Mineral chemistry of mica-pyroxenite xenoliths in NeoProterozoic ultrapotassic syenitic magmas, NE Brazil

VALDEREZ P. FERREIRA¹, ALCIDES N. SIAL¹ and MANOEL J. M. CRUZ²

¹Federal University of Pernambuco, Dept. of Geology, P.O. Box 7852, Recife, PE, 50732-970, Brazil
²Federal University of Bahia, Institute of Geosciences, Salvador, BA, 40210-340, Brazil

ABSTRACT

Mica-pyroxenite xenoliths found in NeoProterozoic ultrapotassic syenitic magmas in northeastern Brazil, present modal phases, textural relationships, bulk-rock and mineral chemistries compatible with metasomatic alteration, they suffered in the mantle source. These inclusions are mainly composed of diopsidesalite (>80% of the bulk rock), with subordinate, in variable amounts, F-rich phlogopite, calcite, hyalophane, sphene, apatite, and traces of barite. Corrosion of pyroxene is observed in the vicinity of these late space-filling minerals.

Xenoliths show high K and Ba contents (K₂O = 1-5 wt%; Ba = 1430-5100 ppm), variable Cr (21-3000 ppm) and Ni (20-280 ppm), very low Nb (≤ 6 ppm), and high H₂O⁺ (1.3-7.2 wt%) and F (0.4-1.8 wt%) contents, being LREE-enriched.

Phlogopite is Cr- and Ti-poor, and F-rich; diopsid-salite show slight to major, complex, patchy zoning, characterized by no systematic elemental variation. These minerals are chemically similar to micas and pyroxenes from enriched mantle derived rocks, such as lamproites and lamprophyres.

Some characteristics of the ultrapotassic host magmas, such as enrichment in incompatible elements, could be explained by melting of a phlogopite-bearing pyroxenite source. Petrography and mineral chemistry of these xenoliths suggest invasion of CO₂-saturated fluid or melt, and therefore it is assumed that the mica pyroxenite is a product of partial metasomatism in the mantle source, and whose melting produced ultrapotassic magmas in the Borborema province.

Key words: Mica pyroxenite, phlogopite, mantle xenolith.

INTRODUCTION

Potassic to ultrapotassic plutonic rocks occur in most of the Brasiliano (= Pan African) foldbelts in the Borborema province, northeastern Brazil, usually as elongate plutons emplaced along shear zones. All known occurrences, mostly of syenitic composition, are characterized by remarkably high, although variable, contents of large-ion lithophile elements (e.g. K, Ba, LREE) and high δ¹⁸O and initial Sr ratios, and low εNd values. These chemical signatures could be ascribed to some crustal contamination, but petrographic and chemical characteristics of phlogopite-bearing pyroxenite xenoliths present in some of these plutons argue against it. In this paper, mineral chemistry data for the pyroxenite xenoliths are used to explain the chemical fingerprints of their host syenites, and to show that xenoliths record metasomatism in the mantle source, for the ultrapotassic magmas in this part of Brazil.
GENERAL GEOLOGY

The Borborema province is characterized by a complex network of sinuous and branched shear zones, which divide the province into elongate structural domains. These domains comprise re-worked Archean terrains and Proterozoic sequences, metamorphosed and deformed during the Brasiliano orogenic cycle (700-550 Ma), besides a Phanerozoic unmetamorphosed sedimentary cover (Santos et al., 1984). The NE-trending central segment of the Central structural domain - the Cachoeirinha-Salgueiro foldbelt (Fig. 1) - covers 20,000 km² and consists of a thick sequence of greenschist facies metasiltites, metapelites and fine-grained metarenites, with minor metavolcanic rocks, believed to represent distal turbidites of a flysch-type sequence (Santos et al., op cit.).

Five major granitoid associations are recognized in this belt (Sial & Ferreira, 1988): (1) high-K calc-alkalic (major porphyritic granodiorites associated with K-diorites); (2) calc-alkalic epidote-bearing tonalites to granodiorites; (3) trondhjemitic tonalites, associated with peralkaline syenites; (4) potassic to ultrapotassic peralkaline alkali feldspar syenites to alkali feldspar granites, some of them associated with alkali clinopyroxenites and (5) shoshonitic monzogranites to granites. The bulk of the intrusions was emplaced between 650 and 500 Ma. Ages of ultrapotassic rocks, in particular, vary from 600 to 510 Ma, attesting to that this kind of magmatism was peculiar to the stage when the Brasiliano cycle was coming to a close in the Borborema province.

