly and Tirone, 1999), becoming less susceptible than the smaller ones to further compositional adjustments that result from later decrease in cooling rates at lower temperatures. Therefore, T_c – log (biotite diameter) variation trends characteristic of decreasing cooling rate thermal regimes will be steeper than those depicting constant cooling rates. Following this reasoning, a wide variety of thermal histories have been explored by diffusion modeling. These numerical experiments indicate that the garnet-biotite (inclusion) data can be reasonably explained if the San Felix granulites cooled at a rate approaching 50 – 100 °C/Ma over the first 150 °C (800 – 650 °C), followed by much slower cooling (10 – 1 °C/ma), as indicated in Figure 3.

Geochronology

Thirty two new results of U/Pb and Sm/Nd, Rb/Sr isotopic analyses on whole rock and mineral (zircon, garnet, pyroxene, biotite and feldspar) separates from the Imataca Complex have been obtained during this study (Tables I and II). We will first address SHRIMP U/Pb-zircon and Sm/Nd whole rock data in order to unravel protolith ages and crustal residence times. We will then describe mineral ages that are pertinent on the thermochronological characterization of Imataca granulites.

Protolith ages

Previous Rb/Sr and Pb/Pb whole rock analyses on the Imataca Complex (Hurley et al., 1972, 1973, 1976; Montgomery and Hurley, 1978; Montgomery, 1979) suggest that protolith ages go back to at least 3.1 Ga and might be as old as 3.4 – 3.7 Ga. Also, early high-grade metamorphic reworking (La Ceiba migmatites; Figure 1) could have taken place at about 2.8 Ga ago (Teixeira et. al., 1989).

Mineral separates from garnet-granulite V6 and felsic segregation V9 yielded predominantly prismatic zircons, showing, (through cathodoluminescence imagery) fine scale oscillatory zoning and homogeneous (rim) overgrowths. Oscillatory-zoned zircon of this type is interpreted as to have grown out of felsic magmas (Pidgeon et al., 1998). Our study of nine analyses focused on zircon sites with well preserved zoning (typical of magmatic zircon). The purpose of this reconnaissance-style work was not to provide precise ages on any event, but to give indications of timing of high temperature geological events from the U/Pb zircon perspective, to be integrated with Sm-Nd and Rb-Sr data. SHRIMP U/Pb-zircon isotopic data are plotted on a $^{207}Pb/^{206}Pb$ – $^{238}U/^{206}Pb$ diagram in Figure 4. Reported $^{204}Pb/^{206}Pb$ ratios (Table I) are below 0.0004, which gives a maximum $206Pb_{\text{non-radiogenic}}/^{206}Pb_{\text{total}} = 0.50$ %, and only minor corrections for common lead. Except for spot analyses V9/1.1, zircon cores are not strongly discordant. Zircon sites of well-preserved oscillatory-zoning (apart from V6-1.1 displaying recrystallization) from the middle and ends of the grains from sample V6 yielded

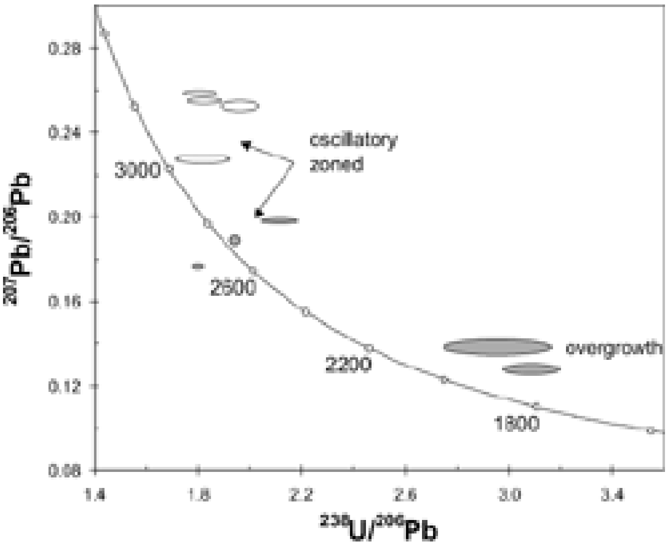

Figure 4** **SHRIMP $^{238}U/^{206}Pb$-$^{207}Pb/^{206}Pb$ Tera-Wasserbourg diagram for zircons from San Felix-Upata samples, V6 (filled symbols) and V9 (open symbols).

