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COMPOSITION OF PLUTONIC MUSCOVITE: GENETIC IMPLICATIONS

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ABSTRACT

"Muscovite" (white mica) is the most common mineralogical indicator of strongly peraluminous composition in plutonic rocks and, by inference, in their parental magmas. Its presence has been used to constrain depth of crystallization; based upon experimental data, approximately 3 kbar (11 km) is commonly considered the minimum pressure at which primary igneous muscovite can crystallize. Recent suggestions that independent criteria require depths < 11 km for emplacement of some granites with texturally primary-looking muscovite, so that such mica would in fact be secondary, raise questions about the use of apparently primary muscovite as an indication either of depth or of magma composition. New data from 41 samples representing 16 plutons in North America and Europe are relevant to the paragenesis of muscovite in igneous rocks. Formulas of the analyzed micas are typically about $K_{0.91}Na_{0.07}Fe^{3+}{}_{0.20}Fe^{2+}{}_{0.05}Mg_{0.10}Ti_{0.03}Al_{2.55}Si_{3.10}O_{10}(OH_{1.93}F_{0.07})$, with very slight trioctahedral substitution (2.00 to 2.04 octahedral cations). Primary- and secondary-looking grains are generally similar, but primary ones are richer in Ti, Na and Al and poorer in Mg and Si. Plutonic muscovite is so far from ideal $KAl_3Si_3O_{10}[OH]_2$ in composition that it is difficult to evaluate its paragenesis in terms of existing experimental data. The many additional components may enhance the stability field sufficiently to explain occurrences of primary muscovite at surprisingly shallow depths.

Keywords: muscovite, paragenesis, mineral chemistry, granitic rocks, peraluminous granites.

SOMMAIRE

La "muscovite" (dite mica blanc) est, dans une roche plutonique, l'indice le plus constant d'une composition fortement hyperalumineuse de la roche et, par induction, du magma originel. Sa présence sert à assigner des limites à la profondeur de cristallisation; d'après les résultats expérimentaux, une pression d'au moins 3 kbar (correspondant à 11 km) serait nécessaire pour que la muscovite primaire puisse cristalliser au sein du magma. Toutefois, des critères indépendants semblent indiquer une mise-en-place, à une profondeur inférieure à 11 km, de certains granites à muscovite d'apparence primaire. Si telle muscovite est réellement secondaire, elle ne peut indiquer ni la profondeur, ni la composition du magma. Des données nouvelles, établies sur 41 échantillons tirés de 16 plutons d'Amérique du Nord et d'Europe, contribuent à préciser la paragenèse de la muscovite plutonique. Une composition typique, $(K_{0.91}Na_{0.07})_{\Sigma 0.98}(Fe^{3+}{}_{0.20}Fe^{2+}{}_{0.05}Mg_{0.10}Ti_{0.03}Al_{1.65})_{\Sigma 2.03}(Al_{0.90}Si_{3.10})_{\Sigma 4}O_{10}(OH_{1.93}F_{0.07})$, montre un léger excès en cations octaédriques (de 2.00 à 2.04 cations). Les cristaux d'aspect primaire ou secondaire se ressemblent, sauf que les muscovites primaires contiennent plus de Ti, Na et Al et moins de Mg et Si. Ces muscovites primaires s'écartent à un point tel de la formule idéale que les données expérimentales sur la stabilité de $KAl_2AlSi_3O_{10}(OH)_2$ n'ont guère de rapport avec les conditions paragénétiques. Les nombreuses composantes qui viennent s'ajouter au système pourraient élargir suffisamment le champ de stabilité pour nous permettre d'expliquer la présence insolite de la muscovite primaire à faible profondeur.

(Traduit par la Rédaction)

Mots-clés: muscovite, paragenèse, composition minérale, roches granitiques, granite hyperalumineux.

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INTRODUCTION

The paragenesis of plutonic muscovite is of considerable petrological importance because the occurrence of apparently primary muscovite is commonly taken as an indication of both magma composition (strongly peraluminous) and depth of crystallization (greater than about 11 km). Although other distinctive aluminous minerals (such as garnet, cordierite, sillimanite and andalusite) may indicate a strongly peraluminous composition (defined here as containing more Al_2O_3 than can be accommodated in feldspars + aluminous biotite), muscovite is the most widespread and could be considered the most characteristic. The presumed pressure constraint imposed by the intersection of the granite solidus and the muscovite breakdown curve has varied from about 3 to 4 kbar (11 to 15 km), depending upon the granite composition and muscovite breakdown reaction used, and upon which experimental results are accepted (*e.g.*, discussion in Thompson 1974), but shallow, "epizonal" depths appear to be precluded.

In recent years many occurrences of primary-looking muscovite have been reported in plutons that have been emplaced at depths interpreted to be in the range 5 to 10 km (*e.g.*, Sylvester *et al.* 1978, Nelson & Sylvester 1971, Benoit 1971, Bradfish 1979, Swanson 1978, Banks 1977). These depth estimates have been based on structural and stratigraphic reconstructions as well as on the metamorphic environment (commonly andalusite-bearing aureoles in unmetamorphosed terranes). If these plutons have in fact been emplaced at such shallow depths, then either the muscovite is secondary in spite of its appearance, or past evaluations of experimental data are inapplicable to plutonic muscovite. In either case, the use of muscovite for depth estimation would be invalidated, and if it is indeed secondary, its use as an indicator of magma composition becomes questionable, because it may have formed metasomatically.