Most of the peralkaline syenitic plutons are emplaced along the southern boundary of the Cachoeirinha-Salgueiro foldbelt, forming the so-called syenitoid line (Sial & Ferreira, 1988). Among these plutons, the 600 km² Triunfo batholith, which is the largest peralkaline pluton known in the Borborema province, was chosen for detailed study, because it provides a large amount of fresh rock exposures, and easy access to the area.

THE SYENITOID LINE AND THE TRIUNFO PLUTON

Plutons along the syenitoid line are ultrapotassic (K₂O up to 13 wt%) syenites. In the Triunfo pluton, the main rock type is medium-grained alkali feldspar syenite composed of perthitic microcline and aegirine-augite (which together comprises over 90% of bulk rock), with apatite, sphene, and very minor magnetite and blue amphibole (formed at the expenses of pyroxene) as accessory phases. The oxidized nature of these syenites is revealed by the association of aegirine augite-sphene-magnetite. The presence of both fine-grained and pegmatic-syenitic dikes, mineralogically and chemically similar to the host Triunfo syenite, suggests that the pluton, in spite of its large volume and great amount of modal feldspar (>80%), has not suffered large-scale differentiation, being compositionally similar to the parental magma, and cannot be attributed to cumulative processes (Ferreira, 1991). A six-point Rb-Sr isochron indicates an age of 583 ± 12 Ma for this syenite, with initial Sr ratios of 0.709, suggestive of enriched-mantle source (Ferreira, op cit.).

Two kinds of magmatic enclaves are found in the syenites. The most common type is of alkaline pyroxenite, which occurs as narrow syn-plutonic dikes or streaky, co-magmatic patches composed mainly of aegirine-augite. Another kind of inclusion, much less common, is of mafic syenite which shows emulsion-like texture, in which syenitic and alkali pyroxenitic materials are mutually interstitial, suggesting that these two contrasting compositions coexisted as liquids (Ferreira et al., 1994). The syenite host as well as the mafic syenitic inclusions are silica-saturated, enriched in K (K₂O/Na₂O up to 5.3 in the Triunfo pluton), characterized by enrichment of incompatible LIL and RE elements.

The fact that host syenite and alkaline pyroxenite bear the same mineral phases, together with (a) field relationships indicating they coexisted as liquids for a certain period of time, and (b) their similar, overlapping chemical compositions of mineral phases, as well as O, S, Sr and Nd isotopic signatures, led Ferreira et al. (1994) to pro-
Fig. 1 — Generalized geological map of the Borborema province, north-eastern Brazil, showing the major granitoid/syenitoid masses (after Santos et al., 1984). Open circle locates the Triunfo pluton.
pose that these two compositionally contrasting magmas were formed by a liquid immiscibility process, from an initially homogeneous liquid of the mafic syenitic composition.

PHLOGOPITE-BEARING XENOLITHS

Xenoliths are usually found as small, up to 10 cm-long, fine-to medium-grained green nodules, rounded to oval in shape, composed of diopsidesalite as major phase, accompanied by variable, but subordinate, amounts of phlogopite, calcite, sphene, apatite, K-feldspar, and traces of barite.

Diopsidesalite shows irregular, patchy zoning, each grain displaying a different pattern. This mineral often forms "mosaics" in which phlogopite, calcite, sphene, apatite, feldspar, and barite are interstitial, forming a network, which replaces clinopyroxene along its grain boundaries. Corrosion of pyroxene grain in the vicinity of these minerals, is indicative that they are late crystallized. In some cases, replacement of earlier mafic mineral (clinopyroxene?) by calcite produces a cryptocrystalline material, which forms "islands" surrounded by a carbonate "sea". The precipitation of calcite suggests the invasion of a fluid saturated in CO₂.

Among the minor phases, mica is the most abundant one, although it is absent in some samples. Mica occurs in two main textural generations; in one of them, mica is associated with and invading clinopyroxene. This earlier fabric of brown mica and pyroxene is penetrated and corroded by late mineral assemblage of sphene, apatite, calcite, K-feldspar and barite, and finer-grained, underformed mica. Calcite and feldspar invades mica and pyroxene along cleavages and narrow fractures. Sphene and apatite occur as anhedral crystals, usually enclosed by mica and poikilitic feldspar.

