MICAS FROM THE PIKES PEAK BATHOLITH AND ITS COGENETIC GRANITIC PEGMATITES, COLORADO: OPTICAL PROPERTIES, COMPOSITION, AND CORRELATION WITH PEGMATITE EVOLUTION

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ABSTRACT

Optical properties are presented for 66 samples of mica covering the range from annite → biotite → zinnwaldite → ferroan lepidolite and ferroan muscovite from occurrences of granitic pegmatite (NYF type) throughout the Pikes Peak batholith (PPB) in Colorado. Chemical composition was determined for 34 of these samples. The optical data are correlated with composition, mode of occurrence, and relation to pegmatite paragenesis. Optical properties of the trioctahedral micas show a consistent trend of decreasing β index of refraction, from an average of 1.693 in annite of the host granite to 1.577 in zinnwaldite and ferroan lepidolite of the miarolitic cavities, which correlates with a progressively decreasing content of Fe. A comparison of optical and compositional data for micas from localities throughout the PPB indicates a variation in geochemical evolution among pegmatites of different districts, and between the Pikes Peak Granite and its late satellite plutons. Analyses of mica samples taken from cross-sections through individual pegmatites reveal a decrease in index of refraction and total iron that unambiguously document a progressive geochemical evolution within a given pegmatite. Such data, in addition to field evidence, indicate that micas enclosed within massive quartz arc paragenetically older than those within miarolitic cavities; minerals within miarolitic cavities represent the final stages of primary crystallization. A general model of pegmatite paragenesis is proposed that hypothesizes formation of miarolitic cavities as a consequence of pegmatite configuration and inclination, as well as early crystallization of massive quartz that confines the silicate melt and volatile phase, resulting in closed-system crystallization with a concomitant increase in pressure, consequent episodic cavity-rupture events, and corresponding changes in mica composition.

Keywords: annite, biotite, iron content, paragenesis, pegmatite, index of refraction, zinnwaldite, Pikes Peak batholith, Colorado.

Nous présentons les propriétés optiques de 66 échantillons de mica représentant l'intervalle annite → biotite → zinnwaldite → lépidolite et muscovite ferrifères, prélevés de pegmatites granitiques de type NYF dans le batholithe de Pikes Peak, au Colorado. Nous présentons aussi des données sur la composition de 34 de ceux-ci. Les données optiques dépendent de la composition, du type de gisement, et de la relation avec les associations paragénétiques. Les échantillons de mica trioctaédriques montrent une diminution progressive de l'indice de réfraction β, d'une moyenne de 1.693 dans l'annite du granite encaissant à 1.577 dans la zinnwaldite et la lépidolite ferrifère des cavités miarolitiques, diminution qui est en corrélation avec la diminution de la teneur en Fe. Une comparaison des données optiques et compositionnelles indique une variation du degré d'évolution géochimique parmi les pegmatites des différents districts, et entre le batholithe de Pikes Peak et les plutons tardifs satellites. Pour le cas d'échantillons prélevés le long de coupes à travers des massifs individuels de pegmatite, nous documentons une diminution de l'indice β et de la teneur en Fe total qui illustre de façon non ambiguë la direction de l'évolution. De telles données, évaluées en considérant les relations de terrain, montrent que le mica piégé dans le noyau de quartz massif est antérieur au mica des cavités miarolitiques. Les minéraux de ces cavités représenteraient le stade final de cristallisation primaire. Nous décrivons ces tendances au moyen d'un modèle général de cristallisation. La formation de cavités dépend de l'attitude et de l'inclinaison de la pegmatite, et la cristallisation précède de quartz massif, qui pêche le magma silicaté et la phase volatile. Il en résulte une cristallisation en système fermé, avec comme conséquences une augmentation de la pression, des ruptures épisodiques des cavités, et des changements correspondants dans la composition du mica.

Mots-clés: annite, biotite, teneur en Fe, paragenèse, pegmatite, indice de réfraction, zinnwaldite, batholithe de Pikes Peak, Colorado.

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INTRODUCTION

The Pikes Peak batholith (PPB) in central Colorado is noted for its granitic pegmatites containing small- to medium-sized miarolitic cavities (as much as 2.5 m across) that yield fine specimens of amazonitic to medium-sized miarolitic cavities (as much as 2.5 m across) that yield fine specimens of amazonitic to medium-sized miarolitic cavities. These occur in such well-known pegmatite-bearing areas (Fig. 1) as Devils Head, the Crystal Peak and Lake George areas that encompass the Lake George ring (LGR) complex, the Tarryall area (Spruce Grove Campground and Matucat Road) that encompasses the Redskin stock, the Wigwam Creek area (trailhead area and Sugarloaf Peak), the Sentinel Rock, Crystal Park, Cameron Cone, and Stove Mountain areas west of Colorado Springs, and the Glen Cove area on the northwest side of Pikes Peak.

Within the host granite and the graphic-textured pegmatite, crystals of the early-formed micas are unzoned, but the mica that crystallized later, within miarolitic cavities, may show distinct color zoning that becomes readily visible when the “books” of mica are separated along the (001) cleavage (Figs. 2–5). Some specimens show complex zoning (e.g., Fig. 5), whereas other specimens show simple zoning, with only a few “color bands” (Figs. 3, 4), or a distinctive epitactic overgrowth (Fig. 6), in which a polycrystalline core of ferroan zinnwaldite is surrounded by a rim of monocristalline zinnwaldite. Mica from the Wigwam Creek area is more likely to show distinct and complex color-zonation than is that from other pegmatite districts. Flakes of mica that appear to be relatively homogeneous in plane-polarized light may display a well-defined oscillatory zonation under crossed-polarized light (e.g., Fig. 7), a result of slight rotation of the optic plane relative to the b axis. Such zones are presumably a manifestation of subtle compositional variation and concomitant structural changes. Most flakes of color-zoned mica do not show an appreciable corresponding “optical” zonation under crossed-polarized light.

There are almost no published data on optical properties of micas from the PPB, and chemical data on a limited set of micas from granitic pegmatites in the PPB have only recently been reported (Foord et al. 1995). These authors documented a progression in composition, with decreasing Fe and Ti and increasing Li and F, from annite in the host granite, through siderophyllite, to lithian biotite, and finally to ferroan zinnwaldite, zinnwaldite and ferroan lepidolite in miarolitic cavities.

