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The Tregonning granite: petrogenesis of Li-mica granites in the Cornubian batholith

MAURICE STONE*

Geologisk Institut, Aarhus Universitet, C.F. Møllers Allé, 8000 Aarhus C, Denmark

Abstract

Li-mica (zinnwaldite and/or lepidolite)—topaz—albite granites in the Tregonning—Godolphin pluton and similar rocks in the St. Austell pluton appear to be petrogenetically unrelated to the spatially associated biotite granites. Evidence is provided by lack of development of Li-mica granites at roof zones of biotite granites and markedly different trends and composition fields in bivariate plots such as Li *vs.* Cs, Rb *vs.* Sr and Nb *vs.* Zr. Thus, differentiation of biotite granite magma is unlikely to have generated Li-mica granite magma, as also, on its own, is partial melting of biotite granite or biotite-absent residual lower crust. However, partial melting of biotite-rich residual rocks involving biotite breakdown could yield a trace alkali- and F-enriched melt, although this would require marked femic mineral, K-feldspar and anorthite fractionation, and Na-enrichment. It is proposed that volatiles derived from either a mantle source or the crust/mantle interface have aided metasomatism of either residual S-type crust that earlier provided S-type biotite granite magma, or basic (biotite-rich) granitoid, to produce a low-temperature, low-viscosity Li-mica granite melt that rose rapidly in the crust soon after the emplacement of associated biotite granites.

KEYWORDS: lithium, mica, granite, Tregonning, Cornubian batholith, zinnwaldite, lepidolite.

Introduction

LI-MICA granites occupy only a small part of the exposed Cornubian batholith (<1%) and form an essential part of only two of its plutons, namely St. Austell and Tregonning–Godolphin. Field evidence from the St. Austell pluton (Exley and Stone, 1982; Manning and Exley, 1984) and, in particular, the Tregonning–Godolphin pluton (Stone, 1975) indicates that the bodies of Li-mica granite were intruded passively and post-date their spatially associated biotite granites. In the Tregonning–Godolphin pluton, the earlier Godolphin intrusion is composed of poorly megacrystic biotite granite, whilst the Tregonning intrusion consists of Li-mica granites (Fig. 1). The Li-mica granites are readily distinguished from the biotite granites by the occurrence of the mineral suite zinnwaldite (or lepidolite), topaz and albite. The term 'Li-mica granite' is preferred to the term 'topaz granite' (Pichavant and Manning, 1984; Manning and Hill, 1990) in the present context, because biotite granites can also contain topaz (e.g. Lundy, the Mourne Mountains and Northern Arran granites).

Previous work suggested that Li-mica granites have been derived by extreme differentiation of biotite granites (Manning, 1981), partial fusion of residual lower crustal rocks (Manning and Hill, 1990) or metasomatism and melting of earlier-formed biotite granite (Stone, 1975; 1984). The principal objective of this study is to compare the Tregonning granite and its roof complex with the biotite granites in an attempt to find a petrogen-

* Present address: Earth Resources Centre, University of Exeter, North Park Road, Exeter, Devon EX4 4QE, England.

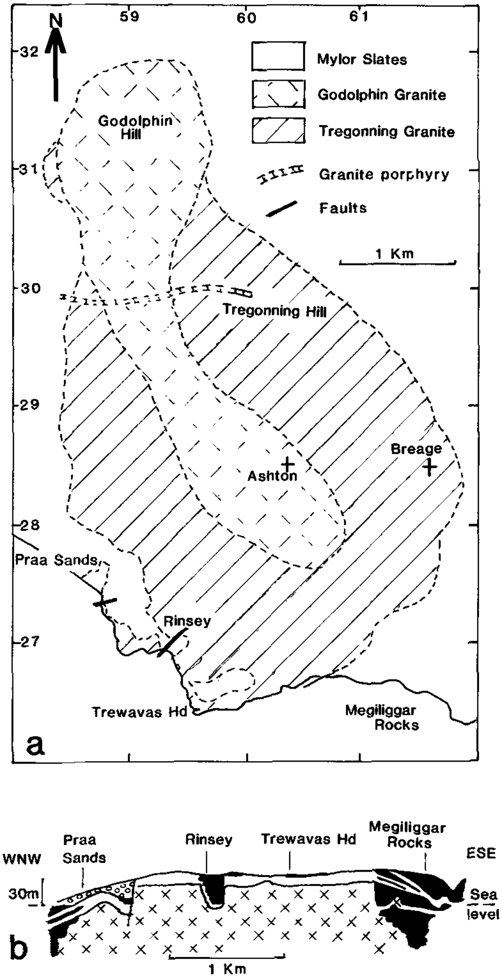

FIG. 1(*a*) Geological sketch map of the Tregonning–Godolphin pluton showing the extent of outcrop of the Li-mica granites. Modified after Taylor and Wilson (1975). Numbers round the margin refer to the National Grid. (*b*) Diagrammatic section across the Tregonning granite intrusion. From Exley and Stone (1982), in Sutherland, D. (Ed.) Igneous Rocks of the British Isles. Copyright 1982, John Wiley and Sons Ltd. Reprinted by permission of John Wiley and Sons Ltd. Key: black = Mylor Slate Formation; open circles (○) = Recent deposits; crosses (×) = zinnwaldite (Tregonning) granite overlain by layered roof complex (without ornament).

esis consistent with one or more of these hypotheses.

Data sources for the Li-mica granites include published (Stone, 1975, 1982, 1984; Manning and Hill, 1990) and unpublished work. Comparisons with the biotite granites use data from Darbyshire and Shepherd (1985), Manning and Hill (1990) and Heath (1982), together with raw material used in the summary data of Stone (1987) and Stone and Exley (1989). The term biotite granite includes all coarse- to medium-grained (i.e. average grain size >1 mm), commonly megacrystic, biotite granite (Type B of Exley and Stone, 1982). The term microgranite is used for associated fine-grained (<1 mm), usually later, rocks (aplites and biotite microgranites) that occur as dykes and sheets in the biotite granites.

Tregonning granite

Mineralogy. Most of the mineral data apply to the Li-mica granites of the Tregonning–Godolphin pluton, but are broadly consistent with data observed or reported from the Li-mica granites of the St. Austell pluton. Microprobe data for *feldspars* (Stone, 1984, and unpublished data) reveal almost pure unzoned albite (typically $Or_{0.6}Ab_{99.2}An_{0.2}$) and orthoclase/microcline with high Or ($Or_{95}Ab_5$). The *trioctahedral micas* have been fully described elsewhere (Stone, 1984; Stone *et al.*, 1988; Henderson *et al.*, 1989). The Tregonning granite itself contains zinnwaldite, but its roof complex contains lepidolite in leucogranites and zinnwaldite in pegmatites. However, there is a transition between zinnwaldite and lepidolite and, at hornfels contacts, between biotite (siderophyllite) and zinnwaldite. *Muscovite* is common in roof granite/pegmatite where compositions broadly compare with those of the biotite granites but are richer in total Fe, Rb and F and poorer in Ti and Mg (Table 1; also compare with Stone and Exley, 1989, Table 1).