Similar textural relationships to these are described in alkali clinopyroxene xenoliths of ultrapotassic lavas in Uganda, which were ascribed by Lloyd et al. (1987) to metasomatism in the mantle and regarded as fragments from the source of their host lavas.

CHEMISTRY

ELEMENTAL CHEMISTRY OF THE MICA-PYROXENE XENOLITHS

Pyroxene xenoliths are K- and Ba-rich (K₂O = 1-3.2 wt%; Ba = 1430-5000 ppm), being characterized by intermediate to high Sr contents (450 to 1170 ppm), very variable Cr (21 to 3000 ppm), low to high Ni (20 to 280 ppm), low to intermediate Rb (33 to 220 ppm), low Y (16 to 36 ppm) and very low Nb (5-6 ppm) (Ferreira & Sial, 1993). High H₂O⁺ (1.3-7.2 wt%) and F (0.4-1.8 wt%) values are even higher than ranges for lamproites, the ultrapotassic group of rocks supposed to have the highest values for these volatiles (Foley, 1992).

Incompatible element concentrations compared to mantle compositions are significantly enhanced, particularly Ba, La, K, Ce, Nd, Sm, Sr and Y, all over 10 times the corresponding abundance in mantle peridotite (Fig. 2). Nb, P, Eu, Ti, and Yb form with troughs in this diagram. Spidergrams for the host syenite are also characterized by negative anomalies in Nb, P, and Ti, a characteristic interpreted as inherited from the source (Ferreira et al., 1994).

![Fig. 2 — Mantle composition normalized incompatible element envelope patterns for Triunfo mica pyroxene xenoliths. Values used for normalizaton are from Menzies et al. (1987).](image-url)
Chondrite-normalized REE patterns have steep negative slope with LREE enrichment and negative to almost absent Eu anomalies. Total REE varies from 56 to 328 ppm, the more enriched samples presenting more pronounced negative Eu anomaly (Fig. 3). Patterns are broadly similar to the patterns for enriched kimberlites and lamproites, according to Haggerty (1994). The host syenites are REE-enriched, exhibiting fractionated chondrite-normalized REE patterns, slightly more LREE-enriched, and usually lacking Eu anomaly (Ferreira et al., op cit.).

MINERAL CHEMISTRY

Mineral compositions were determined by wave-length-dispersive electron microprobe using a SX Cameca electron microprobe, equipped with a link energy dispersive system detector, at the Federal University of Bahia. The operating conditions were: 15 kV, 25 nA current, counting time 10-40s, for the elements Si, Ti, Fe, Al, Mg, Ca, Mn, Na, K, Sr and Ba. Values were corrected using the ZAF systematic. Representative chemical analyses are presented in Table I. Structural formulae for micas and pyroxenes were calculated on the basis of 24 and 6 oxygens, respectively, using the computer program "MINFILE". For micas, all iron was considered to be Fe$^{2+}$.

MICA

All mica grains analyzed are phlogopite in composition (Fig. 4), however, the two identified textural types of mica have some compositional differences. Tetrahedral site in both fine-grained and coarse-grained mica has 8.0 atoms, filled up with Si and Ti. Coarse-grained mica (CG) shows a greater number of tetrahedrally coordinated aluminum ions (1.87-1.89) than do fine-grained one (FG) (1.61-1.65). Both types of mica do not have excess of octahedral cations, but the vacancy in the FG mica (5.77-5.78) is greater than that in the CG one (5.86-5.88).

Both types of mica are characterized by small FeO variation (6.0-6.7 wt%), and low Ti (TiO$_2$ from 0.48 to 0.74 wt%) and Cr (Cr$_2$O$_3$ from 0.0 to 0.22 wt%) contents, although the concentrations of these two latter elements are distinct in the two types, CG mica being poorer in Ti, with Cr content zero. Moreover, CG mica has higher Al$_2$O$_3$, MnO and Na$_2$O than FG mica. As a whole, however, they are in the range exhibited by micas from MARID (mica, amphibole, rutile, ilmenite, diopside), in terms of Ti and Cr contents, according to Haggerty (1994) (Fig. 5). There is no systematic correlation of Cr$_2$O$_3$, TiO$_2$ and FeO among samples, as it is also observed in kimberlitic phlogopite compositions, which show considerable scatter on inter-element plots (Hunter et al., 1984).

