Cation partitioning and substitution mechanisms in 1M phlogopite: A crystal chemical study

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ABSTRACT

The crystal chemistry of 1M phlogopite was studied by single-crystal X-ray diffraction and microprobe analysis. Structural refinement was carried out in space group C2/m (R values between 0.021 and 0.052) on 24 samples from basic lamproites, leucitites, kimberlite, and carbonatites.

The geometric and chemical features of the octahedral sites show the preferential partitioning of high-charge cations (Ti⁴⁺, Al³⁺, Cr³⁺, Fe³⁺) in the cis M2 sites. Bond-length and edge-length distortion parameters show the remarkably different behavior of the two octahedral sites when the high-charge cation content increases. The external shape of the trans M1 polyhedron is heavily constrained by the average field strength in the M2 sites.

In Ti-rich phlogopite the central cations of M2 sites show a significant off-center shift toward the O4 position. The substitution mechanism \[ ^{106}(R^{2+}) + 2(OH)^- = ^{106}(Ti^{4+}) + 2O_2^- \] best explains the observed structural modifications. The loss of a proton linked to O4 is mainly recognized from the shortening of the c lattice dimension.

The consequences of octahedral cation partitioning and structural adjustments related to the Ti substitution may be important when dealing with the thermal stability field of phlogopite and mica-based geothermometry or geobarometry and can help in understanding the dependence of Ti solubility on temperature, \( f_{O_2} \) and \( a_{H_2O} \).

INTRODUCTION

Triocahedral micas along the phlogopite-annite join occur as rock-forming minerals over a wide range of petrogenetic environments. Much systematic chemical study has been devoted to evaluating the most effective substitution mechanisms on both natural and synthetic micas. Some problems of the crystal chemical modeling of micas arise from the chemical substitution of high-charge cations for Mg²⁺ and Fe²⁺. As these cations are probably nonrandomly distributed over the sixfold-coordinated sites, this implies that a complete description of the population of each site is possible only by assuming a reliable cation ordering model. As discussed recently by Guidotti and Dyar (1991), some simplifications to the site substitution schemes used in formulating the solution and activity models may greatly affect conclusions from geothermometric and geobarometric studies.

Different cation ordering patterns for the end-member micas have been extensively investigated by means of single-crystal X-ray diffraction. The basic geometric features of the various arrangements and relative stabilities are discussed and reviewed in Toraya (1981), Bailey (1984), Guggenheim (1984), and Guggenheim and Eggleton (1987).

This work is a systematic investigation of 1M natural rock-forming Mg-rich micas (phlogopite sensu lato) by means of single-crystal X-ray diffraction and microprobe analysis. The aim is to estimate to what extent high-charge cations can control ordering mechanisms within the octahedral layer. The octahedral cation partitioning model based on the normal ordering scheme (Bailey, 1984; Guggenheim, 1984) was analyzed, not only to check the validity of the model, but also to obtain a complete description of the population of each octahedral site. This study was performed over a large, although not fully representative, selection of phlogopite forming in a variety of physical and chemical conditions; special interest was devoted to some samples of igneous Mg-rich micas from rocks believed to have formed in the upper mantle.

Ti may be considered representative of the high field strength elements, and its geochemical behavior in phlogopite has been extensively studied by experimental petrologists. The occurrence of three main Ti substitution mechanisms (Ti ⊥, Ti-Tschermak’s, and Ti-oxy) has been proposed and discussed by several authors. Bohlen et al. (1980), Dymek (1983), and Bol et al. (1989) suggested that the loss of proton in micas is a valid mechanism for maintaining the charge balance when bivalent cations are replaced by tri- or tetravalent cations, but the oxy-component is commonly neglected because of the analytical uncertainties in the Fe³⁺ and OH contents. As shown by Hewitt and Abrecht (1986) and Abrecht and Hewitt (1988), evidence for Ti-oxy substitution based only upon chemical analyses may be ambiguous, even when data on Fe³⁺ and OH contents are available. In this work we analyze the structural indications for the occurrence of Ti-oxy substitution in phlogopite.
TABLE 1. Description of mica samples

