



Ductility of garnet as an indicator of extremely high temperature deformation: Reply

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INTRODUCTION

On the basis of field and laboratory (both optical and TEM) data, we interpreted flattened garnet crystals isolated in a quartz matrix of the granulite facies mylonites from the Morin shear zone as deformed most likely by recovery-accommodated dislocation creep. Den Brok and Kruhl, in their Discussion, propose an origin by "grain boundary diffusion creep (e.g. Coble creep or pressure solution)" for these flattened garnets. Coble creep is a model proposed by Coble (1963) for fine-grained *monophase polycrystalline* aggregates in which creep rate is controlled not by diffusion through the lattice but by diffusion along grain boundaries. Pressure solution is a special form of Coble creep during which the diffusivity is higher along fluid-filled grain boundaries than along dry grain boundaries (Nicolas & Poirier 1976). Den Brok and Kruhl use the term Coble creep in a less restrictive sense to refer to mass-transfer processes including dissolution and precipitation of material in *polyphase* rocks.

PRESSURE SOLUTION OR DISLOCATION CREEP?

We reject Den Brok and Kruhl's proposition for the following six reasons:

(1) In the studied garnets, there is no diagnostic evidence for the operation of pressure solution such as: (i) precipitation growth or overgrowth of garnet at local dilation sites such as pressure shadows, voids, veins and boudin gaps, (ii) indented, sutured or interpenetrated grain contacts (stylolites), (iii) relative concentration of insoluble material along dissolved surfaces, and (iv) truncated idiomorphic garnet crystals (Passchier & Trouw 1995). Surprisingly, the tensile fractures which cut and pull-apart the outward-pointing cusps of an initially flattened garnet grain from its central main segment, shown in our fig. 4(e) and described in our original text (p. 986), are misinterpreted by Den Brok and Kruhl as original idiomorphic crystal surfaces (see their fig. 2). An intermediate stage of such a microboudinage

was shown in our fig. 4(f) (Ji & Martignole 1994). Their interpretation concerning our fig. 4(e) obviously cannot be applied to most of the lens-shaped garnets (e.g. figs. 2-3, fig. 4a-d & f) within the studied rocks. Our fig. 4(e) docs suggest that the garnet has been flattened before being fractured and pull-apart, and that during and after the boudinage the garnet was rheologically much stronger than quartz and feldspar. Note that the pressure shadows at the ends of the central segment of garnet (fig. 4e in our original paper) are filled with K-feldspar. Since the solution transfer of K-feldspar generally takes place at high temperatures, we expect that the garnet was plastically deformed at even higher temperatures.

(2) Gray & Durney (1979) published the following mineral sequence according to decreasing mobility by solution transfer: calcite > quartz > feldspar > chlorite > biotite > muscovite > opaques. The solubility of garnet is much less, even, than opaques in the studied mylonites, as demonstrated by the fact that the opaques precipitated in the pressure shadows of garnets. It is difficult to accept that garnet was mobile while its neighboring quartz was not.

(3) There are two key factors controlling the activity of pressure solution: the presence of a free fluid phase (e.g. water) and the activity of microcracking. The garnet-bearing rocks studied in our paper were first dehydrated during a granulite facies metamorphism (~1190 Ma), then subject to a contact metamorphism around the anorthosite intrusion (~1155 Ma), and subsequently to another granulite facies metamorphism (1070-1020 Ma). These rocks are characterized by anhydrous minerals such as garnet and pyroxenes rather than hydrates like micas and amphiboles. Neither fluid inclusions occur along garnet grain boundaries nor cracks are healed or filled with fine garnet grains. Thus, when the garnet was deformed, temperature and effective confining pressure were so high that water could not exist as a free fluid phase at the high concentration which is required for pressure solution. Even if it were present in small amounts, water may have diffused into the mineral lattice and formed water-related defects which result in a reduction in Peierls stress for glide, promotion of climb, or enhanced dislocation multiplication (Paterson 1989).

(4) Unlike quartz and calcite, garnet has a complex chemical composition. Different cations generally have different diffusion rates either in the lattice or along grain

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boundaries, resulting in compositional zoning, that is, newly precipitated material is different in composition from the consumed one. Our microprobe data (Ji & Martignole 1994, p. 986; and Ji *et al.* unpublished data) demonstrate that the studied garnet crystals have identical composition and that there is no compositional zoning from core to rim. This suggests that the diffusion process proposed by Den Brok and Kruhl did not take place in these garnets.

