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THE EAST KEMPTVILLE TOPAZ–MUSCOVITE LEUCOGRANITE, NOVA SCOTIA. II. MINERAL CHEMISTRY

DANIEL J. KONTAK
Nova Scotia Department of Mines and Energy, P. O. Box 1087, Halifax, Nova Scotia B3J 2X1

ABSTRACT

The major- and trace-element chemistry of the mineral constituents of the East Kemptville leucogranite reflects chemical equilibration at both the magmatic (muscovite, bulk alkali feldspar, some biotite, topaz) and postmagmatic (albitic plagioclase, exsolved alkali feldspar, apatite in albite, most biotite, biotite alteration products) stages. Muscovite is characterized by elevated Fe (to 12 wt.% FeO) and F [to >4 wt. % with low Fe^{3+}/Fe^{2+} ratios (0.14)]; absolute abundances of selected trace elements and ratios indicate compositions similar to white mica from pegmatites. The bulk composition ($Or_{82}Ab_{17}An_1$) and trace-element chemistry of alkali feldspar are also more typical of pegmatites. Topaz chemistry [F/(F+OH) = 0.80 ± 0.05] is consistent with high temperatures of formation, the inferred mineral assemblage and relatively dry nature of the melt. Biotite compositions vary considerably from freshest [Fe/(Fe+Mg) > 0.95, low Ti (<1.14 wt.% TiO_2), 2.7 wt.% F] to chemistries that deviate considerably from ideal trioctahedral chemistry due to postmagmatic alteration. The albitic (Ab_{99}) composition of plagioclase containing abundant apatite inclusions reflects a late-magmatic or early postmagmatic albitization of precursor Ca-bearing plagioclase. The mineral chemistry indicates a highly reduced nature for the melt, similar to that in the case of topaz rhyolites. Collectively, (1) the presence of muscovite, (2) feldspar thermometry and (3) covariation of K and Rb between alkali feldspar and muscovite indicate a T of 500–700°C, whereas primary muscovite suggests that P was $\geq$ 1 kbar. Postmagmatic equilibration continued to $\leq$350°C, on the basis of (1) the presence of chloritic alteration of biotite and (2) two-feldspar thermometry of exsolved phases in alkali feldspar. Postmagmatic alteration proceeded under low $f(O_2)$, as the secondary biotite is characterized by elevated Fe/(Fe+Mg) ratios.

Keywords: topaz–muscovite leucogranite, magmatic topaz, albitization, tin deposit, East Kemptville, Nova Scotia.

SOMMAIRE

La composition des minéraux du leucogranite de East Kemptville (Nouvelle-Écosse), en termes des éléments majeurs et des éléments traces, résulte d'un équilibre aux stades magmatique (muscovite, composition globale du feldspath alcalin, une partie de la biotite, topaze) et postmagmatique (albite, exsolution dans le feldspath alcalin, apatite dans l'albite, la plupart de la biotite, et les produits de l'altération de la biotite). La muscovite contient des fortes concentrations de Fe (jusqu'à 12% FeO en poids) et F (dépassant même 4% en poids), mais le rapport Fe^{3+}/Fe^{2+} est faible (0.14). Les concentrations et les rapports de certains éléments traces indiquent des compositions semblables au mica blanc de milieux pegmatitiques. La composition du feldspath alcalin ($Or_{82}Ab_{17}An_1$) et les teneurs en éléments traces aussi sont typiques d'un milieu pegmatitique de cristallisation. La composition du topaze [F/(F+OH) = 0.80 ± 0.05] concorde avec une température élevée de formation, l'assemblage des minéraux à ce stade, et la faible teneur du magma en eau. La composition de la biotite est très variable; la fraction la plus saine possède un rapport Fe/(Fe+Mg) supérieur à 0.95 et une faible teneur en Ti (<1.14% TiO_2, 2.7% F). La fraction qui s'est ré-équilibrée au stade post-magmatique s'écarte considérablement de la stoechiométrie trioctaédrique idéale. La composition albitique (Ab_{99}) du plagioclase qui contient une abondance d'inclusions d'apatite résulterait d'une albitisation tardi-magmatique ou post-magmatique précoce d'un précurseur à teneur en Ca plus élevée. La composition des minéraux indique un magma fortement réduit, tout comme dans le cas des rhyolites à topaze. La présence de la muscovite, la thermométrie fondée sur les feldspaths coexistants, et la covariation de K et de Rb entre feldspath alcalin et muscovite indiquent une température entre 500° et 700°C, tandis que la présence d'une muscovite primaire indique une pression supérieure ou égale à 1 kbar. Le ré-équilibrage a continué même au delà de 350°C, comme en font foi l'altération chloritique de la biotite et la thermométrie fondée sur les deux feldspaths coexistants dans la perthite. L'altération post-magmatique a procédé sous conditions de faible $f(O_2)$, comme en témoignent les valeurs élevés du rapport Fe/(Fe+Mg).

(Traduit par la Rédaction)

Mots-clés: leucogranite à topaze + muscovite, topaze magmatique, albitisation, gisement d'étain, East Kemptville, Nouvelle-Écosse.

INTRODUCTION

The geological setting, petrography and whole-rock geochemistry of the East Kemptville topaz–muscovite leucogranite, host rock to the East Kemptville tin – base metal deposit (56 million tonnes, 0.165% Sn), have been discussed in detail in Part I (Kontak 1990). Kontak concluded that the leucogranite is the product of dominantly magmatic processes, albeit with a certain degree of metasomatic modification, and that it represents an example of a volatile-rich (*i.e.*, F), highly evolved magma with affinities to the Mongolian ongonites (Kovalenko *et al.* 1971) and topaz rhyolites of the western United

States (Burt *et al.* 1982, Christiansen *et al.* 1983, 1984, 1986).

The postmagmatic overprinting of highly evolved felsic rocks by either cognate or exotic fluids is a well-established phenomenon, generally advocated on the basis of textural observations (*e.g.*, Clarke & Taylor 1985, Pijpekamp 1982, Pollard 1983, 1984, Pollard & Taylor 1986, Taylor & Pollard 1988, Stone & Exley 1986, Manning & Exley 1984, Witt 1988). There remain, however, few detailed geochemical studies of the mineral constituents of such rocks (*e.g.*, ongonites, topaz granites) to document the transition from magmatic through to metasomatic-dominant mineralogy. Until such information is available, it will not be possible to fully appreciate the relationship among magmatic, orthomagmatic and hydrothermal processes.

Although geochemical compilations of specialized granitic suites exist (*e.g.*, Tischendorf 1977), it is not clear whether elements such as Sn, W, Li, Rb and F represent primary or secondary enrichments; few authors discuss these conflicting processes. The same may be said for individual intrusive complexes that show progressive evolution toward highly evolved compositions, in some cases culminating in mineralized centers. For example, a well-documented case of this is the *ca.* 374 Ma Ackley Granite, in southeastern Newfoundland (Tuach 1987, Tuach *et al.* 1986, Kontak *et al.* 1988). Although the work of Tuach and coworkers has demonstrated the chemical evolution of the magmatic–hydrothermal system in the Ackley Granite, the relationship of the two processes in rocks proximal to mineralized centers remains poorly constrained.

Recent experimental work on naturally occurring topaz-bearing granites from St. Austell, Cornwall, and Beauvoir, France, by Weidner & Martin (1987) and Pichavant *et al.* (1987), respectively, has demonstrated that topaz may be interpreted as a primary liquidus phase. However, these two studies also indicated additional important information with respect to other minerals. Firstly, Weidner & Martin (1987) concluded that the albitic plagioclase in the St. Austell granite was of magmatic origin and, hence, that the igneous body represented an excellent example of a subsolvus granite. In addition, these authors found that muscovite was stable to much higher temperatures than indicated by previous experimental work on end-member OH-muscovite. On the other hand, the work of Pichavant *et al.* (1987) showed that the experimentally produced plagioclase is more Ca-rich than that in the natural specimen, thus confirming petrographic and geochemical data that suggest that all features of the Beauvoir granite cannot be explained solely by magmatic processes. Thus, although having contrasting views on the nature of albite in topaz-bearing granites, these two studies indicate the importance of providing detailed mineralogical information on such rocks.

In this paper, the chemistry of the major mineral phases of the leucogranite (plagioclase, alkali feldspar, muscovite, topaz, biotite, apatite) is presented and discussed. Although petrographic features of the minerals indicate that both magmatic and postmagmatic processes have left their signature, the degree to which these processes are developed is best examined *via* mineral chemistry. It will be demonstrated that different minerals have re-equilibrated to vary-

TABLE 1. REPRESENTATIVE COMPOSITIONS OF PLAGIOCLASE

	EK-13A		EK-60		EK-23		EK-69		TM-86	EK-71	TM-86
	1C	1R	2C	2R	3C	3R	4C	4C	5L	6P	7P
SiO_2	68.62	69.18	70.27	70.55	69.53	69.17	68.16	69.04	68.59	68.61	68.66
Al_2O_3	19.05	19.23	19.29	19.24	19.92	20.12	19.63	19.73	19.71	19.75	19.76
FeO	0.16	0.08	0.14	0.18	0.01	0.01	0.08	0.00	0.15	0.13	0.13
MnO	0.20	0.19	0.00	0.00	0.00	0.00	0.04	0.05	0.08	0.05	0.06
CaO	0.00	0.06	0.06	0.05	0.00	0.03	0.03	0.00	0.12	0.10	0.10
Na_2O	11.45	11.91	11.09	10.96	10.41	11.11	11.57	11.76	10.96	10.79	11.24
K_2O	0.09	0.09	0.04	0.00	0.00	0.00	0.00	0.03	0.08	0.06	0.06
Σ wt.%	99.55	100.74	100.89	100.97	99.86	100.43	99.62	100.65	99.69	99.49	100.01
Structural Formulae Based on 8 Oxygen Atoms											
Si	3.008	3.003	3.027	3.033	3.016	2.995	2.986	2.993	2.992	2.995	2.989
Al	0.984	0.983	0.979	0.975	1.018	1.026	1.013	1.008	1.012	1.016	1.013
Fe	0.005	0.002	0.005	0.006	0.000	0.000	0.002	0.000	0.004	0.004	0.004
Mn	0.007	0.003	0.000	0.000	0.000	0.000	0.000	0.001	0.002	0.001	0.001
Ca	0.000	0.002	0.002	0.001	0.000	0.000	0.000	0.000	0.004	0.004	0.004
Na	0.972	1.002	0.926	0.913	0.875	0.932	0.983	0.988	0.926	0.912	0.948
K	0.004	0.004	0.002	0.000	0.000	0.000	0.000	0.001	0.004	0.003	0.003
End-Member Feldspar, Mole %											
Ab	99.6	99.4	99.6	99.9	100.0	100.0	100.0	99.8	99.1	99.2	99.2
Or	0.4	0.4	0.2	0.0	0.0	0.0	0.0	0.2	0.4	0.3	0.3
An	0.0	0.2	0.2	0.1	0.0	0.0	0.0	0.0	0.4	0.4	0.4

C = core; R = rim; L = albite lamellae in alkali feldspar;
P = patch of remnant plagioclase in alkali feldspar

ing degrees, such that a spectrum of magmatic through to postmagmatic signatures is preserved.

ANALYTICAL TECHNIQUES

Major-element analyses were obtained using the JEOL 733 Superprobe at Dalhousie University, and a combined wavelength- and energy-dispersion analytical technique. An accelerating voltage of 15 kV, a probe current of about 5 nA and a beam diameter of about 1 μm (defocused to 10 μm for feldspars) were used. Data reduction of raw counts to correct weight % of oxides was done using the ZAF software program.

Bulk analyses of pure mineral separates of alkali feldspar and muscovite were obtained using conventional wet-chemical techniques for major elements at the Technical University of Nova Scotia (TUNS), Halifax. Trace-element concentrations, including those of the rare-earth elements (*REE*), were acquired using the ICP–MS facility at the Department of Earth Sciences, Memorial University, Newfoundland (Strong & Longerich 1985, Jenner *et al.* 1990, Longerich *et al.* 1990). The only exception to this pertains to the analyses of muscovite separates for Sn, which were obtained at TUNS (details below).

RESULTS

Plagioclase

Representative compositions of plagioclase are presented in Table 1. There is very little variation of the chemical data for coarse plagioclase grains, with a mean composition of about Ab_{99}, including data on core and rim (Table 1). Additional textural varieties of plagioclase are of similar composition, *e.g.*, (1) albite lamellae in perthitic alkali feldspar and (2) coarse patches of plagioclase in alkali feldspar, which are considered to represent remnant plagioclase grains replaced during a period of K-feldspathization.