<table>
<thead>
<tr>
<th>Sample</th>
<th>Host rock</th>
<th>Locality</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Leucocratic dikes in garnet lherzolite</td>
<td>La Gallega (Ronda, Spain)</td>
<td>Gervilla and Leblanc (1990)</td>
</tr>
<tr>
<td>2</td>
<td>Nodule in carbonatite</td>
<td>Polino (Umbria, Italy)</td>
<td>Stopa and Lavecchia (1992)</td>
</tr>
<tr>
<td>3</td>
<td>Pegmatitic vein in olivine melilitite</td>
<td>S. Venanzo (Umbria, Italy)</td>
<td>Cundari and Ferguson (1991)</td>
</tr>
<tr>
<td>4, 5</td>
<td>Ejecta in ultrapotassic pyroclastic rock</td>
<td>Cabezale (Latiun, Italy)</td>
<td>Gianfagna and Tuzi (1988)</td>
</tr>
<tr>
<td>6</td>
<td>Lamproite</td>
<td>Cabezale (Latiun, Italy)</td>
<td>Fuster (1956)</td>
</tr>
<tr>
<td>7</td>
<td>Kaolinite</td>
<td>Torre Alfina (Latiun, Italy)</td>
<td>Cundari and Ferguson (1991)</td>
</tr>
<tr>
<td>8</td>
<td>Ultramafic xenolith in lamproite</td>
<td>Torre Alfina (Latiun, Italy)</td>
<td>Di Girolamo (personal communication)</td>
</tr>
<tr>
<td>9</td>
<td>Ejecta with mica + cpx + ol</td>
<td>Vescovio (Campania, Italy)</td>
<td>Dolfi (personal communication)</td>
</tr>
<tr>
<td>10</td>
<td>Leucitic tephrite</td>
<td>Cundari and Ferguson (1991)</td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>Phlogopitic clinopyroxenite nodule</td>
<td>Roccamonfina (Campania, Italy)</td>
<td>Di Girolamo (personal communication)</td>
</tr>
<tr>
<td>12</td>
<td>Pyroclastic rock with sandine</td>
<td>Acquasparta (Umbria, Italy)</td>
<td>Giannetti and Lufi (1990)</td>
</tr>
<tr>
<td>13</td>
<td>Lamproite</td>
<td>Torre Alfina (Latiun, Italy)</td>
<td>Stopa and Lavecchia (1992)</td>
</tr>
<tr>
<td>14</td>
<td>Metagabbro</td>
<td>Priestley Glacier (Antarctica)</td>
<td>Conticelli and Peccei (1990)</td>
</tr>
<tr>
<td>15</td>
<td>Carbonatite (sövite)</td>
<td>Eifel (Germany)</td>
<td>Lombardo et al. (1987)</td>
</tr>
<tr>
<td>16</td>
<td>Carbonatite (sövite)</td>
<td>Eifel (Germany)</td>
<td>Keller and Schleicher (1990)</td>
</tr>
<tr>
<td>17, 19</td>
<td>Kimberlite</td>
<td>Finch Pipe (South Africa)</td>
<td>Fraser and Hawkesworth (1992)</td>
</tr>
<tr>
<td>18</td>
<td>Lamproite</td>
<td>S.E. Yugoslavia</td>
<td>Hoy (1990), Hoy and Altherr (1991)</td>
</tr>
<tr>
<td>21-23</td>
<td>Lamproite</td>
<td>S. Venanzo (Umbria, Italy)</td>
<td>Cundari (1973)</td>
</tr>
<tr>
<td>24</td>
<td>Carbonatite (sövite)</td>
<td>Araxa (Minas Gerais, Brazil)</td>
<td>da Silva et al. (1979)</td>
</tr>
<tr>
<td>25</td>
<td>Carbonatite (sövite)</td>
<td>Araxa (Minas Gerais, Brazil)</td>
<td>da Silva et al. (1979)</td>
</tr>
<tr>
<td>26</td>
<td>Carbonatite</td>
<td>Fort Portal (Uganda, Africa)</td>
<td>Barker and Nixon (1989)</td>
</tr>
</tbody>
</table>

Note: samples suppliers (sample no.): F. Gervilla (University of Granada, Spain) (1); F. Stopa (University of Perugia, Italy) (2, 3, 7, 8, 12, 14, 16); A. Gianfagna (University of Rome, Italy) (4, 5); collected by G.C. (6); P. Di Girolamo (University of Naples, Italy) (9); D. Dolfi (University of Rome, Italy) (10); B. Giannetti (University of Rome, Italy) (11); B. Lombardo (University of Turin, Italy) and F. Talarico (University of Siena, Italy) (15); A. Cundari (University of Melbourne, Australia) (17, 19, 21, 22, 23); A. Holt and R. Altherr (University of Karlsruhe, Germany) (18); P. Comin-Chiaramonti (University of Palermo, Italy) (24, 25); D.S. Barker (University of Austin, Texas) (26).

EXPERIMENTAL

Sample description

The 24 specimens of mica used in this study are tabulated in Table 1. The geological and petrological setting is well characterized for most of the host rocks. Given references provide specific information on mineral assemblages and other petrological details.

X-ray structural study

Several mica crystals from each rock specimen were carefully hand-picked from crushed material. Only crystals with sharp optical extinction were considered suitable for the diffraction measurements.

Zero-level and first-level Weissenberg photographs were taken on crystals, mounted along the a-axis, to evaluate the crystal mosaic structure and periodicity of stacking along c*. The distribution of intensities along rows 13/ and 02/ was studied to determine polytypes according to Bailey (1988). All crystals analyzed on the diffractometer for data collection were 1M polytypes (space group C2/m), although 3T and 2M polytypes were also found in samples from lamproites.

Intensity data were collected on a Philips PW1100 four-
circle automated diffractometer using graphite-monochromatized MoKα radiation. Unit-cell parameters were refined by the least-squares method applied to fit the setting angles of 25 reflections in the range 8° ≤ θ ≤ 15°. For each crystal, the equivalent hkl and hkl pairs were measured in the θ range 2–30°, using the ω scan mode (speed scan of 1°/min, 3–5° width, and time of 20 s on backgrounds). Data were corrected for Lorentz and polarization factors. Absorption correction was applied according to the semiempirical method of North et al. (1968). Only reflections in which I > 3σ(I) [in a few cases I > 5σ(I)] were employed for the least-squares refinement of each sample.