(5) It is well known that pressure solution and solution transfer are dominant at diagenetic to low-grade metamorphic conditions (e.g. greenschist facies) where fluids are abundant and intragranular deformation mechanisms favored at higher temperatures, are relatively difficult (e.g. Twiss & Moores 1992, Passchier & Trouw 1995). If garnet could be deformed by pressure solution, lens-shaped garnet would be common in low to moderate grade metamorphic rocks.

(6) No experimental data have shown the prevalence of Coble creep in polycrystalline aggregates of garnets. Parthasarathy *et al.* (1992) studied experimentally the deformation behavior of polycrystalline aggregates of yttrium aluminum garnet with an average grain size of only 3 μm in the temperature range 1400–1610° C. Their data suggest that the deformation is controlled by Nabarro–Herring creep rather than by Coble creep.

In the first paragraph of their Discussion, Den Brok and Kruhl suggest that the garnets were deformed by “grain boundary diffusion creep (i.e. Coble creep or pressure solution) at a temperature in the range 550 to 750° C” rather than by recovery-accommodated dislocation creep at high temperatures (e.g. $\sim 900^\circ\text{C}$) as suggested in Ji & Martignole (1994). In the last paragraph of their Discussion, however, they state that the “garnet can be deformed by dislocation creep in the range 550 to 750° C at geologically realistic strain rates”. Obviously, there is a contradiction between these propositions. Furthermore, they do not explain why garnets cannot be deformed by dislocation creep at high temperatures (e.g. $\sim 900^\circ\text{C}$). Yet, they fail to justify their statements: “the total amount of strain by dislocation motions might have been very small” although our TEM “observations are consistent with significant amounts of strain by dislocation processes”. Their ideas are in conflict with previous data on experimentally deformed non-silicate garnets (Rabier *et al.* 1979, Karato *et al.* 1994, Wang *et al.* 1996) and TEM results of naturally deformed silicate garnets from peridotites and eclogites (Ando *et al.* 1993) and from ultradeep mantle xenoliths (Doukhan *et al.* 1994). Like us, all these authors concluded that at high temperature garnet is deformed plastically by dislocation climb creep.

Based on our own studies on the mylonites and on the regional geology, we carefully postulated that the garnet recorded an early ($\sim 1155\text{ Ma}$), high temperature (about 900° C) flattening deformation related to the diapiric emplacement of the Morin anorthosite massif, while quartz and feldspar recorded a later ($\sim 1020\text{ Ma}$), lower temperature (750–550° C), non-coaxial deformation related to dextral strike-slip movement along the Morin

shear zone (Ji & Martignole 1994 p. 994). However, Den Brok and Kruhl, in their Discussion, propose that the deformations of garnet, quartz, K-feldspar and plagioclase might be caused by a single deformation event (i.e. the dextral strike-slip movement) “at a temperature in the range 550 to 750° C”. At such a temperature, garnet is rheologically much stronger than feldspar and quartz (fig. 8 in Ji & Martignole 1994 and their fig. 3). If their proposition were true, we might expect to observe garnets more prolate than quartz and feldspar. Contrary to their first proposition, Den Brok and Kruhl also speculate on two strain events; an oblate strain at higher temperature and a prolate one at lower temperature, but neither field nor regional geological data support such a proposition.