The purpose of this paper is to present data on the composition of plutonic muscovite that have a bearing upon its paragenesis and petrological interpretation.

PROCEDURE

Samples investigated

A total of 186 muscovite grains from 41 samples representing 16 plutons were analyzed (Table 1). A majority of the samples came from the Old Woman–Piute Range in southeastern California and the Teacup granodiorite of southern Arizona.

Analytical methods

The results of electron-microprobe mineral analyses presented here are from three different laboratories. Analyses from the Teacup granodiorite were done at the University of Arizona; the remainder were done at University of California at Los Angeles (UCLA) and at Virginia Polytechnic Institute. Accelerating voltage was 15 kV in all three laboratories; sample current was 20 nA (U. Ariz.) or 15 nA. Although some details of data collection varied among the laboratories, all used the correction factors of Albee & Ray (1970) in calculating weight percent oxides. Analytical accuracy and precision are estimated to be within ± 3% of the amount present for major elements, ± 5% for most minor elements (including Na) and ± 10% for F and Ba. For the determination of Fe^{3+}/Fe^{2+}, ^{57}Fe Mössbauer spectra were measured at UCLA on powdered, handpicked mineral separates of coarse white mica. Further details of analytical techniques are available from the authors.

Primary versus secondary micas: textural criteria

One of our goals in this study was to determine whether there were distinct compositional differences between primary and secondary muscovite in plutonic rocks or, conversely, to see whether compositional criteria were sufficient to distinguish between primary and secondary grains. To do this, we established textural criteria that would distinguish a primary (P) from a secondary (S) origin. For a P origin (*cf.*, Saavedra 1978), a mica grain must (1) have relatively coarse grain size, comparable to obviously primary phases; (2) be cleanly terminated, ideally with subhedral or euhedral form; (3) not be enclosed by, or raggedly enclose, a mineral (or other possible alteration products of a mineral) from which the muscovite may have formed by alteration (*e.g.*, feldspar, aluminum silicate); (4) be in a rock with clean, unaltered, igneous (in most cases hypidiomorphic-granular) texture.

We interpret any grain that meets all the above criteria as primary [but see, for example, Benoit (1971) for an opposing view]. Some grains that do not meet all of these criteria may also be primary; in particular, there may be coarse, subhedral primary muscovites remaining in an altered rock with secondary

TABLE 1. SAMPLES ANALYZED

Locality	Number of samples/ grains analyzed	Mineral Assemblages*	Representative Muscovite Formulas**	References
Old Woman-Piute Range, Mojave Desert, CA (5 plutons)	21/71	mb (± mt ± ilm), mbg (± mt), mg, mgbs	$K_{0.88}Na_{0.09}Fe_{0.23}Mg_{0.06}Ti_{0.03}$-$Al_{2.62}Si_{3.12}O_{10}(OH_{1.90}F_{0.10})$:P $K_{0.93}Na_{0.04}Fe_{0.27}Mg_{0.11}Ti_{0.01}$-$Al_{2.49}Si_{3.19}O_{10}(OH_{1.89}F_{0.11})$:S	Miller et al. 1980; Miller & Stoddard 1978
Teacup granodiorite, South-central AZ	7/58	m (± mt), mb (± mt ± ru), mbg (± mt), mg	$K_{0.94}Na_{0.05}Fe_{0.29}Mg_{0.12}Ti_{0.05}$-$Al_{2.50}Si_{3.12}O_{10}(OH_{2}F_{na})$:P $K_{0.95}Na_{0.04}Fe_{0.32}Mg_{0.16}Ti_{0.02}$-$Al_{2.40}Si_{3.19}O_{10}(OH_{2}F_{na})$:S	Bradfish 1979
Papoose Flat pluton, Inyo Mtns., eastern CA	1/3	mb (mt ± ilm)	$K_{0.92}Na_{0.05}Fe_{0.28}Mg_{0.12}Ti_{0.04}$-$Al_{2.46}Si_{3.17}O_{10}(OH_{1.87}F_{0.13})$:P	Sylvester et al. 1978
Birch Creek pluton, White Mtns., eastern CA	1/5	mb (mt)	$K_{1.03}Na_{0.06}Fe_{0.24}Mg_{0.09}Ti_{0.04}$-$Al_{2.63}Si_{3.05}O_{10}(OH_{1.96}F_{0.04})$:P	Nelson & Sylvester 1971
La Raya pluton, Peninsular Ranges, Baja Calif., Mexico	1/7	mb	$K_{0.86}Na_{0.08}Fe_{0.10}Mg_{0.05}Ti_{0.02}$-$Al_{2.86}Si_{3.04}O_{10}(OH_{1.98}F_{0.02})$:P	Gastil 1975
Stone Mountain granite, Inner Piedmont, GA	1/7	mb	$K_{0.92}Na_{0.06}Fe_{0.18}Mg_{0.11}Ti_{0.05}$-$Al_{2.57}Si_{3.13}O_{10}(OH_{1.91}F_{0.09})$:P	Whitney et al. 1976
Whiteside pluton, Blue Ridge, NC	1/1	mb	$K_{0.96}Na_{0.05}Fe_{0.25}Mg_{0.17}Ti_{0.04}$-$Al_{2.48}Si_{3.14}O_{10}(OH_{1.99}F_{0.01})$:P	Miller and Kish 1980
Spruce Pine Alaskite, Blue Ridge, NC	1/3	mbg	$K_{1.00}Na_{0.06}Fe_{0.31}Mg_{0.09}Ti_{0.00}$-$Al_{2.53}Si_{3.12}O_{10}(OH_{1.97}F_{0.03})$:P	Brobst 1962
Wyatt Station pluton, Raleigh Belt, NC	1/5	mbg (mt)	$K_{0.94}Na_{0.05}Fe_{0.29}Mg_{0.09}Ti_{0.06}$-$Al_{2.49}Si_{3.14}O_{10}(OH_{1.93}F_{0.07})$:P	Parker 1979
Sims pluton, Eastern Slate Belt, NC	1/7	m (mt?)	$K_{0.94}Na_{0.02}Fe_{0.30}Mg_{0.29}Ti_{0.03}$-$Al_{2.23}Si_{3.26}O_{10}(OH_{1.91}F_{0.09})$:S	Wedemeyer & Spruill 1980
Gresham's Lake pluton, Raleigh Belt, NC	3/9	mbg (mt), mg (± mt)	$K_{0.91}Na_{0.05}Fe_{0.27}Mg_{0.07}Ti_{0.02}$-$Al_{2.60}Si_{3.12}O_{10}(OH_{1.96}F_{0.04})$:P	Parker 1979
Argentera Massif, Maritime Alps, France - Italy	2/10	mg, mbg	$K_{0.83}Na_{0.10}Fe_{0.18}Mg_{0.08}Ti_{0.02}$-$Al_{2.70}Si_{3.09}O_{10}(OH_{2}F_{na})$:P	De Pol et al. 1968