Least-squares refinement was carried out in space group C2/m starting from the atomic coordinates of the end-member phlogopite (Hazen and Burnham, 1973). The full matrix program of the SHELX-76 crystallographic package was used for computation (Sheldrick, 1976). Neutral atomic scattering factors were used (Ibers and Hamilton, 1974). Fe and Mg scattering curves, with the constraint that X_Mg + X_Fe = 1, were used to fit the experimental electron density in the octahedral sites. Anisotropic displacement parameters and occupancy were refined in alternate cycles. In general, the composition of the tetrahedral site was fixed as 0.75 Si and 0.25 Al. Only for samples 17 and 19 (from kimberlite) and samples 24 and 25 (from carbonatite) was it possible to refine the occupancy factors of Si vs. Fe within the tetrahedra. Refinements converged to R values in the range 0.021–0.052 (mean 0.031). Unit-cell parameters and some details regarding data collection and refinement procedure are given in Table 2. Final atomic coordinates and temperature parameters are listed in Table 3; the atomic labeling is the same as in Hazen and Burnham (1973). Observed and calculated structure amplitudes are reported in Table 4; selected interatomic bond distances and site volumes are shown in Table 5.1 Table 6 contains some other structural parameters describing the geometry of the mica layer. The electron densities of the cation sites, as obtained from the refinement of the occupancy factors, are given in Table 7 and are compared with those calculated from microprobe analyses.

Microprobe analyses

Quantitative chemical analyses were acquired from the same crystals used for structural refinements (except for sample 3, where analyses were taken from six other crystals from the same rock). Samples were analyzed on a Cameca Camebax microprobe operating in WDS. Working conditions were those employed by Foley (1990) on lamproitic phlogopite using similar equipment. Four to six point analyses were taken for each crystal and averaged. The amount of Li was analyzed on an ion microprobe.

For formula calculations, several normalization schemes were tried and compared. The first scheme (11 atoms) is based on a fixed number [O, (OH, Fe)], and 22 negative charges; in the second scheme the complete occupancy of octahedral and tetrahedral sites (seven cations) is assumed; the third scheme followed the iterative method of Dymek (1983), which is based on the assumption that only the (R3+, R4+)-△ substitution occurs. It was found that each of these methods has severe limitations, depending on the specific assumption on which it is based. Both orders of preference Ti > Fe3+ and Fe3+ > Ti were assumed in filling the tetrahedral sites when the sum of Si + Al was <4 apfu, and the latter was chosen on the basis of the crystal chemical results. We finally adopted the following multistep procedure: (1) The normalization scheme using 11 O atoms was employed for the first calculation of the numbers of ions, and the method of Dy-

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1 A copy of Tables 3, 4, 5, and 9 may be ordered as Document AM-94-547 from the Business Office, Mineralogical Society of America, 1130 Seventeenth Street NW, Suite 330, Washington, DC 20036, U.S.A. Please remit $5.00 in advance for the microfiche.

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Fig. 1. (a) Plot of c lattice parameter vs. estimated OH content, expressed as OH groups per formula unit (g.f.u.). Line refers to the equation OH = 8.3(9)·c – 83.7(1), R2 = 0.88, by regression analysis on data from Brigatti and Davoli (1990) and Brigatti et al. (1991). (b) Plot of c parameter vs. Ti content (a.f.u. = atoms per formula unit). Solid symbols are for micas in this work; open symbols refer to data from Tables 2 and 3 of Abrecht and Hewitt (1988) on Mg-rich synthetic micas (boxes) and Fe-rich synthetic micas (boxes with crosses).
consistent and the accuracy of the calculations were evaluated by taking into account several well-defined relationships, involving structural parameters, observed in this study and from the literature. Best-fit calculations were achieved by trial and error. The complete description for the normalization procedure and discussion of results are reported in Cruciani (1993). Oxide percentages resulting from chemical analyses and numbers of ions content was obtained from the c-cell parameter (Fig. 1a) by the equation \( \text{OH} = 8.3(9) \cdot c - 83.7(1) \), \( R^2 = 0.88 \), which has been calculated by regression analysis from data from Brigatti and Davoli (1990) and Brigatti et al. (1991). Noda and Ushio (1965) found a similar equation for the \( \text{OH} + \text{F} \) substitution in synthetic phlogopite. This relationship is explained by decreased coulombic repulsion between the proton and the interlayer cation in OH-poor micas (McCausley et al., 1973; Giese, 1984; Munoz, 1984). (3) Formulas were calculated again on the basis of the anionic charge inferred by \( \text{OH} + \text{F} \). (4) The consistency and the accuracy of the calculations were evaluated by taking into account several well-defined relationships, involving structural parameters, observed in this study and from the literature. Best-fit calculations were achieved by trial and error. The complete description for the normalization procedure and discussion of results are reported in Cruciani (1993). Oxide percentages resulting from chemical analyses and numbers of ions calculated by the multistep procedure are given in Table 8.

