Sm–Nd disequilibrium in high-pressure, low-temperature Himalayan and Alpine rocks

Béatrice Luais*a,†, Stéphanie Duchènea, Julia de Sigoyerb,2

Laboratoire de Sciences de la Terre, Ecole Normale Supérieure, CNRS–UMR 5570 46, allée d’Italie, 693645 Lyon Cedex 07, France
Laboratoire de Pétrologie et Tectonique, CNRS–UMR 5570, Université Claude Bernard, 27 – 43 Bld du 11 Novembre, 69622 Villeurbanne, France

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Abstract

In order to decipher the causes of Sm–Nd isotopic disequilibrium in high-pressure, low-temperature rocks, Sm–Nd isotopic analyses were carried out on minerals from four Himalayan (Tso Morari unit) and four Alpine (Dora-Maira, Monte Viso, Sesia Lanzo) eclogitic rocks of different lithologies and different intensities of post-eclogitic metamorphism. In most of these samples, garnets show the following striking features: either 147Sm/144Nd ratios lower than whole-rock, or 143Nd/144Nd ratios lower than whole-rock for higher 147Sm/144Nd ratios. In both cases, no age can be calculated. Two hypotheses can be proposed to explain the strong isotopic disequilibrium in garnet separates: (1) the preservation of the isotopic signature of the plagioclase from which garnet formed, (2) the occurrence in garnet of inherited sub-microscopic crustal inclusions with low 147Sm/144Nd and 143Nd/144Nd ratios. Knowing the Lu–Hf eclogitisation age for two samples, these two hypotheses are simulated: (1) the composition of plagioclase before eclogitization is recalculated from the measured garnet composition. The plagioclase/whole-rock age obtained with this simulation is geologically meaningless, showing that this hypothesis is invalid, (2) the 143Nd/144Nd ratio of sub-microscopic inclusions is calculated considering that the garnet separate, which was analysed, is a mixture between inclusions with low 147Sm/144Nd ratios (0.05–0.1) and pure garnet with a 147Sm/144Nd of 2. The inclusion/whole-rock ages recalculated for both samples suggest that inclusions have an old, crustal origin. The hypothesis of sub-microscopic inclusions in garnet can therefore explain the observed isotopic disequilibrium provided that the inclusions have a low Sm/Nd ratio, like allanite, apatite (and monazite), and that they are inherited from crustal contamination processes. In the case of a magmatic protolith, contamination can be achieved through crustal assimilation or a first stage of circulation of crustal fluid. In the case of Dora-Maira, the possible sedimentary protolith contains components from an old contaminant crust. In the Himalayan samples, the inverse correlation between the degree of retrogression in blueschist and then amphibolite facies, and the extent of disequilibrium between garnet and other mineral phases suggest a second stage of fluid circulation associated with retrograde metamorphism. This fluid circulation would result in the dissolution of these sub-microscopic crustal inclusions and partial
leaching of their rare-earth elements (REEs) from the rock, which would scavenge the crustal $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic signature from the rock. © 2001 Elsevier Science B.V. All rights reserved.

Keywords: Sm–Nd dating; HP-LT metamorphism; Isotopic disequilibrium; Inherited inclusions; Fluid circulation; Himalaya; Alps

1. Introduction

The dating of high-pressure metamorphic rocks is critical for an understanding of the dynamic conditions of exhumation rates of HP rocks, such as the exhumation rates. Omphacitic clinopyroxene and garnet, which are the main minerals constituting eclogites, have distinct Sm/Nd ratios. Whole-rock or omphacite have low $^{147}\text{Sm}/^{144}\text{Nd}$ ratios ($^{147}\text{Sm}/^{144}\text{Nd} = 0.1–0.3$, e.g. Thöni and Jagoutz, 1992) with respect to garnet, which is preferentially enriched in MREE compared to LREE. Bulk analyses of $^{147}\text{Sm}/^{144}\text{Nd}$ ratios in garnet from 0.3 to up to 4 have been reported by Stosch and Lugmair (1990), Argles et al. (1999), Prince et al. (2000), but in-situ analyses can reveal extremely high $^{147}\text{Sm}/^{144}\text{Nd}$ (up to 9–12; Prince et al., 2000). The large spread in Sm/Nd ratios between mineral phases make the Sm–Nd isotopic system a suitable geochronological tool for dating garnet-bearing rocks. The Sm–Nd method has been successful in dating moderate (Burton and O’Nions, 1991; Vance et al., 1998, Vance and Harris, 1999) and high-pressure pelites (Thöni and Jagoutz, 1992, de Sigoyer et al., 2000). However, in low-temperature mafic eclogites, garnet does not equilibrate isotopically with clinopyroxene and other accessory mineral phases at the time of crystallization. Age indetermination results from either scatter of data (Thöni and Jagoutz, 1992) or a negative slope of the regression line, because of the unradiogenic $^{143}\text{Nd}/^{144}\text{Nd}$ signature of a high Sm/Nd garnet. Disequilibrium resulting in an apparent age that is often younger or older than the geological one (Hensen and Zhou, 1995) will be difficult to identify on isochron diagrams.

In some cases, the disequilibria have been explained by incomplete redistribution of the $^{143}\text{Nd}/^{144}\text{Nd}$ ratios between the major minerals during metamorphic recrystalization. Mørk and Mearns (1986) made a systematic isotopic analysis of a gabbro–eclogite transition in Norwegian coronites. They demonstrated that the disequilibrium was greatest in those rocks preserving magmatic textures, and that equilibrium was only achieved in well recrystallized eclogites. In their interpretation, garnet and clinopyroxene would (more or less) preserve the isotopic signature of the minerals from which they grew in situ, plagioclase and augite, respectively. Thöni and Jagoutz (1992) also interpret the disequilibrium observed in Austroalpine eclogites as the preservation of ancient isotopic signatures by pseudomorph minerals. Another type of isotopic disequilibrium that has been identified corresponds to multi-stage growth of garnet, in which cores are older than rims, but the existing studies concern amphibolite facies rocks in the Sm/Nd (Vance and O’Nions, 1990; Burton and O’Nions, 1991; Getty et al., 1993) and the Rb/Sr (Christensen et al., 1989) systems rather than Sm/Nd in eclogitic rocks.

The combined knowledge of metamorphic temperatures and Sm–Nd mineral equilibria of HP granulitic and eclogitic rocks has been used to estimate a Sm–Nd closure temperature ($T_c$) of 600–800 °C in garnets from various tectonic settings (Cohen et al., 1988; Mezger et al., 1992; Burton et al., 1995; Hensen and Zhou, 1995). Low-temperature eclogite facies rocks (peak $T$ at 450–650 °C) are therefore likely to form below the closure temperature of the system, that is at a temperature where volume diffusion is inefficient at the scale of a single crystal. This case is a priori favourable for the persistence of the elemental and the isotopic heterogeneities discussed above. By analogy with major element equilibration, the equilibration of the radiometric system should be influenced by other kinetic factors such as deformation (Mørk and Mearns, 1986). In the particular cases under discussion, textural equilibration is achieved however, and demonstrates major element mobility at the centimetre scale.