*varietal minerals: M = muscovite, b = biotite, g = garnet, s = sillimanite; (accessory minerals): (mt) = magnetite, (ilm) = ilmenite, (ru) = rutile. **P = Primary-looking, S = secondary-looking (see text); all Fe calculated as Fe (2+); minor Mn, Cr, Ca, Ba, and Cl omitted; OH+F assumed = 2; na = not analyzed.

muscovite; in fact, this is the most interesting case for analysis. Therefore, we assign those grains that meet the first three criteria, even those found in mildly altered rocks, to the "P-mica" category, whereas small, ragged grains, particularly those that are confined within or at the margins of certain minerals, are considered "S-micas", or secondary. Figure 1 shows typical examples.

Since by no means were all grains easily placed in one or the other category, some of the designations as P- or S-mica are somewhat uncertain.

RESULTS

The few published analyses of plutonic muscovite are commonly high in total Fe + Mg (0.2 to 0.5 cations *per* formula based on 11 oxygen atoms), contain noteworthy Ti, and

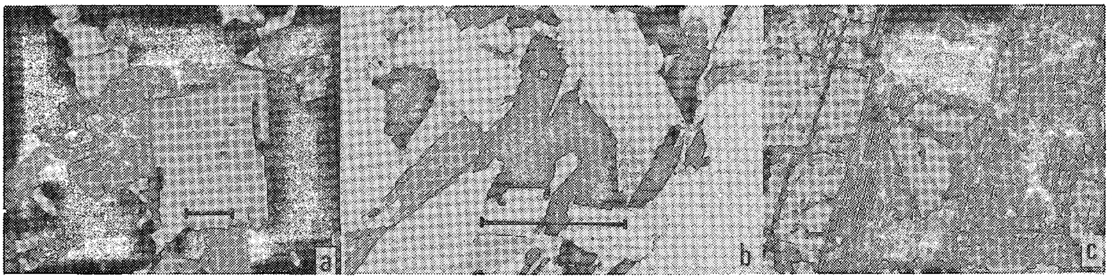

FIG. 1. Typical P- and S-micas; bars indicate 1 mm. (a) P: coarse, subhedral, cleanly terminated grain, Old Woman-Piute Range (crossed polars); (b) P: coarse, subhedral, randomly oriented grains partially enclosed by (later?) biotite; Old Woman–Piute Range (plane polars); (c) S: small grains enclosed within plagioclase (as alteration?), Teacup granodiorite (crossed polars).