Table I** **Zircon SHRIMP data

sample	grain type	U ppm	Th ppm	Th/U	Pb*	204 ppb	238/206	1σ	207/206	1σ	age 207/206 Ma	1σ
V6-1.1	m,osc/hb,p	343	125	0.37	210	9	1.7989	0.0123	0.1774	0.0005	2629	5
V6-4.1	e,osc,p	327	92	0.28	184	2	1.9408	0.0154	0.1896	0.0016	2739	14
V6-5.1	r,h,p	219	96	0.44	81	24	2.9552	0.1730	0.1390	0.0030	2215	37
V6-6.1	c,osc,anh	339	150	0.44	181	8	2.1163	0.0594	0.1988	0.0008	2816	7
V6-7.1	r,h,anh/p	137	135	0.99	54	12	3.0979	0.0938	0.1269	0.0020	2055	28
V9-1.1	e,osc,p	163	102	0.63	102	3	1.9605	0.0588	0.2528	0.0020	3203	12
V9-2.1	m,osc,p	89	52	0.58	60	7	1.8205	0.0534	0.2556	0.0011	3220	7
V9-3.1	m,osc,p	487	126	0.26	299	14	1.8167	0.0858	0.2277	0.0013	3036	9
V9-4.1	m,osc,p	156	78	0.50	103	2	1.8069	0.0518	0.2589	0.0009	3240	5

anh–anhedral; c–core; e–end; h–homogeneous; m–middle; osc–fine scale zoning; p– prismatic; r–overgrowth.

Table II Sm-Nd and Rb-Sr data

samp.	mat.$^{(1)}$	Sm	Nd	^{147}Sm/ ^{144}Nd	2σ	^{143}Nd/ ^{144}Nd	2σ	Mineral Ages Ma $^{(2)}$	(143/144) (2Ga)	$_{0}\varepsilon_{N\delta(2Ga)}$	$T_{(DM)}$ Ga
V6	gt	6.676	5.237	0.7709	0.0031	0.519748	0.000013	2009±11			
V7	gt	7.951	6.829	0.7040	0.0026	0.518548	0.000048	1896±17			
V8	px	4.025	12.457	0.1954	0.0007	0.512348	0.000015	1976±49			
V-1	wr	4.947	20.291	0.1474	0.0005	0.511575	0.000012		0.509634	-8.1	2.89
V-2	wr	13.300	101.650	0.0791	0.0003	0.510654	0.000011		0.509613 -	8.5	2.92
V-3	wr	2.659	10.664	0.1508	0.0005	0.511705	0.000030		0.509720 -	6.4	2.77
V-4	wr	19.602	107.531	0.1102	0.0004	0.511218	0.000013		0.509767	-5.5	2.70
V-5	wr	4.705	18.802	0.1513	0.0005	0.511778	0.000011		0.509786	-5.1	2.67
V-6	wr	5.517	17.732	0.1881	0.0006	0.512041	0.000010		0.509565 -	9.4	2.98
V-7	wr	5.180	19.294	0.1623	0.0006	0.511789	0.000018		0.509652 -	7.7	2.86
V-8	wr	1.744	8.914	0.1183	0.0004	0.511345	0.000018		0.509788 -	5.1	2.67

sample	material$^{(1)}$	Rb	Sr	^{87}Rb/^{86}Sr	2σ	^{87}Sr/^{86}Sr	2σ	Mineral Ages Ma $^{(2)}$	$(87Sr/86Sr)_0$ 2Ga
V6	bio	614	13.37	198.928	1.621	5.789510	0.005280	1769±14	
V6	pl+KF	43.5	201	0.628	0.011	0.743428	0.000074		
V7	bio	624	13.27	203.529	1.635	5.771100	0.005620	1724±14	
V7	opx	3.56	8.90	1.162	0.009	0.751820	0.000060		
V10	bio	1152	16.04	350.899	2.805	7.750270	0.001780	1389±11	
V10	KF	171	135	3.701	0.030	0.835180	0.000033		
V10	wr	273	151	5.317	0.049	0.867518	0.000078		
V6	wr	38	183	0.603	0.005	0.744450	0.000130		0.72709
V7	wr	184	284	1.887	0.007	0.775077	0.000132		0.72072
V8	KF	9.2	454	0.057	0.003	0.711904	0.000057		0.71026
V8	px					0.712279	0.000050		

(1) - bio: biotite, KF: K-feldspar, gt: garnet, pl: plagioclase, opx: orthpyroxene, px: pyroxene, wr: whole rock
(2) - mineral - whole rock ages.