Within individual mica grains, a limited, normal zoning, is observed for Ni, Cr, Mn and Fe, as there is a decrease from core to rim in the concentrations of these elements, and Sr and K tend to increase from core to rim of the grains. Reverse zoning is observed for Mg, which increases from core to rim, and as a consequence, Mg$#$ also increases in the same sense. Na and Ti have no systematic variation, as they can either increase or decrease from core to rim in different grains, in a single thin section.
TABLE I
Representative microprobe analyses for (a) mica; (b) pyroxene; (c) calcite; (d) barite (semiquantitative) and (e) feldspar from Triunfo mica-pyroxene xenoliths. Structural formulae based on 23 and 6 oxygens, for mica and pyroxene, respectively.

<table>
<thead>
<tr>
<th></th>
<th>a) Mica</th>
<th></th>
<th>b) Pyroxene</th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>2c</td>
<td>3b</td>
<td>5c</td>
<td>6b</td>
<td>6c</td>
<td>7b</td>
<td>12c</td>
<td>13b</td>
</tr>
<tr>
<td>SiO₂</td>
<td>42.16</td>
<td>42.12</td>
<td>43.88</td>
<td>44.01</td>
<td>SiO₂</td>
<td>53.82</td>
<td>53.70</td>
<td>51.85</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.55</td>
<td>0.48</td>
<td>0.63</td>
<td>0.74</td>
<td>TiO₂</td>
<td>0.09</td>
<td>0.01</td>
<td>0.28</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>12.69</td>
<td>12.47</td>
<td>10.96</td>
<td>10.64</td>
<td>Al₂O₃</td>
<td>0.69</td>
<td>0.73</td>
<td>3.47</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.00</td>
<td>0.00</td>
<td>0.22</td>
<td>0.12</td>
<td>Cr₂O₃</td>
<td>1.06</td>
<td>0.38</td>
<td>0.00</td>
</tr>
<tr>
<td>FeO</td>
<td>6.75</td>
<td>6.29</td>
<td>6.59</td>
<td>6.43</td>
<td>Fe₂O₃</td>
<td>5.46</td>
<td>6.70</td>
<td>1.13</td>
</tr>
<tr>
<td>MnO</td>
<td>0.22</td>
<td>0.17</td>
<td>0.13</td>
<td>0.12</td>
<td>FeO</td>
<td>3.56</td>
<td>2.89</td>
<td>4.40</td>
</tr>
<tr>
<td>MgO</td>
<td>21.79</td>
<td>21.95</td>
<td>21.57</td>
<td>21.96</td>
<td>MnO</td>
<td>0.23</td>
<td>0.19</td>
<td>0.51</td>
</tr>
<tr>
<td>CaO</td>
<td>0.12</td>
<td>0.00</td>
<td>0.04</td>
<td>0.00</td>
<td>NiO</td>
<td>0.03</td>
<td>0.00</td>
<td>0.08</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.25</td>
<td>0.31</td>
<td>0.21</td>
<td>0.20</td>
<td>MgO</td>
<td>12.14</td>
<td>12.58</td>
<td>13.09</td>
</tr>
<tr>
<td>H₂O</td>
<td>2.72</td>
<td>2.75</td>
<td>2.40</td>
<td>2.50</td>
<td>Na₂O</td>
<td>2.65</td>
<td>2.55</td>
<td>0.89</td>
</tr>
<tr>
<td>F</td>
<td>2.77</td>
<td>2.67</td>
<td>3.38</td>
<td>3.19</td>
<td>K₂O</td>
<td>0.02</td>
<td>0.00</td>
<td>0.01</td>
</tr>
<tr>
<td>Cl</td>
<td>0.00</td>
<td>0.00</td>
<td>0.01</td>
<td>0.04</td>
<td>Total</td>
<td>100.53</td>
<td>100.90</td>
<td>98.93</td>
</tr>
<tr>
<td>Total</td>
<td>98.29</td>
<td>97.78</td>
<td>98.13</td>
<td>98.24</td>
<td>Si</td>
<td>1.99</td>
<td>1.97</td>
<td>1.93</td>
</tr>
<tr>
<td>Al</td>
<td>1.89</td>
<td>1.87</td>
<td>1.65</td>
<td>1.64</td>
<td>Al⁴⁺</td>