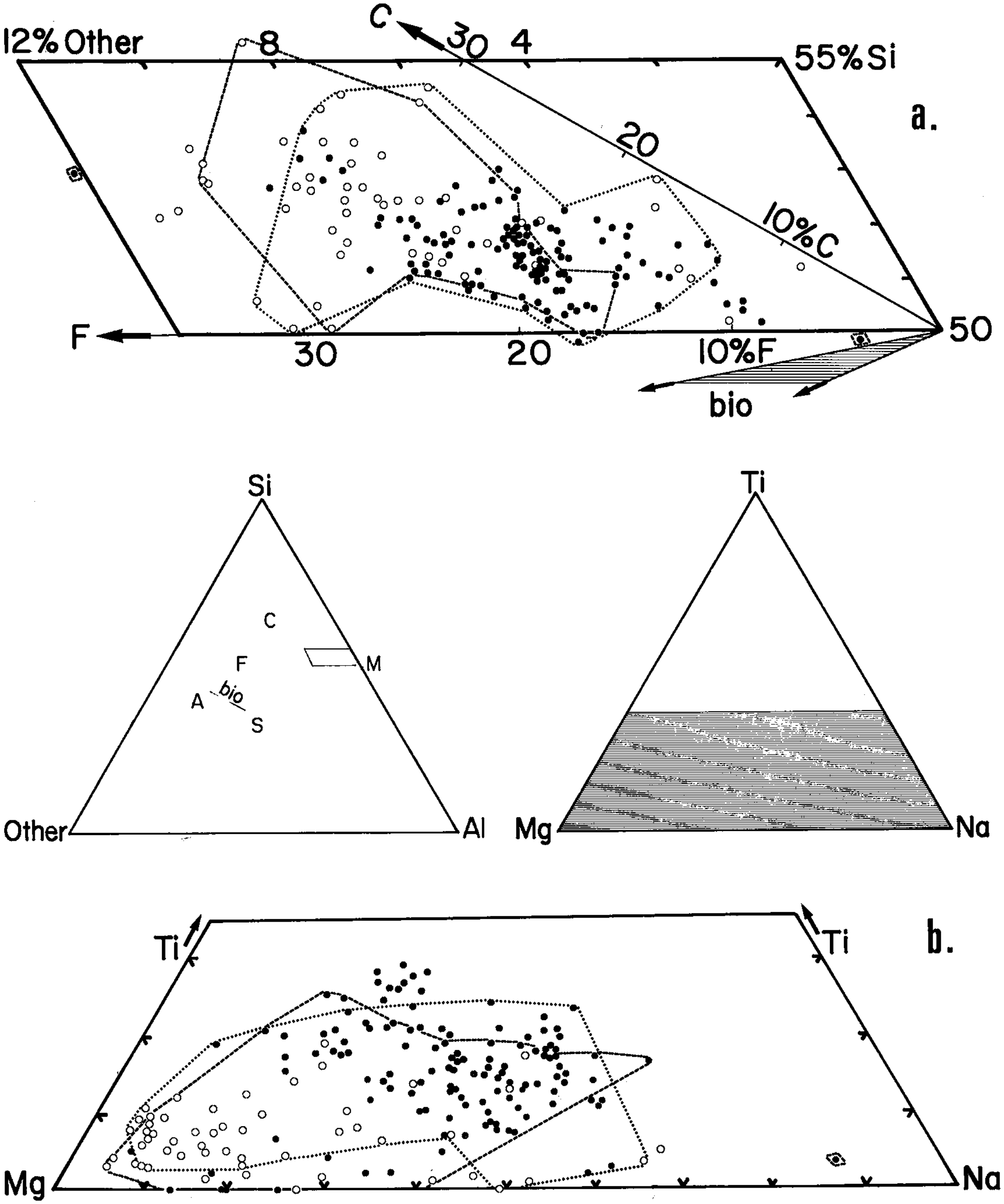

FIG. 2. Compositions of analyzed muscovite grains. Closed circles = P-micas; open circles = S-micas; dotted lines enclose fields of Old Woman–Piute Range micas; dashed lines enclose Teacup granodiorite micas. (a) Compositions in terms of Si, Al and other octahedral cations (Fe, Mg, Ti, Mn, Cr) (atomic proportions). The inset shows the location of mica end-members: M muscovite, fluormuscovite and paragonite, F ferrimuscovite, C celadonite, T end-member formed by total substitution Ti Mg^{vi} → 2 Al^{vi}, A annite–phlogopite, and S siderophylite–eastonite (see text for formulas). Ruled area shown in figure. (b) Compositions in terms of Ti, Mg and Na (atomic proportions). Ruled area in inset is the area shown in the main figure.

have excess Si and deficient Al compared with ideal muscovite (Joyce 1973, Best *et al.* 1974, Neiva 1975, Mohon 1975, Becker 1978, Anderson *et al.* 1980, Anderson & Rowley 1981). Muscovites investigated by Castle & Theodore (1972), Harris (1974) and Guidotti (1978a) approach ideality more closely but still have about 0.15 (Fe + Mg)/formula unit.

Our data confirm that plutonic "muscovites" differ significantly from the ideal stoichiometry $\mathrm{KAl_3Si_3O_{10}[OH]_2}$. The average formula (Table 1) is approximately $\mathrm{K_{0.91}Na_{0.07}Fe^{3+}_{0.20}Fe^{2+}_{0.05}Mg_{0.10}Ti_{0.03}Al_{2.55}Si_{3.10}O_{10}(OH_{1.93}F_{0.07})}$. The elements Ca, Ba, Mn, Cr, Cl and P are present in negligible quantities (< 0.01/formula).