Because the composition of micas has been shown by Foord et al. (1995) to be related to geochemical evolution, they are excellent indicators of the differentiation and development of miarolitic and other types of pegmatites. A correlation of index of refraction to chemical composition (especially iron content) in micas permits the use of optical properties as an efficient means of assessing such parageneses. An advantage of optical methods is that exceedingly small quantities of sample can be examined, and a visual evaluation of alteration is possible. Consequently, a comparative evaluation of the changes in optical (and compositional) properties across color zones within single crystals of mica and in micas from pegmatite cross-sections, as well as from localities throughout the PPB, allows an assessment of paragenesis of individual occurrences of pegmatite. The present study is based on an extensive suite of micas that have been well characterized as to mode of occurrence, and that represent a thorough sampling of the various occurrences of pegmatite in the Pikes Peak batholith; studies of micas from cross-sections of pegmatite bodies have not previously been reported. These samples document a close relation between the optical and chemical properties of micas and their mode of occurrence within a given pegmatite.

GEological SETTING AND MODE OF OCCURRENCE

The PPB is a composite, epizonal, primarily granitic (greater than 90%) to quartz monzonitic body that is exposed over an area of about 5000 km² in the Front Range of central Colorado. It is an anorogenic pluton of Precambrian age, approximately 1092 ± 2 to 1074 ± 3 Ma, on the basis of a recent study of U–Pb systemsatics (zircon) by Unruh et al. (1995), and represents the last of three major intrusive episodes in the Front Range of Colorado. The plutons were emplaced at successively shallower depths (Simmons et al. 1987, Wobus & Hutchinson 1988), and mineralogical studies by Barker et al. (1975), Foord & Martin (1979), and Blasi et al. (1984) support an epizonal emplacement of the PPB. Hawley & Wobus (1977), Wobus & Anderson (1978), Wobus (1986), and Hutchinson (1976, 1988) provided further detail on the Pikes Peak Granite.

Numerous smaller and younger intrusions of both sodic and potassic affinity (Wobus & Anderson 1978, Wobus & Hutchinson 1988) are clustered along two parallel NW–SE trends within the batholith; the most prominent of these intrusions are illustrated in Figure 1. These late plutons are shown in detail on the Denver geological map by Bryant et al. (1981). The sodic plutons consist of gabbro, syenite, fayalite syenite, fayalite granite, and mildly peralkaline riebeckite granite. Late potassic plutons are metaluminous to peraluminous, and compositionally similar to the Pikes Peak Granite, but are generally finer grained. The largest of these potassic plutons, the Redskin stock, is enriched in Li, Be, and Sn, and hosts topaz-bearing pegmatites as well as Be-rich greisen deposits, such as at the Boomer mine (Hawley 1969, Hawley & Wobus 1977, Desborough et al. 1980). Pegmatites throughout the batholith are concentrated in fractures both in and around the late-stage plutons and large intrusive centers. Pegmatites associated with
both sodic and potassic groups are generally of the NYF (Nb-Y-F) affiliation (Cerný 1991), although the youngest pegmatites show slight enrichment in Li and Rb, i.e., zinnwaldite and ferroan lepidolite are present in miarolitic cavities. The pegmatites are shallow to subvolcanic; those in the southern part of the batholith host miarolitic cavities that provide the specimens of amazonitic microcline and smoky quartz for which the region is famous. Recent studies by Unruh et al. (1995) have elucidated the age of emplacement of late intrusive units and pegmatites of the PPB. These authors found, on the basis of $^{40}$Ar/$^{39}$Ar plateau dates on riebeckite, biotite, lithian biotite, and miarolitic-cavity zinnwaldite, that the shallow-level pegmatites in the western and southern portions of the batholith were emplaced between 1062 ± 2 and 1059 ± 2 Ma, and that deeper-level bodies of pegmatite in the north (South Platte district) were emplaced between 1077 ± 2 and 1066 ± 2 Ma. In each area, U–Pb ages for the late potassic plutons are similar to those of the host

Fig. 1. Geological index-map of the Pikes Peak batholith (shaded areas), showing locations of major late-stage potassic plutons and other geographic features. LGR: Lake George ring, RS: Redskin stock, DH: Devils Head, LRP: Lone Rock pluton. Map taken from Foord et al. (1995).
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Fig. 2. Cleavage fragment of zinnwaldite (sample no. 4), 2.4 cm across, showing complex zonation, from a miarolitic cavity, Sugarloaf Peak, Wigwam Creek area. Transition from core to rim shows a decreasing $\beta$ index of refraction, with the decrease being attributed to a lower Fe content. The composition of the rim corresponds to ferroan lepidolite.

Fig. 3. Cleavage fragment of zinnwaldite (sample no. 6), 2.0 cm across, showing distinct zonation, from the Lake George area, northern Lake George ring. Rim shows a substantial Ti content (0.36 wt. % TiO$_2$) that accounts for its dark brown color.

Fig. 4. Cleavage fragment of zinnwaldite (sample no. 3), 2.6 cm across, showing distinct zonation, from a miarolitic cavity, Sugarloaf Peak, Wigwam Creek area.

Fig. 5. Cleavage fragment of zinnwaldite (sample no. 52), 4.4 cm across, showing complex zonation, from a miarolitic cavity, Wigwam Creek trailhead area.

Fig. 6. Crystal of zinnwaldite (sample no. 54), 7 x 5.8 cm, showing distinct zoning, from a miarolitic cavity, Harris Park area (collected 1990). Transition from a polycrystalline core to a monocristalline rim shows decreasing $\beta$ index of refraction and increasing optic angle. The lighter color of the core (compared to the rim) is attributed to light scattering by a non-uniform polycrystalline texture, in spite of its higher Ti (0.30 wt. % TiO$_2$) relative to that in the rim (0.23 wt. % TiO$_2$).

Pikes Peak Granite. The ages of emplacement of the sodic plutons thus far analyzed by Unruh et al. (1995) are all $\leq 1085 \pm 4$ Ma.

A typical miarolitic pegmatite in the PPB is characterized by an outer zone of graphic pegmatite that shows increasingly large and elongate crystals toward the center, and a clay-filled cavity that commonly is centrally located. The “pocket stage” of pegmatite development represents the final result of extended fractional crystallization of the felsic magma (Foord & Martin 1979). Zinnwaldite crystals that formed in miarolitic cavities may be as much as 10 cm across and
show a characteristic pseudohexagonal morphology, in some cases with conspicuous zonation evident in cleaved sections. Examination of mineral intergrowths shows that zinnwaldite crystallized over an extended period in the miarolitic cavity, being contemporary with early-formed feldspars but preceding the formation of lower-temperature minerals such as fluorite and goethite. Mica crystals from miarolitic cavities in and near the Lake George ring complex are much less common than elsewhere in the PPB. This may be due either to movement of the Becke line in ambiguous situations is problematic because of unfavorable geometry of the grains, pleochroism, alteration, and inclusions. Observation of the Becke line for determination of the indices of refraction proved to be problematic, as the indices of the darker and more pleochroic samples, whereas Table 4 gives empirical formulas of the micas. Identical areas from zoned micas were used for both optical and compositional analyses. Nine samples represent mica collected from different units within a single pegmatite and adjacent host granite that was excavated in 1986, whereas 19 others represent mica collected from cross-sections through six other pegmatites or host granites (or both). Table 5 summarizes optical properties and compositions of the micas representing cross-sections through individual pegmatites.