**Table 6. Structural parameters for the 1M phlogopite samples**

<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
</tr>
</thead>
<tbody>
<tr>
<td>( a )</td>
<td>5.368(3)</td>
<td>5.625(2)</td>
<td>5.572(2)</td>
<td>5.561(2)</td>
<td>5.562(2)</td>
<td>5.561(2)</td>
<td>5.562(2)</td>
<td>5.561(2)</td>
<td>5.562(2)</td>
<td>5.561(2)</td>
<td>5.562(2)</td>
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<tr>
<td>( b )</td>
<td>9.354(1)</td>
<td>9.354(1)</td>
<td>9.354(1)</td>
<td>9.354(1)</td>
<td>9.354(1)</td>
<td>9.354(1)</td>
<td>9.354(1)</td>
<td>9.354(1)</td>
<td>9.354(1)</td>
<td>9.354(1)</td>
<td>9.354(1)</td>
</tr>
<tr>
<td>( c )</td>
<td>3.958(2)</td>
<td>3.958(2)</td>
<td>3.958(2)</td>
<td>3.958(2)</td>
<td>3.958(2)</td>
<td>3.958(2)</td>
<td>3.958(2)</td>
<td>3.958(2)</td>
<td>3.958(2)</td>
<td>3.958(2)</td>
<td>3.958(2)</td>
</tr>
</tbody>
</table>

**Note:** \( \text{OH} \) (A): tetrahedral sheet thickness calculated from z coordinates of basal and apical O atoms; \( \text{Ti} \) (A): octahedral sheet thickness as defined by Hazen and Burnham, 1973; \( \text{K} \) (A): projection of K-O distance along c; \( \psi_{\text{OH}} \) and \( \psi_{\text{F}} \): octahedral flattening angles (as in Hazen and Burnham, 1973); \( \gamma \): tetrahedral rotation angle (as in Hazen and Burnham, 1973); \( \delta_{\text{OH}} \): (K-O,F) - (K-O,Mg); \( \delta_{\text{Ti}} \): (K-O,F) - (K-O,Mg); D.M. (A): dimensional misfit between tetrahedral and octahedral sheets; \( m_{\text{OH}} \): (K-O,F) - (K-O,Mg); \( m_{\text{Ti}} \): (K-O,F) - (K-O,Mg); \( m_{\text{K}} \): (K-O,F) - (K-O,Mg); \( m_{\text{Fe}} \): (K-O,F) - (K-O,Mg); \( m_{\text{Mg}} \): (K-O,F) - (K-O,Mg); \( m_{\text{Al}} \): (K-O,F) - (K-O,Mg); \( m_{\text{Si}} \): (K-O,F) - (K-O,Mg). Oxygens percentages as calculated from the chemical analysis for the tetrahedral and interlayer cation sites.

**Table 7. Scattering power (electrons) in octahedral and interlayer cation sites**

<table>
<thead>
<tr>
<th></th>
<th>M1</th>
<th>M2</th>
<th>M1</th>
<th>M2</th>
<th>M1</th>
<th>M2</th>
<th>M1</th>
<th>M2</th>
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<tr>
<td></td>
<td>electrons</td>
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<tr>
<td>M1 electrons</td>
<td>15.6(1)</td>
<td>12.8(1)</td>
<td>13.6(1)</td>
<td>15.7(1)</td>
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<td>13.0(2)</td>
<td>14.4(1)</td>
<td>15.2(1)</td>
<td>14.7(1)</td>
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<tr>
<td>M2 electrons</td>
<td>15.9(1)</td>
<td>12.8(1)</td>
<td>13.7(1)</td>
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<td>12.9(1)</td>
<td>13.8(3)</td>
<td>14.5(1)</td>
<td>15.3(1)</td>
<td>14.8(1)</td>
<td></td>
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<tr>
<td>electrons</td>
<td>16.0(1)</td>
<td>12.9(1)</td>
<td>13.6(1)</td>
<td>15.7(1)</td>
<td>14.8(1)</td>
<td>13.1(2)</td>
<td>12.9(1)</td>
<td>13.8(3)</td>
<td>14.5(1)</td>
<td>15.3(1)</td>
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<td>EMPA</td>
<td>48.0</td>
<td>38.4</td>
<td>40.9</td>
<td>46.9</td>
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<td>36.1</td>
<td>40.6</td>
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<td>40.1</td>
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<td>17.4(1)</td>
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<td>18.2</td>
<td>19.2</td>
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<td>17.9</td>
<td>19.3</td>
<td>17.5</td>
<td></td>
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</tbody>
</table>

**Note:** M1 electrons, M2 electrons, and K electrons are from X-ray refinements in M1, M2, and interlayer sites, respectively; estimated standard deviations refer to the last digit; \( \text{M1 electrons} + 2 \cdot \text{M2 electrons} \); \( \text{EMPA} \) and \( \text{EMPA} \) are the electrons calculated from the chemical analysis for the tetrahedral and interlayer cations.
eled as a linear combination of the (Al-O) and (Si-O) bond distances on the basis of the chemical data (Cruciani, 1993). The enlargement of the tetrahedra of samples 17, 19, 24, and 25 implies the presence of a cation larger than Al. These samples have sums of Si + Al < 4 apfu, and reverse pleochroism, suggesting Fe³⁺. The occurrence of Fe³⁺ in sample 24 was confirmed by means of X-ray absorption spectroscopy (XANES) (Cruciani and Quartieri, 1992). A detailed description of single-crystal X-ray characterization of those phlogopite samples is given elsewhere (Cruciani and Zanazzi, 1992).