Tourmaline (Lister, 1978; Power, 1968 and unpublished data) in the Tregonning granite is schorl, richer in Li and F than that in the biotite granites (Table 1). High Li is inferred from high Rb contents in microprobe analyses and an analysis of tourmaline from Li-mica pegmatite at the Megiliggar Rocks (Li_2O = 0.46 wt.%). Pale brown tourmaline in the lepidolite leucogranites and some aplites (Table 1, cols 5 and 6) is richer in Al, Mn and F, and poorer in Fe and Ti than tourmaline in pegmatite and zinnwaldite granite. *Topaz* was considered to be late- or post-magmatic by Stone (1984). Microprobe data indicate F/(F + OH) = 0.7, although (probably less reliable) unit cell data for topaz in Li-mica leucogranite give F/(F + OH) greater than this (Stone and George, 1978). Comparable (slightly higher) F/(F + OH) occurs in topaz in Li-mica granite from the St. Austell pluton (Manning and Exley, 1984). Phosphates include *fluorapatite,* *triplite* (George *et al.*, 1981), *amblygonite*, and others (Stone and George, 1983).

Petrography and geochemistry. Stone (1984) distinguished aphyric zinnwaldite–topaz–albite granite (Type E of Exley and Stone, 1982) that forms the bulk of the Tregonning granite outcrop from overlying lepidolite-bearing leucogranite and associate aplite and pegmatite in the Roof Complex (Fig. 1*b*). Average chemical analyses (Table 2, cols 2 and 3) are compared with the spatially-associated poorly megacrystic biotite granite (Godolphin granite, col. 1), together with various other biotite granites and microgranites.

Some chemical variations in the Tregonning–Godolphin pluton were reported by Stone (1975, 1982) and will not be repeated here. Suffice it to say that there is strong positive covariation between elements/oxides of a 'trace-alkali' suite (includes F, Nb, Mn, P, Sn in addition to Li, Rb and Cs e.g. Manning and Hill, 1990, Fig. 11; Stone, 1982, Table 2 and Fig. 4) and strong negative relationship between members of this suite and a strongly correlated 'femic oxide/element' suite (TiO_2, total iron oxide—as tFeO, MgO, CaO, Zr, Sr, Ba). Such relationships are common elsewhere, but are particularly strong in the Tregonning–Godolphin and St. Austell plutons, owing to the apparently more highly evolved nature of Li-mica granites compared with the most differentiated members (the microgranites) of the biotite granite suite. Within the Li-mica granite suite (Tregonning intrusion) of the Tregonning–Godolphin pluton, the most evolved rocks are leucogranite, pegmatite and aplite that form the Roof Complex (Fig. 1*b*) and have clearly differentiated *in situ* (Stone, 1975).

Comparison with Cornubian biotite granites

The Li-mica granites show a marked enrichment in the 'trace alkali' elements, Nb, Mn, Ga, Sn, F and P_2O_5, and deficiency in most of the 'femic' suite, including K and Sr, compared with the biotite granites (Table 2). This produces very low K/Rb and high Rb/Sr in the Li-mica granites. Comparisons are enhanced in plots taken from the Minitab Oneway analysis of variance diagrams (see also Stone, 1990). The raw data

Table 1. Microprobe analyses of muscovite and tourmaline

	1	2	3	4	5	6	7
	Muscovite		Tourmaline				
Rock	TG3	peg	TG3	TG2	TG3	TG3	peg
Sample	810	845	810	516	815	035	845
				a b			
SiO_2	46.94	44.64	34.90	34.96	38.46	37.94	35.78
TiO_2	0.24	0.40	0.62	0.32 - 0.66	0.36	0.44	0.51
Al_2O_3	34.93	34.14	34.55	34.77	38.56	36.94	35.57
tFeO	2.54	3.38	10.64	12.97 -13.95	4.72	6.85	13.35
MnO	0.05	0.05	0.08	0.04 - 0.12	1.46	0.95	0.09
CaO	nd	0.03	0.50	0.11	0.04	nd	0.13
Na_2O	0.75	0.85	1.82	1.68	2.84	3.07	1.99
K_2O	9.54	9.33	0.07	0.66	nd	nd	nd
Rb_2O	0.37	0.18	---	---	---	---	---
F	---	2.16	---	---	1.40	1.36	0.73
Total	95.36	95.16	83.18	85.51 -86.91	87.84	87.55	88.15
n	3	5	4	1 1	3	6	6

--- = not determined; nd = not detected; n = number of points analysed.

1 and 2. Muscovite in aplite and pegmatite, respectively, at roof contact and adjacent to xenolith, Megiliggar Rocks (GR SW608267).

3. Tourmaline as (1).

4. Tourmaline in zinnwaldite granite at contact, Lesceave Por (GR SW586275): a = centre (pale yellow brown), b = rim (dark brown). Single values are averages of centre and rim where these are close.