Methods and Instrumentation

Sixty-six mica crystals representing localities throughout the mineral-specimen-producing areas of the PPB were studied. Physical descriptions and general localities for these samples are given in Table 1. Of these samples, 38 were collected from miarolitic cavities, 10 from quartz-core occurrences, eight from graphic pegmatite, and seven from host granites. Three samples are representative of "remnant mica" that formed during early stages of miarolitic cavity crystallization and that were subsequently partly resorbed. Complete optical data were obtained for all 66 samples (including core and rim zones of ten crystals). Measured optical properties are tabulated in Table 2 for pegmatites throughout the PPB according to mode of occurrence, i.e., host granite, graphic pegmatite, quartz core, remnant mica, and miarolitic cavity. Thirty-four samples were quantitatively analyzed for major constituents; there was insufficient sample to permit analyses for Li or FeO/Fe₂O₃. Table 3 shows results of quantitative analyses for selected samples, whereas Table 4 gives empirical formulas of the micas. Identical areas from zoned micas were used for both optical and compositional analyses. Nine samples represent mica collected from different units within a single pegmatite and adjacent host granite that was excavated in 1986, whereas 19 others represent mica collected from cross-sections through six other pegmatites or host granites (or both). Table 5 summarizes optical properties and compositions of the micas representing cross-sections through individual pegmatites.

Optical data were determined using a conventional petrographic microscope and calibrated immersion oils. Two of the indices, β and γ, were determined by grain mount on an approximately centered Bxa figure; α was determined using spindle stage methods to attain proper orientation of the X vibration direction. Observation of the Becke line for determination of the indices of refraction in white light proved to be problematic because of unfavorable geometry of the grains, pleochroism, alteration, and inclusions. Movement of the Becke line in ambiguous situations was confirmed with a 589 nm interference filter. Samples showing considerable alteration (e.g., numbers 44, 63, and 68) were carefully examined to locate transparent grains showing little or no alteration for use in determination of the indices of refraction. Such unaltered portions of mica flakes yielded consistent results, in agreement with Peikert (1963), who noted that a moderate degree of chloritization did not affect the indices. Determination of the indices of refraction in the case of lighter-colored micas (zinnwaldite and some biotite) was relatively straightforward, and the error is given as ±0.002, whereas determination of the indices of the darker and more pleochroic samples,
TABLE 1. DESCRIPTION OF SELECTED MICA SAMPLES FROM THE PIKES PEAK BATHOLITH

<table>
<thead>
<tr>
<th>Sample</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>Pale brown zinnwaldite from a mafic pluton, northeastern LGR.</td>
</tr>
<tr>
<td>2.</td>
<td>Zoned mica (Fig. 2) with gray zinnwaldite and coarse pale brown ferro-micaspidite from a mafic pluton, South Peak, Wiggins Creek area. Associated with medium-blue mica (amazonite).</td>
</tr>
<tr>
<td>3.</td>
<td>Zoned ferro-micaspidite with gray core and brown rim (Fig. 3); from a mafic pluton, northwestern LGR.</td>
</tr>
<tr>
<td>4.</td>
<td>Zoned, pale brown zinnwaldite from a mafic pluton near Devils Head. Associated with topaz, smoky quartz, and anthophyllite.</td>
</tr>
<tr>
<td>5.</td>
<td>Dark brown-zoned, hexagonal biotite from quartz core (1979), northwest of the LGR. Associated with samples 99 and 90.</td>
</tr>
<tr>
<td>6.</td>
<td>Dark brown zoned biotite distributed in massive quartz (1986), from a pegmatite in coarse-grained Pliss Peak Granite (Ype), LGR; with samples 133, 72-74, 77, 87 and 88.</td>
</tr>
<tr>
<td>7.</td>
<td>Lath, dark-brown to black, hexagonal biotite enclosed in massive quartz, northern LGR (1978); sample adjacent to no. 44. Associated with medium-to-dark blue mica (amazonite).</td>
</tr>
<tr>
<td>8.</td>
<td>Pseudohexagonal, brown zinnwaldite, 1.4 cm across, from a mafic pluton, northwestern LGR; associated with microcline (variety amazonite) and smoky quartz.</td>
</tr>
<tr>
<td>9.</td>
<td>Biotite showing alternation, with bronze color and submetallic luster. From graphic pegmatite (1978), northeastern LGR; sample adjacent to no. 86.</td>
</tr>
<tr>
<td>10.</td>
<td>Light, brown-zoned biotite from quartz core, north of the LGR. Associated with microcline.</td>
</tr>
<tr>
<td>11.</td>
<td>Pale-brown zinnwaldite showing indistinct zoning, from a mafic pluton near the Wiggins Creek trailhead. Associated with pale-blue mica (amazonite), smoky quartz, and albite (sample 24-1, Ford et al., 1995).</td>
</tr>
<tr>
<td>12.</td>
<td>Pale-to-medium brown, distinctly zoned zinnwaldite, 4.4 cm across (Fig. 5); from area adjacent to the Wiggins Creek trailhead.</td>
</tr>
<tr>
<td>13.</td>
<td>Vivid bluish-green litthian biotite from quartz core near the Wiggins Creek trailhead. Associated with quartz, microcline, and albite.</td>
</tr>
<tr>
<td>14.</td>
<td>Distinctly zoned biotite within a zinnwaldite that has a silvery-submetallic luster, and a dark-brown monoclinic pyroxene rim (Fig. 1B); from an associated tetraxyloite collected near Harris Park in 1992; associated with pale-to-medium blue mica (amazonite).</td>
</tr>
<tr>
<td>15.</td>
<td>Zoned zinnwaldite, with a light-brown core and brown rim; from a mafic pluton near Devil's Peak, Wiggins Creek area (samples 1a and 1b, Ford et al., 1995).</td>
</tr>
<tr>
<td>16.</td>
<td>Hexagonal, lath dark-brown to black zinnwaldite from a mafic pluton (1986), associated with medium-blue mica (amazonite), smoky quartz, and albite. From a pegmatite in coarse-grained (Ype) Pliss Peak Granite, LGR; associated with samples 133, 72-74, 77, 87 and 88.</td>
</tr>
<tr>
<td>17.</td>
<td>Pale brown-grain mica zinnwaldite, 2.6 cm across, from the Devils Head area.</td>
</tr>
<tr>
<td>18.</td>
<td>Partly dark-brown to black-biotite, from graphic pegmatite north of LGR; sample adjacent to no. 64.</td>
</tr>
<tr>
<td>19.</td>
<td>Lath, black, annedral mica from coarse-grained Pliss Peak Granite (Ype), northeastern LGR.</td>
</tr>
<tr>
<td>20.</td>
<td>Irregular, brown biotite showing partial dissolution of early microcline; from a mafic pluton, near Devil's Peak, Wiggins Creek area. Associated with pale-to-medium blue mica (amazonite), microcline overgrowth, quartz, and albite.</td>
</tr>
<tr>
<td>21.</td>
<td>Annedral zinnwaldite from an irregular cavity (remaining after partial dissolution of early microcline) along the base of a mafic pluton, platt rock, Wiggins Creek area. From the 1986 pegmatite in coarse-grained Pliss Peak Granite (Ype), LGR; Associated with samples 33, 56, 73, 74, 77, 87 and 88.</td>
</tr>
<tr>
<td>22.</td>
<td>Remnant ferro-zinnwaldite from an edgewise decay (remaining after partial dissolution of early microcline) showing part of a hexagonal outline in massive quartz and microcline (variety amazonite) from the wall of a mafic pluton, Wiggins Creek area. From the 1986 pegmatite in coarse-grained Pliss Peak Granite (Ype), LGR; Associated with samples 33, 56, 73, 74, 77, 87 and 88.</td>
</tr>
<tr>
<td>23.</td>
<td>Intergrown crystal of brown zinnwaldite from mafic pluton of 1986 pegmatite in Castle Creek Granite (Ype), LGR. Associated with samples 33, 56, 72, 74, 77, 87 and 88.</td>
</tr>
<tr>
<td>24.</td>
<td>Euhedral, dark-brown to black zinnwaldite showing distinct core and rim zonation; perthite replaced by smoky quartz crystals. From the 1986 pegmatite, coarse-grained Pliss Peak Granite (Ype), LGR; Associated with samples 33, 56, 72-74, 77 and 88.</td>
</tr>
<tr>
<td>25.</td>
<td>Euhedral, colorless to gray mica, 1 mm across, on microcline. From a mafic pluton near the Spruce Grove Campground, Tarryall area.</td>
</tr>
<tr>
<td>26.</td>
<td>Annedral, lath-shaped, dark brown biotite from fine-grained pegmatite, near Devil's Peak, Wiggins Creek area. From the 1986 pegmatite, coarse-grained Pliss Peak Granite (Ype), LGR; Associated with samples 33, 56, 72-74, 77 and 88.</td>
</tr>
<tr>
<td>27.</td>
<td>Annedral, colorless to gray mica, 1 mm across, on microcline. From a mafic pluton near the Spruce Grove Campground, Tarryall area.</td>
</tr>
<tr>
<td>28.</td>
<td>Lath, dark-brown to black zinnwaldite from coarse-grained Pliss Peak Granite (Ype), near Devil's Peak, Wiggins Creek area. From the 1986 pegmatite. Associated with samples 33, 56, 72-74, 77 and 88.</td>
</tr>
</tbody>
</table>