In a number of cases however, no petrologic and textural observations support any explanation of the disequilibrium. In this paper, we report a Sm–Nd isotopic study on high pressure–low temperature metamorphic rocks from the Himalayan belt (Tso Morari unit, East Ladakh) and from the Alpine belt (Dora-Maira, Sezia-Lanzo and Monviso units, Western Alps), in which isotopic disequilibrium at the time
of eclogitization has been observed in samples which were texturally equilibrated under eclogitic facies. By contrast, Lu–Hf studies carried out on the same samples, and Sm–Nd studies carried out on neighbouring HP-LT metapelitic rocks (de Sigoyer et al., 2000) did not show such isotopic disequilibria. We will discuss the possible explanations for the Sm–Nd disequilibria and emphasise the role of inherited submicroscopic inclusions that could lower the Nd isotopic ratio of the host mineral (Zhou and Hensen, 1995).

2. Geological context

2.1. Himalayan samples

Four mafic samples were collected in the high-pressure metamorphosed Tso Morari massif, located in the internal part of the northwest Himalaya (de Sigoyer et al., 2000). This crystalline massif, which belongs to the Indian margin, is located in the internal part of the Himalayan belt, south of the Indus suture zone and North of the Tethyan sedimentary cover of the Indian margin. The Tso Morari massif consists of a Precambrian to Triassic sedimentary cover. An Ordovician granite has been intruded into the Precambrian and Cambrian sediments (Girard and Bussy, 1999), before being deformed and transformed into the orthogneissic core of the Tso Morari dome. Mafic lenses are observed in this unit: some dolerite lenses (Ch171c) in the orthogneiss may be related to Ordovician magmatism and other basaltic lenses (Ts34, Ts51, Ts45) are associated with the Permian dolomite or metapelitic cover. The tholeiitic to alkalic chemical composition of all these basaltic lenses can be correlated to the Permian Panjal Trap basalts (de Sigoyer, 1998). The Panjal Traps represent continental flood basalts, and their eruption has been related to the late Paleozoic rifting phase of the Neo-Tethys along the northern Indian margin (Papritz and Rey, 1989). All the rocks observed in this unit were subjected to an HP-LT metamorphic event.

Four samples of metamorphic mafic rocks were selected for the isotopic study, according to their different metamorphic overprints. Two samples (Ts34 and Ch171c), collected in the central part of the massif, are fresh eclogites. They are characterised by garnet, clinopyroxene, rutile, quartz, zoisite, + glaucoephane, phengite, paragonite, and carbonates. The metamorphic conditions estimated for these eclogitic assemblages are 20±3 kbar, 580±60 °C (de Sigoyer et al., 1997; Guillot et al., 1997). The third sample (Ts51) was collected in the southern limb of the massif, where the post-eclogitic deformation is more well developed. This sample underwent the first stage of an isothermal decompression through the blueschist facies after eclogitization, as shown by the slight symplectic reaction due to the transformation of the omphacites (0.41 ≤ XJd ≤ 0.50) in sodic augite (0.13 ≤ XJd ≤ 0.18) associated with plagioclase lamellae (0.85 ≤ XAb ≤ 0.98) (de Sigoyer et al., 1997). The last sample (Ts45) has been taken from the border of a deformed mafic lens located in the central part of the massif. In this sample, omphacite has been totally replaced by calcic amphibole and plagioclase. This reaction suggests the recrystallization of this rock under amphibolitic conditions (9±3 kbar, 610±70 °C) during the exhumation of the massif (de Sigoyer et al., 1997).

A geochronological study has been carried out on metapelitic and metabasaltic rocks of the Tso Morari massif (de Sigoyer et al., 2000), using several methods (U–Pb, Lu–Hf, Sm–Nd, Rb–Sr and Ar–Ar). In order to obtain information about the exhumation rates, the geochronological analyses were carried out on specific mineral associations, which characterise the different stages of the Tso Morari metamorphic evolution. A ca. 55-Ma age for the eclogitization has been obtained by Lu–Hf for the mafic eclogite Ts34, by Sm–Nd and U–Pb on allanite for high-pressure metapelites. The recrystallization under amphibolitic conditions has been dated by Sm–Nd, Rb–Sr and Ar–Ar on recrystallized rocks at 47±2 Ma. Finally, the end of the Tso Morari exhumation (3–5 kbar) has been dated by Ar–Ar on muscovite and biotite at about 29±1 Ma.

2.2. Alpine samples

Four eclogitic samples from the Western Alps have been dated using the Sm–Nd method. They come from three different geodynamic contexts, namely the former Tethyan Ocean in the Monviso area, the former European margin in the Dora-Maira Massif, and the former Apulian margin in the Sesia Lanzo zone.
They are all fresh eclogitic rocks, free from any retrograde mineral phases. The Monviso area is a metamorphic ophiolite. The conditions of eclogitization estimated from thermobarometric studies are 450 °C for a pressure of 15–20 kbar (Lardeaux et al., 1987; Cliff et al., 1998). The exhumation path goes through the blueschist and the greenschist facies, but is not recorded in our sample. The age of eclogitization in the Monviso area has been consistently estimated from 39Ar/40Ar (Monié and Philippot, 1989) and Lu–Hf analyses (Duchêne et al., 1997) at 48–51 Ma. The Sm–Nd data published by Cliff et al. (1998) favour a 60-Ma age, but the large uncertainty (± 10 Ma) associated with these data makes it compatible within errors. Sample Vi8 is a mafic eclogite (ferrogabbro) taken from the Lago Superiore Unit, and consists of a fine-grained (200 μm), banded assemblage of omphacite, garnet, rutile and, to a lesser extent, glaucophane and zoisite.

The Dora-Maira massif is one of the Internal Crystalline Massifs of the Alpine belt. The area has been the subject of numerous studies, in particular because of the discovery of coesite in eclogitic samples (e.g., Chopin, 1984; Schreyer et al., 1987; Chopin et al., 1991). Sample DM is a fine grained pyrope–coesite bearing quartzite, collected in the ultra-high pressure unit, which records peak metamorphic conditions of 30 kbar and 700 °C (see Chopin, 1984 for the complete petrographic description). Petrological and geochemical studies generally attribute a sedimentary origin to the protolith, which would therefore represent part of the cover of the crystalline basement (Schreyer, 1977; Chopin, 1984). However, a metasomatic origin for the protolith has also been suggested (Chopin and Schertl, 1999). The dating of ultra-high pressure metamorphism in the Dora-Maira is essential for an understanding of Alpine geodynamics, but the question remains open. Early studies favoured a Cretaceous age of eclogitization (Paquette et al., 1989; Monié and Chopin, 1991), but these ages where questioned by Tilton et al. (1991) who published the first Eocene ages (35 Ma), later confirmed by SHRIMP U–Pb (Gebauer et al., 1997) and Lu–Hf (Duchêne et al., 1997) studies.