**Tetrahedral sheet and interlayer site geometry**

Tetrahedral rotation (α) (Newnham and Brindley, 1956; as in Hazen and Burnham, 1973) and out-of-plane tilting of basal O atoms (i.e., basal corrugation, Δz (Güven, 1971) are the most common parameters for describing interpolyhedral geometry in the tetrahedral sheet and the ditrigonality of the interlayer site. A close linear relationship exists between the tetrahedral rotation and the difference Δ_{K-O} = (K-O_{outer}) - (K-O_{inner}) (McCausley and Newnham, 1971). For the studied micas, linear regression yields Δ_{K-O} = 0.009(3) + α.Δ_{K-O} - 10^{-3}; N_{obs} = 24, R² = 0.997. As discussed by several authors (see Bailey, 1984, for a review) the in-plane rotation of adjacent tetrahedra is generally the most effective mechanism for the fitting of the lateral dimensions of the tetrahedral and octahedral sheets. This is confirmed by the positive trend between the tetrahedral rotation and the dimensional misfit parameters (Fig. 2). The dimensional misfit is a function of the chemical composition of octahedral and tetrahedral sheets. Mg-rich micas generally have higher α, Δ_{K-O}, and D.M. values (as defined in Table 2) than Fe-rich micas. The occurrence of some Mg-rich micas in a field of low α-Δ_{K-O} values may be explained by their high F content. In Figure 3, the increase in the estimated OH content is accompanied by a significant increase in the longer mean K-O bond distance ((K-O_{outer})), and by a very slight decrease in the shorter mean K-O bond distance ((K-O_{inner})), resulting in a greater difference Δ_{K-O}. The substitution (F, O) → OH on O4 may be associated with bond length variations of the interlayer site on the basis of the weak electrostatic interactions between the H and the O atoms of the interlayer site. Guggenheim et al. (1987) concluded that the proton in muscovite is able to weaken (lengthen) the longer K-O bonds with respect to the others. In phlogopite, the loss of H may be a complementary explanation, together with the decrease in the dimensional misfit, for the increased regularity of the interlayer cavity. The above consideration is also in agreement with the Δ_{K-O} values found for the oxybiotite (Ohta et al., 1982), which are smaller than those for the hydrogenated sample (Takeda and Ross, 1975). Samples featured by the lowest α, Δ_{K-O}, and D.M. values also have relatively high Ba content, which can be favored by or responsible for a decreased ditrigonality. Further discussion on this point is given in a later section on the basis of the bond valence theory.

The corrugation of basal O atoms, Δz, has positive and negative values, depending on the different ways the tetra-
has been experimentally found for several mica structures. This kind of cation ordering pattern agrees with octahedral tilt, which are in turn linked to the movement of the apical O. Enlargement of the M1 site with respect to the M2 site causes a shift of the apical O, which is consistent with $\Delta(z) > 0$, a common finding for most of our samples.

**Octahedral site geometry**

For all studied micas, the volumes and flattening angles of the trans M1 octahedral site are larger than those of the cis M2 site. These features can be referred to as the so-called normal ordering scheme, as defined by Bailey (1984) and Guggenheim (1984).

The normal ordering scheme has been described as an arrangement in which enlargement of the M1 site with respect to the M2 site and partitioning of high-charge, small-radius cations within the M2 site are favored by geometric and energy factors (Radoslovich, 1963; Toraya, 1981; Guggenheim, 1984; Guggenheim and Eggleton, 1987). Furthermore, the large size of the M1 site and the constraint of local charge balance agree with octahedral vacancies in M1. This kind of cation ordering pattern has been experimentally found for several mica structures (Lin and Guggenheim, 1983; Ohta et al., 1982, and references therein).

To characterize the distortion of M1 and M2, bondlength and edge-length distortion (BLD and ELD) parameters (Renner and Lehmann, 1986; Kunz et al., 1991) were calculated. These parameters describe both the geometry of the octahedron (ELD) and the arrangement of the cation within it (BLD), i.e., the external distortion of polyhedra and displacement of the central atom from the geometrical center of the polyhedron.

The ELD values of the M1 site are higher than those of the M2 site. This is consistent with the greater flattening angles and angle variance of the M1 site. The BLD values are generally lower for M1 than for M2. These observations are in good agreement with those of Renner and Lehmann (1986) and Kunz et al. (1991): the smaller polyhedron appears to have relatively more distorted bond lengths and fewer distorted edge lengths.

The magnitude of the BLD parameter of the M2 site shows a clear increase with increasing $^{60}$Ti content, whereas the BLD of the M1 site is quite constant (Fig. 4). The high bond-length distortion of the M2 sites is also recognized by comparison of the individual M2-O bond distances: M2-OH and M2-O3 bond lengths show the prevalent displacement of the M2 cation from the geometric center of the octahedra toward O4, along the two-fold axis. This off-center shift in M2 is correlated with Ti content (Fig. 5). The off-center shift of Ti atoms within octahedra is well known for a wide group of Ti-rich compounds (Kunz et al., 1991, and references therein); it is
noteworthy that this kind of distortion may occur even in cases of relatively low Ti content (e.g., micas). A similar feature has also recently been observed in amphiboles (Oberti et al., 1992).

The magnitude of ELD values for M1 shows a fairly good positive relationship with Ti content, although there is no relation between Ti and ELD values for M2 (Fig. 6).