5. Tourmaline in lepidolite leucogranite, Megiliggar Rocks: pale yellow unzoned grains.

6. Tourmaline in zinnwaldite aplite, Roof Zone, Trequean Cliff (GR SW603266): pale unzoned skeletal grains.

7. Tourmaline in pegmatite, as (2).

Table 2. Average analyses of Cornubian granites

	1 TG1	2 TG2	3 TG3/4	4 CM1	5 CM3	6 CM4	7 SC2	8 SC3/4	9 DT1	10 DT2
Wt.%										
SiO_2	72.28	71.13	72.52	71.74	72.84	75.71	71.52	73.28	71.81	74.86
TiO_2	0.28	0.07	0.03	0.24	0.17	0.07	0.24	0.08	0.38	0.17
Al_2O_3	14.88	15.91	15.53	15.20	14.57	14.42	14.84	14.36	14.46	13.11
tFe_2O_3	2.18	1.39	0.73	1.82	1.46	1.10	1.66	0.99	3.05	1.96
MgO	0.38	0.10	0.07	0.43	0.38	0.13	0.36	0.20	0.50	0.19
CaO	0.57	0.60	0.31	0.93	0.68	0.57	0.81	0.52	1.17	0.40
Na_2O	2.42	3.79	4.93	3.00	3.18	3.76	2.94	3.29	3.23	3.19
K_2O	5.97	4.71	3.60	5.19	4.91	4.29	5.42	5.03	5.06	4.89
P_2O_5	0.27	0.50	0.45	0.23	0.22	0.26	0.23	0.23	0.21	0.19
F	0.13	1.28	1.57	0.40	0.24	0.49	0.24	0.17	--	--
ppm										
Nb	15	65	68	12	14	18	12	14	18	20
Zr	140	41	15	110	74	25	117	29	152	97
Y	23	28	19	17	14	5	16	17	39	32
Sr	99	50	47	86	78	27	108	37	88	19
Rb	481	1262	1627	505	469	603	441	493	420	558
Mn	241	604	849	358	322	259	236	242	484	523
Ba	318	43	15	257	223	111	420	255	280	67
La	42	13	21	36	27	1	25	4	36	18
Ce	90	12	8	78	59	19	78	8	66	35
U	9	11	7	15	10	12	7	5	13	19
Th	21	9	6	14	8	3	27	6	31	14
Pb	23	6	4	33	37	15	37	36	35	24
As	17	23	22	15	48	49	9	5	8	21
Ga	19	31	34	21	21	23	21	20	18	23
Zn	67	50	37	40	57	48	37	33	49	40
Cs	25	121	183	56	58	46	30	35	46	75
Sn	15	29	30	14	17	15	9	12	16	19
Li	174	1381	2301	434	412	295	293	200	306	--
K/Rb	103.5	31.9	19.0	85.4	87.3	64.7	102.5	87.9	101.2	69.2
Rb/Sr	5.4	40.5	85.8	7.9	6.0	33.9	4.3	17.4	5.3	30.7
ASI	1.38	1.29	1.24	1.30	1.30	1.29	1.29	1.28	1.17	1.21
N	4	14	20	23	10	14	22	15	16	4

1-3 = Godolphin (TG1), Tregonning (TG2) and Roof Complex (TG3/4) granites.
4-6 = Carnmenellis outer (CM1), inner (CM3) and microgranites (CM4) (Stone, 1987)
7,8 = Isles of Scilly outer Granite (SC2) and inner granites/microgranites (SC3/4) (Stone and Exley, 1989)
9,10 = Dartmoor megacrystic (DT1) and poorly megacrystic (DT2) granites (Darbyshire and Shepherd, 1985; Heath, 1982 and Stone, unpublished).
N = number of samples; -- = not determined; tr = trace amounts; ASI = Aluminium saturation index (mol Al_2O_3/mol(CaO^*+Na_2O+K_2O), where CaO^* has been corrected for P_2O_5 in apatite); tFe_2O_3 = total Fe as Fe_2O_3.
Analysis: as in Stone *et al.* (1988).

used in Fig. 2 are those used in compiling Table 2, together with St. Austell data (Darbyshire and Shepherd, 1985; Manning and Hill, 1990). The clearest separations between evolved and relatively primitive granites are shown by several members of the 'femic oxide/element' suite (Stone, 1987). TiO_2 (not shown), Zr (Fig. 2*a*), and tFeO (Fig. 2*b*) show clear trends towards lower values with evolution, and little or no separation of Li-mica granites from the microgranites. However, K/Rb (Fig. 2*c*) shows similar decreasing trends in all plutons together with a marked separation of the Li-mica granites from the microgranites. Nb, Sn and Ga (not shown), and Li (Fig. 2*d*) (i.e. elements of the 'trace alkali' suite) also show clear separation of the two granite suites and marked increases in Li-mica granites with little or no increase in the microgranites compared with the biotite granites.

Simple covariation is illustrated in bivariate plots of several members of each of the element/oxide suites. Zr *vs.* TiO_2 (Fig. 3*a*) and TiO_2 *vs.* tFeO (not shown) are typical of 'femic' suite covariation and suggest a simple evolutionary sequence from biotite granite to Li-mica granite. These plots do not discriminate between Li-mica granites and microgranites. Covariation between members of the 'trace-alkali' suite is illustrated in the Cs *vs.* Li plot (Fig. 3*b*). The wide variation within the Li-mica granites is likely to arise from late/post-magmatic events, resulting in low K, Cs and Rb in micas, and scatter away from a linear pattern with the biotite granites. Of considerable importance in this plot is the field of very low values shown by the Isles of Scilly and Carnmenellis microgranites, giving a trend from biotite granites (fields A and B) in the *opposite* direction to that of the Li-mica granites. K/Rb *vs.* TiO_2 (Fig. 3*c*) also largely separates Dartmoor data from the biotite granites of the other plutons, despite overall large scatter. A broad trend of diminishing TiO_2 is accompanied by falling K/Rb to values of 60–90 (and no lower than 40). Li-mica granites compare in their low TiO_2 contents with the microgranites, but have lower values of K/Rb (below 40). A Nb *vs.* Zr plot (Fig. 3*d*) also separates the two groups (see also Manning and Hill, 1990, Fig. 9). Carnmenellis data show almost constant Nb content with decreasing Zr in the time series outer granite, inner granite, microgranite (Stone, 1987), despite a wide scatter in the microgranites. All Li-mica granites occupy a different field at high Nb, low Zr content.

Discriminant plots (Fig. 4) based upon Pearce *et al.* (1984) also separate the biotite granites from the Li-mica granites. As pointed out by Stone (1990), Dartmoor biotite granites plot on the WPG/syn-COLG boundary or much closer to the WPG field than the Isles of Scilly and Carnmenellis data points. In the Nb *vs.* Y plot (Fig. 4*a*) the Li-mica granites cross the syn-COLG/WPG field boundary, with Tregonning data points well within the WPG field as a result of enhanced Nb contents compared with the biotite granites, but similar ranges of Y. All data plot within the syn-COLG field in the Rb *vs.* (Nb + Y) diagram (Fig. 4*b*), although the three groups referred to above occupy distinct composition fields, with Li-mica granites, having both higher Rb and Nb + Y, well separated from the biotite granite.