Notes: Dates indicate the year collected, and correspond to dates given in Table 2. LGR: Lake George ring.
First, the text describes the universal stage measurements. The structure, for example, was determined from a stereonet plot derived from large bodies of pegmatite where annotated Bith a data; these nica sequences did different locations within the specified part of the pegmatite.

Then, the text provides specific values, such as:

- For the optic axial angle, Mallard's method was used.
- Details on the mica samples taken from the same geographic locality.
- Values for the number and locality, including:
  - 3.2 t.94 t.617 t.6t7 0.93
  - 21.0 l.611 l.695 l.695 0.984
  - 15.1 l.612 l.705 l.705 0.993
  - And so on...

The text also mentions specific locations and geographic features, such as:

- LGR, northwestern: Yp
- LGR, southeastern: Yp,
- and others.

The conclusion discusses the implications of these findings, noting that the average optic axial angle was determined in Mallard's method. Optical properties and composition of mica, Pikes Peak Batholith.

Finally, it is noted that for dioctahedral – triocahedral of selected samples was determined by optic plane orientation relative to [010] using universal stage methods. Polytype was also ascertained for three of these samples (114, 129, and 147) by X-ray diffraction (XRD) using a Siemens Kristalloflex 805 operating at 40 kV, 25 mA, or a Philips XRD 3000 generator operating at 40 kV, 25 mA. (Use of trade names is for identification purposes only and does not constitute endorsement by the U.S. Geological Survey). A thorough discussion of optical...
analytical methods applied to mica is given by Wilcox (1984), who also summarized optical data for the different species of mica.

Electron-microprobe analyses were done at the U.S. Geological Survey in Denver using a JEOL 8900 Superprobe with five automated WDS spectrometers. Analyses were done using an accelerating voltage of 15 kV and a sample current of 20 nA. The following elements were sought: Na, Mg, Al, Si, Ca, Ti, Fe, Mn and K, using K lines in each case. Natural and synthetic mineral standards were used: albite (Na, Si), forsterite (Mg), anorthite (Al, Ca), rutile (Ti), fayalite (Fe), fluor-phlogopite (F), spessartine (Mn), and orthoclase (K). The concentration of Na, F, Al, Mg and Si was established using TAP crystals, that of K and Ca, using PET, and that of Fe, Mn and Ti, using LIF. For all elements except F, background counts were accumulated for 10 seconds and peak counts for 80 seconds; for F, background was accumulated for 40 seconds and peak for 80 seconds. Matrix corrections were performed using JEOL ZAF procedures. All but two samples were analyzed at 4–5 points each; samples 63 and 112 were analyzed at 2 and 6 points, respectively. Li content (not measured on account of small size of the samples) was calculated using a regression equation given by Tindle & Webb (1990) that is based on an empirical relation between Li2O and SiO2.