^{207}Pb/^{206}Pb ages of 2629 (± 5) – 2739 (± 14) Ma, whereas a site from the core of grain V6/6 yielded the oldest age at 2816 ± 7 Ma; this age is in close agreement with La Ceiba migmatite Rb/Sr whole rock isochron at 2.78 Ga (Hurley et al., 1973). These results suggest a period of high-grade metamorphism, extensive melting and migmatite injection in the Imataca Complex during the late-Archean. Zircon analyses from sample V9 were also of the dominant oscillatory-zoned (middle/end) grain sites. Three of these sites yielded a weighted mean ^{207}Pb/^{206}Pb age of 3229 ± 39 Ma (MSWD = 5.2) and a fourth site yielded 3036 ± 9 Ma that might reflect partial lead loss during younger thermal event(s). The dates are consistent with a mid-Archean age (≥ 3.2 Ga) for at least some Imataca protolith(s).

Nd isotopic systematics

The Nd (and Sr) isotopic data are summarized in Table II. There is some debate as to whether high-grade metamorphic differentiation of continental crust may, or may not, involve fractionation of Sm and Nd (e.g., Ben Othman et al., 1984; Burton and O'Nions, 1992). ^{147}Sm/^{144}Nd ratio in the Imataca samples ranges from 0.08 to 0.19, largely overlapping the typical range for felsic crust with an average ^{147}Sm/^{144}Nd ratio of ~ 0.11 (Taylor and McLennan, 1985). Nevertheless, Figure 5 suggests that there was significant Nd isotopic resetting during Transamazonian metamorphism (see also, Montgomery and Hurley, 1978). Accordingly, $\varepsilon_{Nd(t)}$ (DePaolo, 1988) values have been calculated for 2.0 Ga, the inferred time of peak granulite-facies metamorphism (see bellow), in order to minimize the effects of possible alteration of the Sm/Nd ratio by metamorphism. The $\varepsilon_{Nd(2\ Ga)}$ values for Imataca granulites show a limited range from - 5.1 to - 9.4 (average - 7.0 ± 1.7) and are inversely correlated with $(^{87}Sr/^{86}Sr)_{(2Ga)}$ = 0.7103 – 0.7271, being consistent with a previous long-term residence in LREE and Rb/Sr enriched upper crustal reservoir(s). Nd model ages, which are actually crustal residence times (e.g., O'Nions et al., 1983; Arndt and Goldstein, 1987), were calculated by using the measured Sm/Nd ratio for the isotopic evolution until 2 Ga ago and then assuming a typical crustal ^{147}Sm/^{144}Nd ratio of 0.11 for calculation of the intersection of the sample evolution line and the depleted mantle evolution curve (e.g., Ben Othman et al., 1984; Liew and Hofman, 1988). The corresponding mean Nd crustal residence ages of 2.8 ± 0.1 Ga (Table II) strongly support the U/Pb-zircon isotopic data (see above), and all indicate that the late-Archean (~ 2.8 Ga) was a period of major crustal reworking in the Imataca terrane. Convergence between Nd model ages and U/Pb-zircon data suggests that this event does not only involved internal differentiation of the pre-existing crustal rocks, but that a large fraction of new mantle derived material must have been added to pre-existing continental crust (McCulloch and Wasserburg, 1978; Veizer and Jansen, 1979; Allègre and Ben Othman, 1980; O'Nions et al., 1983). Thus, the "proto"-Imataca (≥ 3.2 Ga) continental block must have grown considerably at that time. From ~ 2.8 Ga to ~ 2.2 Ga the Imataca terrane appears to have undergone a period of relative tectonic quiescence, which did however involved insipient continental rifting (Tassinari et al., 2000), during the early passive stages of the Transamazonian orogenic cycle.