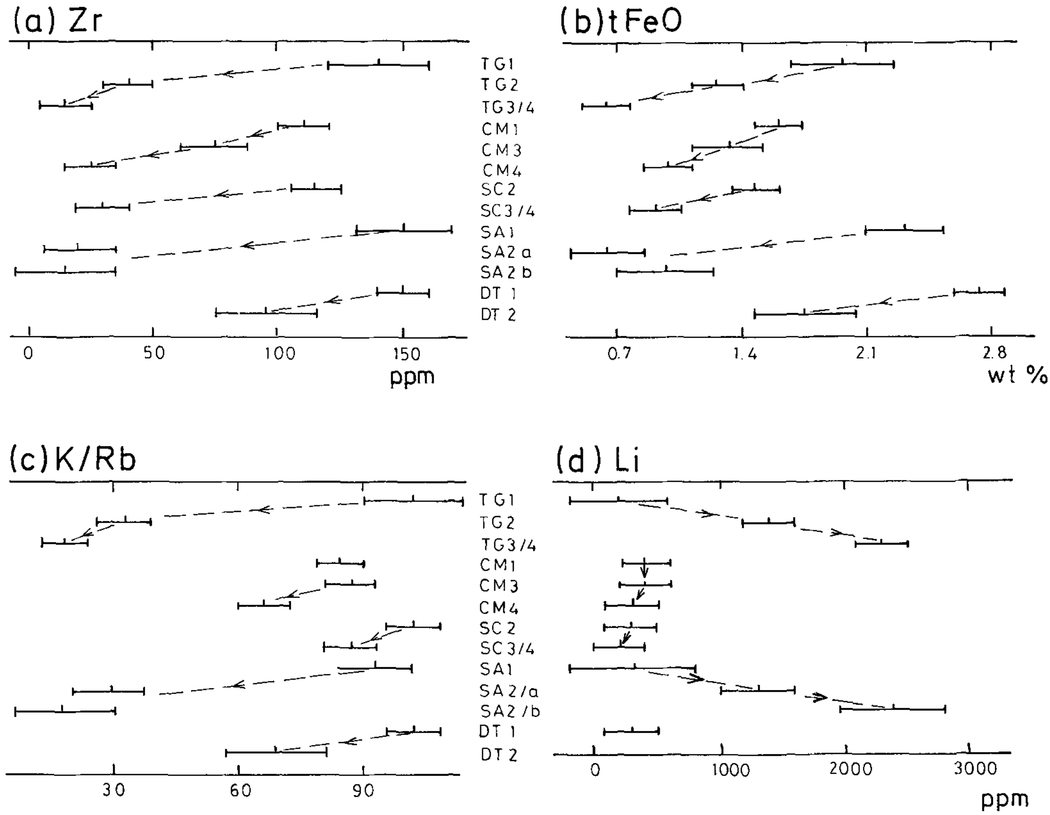

FIG. 2. Means and 95% confidence intervals of rock types of Table 2 together with St. Austell data (SA), grouped according to pluton in order to illustrate possible evolutionary trends from older to later rock types (indicated by dashed lines and arrows). (*a*) Zr; (*b*) tFeO (total iron as FeO); (*c*) K/Rb; (*d*) Li. Symbols as in Table 2 and, in addition, for St. Austell data: SA1 = megacrystic biotite granite (Darbyshire and Shepherd, 1985; Manning and Hill, 1990); SA2a and b = Li-mica granites from Nanpean and Hensbarrow stocks respectively (Manning and Hill, 1990). The figures are taken directly from the Minitab output in the oneway AOV command.

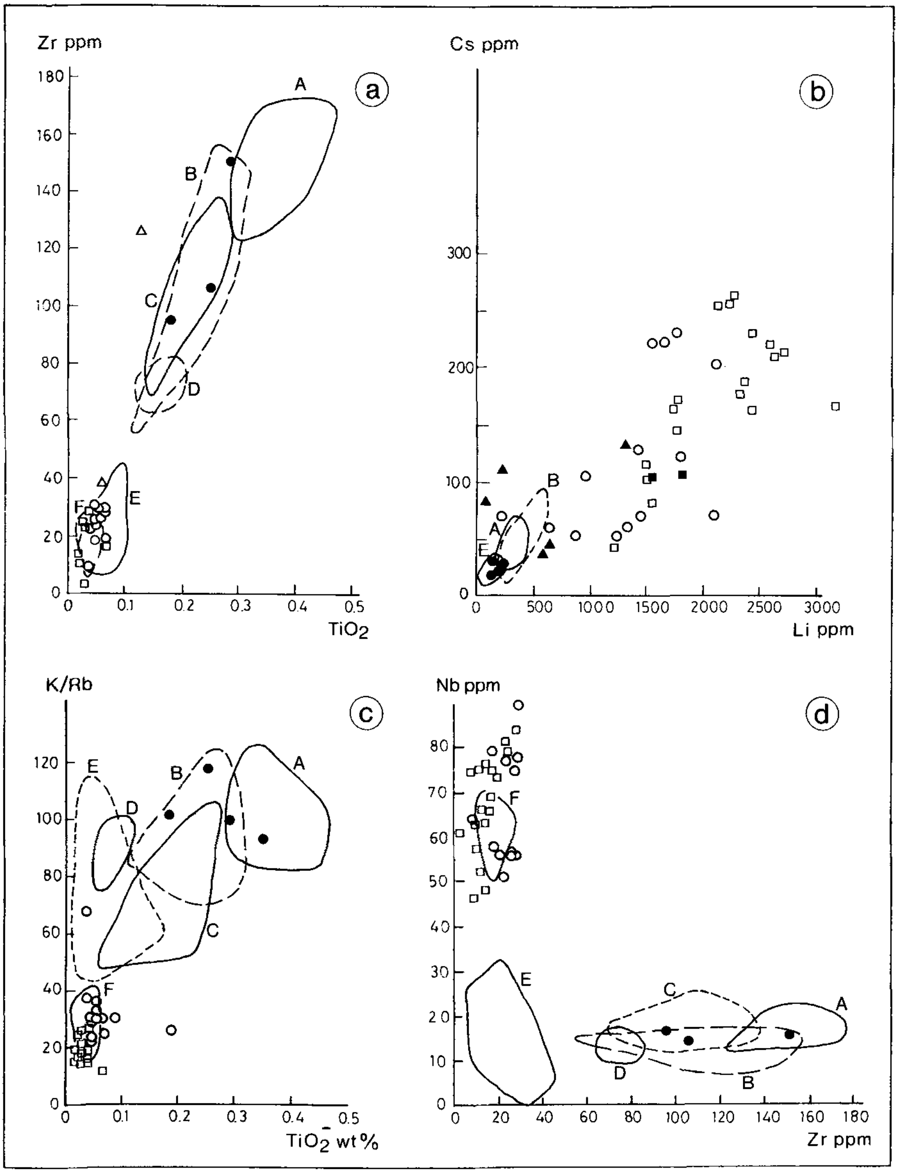

FIG. 3. Bivariate plots of Tregonning Li-mica granites and the composition fields of Li-mica granites from St. Austell and various biotite granites. Biotite granites: A = Dartmoor and St. Austell megacrystic granites; B = Carnemenellis and Isles of Scilly megacrystic granites; C = Dartmoor poorly megacrystic and femic-poor granites; D = Inner granite of Carnmenellis; E = microgranites from Carnmenellis and Isles of Scilly. Filled circles = Godolphin granite. Li-mica granites: F = St. Austell; open circles = Tregonning granite; open squares = aplites and lepidolite leucogranites in roof zone of Tregonning granite. (*a*) Zr (ppm) *vs.* TiO_2 (wt.%)—points belonging to femic-poor granites not included in field C are marked by open triangles; (*b*) Cs (ppm) *vs.* Li (ppm)—filled squares are Li-mica granites from St. Austell and filled triangles are widely scattered data points of microgranites not enclosed by field E; (*c*) K/Rb *vs.* TiO_2 (wt.%); (*d*) Nb (ppm) *vs.* Zr (ppm).