**TABLE 3. COMPOSITIONAL DATA FOR SELECTED MICAS FROM THE PIKES PEAK BATHOLITH**

<table>
<thead>
<tr>
<th>number</th>
<th>locality</th>
<th>SiO2</th>
<th>Al2O3</th>
<th>TiO2</th>
<th>FeO</th>
<th>MnO</th>
<th>MgO</th>
<th>Li2O</th>
<th>CaO</th>
<th>Na2O</th>
<th>K2O</th>
<th>F</th>
<th>O=O-F</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>63</td>
<td>LGR, NE</td>
<td>34.6</td>
<td>12.9</td>
<td>4.00</td>
<td>32.8</td>
<td>0.53</td>
<td>2.29</td>
<td>0.38</td>
<td>0.00</td>
<td>0.06</td>
<td>0.0</td>
<td>9.0</td>
<td>0.36</td>
<td>97.1</td>
</tr>
<tr>
<td>68</td>
<td>LGR, SW</td>
<td>34.3</td>
<td>9.8</td>
<td>2.37</td>
<td>34.9</td>
<td>0.81</td>
<td>0.14</td>
<td>0.87</td>
<td>0.00</td>
<td>0.06</td>
<td>0.0</td>
<td>3.5</td>
<td>1.46</td>
<td>96.3</td>
</tr>
<tr>
<td>88</td>
<td>LGR, 1986</td>
<td>33.4</td>
<td>12.6</td>
<td>3.96</td>
<td>33.3</td>
<td>0.56</td>
<td>1.73</td>
<td>0.63</td>
<td>0.11</td>
<td>0.05</td>
<td>0.8</td>
<td>9.6</td>
<td>0.27</td>
<td>95.0</td>
</tr>
<tr>
<td>117</td>
<td>RGS, Tarryall</td>
<td>34.5</td>
<td>17.2</td>
<td>2.43</td>
<td>28.5</td>
<td>1.14</td>
<td>0.77</td>
<td>0.18</td>
<td>0.00</td>
<td>0.15</td>
<td>9.1</td>
<td>2.7</td>
<td>1.14</td>
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<td>12.2</td>
<td>3.84</td>
<td>34.2</td>
<td>0.63</td>
<td>1.21</td>
<td>0.38</td>
<td>0.00</td>
<td>0.04</td>
<td>0.0</td>
<td>9.0</td>
<td>0.27</td>
<td>96.5</td>
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**II. Graphic Pyrometamorphic biotite**

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<tr>
<th>number</th>
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<th>SiO2</th>
<th>Al2O3</th>
<th>TiO2</th>
<th>FeO</th>
<th>MnO</th>
<th>MgO</th>
<th>Li2O</th>
<th>CaO</th>
<th>Na2O</th>
<th>K2O</th>
<th>F</th>
<th>O=O-F</th>
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<tr>
<td>117</td>
<td>RGS, Tarryall</td>
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<td></td>
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<td>Wigum Cr., SP</td>
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</tbody>
</table>

**IV. Rhexitic, Early Micaofitic Cavities**

**V. Micaofitic Cavities**

**VI. Micaofitic Cavities**

**Notes:** Li2O calculated from Li2O = (0.287 x SiO2) - 9.552 (Tindle & Webb, 1990).

Abbreviations: LG = Lake George ring; RGS = Redskin Granite; SP = Superlith Peak; TH = trailhead.

RELATION OF OPTICAL DATA TO COMPOSITION

In the earliest comprehensive study of the relation between indices of refraction of a mica and its composition, Hall (1941) found nR to be related to the concentration of both Fe and Ti (with 1% TiO2 causing an increase in nR of 0.0046), and also to the oxidation state of Fe (with Fe3+ giving a higher index than Fe2+).
Heinrich (1946) subsequently found a good correlation between \( n_t \) and the concentration of Fe\(^{3+} \) and Ti. Rieder et al. (1971) later showed a correlation between indices of refraction and Fe in the sheets of octahedra in Li–Fe-rich micas, whereas Gottesmann & Tischendorf (1978) found the indices in trioctahedral micas to be proportional to the Fe\(^{3+} \) content, attaining maximum values with a simultaneously high Fe\(^{3+} \) content.

The suite of micas from the PPB in the present study shows an excellent linear relation between the total Fe (expressed as wt.% FeO) and \( n_t \). Figure 8 gives a plot of total FeO versus \( n_t \) for selected samples from this study that have been quantitatively analyzed. The regression equation for these data is \( y = 0.004725x + 1.519 \) (\( r^2 = 0.94 \)), where \( y \) is \( n_t \) and \( x \) is total Fe. A less precise correlation is noted between \( n_t \) and Ti (\( r^2 = 0.83 \)), as well as between total Fe and 2\( V_t \), where the Fe content is generally inversely proportional to 2\( V_t \) (\( r^2 = 0.35 \)). The relatively poor correlation between 2\( V_t \) and composition is presumably due to interlayering of twinned crystals (Bloss 1965, Axelrod & Grimaldi 1949), or to effects of polytypism (Rieder et al. 1971), either of which can lower the axial angle.

Zinnwaldite and ferroan lepidolite studied by Foord et al. (1995) are 1\( M \) polytypes, whereas lithian biotite was found either to be a 3\( T \) or 2\( M \) polytype; the single sample of muscovite consists of the 2\( M \)\(_1 \) polytype. Sample 114 from this study (miarolitic-cavity zinnwaldite from Sentinel Rock) was determined by XRD to be a 1\( M \) polytype, whereas samples 129 and 147 (muscovite from Spruce Grove Campground and Glen Cove, respectively) were confirmed to be dioctahedral and to consist of the 2\( M \)\(_1 \) polytype.
The sample is a high-Fe zinnwaldite to low-Fe biotite with a Ti-rich rim that shows higher indices of refraction for both core and rim than is typical for most cavity zinnwaldite. This is presumably due to its higher Fe and Ti content relative to other samples of zinnwaldite. The rim of this sample shows higher indices relative to the core, despite a lower Fe content in the rim; this is likely attributable to the substantially higher Ti content in the rim. This sample also shows a much lower 2V, than expected, likely on account of its being a 3T polytype (determined by M.F. Brigatti, pers. commun., 1997).

Micas from the host Pikes Peak Granite have been defined as annite by Barker et al. (1975), on the basis of an average Fe content of 32.6% FeO. The high indices of refraction and Fe content for annite from the coarse-grained Pikes Peak Granite (Ypc) studied here (average 39 = 1.701 and 33.7% FeO) fall within this range. The high index of refraction, compared to an upper limit of 1.690 given by Wilcox (1984), may be attributable to a compositional transition toward ferri-annite, which has a maximum 39 of 1.720 (Wilcox 1984). Optical data for miarolitic-cavity samples of zinnwaldite (average 39 = 1.579) fall well within published values, which range from 1.570 to 1.590 (Gottesmann & Tischendorf 1978, Némec 1983, Wilcox 1984). Optical properties for muscovite samples, with an average 39 of 1.595, are also within established values given by Wilcox (1984).

Figure 9 illustrates the relation between composition and 39 for selected constituents. This figure clearly shows the geochemical evolution of the micas, with decreasing FeO and TiO2, and concomitantly increasing SiO2, Al2O3, and F. K2O (data not shown) shows a consistent value throughout the range of 39.

**Relation of Optical Data to Regional Geochemical Evolution**

The relation between 39 and FeO (Fig. 8) and the correlation of Fe in mica with geochemical evolution in the PPB shown by Foord et al. (1995) permit the use.