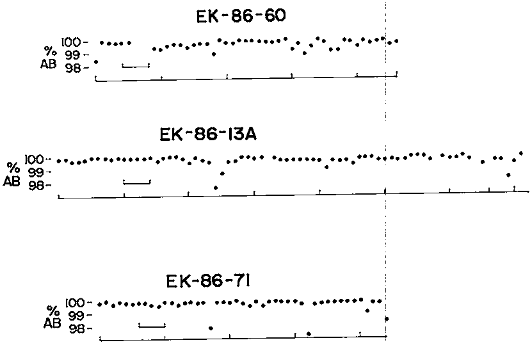

FIG. 1. Chemical profiles (in mole % Ab) of plagioclase grains progressing from margin (left side) toward core (right side). Analyses done in increments of 10 μm steps. Scale bar is 40 μm in each case. The few departures from the *ca.* Ab_{100} composition probably reflect the influence of microinclusions (*e.g.*, apatite) in the analysis.

Detailed traverses of three plagioclase grains (Fig. 1) illustrate the apparent lack of chemical zonation, and essentially flat profiles of Ab_{100} composition.

TABLE 2. REPRESENTATIVE COMPOSITIONS OF ALKALI FELDSPAR

	1	3	17	27	31	35	41	47	52	53
SiO_2	63.49	63.29	62.80	63.60	64.37	64.19	64.47	64.30	63.31	63.25
Al_2O_3	18.61	18.81	18.74	18.80	19.21	18.57	18.68	18.80	18.84	19.09
FeO	0.18	0.18	0.25	0.20	0.20	0.19	0.18	0.22	0.19	0.19
MnO	0.07	0.07	0.15	0.12	0.09	0.06	0.08	0.11	0.07	0.07
CaO	0.10	0.10	0.13	0.09	0.09	0.11	0.11	0.12	0.11	0.10
Na_2O	0.43	0.42	0.45	0.55	0.50	0.39	0.42	0.30	0.34	0.32
K_2O	16.41	16.34	16.34	16.24	16.16	16.80	16.58	16.66	16.41	16.22
Σ wt.%	99.29	99.21	98.86	99.60	100.62	100.21	100.52	100.51	99.28	99.24
Structural Formulae Based on 8 Oxygen Atoms										
Si	2.961	2.952	2.948	2.956	2.952	2.967	2.968	2.964	2.953	2.946
Al	1.023	1.034	1.036	1.029	1.039	1.011	1.013	1.021	1.036	1.048
Fe	0.006	0.002	0.009	0.007	0.007	0.007	0.006	0.008	0.007	0.007
Mn	0.002	0.004	0.005	0.004	0.003	0.002	0.003	0.004	0.002	0.002
Ca	0.004	0.004	0.004	0.004	0.004	0.004	0.005	0.005	0.004	0.004
Na	0.038	0.037	0.040	0.048	0.044	0.035	0.036	0.026	0.030	0.028
K	0.976	0.973	0.979	0.964	0.946	0.991	0.974	0.980	0.976	0.964
End-Member Feldspar, Mole %										
Ab	3.7	3.6	3.9	4.7	4.4	3.4	3.5	2.5	2.9	2.8
Or	95.8	95.6	95.6	94.8	95.1	96.2	95.9	96.9	96.0	96.7
An	0.5	0.8	0.5	0.5	0.5	0.4	0.6	0.6	1.1	0.5

#3 represents replacement-type alkali feldspar

TABLE 3. CHEMICAL COMPOSITION OF MINERAL SEPARATES

	Alkali Feldspar			Muscovite			
	EK-86-23C	EK-86-161	EK-86-1075b	EK-86-23C	EK-86-132	EK-86-161	EK-86-1075b
SiO_2	64.25	64.12	64.21	47.54	48.45	49.65	49.85
TiO_2	-	-	-	0.34	0.12	0.38	0.37
Al_2O_3	19.21	19.20	19.44	25.91	31.27	26.27	27.96
Fe_2O_3	-	-	-	0.79	1.27	0.31	0.59
FeO	-	-	-	7.39	3.40	7.73	5.93
MgO	-	-	-	0.25	0.27	0.22	0.25
CaO	0.26	0.15	0.34	-	-	-	-
Na_2O	1.81	1.60	1.81	0.40	0.66	0.61	0.47
K_2O	13.02	13.10	13.13	10.10	10.40	9.78	10.22
Σ wt.%	98.55	98.17	98.93	92.72	95.84	94.95	95.64
	Structural Formulae Based on 8 O Atoms			Structural Formulae Based on 22 O Atoms			
Si	2.974	2.977	2.964	6.718	6.486	6.814	6.742
${}^{IV}Al$	1.049	1.051	1.058	1.282	1.514	1.186	1.258
${}^{VI}Al$	-	-	-	0.034	3.422	3.065	3.200
Ti	-	-	-	0.036	0.012	0.039	0.038
Fe^{3+}	-	-	-	0.084	0.128	0.032	0.060
Fe^{2+}	-	-	-	0.873	0.381	0.887	0.671
Mg	-	-	-	0.053	0.054	0.045	0.050
Ca	0.013	0.007	0.016	-	-	-	-
Na	0.162	0.144	0.162	0.110	0.171	0.162	0.123
K	0.769	0.776	0.773	1.821	1.776	1.712	1.763
	End-Member Feldspar, Mole %						
Or	81.4	83.7	81.2				
Ab	17.2	15.5	17.0				
An	1.4	0.8	1.8				

Alkali feldspar

Representative compositions of alkali feldspar are presented in Table 2. The average composition of apparently unexsolved parts of grains is Or_{96}; this is independent of both grain size and volume of perthite. In addition, no apparent chemical difference exists between different textural varieties of alkali feldspar. For example, grains that show textural evidence of having replaced plagioclase are similar compositionally to those of non-replacement origin. Bulk compositions of three alkali feldspar separates (Table 3) from granite samples containing a minimal extent of K-feldspathization indicate an average composition of $Or_{82}Ab_{17}An_1$.

Trace-element concentrations in a single alkali feldspar separate are presented in Tables 4 and 5, and its chondrite-normalized profile is shown in Figure 2. Although it is difficult to interpret such data in isolation, comparison to a much larger dataset for alkali feldspar from the eastern part of the South Mountain Batholith (Kontak 1988a, Kontak & Strong 1988, Kontak, unpubl. data) permits some conclusions: (1) the Rb content (1729 ppm) is similar to concentrations found in pegmatites rather than granites (*sensu lato*), (2) the Sr content (139 ppm) is extremely anomalous and is more typical of concentrations found in alkali feldspar from granodiorites and primitive monzogranites. However, it should be noted that elevated Sr contents of whole-rock samples are characteristic of the East Kemptville leucogranite (Kontak 1990), and (3) Cs, Li, Pb and Ba contents are typical of pegmatitic alkali feldspar in the South Mountain Batholith. Comparison to data for alkali feldspar not from the South Mountain Batholith also indicates an evolved or fractionated trace-element chemistry. For example, the K/Rb ratio (Table 4) is comparable to values reported by

TABLE 4. TRACE-ELEMENT DATA FOR ALKALI FELDSPAR AND MUSCOVITE

	ALKALI FELDSPAR		MUSCOVITE							
	EK-23C	NS-18	EK-23C	EK-132	EK-1075	EK-161	EK-13A	EK-98	EK-116	EK-26
(ppm)										
Li	103	57.2	7386	947	3814	8458	5543	1460	2689	940
Be	6.6	17.3	11.3	17.8	33.5	23.3	24.0	20.8	14.2	-
Sc	0.5	-	10.0	7.3	15.0	15.7	16.1	16.1	1.5.	3.0
Rb	1729	1810	6142	1811	3226	3101	5499	2714	4467	2491
Sr	139	4.8	50.1	29.3	16.0	13.1	35.0	17.8	4.1	65.8
Y	0.7	0.1	7.9	2.8	5.2	5.5	2.2	5.4	0.9	2.0
Zr	10.9	6.1	37.2	22.2	21.1	29.2	32.3	11.5	15.2	27.1
Nb	2.7	-	153	86.3	126	84.6	129	106	33.9	42.7
Cs	23.3	14.4	354	119	250	717	388	186	237	74.6
Ba	29.6	36.0	7.3	157	22.5	7.9	38.5	59.7	62.1	172
Hf	0.6	-	2.7	1.6	1.43	1.9	2.3	0.6	0.9	1.9
Ta	1.3	-	45.3	22.2	13.5	9.2	31.9	9.3	7.1	8.5
W	19.5	-	82.3	91.9	66.9	51.6	83.7	81.3	72.0	58.0
Pb	15.9	11.1	10.9	5.4	4.7	3.8	6.0	10.5	2.5	5.1
Th	0.4	0.2	9.1	3.3	6.9	5.5	8.2	4.4	2.4	4.1
U	6.0	1.7	36.9	12.3	15.1	17.5	9.5	23.0	9.4	29.8
Tl	12.0	11.7	-	11.1	-	23.7	-	4.9	-	-
Rb/Sr	12.4	37.7	122	61.8	201	236	157	152	1089	37.8
K/Rb	50.3	48.1	14.1	48.1	27.0	28.0	15.8	32.1	19.5	34.9
U/Th	15.0	8.5	4.0	3.7	2.1	3.8	1.1	5.2	3.9	7.2
Zr/Hf	18.1	-	13.7	13.8	14.7	15.3	14.0	19.1	16.8	14.2
Ta/W	0.06	-	0.55	0.24	20.1	0.17	0.38	0.11	0.09	0.14
Ta/Cs	0.05	-	0.12	0.18	0.05	0.01	0.08	0.05	0.02	0.11
K/Cs	3737	6047	246	732	348	121	224	468	367	1161

Note that EK-98 is a pegmatite-aplite and EK-116, 26 are greisens.

Shearer *et al.* (1987) for pegmatitic feldspar from the Black Hills, South Dakota, whereas the Li content is similar to values Černý *et al.* (1985) reported for feldspar from a variety of localities for rare-element granitic pegmatites. The elevated contents of elements such as W, U, Zr and Nb may be suspect because of the possibility of microinclusions of accessory phases.

The *REE* profile indicates low absolute totals and an overall flat, unfractionated pattern that is similar in shape, although not absolute abundance, to that of the whole-rock chondrite-normalized pattern (Kontak 1990). The pattern is somewhat similar to chondrite-normalized profiles for alkali feldspar in pegmatites within the South Mountain Batholith (see Fig. 2), although these have slightly fractionated patterns [$3 < (La/Lu)_N <$]. The absence of a positive Eu anomaly is not unusual for alkali feldspar in pegmatites from the South Mountain Batholith, which has similarly low absolute concentrations of *REE* (Kontak 1988a, unpubl. data).

TABLE 5. RARE-EARTH-ELEMENT DATA FOR ALKALI FELDSPAR AND MUSCOVITE

(ppm)	EK-23C(K)	EK-23C	EK-132	EK-1075	EK-161	EK-13A	EK-98
La	0.22	2.40	1.12	2.56	1.57	1.77	2.09
Ce	0.40	7.23	2.51	5.77	4.52	4.59	4.47
Pr	0.03	0.93	0.31	0.76	0.57	0.57	0.99
Nd	0.13	3.31	1.06	2.57	2.04	2.05	2.93
Sm	0.06	1.43	0.46	0.94	0.89	0.79	1.76
Eu	ND	0.01	0.07	0.02	ND	ND	0.16
Gd	0.09	1.48	0.45	0.95	0.88	0.54	0.94
Tb	0.02	0.37	0.12	0.23	0.23	0.12	0.23
Dy	0.14	1.96	0.67	1.32	1.23	0.57	1.34
Ho	0.01	0.25	0.10	0.17	0.14	0.07	0.22
Er	0.04	0.50	0.28	0.41	0.32	0.14	1.21
Tm	0.01	0.06	0.04	0.05	0.03	0.02	0.19
Yb	0.06	0.39	0.21	0.29	0.23	0.13	0.47
Lu	ND	0.04	0.02	0.03	0.02	0.01	0.06

Sample EK-23C(K) is an alkali feldspar, remainder are muscovites. All samples from leucogranite except for 98 (aplite-pegmatite)
ND = not detected

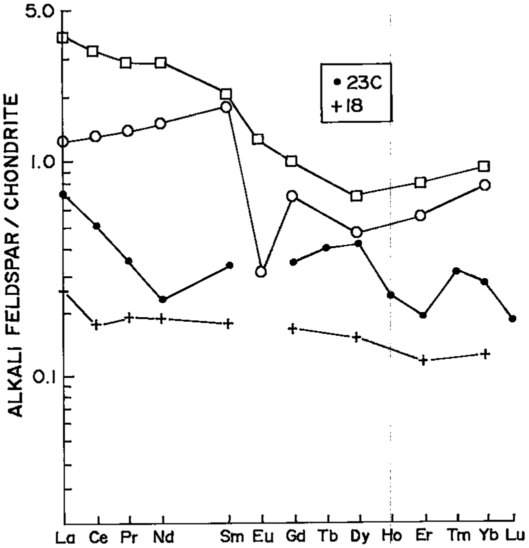

FIG. 2. Chondrite-normalized profiles for samples of alkali feldspar from East Kemptville leucogranite (23C; Table 5) and pegmatites of the South Mountain Batholith [from Kontak (1988a) and unpubl. data of the author].