<td>0.01</td>
<td>0.03</td>
<td>0.07</td>
</tr>
<tr>
<td>Al⁶⁺</td>
<td>0.27</td>
<td>0.26</td>
<td>0.22</td>
<td>0.17</td>
<td>Al⁵⁺</td>
<td>0.02</td>
<td>0.01</td>
<td>0.08</td>
</tr>
<tr>
<td>Ti</td>
<td>0.06</td>
<td>0.05</td>
<td>0.07</td>
<td>0.08</td>
<td>Ti</td>
<td>0.00</td>
<td>0.00</td>
<td>0.01</td>
</tr>
<tr>
<td>Cr</td>
<td>0.00</td>
<td>0.00</td>
<td>0.03</td>
<td>0.01</td>
<td>Cr</td>
<td>0.03</td>
<td>0.01</td>
<td>0.00</td>
</tr>
<tr>
<td>Fe</td>
<td>0.82</td>
<td>0.76</td>
<td>0.80</td>
<td>0.78</td>
<td>Fe³⁺</td>
<td>0.15</td>
<td>0.19</td>
<td>0.03</td>
</tr>
<tr>
<td>Mn</td>
<td>0.03</td>
<td>0.02</td>
<td>0.02</td>
<td>0.02</td>
<td>Fe²⁺</td>
<td>0.11</td>
<td>0.09</td>
<td>0.14</td>
</tr>
<tr>
<td>Mg</td>
<td>4.70</td>
<td>4.76</td>
<td>4.65</td>
<td>4.73</td>
<td>Mn</td>
<td>0.01</td>
<td>0.01</td>
<td>0.02</td>
</tr>
<tr>
<td>Ca</td>
<td>0.02</td>
<td>0.00</td>
<td>0.01</td>
<td>0.00</td>
<td>Mg</td>
<td>0.67</td>
<td>0.69</td>
<td>0.73</td>
</tr>
<tr>
<td>Na</td>
<td>0.07</td>
<td>0.09</td>
<td>0.06</td>
<td>0.06</td>
<td>Ca</td>
<td>0.82</td>
<td>0.83</td>
<td>0.93</td>
</tr>
<tr>
<td>K</td>
<td>1.74</td>
<td>1.80</td>
<td>1.76</td>
<td>1.78</td>
<td>Na</td>
<td>0.19</td>
<td>0.18</td>
<td>0.06</td>
</tr>
<tr>
<td>O</td>
<td>20.11</td>
<td>20.10</td>
<td>20.13</td>
<td>20.12</td>
<td>K</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>OH</td>
<td>2.62</td>
<td>2.67</td>
<td>2.32</td>
<td>2.41</td>
<td>c) Calcite</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>F</td>
<td>1.27</td>
<td>1.23</td>
<td>1.55</td>
<td>1.46</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cl</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.01</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>5</th>
<th>10</th>
<th>d) Barite</th>
<th></th>
<th></th>
<th>e) Feldspar</th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Si (CO₃)₂</td>
<td>0.035</td>
<td>0.000</td>
<td>SO₂</td>
<td>19.45</td>
<td>19.57</td>
<td>SiO₂</td>
<td>61.37</td>
<td>61.87</td>
<td></td>
</tr>
<tr>
<td>Al₂ (CO₃)₂</td>
<td>0.000</td>
<td>0.060</td>
<td>BaO</td>
<td>71.81</td>
<td>70.19</td>
<td>Al₂O₃</td>
<td>19.73</td>
<td>19.76</td>
<td></td>
</tr>
<tr>
<td>Mg (CO₃)₂</td>
<td>0.004</td>
<td>0.000</td>
<td>CaO</td>
<td>0.04</td>
<td>0.06</td>
<td>CaO</td>
<td>0.06</td>
<td>0.05</td>
<td></td>
</tr>
<tr>
<td>Ca (CO₃)₁</td>
<td>98.894</td>
<td>98.169</td>
<td>SrO</td>
<td>0.01</td>
<td>0.10</td>
<td>FeO</td>
<td>0.00</td>
<td>0.08</td>
<td></td>
</tr>
<tr>
<td>Mn (CO₃)₁</td>
<td>0.000</td>
<td>0.033</td>
<td>total</td>
<td>91.31</td>
<td>89.92</td>
<td>SrO</td>
<td>0.60</td>
<td>0.47</td>
<td></td>
</tr>
<tr>
<td>Fe (CO₃)</td>
<td>0.000</td>
<td>0.005</td>
<td></td>
<td></td>
<td></td>
<td>BaO</td>
<td>4.44</td>
<td>4.74</td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>98.933</td>
<td>98.267</td>
<td></td>
<td></td>
<td></td>
<td>Na₂O</td>
<td>2.28</td>
<td>2.25</td>
<td></td>
</tr>
</tbody>
</table>