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**Table 5. Index of Refraction \( \beta \) and Total Fe Content for Selected Occurrences of Mica, Pikes Peak Batholith**

<table>
<thead>
<tr>
<th>Sample Type</th>
<th>( n_\beta )</th>
<th>FeO</th>
<th>Sample Type</th>
<th>( n_\beta )</th>
<th>FeO</th>
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<tbody>
<tr>
<td>Host granite</td>
<td>1.705</td>
<td>31.3</td>
<td>Miharolitic cavity, core</td>
<td>1.585</td>
<td>15.7</td>
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<tr>
<td>Pegmatite</td>
<td>1.571</td>
<td>31.0</td>
<td>Miharolitic cavity, rim</td>
<td>1.577</td>
<td>15.5</td>
</tr>
<tr>
<td>Remnant mica</td>
<td>1.667</td>
<td>nd</td>
<td>Host granite</td>
<td>1.697</td>
<td>34.2</td>
</tr>
<tr>
<td>Crystal core</td>
<td>1.659</td>
<td>15.5</td>
<td>Miharolitic cavity, core</td>
<td>1.583</td>
<td>14.5</td>
</tr>
<tr>
<td>North of LGR</td>
<td>1.641</td>
<td>24.0</td>
<td>Miharolitic cavity, rim</td>
<td>1.583</td>
<td>14.5</td>
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<tr>
<td>Quartz core</td>
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<td>27.1</td>
<td>Pegmatite</td>
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<td>30.9</td>
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<tr>
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<td>1.659</td>
<td>nd</td>
<td>Quartz core</td>
<td>1.653</td>
<td>32.5</td>
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</table>

The occurrences are listed in order of increasing \( n_\beta \).
of optical data for an assessment of paragenesis. On a regional level, optical and compositional data of the micas illustrate a comparative geochemical evolution of granitic rocks within the PPB. For example, the lower $n_\beta$ and FeO for a mica from the medium-grained granite composing a part of the Lake George ring complex (Ypm, sample 68), relative to those from the coarse-grained Pikes Peak Granite (Ypc), correlate with the later emplacement of this potassic pluton. Similarly, a mica from the Redskin Granite (sample 117), a highly evolved intrusive body (Hawley & Wobus 1977, Desborough et al. 1980), shows a considerably lower $n_\beta$ (1.663) and FeO than do micas from the coarse-grained Pikes Peak Granite. The high degree of fractionation of the Redskin stock and associated pegmatites is further evidenced by the presence of muscovite in all three samples of miarolitic-cavity mica thus far studied.

Substantial differences in geochemical evolution are shown among pegmatites from different localities within the batholith. This is illustrated by comparatively low $n_\beta$ and corresponding FeO values for quartz-core and miarolitic-cavity micas from the Sugarloaf Peak locality relative to similarly occurring micas from the 1986 locality in the Lake George ring complex. This finding suggests that intermediate stages of crystallization in the Sugarloaf Peak pegmatite are more highly evolved.

Optical and compositional data also correlate with mode of occurrence throughout the PPB, illustrating a progressive geochemical evolution from host granite → graphic pegmatite → quartz core → miarolitic cavity, with a general trend of decreasing indices of refraction and decreasing FeO (Fig. 8). The highest indices are found in mica from the host granite (average $n_\beta = 1.693$), and the lowest are represented by zinnwaldite to ferroan lepidolite (average $n_\beta = 1.579$) contained within miarolitic cavities. Although index of refraction (and FeO) is a more reliable indicator of mode of occurrence than is $2V_\alpha$, the optic angle for micas from different modes of occurrence throughout the PPB is reasonably consistent, showing for the most part a low $2V_\alpha$ expected for biotite and annite in granite, graphic pegmatite, and quartz-core occurrences, and a higher optic angle for the zinnwaldite – ferroan lepidolite from miarolitic cavities. Average values of optic angle for the micas shown in Table 2 range from 13.4° for host granite, 7.7° for graphic pegmatite, 9.5° for quartz core, and 28.4° for miarolitic-cavity zinnwaldite (calculated from rim samples where applicable, excluding sample 6), to 36.6° for miarolitic-cavity ferroan muscovite. Determinations of optic angle on 37 additional samples of mica (unpubl. data) corroborate these results.

Mica of intermediate composition, typical of quartz-core occurrences, has generally lower indices of refraction (average $n_\beta = 1.631$) compared to mica from surrounding pegmatites (average $n_\beta = 1.654$). However, the high $n_\beta$ and corresponding FeO content for quartz-core micas compared to those in miarolitic cavities suggest a paragenetically earlier crystallization within the quartz core than in the miarolitic cavities. This sequence agrees with field observation, e.g., late-formed minerals such as fluorite are not found enclosed
in massive quartz, but early-formed species (such as zircon and ferrocolumbite) are. The final period of microcline crystallization within the quartz core was roughly contemporary with that of crystallization of the massive quartz, as shown by varying degrees of irregular surfaces (a result of growth interference) on microcline.

The greater range of $n_b$ (based on two standard deviations) for samples of quartz-core mica throughout the PPB (1.631 ± 0.029) relative to samples of miarolitic-cavity mica (1.579 ± 0.009) attests to a greater variability in degree of fractionation at the quartz-core stage of pegmatite development. Alternatively, the comparatively consistent $n_b$ for miarolitic-cavity micas throughout the PPB suggests a relatively uniform "endpoint" of geochemical evolution for the trioctahedral micas.

**Paragenesis Within Individual Pegmatites**

Although these optical and compositional data allow a comparison of geochemical differences for a given mode of occurrence among pegmatites from different localities, an unambiguous assessment of the paragenetic sequence within individual pegmatites can only be determined with data from proximate occurrences across a single pegmatite (i.e., a cross-section through host granite, graphic pegmatite, quartz core, and miarolitic cavity). Figure 10 illustrates the relation between optical properties and paragenesis for four granitic pegmatites within the PPB, whereas Table 5 summarizes optical and compositional data for these and three other pegmatite cross-sections. Index of refraction, rather than % FeO, is shown in this figure because it was possible to obtain a complete set of data throughout the pegmatite cross-sections. Although this figure shows overlapping ranges of index of refraction ($n_b$) in mica samples from the quartz core and graphic pegmatite zones from different localities, there is nevertheless a continuous evolutionary trend within a given pegmatite toward a lower $n_b$ index. Figure 10 further illustrates the early paragenesis of mica in the quartz core relative to the miarolitic-cavity stage of crystallization.

The consistent decrease in index of refraction within a series of mica samples from a pegmatite is not always paralleled by a decrease in FeO (Table 5), but the discrepancies fall within limits of analytical error. Such discrepancies may also be attributable to electron-microprobe data acquired on slightly altered or limonite-coated mica grains, whereas polarized light microscopy, by virtue of a better visual estimate of such features, seems to give more consistent results.