Such different behavior of the internal and the external distortions between the M1 and M2 can be ascribed to the preference of Ti for M2. The correlation between the distortions from the M1 and M2 can be ascribed to the constraints within the mica octahedral sheet. When high-charge cations are partitioned into six smaller M2 octahedra surrounding one larger M1 site, the external distortion of the M1 polyhedron is expected to be strongly controlled. Lin and Guggenheim (1983) observed that the flattening of each octahedron is mainly dependent on the sum of the field strengths of the neighboring sites. Weiss et al. (1985) also deduced that the distortions in any particular octahedron are due to interaction in the whole sheet rather than in the octahedron alone. Consequently the increase in Ti on M2 may be responsible for external polyhedral distortions (ELD) of the isolated M1 polyhedral more than the effect exerted by the mean charge of the M1 site. This distortion behavior of the M1 and M2 sites was misinterpreted by Brigatti et al. (1991) as an indication of Ti preferential partitioning on M1.

Least-squares analysis

A least-squares program was constructed to merge the information obtained from the X-ray structural study and chemical analysis. The aim was to have a reliable quantitative description of the population of each octahedral site to evaluate the consistency of the observed X-ray diffraction and chemical data with the cation partitioning model. A system of eight equations relating the experimental data was developed according to the following assumptions: (1) all high-charge, small-radius cations (Ti$^{4+}$, Al$^{3+}$, Cr$^{3+}$, Fe$^{3+}$) are located on M2; (2) full occupancy is assumed for the M2 site; whereas octahedral vacancies are considered on M1; (3) modeling of the observed mean bond lengths, as commonly found in the literature, is carried out by a linear combination of the ideal cation to O bond distances of the type: $D_{n} = \sum x_i D_{ij}$, where $x_i$ is the fraction of the $n$th cation. The $D_{ij}$ ideal distances used here and found below were taken from Weiss et al. (1992); only the (Mg-O) value, closer to the distance found by Hazen and Burnham (1973) for the
end-member phlogopite, was chosen to optimize the agreement between observed and calculated mean bond distances for all samples studied.

The eight equations are as follows:

\[
26\text{Fe}^1 + 12\text{Mg}^1 = \text{e-M}1_{\text{obs}} - 25\text{Mn} - 3\text{Li} \quad (1)
\]

\[
26\text{Fe}^2 + 12\text{Mg}^2 = \text{e-M}2_{\text{obs}} - (13\text{Al} + 26\text{Fe}^3+ + 22\text{Ti} + 24\text{Cr})/2 \quad (2)
\]

\[
\text{Fe}^1 + 2\text{Fe}^2 = \text{Fe}_{\text{tot}} \quad (3)
\]

\[
\text{Mg}^1 + 2\text{Mg}^2 = \text{Mg}_{\text{tot}} \quad (4)
\]

Fe\(1 + \text{Mg}^1 = 1 - \text{Mn} - \text{Li} - \text{Mg}_0 \quad (5)
\]

Fe\(2 + \text{Mg}^2 = 1 - \text{Mg}_0 - \text{Fe}^3+ - \text{Ti} + \text{Cr}/2 \quad (6)
\]

\[
(\text{Fe}_{2+}^+\text{O})\text{Fe}^1 + (\text{Mg}\text{O})\text{Mg}^1
\]

\[
= (\text{M}1\text{O})_{\text{obs}} - (\text{Mn}\text{O})\text{Mn} - (\text{Li}\text{O})\text{Li} - (\Box\text{O}) \quad (7)
\]

\[
(\text{Fe}_{2+}^+\text{O})\text{Fe}^2 + (\text{Mg}\text{O})\text{Mg}^2
\]

\[
= (\text{M}2\text{O})_{\text{obs}} - (\text{Al}\text{O})\text{Al} - (\text{Fe}_{3+}\text{O})\text{Fe}^3+
\]

\[+ (\Box\text{O})\text{Ti} - (\text{Cr}\text{O})\text{Cr}/2 \quad (8)
\]

where (1) Mg\(_{\text{tot}}\), Fe\(_{\text{tot}}\), Ti\(_{\text{tot}}\), Al\(_{\text{tot}}\), Fe\(^{3+}\), Cr, Mn, Li are expressed as number of atoms per formula unit as calculated in Table 8; (2) \(\text{Mg}_0\) is calculated as (3 minus the sum of octahedral cations); (3) \(\text{Mg}\text{O}\): 2.067, \(\text{Fe}_{2+}\text{O}\): 2.053, \(\text{Al}\text{O}\): 1.920, \(\text{Ti}\text{O}\): 2.073, \(\text{Cr}\text{O}\): 2.040, \(\text{Fe}_{3+}\text{O}\): 2.11, \(\text{Mn}\text{O}\): 2.014, \(\text{Li}\text{O}\): 2.106, and \(\Box\text{O}\): 2.21 Å are the "ideal" values for \(\text{M}\) (or vacancy)—O average bond distances; (4) Fe\(^1\) and Fe\(^2\) are \(\text{Fe}_{2+}\) contents in M\(^1\) and M\(^2\); (5) Mg\(^1\) and Mg\(^2\) are Mg contents (atoms per site) in M\(^1\) and M\(^2\).

The results of the least-squares calculations are reported in Table 9. Calculated and observed values for bond distances and electron densities show the best agreement compared with several other systems of equations based on various ordering schemes that we tried.