The Sesia-Lanzo zone consists of continental material of Apulian origin, metamorphosed under eclogitic conditions at 550 °C for a minimum pressure of 15 kbar (Lardeaux et al., 1982). The same debate on the eclogitization age exists in the Sesia-Lanzo zone as in the Dora-Maira Massif. Cretaceous ages of 129 Ma were first recognised by Oberhansli et al. (1985), but SHRIMP U–Pb (Rubatto et al., 1999), Lu–Hf (Duchêne et al., 1997), Rb–Sr and Ar–Ar (Ruffet et al., 1997) studies all result in consistent ages of around 60–65 Ma. Two samples were collected in the Aosta Valley. Sample Li6 was taken from the outcrop of Lillianes-Fontainemore. It is well equilibrated under eclogitic facies conditions and has a coarse grained (cm) assemblage of garnet, omphacite, quartz, phengite and rutile. REE patterns and major element chemistry indicate a possible andesitic origin for this sample. Sample Qu comes from the region of Quincinetto. It is a mafic eclogite, which represents an enclave in a metasedimentary lithology, with large (cm) clinopyroxene, glaucophane and phengite, garnets of 500 μm in diameter, and accessory rutile.

3. Analytical techniques

The main mineral phases from eclogites, such as garnet, pyroxene, amphibole and muscovite were first separated using heavy liquids, then using a Frantz magnetic separator. The separation was refined by hand-picking, under a binocular microscope equipped with cold light, to ensure that the minerals did not contain any visible inclusions. Mineral separates were cleaned with de-ionised water and then hand-crushed in an agate mortar.

In order to avoid any problems of sample heterogeneity, the same purified fraction was used to measure REE contents by quadrupole ICPMS, Sm and Nd contents by isotopic dilution, and Nd isotopic composition. These fractions are also the same as those on which Lu–Hf analyses were made (Duchêne et al., 1997; de Sigoyer et al., 2000).

REE concentrations in separated minerals and whole-rock have been determined on about 3-mg aliquots using a quadrupole ICP-MS (Tables 1A and 1B). The Sm and Nd analyses by ICP-MS are not precise enough for geochronological purposes, but can be used as estimates to ensure an adequate Sm–Nd spiking, for the subsequent accurate determination of Sm and Nd abundances by the isotope dilution method.
## Table 1A
Major- and trace-elements analyses for Himalayan samples (major elements in wt.% and trace elements in ppm)

<table>
<thead>
<tr>
<th>Element</th>
<th>Whole-rocks</th>
<th>Garnets</th>
<th>Omphacites</th>
<th>Amphibole</th>
<th>Muscovite</th>
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<tr>
<td></td>
<td>Ch171c</td>
<td>Ts34</td>
<td>Ts51</td>
<td>Ts45</td>
<td>Ch171c</td>
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<tr>
<td>SiO₂</td>
<td>46.19</td>
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<td>48.21</td>
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<td>Al₂O₃</td>
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<td>CaO</td>
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<td>Na₂O</td>
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<tr>
<td>TiO₂</td>
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<td>Total</td>
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<td>100.8</td>
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Cu  64.7  59.1  58.8
Ni  95.5  96.8  98.5
Co  54.7  59.2  55.8
Cr  209.3 180.1 178.9
V   389.4  398  373.8
Sc  33.2  33.8  31.4
Ba  70.2  70.0  60.6
Rb  61.4  4.4  20
Sr  217.9  88.1 261.8
Y   32.5  36.8  25.1
Zr  171.8 176.7 144.7
Nb  8.1  9.5  7.1

La  8.203 11.68 10.77 16.24 0.422 2.443 1.398 0.989 6.104 6.049 4.430 1.501 3.909
Ce  16.81 30.01 28.41 41.18 0.990 6.170 3.595 2.521 12.42 15.38 11.92 3.695 10.29
Pr  2.245 4.457 4.288 6.202 0.109 0.876 0.509 0.379 1.690 2.280 1.821 0.565 1.526
Sm  2.612 5.006 4.840 6.945 0.590 2.028 1.497 1.021 1.904 2.439 2.164 0.903 1.319
Eu  0.902 1.756 1.665 2.355 0.635 1.714 1.295 0.913 0.579 0.738 0.651 0.548 0.399
Gd  2.938 5.024 4.525 6.616 3.634 6.763 4.982 4.341 1.427 1.612 1.319 2.331 0.722
Tb  0.547 0.829 0.709 1.002 1.172 1.555 1.130 1.369 0.104 0.129 0.100 0.669 0.045
Dy  3.365 4.814 3.948 5.377 7.628 9.082 6.608 9.859 0.243 0.345 0.263 4.895 0.095
Ho  0.719 0.982 0.803 1.072 1.516 1.750 1.273 2.119 0.028 0.045 0.033 1.011 0.013
Er  1.929 2.541 2.031 2.696 3.830 5.178 3.837 6.599 0.048 0.102 0.070 2.741 0.024
Tm  0.291 0.363 0.290 0.388 0.552 0.606 0.441 0.771 0.006 0.010 0.006 0.269 0.003
Yb  1.761 2.184 1.746 2.304 3.331 3.653 2.676 4.599 0.038 0.082 0.046 1.445 0.024
Lu  0.263 0.311 0.255 0.339 0.494 0.552 0.392 0.666 0.005 0.010 0.006 0.217 0.003

La/Sm₀
La/Yb₀
Sm/Nd₀
La/Gd₀

Major and trace elements (Y – Ba) by X-ray fluorescence (P. Capiez, Laboratoire Sciences de la Terre, Univ. Claude Bernard, Lyon). Analytical precision is within ±1%.
Trace and REE analyses by ICP-MS (F. Keller, Institut Dolomieu, Grenoble). Precisions are better than 5%. Normalisation to chondrites from Nakamura (1974).
## Table 1B
Major- and trace-element analyses for Alpine samples (major elements in wt.% and trace elements in ppm)

<table>
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<tr>
<th>Component</th>
<th>Whole-rocks</th>
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<th>Omphacites</th>
<th>Phengite</th>
<th>Glauconides</th>
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<td>45.19</td>
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<tr>
<td>Ba</td>
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<td>0.789</td>
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</table>

Major element compositions by X-ray Fluorescence (P. Capiez, Laboratoire Sciences de la Terre, Univ. Claude Bernard, Lyon).

Analytical precision is within ± 1%.