Formulas calculated for muscovites assuming all Fe as $\mathrm{Fe^{2+}}$ have 2.04 to 2.10 octahedral cations *per* formula, suggesting that much of the Mg and Fe might be present as a trioctahedral (biotite) component. However, in most of the samples analyzed by Mössbauer spectroscopy, over 70% of the Fe is present as $\mathrm{Fe^{3+}}$ (total range was 27 to 88%). Recalculating the formulas of these samples using the $\mathrm{Fe^{3+}/Fe_{total}}$ ratios determined indicate a far smaller octahedral excess (Table 2); with a reasonable estimate of $\mathrm{Fe^{3+}/Fe_{total}}$ (0.75), octahedral site occupancy in all samples is 2.02 ± 0.02 (1σ) *per* formula. If all Fe and Mg were trioctahedral, the octahedral site would have [2 + (Fe + Mg)/3] cations, neglecting minor effects of Ti. Because Fe + Mg typically total more than 0.3 cations/formula, it is clear that trioctahedral substitution is very minor.

TABLE 2. EXAMPLE OF CORRECTION OF MUSCOVITE COMPOSITION FOR FERRIC IRON*

	Fe(3+)/Fe(total)=0.80			
	uncorrected		corrected	
SiO2	45.83		45.83	
Al2O3	32.53		32.53	
TiO2	0.47		0.47	
MgO	0.63		0.63	
Fe2O3	--		3.56	
FeO	4.01		0.80	
MnO	0.03		0.03	
Cr2O3	0.01		0.01	
CaO	0.02		0.02	
Na2O	0.62		0.62	
K2O	10.25		10.25	
BaO	0.05		0.05	
F	0.48		0.48	
O = F	-0.20		-0.20	
Total	94.73		95.08	
Formula, O = 11:				
Si	3.129		3.103	
Al(iv)	0.871		0.897	
Al(vi)	1.743		1.695	
Ti	0.024		0.024	
Mg	0.064	2.062	0.063	2.011
Fe(3+)	--		0.181	
Fe(2+)	0.229		0.046	
Mn	0.002		0.002	
Cr	0		0	
Ca	0.002		0.002	
Na	0.082	0.976	0.081	0.968
K	0.891		0.884	
Ba	0.001		0.001	
F			0.102	
ΣAl	2.614		2.592	

*Sample SW-8, muscovite-biotite granite, Old Woman Mountains; Fe(3+)/Fe(total) from coarse muscovite separate, rest of analysis = average of P-mica analyses.

Our data do not allow formulas to be recast uniquely as combinations of ideal end-members because (1) the nature of Ti substitution is not fully understood and (2) for most samples $\mathrm{Fe^{3+}/Fe_{total}}$ is not known. End members that must be present include ideal muscovite $\mathrm{KAl_3Si_3O_{10}[OH]_2}$, ferrimuscovite $\mathrm{KFe^{3+}_2AlSi_3O_{10}[OH]_2}$, celadonite $\mathrm{K[Mg,Fe]AlSi_4O_{10}[OH]_2}$, paragonite $\mathrm{NaAl_3Si_3O_{10}[OH]_2}$, fluor-muscovite $\mathrm{KAl_3Si_3O_{10}F_2}$, biotite $\mathrm{K[Fe,Mg]_3AlSi_3O_{10}[OH]_2}$–$\mathrm{K(Fe,Mg)_{2.5}Al_2Si_{2.5}O_{10}[OH]_2}$ and a Ti-bearing species. The Ti substitution may involve (1) $\mathrm{[(Mg,Fe) + Ti]^{vi} \rightleftharpoons 2Al^{vi}}$, giving $\mathrm{K(Mg,Fe)TiAlSi_3O_{10}(OH)_2}$ as the end member; (2) $\mathrm{Ti^{vi} + Al^{iv} \rightleftharpoons Al^{vi} + Si^{iv}}$, giving $\mathrm{KTi_2Al_3SiO_{10}(OH)_2}$; or (3) $\mathrm{Ti^{vi}} + \square \rightleftharpoons \mathrm{2(Mg,Fe)}$, leading to $\mathrm{KTi_{0.5}AlSi_4O_{10}(OH)_2}$ (Dymek & Albee 1977). Guidotti (1978b) considered substitution (1) to be most important in muscovite.

In Figure 2a our analyses are plotted in terms of Si,Al and the sum of octahedral substituents Fe, Mg, Ti, Mn and Cr. Although the presence of minor amounts of trioctahedral (biotite) and Ti-bearing end-members complicates the interpretation of the diagram, it is clear that ferrimuscovite and celadonite are important components, and that our analyses average approximately 75% muscovite + paragonite + fluor-muscovite. Paragonite and fluor-muscovite typically total nearly 10%, leaving approximately 65% of the pure muscovite end-member.

Although our data reveal general consistency among plutonic muscovite compositions, Figure 2 and Table 1 illustrate considerable variation from pluton to pluton and from sample to sample. The variability is not clearly linked to the coexisting mineral assemblage (Table 3). Table 4 reveals that there are also large differences among grains within a single sample. Much of the within-sample variation seems attributable to paragenesis; we believe that differences in composition may arise depending upon the stage at which the individual mica grain crystallized or last equilibrated (see below).