A complete evolutionary trend, from host granite to miarolitic cavity, is illustrated by a series of mica samples from the Sugarloaf Peak and the 1986, Lake George ring-complex localities. Specimens from both occurrences show a progressive decrease in $n_b$ (and corresponding decrease in FeO) throughout the sequence of crystallization. The Sugarloaf Peak (Wigwam Creek) occurrence is an extensive, zoned, nearly horizontal pegmatite that exceeds 100 m in length; it is more than 1 m thick in places, with numerous quartz lenses and small miarolitic cavities occurring throughout the structure. The high degree of evolution of micas in quartz-core and miarolitic-cavity occurrences in this pegmatite (shown by their low values of $n_b$), relative to other localities such as the 1986 pegmatite (discussed in detail below), is clearly illustrated in Figure 10. It is important to note that although the above data show consistent evolutionary trends within individual pegmatites throughout the PPB, they also illustrate fundamental differences in evolution among different pegmatites as well as pegmatite districts.

Remnants of early-formed, partly resorbed micas line the walls of small irregular cavities or depressions that remain following partial to nearly complete dissolution of the original crystal. These small voids, typically situated near the base of microcline crystals or in the graphic pegmatite wallrock adjacent to the miarolitic cavity, retain their original, usually irregular pseudohexagonal form, typically with a characteristic lining of quartz microcrystals. The "remnant" flakes retain their original orientation in the nearly hollow cavity, i.e., with (001) being perpendicular to the long axis of the cavity. Optical properties of such mica (samples 72, 73, and 153) substantiate their early formation in the miarolitic cavity, and suggest a paragenesis that is equivalent to the early polycrystalline core of mica crystals from other pegmatites, e.g., samples 54 and 77.

The last stages of fractionation within individual miarolitic cavities are shown by zoned crystals of zinnwaldite (samples 4, Fig 2; 54, Fig. 6; 77 and 106) that show a decrease in $n_b$ and FeO from core to rim. Electron-microprobe traverses across three zoned crystals of miarolitic-cavity zinnwaldite show a fluctuating Ti content that closely correlates with color zonation (darker zones have higher Ti), whereas total iron shows a consistent decrease from core to rim (unpubl. data). Later stages of sequential differentiation are particularly well represented by the specimen shown in Figure 11, from the Sugarloaf Peak locality. This specimen (sample 136) exhibits a continuous sequence of mica crystallization with a corresponding decrease in $n_b$ and FeO, from a polycrystalline base ($n_b = 1.586, \text{FeO} = 13.8\%$) partly enclosed by graphic pegmatite, to an epitactic open-cavity zinnwaldite that shows zonation from a brown interior ($n_b = 1.573, \text{FeO} = 9.8\%$) to a pale-brown rim ($n_b = 1.569, \text{FeO} = 9.7\%$).

**Paragenesis of the 1986 Pegmatite**

A detailed example for the geochemical evolution of mica within a single body of pegmatite is provided by an occurrence in the coarse-grained Pikes Peak Granite (Ypc) near the Lake George ring complex, in
Fig. 10. Sequential differentiation of micas through cross-sections within individual pegmatites in the Pikes Peak batholith. The figure shows optical progression in the sequence host granite → graphic pegmatite → quartz core → miarolitic cavity. Sample numbers given on plot; LGR: Lake George ring complex.

Fig. 11. Matrix specimen from Sugarloaf Peak, Wigwam Creek area, illustrating early-generation polycrystalline mica (showing evidence of partial dissolution) that is partly enclosed by the wallrock of the miarolitic cavity, and later-generation, open-cavity epitactic growth of monocrystalline, zoned zinnwaldite (sample no. 136, Table 2), associated with microcline, variety amazonite. Specimen is 11 cm across.
which the presence of mica in all modes of occurrence within the pegmatite, and comprehensive field data, allow a complete assessment of its paragenesis. A diagram of this pegmatite is shown in Figure 12; it illustrates a miarolitic cavity, adjacent bodies of massive quartz, and surrounding granite; mica sample locations also are indicated. Figures 13 and 14 illustrate in situ cross-sectional views of this occurrence. This pegmatite, excavated in 1986, represents a zonal structure and mineralogy that are characteristic of the Crystal Peak area.

Paragenesis of minerals adjacent to and within the miarolitic cavity is shown in Figure 15. Amazonitic microcline, albite, and biotite are enclosed by massive quartz at the extremities of the elongate miarolitic cavity. The massive quartz becomes increasingly fragmented and more loosely consolidated toward the pocket margin, with large partial prism faces of milky to gray quartz facing the adjacent cavity. Crystallization within the central miarolitic cavity consisted of early amazonitic microcline and platy albite, followed by extensive crystallization of euhedral quartz that formed mostly on the ceiling of the pocket. Later-formed minerals within the cavity include (in paragenetic order) fluorite, hematite, and calcite.

The mica samples collected from this pegmatite and surrounding host granite (summarized in Table 5) show a progressive decrease in $n_g$ and FeO, from annite in the host granite, to biotite in the pegmatite and quartz core, to ferroan zinnwaldite and zinnwaldite in the miarolitic cavity. The decrease in index of refraction and FeO suggests that the formation of the massive quartz and its enclosed mica (and feldspar minerals) at the distal ends of this pegmatite preceded major crystallization in the centrally located miarolitic cavity. Formation of the euhedral quartz crystals in the miarolitic cavity is not likely to have been either earlier than or contemporaneous with the formation of the massive quartz because: (1) massive quartz is not seen enclosing late-formed fluorite crystals, (2) mica crystals enclosed in massive quartz have higher indices of refraction (and a corresponding higher Fe content) than those within the miarolitic cavity, and (3) euhedral quartz crystals (i.e., smoky quartz) do not enclose biotite, but commonly partly enclose crystals of zinnwaldite and amazonitic microcline.

Remnant ferroan zinnwaldite, showing an optic angle and indices of refraction intermediate to those of biotite and zinnwaldite, was noted as minute flakes (sample 72) on the walls of an irregular cavity alongside the base of a microcline crystal, and as fragments (sample 73) in an elongate depression (showing part of the original pseudo-hexagonal outline) in massive quartz and microcline adjacent to the miarolitic cavity. Continuing fractionation within the miarolitic cavity is shown by a distinct zonation between the core and rim of sample 77, evidenced by decreasing $n_g$ and increasing $2V_r$. This zonation is possibly a result of a pocket-rupture event (shown in Fig. 15); broken quartz crystals that show varying degrees of regrowth as well as the presence of fine-grained crystals also are probable manifestations of one or more pocket-rupture events.

A Proposed Model of Evolution for Pegmatites of the Pikes Peak Batholith

Studies of pegmatite genesis [Jahns & Burnham (1969), Jahns (1979, 1982), London (1986, 1990, 1992)] have established two general models for the development of miarolitic-cavity-bearing pegmatites. These are largely based on studies of LCT-type pegmatites in the Peninsular Ranges batholith of southern California. Few studies, however, have focused on miarolitic-cavity-bearing pegmatites of the PPB (e.g., Wobus et al. 1988). The optical and compositional data presented here, as well as field evidence, provide new details concerning the evolution of pegmatite bodies and miarolitic cavities in the PPB, and are in agreement with the major tenets of the general models proposed in earlier research. Much of this proposed model is based on the early crystallization of massive quartz, which has specific implications regarding pegmatite genesis.