The comparison of the calculated Fe and Mg contents in M\(^1\) and M\(^2\) reveals overall enrichment of Fe\(^{2+}\) within M\(^1\) and a random distribution of Mg over the two octahedral sites (Fig. 7). Ordering of Fe\(^{2+}\) on M\(^1\) may be considered as an indirect consequence of partitioning of the high-charge, small-radius cations on M\(^2\), but it may also be favored by the greater distortion of the M\(^1\) coordination polyhedron. In fact, the strong differences in the crystal-field stabilization energy and covalency degree of bonds are known as acting as possible ordering factors.
of Mg and Fe$^{2+}$ cations in ferromagnesian silicates. The general tendency of Fe$^{2+}$ to be ordered into the trans M1 site of trioctahedral micas is better recognized in Fe-rich micas (cf. Brigatti and Davoli, 1990) in which electron densities on M2 are generally lower than those on M1. In Mg- and Ti-rich micas, the electron density at M2 is generally higher than that at M1, despite the enrichment of Fe in M1; this feature can be ascribed to partitioning of Ti into M2.

**Bond valence and field strength calculations**

A full description of the cation site population allows the determination of bond valences and mean field strengths.

The bond valences of each bond ($s_j$) from the observed bond length distances ($R_i$) were calculated by the equation $s_j = \exp[R_{\text{min}} - R_i]/0.37$, by assuming as $R_{\text{min}}$ value the linear combination of the $R_0$ constants given by Brown and Altermatt (1985), weighted according to the composition of each site. Mixed occupancy (OH, F, O) was also approximately considered for bonds involving O4. The sums of bond valences for each site are reported in Table 10 and are in good agreement with the electrostatic balance requirements. In order to obtain the total charge at O4, the contribution of the H fraction should be added to the bond valence sum. The smallest value found for sample 3 agrees with its high F content, whereas the higher values for samples 21, 22, and 23 may be related to the shift of Ti toward O4 to compensate for the loss of the proton.

For Ti-rich samples, M2 bond valence sums are significantly lower than the average valence of the M2 cation obtained from its chemical composition. Still lower values are found if bond valence sums are calculated in the geometric center of the M2 polyhedron (see Table 10). In agreement with these observations, the off-center shift of Ti may be explained on the basis of the distortion theorem (Brown, 1992; Kunz and Armbruster, 1992), which derives from the concave shape of the correlation curve formed by the bond valence and bond lengths.

On the basis of this exponential curve, the increase in regularity of Ba-containing interlayer sites may also be explained. In fact, a net increase in positive charge on the interlayer central position, which is fixed by symmetry constraints, probably leads to a shortening of the K-O$_{\text{outer}}$ distances.

Mean field strengths (F.S.) (cf. Lin and Guggenheim, 1983) of the cation sites were calculated as the average of the valence to ionic radius ratios (from Shannon, 1976) weighted according to site population. Significant rela-
TABLE 10. Bond valence sums for refined phlogopite

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Note: values for each site are \( \sum s_i = \exp[(R_{i-} - R_t)/0.37] \) (Brown and Altermatt, 1985). \( \text{M2}_{\text{ave}} \): average valence of the M2 cation calculated from its chemical composition as \( \text{M2}_{\text{ave}} = \sum v_i/m_i \), where \( v_i \) = valence of \( i \)th cation and \( m_i \) = number of atoms per site of \( i \)th cation. \( \text{M2}_{\text{c}} \): bond valence sums calculated in the geometrical center of the M2 octahedron.

![Fig. 8](image-url)  
(a) Variation of average field strengths (F.S.) of M1 sites (open symbols) and M2 sites (solid symbols and labels) vs. edge-length distortion of the M1 site (M1-ELD); (b) variation of average field strengths of M1 sites (open symbols) and M2 sites (solid symbols and labels) vs. bond-length distortion of the M2 site (M2-BLD).

Tensions were found between the variation of (F.S.)\(_{\text{M1}}\) and (F.S.)\(_{\text{M2}}\) vs. the M1 and M2 distortion parameters. The external polyhedral distortion of M1 (M1-ELD) and the internal distortion of M2 (M2-BLD) increased strongly with (F.S.)\(_{\text{M2}}\) and decreased slightly with (F.S.)\(_{\text{M1}}\) (Fig. 8). These trends confirm the preference of the high-charge cations for M2.

**Substitution mechanisms in phlogopite**

This crystal chemical study may provide specific insights into some petrological aspects of micas. They mainly deal with the effects of Ti (as representative of high-charge, high-field elements) on the structure of mica, which is still poorly understood.