Biotite

It should first be noted that biotite occurs only as a trace constituent in the East Kemptville leucogranite, generally as small (0.05 mm) grains within quartz (Kontak 1990). Although the biotite grains exhibit typical pleochroism and birefringence, optical observations clearly indicate alteration to other phases. Some of the alteration products are chlorite and an unidentified clay (?) mineral; representative compositions of these are presented in Table 6. Note that because the alteration of biotite may occur on a fine scale (*i.e.*, a few tens of Å or less: Ahn & Peacor 1987, Eggleton & Banfield 1985), the composition of the alteration phase may represent a mixture of distinct phases unresolved at the micrometer scale of analysis.

The chlorite is Fe-rich [Fe/(Fe+Mg) = 0.99], reflecting the composition of its precursor (*e.g.*, Eggleton & Banfield 1985), and corresponds to the delessite field in the classification scheme of Hey (1954).

The biotite (representative compositions in Table 7) is iron-rich, with 13–25 wt.% FeO (a single exception is represented by analysis 13 in Table 7; see below) and a mean Fe/(Fe+Mg) value of 0.96 ± 0.014 ($n = 15$). Contents of the more refractory elements are typically low: Mg (<0.66 wt.% MgO), Ti (<1.19 wt.% TiO_2) and Mn (<0.40 wt.% MnO); trace amounts of Ca, Na and Cr occur. Compared to biotite compositions in the South Mountain Batholith (Allan & Clarke 1981, Kontak & Corey 1988, Corey 1988) and the Davis Lake Pluton (Chatterjee *et al.* 1985), the biotite is characterized by elevated Fe, lower Mg and Ti, and highly variable Si. In Figure 3, binary plots of ^{IV}Al *versus* Fe/(Fe+Mg) and ^{VI}Al *versus* Ti, the highly unusual chemistry of the biotite is illustrated, particularly its enrichment in Fe, depletion in ^{IV}Al (*i.e.*, reflecting enrichment in silica) and highly evolved nature in terms of Ti content. Of particular importance is the marked divergence from the suite of biotite samples analyzed from granitic rocks of the Davis Lake Pluton (Fig. 3), considered by some (*e.g.*, Richardson 1988a, b, Richardson *et al.* 1989) to be parental to the East Kemptville leucogranite.

TABLE 6. REPRESENTATIVE COMPOSITION OF CHLORITE AND ALTERATION PHASE

	15	16	14	10	17	18	8
SiO_2	38.84	43.41	43.68	46.42	42.64	44.49	28.29
TiO_2	0.07	0.06	0.10	0.05	0.06	0.06	0.12
Al_2O_3	27.92	32.35	31.97	36.31	33.65	33.66	17.21
Cr_2O_3	0.06	0.06	0.09	0.05	0.00	0.05	0.11
FeO	12.08	6.99	9.07	5.12	8.66	8.29	37.68
MnO	0.26	0.26	0.33	0.21	0.19	0.17	1.08
MgO	0.15	0.15	0.21	0.11	0.08	0.14	0.18
CaO	0.35	0.30	0.28	0.14	0.13	0.23	0.32
Na_2O	0.28	0.15	0.11	0.10	0.09	0.21	0.17
K_2O	0.09	0.12	0.25	0.04	0.05	0.12	0.11
Σ	80.10	83.85	86.09	88.54	85.56	87.42	85.28
	Structural Formulae based on 28 Oxygen Atoms						
Si	7.852	8.048	7.999	8.019	7.820	7.962	6.567
^{IV}Al	0.148	0.000	0.001	0.000	0.180	0.038	1.133
^{VI}Al	5.082	6.612	6.902	7.394	7.096	7.062	3.578
Ti	0.010	0.008	0.014	0.006	0.008	0.007	0.020
Cr	0.010	0.008	0.012	0.006	0.000	0.007	0.020
Fe^{2+}	2.042	1.084	1.388	0.739	1.328	1.240	7.318
Mn	0.044	0.040	0.050	0.030	0.029	0.025	0.212
Mg	0.045	0.042	0.057	0.028	0.021	0.036	0.062
Ca	0.076	0.059	0.054	0.025	0.025	0.044	0.080
Na	0.109	0.053	0.039	0.025	0.031	0.072	0.076
K	0.022	0.028	0.058	0.008	0.011	0.028	0.033

Chlorite: 8; Unidentified alteration phases of biotite: 15, 16, 14, 10, 17, 18

TABLE 7 REPRESENTATIVE COMPOSITIONS OF BIOTITE

	EK-71	EK-91			EK-23A		
	2	5	6	13	26	28	20
SiO_2	39.20	39.12	44.94	48.11	39.25	41.33	46.99
TiO_2	0.72	1.14	0.39	0.11	1.05	0.94	0.34
Al_2O_3	21.40	18.70	20.06	27.57	21.39	21.52	20.50
Cr_2O_3	0.12	0.11	0.09	0.07	0.12	0.12	0.08
FeO	18.36	24.99	15.54	7.48	21.69	19.91	13.87
MnO	0.26	0.40	0.35	0.21	0.20	0.19	0.21
MgO	0.50	0.39	0.37	0.32	0.62	0.66	0.32
CaO	0.10	0.13	0.10	0.20	0.11	0.09	0.09
Na_2O	0.13	0.15	0.11	0.20	0.16	0.18	0.20
K_2O	9.38	9.54	10.04	9.15	9.84	10.02	10.24
F	2.84	2.21	3.19	1.34	2.84	3.26	4.47
	93.00	96.86	95.18	94.76	97.25	98.20	97.32
F=O	1.19	0.92	1.33	0.56	1.19	1.36	1.87
Σ wt.%	91.81	95.94	93.75	94.20	96.06	96.84	95.45
	Structural Formulae Based on 22 Oxygen Atoms						
Si	5.985	5.953	6.534	6.603	5.835	5.999	6.433
^{IV}Al	2.015	2.047	1.396	1.397	2.165	2.001	1.567
^{VI}Al	1.828	1.391	3.064	3.063	1.584	1.682	1.738
Ti	0.082	0.130	0.042	0.011	0.117	0.102	0.035
Cr	0.014	0.013	0.009	0.007	0.014	0.013	0.008
Fe	2.344	3.180	1.889	0.858	2.696	2.418	1.587
Mn	0.033	0.050	0.042	0.024	0.024	0.023	0.024
Mg	0.063	0.049	0.044	0.036	0.077	0.080	0.036
Ca	0.015	0.023	0.001	0.029	0.025	0.027	0.029
Na	0.038	0.044	0.030	0.052	0.050	0.050	0.052
K	1.825	1.850	1.861	1.601	1.865	1.855	1.788
F	1.367	1.063	1.466	0.581	1.335	1.496	1.935

The biotite compositions have a large deficiency in the *Y* sites (4.177 to 5.063 based on 22 O p.f.u.) compared to the ideal value (6) in end-member trioctahedral micas (Fig. 4). In magmatic biotite and muscovite, the total cationic abundances are 16 and 14, respectively (for 22 O p.f.u.); the deficiency for the biotite and degree of muscovite substitution are clearly illustrated. The trends in this figure correspond to the following substitutions suggested by Konings *et al.* (1988):

$$3R^{3+} = 2R^{3+} + \square \qquad (1)$$

$$2T^{3+} + R^{2+} = 2\mathrm{Si} + \square \qquad (2)$$

where □ represents a vacancy in the octahedral or R sites, R^{2+} = (Mn, Mg, Fe), T^{3+} = ^{IV}Al, and R^{3+} = (^{VI}Al, Fe). The large range in Si values for the biotite suggests that substitution (2) is probably dominant.

The degree of substitution toward muscovite also may be evaluated using the triangular Al–Si–M^{2+} plot of Monier & Robert (1986; Fig. 5). The deviation of the data from the muscovite – (phlogopite – annite) join toward phengite is consistent with substitution (2) referred to above, involving the tetrahedral sites. The degree to which the substitution toward muscovite has modified primary biotite (analysis #4 in Table 7) is represented by analysis 13 in Table 7 and is highlighted in Figures 3, 4, 5 and 6; this composition is not unlike that of some muscovite found in the leucogranite (Fig. 3 and see below). In the triangular (Al–Si–M^{2+}) plot of Monier & Robert (1986), the data points fall well outside even the low-temperature isotherms (see Monier & Robert 1986), suggesting therefore that postmagmatic processes are responsible for the highly variable chemistry of the biotite.

The coherent chemistry of individual grains of biotite is illustrated in Figure 6, where binary plots of FeO *versus* SiO_2 and TiO_2 demonstrate that the chemical variation within single grains is minimal. Note, however, that groups 1, 2 and 3 in Figure 6 refer to separate grains in the same sample; hence, the degree of re-equilibration even on a thin section scale is not homogeneous.

The F contents of the biotite vary from a low of 2.21 wt.% to a high of 4.47 wt.%; there is a strong inverse correlation with FeO (Fig. 6). Note that the sole exception to the trend is represented by analysis #3 (Fig. 6), which has been referred to previously because of its apparently anomalous chemistry. Since there is an apparent increase of F content with greater substitution toward muscovite, the best estimate of primary F concentration is based on those samples that show minimal amounts of substitution toward muscovite (#5 in Table 7); a value of 2.72 ± 0.25 wt.% F ($n = 5$) is calculated using the appropriate analyses. It should be noted that because of the Fe–F avoidance principle, the apparent trend of increasing F content in biotite, which may record equilibration to lower temperatures, does not necessarily reflect increasing activity of F in the fluid.

Muscovite

Muscovite (representative compositions in Table 8) is characterized by its variable color in hand specimen. The most common variety has a faint brownish or amber tone. In thin section, a subtle pleochroism (light brownish) is present, which generally correlates with high iron content. The elevated iron content distinguishes this muscovite from that in the South

Mountain Batholith (*cf.* Ham & Kontak 1988). Muscovite of inferred primary origin in the South Mountain Batholith contains about 45 wt.% SiO_2, 35 wt.% Al_2O_3 and <2 wt.% FeO, which contrasts with the representative compositions in Table 8. Although fine grained, clearly secondary muscovite occurs in some samples; the data reported herein pertain to coarser-grained mica of inferred primary origin.

The wt.% FeO contents of muscovite vary from about 7 to 10 wt.%, with some samples showing extreme enrichment to 12 wt. % (Fig. 7). In the same diagram, data for greisen muscovite (unpubl. data of Kontak) from East Kemptville are shown for comparative purposes. The figure shows relatively good discrimination for the two paragenetic types on the basis of wt.% FeO and Si (atomic proportions). Analyses of muscovite separates (Table 3) indicate that this phase is characterized by very low Fe^{3+}/Fe^{2+} ratios of about <0.1, values typical of S-type or strongly peraluminous granites (Neiva 1982, 1987).

The elevated iron contents of the muscovite are unusual for granites (*sensu lato*) in general (*e.g.*, Miller *et al.* 1981, Anderson & Rowley 1981, Clarke 1981, Neiva 1982, 1987), although this is partly a reflection of the absence of ferromagnesian minerals (*e.g.*, significant biotite, garnet) in the leucogranite such that Fe does not have to be partitioned among coexisting phases. This latter point was emphasized, for example, by Guidotti (1978a, b), who showed that the celadonitic component of muscovite changed with mineral paragenesis in both metamorphic and igneous (*i.e.*, granitic) rocks (see also Anderson & Rowley 1981 and Miller *et al.* 1981).

The magnitude of the iron enrichment noted herein is comparable to that in zinnwaldite analyzed by Stone *et al.* (1988) from Cornish granites, although their analyses contain 2.6–4.0 wt.% Li_2O, well in excess of the *ca.* 0.5 wt.% for the East Kemptville muscovite (see below). In addition, compositions of zinnwaldite from rare-element pegmatites in Colorado and India (Hawthorne & Černý 1982, their Table 8) are similar to those of the East Kemptville muscovite with respect to most major elements except for the enrichment in Li. Muscovite from other environments that show similar iron enrichment include phengite in high-pressure metamorphosed granitic orthogneiss from the Seward Peninsula, Alaska (Evans & Patrick 1987), and some phengite in metamorphosed Triassic volcanic rocks from western Greece (Pe-Piper 1985). These latter two studies also are important in emphasizing muscovite composition as a function of the host-rock composition, mineral assemblage and attendant *P–T* conditions during growth.