Thermochronology of Imataca Transamazonian granulites (geochronological cooling rates)

Previous thermochronological studies in Imataca granulites (Onstott et al., 1989) have been concerned mainly with argon cooling ages within the lower metamorphic temperature range of ~ 550 °C to ~ 150 °C. However, in high-grade rocks, such as granulites, only phases with extremely slow diffusivities for the isotopes of interest will be able to preserve ages that are close to the thermal peak during metamorphism. In the Imataca case the available phases with such characteristics are zircon (U/Pb; Heaman and Parrish, 1991; Cherniak et al., 1997) and pyroxene (Sm/Nd; Van Orman et

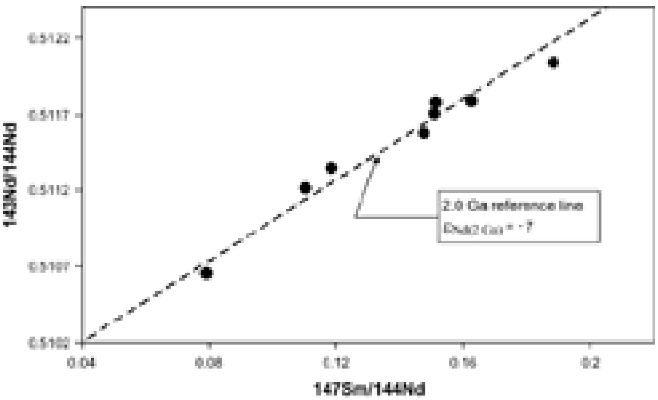

Figure 5 Plot of 147Sm/144Nd - 143Nd/144Nd relationships for whole-rock samples of San Felix-Upata granulites.

al., 2001). Thus the higher-temperature metamorphic and cooling chronology is described on the basis of U/Pb-zircon, Sm/Nd-pyroxene dating and the previously estimated petrological cooling rates. Moreover, new Sm/Nd-garnet and Rb/Sr-biotite mineral ages are presented to complement the Onstott et al., (1989) ^{40}Ar/^{39}Ar data.

U/Pb — zircon and Sm/Nd, Rb/Sr whole rock — mineral ages are summarized (together with isotopic data) in Tables I, II. ^{40}Ar/^{39}Ar closure temperatures discussed by Onstott et al. (1989) were generally adopted in this study (hornblende ~ 550 °C; biotite: ~ 300 °C; K-feldspar ~ 400 – 150 °C). In addition, closure temperatures of 600 °C for garnet (Sm/Nd; Mezger et al., 1992) and 350 °C for biotite (Rb/Sr; Harrison and MacDougall, 1980) were assumed.

Two reconnaissance analyses of relatively homogeneous overgrowths on oscillatory-zoned zircon grains from sample V6 (sites 5.1 and 7.1; Table I) yielded 207Pb/206Pb ages of 2215 ± 37 and 2055 ± 28 Ma. The thin metamorphic overgrowths and (apparent) 'zoned cores' are extremely difficult to separate with the SHRIMP 30 μm size spot and may give composite zircon ages, which are intermediate between the values of the respective cores and metamorphic grains. The site with the least discordant, younger date seems most free of core material and is interpreted as the closest indication the metamorphic peak age. This date, taken together with the 1976 ± 49 Ma Sm/Nd pyroxene age (Table II) and Montgomery and Hurley (1978) Rb/Sr thin-slab whole-rock data (~ 2.02 Ga), give a coherent picture of the chronology of metamorphism, and indicate that the thermal peak was reached at ~ 2.0 Ga ago. Nevertheless, the beginning of high-grade metamorphism may have begun somewhat earlier, being induced by pervasive plutonic magmatism in the Imataca Complex at 2.28 – 2.09 Ga (Posadas and Kalliokoski, 1967).

The cooling history can be assessed using the age determinations of minerals that were open systems and subsequently passed through their closure temperatures during cooling. In Figure 6, these closure temperatures are plotted against the cooling ages of garnet, hornblende, biotite and K-feldspar. As seen in Figure 6, the detailed thermal history of the Imataca metamorphic rocks is highly complex. Notwithstanding this, a major trend is apparent, and indicates initial fast cooling followed by slower cooling rates of 1 – 2 °C/Ma, from ~ 600 °C to ~ 350 °C. Petrologic cooling rates calculated near the metamorphic peak, are much higher and are in the range of 50 – 100 °C/Ma. Comparison of Figures 3 and 5 do indeed suggests that there is general agreement between the cooling rates obtained from garnet-biotite diffusion modeling and those obtained by thermochronology. Our data generally support Swapp and Onstott (1989) forward heat flow model for Imataca Complex. Thus, combining the model retrograde P-T path (see Figure 2) with the estimated cooling rates (Figs. 3 and 6), suggest that after peak metamorphism large portions of the Imataca terrane were exhumed from 30 to 17 km at a rate of 7 – 2 km/Ma after which exhumation rates progressively decreased (e.g., 15 → 10 km at a rate of 0.06 – 0.03 km/Ma) as the rocks approached the surface. Rapid uplift/erosion had ceased before the rocks passed below 600 – 550 °C (2.01 – 1.96 Ga ago), and the remaining Temperature — time trend shown in Figure 6 reflects thermo-mechanical recovery of the thinned crust on approaching isostatic equilibrium.