Coexisting phases

In all samples examined, muscovite coexists

TABLE 3. MEAN MUSCOVITE COMPOSITIONS IN DIFFERENT MINERAL ASSEMBLAGES**

Assemblage	mu + bi	mu + bi + gt	mu +gt	mu + bi + gt + sil
# samples	6	10	4	1
# grains	19	16	12	4
SiO2	45.72 ± 0.46 (1σ)	45.85 ± 0.66	45.30 ± 0.85	45.16 ± 0.24
Al2O3	32.43 ± 1.13	32.45 ± 1.34	32.42 ± 0.82	34.09 ± 0.24
TiO2	0.67 ± 0.18	0.45 ± 0.21	0.64 ± 0.19	0.08 ± 0.07
MgO	0.73 ± 0.16	0.75 ± 0.19	0.73 ± 0.07	0.61 ± 0.07
FeO*	3.92 ± 0.74	4.04 ± 0.77	3.70 ± 0.44	2.15 ± 0.22
MnO	0.04 ± 0.03	0.05 ± 0.02	0.05 ± 0.02	0.01 ± 0.01
CaO	0.04 ± 0.04	0.09 ± 0.04	0.08 ± 0.04	0.09 ± 0.03
Na2O	0.58 ± 0.18	0.41 ± 0.12	0.70 ± 0.10	0.59 ± 0.08
K2O	10.44 ± 0.23	10.43 ± 0.36	10.38 ± 0.40	10.23 ± 0.07
F	0.36 ± 0.10	NA	0.26 ± 0.05	NA
O = F	-0.15		-0.11	
Anhydrous Total	94.78	94.52	94.15	93.01
Formula, O = 11***				
Si	3.115 ± 0.032	3.124 ± 0.022	3.104 ± 0.042	3.093 ± 0.010
Al(iv)	0.885	0.876	0.996	0.907
Al(vi)	1.718	1.730	1.623	1.846
Ti	0.034 ± 0.010	0.023 ± 0.011	0.033 ± 0.010	0.004 ± 0.004
Mg	0.074 ± 0.016 2.052	0.076 ± 0.020 2.063	0.074 ± 0.007 2.045	0.062 ± 0.008 2.035
Fe	0.224 ± 0.044	0.231 ± 0.046	0.212 ± 0.025	0.123 ± 0.013
Mn	0.002 ± 0.002	0.003 ± 0.001	0.003 ± 0.001	0
Ca	0.003 ± 0.003	0.007 ± 0.003	0.006 ± 0.003	0.007 ± 0.001
Na	0.076 ± 0.023 0.988	0.054 ± 0.017 0.968	0.093 ± 0.014 1.006	0.078 ± 0.010 0.979
K	0.907 ± 0.020	0.907 ± 0.034	0.907 ± 0.036	0.894 ± 0.007
F	0.080 ± 0.020	--	0.060 ± 0.012	--
ΣAl	2.603 ± 0.070	2.606 ± 0.074	2.619 ± 0.055	2.753 ± 0.010

*Fe as FeO. **Old Woman-Piute Range P-micas. ***All Fe Calculated as Fe(2+). NA not analyzed.

with quartz, K-feldspar and sodic plagioclase. Most also contain Al-rich (> 1.5 Al/11 oxygen atoms) biotite, and many have almandine–spessartine garnet. The most common accessory minerals are apatite, zircon and magnetite; ilmenite, rutile and primary-looking epidote are present in a few samples. Sillimanite was identified in one sample.

The occurrence of Al-rich biotite and garnet is consistent with a strongly peraluminous magma composition.

Previous interpretations of compositional variation

In recent years, celadonite-rich micas or phengites commonly have been considered to be restricted to relatively high-P and low-T environments. There are considerable analytical data that support the occurrence of celadonitic micas in such environments (Miyashiro 1973), and some experimental data suggest that these micas are unstable in igneous or high-grade metamorphic environments [*e.g.*, Velde (1965), but see Crowley & Roy (1964) for a different interpretation]. Brimhall (1972) reported celadonite-rich compositions similar to ours for clearly secondary, sericitic mica, but these grains are not low-T alterations: he estimated that they were formed in the range 550–690°C. White mica in high-grade metapelites, though higher in the muscovite component than our samples, is also not ideal in composition; for example, Guidotti (1978b) reported upper-sillimanite-zone muscovite that has more than 10% celadonite or celadonite + Ti-bearing component. Anderson & Rowley (1981) propose, on the basis of thermodynamic considerations, that celadonitic muscovite should be stable to *higher* T than pure muscovite.

Composition as a function of coexisting assemblage

Table 3 illustrates the variation in muscovite composition as a function of coexisting mineral assemblage for samples from the Old Woman–Piute Range. Muscovites coexisting with biotite, garnet, and garnet + biotite are almost identical. This is somewhat surprising, because whole-rock and biotite chemistry do vary systematically. Muscovites from the single sillimanite-bearing sample are distinctly different from the rest, with higher Al and lower Fe, Mg, Ti and Si. Analyses from this sample differ from all others determined during this project, with the exception of those from the La Raya pluton, Baja California. The La Raya sample does not contain sillimanite, nor has it been found in the pluton as a whole (R.G. Gastil, pers. comm. 1980), but similar analyses have been reported for other sillimanite-bearing plutonic rocks (Guidotti 1978a).