REE data

New data from the Tregonning–Godolphin granite (Table 3) are plotted in Fig. 5. Total *REE* show a significant drop in the series Godolphin granite–Tregonning granite–lepidolite leucogranite, a sequence that compares with the Isles of

Scilly sequence outer granite (and inclusions)–inner granite–microgranite (Stone and Exley, 1989). Ce_N/Yb_N ratios in the Li-mica granites (4.3–1.4) are much lower than the ratio in the Godolphin granite (8.5) which, in turn, is lower than the ratios in the Carnmenellis and Isles of Scilly granites (Stone and Exley, 1989). These low ratios are reflected also in the Li-micas compared with biotites in the Carnmenellis and Isles of Scilly granites, and compare more closely with values in the microgranites.

Discussion

The three possible modes of origin of Li-mica granites referred to at the beginning of this paper, together with modifications, can now be considered in the light of the data and results examined above.

(1) *Magmatic differentiation of biotite granite.* Geochemical data for the already highly evolved biotite granites lead to the conclusion that the series from outer granites to late-stage microgranites resulted from relatively small degrees of fractionation of feldspars, biotite and accessory minerals (Stone, 1987; Stone and Exley, 1989). The question that now arises is, 'Did processes that began in the biotite granites continue also into the Li-mica granites?' Uncertainties attached to the *Kd* values of trace elements like Rb, Sr and Ba (i.e. those occurring in the common minerals—biotite and the feldspars) inhibit quantitative modelling of biotite granite magma fractionation. These uncertainties arise from the diverse values found in the literature, the marked chemical differences between the Fe- and trace alkali-rich S-type Cornubian granites and most of the glasses and fine-grained rocks used to estimate *Kd*s, and the evidence for considerable post-magmatic subsolidus changes. However, the typically high *Kd*s for Sr and Ba in orthoclase and high Kd of Sr in plagioclase will give rise to bulk distribution coefficients (D) that are >1 for a fractionating assemblage composed of these minerals, quartz and a few percent of biotite, e.g. *D*-values of *c.* 2.7 for Ba and 2–5 for Sr using *Kd*s from various sources (Hanson, 1978; Henderson, 1982; Mackenzie *et al.*, 1988; and others) and modal data for TG1 (cf. this paper, Table 2, col. 1) from Stone (1975) and CM1 (cf. this paper, Table 2, col. 4) from Al-Turki (1972). Consequently, Sr and Ba would be expected to fall significantly with marked feldspar and some

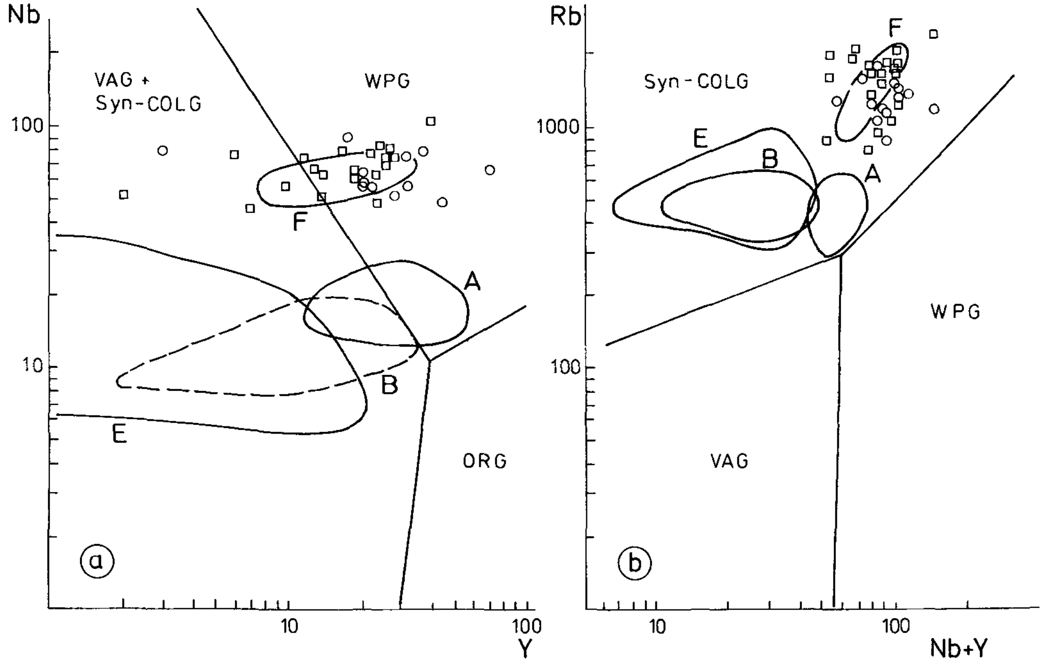

FIG. 4. 'Discriminant plots' based upon Pearce *et al.* (1984). (*a*) Nb (ppm) *vs.* Y (ppm); (*b*) Rb (ppm) *vs.* Nb + Y (ppm). A, B, E, F and other symbols as in Fig. 3. Straight lines separate fields of syn-collision granites (syn-COLG), within plate granites (WPG) and orogenic (ORG) or volcanic arc (VAG) granites.

Table 3. REE data: rocks and trioctahedral Li-micas

	1	2	3	4	5	6	7	8
	Granites					Micas		
Rock Type	TG1	TG2	TG2	TG3	TG3	ZTG2	ZTG2	LTG3
Spec. No.	050	004	051	044	042	004	045	034
La	22.40	7.29	4.40	1.15	0.48	0.79	2.17	0.85
Ce	52.40	6.82	11.50	3.30	0.77	1.56	1.70	3.44
Pr	6.90	1.76	0.90	0.24	0.00	0.00	0.49	0.23
Nd	26.00	6.72	7.60	2.70	0.24	1.36	2.30	2.00
Sm	5.30	1.51	1.50	0.13	0.11	0.30	0.35	0.33
Eu	0.47	0.15	0.05	0.01	0.01	0.03	0.06	0.01
Gd	4.60	1.29	1.50	0.25	0.11	0.45	0.50	0.00
Dy	4.30	1.29	2.00	0.34	0.09	0.25	0.39	0.00
Ho	0.52	0.21	0.31	0.05	0.02	0.05	0.03	0.02
Er	1.30	0.51	0.90	0.21	0.09	0.39	0.00	0.10
Yb	1.60	0.60	1.30	0.20	0.14	0.24	0.12	0.16
Lu	0.38	0.08	0.18	0.03	0.02	0.04	0.02	0.04
tREE	126.17	28.23	32.14	8.61	2.08	5.46	8.13	7.18
Ce_N/Yb_N	8.48	2.94	2.29	4.27	1.42	1.68	3.67	5.56
Sm_N	34.42	9.81	9.74	0.84	0.71	1.95	2.27	2.14
Eu_N/Eu_N^*	0.29	0.32	0.10	0.17	0.28	0.25	0.44	0.18