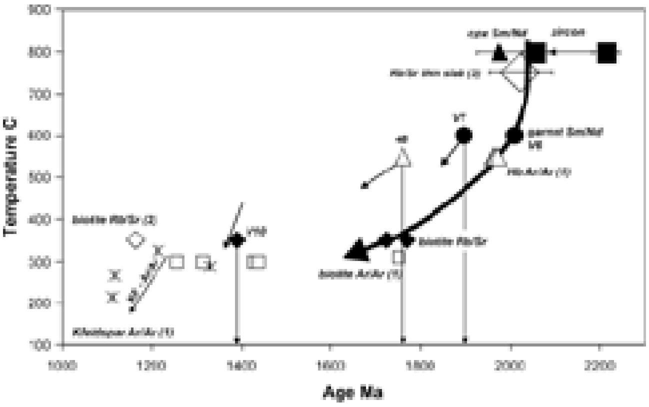

Figure 6 Plot of temperature vs. time for thermal history of Imataca granulites, based on mineral ages reported in Table II (filled symbols) and (1) - Onstott et al., (1989), (2) - Montgomery et al., (1977), (3) - Montgomery and Hurley (1978). Continuous (arrowed) thick line shows the main cooling trend for Imataca granulites. Inferred (maximum) ages of shear-zone re-activation in the Imataca Complex are illustrated by vertical thin (arrowed) lines (see text).

Despite the overall cooling trend, it is obvious from Figure 6 that large variations in cooling age occur between identical minerals (and isotopic systems) from neighboring areas in the Imataca Complex. The banded garnet-granulite sample V7 must have cooled below the garnet Sm/Nd closure temperature (~ 600 °C) about 100 Ma later (~ 1900 Ma; Table II) than the adjacent isotropic San Felix granulites. The amphibolite sample 48 (Onstott et al., 1989) from the La Ceiba migmatites (see Figure1) yields a hornblende ^{40}Ar/^{39}Ar cooling age (1760 Ma) that is ~ 200 Ma younger than those of well preserved two-pyroxene granulites (1960 Ma; Onstott et al., 1989). An even more dramatic age range is provided by biotite Rb/Sr cooling ages, with biotites from San Felix (sample V10; Table II) and Guri Dam (Montgomery et al., 1977; see Figure 1) shear zones yielding 1389 Ma and 1165 Ma, respectively, ~ 350 to 600 Ma younger than the remaining biotites (1724 – 1769 Ma; Table II). Accepting the retrograde path (≤600 °C) of 30 °C/km (as discussed above) and a constant cooling rate of ~ 2 °C/Ma (see Figures 3 and 6), the 100 Ma cooling age difference between the analyzed garnets, in Table II, corresponds to a burial depth difference of 7 km at the time of Nd isotopic system closure in granulite sample V7. This is due to the fact that these samples would have been juxtaposed at ~ 1750 Ma ago (as indicated by similar Rb/Sr biotite ages; see Figure 6). Therefore, a differential uplift of up to 7 km is inferred to have occurred between San Felix granulite samples V7 and V6 sometime after ~ 1900 Ma and before 1750 Ma. Using the same reasoning (1 – 2 °C/Ma and 35 °C/km at T ≤ 550 °C; see Figs. 3 and 5), the 200 Ma cooling age difference between the hornblendes analyzed by Onstott et al. (1989) implies that a differential uplift of 6 – 12 km has occurred between La Ceiba (amphibolite) rocks and neighboring domains of the Imataca Complex sometime after 1760 Ma and before 1100 Ma (when all samples should have been close, < 4 km, to the surface; see Figure 6). Considering these results, a reasonable interpretation for the thermochronological data is that repeated tectonic movements played an important role on the thermal structure of the Imataca Complex.