This model of evolution postulates that graphic pegmatites formed within fracture zones of the granite that were caused by tectonic features related to the emplacement of late plutons within the parent granite. Attitude and configuration of the fractures, and of the resultant bodies of pegmatite, likely played a role in the development of late-stage mineralization. Irregularities in inclination may have facilitated entrapment of a residual fluid phase that exsolved from the silicate melt as it crystallized during cooling of the parent granite. Segregation and coalescence of vapor within the pegmatite-forming melt resulted in the appearance of miarolitic cavities. In the PPB, miarolitic cavities are most abundant in pegmatites with dips less than about 30°, but rare in vertical or near-vertical pegmatites, from which the volatiles presumably more readily escaped. It is possible, however, that some miarolitic cavities noted in steeply inclined pegmatites formed as a result of escaping volatiles that were trapped either at an interface, such as a diorite porphyry, or at a granodiorite xenolith caprock (Wobus et al. 1988). Alternatively, it is possible that early crystallization of massive quartz within the miarolitic cavities formed a constriction that trapped the volatile phase, which may account for cavities seen in many pegmatites in the PPB that show no conspicuous inflection in dip. This is in contrast to the pegmatite dikes of San Diego County, California, where pegmatite configuration appears to be a controlling factor for the location of miarolitic cavities.

Whereas crystallization of massive quartz adjacent to miarolitic cavities preserved the enclosed biotite, crystals of ferroan zinnwaldite that subsequently
Fig. 12. Diagram of a pegmatite with enclosed miarolitic cavity and adjacent massive quartz, excavated near the Lake George ring complex in 1986. Ypc: coarse-grained Pikes Peak Granite, peg: pegmatite showing graphic structure near cavity border, Q: massive quartz, MC: miarolitic cavity. The position of mica samples collected from within this pegmatite is indicated by sample numbers in the diagram. Boundaries dashed where inferred. A–A': line of cross-section, B–B': level of exposure for plan view.
formed early within the remaining central miarolitic cavities and adjacent wallrock were resorbed to varying extents, leaving pseudohexagonal cavities that may contain remnant mica. The instability of the early-cavity mica may result from increased pressure within the miarolitic cavity and concomitant changes in oxygen fugacity, partial pressures of H₂O, and activities of chemical constituents in the vapor phase, all of which have been shown to affect the composition and field of stability of micas in silicate melts (Eugster 1957, Eugster & Wones 1958, 1962, Peikert 1963, Wones & Eugster 1965, Rieder 1971, Syritso et al. 1996). This increase in pressure likely resulted from exsolution of an aqueous vapor phase due to crystallization of the silicate melt (Burnham 1979a, b, 1983). Internal pressures exceeding lithostatic pressure could have resulted in pocket rupture and a subsequent reduction in pressure that would cause abrupt changes
in fluid composition (e.g., Fe and Ti) and account for some of the simple zonation seen in mica. However, the presence of numerous such zones within a small linear span (e.g., Figs. 5, 7), which suggests rapid compositional changes within a relatively short time, is likely due to a chemical diffusion process initiated by crystallization.

The model proposed here requires that a silicate melt phase persists throughout most of the crystallization from aqueous fluid within the miarolitic cavity to account for continuing generation of pressure. Coexistence of a silicate melt with an aqueous vapor phase at temperatures compatible with formation of α-quartz may be possible under metastable, disequilibrium conditions (London et al. 1989, London 1992, 1996), or in the presence of F [found at an average concentration of 0.47% in the Pikes Peak Granite by Hawley & Wobus (1977)]. Fluorine is known to reduce the solidus of granitic systems (Manning 1981, London et al. 1989, London 1992).

**SUMMARY AND CONCLUSIONS**

The composition of mica provides an excellent indicator of geochemical differentiation within the PPB. Because the index of refraction of the micas shows a close linear relation to FeO (which is correlated to the degree of fractionation of the magma), determination of $n_0$ can provide an unambiguous assessment of paragenesis on both regional and local scales.

A continuum of fractionation, based on indices of refraction and composition of the micas, is shown throughout the PPB, from the least-evolved pegmatites in the Lake George ring complex, to the ferroan lepidolite-bearing pegmatite in the Sugarloaf Peak (Wigwam Creek) area, and ultimately to the muscovite-bearing pegmatites at Glen Cove and within the Redskin Stock. The transition of mica composition, from quartz-enclosed biotite crystals at the distal ends of the miarolitic cavity, to resorbed crystals of ferroan zinnwaldite at the margin of the cavity, and finally to euhedral, late-formed zinnwaldite to ferroan lepidolite within the miarolitic cavity, substantiates a sequence of crystallization from the quartz core inward toward the adjacent miarolitic cavity. Crystallization within miarolitic cavities represents the final stage of fractionation, and reaches a uniform compositional “endpoint”, as evidenced by the relatively consistent value of $n_0$ and FeO content of the trioctahedral mica in the miarolitic cavities. Crystallization within the cavities ceased when the silicate melt was exhausted and the vapor phase was depleted of all dissolved constituents.
The general sequence of development of miarolitic cavities proposed here is partly based on cavity-rupture events due to the hydrostatic pressure exerted by the aqueous phase that results from closed-system crystallization from the silicate melt phase, with subsequent resealing and continuing crystallization. This model accounts for the simple zonation of pocket mica and the shock damage and regrowth of other crystals. It is in agreement with the models advanced by Jahns (1979, 1982), largely based on studies of pegmatite bodies in San Diego County, California, and by London (1990, 1992, 1996). However, additional details are added that are specific to the PPB pegmatites, i.e., the early crystallization of massive quartz that may account (in part) for the formation of miarolitic cavities and results in closed-system conditions. It should be noted that some aspects of paragenesis, as expressed in the above model, may not apply to all pegmatite occurrences in the PPB, but general elements of this model should be pertinent to miarolitic-cavity-bearing pegmatites throughout the PPB. Details of the paragenetic sequence will vary depending on local conditions of formation, such as attitude and size of the pegmatite, and local variations in composition of the silicate melt and aqueous fluids. Although this model of pegmatite paragenesis contradicts aspects of an earlier report, in which the quartz core mode was postulated to be the last-formed phase within a pegmatite (Wobus et al. 1988), field evidence, and optical and chemical data reported here appear to best support the proposed interpretation.

Finally, it appears that polarized light microscopy provides an economical and reliable method for assessing geochemical evolution in igneous bodies, particularly if one must rely on partly altered samples.

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