The iron contents of the muscovite show strong positive and negative correlations with, respectively, Si (Fig. 8) and ^{VI}Al (Fig. 9); a poor correlation is noted between the remaining M^{2+} cations and Si (*e.g.*, Ti in Fig. 8). These correlations suggest that

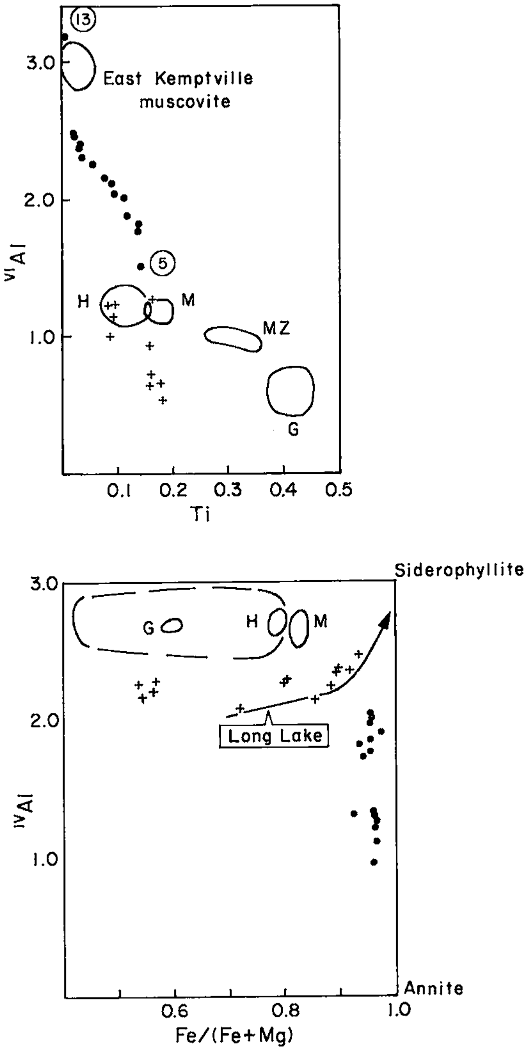

FIG. 3. Chemical data for biotite plotted in ^{VI}Al *versus* Ti and ^{IV}Al *versus* Fe/(Fe + Mg) plots. Results for biotite from the East Kemptville leucogranite (•) are compared to fields for biotite from (1) the Big Indian Lake polyphase intrusive complex and associated alteration (G: granodiorite, MZ: monzogranite, M: microgranite, H: high-alumina alteration zone; data from Corey 1988, Kontak & Corey 1988), (2) peraluminous granites in general (dashed lines in the lower plot; from Clarke 1981), (3) Long Lake area of the South Mountain Batholith (from Kontak & Corey 1988) and (4) leucomonzogranites of the Davis Lake complex (+; data from Chatterjee *et al.* 1985). Also shown in the top figure is the field for muscovite compositions from the East Kemptville leucogranite (data from this study). Analyses #5 and #13 are discussed in the text, and the data are given in Table 7.

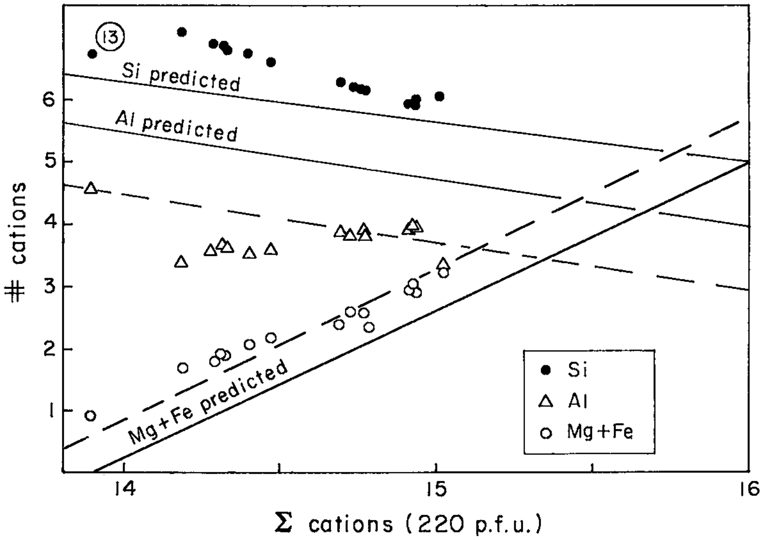

FIG. 4. Biotite data plotted as cations (Si, Al, Fe+Mg) *versus* total cations (based on 22 O p.f.u.). Diagram adapted from Konings *et al.* (1988). The solid lines refer to the exchange predicted in compositionally ideal magmatic biotite (see Konings *et al.* 1988 for discussion). #13 refers to analysis in Table 7 and is discussed in the text.

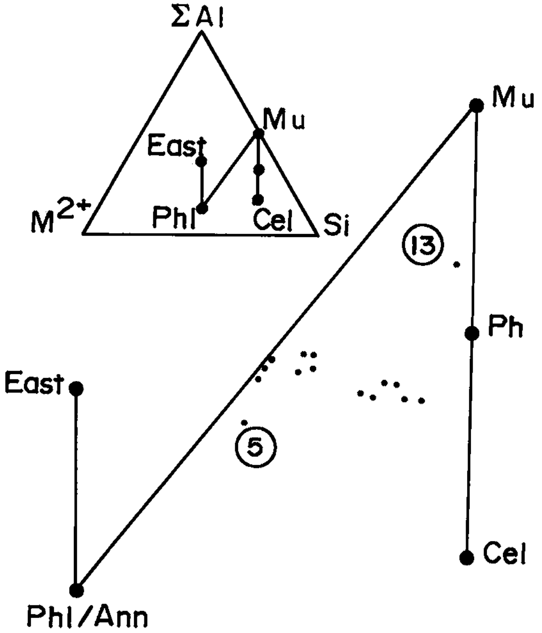

FIG. 5. Biotite data plotted in the triangular diagram of Monier & Robert (1986), showing deviation of the compositions from ideal end-member biotite due to substitution toward muscovite. Numbers (5, 13) refer to data in Table 7 and are discussed in the text.

the dominant mechanism of substitution is represented by the phengite series of Monier & Robert (1986), in which the following exchange operates:

$$^{VI}Al,\ ^{IV}Al = (M^{2+}),\ ^{IV}Si \qquad (3)$$

where Fe is the major M^{2+} cation involved in this case. Analysis of muscovite separates (Table 3) indicates that most of the iron is in fact present in the Fe^{2+} state. Individual muscovite compositions from five samples of leucogranite are shown separately in Figure 10, a plot of Si-Al-M^{2+} adapted from Monier & Robert (1986) in order to illustrate this mechanism of substitution. As the data indicate, there is deviation of the chemistry from pure end-member muscovite. Although considerable substitution toward celadonite occurs, there is an additional substitution that causes a shift from the muscovite-celadonite tie line. This second substitution, which is referred to as "biotitic" by Monier & Robert (1986) is described by these authors as follows:

$$\tfrac{2}{3}\,^{VI}Al,\ \tfrac{1}{3}\,^{VI}\square = \,^{VI}(M^{2+}) \qquad (4)$$

Because of the very low Fe^{3+}/Fe^{2+} ratios of the muscovite (Table 3), the deviation of data points from the muscovite-celadonite join cannot be attributed to overestimation of M^{2+} site occupancy.

Thus, the apparent inference that significant substitution toward biotite occurs in muscovite appears to be real.

The F content of the muscovite is highly variable, between about 1.0 and 4.2 wt.%. In Figure 11, the data have been plotted as a function of Fe/(Fe + Mg) ratio and grouped according to the sample analyzed; it appears that for any given sample, a restricted range occurs. Sample EK-098 represents a pegmatite–aplite segregation, and EK-105 has an anomalous whole-rock chemistry (see Kontak 1990); hence, the remaining data for samples EK-69 and EK-23C are considered to give the best approximation of the F content of the muscovite. The F contents are comparable to those reported for muscovite from highly evolved magmatic systems, such as represented by the host granite at the Henderson molybdenite deposit, Colorado (Gunow *et al.* 1980), evolved phases of the Bob Ingersoll pegmatite, South Dakota (Jolliff *et al.* 1987), zinnwaldite from Cornwall (Stone *et al.* 1988), and rare-element pegmatites (Hawthorne & Černý 1982). In contrast, the F contents of muscovite from even the most evolved phases of the South Mountain Batholith rarely attain such elevated values (Ham & Kontak 1988). The F content of muscovite in the leucogranite and greisens also is compared in Figure 11. Except for sample EK-98, an aplite–pegmatite segregation, there is a distinct enrichment of F in granite-hosted *versus* greisen-hosted muscovite.

Chlorine contents of representative samples of muscovite are in the range ≤ 0.05 wt.%. There is no consistent relationship observed between Cl and F or any of the cations (*e.g.*, Fe content).

Trace-element contents, including *REE* data, have been obtained for six muscovite separates (5 leucogranite, 1 pegmatite–aplite), and results are presented in Tables 4 and 5. For many of the elements considered, there remains the problem of what proportion of the elemental abundance may be contributed by microinclusions. Although it is difficult to assess this problem without direct, comparative microanalysis of the same material, note that combined qualitative EDS electron-microprobe analyses and SEM studies have confirmed the presence of small (*i.e.*, generally <30–50 μm) inclusions of zircon, triplite, apatite, rutile, cassiterite, Nb–Ta oxide, uraninite, Fe-rich wolframite (*i.e.*, ferberite), sphalerite and Fe–Cu sulfide phases, albeit in very small proportions. As discussed by Kontak (1990), it is not possible to unequivocally assign a magmatic, orthomagmatic or hydrothermal origin to these phases, although the euhedral morphology and absence of alteration of the muscovite host (*e.g.*, rutilization, recrystallization) would be consistent with a magmatic origin. Hence, it is with a certain degree of caution that the concentrations of some elements must be interpreted when discussed in isolation or when compared to data from other areas.

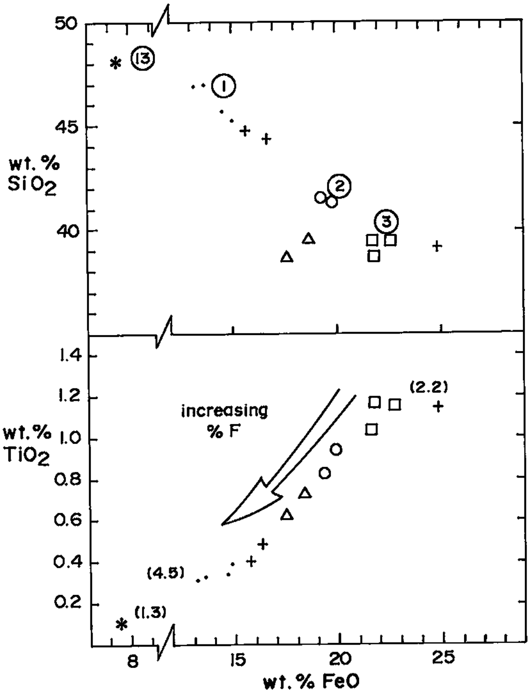

FIG. 6. Biotite data plotted as a function of wt.% SiO_2 and TiO_2 *versus* wt.% FeO (total iron). Sample #13 refers to data in Table 7 (discussed in the text), whereas numbers 1, 2, 3 refer to groupings of multiple analyses of single grains in the same polished section (see text for discussion). Numbers in the lower figure refer to wt.% F in the biotite.

TABLE 8. REPRESENTATIVE COMPOSITIONS OF MUSCOVITE

	EK-23C		EK-98		EK-69		TM-86-1	EK-105
SiO_2	50.84	48.47	49.31	48.57	49.43	48.79	48.55	48.82
TiO_2	0.08	0.13	0.00	0.24	0.09	0.21	0.25	0.20
Al_2O_3	24.18	24.90	30.22	23.93	24.59	23.57	24.71	28.87
FeO	8.50	9.96	6.99	8.63	8.08	10.74	10.99	6.48
MnO	0.09	0.00	0.19	0.08	0.11	0.00	0.00	0.30
MgO	0.24	0.25	0.16	0.28	0.31	0.28	0.33	0.26
CaO	0.00	0.03	0.01	0.00	0.00	0.004	0.00	0.01
Na_2O	0.05	0.14	0.21	0.15	0.14	0.11	0.13	0.17
K_2O	10.69	11.36	10.96	11.00	11.11	11.14	10.82	10.40
F	3.28	3.67	1.43	4.31	3.37	NA	NA	2.01
	97.95	98.91	99.48	97.19	97.22	94.88	95.83	97.52
F≡O	1.37	1.54	0.60	1.81	1.41	-	-	0.84
Σ wt.%	96.58	97.37	98.88	95.38	95.81	94.88	95.83	96.68
Structural Formulae Based on 22 Oxygen Atoms								
Si	6.768	6.542	6.469	6.620	6.701	6.872	6.765	6.507
^{IV}Al	1.232	1.458	1.531	1.380	1.299	1.128	1.235	1.493
^{VI}Al	2.556	2.500	3.141	2.465	2.628	2.784	2.825	3.042
Ti	0.008	0.012	0.000	0.024	0.008	0.022	0.025	0.020
Fe	0.944	1.119	0.764	0.983	0.912	1.261	1.273	0.721
Mn	0.008	0.000	0.020	0.009	0.012	0.000	0.000	0.033
Mg	0.047	0.050	0.030	0.056	0.061	0.058	0.067	0.051
Ca	0.000	0.004	0.000	0.000	0.000	0.005	0.000	0.000
Na	0.012	0.035	0.052	0.039	0.034	0.028	0.048	0.043
K	1.814	1.946	1.829	1.900	1.921	2.002	1.923	1.772
F	1.376	1.565	0.591	1.851	1.440	0.000	0.000	0.847

The muscovite is characterized by elevated absolute contents of Li, Rb, Be, Ta, W, U and Tl, and extremely low K/Rb and K/Cs ratios. Direct com-

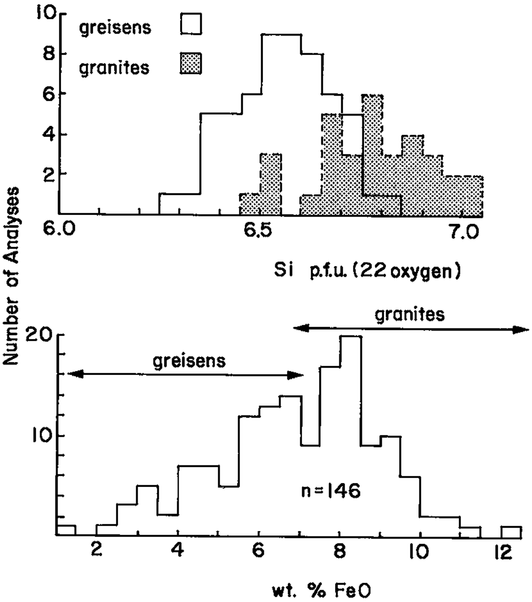

FIG. 7. Histograms of Si (cations p.f.u. based on 22 O) and wt.% FeO (total iron) for muscovite from the East Kemptville leucogranite compared to muscovite from greisens at the same locality (unpubl. data of the author). Note that there is no overlap in terms of wt.% FeO between the two populations.