As shown, increased Ti content in phlogopite leads to the decrease of the \( c \) lattice parameter. Figure 9 shows that the shortening of \( d_{001} \) is tightly linked to the decrease in K-O\(_4\) distances. The modifications of thickness associated to the three main mechanisms involved in Ti substitution in micas may be roughly estimated on the basis of the size differences in cations. For the substitution \( 8^rR_2^++2^r4rAl_3^+ = 8^rTi_4^++2^r4rAl_3^+ \) (Robert, 1976) a relatively small thinning of the octahedral sheet is expected from the replacement of Mg and Fe by Ti, whereas the replacement of Si\(_4^+\) by Al\(_3^+\) or Fe\(_3^+\) should increase the tetrahedral thickness. In the substitution \( 2^r6r(R_2^+-) = 8^r(Ti_4^++ 8^r(Ti_4^++ 8^r(Ti_4^+)) \) (Forbes and Fowler, 1974), the decrease of octahedral thickness due to the smaller radius of Ti is partially or completely compensated by the presence of the large vacancy. Only the substitution \( 8^r(Ti_4^++ 2(OH)^- = 8^r(Ti_4^++ 2O^2- \) (Ti-oxo) implies shortening of the K-O\(_4\) distance because of the decreased interaction between the proton and the interlayer cation without effects on tetrahedral and octahedral sheet thicknesses. Moreover the off-center shift of Ti cations within M2 sites toward O\(_4\) is consistent with the loss of the proton bonded to O\(_4\). It follows that the Ti-oxo substitution mechanism best explains the observed structural adjustments. Evidence for the Ti-oxo substitution has also been recognized in amphiboles by means of single-crystal X-ray diffraction (Ob-
tent estimated by the difference \( \text{OH} : 2 - F \), suggesting the occupancies of the tetrahedral and octahedral sites. The observed tetrahedral bond lengths.

These samples a considerable Al deficiency, in contrast to charges (i.e., neglecting the oxy-component) yields for the existence of a relatively large oxy-component. For-...ship in Figure 1a is remarkably lower than the OH content. For instance, the OH- contents of Ti-rich phlogopite from samples 21, 22, 23) calculated by the relation-ship in Figure 1a is remarkably lower than the OH content estimated by the difference \( \text{OH} = 2 - F \), suggesting the existence of a relatively large oxy-component. Formula calculation based on the assumption of 22 negative charges (i.e., neglecting the oxy-component) yields for these samples a considerable Al deficiency, in contrast to the observed tetrahedral bond lengths.

**Implications for Phlogopite Stability**

The higher thermal stability of phlogopite when F and Ti increases is well documented (Forbes and Flower, 1974; Foley, 1990; Peterson et al., 1991; Thibault et al., 1992). Synthesis experiments also show that the entry of Ti into Mg-rich micas is favored by increased \( f_{\text{H}} \) (Arima and Edgar, 1981; Foley, 1989) and by decreased \( H_2O \) activity (Tronnes et al., 1985; Foley, 1989). The structural and charge-balance requirements described for Ti substitu-tion help us to understand these behaviors.

In general, as discussed from a geometric point of view by Hazen and Wones (1978) and Hazen and Finger (1982), in phlogopite with Si\textsubscript{4}Al tetrahedral sheets the allowed range of tetrahedral rotation (\( 0^\circ \leq \alpha \leq 12^\circ \)) imposes an upper limit to the octahedral sheet dimensions. The authors showed that micas containing large bivalent cations reach the geometrical limit \( \alpha = 0^\circ \) at temperature lower than micas with small-radius octahedral cations. In addition, a relatively small thermal expansion is expected for Ti\textsuperscript{4+}-containing octahedra, from its high ratio of valence to coordination number (Hazen and Prewitt, 1977).

Furthermore, Toraya (1981) pointed out that the thermal

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**Table 10.—Continued**

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**Fig. 9.** Plot of \( d_{001} \) vs. (a) K-O4 distance projected along c*, (b) tetrahedral sheet thickness, (c) octahedral sheet thickness. Note the same range of variation on the vertical axes.
decomposition of octahedral layers in trioctahedral mica takes place along the chains of M2 sites, in agreement with the greater expansion rate of M2, as found from the high-temperature refinement by Takeda and Morosin (1975).

According to the above considerations, even low contents of $\text{Ti}^{4+}$, partitioned into M2 sites, can increase remarkably the thermal stability of mica.

Several authors (Bol et al., 1989; Tracy, 1991, and references therein) described the occurrence of Ba and Ti-rich phlogopite in high-temperature metamorphic and igneous rocks. Our Ba- and Ti-rich phlogopite (samples 21, 22, 23) has small $\alpha$ (and $\delta_{K,1\alpha}$) values. The enhanced regularity of the interlayer cavity may be ascribed to the loss of the proton at O4 resulting from the Ti-oxy substitution. This structural adjustment and the decreased repulsion should favor the entry of Ba in 12-fold coordination and results in a more homogeneous distribution of positive charge over the neighboring O atoms. Since thermal decomposition of mica may result from the breaking of the weak bonds between the anions around the interlayer cation (Takeda and Morosin, 1975), Ba substitution may also contribute to thermal stabilization of the whole structure.

**IMPLICATIONS FOR MICA-BASED GEOTHERMOMETRIC AND GEOBAROMETRIC STUDIES**

Preferential partitioning of high-charge cations into M2 may have important consequences for the formulation of the biotite-garnet geothermometer. Multisite substitutions provide evidence of nonideality, which has to be considered in the mixing model for phlogopite-annite solid solutions (Guidotti and Dyar, 1991). In fact, even low contents of Ti may constrain both the polyhedral geometry of M2 and the distortion of M1, exerting a considerable degree of interference in the Fe-Mg intracrystalline ordering and the intercrystalline distribution between garnet and mica. Evidence of deviation from ideality arising from the occurrence of Ti and Al in the octahedral sites of biotite is well documented (Indares and Martignole, 1985; Dasgupta et al., 1991, and references therein).

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