Trace elements and REE by ICP-MS (ISTEEM, Montpellier). Reproducibility is better within 2 – 5%, except for Yb, Lu, Tb and Pb for which it is 6 – 8%. Normalisation to chondrites from Nakamura (1974).
Sm and Nd contents by isotopic dilution and Nd isotopic compositions were determined on the remaining 50–100 mg of the same purified separates (Tables 2A and 2B). Contents and isotopic compositions were measured on the same LREE + MREE fraction after one step chemistry using the Plasma 54 (VG Instruments, ENS-Lyon) according to the method described by Luais et al. (1997). Over the period of analysis, isotopic measurements of a 0.5-ppm diluted JMC Nd standard with a typical internal precision of 18 ppm ($2\sigma_m$) gave a mean value and external reproducibility in $^{143}$Nd/$^{144}$Nd ratios of 0.512241 ± 0.000036 (unweighted $2\sigma$ S.D., $n = 16$), whilst one La Jolla measurement with the internal precision gives $^{143}$Nd/$^{144}$Nd = 0.511835 ± 13 ($2\sigma$ S.E.). Isotopic measurements in low Nd abundance samples (e.g. garnet samples with less than 80 ng of Nd) were achieved with a good internal precision of ±17 ($2\sigma_m$). The values were corrected, using the exponential law, for instrumental mass fractionation using $^{146}$Nd/$^{144}$Nd = 0.7219 (Wasserburg et al., 1981).

For samples Ts34 whole-rock and Ch171c whole-rock, replicates from the same dissolution were performed and were reproducible within the analytical error (Table 2A).

Sample Qu was analysed in the Cosmochemistry Department of the Max Planck Institute of Mainz (Germany). Analytical techniques have been described by Thöni and Jagoutz (1992). $^{147}$Sm/$^{144}$Nd ratios by isotope dilution and Nd isotopic compositions were measured by thermal-ionisation mass spectrometry.

### 4. Results

#### 4.1. Himalaya

#### 4.1.1. Major elements (Table 1A)

Samples Ts34, Ts51 and Ts45 are alkalic to tholeiitic basalts with SiO$_2$ contents varying from 46.19%

<table>
<thead>
<tr>
<th>Table 2A</th>
<th>Sm–Nd isotope data for Himalayan samples</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Sm (ppm)</td>
</tr>
<tr>
<td>Ch171c (Fresh eclogite)</td>
<td></td>
</tr>
<tr>
<td>Ch171c whole-rock 1</td>
<td>2.90</td>
</tr>
<tr>
<td>Ch171c whole-rock 2</td>
<td>2.90</td>
</tr>
<tr>
<td>Ch171c garnet</td>
<td>0.86</td>
</tr>
<tr>
<td>Ch171c omphacite</td>
<td>2.52</td>
</tr>
<tr>
<td>Ts34 (Fresh eclogite)</td>
<td></td>
</tr>
<tr>
<td>Ts34 whole-rock 1</td>
<td>5.46</td>
</tr>
<tr>
<td>Ts34 whole-rock 2</td>
<td>5.46</td>
</tr>
<tr>
<td>Ts34 garnet</td>
<td>2.38</td>
</tr>
<tr>
<td>Ts34 omphacite</td>
<td>2.54</td>
</tr>
<tr>
<td>Ts51 (Slightly retrogressed)</td>
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<tr>
<td>Ts51 whole-rock</td>
<td>5.02</td>
</tr>
<tr>
<td>Ts51 garnet</td>
<td>2.47</td>
</tr>
<tr>
<td>Ts51 omphacite</td>
<td>2.92</td>
</tr>
<tr>
<td>Ts51 phengite</td>
<td>1.67</td>
</tr>
<tr>
<td>Ts45 (Amphibolite)$^b$</td>
<td></td>
</tr>
<tr>
<td>Ts45 whole-rock$^b$</td>
<td>7.44</td>
</tr>
<tr>
<td>Ts45 garnet$^b$</td>
<td>1.61</td>
</tr>
<tr>
<td>Ts45 amphibole$^b$</td>
<td>1.05</td>
</tr>
</tbody>
</table>

The numbers in parentheses are twice the standard error.

$\varepsilon_{Nd}$ calculated relative to chondritic uniform reservoir (CHUR) at $T = 55$ Ma (except $T = 47$ Ma for Ts45).

$T_{DM}$ is model age for depleted mantle following DePaolo (1981).

$^a$ Lu–Hf ages from Duchêne et al. (1997).

$^b$ From de Sigoyer et al. (2000).
to 48.21%, fresh eclogite having the lowest contents of 46.19%. Contents in Al₂O₃, FeO, MgO and TiO₂ are similar whatever the metamorphic facies. Only K₂O contents differ, with a decrease in K₂O contents with increasing degree of retrograde metamorphism, from 0.91% in fresh eclogite (Ts34) to 0.08% in amphibolite facies (Ts45).

4.1.2. Rare earth elements (Table 1A and Fig. 1A,B)

Total REE contents in whole-rock (Fig. 1A) and in garnet (Fig. 1B) are much lower in fresh eclogite Ch171c associated to the orthogneiss core, than in samples Ts34, Ts51, Ts45 associated with sedimentary cover, whatever their degree of metamorphism. Ch171C also differs by its high LREE/MREE fractionation (\((La/Sm)_n = 1.94\)) compared to samples Ts34, Ts51 and Ts45 with \((La/Sm)_n = 1.37–1.44\). This shows that lithological changes induce much larger variations than metamorphic grade.

The degree of retrograde metamorphism induces LREE/HREE enrichment in whole-rock, from \(La/Yb_n = 3.58\) in fresh eclogite facies Ts34, to \(La/Yb_n = 4.12\) in blueschist facies Ts51, then to \(La/Yb_n = 4.72\) in amphibolite facies Ts45. In garnet, rare-earth patterns cross over in the region of Gd, resulting in coherent LREE–MREE variations with the degree of metamorphism: decreasing pressure results in a decrease in LREE and MREE contents, and a preferential depletion in LREE over MREE, inducing a fractionation of the Sm/Nd ratio. Garnets in the eclogite facies sample (Ts34) are enriched in LREE (Laₙ to Ndₙ = 6–7.5 ppm) and MREE (Smₙ to Gdₙ = 10–25 ppm) and have high La/Gd and low Sm/Nd ratios (La/Gdₙ = 0.303, Sm/Ndₙ = 1.66) with respect to garnets in the blueschist facies (Ts51: Laₙ to Ndₙ = 3.5–4.4 ppm, Smₙ to Gdₙ = 7–18 ppm, La/Gdₙ = 0.235, Sm/Ndₙ = 2.02) and then to garnets in the amphibolite facies (Ts45: Laₙ to Ndₙ = 2.7–3.3).
ppm, Sm<sub>n</sub> to Gd<sub>n</sub> = 5–16 ppm, La/Gd<sub>n</sub> = 0.191, Sm/Nd<sub>n</sub> = 1.865).

4.1.3. Nd isotopic compositions

Sm and Nd contents, 147Sm/144Nd and 143Nd/144Nd ratios for the fresh eclogites (Ts34 and Ch171c), the eclogite slightly recrystallized under blueschist facies (Ts51), and the eclogite retrofomed under amphibolite facies (Ts45) are reported in Table 2A.