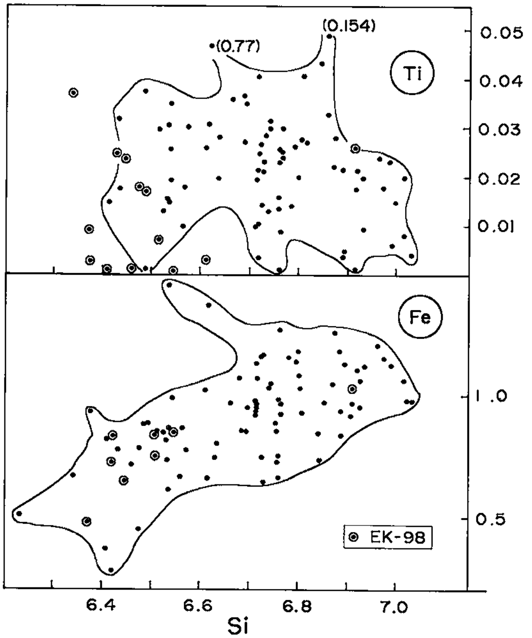

FIG. 8. Muscovite data plotted in terms of Ti and Fe *versus* Si. EK-98 represents muscovite from a pegmatite–aplite sample. Note the strong correlation between Fe and Si, which suggests a phengitic substitution.

parison to trace-element contents of muscovite from granitoid rocks and greisens of the South Mountain Batholith (Ham & Kontak 1988) reveal relatively strong to exceptionally higher Li, U, Th, Cs, Y and Sr contents, whereas Ba, Ta and Zr show comparable values, and Sc is lower.

In Figure 12, data for selected trace elements in muscovite are compared to data for muscovite hosted by granites, pegmatites–aplites and mineralized (Sn, W) quartz veins from a variety of localities. As the diagram illustrates, there is such a large range for individual elements that it is difficult to assign specific levels of enrichment or depletion for any element for a given host. This spread is interpreted to reflect three processes acting alone or in concert. First, within an individual magmatic or hydrothermal system, there is a progressive evolution that depends in part on the coexisting phases (Ham & Kontak 1988, Jolliff *et al.* 1987, Černý *et al.* 1985). In the absence of a similar paragenesis and mineral assemblage for the analyzed samples used to construct Figure 12, some variation is to be expected. Second, the primary elemental concentrations between individual magmatic or hydrothermal systems vary, such that absolute abundances will vary for muscovite from different areas (*e.g.*, data in Möller & Morteani 1987). And third, the potential problem of microinclusions must be considered as an unknown factor. With these potential caveats in mind, note that the East Kemptville muscovite has concentrations equivalent to or greater than those for granite-hosted muscovite for W, Nb, Ta, Sn and Cs, whereas the Ba contents are more comparable to levels found in magmatic rather than hydrothermal (*i.e.*, quartz-vein hosted) muscovite.

Tin contents of four granite- and two greisen-hosted muscovite separates are presented in Table 9. The analyses obtained are such that the difference between the ammonium iodide fusion (dissolves Sn in oxide and sulfide phases) and the sodium peroxide fusion (total dissolution of all phases) indicates the amount of tin that is structurally bound in the muscovite structure. Although there is a large variance and the data set is small, the results indicate that the granitic muscovite contains about 100 ppm structurally bound tin, a value considerably below that for greisen muscovite. The data are comparable to values given by Neiva (1982, 1987) for muscovite from a variety of granite-, aplite-, greisen- and quartz-vein-hosted muscovite from Portuguese Sn and W deposits. Unfortunately, there are no comparable data for muscovite from the South Mountain Batholith.

In Figure 13 some chemical parameters commonly used to assess the degree of evolution in pegmatite–aplite systems are examined for muscovite from East Kemptville and compared to the fields for data from the South Mountain Batholith and rare-element peg-

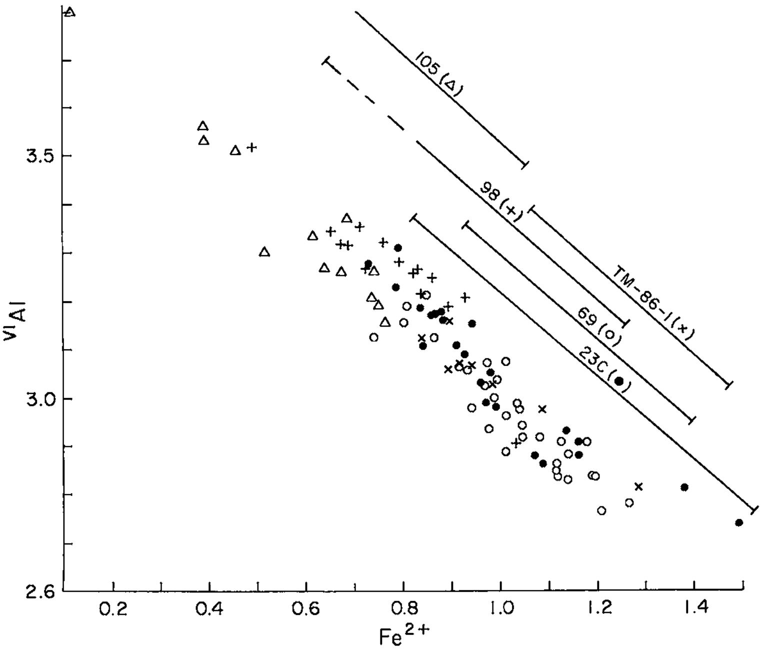

FIG. 9. Muscovite data plotted as a function of ^{VI}Al *versus* Fe (total iron). The range for muscovite compositions in individual samples is shown.

matites. Data for the East Kemptville muscovite overlap the fields as defined by Möller & Morteani (1987) for Ta-enriched pegmatites based on the covariance of K/Rb–Cs (Fig. 13a) and Ta–K/Cs (Fig. 13c). In addition, the data are compared to the fields defined for beryl + columbite- and spodumene-bearing pegmatites at Cross Lake, Manitoba, based on covariance of K/Rb *versus* K/Cs (Fig. 13b); data for the East Kemptville muscovite correspond to the former field. In addition, Rb/Tl values for the East Kemptville muscovite (Table 4) are comparable to those in muscovite from rare-metal pegmatites from Mongolia (Černý 1988). In all aforementioned diagrams, the East Kemptville muscovite generally represents more evolved compositions than (1) greisen muscovite from East Kemptville and (2) muscovite data for the South Mountain Batholith.

Chondrite-normalized patterns for muscovite from East Kemptville (Fig. 14) are characterized by flat *LREE* profiles, moderate overall fractionation [$(La/Lu)_N$ averages 5.9 + 2.3], negative Eu anomalies (Eu/Eu* varies from 0.02 to 0.5), and a distinct inflection in the *MREE* at Tb, with strongly fractionated *HREE* patterns. This latter observation contrasts markedly, for example, with the overall flat profiles for granites from highly evolved, LILE-enriched systems at Mount Pleasant (Taylor *et al.* 1985) and True Hill (Lentz *et al.* 1988), New Brunswick, and the Ackley Granite, Newfoundland (Tuach 1987). The muscovite patterns contrast markedly with those found for muscovite from a variety of granites (Arniaud *et al.* 1984, Neiva 1982, Alderton *et al.* 1980), including the South Mountain Batholith (Ham & Kontak 1988), which in general have higher absolute abundances (up to several hundred for the *LREE* in the case of the Auriat Granite, France: Arniaud *et al.* 1984) and strongly fractionated patterns [$(La/Lu)_N$ = 10]. In some cases, muscovite from mineralized vein samples from Portuguese Sn–W deposits (Neiva 1982) and greisens in the South Mountain Batholith (Ham & Kontak 1988) have similar *REE* abundances, but again they have much more strongly fractionated patterns.

Apatite

Apatite has not been identified as a primary magmatic phase in the leucogranite; it occurs instead as ubiquitous microinclusions (of secondary origin) in albitic plagioclase. The apatite (representative compositions in Table 10) is enriched in manganese (up to 2.53 wt.% MnO) and fluorine (up to 4.03 wt.% F); the highest manganese contents are found in the core of the grains. There is a strong correlation of Ca with Mn (Fig. 15), suggesting that the major substitution in this mineral is accommodated by direct exchange of these two cations. Based on a larger data-base ($n = 37$), O'Reilly (1988) found that for

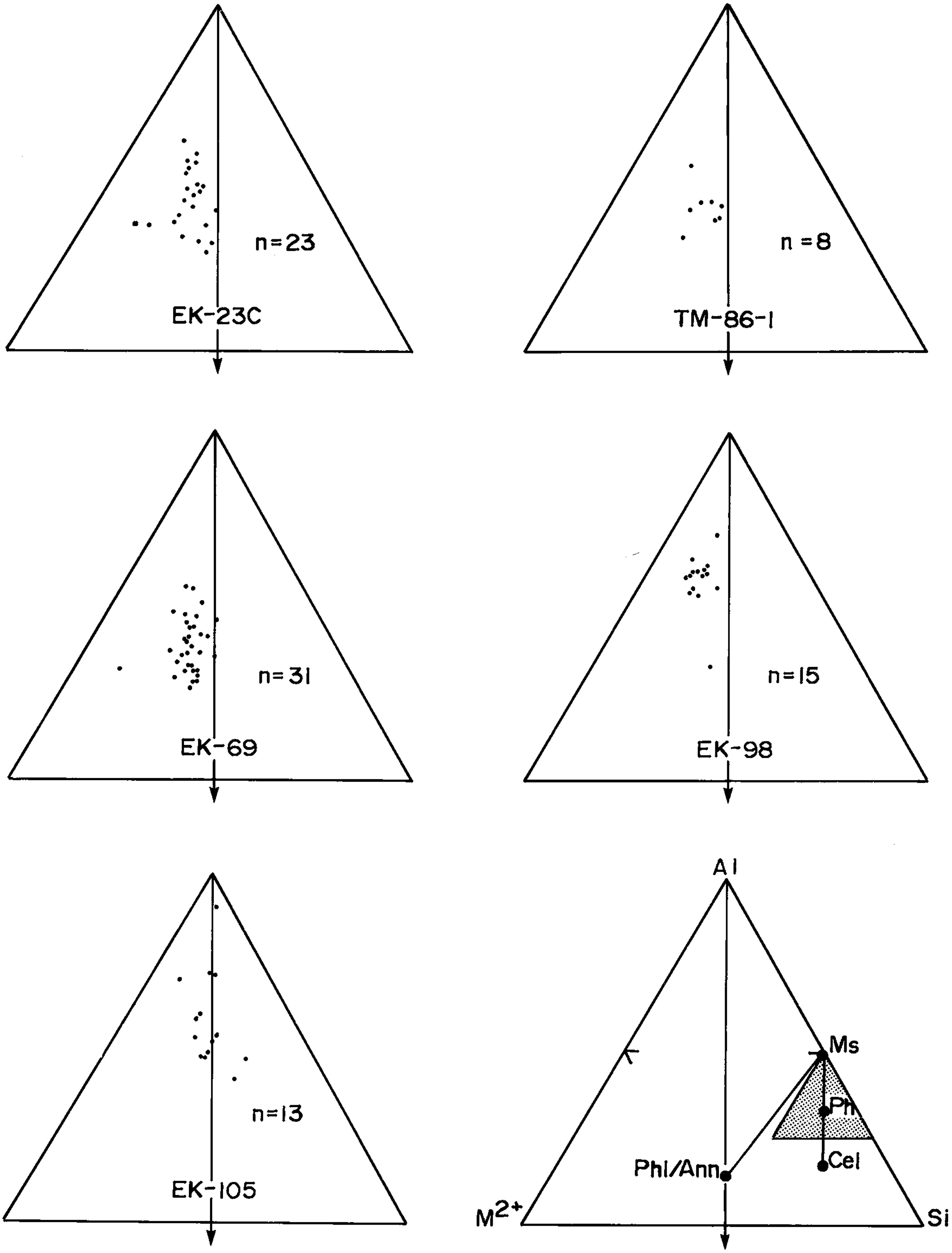

FIG. 10. Muscovite data from individual samples plotted in the triangular diagram of Monier & Robert (1986). Note the exchange along the muscovite – phengite + celadonite tie line and apparent deviation because of variable amounts of substitution toward biotite. The muscovite–celadonite tie line is shown in each triangular plot.

secondary apatite (containing up to 7 wt.% MnO) in the Devonian Sangster Lake pluton of the eastern Meguma Terrane, the same exchange had a Spearman Rank Correlation Coefficient of 0.928. The data reported here essentially correspond to the same trend (inset in Fig. 15).