In both types of micas Cl contents is very low (0.0-0.04 wt%) and F is high (3.2-3.4 wt% in the FG mica; around 2.7 wt% in the CG one). High F contents have been recorded in phlogopite crystallized from ultrapotassic rock compositions (Edgar & Arima, 1985), which according to these authors, indicate that micas may be one of the reservoirs, besides amphibole and apatite, for halogens in the source regions for mantle-derived magmas. The role of F in the genesis of ultrapotassic magmas has been debated by different authors (e.g. Foley et al., 1987; Edgar & Arima, 1985). In mantle-derived xenoliths, phlogopite, apatite and amphibole, in this sequence, are likely the F reservoirs (Smith et al. 1981); being generally true for both lamproites and kamafugites.
In terms of Ti, Al, Mg and Fe contents, the studied micas are similar to the phlogopites from kimberlites and high-P xenoliths in kimberlites, as described by Bachinski & Simpson (1984) (Fig. 6), and from lamproites and ultramafic lamprophyres studied by Rock (1991) (Fig. 7).

**PYROXENE**

All clinopyroxenes analyzed so far in these xenoliths are Ca-rich, ranging in composition from diopside to salite in the quadrilateral of pyroxenes.

Some samples fall above the di-hd line, in the compositional field for pyroxenes from alkalic and ultramafic lamprophyres, described by Rock (1991) (Fig. 8). They are characterized by very low TiO$_2$ (0.05-0.37 wt%), with TiO$_2$ < Al$_2$O$_3$, a feature typical of diopsides from shoshonitic lamprophyres, according to Sheppard & Taylor (1992) (Fig. 9).

Two groups of pyroxenes are distinct in terms of Ca contents. In one group, CaO < 22 wt%, the ones CaO > 23 wt% (Fig. 10). Both groups have Al$_2$O$_3$ < 4 wt%, characteristic of clinopyroxenes in kimberlites, according to Haggerty (1994).

Grains show slight to major, complex, patchy zoning, characterized by no systematic elemental variation, as Na, mg#, Al (t) and Fe* may increase or decrease from core to rim, even in grains from same thin section, as illustrated in Figure 11. This behavior is similar to what is observed in pyroxenes from alkalic and ultramafic lamprophyres as studied by Rock (1991).

**K-FELDSPAR**

The proportion of KAlSi$_3$O$_8$ component in alkali feldspar in the xenoliths is high, with Or molecule ranging from 75.4 to 87.6, Ab from 12.2 to 24.3 and An from 0.05 to 0.33. Feldspar is characterized by low Fe (FeO < 0.4 wt%), high Sr contents (SrO from 0.03 to 0.83 wt%), and very high

*Fig. 6 — Micas of Triunfo mica-pyroxene xenoliths expressed in terms of TiO$_2$ and mg#. Unpattern field represents phlogopites from kimberlites and high-P xenoliths in kimberlites, as described by Bachinski & Simpson (1984).*

*Fig. 7 — Micas from Triunfo mica-pyroxene xenoliths plotted on a TiO$_2$-Al$_2$O$_3$ diagram. Fields for phlogopites from ultramafic lamprophyres and lamproites are from Rock (1991).*
concentrations of Ba, which varies from 1.9 to 4.9 wt% (average 4.1 wt%). This latter feature allows the classification of these feldspars as barium feldsparhyalophane – according to the criteria of Deer et al. (1966). Hyalophane has been reported in ultramafic lamprophyres (e.g. Rock, 1991). In terms of orthoclase and FeO contents, these feldspars are similar to feldspars from alkalic lamprophyres (Rock, 1991), and lamproites from Australia (Sheppard & Taylor, 1992) (Fig. 12).