Accordingly, the most significant mineralogical and age variations in the Imataca Complex seem to occur across shear zones which were active several times during the Proterozoic (2.0 – 1.1 Ga; Ascanio, 1985; Swapp and Onstott, 1989; Gibbs and Barron, 1993) and may have been responsible for differential displacements of Imataca metamorphic rocks (e.g., Copeland et al., 1995; Dunlap et al., 1997). The thermochronological data are consistent with geological observations, also suggesting that differential tectonic exhumation was episodic on the shear zones. Thus long lasting periods of very low cooling rates (e.g., ~ 1700 Ma → 1400 Ma and ~ 1350 Ma → 1250 Ma; see Figure 6) alternate with relatively faster exhumation events (e.g., ≤1900 Ma, ≤1760 Ma, ≤1400 Ma and ≤1170 Ma; see Figure 6) that proceeded to progressively shallower depths (as required by the overall, main cooling trend of the Imataca Complex in Figure 6). Interestingly, the inferred (maximum) ages of shear zone re-activation in the Imataca Complex generally coincide with major, continental accretion tectonic-thermal events in the Amazonian Craton (e.g., Ventuari-Tapajos: 1.95 – 1.8 Ga; Rio Negro– Juruena: 1.8 – 1.55 Ga; Rondonian – San Ignácio: 1.5 – 1.3 Ga; and Sunsás: 1.25 – 1.0 Ga; see Tassinari et al., 2000). During these (late) collision events, the Imataca terrane(s) should have behaved like a rigid body, and the resulting deformation was mostly concentrated on pre-existing fault systems. Thus, renewed differential displacements along the main shear zones allowed relatively faster uplift of the intervening blocks, while permitting slower exhumation on the remaining Imataca Complex.

Conclusions

SHRIMP U/Pb-zircon data, coupled with Nd mean crustal residence ages, indicate that at least some of the Imataca Complex developed from mid-Archean (≥ 3.2 Ga) continental protoliths which underwent considerable reworking and juvenile accretion additions during late-Archean (~ 2.8 Ga. This was followed by a long period of relative tectonic quiescence, from ~ 2.8 Ga to ~ 2.2 Ga, before the onset of the Transamazonian orogeny which is the major event preserved in Imataca rocks.

Imataca Transamazonian granulites experienced peak metamorphic conditions of 750 – 800 °C, 6 – 8 kbar with associated transpressive shearing that leaded to northward directed thrusting and tectonic imbrication. Geochronology on zircon, pyroxene, garnet, hornblende (Onstott et al., 1989) and biotite has been used to constrain the timing of peak metamorphism at 1.98 – 2.05 Ga and the average initial cooling rate of ~ 30 °C/Ma (from 800 °C to 600 °C). Diffusion modeling of Fe-Mg exchange between biotite inclusions and host garnet yields (near metamorphic peak) cooling rates of 50 – 100 °C/Ma, that are generally consistent with cooling rates determined from geochronology. Combining the inferred retrograde P-T path (30 – 40 °C/km) with the estimated (petrological/geochronological) cooling rates, suggests that after peak metamorphism large portions of the Imataca Complex were exhumed from 30 to 17 km at a rate of 7 – 2 km/Ma after which exhumation rates progressively decreased (15 → 10 km at 0.06 – 0.03 km/Ma) as the rocks approached the surface. Rapid overall uplift/erosion had ceased before the rocks passed below 600 – 550 °C (2.01 – 1.96 Ga ago) and the remaining slow cooling represents thermo-mechanical recovery of the thinned crust on approaching isostatic equilibrium.

Large variations in mineral cooling ages seem to occur across major fault systems in the Imataca Complex, which are interpreted as to reflect episodic differential tectonic exhumation within those shear zones. Geological evidence indicates that movements along the shear zones occurred for several times during the Proterozoic and the inferred (maximum) ages of re-activation generally coincide with major continental accretion events in the Amazonian Craton. The thermochronological data, therefore, reflects the long-term thermal evolution of Imataca Complex, as conditioned by variable response to continued continental development into the Amazonian Craton during the Proterozoic.

Acknowledgements

This project was partially supported by grants from SOPEMI, Pesquisa e Exploração de Minérios S/A and PRONEX project (# 41.96.0899.00). J.Munhá acknowledges support from Centro de Geologia (FCUL) and FAPESP (# 03/00027-4). The authors also acknowledge Compania Minera SOPEVEN for his helpful support during field work and Veridiana T.S. Martins for a review of the manuscript. This paper is a contribution to research project PETROLOG (FCT-Portugal; POCTI/FEDER) and FAPESP (Thematic project # 95/4652-2).

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