Although the manganese contents are high compared to data reported by Roeder *et al.* (1987) and Pichavant *et al.* (1988) for apatite representing a wide variety of occurrences, including hydrothermal veins and highly evolved felsic melts, comparable or higher levels of manganese enrichment have been reported in apatite from pegmatites (*e.g.*, Jolliff *et al.* 1989).

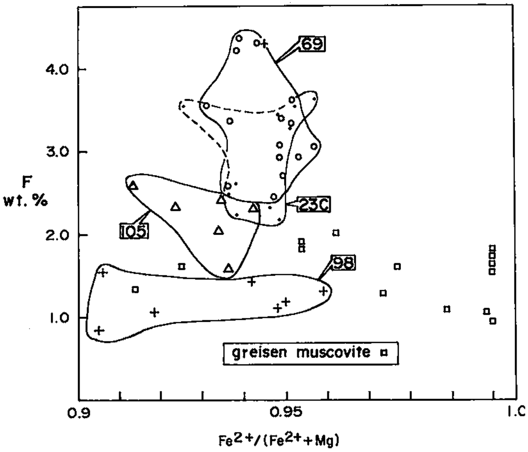

FIG. 11. Muscovite data plotted as a function of wt.% F *versus* Fe/(Fe + Mg). Muscovite from greisens represent unpublished data of the author.

Topaz

Core and rim compositions of topaz (representative compositions in Table 11) have been determined for both granite and greisen host-rock; no chemical variation was detected; average F contents for granitic- and greisen-hosted topaz are 16.47 ± 0.98 ($n = 14$) and 16.72 ± 0.66 ($n = 6$), respectively. The results of the analyses reported are somewhat at variance with the results of Beneteau & Richardson (1989), who reported F contents of 17.3–20.6 wt.%; the reason for the discrepancy remains unexplained, but it is noted that A.K. Chatterjee (pers. comm., 1989) obtained topaz compositions for material from East Kemptville that are similar to those reported here. The topaz described here is similar chemically to topaz from the St. Austell granite reported by Manning & Exley (1984), whereas topaz from topaz–

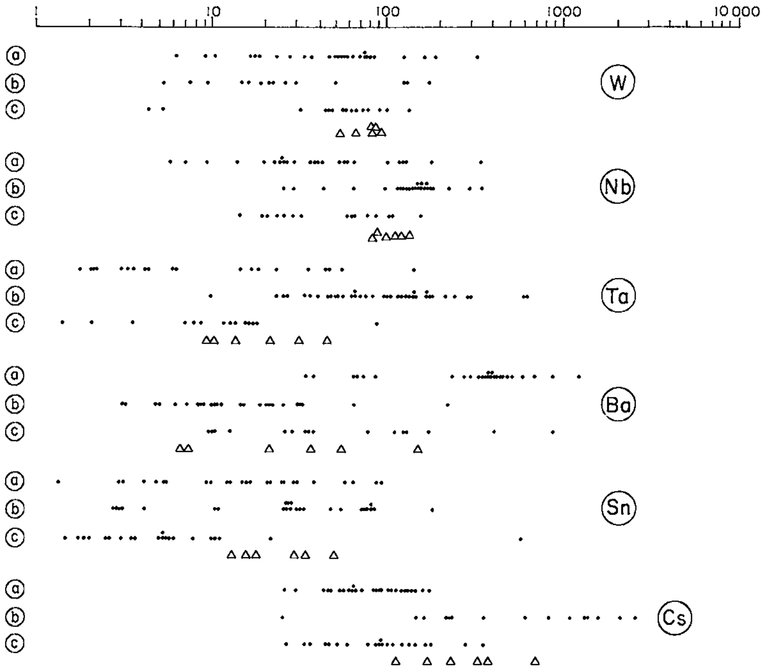

FIG. 12. Trace-element data of East Kemptville muscovite (open triangles) compared to data for muscovite from (a) granitoid rocks, (b) pegmatites and aplites and (c) greisens and veins. Sources of data: Neiva (1982, 1987), Jolliff *et al.* (1987), Arniaud *et al.* (1984), Möller & Morteani (1987).

TABLE 9. Sn CONTENTS OF MUSCOVITE SEPARATES

Sample	Ammonium Iodide Fusion	Sodium Peroxide Fusion	Sn in Muscovite Lattice
(ppm)			
EK-23C	270	335	65
EK-132	75	335	260
EK-161	330	385	55
EK-1075B	2050	2060	10
EK-1075B	50	168	118
EK-117	35	260	225
EK-118	350	627	277
Primary Avg.(n=5)	555 ± 755	656 ± 705	101 ± 86
Greisen Avg.(n=2)	192	443	251

magnetite alteration zones at the Climax–Henderson molybdenum deposit is relatively enriched in F: 17.8–19.8 (±0.6) wt.% F (Gunow *et al.* 1980).

DISCUSSION

The foregoing discussion indicates that both magmatic and postmagmatic equilibria are represented in the mineral chemistry of the East Kemptville leucogranite. Whereas the chemistries of muscovite, bulk alkali feldspar and topaz can be argued to result from mineral–melt equilibria, those of plagioclase, exsolved phases within alkali feldspar and biotite are more easily explained in terms of mineral–fluid equilibria. The following discussion is, therefore, focused on what information may be extracted from these minerals in terms of magmatic *versus* postmagmatic conditions.

Similar approaches have been applied to a vari-

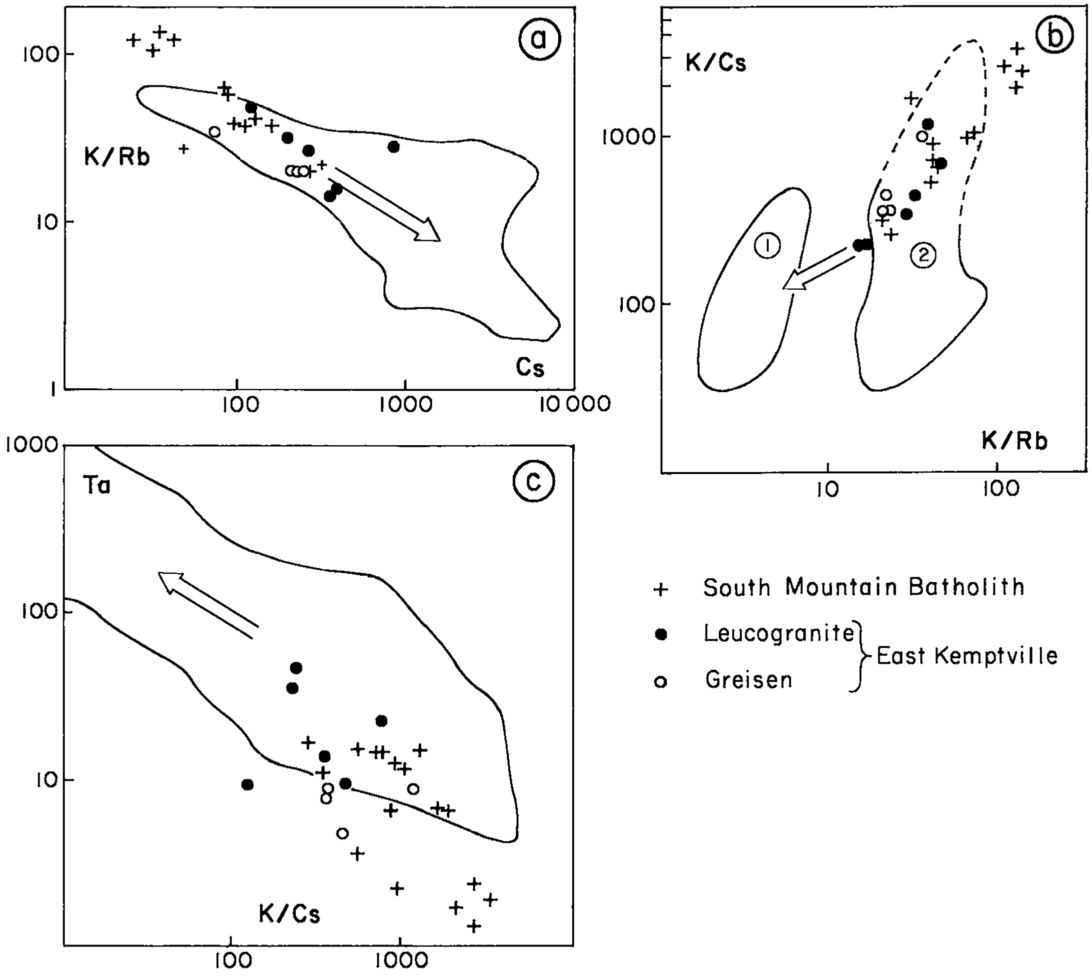

FIG. 13. Muscovite data plotted in binary plots of (a) K/Rb *versus* Cs, (b) K/Cs *versus* K/Rb and (c) Ta *versus* K/Cs. Data for the South Mountain Batholith from Ham & Kontak (1988; data on muscovite from East Kemptville greisen are the author's unpublished data. Fields in (a) and (c) outline the compositions of white mica in Ta-bearing pegmatites (from Möller & Morteani 1987). Fields in (b) correspond to compositions of white mica from (1) spodumene- and (2) columbite-bearing pegmatites of the Cross Lake pegmatite field, Manitoba (Černý *et al.* 1985).

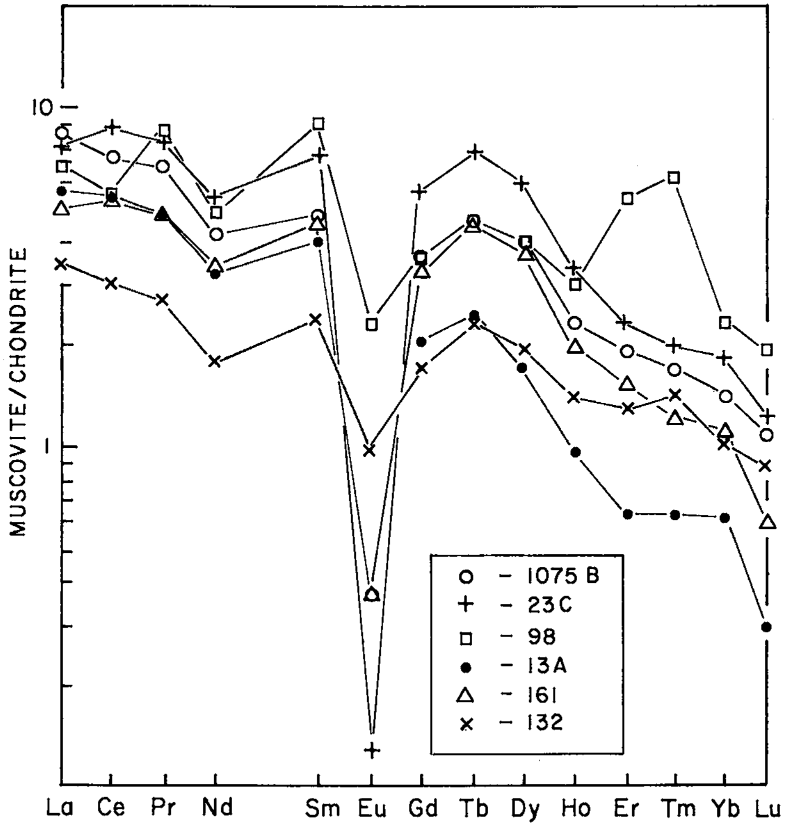

FIG. 14. Chondrite-normalized profiles for muscovite from East Kemptville leucogranite. Note the presence of a strong inflection in all profiles at Tb (discussed in the text).

ety of intrusive complexes, including unmineralized granitic rocks of Maine and the Isle of Skye (Ferry 1978, 1979, 1985) and base- and precious-metal mineralized granitic complexes in the Wasatch Mountains, Utah (John 1989), and the Yarington Batholith, Nevada (Dilles 1987). More relevant to the present investigation are the studies of background alteration in lithophile-element-mineralized systems in, for example, Australian (Witt 1985, 1988, Pollard 1984, Pollard & Taylor 1986, Taylor & Pollard 1988, Clarke & Taylor 1985) and South African (Taylor & Pollard 1988) granitic rocks, and the synopsis of Pollard (1983) for rare-element-enriched granitic rocks worldwide.