![Diagram](image)

**Fig. 8** — Pyroxenes from Triunfo mica-pyroxenite xenoliths plotted on a Mg-Ca-Fe diagram. Fields for ultramafic (stippled) and alkalic (unpatterned) lamprophyres are from Rock (1991).

![Graph](image)

**Fig. 9** — Al₂O₃-TiO₂ compositions of pyroxenes from Triunfo mica-pyroxenite xenoliths. Fields shown are from Sheppard & Taylor (1992). Dashed line is TiO₂:Al₂O₃ 1:1 line.
Fig. 10 — CaO-Al2O3 compositions of clinopyroxene from Triunfo mica-pyroxene xenoliths (closed circle = core; open circle = rim; tie lines connect core and rim of a same grain). Field for pyroxenes in peridotites (stippled), and line of separation of pyroxenes in alkali basalts and in kimberlites are from Haggerty (1994).

CALCITE

In the two grains analyzed, carbonate mineral is almost pure calcite, with CaCO3 averaging 98.5 wt%, SiO3(CO3)2 ranging from 0.0 to 0.035 wt%, Al2(CO3)3 from 0.0 to 0.06 wt%, MnCO3 from 0.0 to 0.033 wt% and FeCO3 from 0.0 to 0.005 wt%. Pure calcite is reported to occur in kimberlites (Mitchell, 1986).

BARITE

Barite was detected by EDS in one sample. This mineral occurs associated to calcite, in contact with feldspar and pyroxene. Semiquantitative analyses for this mineral, stoichiometrically recalculated as SO2, indicate very high BaO contents (from 70.2 to 71.8 wt%), and trace amounts of CaO (0.04 to 0.23 wt%) and SrO (0.014 to 0.10 wt%).

CONCLUDING REMARKS

The intimate association of phlogopite-pyroxene xenoliths with ultrapotassic magmas in the Borborema province leads to the assumption that the xenoliths come from the magma source. The bulk chemistry of these xenoliths approaches that of pyroxene xenoliths in ultrapotassic lavas from Uganda, studied by Lloyd et al. (1987), considered to be derived from a metasomatized mantle source. This chemical similarity implies formation under similar conditions and also that their partial melting could produce liquids of similar composition.

Some chemical characteristics of the parental, mafic syenite of the Triunfo batholith, such as high K (K2O = 6.5 wt%), SiO2 saturation (SiO2 ≈ 56 wt%), and enrichment in incompatible elements, could be explained by partial melting of a phlogopite-bearing pyroxenite source.

Petrographic observations together with mineral chemistry of the xenoliths, suggest invasion of
fluid (CO2?) or melt (?), probably generating a locally veined lithosphere, or hybridization of the mantle source. This would explain the observed corrosion of clinopyroxene, followed by precipitation of calcite and K-feldspar. Therefore, we assume that the xenoliths have suffered a metasomatism in the mantle source.

Lithospheric mantle underneath the Borborema province in NE Brazil apparently received substantial continental crustal input (via subduction ?) during the Transamazonian cycle, as suggested by Nd model age for the host ultrapotassic syenite, around 2.3 Ga (Ferreira et al., 1994), and for all syenites along the syenitoid line (Ferreira, 1991; Van Schmus et al. in press). High initial Sr ratios (0.710) and very negative εNd values (ca. -16) for the syenite (Ferreira et al., op cit.), could be explained by this hypothesis.
Enriched mantle-source has been proposed for Brasiliano shoshonitic granitoids, K-rich diorites and syenites in the southernmost (e.g. Bitencourt & Nardi, 1993) and southeastern (e.g. Janasi et al., 1993) regions of Brazil, and even for Transamazonian syenites in eastern Brazil (e.g. Conceição, 1993). Therefore, such a feature by no means is restricted to the NeoProterozoic ultrapotassic rocks in the Borborema province. On the contrary, it seems to be a characteristic of the lithospheric mantle during Precambrian times, under a major part of Brazil.

ACKNOWLEDGMENTS

This research was made possible by grants of FACEPE, process n. APQ 3071.07/91 and FINEP/PADCT, process n. 65.930.619-00. This is the contribution n. 72 of the NEG (Laboratory Nucleus for Granite Studies), Dept. of Geology, UFPE.

REFERENCES