Composition as an indicator of primary or secondary origin

Several compositional characteristics distinguish most of our texturally primary-looking (P) from secondary-looking (S) micas. Most textural P-micas are considerably richer in Ti, Al and Na and poorer in Mg and Si than S-micas (Fig. 2b) [*cf.*, Anderson & Rowley (1981) for rather different result]. P-micas are typically somewhat closer to ideal muscovite than their S-mica counterparts, primarily because of lower Mg (Fig. 2a).

Table 4 illustrates compositional differences within a single sample, and Figure 3 shows typical P- and S-micas from that sample.

Imperfect correlation between composition and our textural types is probably a function of (1) our inability to evaluate paragenesis of all grains on a textural basis and (2) the complex physical–chemical environment and history of these grains. Secondary muscovite probably forms under a variety of conditions, and primary muscovite may be susceptible to subsolidus re-equilibration. Still, the consistency of compositional distinctions is striking and may prove useful in the interpretation of other muscovite granites.

The high Ti of apparently primary muscovite has been noted by Anderson (*e.g.*, Anderson & Rowley 1981); he points out the analogy with high-grade metamorphic muscovite, which is also enriched in Ti compared with low-grade muscovite (Guidotti 1978b). The significance of high Na is difficult to evaluate because maximum Na in muscovite increases with T at low to moderate metamorphic grades, but near magmatic T, where the assemblage muscovite + K-feldspar + Al-silicate is stable, the trend reverses (*cf.*, Thompson 1974).

The lower celadonite component of P-type micas would qualitatively agree with Velde's (1965, 1967) conclusion that celadonite abundance should diminish with increasing T. However, according to Velde's experimental data, all of our micas, P or S, have higher celadonite than should be tolerated at temperatures approaching those of magmas (see also plot in Miyashiro 1973, p. 202). The discrepancy between our apparently igneous mica compositions and those predicted by Velde may be a function of an oversimplified experimental environment. It is conceivable, though in our view less likely, that the P-micas are either partially re-equilibrated primary micas or coarsely crystallized secondary micas (Bradfish 1979). If so, the differences between P and S compositions must reflect either imperfect equilibration or crystallization under different subsolidus conditions.

Pressure constraints implied by primary muscovite

The minimum pressure at which muscovite can coexist with silicate liquid is dependent upon so many factors that are currently poorly understood that we regard precise estimation as impossible. Both the granite solidus and the muscovite-breakdown (muscovite + albite + in either lead to large uncertainties in their

TABLE 4. TYPICAL P- AND S-MICAS FROM A SINGLE SAMPLE**

Grain, type -	A(S)		D(P)	
SiO2	46.43		45.55	
Al2O3	29.18		32.32	
TiO2	0.31		0.77	
MgO	1.64		0.74	
FeO*	5.56		4.12	
MnO	0.18		0.07	
Cr2O3	0.02		0.02	
CaO	0.01		0.01	
Na2O	0.19		0.75	
K2O	10.93		10.20	
BaO	0.04		0.02	
F	0.51		0.28	
O = F	-0.21		-0.12	
Anhydrous Total	94.79		94.73	
Formula, O = 11***				
Si	3.205		3.105	
Al(iv)	0.795		0.895	
Al(vi)	1.579		1.704	
Ti	0.016		0.039	
Mg	0.169	2.097	0.076	2.059
Fe	0.321		0.235	
Mn	0.011		0.004	
Cr	0.001		0.001	
Ca	0.001		0	
Na	0.025	0.999	0.099	0.987
K	0.962		0.888	
Ba	0.001		0	
F	0.111		0.061	
ΣAl	2.374		2.599	

*Fe as FeO;
**P15, muscovite-biotite granite, Old Woman-Piute Range (grains illustrated in Figure 3)
***All Fe calculated as Fe(2+)

placement. An error of 10°C in either will change their intersection by about 0.4 kbar (1.5 km). Most workers use the intersection of the curve muscovite + quartz → K-feldspar + sillimanite + H_2O with the water-saturated solidus in the system Ab + Or + Qz to define quartz → K-feldspar + Al-silicate + H_2O) curve are steep, and therefore small uncertainties the minimum P. The exact location of the former curve is open to some question (*cf.*, Day 1973, Althaus *et al.* 1970) and should in any case be at lower T because of the involvement of the Ab component in the reaction (Thompson 1974), leading to an intersection at higher pressures. The granite solidus will be raised by the presence of the An component and by $a(H_2O) < 1$ (therefore higher P intersection) and lowered by the presence of mafic components, excess Al_2O_3 (Abbott & Clarke 1979) and by the presence of boron (Chorlton & Martin 1978), thus lowering the P of intersection. The effect of boron may be particularly important.

The nonideal composition of plutonic muscovite also must influence the position of the granite solidus/muscovite breakdown intersection. Extrapolation of Velde's (1965) data suggests that muscovites with the compositions we find could only be stable at unreasonably high P (probably at least 7 or 8 kbar, or 30 km). Thermodynamic considerations lead Anderson & Rowley (1981) to an opposite conclusion, that their muscovite, which is within our compositional range, has a higher T stability limit than ideal muscovite and could have crystallized from granodioritic melt at 2 kbar (7.6 km).