TABLE 10. REPRESENTATIVE COMPOSITIONS OF APATITE

	1	2	3	4	5	12	13
CaO	53.94	54.95	53.59	53.09	54.34	54.28	56.77
Na_2O	0.00	0.01	0.01	0.00	0.03	0.00	0.01
MnO	1.67	0.53	2.10	2.01	1.47	1.24	0.30
P_2O_5	42.20	41.84	41.75	42.21	42.30	41.88	42.29
F	3.45	3.04	2.94	4.03	2.80	3.10	2.42
	101.25	100.36	100.39	101.34	100.92	100.49	101.79
F≡O	1.44	1.27	1.23	1.69	1.17	1.30	1.01
Σ wt.%	99.81	99.09	99.16	99.65	99.75	99.19	100.78
Structural Formulae Based on 26 Oxygen Atoms							
Ca	9.760	10.025	9.807	9.580	9.865	9.899	10.121
Na	0.000	0.001	0.001	0.000	0.004	0.000	0.001
Mn	0.238	0.075	0.303	0.299	0.210	0.177	0.042
P	6.031	6.031	6.037	6.018	6.067	6.035	5.959
F	1.841	1.637	1.587	2.146	1.499	1.668	1.273

Analyses 3, 4, 5 are for various parts of the same grain.

Magmatic conditions

The presence of muscovite of inferred primary, magmatic origin indicates a strongly peraluminous nature for the melt and provides some constraints on the level of emplacement because of the positive d*P*/d*T* slope for the stability of this mineral (*e.g.*, Clarke 1981). The stability of muscovite in natural systems has been examined by several investigators, with conflicting results. For example, whereas Anderson & Rowley (1981) argued from thermodynamic data that impurities (*i.e.*, celadonitic component) expand the stability of end-member

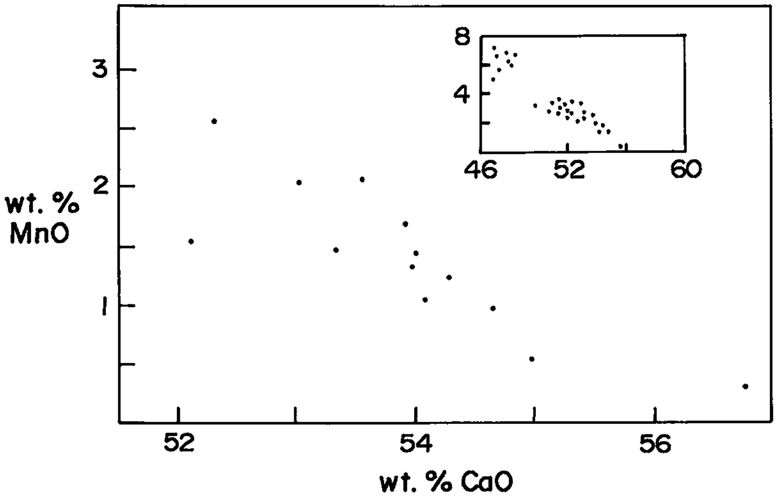

FIG. 15. Apatite compositions plotted as a function of wt.% MnO *versus* wt.% CaO. The inset diagram shows a similar plot for apatite from the Larrys River and Sangster Lake plutons of the Meguma Terrane, Nova Scotia (data from O'Reilly 1988).

TABLE 11. REPRESENTATIVE COMPOSITIONS OF TOPAZ

	1C	1R	2C	3R	4C	4R	5C	5R	STA
SiO_2	33.00	33.36	32.94	32.54	32.82	33.59	33.26	32.46	33.17
Al_2O_3	56.54	54.51	55.41	55.37	56.89	57.23	57.36	55.28	55.33
FeO	0.11	0.11	0.02	0.08	0.03	0.05	0.01	0.01	NA
F	16.08	17.20	17.57	17.92	15.81	15.09	17.59	17.81	15.78
H_2O	1.03	2.06	1.43	1.63	1.10	0.39	0.00	1.93	2.36
	106.76	107.24	107.37	107.54	106.64	106.35	108.22	107.49	106.64
F=O	6.76	7.24	7.37	7.54	6.64	6.35	7.40	7.49	6.64
Σ wt.%	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
Structural Formulae Based on 24 (O, OH, F)									
Si	4.062	4.058	3.563	3.510	3.611	3.718	4.077	3.948	3.651
^{IV}Al	0.000	0.000	0.437	0.490	0.389	0.282	0.000	0.052	0.349
^{VI}Al	8.205	7.817	6.629	6.551	6.990	7.186	8.288	7.875	6.830
Fe	0.011	0.010	0.001	0.007	0.002	0.004	0.000	0.000	0.000
F	6.261	6.617	6.011	6.114	5.502	5.283	6.819	6.852	5.493
OH	0.844	1.671	1.031	1.172	0.806	0.287	0.000	1.565	1.731
F/(F+OH)	0.881	0.798	0.853	0.839	0.872	0.948	1.000	0.814	0.760

C and R refer to core and rim analyses, respectively.
STA: topaz analysis from St. Austell granite, Cornwall (Manning & Exley 1984).

muscovite to higher *T* and, hence, lower *P* in felsic melts, the recent experimental work of Miller *et al.* (1989) indicates that such impurities have little effect on muscovite stability. However, these authors did note that the stability is enhanced under fluorite-saturated conditions. In contrast, experimental studies on naturally occurring volatile-rich (F, Li, B) samples of felsic rocks by Weidner & Martin (1987), Pichavant *et al.* (1987) and London (1987) indicate that the stability field of muscovite is expanded to higher *T* compared to end-member OH-muscovite for similar *P*, probably as a result of the influence of F (Munoz 1984). The work of Weidner & Martin (1987), and to a lesser extent that of Pichavant *et al.* (1987), is most applicable to the East Kemptville leucogranite and suggests that muscovite may have been stable to 1 kbar and 650–700°C.

The observed chemical variation of the muscovite, particularly with respect to the amount of celadonitic component (Fig. 10), remains unexplained. In terms of petrographic observations, the three samples with the most pristine textural features have the highest iron contents. Even in these samples there is a large range in composition (*i.e.*, 0.7–1.4 Fe p.f.u.), which suggests that muscovite compositions may reflect changing physicochemical parameters within the melt during crystallization [*e.g.*, $f(O)_2$, *T*, $P(H_2O)$, $a(HF)$]. For example, recent experiments by Miller *et al.* (1989) indicate that the iron content of muscovite is strongly dependent on $f(O)_2$. In any case, the chemistries of granite-hosted *versus* greisen-hosted muscovite are distinctly different with respect to both major- and trace-element abundances (Figs. 7, 11, 13), which is interpreted to reflect a primary magmatic origin for the granite-hosted muscovite.

The trace-element data for muscovite, if indeed reflecting the primary partitioning of elements between crystal and melt, imply primary enrichments of the magma in many elements generally associated with specialized granites. Conversely, the low absolute levels of all the *REE*, to <10 times chondritic abundances, indicates depletion of these elements in the melt.

The bulk composition of alkali feldspar determined ($Or_{82}Ab_{17}An_1$) indicates a value more typical of pegmatitic bulk compositions for the South Mountain Batholith than late-stage leucogranites (Kontak 1988a, Kontak & Strong 1988, Kontak, unpubl. data); however, the composition is not unlike values reported for ongonites (Kovalenko *et* *al.* 1971). Although moderately low temperatures of crystallization would be inferred from this bulk composition (*ca.* 500–600°C for coexisting plagioclase feldspar of An_{10} composition), such temperatures would be consistent with the elevated F content and presumably depressed solidus temperature of the melt (Manning 1981, Pichavant *et al.* 1987, Weidner & Martin 1987, Webster *et al.* 1987).

The exchange equilibria of K and Rb between K-feldspar and muscovite was investigated by Beswick (1973), and the relationship used to establish a geothermometer. Results of temperature determinations for five K-feldspar - muscovite pairs (Fig. 16) indicate a range in temperature from <400°C to 750°C, values similar to what Kovalenko *et al.* (1978) obtained for ongonites using the same technique. Although the lower temperature range (indicated by two samples) is unrealistic for magmatic conditions, the three higher values indicate a more reasonable temperature of *ca.* 650°C.

The presence of topaz of inferred magmatic origin in the leucogranite (Kontak 1990) indicates elevated F contents in the melt in order to stabilize this phase (Pichavant *et al.* 1987, Weidner & Martin 1987). The elevated F/(F + OH) value of the topaz (*ca.* 0.80 ± 0.05, Table 11) is consistent with the experimental studies of Barton (1982), which indicate that such conditions are required in order to stabilize the assemblage topaz + K-feldspar over quartz + muscovite. With decreasing F/(F + OH) ratios and temperature, the latter assemblage is preferred (Barton 1982, Burt 1981), such as is observed in the greisen assemblages at East Kemptville.

In Table 12 the results of mass-balance calculations and estimates of partition coefficients for Rb and Sr are presented for muscovite, plagioclase and K-feldspar from the East Kemptville leucogranite. The data were obtained during an Rb–Sr isotopic study of the leucogranite (Kontak *et al.* 1989, Kontak & Cormier, in press.). Results indicate consistent values for partition coefficients, the only exception being the large variation for K_{Sr}(muscovite). With the exception of the plagioclase data, calculated values for the leucogranite are comparable to other felsic suites (Table 13), particularly those with elevated volatile contents (Auriat Granite, Twin Peaks rhyolite, ongonites). The apparently low K_{Sr} value for plagioclase is considered to reflect its albitic composition compared to the more calcic composition of plagioclase from the Twin Peaks rhyolite and to the data used in Arth's (1976) compilation. In addition, the albitic plagioclase is considered to have formed from metasomatic processes (see below) and, hence, the trace-element chemistry may not reflect magmatic values. However, for muscovite and K-feldspar, the data are consistent with published values. Some variation between the East Kemptville data and those of less volatile-rich and less evolved suites would be consistent with the findings of Mahood & Hildreth (1983), who found that large partition coefficients characterize high-silica rhyolites.

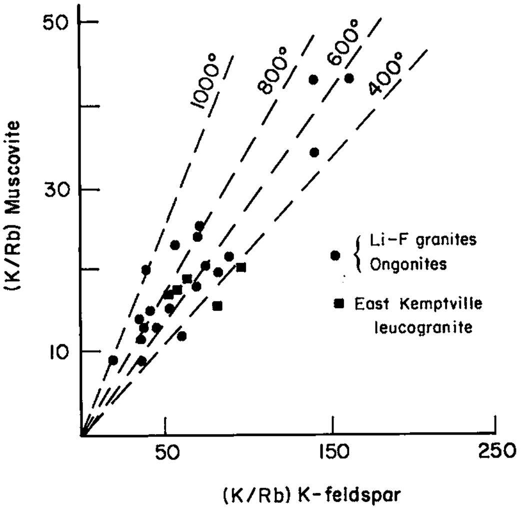

FIG. 16. Plot of the covariation of K/Rb in muscovite *versus* K-feldspar for mineral separates from the East Kemptville leucogranite (data obtained from Rb–Sr isotopic analyses: Kontak & Cormier, 1991, Kontak *et al.* 1989) compared to isotherms calculated by Beswick (1973). Data are compared to results for ongonites and Li–F granites from Kovalenko *et al.* (1978).