DYNAMIC RECRYSTALLIZATION

By analogy with other crustal minerals such as feldspar, Den Brok and Kruhl infer that dynamic recrystallization should have taken place in garnet crystals from the Morin shear zone if they were flattened by about 50% on average. Such a reasoning does not seem to be appropriate because the critical strain required to initiate dynamic recrystallization depends on the intrinsic nature of the material. *Not all materials recrystallize dynamically.* According to Gottstein & Mecking (1985), the activity of dynamic recrystallization increases with increasing rate of dislocation storage during deformation and with increasing grain-boundary mobility. In low symmetry minerals such as feldspar, there are only one or two easy slip systems available. Because intracrystalline diffusion of one or more ions is very slow and because stacking fault energy is low, it is difficult for extended dislocations to climb and cross-slip, and so to annihilate and rearrange (Tullis & Yund 1985, Ji & Mainprice 1990). Deformation can rapidly result in such a high density of immobile (tangled) dislocations that grain-boundary migration recrystallization begins. In contrast, during high temperature deformation of materials with a high stacking fault energy, like aluminum or b.c.c. iron, the operation of dynamic recovery alone leads to a high enough dislocation annihilation rate so that the stored dislocation density remains insufficient to initiate recrystallization (McQueen & Jonas 1975). Even for a given material, the critical strain required to initiate dynamic recrystallization depends on deformation conditions such as temperature and strain-rate. For example, Rossard (1960) showed that the critical strain in austenitic steel at 1100° C increases with increasing strain-rate. Blaz *et al.* (1983) demonstrate that the critical strain in OFHC copper deformed at $2 \times 10^{-3}\text{ s}^{-1}$ increases with decreasing temperature. So far, TEM evidence for dynamic recrystallization has not been found in plastically deformed garnets from peridotites, eclogites (Ando *et al.* 1993) and ultradeep mantle xenoliths (Doukhan *et al.* 1994). Dynamic recrystallization is also absent during experimental plastic deformation of non-silicate garnets (e.g. Rabier *et al.* 1979 and other references cited in our original paper).

To support their idea that strains of less than 50% can lead to intense recrystallization of garnet, Den Brok and Kruhl incorporate a microphotograph in their Discussion. However, they did not provide any TEM evidence showing that the shape of the *porphyroblast* shown in their fig. 1 is due to dislocation creep rather than to metamorphic reaction (Schenk 1990) and that the garnet is dynamically recrystallized rather than fractured. Since it is very hazardous to speculate from a microphotograph, we simply note some differences between their garnet *porphyroblast* from the central metapelite unit of the Calabrian lower crustal section and our garnet porphyroclasts from the Morin shear zone. (i) Their porphyroblast contains inclusions of fibrolitic sillimanite, quartz and graphite. (ii) The boundary of this porphyroblast is irregular rather than smooth. (iii) The porphyroblast is cut by at least three sets of fractures. The Discussion did not explain these differences nor explore the possible implications for petrogenesis and strain history.

CRYSTALLOGRAPHIC PREFERRED ORIENTATION

Den Brok and Kruhl claim that dislocation creep should, while diffusion creep should not, lead to a significant crystallographic preferred orientation (CPO) of garnet. It is common knowledge in material science, that the CPO development of a material (mineral or metal) depends not only on the prevailing deformation mechanism, but also on many other factors such as the crystal nature and volume fraction of the material in polyphase aggregates as well as deformation conditions (e.g. temperature and strain-rate). For a given finite strain produced by dislocation slip, crystals with more active independent slip systems generally develop weaker CPOs than those with fewer active slip systems even though critical shear strength of each slip system is almost equal. During slip, an unconstrained crystal with five or more independent slip systems changes its shape but its lattice is not subject to significant rotation with respect to the instantaneous stretching axes of bulk flow. We will not be surprised if the garnets show a random CPO in the studied mylonites even though each garnet was deformed by dislocation processes. The reasons for this are simple. (i) Garnet is of cubic symmetry and has more than 5 independent slip systems. (ii) Garnet, which represents only 2–3 vol. % of the studied rocks, is isolated in the matrix of quartz and feldspar. (iii) As interpreted in our original paper (Ji & Martignole 1994 p. 994), the garnet crystals flattened during an early high-temperature deformation ($\sim 900^\circ\text{C}$) were subsequently rotated as rigid inclusions in the flowing quartz matrix during a later low temperature ($550\text{--}750^\circ\text{C}$), non-coaxial deformation. Furthermore, grain boundary diffusion, pressure solution or dissolution–precipitation may also result in CPO; this is true at least in the cases of layer silicates and quartz (Cox & Etheridge 1989 and references cited therein). On the other hand, the presence of a CPO is

not necessary to indicate the absence of diffusion-accommodated grain boundary sliding (Rutter *et al.* 1994). Thus, the absence of a garnet CPO in the studied mylonites will not provide unequivocal evidence against the action of dislocation glide.

CLARIFICATION OF OTHER POINTS

A number of other comments require brief responses:

(1) The disputers question why “there are no remarkable optical signs of heterogeneous intracrystalline strain” (e.g. lattice rotation, kinks, undulatory extinction, subgrains in the optical scale). Reasons for this are simple: (i) garnet is optically isotropic; (ii) according to the von Mises (1928) criterion, a material with five or more independently active slip systems can deform homogeneously by slip alone; and (iii) at least some of the strain features may be recovered or even erased by annealing.