TG1 = Godolphin biotite granite; TG2 = Tregonning zinnwaldite granite; TG3 = Roof Complex lepidolite leucogranite. Prefixes Z and L refer to zinnwaldite and lepidolite respectively.
tREE = total rare earths; subscript N refers to chondrite normalized values (Evensen et al., 1978). Eu_N^* is the normalized value of Eu obtained by interpolation between Sm_N and Gd_N.
Analysis: by ICP at Department of Geology, Royal Holloway and Bedford New College.

biotite fractionation, whilst Rb, with low *D* (0.27–0.49) would be expected to increase in the residual fluid. Thus, qualitatively, in terms of these elements, it is possible to envisage an evolution from biotite granite magma to Li-mica granite magma, although the Rb *vs.* Sr diagram (Fig. 6), considered below, does not support this. Further, field and other geochemical evidence do not support an origin of Tregonning granite by magmatic differentiation of biotite granite or its microgranite differentiates.

(*a*) Field evidence indicates a major break between biotite granite and Li-mica granite, though such breaks and breaks in variation diagrams do not preclude continuous differentiation below the present exposure level. However, where the roof of biotite granite is exposed, e.g. at Blackadon Tor (GR SW712735) in the Dartmoor pluton, Porthledden (GR SW355321) in the Land's End pluton and elsewhere, and in the mini-pluton of Porthmeor Cove (Stone and Exley, 1984), the end-result of roof differentiation *in situ* is exposed and is never Li-mica granite. At Porthmeor (GR SW425376), there is only slight metasomatic enrichment of the roof hornfelses in trace alkali elements and F, in marked contrast with hornfelses at Tregonning granite contacts (Stone and Awad, 1988).

(*b*) In the St. Austell pluton, the megracrystic Li-mica granite (Type D of Exley and Stone, 1982) does not appear to be an intermediate differentiate between biotite granite and aphyric Li-mica granite, as formerly thought (Exley, 1959). It apparently forms a Li-rich aureole around the aphyric Li-mica granites (Dangerfield *et al.*, 1980; Manning and Exley, 1984) that marks late-stage metasomatic modification of earlier biotite granite. More recently, Hill and Manning (1987) have demonstrated the occurrence of additional rock types and intrusions to those formerly recognised and shown that relations between them are complex. However, despite their evidence for a magmatic sequence from biotite granite, through equigranular biotite granite, globular quartz granite, tourmaline granite, aphyric granite, to topaz (i.e. Li-mica) granite, the intermediate members between the biotite granites and the Li-mica granite show evidence of metasomatism and the occurrence of brown mica and topaz: also, as pointed out in the following paragraph, Manning and Hill (1990) now consider that the Li-mica granites are not petrogenetically related to the biotite granites.

(*c*) Manning and Hill (1990) give two lines of evidence against the derivation of Li-mica granite from biotite granite magma by fractional crystallisation. Firstly, they point out that experimental data indicate that fractionation of biotite or other hydrous aluminosilicates will remove F and leave an evolved liquid depleted rather than enriched in F. Of course, this will be true where biotite or other minerals with *Kd*s greater than 1, e.g. topaz and tourmaline, are the principal phases involved: whether bulk *D* is greater or less than 1 will depend upon the total assemblage undergoing fractionation, so that dominant quartz–feldspar fractionation would be expected to result in D < 1 and an increase in F in the residual melt. Secondly, and of greater importance, there is an overall lack of consistent trends in many bivariate plots and the common absence of a continuum of compositions, e.g. Li_2O *vs.* SiO_2, Nb *vs.* Ga and Zr *vs.* Nb. Stone (1975) also considered that chemical and petrographic differences between biotite and Li-mica granites were so great they they must have had separate petrogeneses.

Some of the variation patterns examined in the this study (e.g. Zr *vs.* TiO_2, Fig. 3*a*) could suggest a continuum between biotite granites and Li-mica granites, but most indicate major breaks and/or

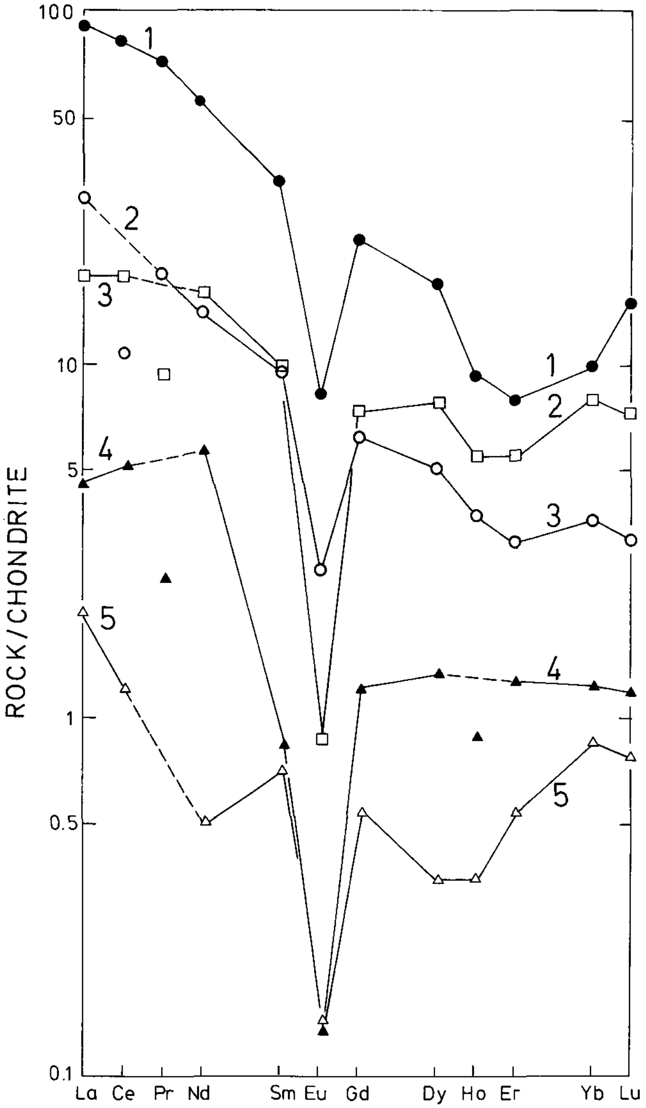

FIG. 5. Chondrite-normalised *REE* plots for rocks from the Tregonning–Godolphin pluton using chondrite values in Evensen *et al.* (1978). Numbers correspond with those in Table 3.