TABLE 12. Rb AND Sr MASS-BALANCE CALCULATIONS AND PARTITION COEFFICIENTS

	Rb				Sr			
Sample	M^{MS}_{wr}	M^{KF}_{wr}	M^{PL}_{wr}	Σ	M^{MS}_{wr}	M^{KF}_{wr}	M^{PL}_{wr}	Σ
EK-86-161	0.38	0.464	0.159	1.003	0.122	0.497	0.321	0.940
EK-86-1075B	0.515	0.201	0.220	0.936	0.322	0.140	0.422	0.884
EK-86-132	0.459	-	-	-	0.136	-	-	-
EK-86-23C	0.154	0.461	0.168	0.783	0.104	0.461	0.307	0.872
EK-89-1	0.569	0.328	0.222	1.119	0.055	0.330	0.505	0.890
EK-86-156	0.453	0.300	0.229	0.982	0.043	0.241	0.424	0.708
EK-86-124B	-	0.145	0.231	-	-	0.094	0.403	-
EK-86-9	-	0.528	0.339	-	-	0.320	0.484	-
MASS BALANCE	=			0.964 ±0.109				0.858 ±0.078

	Rb			Sr		
Sample	K^{MS}_{wr}	K^{KF}_{wr}	K^{PL}_{wr}	K^{MS}_{wr}	K^{KF}_{wr}	K^{PL}_{wr}
EK-86-161	5.34	2.00	0.433	1.94	2.14	0.876
EK-86-1075B	5.25	1.86	0.824	3.29	1.29	1.58
EK-86-132	3.58	-	-	1.07	-	-
EK-86-23C	4.98	1.87	0.572	3.36	1.87	1.04
EK-89-1	7.30	1.69	0.740	0.708	1.70	1.68
EK-86-156	5.59	1.36	0.700	0.531	1.09	1.29
EK-86-124B	-	1.34	0.868	-	0.872	1.51
EK-86-9	-	2.10	1.03	-	1.27	1.46
PARTITION COEFFICIENTS	5.34 ±1.09	1.74 ±0.27	0.738 ±0.181	1.81 ±1.15	1.46 ±0.42	1.34 ±0.27

wr = whole rock, MS = muscovite, KF = K-feldspar, PL = plagioclase

The evolved nature of the East Kemptville leucogranite is indicated from both the trace-element and *REE* contents of mineral phases (K-feldspar, muscovite; see above) and the composition of the

TABLE 13. CALCULATED PARTITION-COEFFICIENTS FOR EAST KEMPTVILLE LEUCOGRANITE COMPARED TO OTHER FELSIC SUITES

	Rb	Sr	Reference
East Kemptville			
muscovite	5.34 ± 1.09	1.81 ± 1.15	this study
K-feldspar	1.74 ± 0.27	1.46 ± 0.42	
plagioclase	0.738 ± 0.181	1.34 ± 0.27	
Yellowstone rhyolite			Leeman & Phelps (1981)
K-feldspar	0.40 ± 0.03	NA	
Auriat Granite			Arniaud *et al.* (1984)
muscovite	3.43	1.95	
K-feldspar (Or_{77})	2.25	0.93	
plagioclase	0.92	1.18	
Tuscany rhyolites			Giraud *et al.* (1986)
K-feldspar (Or_{76})	0.8	7.2	
Bishop Tuff			Hildreth (1977)
K-feldspar	0.55-0.65	NA	
Twin Peaks rhyolite			Nash & Crecraft (1985)
K-feldspar (Or_{60})	1.2-1.8	4.5-7.3	
plagioclase (An_{30-42})	0.06-0.19	6.8-33.0	
Rhyolitic rocks			Arth (1976)
K-feldspar	0.34	3.87	
plagioclase	0.041	4.4	
Ongonites			Kovalenko *et al.* (1978)
muscovite	5.48 ± 1.61		
K-feldspar	3.05 ± 1.22	NA	
Granites			
K-feldspar (Or_{60-80})	0.77 - 1.1	1.2 - 5.0	Long (1978)

Partition coefficient used here is defined as $K_d = C_m/C_l$, where K_d = partition coefficient, C_m = concentration of element of interest in mineral, C_l = concentration of element of interest in liquid (or rock in the case of granitic suites referred to above).

most primitive biotite (analysis 5 in Table 7 and Figs. 3, 4, 6). The elevated Fe/(Fe + Mg) value of biotite also characterizes evolved ilmenite-series granites of Japan (Czamanske *et al.* 1981) and F-rich igneous suites (Rubin *et al.* 1987, Christiansen *et al.* 1986, Webster *et al.* 1987, Nash *et al.* 1985). In addition, the trace element and *REE* data for K-feldspar and muscovite are comparable to data of the same mineral phases in pegmatites from the eastern part of the South Mountain Batholith and fields defined for rare-element pegmatites by Černý and coworkers (as discussed above).

Finally, the low Fe^{3+}/Fe^{2+} ratios of muscovite and elevated Fe/(Fe + Mg) values of biotite are both consistent with low $f(O_2)$ in the magma. Christiansen *et al.* (1986) also inferred such reducing conditions based on similar compositions of biotite, which they confirmed using two-oxide equilibria (lacking in the East Kemptville leucogranite) to calculate absolute values of $f(O_2)$.

Postmagmatic conditions

Several of the mineral phases (biotite, K-feldspar, plagioclase) have chemistries (also textures in the case of K-feldspar) that reflect re-equilibration at the postmagmatic stage, presumably with deuteric fluids.

The chemistry of biotite reflects considerable departure from typical magmatic values in terms of major cations (*i.e.*, Al, Si, Fe) compared to trends established for the South Mountain Batholith and Davis Lake Pluton. The extensive departure toward muscovite (Fig. 5) reflects temperatures below 600°C, on the basis of the triangular plot of Monier & Robert (1986; Fig. 7). The low Ti contents of the biotite similarly suggest low temperatures of equilibration (Le Bel 1979, Taner *et al.* 1986, Hewitt & Wones 1984, Dilles 1987, Czamanske *et al.* 1981), as does the chloritic alteration of biotite. For example, Eggleton & Banfield (1985) suggested a temperature of *ca.* 330–340°C for the alteration of biotite to chlorite in Australian granites, whereas Ferry (1979) calculated a temperature of 425 ± 25°C for similar alteration in granites from south-central Maine. Further alteration of biotite and chlorite to an unidentified clay(?) phase probably reflects equilibration to even lower temperatures (Ferry 1985).

An important adjunct to the alteration of biotite is the considerable Fe enrichment observed. It is not uncommon for apparent reverse differentiation to occur in biotite because of the sensitivity of this mineral to changes in $f(O_2)$ and, consequently, preferential incorporation of Mg rather than Fe (as Fe^{3+}) if $f(O_2)$ should increase. For example, Corey (1988) documented such a relationship in some alteration zones within the South Mountain Batholith, and Czamanske & Wones (1973) demonstrated such an occurrence in the Finnmarka Complex, Norway. Hence, the lack of a similar relationship suggests that excursions of late-stage fluids to higher $f(O_2)$ conditions did not prevail at East Kemptville.

The presence of coarse perthitic textures in K-feldspar indicates subsolidus reaction of once homogeneous, unexsolved alkali feldspar with a fluid phase (Parsons 1980, Parsons & Brown 1984). The compositions of the potassium- and sodium-rich domains can be used, albeit only in a general way, to estimate the temperature of final re-equilibration using two-feldspar geothermometry. For compositions determined for the respective domains (*i.e.*, Ab_{99} and Or_{96}), temperatures of *ca.* 350–400°C are indicated. Ferry (1979, 1985) found similar compositions for the potassium- and sodium-rich phases of exsolved alkali feldspar in deuterically altered granites from south-central Maine and the Isle of Skye and inferred similar temperatures of feldspar re-equilibration. Structural determinations (unpubl. data, Kontak; R.F. Martin, pers. comm. 1989) of the two intergrown phases indicate the presence of fully ordered microcline and low albite. The stability fields of these phases are consistent with re-equilibration at 350–400°C.

Origin of albite

The presence of pure end-member albite in the East Kemptville leucogranite strongly suggests a period of fluid:rock interaction, during which albite was the stable product. Important also is the presence of abundant inclusions of zoned apatite, which are considered to have been generated during albitization of a more Ca-rich plagioclase. Implicit, therefore, in this reaction is the availability of P in order to promote apatite crystallization. The recent experimental work of London (1987) has clearly demonstrated that late-stage fluids in evolved, volatile-rich felsic systems commonly are enriched in P, among other volatile species, because they remain unbuffered throughout the crystallization history of the melt. The absence of a primary phosphate phase in the leucogranite (*e.g.*, apatite) is consistent with this proposal. It is important to note that the albitization referred to here does not include replacement of alkali feldspar, but only replacement of precursor plagioclase.

The presence of magmatic albite in felsic rocks is rare; perhaps one of the only convincing cases is that of the topaz-bearing quartz keratophyres or ongonites described by Kovalenko *et al.* (1971). In this instance albite of An_5 composition is described as containing small glass inclusions. More recently, Weidner & Martin (1987) have argued for a magmatic origin for albite in the fluorine-rich leucogranite from St. Austell, Cornwall, based on the lack of "residual" ordering in the albite. In contrast, Pichavant *et al.* (1987) have shown that albite in the Beauvoir albite–lepidolite granite originated from late-stage modification of a Ca-rich precursor; experimental runs on powdered granite produced plagioclase of An_3 composition, not pure end-member albite as found in the granite. Of relevance also is the experimental work of Weber & Pichavant (1986), which demonstrated that Ca-bearing plagioclase will preferentially form even from melts with low bulk CaO contents (*i.e.*, < 1 wt. %). The exception to this is favored when an additional Ca-bearing phase such as fluorite is present and, hence, competes for the calcium (*e.g.*, Weidner & Martin 1987). Also important is the fact that the plagioclase in topaz rhyolites is generally sodic oligoclase, although calcic albite may be present in the most evolved rocks (Christiansen *et al.* 1986, Webster *et al.* 1987). Where late-stage fluid interaction has been demonstrated, such as in the cryolite-bearing rhyolites of Texas (Rubin *et al.* 1989), end-member albite occurs (Rubin *et al.* 1987).

In the case of the East Kemptville leucogranite, the primary magmatic assemblage is considered to have been quartz–sanidine–plagioclase–topaz–muscovite, with the plagioclase probably of oligoclase composition from analogies with fresh volcanic rocks of similar composition (*e.g.*, Pichavant *et al.* 1988, Christiansen *et al.* 1986). The formation of albite resulted from replacement of precursor Ca-bearing plagioclase during either the late-magmatic or early postmagmatic stage. The first possibility is suggested because of the F-rich nature of the melt, which would have become more Na-rich during the terminal stages of crystallization (Manning 1981) because of internal diffusion and convective circulation. Pollard (1983) favored such a mechanism for albitization or Na-feldspathization at a late-magmatic stage. Concerning the second possibility, the late-stage exsolution of a Na-rich fluid phase, common in felsic rocks (Burnham 1979), might promote postmagmatic albitization. The presence of hypersaline, halite-bearing fluid inclusions in the quartz grains of the leucogranite (Kontak 1988b) strengthens this case. The latter observation is consistent with the experimental study of albitization of plagioclase by Orville (1972), who found that the reaction proceeds only under very low ratios of $[CaCl_2/(CaCl_2 + NaCl)]$. The presence of phosphate minerals (apatite, triplite) in quartz veins at East Kemptville indicates that a P-rich hydrothermal fluid was available within the system.

Since the composition of end-member albite is not diagnostic of its temperature of formation (Wenk & Wenk 1977, Moody *et al.* 1985) and no structural data for albite are presently available (cf. Martin 1984, Weidner & Martin 1987), it is not possible to say unequivocally whether albitization was a late-magmatic or early postmagmatic event.

CONCLUSIONS

A chemical study of the mineral phases of the fluorine-rich, topaz–muscovite leucogranite at the East Kemptville tin deposit reveals variable amounts of magmatic and postmagmatic equilibration. Whereas topaz, muscovite and bulk alkali feldspar chemistries are consistent with a magmatic origin, as argued using both major- and trace-element data, biotite and plagioclase reflect re-equilibration at a late-magmatic or postmagmatic stage. Collectively, the chemistries of muscovite and K-feldspar indicate a highly evolved, fluorine-rich bulk composition for the melt, with concentrations of some trace elements more typical of pegmatites than granites. Partition coefficients for Rb and Sr calculated for these minerals are similar to values reported in the literature and are consistent, therefore, with magmatic signatures. Both the low Fe^{3+}/Fe^{2+} ratios and the extremely Fe-rich nature of muscovite are consistent with a reduced, highly evolved melt. Compositions of the most pristine biotite also are consistent with these conclusions. Estimates of magmatic temperatures vary from 700 to 600°C based on (1) the stability of muscovite, (2) K and Rb partitioning between muscovite and K-feldspar, and (3) general estimates using two-feldspar geothermometry.

Chemistries of biotite and plagioclase reflect postmagmatic re-equilibration. The composition of the biotite chemistry (low Ti, Fe and elevated Si, Al) indicates considerable departure toward muscovite and away from compositions typical for evolved rocks in other parts of the South Mountain Batholith.

Depleted Ti contents suggest low temperatures of re-equilibration, possibly $<400°C$, whereas the Mg-poor and Fe-rich nature of the biotite and biotite-like phases is consistent with formation under consistently low $f(O_2)$ conditions, whether of magmatic or postmagmatic origin. Albitization of an original Ca-bearing plagioclase resulted in the assemblage of albite + apatite. This reaction may have proceeded under either late-magmatic or early postmagmatic conditions; the present data do not permit an unequivocal decision.

The alteration of biotite to chlorite and an unidentified clay (?) mineral indicates re-equilibration to temperatures below 400°C. Thermometry on exsolution phases in K-feldspar indicates similar temperatures of re-equilibration (*i.e.,* 350–400°C).

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