We tentatively support the reasoning of Anderson & Rowley, which leads to the conclusion that the effect of celadonitic impurities alone can explain the occurrence of primary muscovite in plutons emplaced at apparently anomalously low pressure. Most importantly, however, we wish to emphasize the uncertainty of the pressure range in which primary igneous muscovite can crystallize and to caution against petrological interpretations based upon the 3–4 kbar (11–15 km) minimum.

CONCLUSIONS

Our conclusions regarding plutonic muscovites may be summarized as folllows:

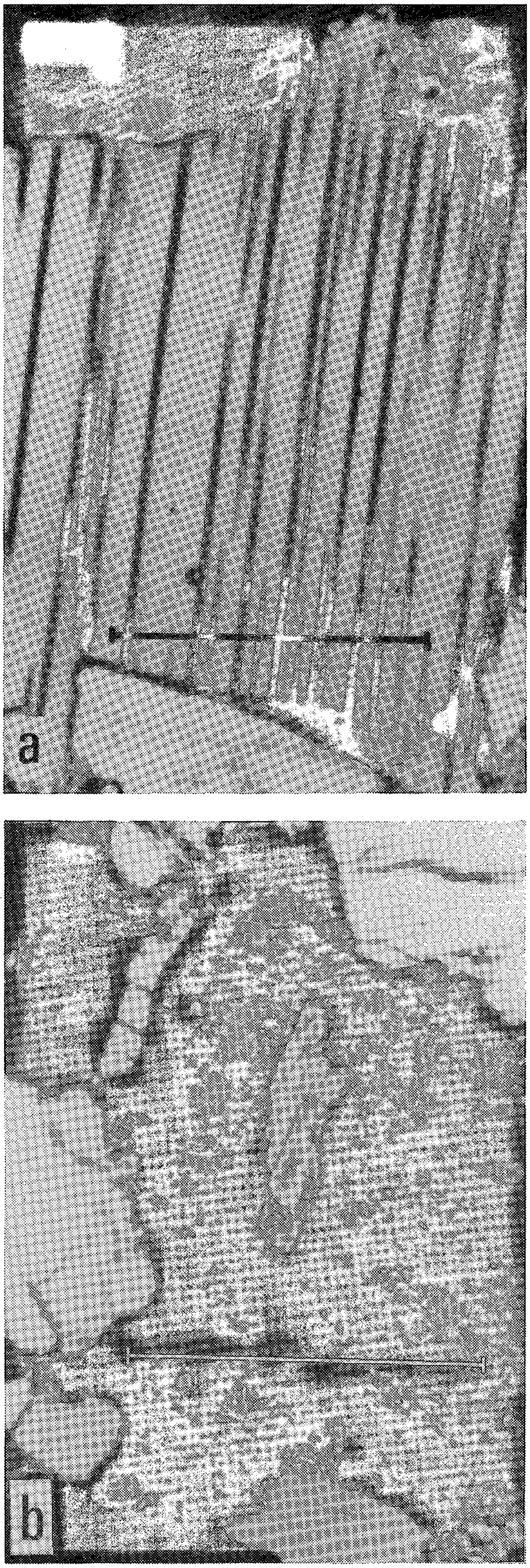

FIG. 3. Examples of analyzed P- and S-micas from sample P15, Old Woman–Piute Range; bar indicates 1 mm. See Table 4 for analyses. (a) grain D (P-mica). (b) grain A (S-mica).

(1) The analyzed muscovites are far from ideal end-member muscovite. They contain appreciable Fe, Mg, Ti, Na and F. They are deficient in Al and have a modest excess of Si.

(2) Because the analyzed micas are so nonideal, experimental data are not directly applicable to their stability. Hence, the intersection of the reaction curve muscovite + albite + quartz = K-feldspar + Al-silicate + H_2O with the granite minimum melting curve is not the true minimum pressure of crystallization of primary igneous muscovite. If, as suggested by Anderson & Rowley (1981), this reaction occurs at higher T for celadonitic muscovite, the minimum P is lower (they suggest 2 kbar, or 7.6 km, for their samples). Alternatively, if one accepts the conclusion of Velde (1965, 1967) that celadonitic mica is confined to lower T than muscovite, either much higher P is required for crystallization of these micas as magmatic minerals, *or* the mica compositions represent subsolidus re-equilibration, *or* the micas are all relatively low-T secondary minerals. The final alternative eliminates plutonic muscovite as an indicator of emplacement pressure and casts doubt on its value as an indicator of primary magma composition. We tentatively agree with the interpretation that the impurities increase the stability field of muscovite (Anderson & Rowley 1981) and that these micas may therefore have crystallized from magmas at relatively shallow depths ($<$ 10 km). We further emphasize the many other uncertainties in placement of both granite-solidus and muscovite-breakdown curves that make estimates of their intersection (minimum P of primary muscovite) very imprecise.

(3) The fact that primary-looking muscovite almost invariably coexists with aluminous biotite ± garnet suggests that such muscovite is indeed restricted to rocks crystallized from primary strongly peraluminous magmas, even if the muscovite itself may be secondary.

(4) Compositional differences may distinguish primary from secondary muscovite. Grains that appear texturally to be primary typically have higher Ti, Na and Al and lower Si and Mg than those that appear to be secondary.

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