(2) We agree with Den Brok and Kruhl that the knowledge of original grain shape is important for finite strain analysis. In the construction of our fig. 6(b) (Ji & Martignole 1994), we assumed that the measured feldspar grains were neoblasts formed during the early high temperature deformation and thus almost equant prior to the last deformation.

(3) The flow laws of minerals used in our original paper follow a conventional power law formula:

$$\dot{\epsilon} = A\sigma^n \exp(-Q/RT) \quad (1)$$

where $\dot{\epsilon}$ is the strain-rate, σ the differential stress, A a constant, n the stress exponent, Q the activation energy and R the gas constant. The flow law parameters for garnet were experimentally determined by Dr. Zichao Wang (written communication, October 2 1992) and included in Liu *et al.* (1992) and particularly in Karato *et al.* (1995) in the following form:

$$\dot{\epsilon} = B(\sigma/\mu)^n \exp(-gT_m/T) \quad (2)$$

where μ is the shear modulus, T_m the melting temperature, and B and g are two constants. Equations (1) and (2) can be intertransformed. Rheological data for various garnets and the experimental conditions were given in detail in Karato *et al.* (1995).

(4) Den Brok and Kruhl dispute the judicious use of the flow law for ‘dry’ quartzite determined by Koch *et al.* (1980). Of course, we are aware of many flow laws of quartzites determined in different laboratories using different apparatus, starting materials, sample assemblies, water fugacities, confining pressures, temperatures and strain-rates (see Paterson & Luan 1990, Kohlstedt *et al.* 1995 for reviews, table 1 in Koch *et al.* 1989 and Wang *et al.* 1994). Our choice was based on the following criteria: (i) The same deformation mechanism (i.e. dislocation creep) operates in nature as in the experiments. According to Den Brok (1991), the microstructures of the deformed ‘wet’ samples reported in Jaoul *et al.* (1984) are indicative of microcracking plus solution–

precipitation process rather than dislocation creep. This is why the data of Jaoul *et al.* (1984) were not used in the construction of our fig. 8 (Ji & Martignole 1994). (ii) The flow strength of quartz should not be higher than that of plagioclase. (iii) The extrapolated flow strength of quartzite given by Koch *et al.* (1980) is located at nearly mid-way between the upper and lower bounds set by Paterson & Luan (1990) using a gas-medium apparatus. (iv) Finally, the flow law determined by Koch *et al.* (1980) has been widely used in geodynamic modeling (e.g. White & Bretan 1985, Kuszniir & Park 1987).

(5) It is premature to debate the relative flow strength change between garnet and quartz at extremely high temperatures ($\sim 900\text{--}1000^\circ\text{C}$) and natural strain-rates (e.g. 10^{-14} s^{-1}) because high quality rheological data for quartz and particularly garnet and feldspar, determined by laboratory experiments under well-controlled conditions (i.e. differential stress, confining pressure, strain, temperature, strain-rate, water fugacity, grain size deformation mechanism etc.) are not available. A recent microstructural study of granulite facies quartzitic rocks from the Highland Complex of Sri Lanka (Kleinschrodt & McGrew 1995) also inferred that differences in flow strength between quartz, feldspar and garnet were minimal when garnet deformed plastically. Theoretically, two materials with different stress exponents should reverse their relative strength with changing differential stress and strain-rate (Ji & Zhao 1993a). Some examples of the strength reverse for metal compounds were given in our original paper (Ji & Martignole 1994 p. 994). In response to the dispute as to whether a strong inclusion is fractured in a soft matrix, we simply invite the disputers to refer to Lloyd *et al.* (1982) and Ji & Zhao (1993b).

CONCLUSIONS

First, there is neither field nor microstructural evidence for dissolution-precipitation of garnet in the studied mylonites. Second, although the problem of flattened garnet is debatable, we consider that at least in the case of the Morin shear zone, dislocation creep is the most plausible mechanism to have been active and we do not find any convincing arguments to discard this possibility. We hope that our work will stimulate the research on the causes and possible geological implications of flattened garnets in high grade metamorphic rocks.

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