4.1.3.1. Eclogites. For each rock, whole-rock, omphacite and garnet have been analysed. Results are plotted in Fig. 2A.

Ts34: Nd and Sm contents range from 4.7 ppm in garnet to 22.6 ppm in whole-rock, and 2.4 ppm in garnet to 5.5 ppm in whole-rock, respectively. Omphacite has intermediate Nd and Sm contents. The resulting 147Sm/144Nd ratios vary from 0.139 in omphacite to 0.307 in garnet, whereas 143Nd/144Nd ratios range from 0.512767 to 0.512746 for whole-rock to 0.512203 for garnet. Omphacite and whole-rock have similar 147Sm/144Nd and 143Nd/144Nd ratios. Garnet, which presents the highest 147Sm/144Nd ratio, exhibits the lowest 143Nd/144Nd ratio. This results in a regression line with a negative slope, and no isochron can be defined.

Ch171c: The same scheme as for Ts34 is observed, with a large range in 147Sm/144Nd from 0.149 in clinopyroxene to 0.554 in garnet, whereas 143Nd/144Nd ratios show large variations from 0.512312 in whole-rock to 0.510299 in garnet. Omphacite and whole-rock plot close to each other. Because of the low 143Nd/144Nd in garnet, the regression line has a negative slope, and here again no isochron can be defined.
4.1.3.2. Eclogite slightly retrogressed in the blueschist facies. Ts51 (Fig. 2A): The large range in \( ^{147}\text{Sm}/^{144}\text{Nd} \), from 0.123 in muscovite to 0.403 in garnet is not accompanied by significant variations in \( ^{143}\text{Nd}/^{144}\text{Nd} \) whose values of 0.512835–0.512904 vary within the analytical error. Muscovite, clinopyroxene and whole-rock \( ^{147}\text{Sm}/^{144}\text{Nd} \) and \( ^{143}\text{Nd}/^{144}\text{Nd} \) ratios cannot be distinguished. Thus, no isochron can be defined.

4.1.3.3. Eclogite recrystallized under amphibolite facies. Ts45 (Fig. 2A): Whole-rock, amphibole and garnet exhibit a large range in \( ^{147}\text{Sm}/^{144}\text{Nd} \) from 0.145 to 0.406 and a significant range in \( ^{143}\text{Nd}/^{144}\text{Nd} \), with garnet having the highest \( ^{147}\text{Sm}/^{144}\text{Nd} \) and \( ^{143}\text{Nd}/^{144}\text{Nd} \) ratios. The regression line with a positive slope gives an isochron with an age of 47 ± 11 Ma (de Sigoyer et al., 2000).

It must be emphasised that, the fresher the eclogite, the greater the Sm–Nd disequilibrium, and the lower the whole-rock \( ^{143}\text{Nd}/^{144}\text{Nd} \) ratio.

4.2. Alps

4.2.1. Major elements (Table 1B)

The four Alpine samples have contrasting compositions. The quartzite from Dora-Maira (DM) is silica-rich (66% SiO₂) and contains almost only magnesium (9.82% MgO), aluminium (17% Al₂O₃) and potassium (3% K₂O) as other constituents. The Monviso sample Vi8 has a mafic composition (45% SiO₂) and the Sesia-Lanzo sample Li6 an intermediate composition (52.29% SiO₂). Sample Qu has not been analysed for major elements but REE patterns suggest a mafic composition.

4.2.2. Rare earth elements (Table 1B and Fig. 1C and D)

Rare earth elements have been analysed in samples Li6 and Qu (Sesia-Lanzo) and sample Vi8 (Monviso). Data for sample DM (Dora-Maira) whole-rocks have already been published by Paquette et al. (1989).

Bulk REE contents are higher in the Dora-Maira quartzite (\( \sum_{\text{REE}} = 193 \) ppm for the whole-rock) than in the meta-andesite Li6 from Sesia-Lanzo, the mafic eclogite Qu (\( \sum_{\text{REE}} = 25 \) ppm for the whole-rock). REE patterns for the mafic samples Qu and Vi8 are MORB-type, with \( (\text{La/Yb})_n = 0.5 \) and 1, respectively, and \( (\text{Sm/Nd})_n = 1.20 \) and 0.90. The two other samples are enriched in LREE with \( (\text{La/Yb})_n = 10 \) and 6 for DM and Li6, respectively, and \( (\text{Sm/Nd})_n = 0.6 \) and 0.7.

4.2.3. Nd isotopic compositions

Sm and Nd contents and \( ^{143}\text{Nd}/^{144}\text{Nd} \) ratios for eclogites are reported in Table 2B.

4.2.3.1. Mafic eclogites from Monviso (Vi8) and Sesia-Lanzo (Qu). Vi8 (Fig. 2B): \( ^{147}\text{Sm}/^{144}\text{Nd} \) ratios range from 0.18 for whole-rock and clinopyroxene to 0.65 in garnet. The garnet fraction has a more radiogenic \( 143\text{Nd}/^{144}\text{Nd} \) ratio (0.513131) than the clinopyroxene (0.513041) and whole-rock (0.513012) fractions. However, because of analytical problems, the whole-rock–garnet age of 40 ± 10 Ma must be regarded with caution (Table 2B).

Qu (Fig. 2B): \( ^{147}\text{Sm}/^{144}\text{Nd} \) ratios range from 0.22 for glaucophane and clinopyroxene to 0.86 in garnet. The garnet is less radiogenic than the clinopyroxene and the glaucophane, which leads to a negative slope, therefore no age can be calculated. It should be noted that the Nd content in the garnet fraction is very low (0.08 ppm).

4.2.3.2. Felsic eclogite from Sezia-Lanzo (Li6) and Dora-Maira (Dm). Li6 (Fig. 2B): Analyses have been performed on three whole-rock powder aliquots, phengite, clinopyroxene, and two garnet fractions. One of these garnet fractions, named hereafter “impure garnet”, was purified by magnetic separation only, and thus may contain some tiny grains of other major mineral phases of the rock, i.e. cpx, phengite, and constitute with garnet an assemblage of “mixed grains”. The “pure garnet” fraction corresponds to...
garnet grains purified by hand-picking and are free from any visible inclusions. Data for mineral phases from Sesia-Lanzo sample Li6 are extremely scattered. They define two groups according to their $^{143}$Nd/$^{144}$Nd ratios: cpx and phengite have high $^{143}$Nd/$^{144}$Nd ratios (0.513154–0.513198); whole-rock and garnet have significantly lower $^{143}$Nd/$^{144}$Nd values of 0.510662 for pure garnet and 0.512093 for whole-rock. The “impure garnet” plots on the line defined by the whole-rock and pure garnet, and close to whole-rock composition, which is easily explained by a mixing between a “pure garnet” fraction (i.e., garnet which could contain sub-microscopic inclusions) and a “whole-rock” component (i.e., major phases present in mixed grains). Although the highest Sm/Nd ratios are found in garnet, it exhibits the lowest $^{143}$Nd/$^{144}$Nd ratios. Clinopyroxene and phengite compositions define a regression line with a positive slope, the resulting isochron giving a poorly defined age of 63 ± 50 Ma. By contrast, whole-rock, “impure garnet,” and “pure” garnet compositions define a regression line with a negative slope from which no geochronological information can be obtained.