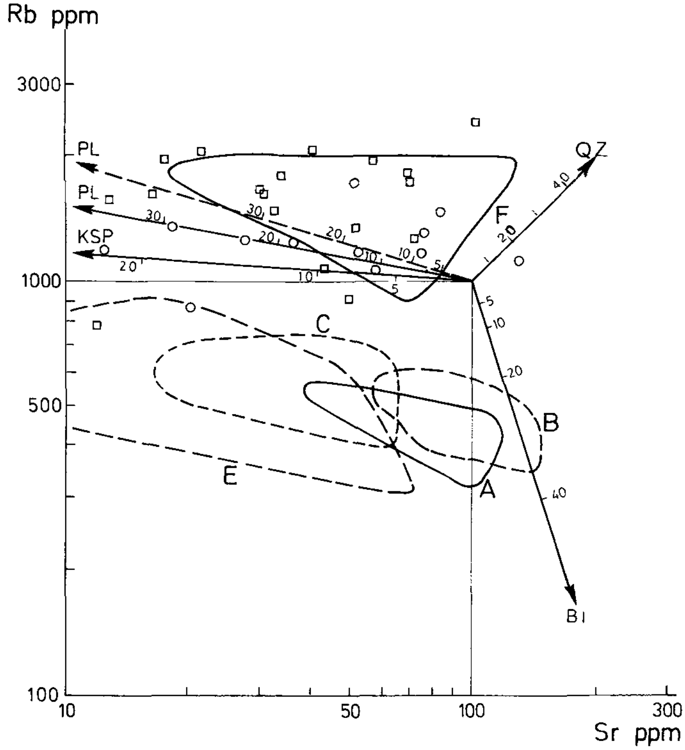

FIG. 6. Rb (ppm) *vs.* Sr (ppm) plot. Symbols as in Fig. 3. Solid straight lines show fractionation vectors using the Rayleigh fractionation equation for biotite (BI), plagioclase (PL) and K-feldspar (KSP). *Kd*s used for Rb and Sr are 3.4 and 0.24 for BI, 0.09 and 6 for PL, and 0.38 and 9.4 for KSP. The dashed line is the PL vector for *Kd*(Rb) = 0.04 and *Kd*(Sr) = 4.4. If the *Kd*(Sr) for KSP is lowered to 4, the fractionation vector almost coincides with the solid line for PL fractionation. Numbers give percent solid phase (1–F).

differences in trend between the two data sets. For example, the completely separate composition fields in the Nb *vs.* Y and Rb *vs.* Nb + Y plots (Fig. 4) and the Nb *vs.* Zr plot (Fig. 3*d*, see also Manning and Hill, 1990, Fig. 9) suggest different evolutions. The trend of diminishing Zr at near constant Nb in the biotite granites suggests zircon fractionation, but not that of Nb-bearing minerals (e.g. rutile, and possibly, columbite–tantalite). Zircon is commonly included, together with monazite, in biotite and, presumably, is fractionated with its host. The marked enrichment of Nb in the Li-mica granites is common in many Li-mica granite pegmatites and must signify increased amounts of tiny grains of Nb-bearing accessory minerals (e.g. columbite–tantalite and/or tapiolite etc.) in the matrix, rather than included in Li-micas (cf. Stone *et al.*, 1988).

Particularly strong evidence for different petrogeneses of biotite and Li-mica granites is indicated by the diminished contents of Cs and Li (Fig. 3*b*) in the microgranites relative to the biotite granites compared with marked enrichment in the Li-mica granites. Rb *vs.* Sr (Fig. 6) shows a typical trend of granitoid differentiation for the biotite granites. This suggests only minor roles of quartz and biotite and reflects dominant feldspar fractionation with, perhaps, some 5–15% biotite fractionation (depending upon the feldspar *Kd*s used). Significant quartz fractionation would tend to increase the negative slope of the biotite granite trend: however, this trend has a slope close to and just below that of the feldspars, so that quartz fractionation is relatively unimportant at this stage. The Li-mica granites plot in an entirely different region at much higher Rb than the biotite granites but have a similar spread of Sr values. Of course, quartz–feldspar fractionation of biotite granite magma could yield Li-mica granite compositions in the Rb *vs.* Sr plot (as already indicated), but this is *not* the trend within the biotite granite suite.

Data points in a Sm_N *vs.* Ce_N/Yb_N plot (Fig. 7) for the Carnmenellis and Isles of Scilly plutons

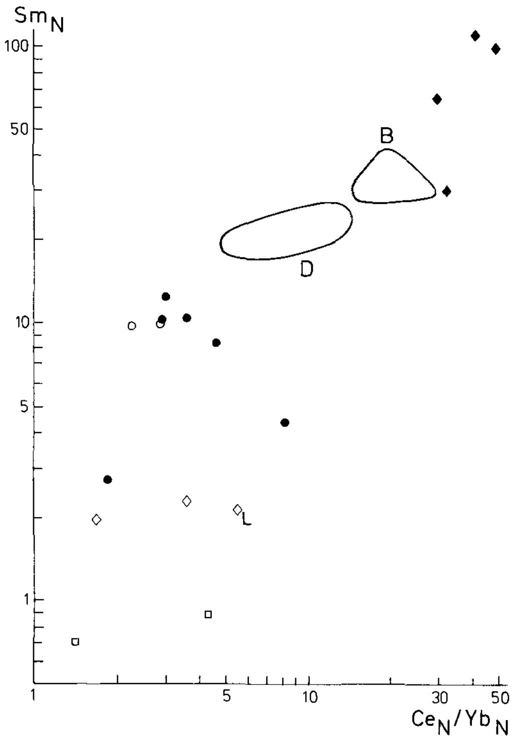

FIG. 7. Chondrite-normalised Sm_N *vs.* Ce_N/Yb_N plot. Symbols B, D, open circles and open squares as in Fig. 3. Filled circles = microgranites; filled diamonds = biotites from the Carnmenellis and Isles of Scilly plutons (Stone, 1987; Stone and Exley, 1989); open diamonds = L-micas zinnwaldite and lepidolite (L).

show a general trend from biotite, through outer megacrystic granites to inner granites and late differentiates (see also Fig. 8 in Stone, 1990) as both Sm_N and total *REE* fall and slope (Ce_N/Yb_N) flattens. This pattern is taken to indicate some biotite fractionation accompanied by *REE* accessory minerals (Stone, 1987; Stone and Exley, 1989). Although data points for zinnwaldite granites fall close to microgranite compositions, their roof differentiates lie at lower total *REE* but with little change in slope. Data points for Li-micas lie at lower values of Sm_N than those of zinnwaldite granites, between the latter and late roof differentiates. It follows that (*a*) because these micas have such low *REE* concentrations, much of the *REE* content of the zinnwaldite (Tregonning) granite occurs in accessory minerals in the matrix rather than any included in the micas, (*b*) the *in situ* differentiation of zinnwaldite granite to lepidolite leucogranite results mainly from direct accessory mineral fractionation rather than Li-mica fractionation; and (*c*) these micas could not have been derived directly from biotite by metasomatism, despite evidence of such changes at contacts between Li-mica granite and pelitic country rocks (Stone, 1984; Stone and Awad, 1988), but probably nucleated within a Li-mica granite at a late- to post-magmatic stage of crystallisation (Stone, 1984; cf. also Henderson *et al.*, 1989, for micas in the megacrystic Li-mica granite in the St. Austell pluton).