**DM** (Fig. 2B): Dora-Maira whole-rock and garnet compositions have no chronological meaning: the regression line has a strong positive slope which would give an age of 2 Ga. Garnet has $^{147}$Sm/$^{144}$Nd and $^{143}$Nd/$^{144}$Nd ratios of 0.089 and 0.511738, respectively, lower than the whole-rock ($^{147}$Sm/$^{144}$Nd = 0.1228 and $^{143}$Nd/$^{144}$Nd = 0.512189).

### 5. Discussion

Metamorphic ages are given by the slope of the regression line between the different minerals, which in the case of high pressure rocks, is strongly controlled by the high $^{143}$Nd/$^{144}$Nd and Sm/Nd ratios of garnet. The above examples on mineral phases from eclogites show that such an approach fails in most cases. In fresh (Ts34, Ch171c) and slightly retrogressed (Ts51) mafic eclogites from Tso Morari, and felsic and mafic eclogites from Sesia-Lanzo (Li6 and Qu), garnet was in strong disequilibrium with whole-rock and other phases (omphacite, phengite...) at the time of eclogitization, as its unradiogenic $^{143}$Nd/$^{144}$Nd composition defines a negative slope. On the other hand, the Sm/Nd ratio in these garnets seems to be adequately and significantly higher than these phases. The case of Dora-Maira sample (DM) is even more striking as both $^{143}$Nd/$^{144}$Nd and Sm/Nd ratios are lower than the whole-rock. We will discuss below the two following hypotheses to explain the low $^{143}$Nd/$^{144}$Nd ratio in garnet: (1) the incomplete equilibration of garnet during the transformation to eclogite, (2) the presence of inherited inclusions with low $^{143}$Nd/$^{144}$Nd. We will then explore the role of retrograde metamorphism on garnet equilibrium during blueschist and amphibolite metamorphism.

#### 5.1. Modeling the low $^{143}$Nd/$^{144}$Nd of garnet

##### 5.1.1. Isotopic composition inherited from precursor plagioclase

Several authors have related such garnet isotopic disequilibrium in eclogites with inefficient re-equilibration at the new P, T conditions of eclogite metamorphism (Mørk and Mearns, 1986; Thöni and Jagoutz, 1992). Garnet crystallizes partly from pre-existing plagioclase, with contrasting REE patterns (low Sm/Nd ratios / C25 0.2) (Fig. 3A). This reaction implies the transfer of major elements between the igneous and the metamorphic minerals. The isotopic ratio of the plagioclase could be preserved in the growing garnet provided that minor elements such as REE are less mobile during mineral reactions than major elements. Even if a higher Sm/Nd ratio is usually observed in garnet (1.05) in disequilibrium, than in plagioclase (0.18) (Thöni and Jagoutz, 1992), the very low Sm/Nd values of 0.5 in garnet from sample Ts34 compared with Sm/Nd values of 0.67 in garnet in equilibrium from Ts45, indicates that the Sm/Nd ratio is not the equilibrium ratio for garnet, but reflects the strong disequilibrium also shown by the extremely low $^{143}$Nd/$^{144}$Nd ratio. Sm and Nd concentrations, as well as Nd isotopic compositions, therefore both reflect disequilibria. For the Himalayan sample Ts34, a simulation of this model is presented in Fig. 3B. The initial $^{143}$Nd/$^{144}$Nd isotopic composition of plagioclase before eclogitization at 55 Ma has been recalculated from the $^{147}$Sm/$^{144}$Nd and $^{143}$Nd/$^{144}$Nd ratios of garnet, assuming a $^{147}$Sm/$^{144}$Nd for plagioclase of 0.1 (Thöni and Jagoutz, 1992). The value that is obtained is a minimum value because the Nd isotopic ratio of the pseudomorphous garnet should be intermediate between the initial plagioclase and the
initial whole-rock compositions (see Fig. 3A). The plagioclase-whole-rock equilibrium gives a meaningless minimum age of about 2000 Ma. This is not consistent with the stratigraphic ages of 258–275 Ma for the protolith (Kapoor, 1977; Nakazawa et al., 1975; Gaetani et al., 1990), which is believed to correspond to the Panjal Traps (de Sigoyer, 1998). The hypothesis that the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio is partly inherited from the plagioclase of the magmatic protolith is therefore not valid in the case of the Tso Morari.
The same can be demonstrated for the Alpine samples of Dora-Maira and the Sesia-Zone, with meaningless minimum ages (>3600 Ma). Another argument for plagioclase inheritance could be that element re-equilibration in the newly formed garnet is faster than isotopic re-equilibration (for a given $^{147}$Sm/$^{144}$Nd of garnet, $^{143}$Nd/$^{144}$Nd is too low and lies off a whole-rock–cpx isochron), which is the opposite of experimental studies results based on silicate melts (Lesher, 1990). Therefore, it can be concluded that the inheritance of the isotopic signature of plagioclase is the not the cause for Nd isotopic disequilibrium in Alpine and Himalayan samples.

5.1.2. Low $^{143}$Nd/$^{144}$Nd and Sm/Nd inherited mineral inclusions

Garnets in eclogites are rarely pure. The inclusions that can be observed under the microscope are either equally distributed in the mineral, or concentrated (1) in the core of the garnet, leaving an inclusion-free rim (Vance and O’Nions, 1990; Burton and O’Nions, 1991; de Sigoyer, 1998), or (2) in the rim of garnet, such as coesite inclusions in garnet from UHP eclogite of the Zermatt-Saas eclogite (Amato et al., 1999), or graphite inclusions in pelitic schists of Scotland (Vance et al., 1998), leaving a inclusion-free core. Although care has been taken in the purification of garnets of Alpine and Himalayan samples, and even in the absence of visible inclusions, sub-microscopic inclusions can still be present in the mineral separates. It has often been observed that the whole-rock composition cannot be defined by the mixing of major mineral phases (Fig. 2A and B), which confirms the occurrence of other minor phases such as inclusions of accessory minerals. In this case, the low measured $^{143}$Nd/$^{144}$Nd and Sm/Nd ratios of garnet separates could result from the mixing between pure garnet and inherited inclusions with Sm/Nd and $^{143}$Nd/$^{144}$Nd ratios that are lower than the measured garnet. Such inclusions must have very high Nd contents in order that a small amount can obliterate the $^{143}$Nd/$^{144}$Nd signature of garnet. Moreover, their Hf contents and the Lu/Hf ratio must be extremely low since the Hf isotopic system does not show any disequilibrium (Duchêne et al., 1997; de Sigoyer et al., 2000). Finally, they must be resistant to high-pressure metamorphism.

Among the potential inherited accessory mineral phases, zircon can be excluded because of its extremely high Hf contents (about 2% Hf, Fujimaki et al., 1984; Fujimaki, 1986). Sphene is present in eclogites in equilibrium with the eclogitic assemblage, or more often, it appears as the product of rutile breakdown. In all cases, it is never described as an inherited phase.

Apatite, monazite, and allanite are the LREE-rich accessory minerals described in eclogite facies rocks (Mørk and Mearns, 1986; Hensen and Zhou, 1995; Getty et al., 1993) which have the required characteristics. REE chemical determination on mineral separates (Fujimaki, 1986; Gromet and Silver, 1983) and in-situ microprobe or SIMS analyses (Bea, 1996; Pan and Fleet, 1996) have revealed a large range of LREE contents, but systematically higher Nd contents in monazite (0.1–1.5%) than allanite (0.016–0.4%) and apatite (150–1700 ppm), and low $^{147}$Sm/$^{144}$Nd
ratios ranging from 0.05 to 0.1. Mixing of these relict inclusions with garnet has been simulated for two samples where garnet is in strong disequilibrium: sample Ts34 from Tso Morari, and sample Li6 from Sesia-Lanzo. Results are presented in the $^{143}\text{Nd} / ^{144}\text{Nd}$--$^{147}\text{Sm} / ^{144}\text{Nd}$ diagram (Fig. 4A and B). The $^{143}\text{Nd} / ^{144}\text{Nd}$ composition of pure garnet in equilibrium with whole-rock has been recalculated, assuming a $^{147}\text{Sm} / ^{144}\text{Nd}$ of 2, a Lu–Hf age of 55 Ma for Ts34 (de Sigoyer et al., 2000), or a mean age U–Pb, Lu–Hf and Sm–Nd (cpx–phengite) of 65 Ma for Li6 (Duchêne et al., 1997). If the analysed garnet separate is assumed to result from a mixture between the pure garnet and the inclusions, the $^{143}\text{Nd} / ^{144}\text{Nd}$ isotopic composition of the latter can be plotted on the measured-garnet/pure-garnet regression line for any $^{147}\text{Sm} / ^{144}\text{Nd}$ ratio. Assuming that the inclusions are cogenetic with the protolith, a whole-rock/inclusion age is calculated. The $^{147}\text{Sm} / ^{144}\text{Nd}$ value of the relict inclusion is critical in the determination of this age. A
minimum $^{147}\text{Sm}/^{144}\text{Nd}$ value of 0 for the inclusion gives a minimum age, e.g. 823 Ma in the case of Ts34, and 2670 Ma in the case of Li6. Allanite inclusions with very low $^{147}\text{Sm}/^{144}\text{Nd}$ of 0.047 (Gromet and Silver, 1983) induce an age of 1163 Ma for Ts34 and 3800 Ma for (Li6). However, $^{147}\text{Sm}/^{144}\text{Nd}$ of 0.1, as found in apatite (Fujimaki, 1986; Gromet and Silver, 1983) or monazite (Bea, 1996), gives ages of 3 Ga (Ts34) and >4.5 Ga (Li6). In all cases, the percentage of inclusions which explains the measured-garnet composition is very low ($<0.05\%$). Note also that if we simulate the same mixing process for the Lu–Hf system, we find that the garnet $^{176}\text{Hf}/^{177}\text{Hf}$ ratio is not modified greatly by the presence of these inclusions since they do not contain appreciable amount of Hf. The difference between the ages calculated from whole-rock/pure-garnet and whole-rock/measured-garnet is less than 2 Ma, which is within the range of the uncertainty on the age. Different $^{147}\text{Sm}/^{144}\text{Nd}$ ratios for pure garnet have been tested in this model, and the resulting whole-rock/inclusion ages are all Precambrian.

The ages obtained from whole-rock/inclusion pairs do not agree with the protolith age of these eclogites. In the Tso Morari, the protolith of eclogitic samples Ts34, Ts45, Ts51 corresponds to the Panjal Traps (de Sigoyer, 1998) with stratigraphic ages of 258–275 Ma (Kapoor, 1977; Nakazawa et al., 1975; Gaetani et al., 1990), as further demonstrated by their similar $\varepsilon_{\text{Nd}}$ at 55 Ma (47 Ma for Ts45) (Fig. 5). The protolith of the eclogite Ch171c could be the Ordovician orthogneiss ($T=458 \pm 14$ Ma, de Sigoyer, 1998). In the Sesia zone, it is related to Hercynian metagranitoids dated at 280 Ma (Oberhänsl et al., 1985; Rubatto et al., 1999). The assumption that inclusions are in equilibrium with these magmatic protoliths is therefore not valid. The low $^{143}\text{Nd}/^{144}\text{Nd}$ of the inclusions suggests their crustal origin, implying that they originate through crustal contamination of the magmatic protolith.

For Alpine and Himalayan samples that show isotopic disequilibrium, crustal contamination of the protolith has been well demonstrated. Indeed, in the Alpine eclogites, both the Sesia-Lanzo zone and the Dora-Maira massif have been recognised as slices of
and continental crust (Compagnoni et al., 1977; Schreyer, 1977; Chopin, 1984). Crustal signature is further demonstrated by their negative $\varepsilon_{\text{Nd}}$ of $-10.15$ and $-8.46$, respectively (Table 2B). Moreover, zircons from the Alpine eclogites have inherited Precambrian cores. The oldest age reported in Rubatto et al. (1999) is 2.36 Ga for the Sesia-Lanzo zone. This age is older than the model ages of 2.08 and 1.48 Ga calculated for Li6 (Sesia Lanzo) and DM (Dora-Maira), respectively (Table 2B). The only sample for which no Sm–Nd disequilibrium is observed is the metagabbro from the Monviso ophiolite (Duchêne et al., 1997), where continental crust has never been involved, in agreement with the positive $\varepsilon_{\text{Nd}}$ value ($\varepsilon_{\text{Nd}} = 7.4$) (Table 2B). In the Himalayas, crustal contamination of the protolith by Himalayan crust is demonstrated by the strongly negative $\varepsilon_{\text{Nd}}$ of both the Panjal Traps ($\varepsilon_{\text{Nd}} = +3.5$ to $-9.9$, Spencer et al., 1995) and the Orдовician orthogneiss ($\varepsilon_{\text{Nd}} = -11$, de Sigoyer, 1998). In addition, the lower the $^{147}\text{Sm}/^{144}\text{Nd}$ of garnet, the stronger the disequilibrium between garnet and whole-rock at 55 Ma, and the lower the $\varepsilon_{\text{Nd}}$ of whole-rock (Table 2A, Fig. 5), which demonstrates that inclusions in garnet represent the contaminant phase in these eclogitic rocks. The $\varepsilon_{\text{Nd}}$ values ($\varepsilon_{\text{Nd}} = -10$ to $-11$) (Fig. 5) and the ages (1.2–3 Ga) of allanite and monazite inclusions, as calculated from the modeling (Fig. 4), are consistent with $\varepsilon_{\text{Nd}}$ values (< −5), and with Sm–Nd ages (2.8–3.3 Ga) and Nd model ages ($T_{\text{DM}} = 1.3–2.8$ Ga) of the western Himalayan crust (Gopalan et al., 1990, Whittington et al., 1999).

The processes by which these REE-rich accessory minerals have contaminated the magmatic protolith can now be evaluated. The magma could have assimilated these crustal inclusions during its ascent through the crust. However, it is questionable if inclusions of REE-rich minerals, especially allanite and monazite, can survive their incorporation in high T magmas, and thus be inherited from the crust. For example, monazite is not stable during metamorphism over $T=750–850$ °C (Bingen and van Breemen, 1998; Rubatto et al., 2001), and it is highly soluble in mafic high temperature melts (Montel, 1993). However, accessory mineral inheritance has already been described in basic rocks. For example, allanite has been described in diorite (Kosterin et al., 1961). Prince et al. (2000) demonstrated that the Sm–Nd systematics of garnet from granulite gneiss of the Southern Bohemian massif, with extreme $P–T$ conditions of 20 kb–1100 °C, necessitate the occurrence of allanite and monazite inclusions.

Searches for these accessory mineral inclusions in the Alpine and Tso Morari samples have been made by careful examination of pure garnet fractions from Ts34, Vi 8 and Li6 samples, by using scanning electron microscopy (SEM) (Fig. 6). Zircon inclusions are often present in both samples. Apatite and (rare) allanite crystals have been found, but no monazite has been observed. These allanite inclusions, and most of the zircons, are submicrometric in size. Their very small size, as well as their anhedral rounded shape, suggest that they have been partially destabilised before incorporation. As suggested by Rubatto et al. (2001) for monazite, a large initial crystal size can be a determinant factor for their preservation: the $P–T$ conditions for their stability may have been reached before their complete dissolution, and relic cores may be preserved. We can therefore suggest that these crustal minerals were assimilated during the later stages of the crystallization of the mafic protolith.

The alternative hypothesis is that the REE-rich minerals recrystallized from REE-rich crustal fluids (Poitrasson et al., 1998). The REE-rich minerals occur as inclusions in metamorphic garnet and must predate the eclogitic metamorphism, and they also have crustal $^{143}\text{Nd}/^{144}\text{Nd}$ signatures. Therefore, the fluids must have interacted with the crust before they percolated through the mafic protolith. This episode of fluid circulation in the protolith, and thus the precipitation of REE-rich minerals, must have occurred during the latter stages of crystallization, for these minerals to be stable. However, it is surprising that allanites are not euhedral, as would be expected in the case of precipitation from late stage fluids.

5.2. The effect of retrograde metamorphism on the Sm–Nd disequilibrium

Tso Morari samples exhibit a decrease in the degree of Sm–Nd isotopic disequilibria with decreasing pressure: $^{144}\text{Nd}/^{144}\text{Nd}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ ratios in garnet increase progressively from eclogite to blueschist, and isotopic equilibrium is reached in
amphibolite facies (Ts45), where an isochron can be defined. Isotopic re-equilibration under amphibolite facies conditions could be enhanced by volume diffusion, which is usually more rapid at low pressure. There are no experimental data on the variation of the Nd diffusion coefficient in garnet with pressure, but we know (1) that the diffusion coefficient is similar to that of Mg at 1 atm (Ganguly et al., 1998a) and (2) that, according to the value of the activation volume,
the drop of pressure between eclogitic and amphibolitic facies may enhance Mg diffusion by a maximum of one order of magnitude (Ganguly et al., 1998b). Chemical zoning profiles of garnet in amphibolitic sample Ts 45 (de Sigoyer et al., 1997) show that garnets are zoned with respect to Mg contents, having Mg-depleted cores, and Mg-rich rims. Nd will almost certainly be distributed in a similar way (Vance and O’Nions, 1990; Ganguly et al., 1998b), and this would demonstrate that volume diffusion was insufficient for the homogenization of Nd at amphibolite temperatures (590–620 °C) and pressures (7–11 kbar) (de Sigoyer et al., 1997). The large size of garnet (about 1300 μm in diameter) precludes any complete re-equilibration (it would take over 1 Ga for Nd to diffuse out of the cores) (Ganguly et al., 1998a).

Isothermal equilibrium between garnet and whole-rock can be explained more easily by a second stage of fluid circulation (during retrograde metamorphism). The eclogite–amphibolite transition is characterised by hydration reactions. In addition, amphibolite metamorphism is associated with deformation which enhances fluid circulation: cracks filled with carbonates are often observed, both on the outcrop and mineral scale, as in the case of sample Ts34 (de Sigoyer et al., 1997). We suggest that isotopic equilibrium under amphibolite conditions is enhanced by fluid circulation at the mineral scale, which would preferentially dissolve inherited crustal micro-inclusions, and leach the mobile LREE with respect to the HREE from garnet. This is well demonstrated by progressive decrease of LREE/MREE (i.e., La/Gd) or LREE/HREE (i.e., La/Yb) ratios in garnet from fresh eclogite (Ts34) to blueschist (Ts51) and then amphibolite (Ts45) facies samples (La/Gd)_n = 0.303–0.191, (La/Yb)_n = 0.447–0.144 (Table 1A). Evidence for dissolution and leaching processes can even be seen in resistant minerals such as monazite (Poitrasson et al., 1996; Crowley and Ghent, 1999) and zircons (Sinha et al., 1992; Wayne et al., 1992; Vavra et al., 1996, 1999). However, the 147Sm/144Nd ratio of garnet in Ts45, and consequently 143Nd/144Nd, although higher than in Ts34, is still low for a fully equilibrated garnet, which suggests that traces of micro-inclusions remains. The presence of these inclusions, although in extremely low amounts, leads to a poorly defined age of 47 ± 11 Ma.

6. Conclusions

(1) Isotopic Sm–Nd disequilibria have been observed in numerous samples of Himalayan and Alpine high pressure rocks. Whole-rock, omphacitic pyroxene and garnet isotopic compositions fail to define isochrons, mainly because of the very low 143Nd/144Nd ratios, and the low Sm/Nd ratios of garnet.

(2) In contrast to previous studies, these disequilibria cannot be related to the preservation of the isotopic signatures of the minerals from which they grew (e.g., plagioclase), because this would imply unreasonable ages for the protolith.

(3) The disequilibria can be easily explained if garnet separates contain sub-microscopic (inherited) REE-rich inclusions, with low Sm/Nd ratios and crustal 143Nd/144Nd signature. Such inclusions must have very low Hf contents, because of the negligible perturbation of the Lu–Hf system. In addition, they must be resistant to metamorphism. SEM images showapatite and sub-microscopic allanite inclusions in garnet. Simulation of mixing between pure garnet compositions and allanite, apatite (and monazite) compositions, with 147Sm/144Nd ratios of 0.05–0.1, require that these inclusions must be older than the protolith, and that they represent the crustal contaminant pole of the protolith. The survival of these inclusions in a high-temperature mafic magma implies that they were assimilated by the magma during the latter stages of crystallisation in the crust. Alternatively, they could have been precipitated from a late crustal fluid (rich in LREE) that percolated through the protolith.

(4) An inverse correlation has been observed (in the Himalayan samples) between the extent of whole-rock–garnet disequilibrium, and the degree of retrogression. In general, the equilibration of the Sm–Nd system can be explained by synchronous fluid circulation and deformation under amphibolite conditions.

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