(2) *Partial melting of earlier-formed biotite granite.* Modelling the partial fusion of a typical Cornubian biotite granite is fraught with the problems referred to above in connection with fractional crystallisation. However, using Rb, Sr, and Ba again, it is clear that partial melting will concentrate Rb ($D < 1$) in the melt and Sr and Ba ($D > 1$) in the solid residue. By using suitable *Kd*s (e.g. taking those given in Henderson, 1982), it is possible to arrive at a melt composition that approaches that of typical Li-mica granite in terms of Rb and Sr, although Ba tends to be too high. For example, simple batch melting (say 10%) of megacrystic biotite granite (early Carnmenellis or Godolphin, Table 2, cols 1 and 4; modal data from Al-Turki, 1972, and Stone, 1975) could generate a melt with 1100–1300 ppm Rb and 20–40 ppm Sr, values that approximate those in the Tregonning granite, although Ba at 110–130 ppm is too high. Smaller degrees of melting lead to even larger amounts of Rb in the melt (*c.* 1500 ppm for 1% melt) but only marginally lower amounts of Sr and Ba. Partial melting of quartz and feldspars alone could produce a melt (removed from residue and prevented from re-equilibrating) with even higher Rb and lower Sr and Ba. However, such a simple model does not account for the high covariation between Rb and Li in biotite (Stone *et al.*, 1988) despite disparate *Kd*s (cf. Henderson, 1982, Table 5.2b), or the high covariations between these elements and F, if biotite breakdown is not involved to any extent, nor would the amount of biotite present in typical biotite granite produce a sufficient volume of trace alkali- and F-enriched melt if it were decomposed.

(3) *Partial fusion of lower crustal residual rocks.* It has been suggested that high-temperature partial melting of a depleted I-type granite source in the lower crust could yield a metaluminous or peralkaline granite magma (Collins *et al.*, 1982; White and Chappell, 1983; Clemens *et al.*, 1986) although some authors think this unlikely (e.g. Creaser *et al.*, 1991). This idea has been extended to Li-mica granite by Manning and Hill (1990), who suggested an origin by partial melting of residual S-type granite source rocks. They noted

that, apart from higher Li and P_2O_5 contents, these rocks have similar chemical features to topaz rhyolites from North America which, in turn, have some chemical characteristics of A-type granitoids (Burt *et al.*, 1982; Christiansen *et al.*, 1983). On the other hand, compared with typical metaluminous A-type biotite granitoids and topaz rhyolites, the Li-mica granites are richer in Al_2O_3, P_2O_5, trace alkali elements and F, and poorer in Zr, Y, Ba, U, Th and *HREE* (Christiansen *et al.*, 1983; Stone, 1990). Of course, peralkaline granites and rhyolites can contain considerable amounts of the trace alkalis and F, although these are usually lower than in the Li-mica granites (Christiansen *et al.*, 1983, Fig. 3). The former also have significantly lower Al_2O_3 (and ASI) and higher, Zr, tFeO and total *REE* than the latter.

Residual rocks in the lower crust left after extraction of S-type granite melt would be depleted in incompatible elements to varying degrees, depending upon the temperature and the melting process (batch, incremental etc.) and its extent. If originally pelitic or semipelite, the residuum would be composed of such minerals as plagioclase, cordierite, biotite (orthopyroxene at higher temperatures), amphibole, sillimanite, garnet (almandine), iron oxides and, perhaps, some quartz (Green, 1976; Winkler, 1979; Clemens and Wall, 1988).

Recent work by Vielzeuf and Holloway (1988) has suggested that substantial melting in H_2O-undersaturated pelitic rocks at deep crustal levels can occur in the region of 850–875 °C as a result of biotite breakdown in the reaction (also Clemens and Wall, 1981):

$$\text{biotite} + \text{Al silicate} + \text{plagioclase} + \text{quartz} = \text{garnet} + \text{K-feldspar} + \text{liquid}$$

This reaction will leave an aluminous granulite residuum of quartz, garnet, sillimanite and plagioclase (or K-feldspar). Vielzeuf and Holloway (1988) also point out that the large amount of melt produced (up to 50%) will tend to buffer the temperature, so that temperatures > 850 °C will be uncommon in the partial fusion of a pelitic lower crust; further partial melting is only possible at higher temperatures after the extraction of an earlier melt. However, once biotite has broken down in the above or similar reactions, the main source of trace alkali elements and F has gone (in this case, into the earlier melt), so that the biotite-free granulite residuum is unlikely to provide, on its own, a fertile source for Li-mica granite.

Thus, differentiation of biotite granite magma, partial fusion of biotite granite (as presently exposed) and partial fusion of biotite-free residual lower crust are unlikely direct sources of Li-mica granite magma. However, two further means of generating such a magma remain: these are (4) partial fusion of residual lower crust enriched in biotite, and (5) trace alkali and F metasomatism accompanying an input of volatiles from an outside source.

(4) *Partial fusion of biotite-enriched lower crustal residual rocks.* Although biotite can crystallise readily from granitic melts, the low solubilities of Mg and Fe in a H_2O-undersaturated peraluminous granite melt (Puziewicz and Johannes, 1990) suggest that much biotite, along with cordierite and almandine–garnet, is restite (White and Chappell, 1977). Of course, most of the biotite and other restite minerals will remain behind and form an important part of the residuum.

Some evidence for biotite granite magma generation close to 800 °C in the lower crust is provided by Fe–Mg distribution in co-existing almandine-rich garnet, cordierite and biotite in the Dartmoor granite (Stone, 1988). These are considered to be restite minerals carried by the rising host magma: if so, it is evident that biotite is stable at such temperatures in the lower lower crust, and that much of it can be generated in a residuum derived particularly from semipelitic rocks or even greywacke. The latter is abundant in orogenic belts in the crust and could provide large volumes of biotite granite melt (Miller, 1985; White and Chappell, 1988) and leave behind a biotite-rich residuum which, in turn, could contribute a fertile source for subsequent extraction of trace alkali- and F-enriched melt.

If the trace alkali elements (particularly Rb and Cs) and F are preferentially taken up by biotite compared with co-existing melt, as suggested by limited experimental work and distribution coefficients, residual biotite could provide an important source for these elements. With fluid-absent breakdown of biotite and movement into granulite facies conditions in a later episode of partial melting, trace alkalis, and F would be released into the liquid (cf. Burt *et al.*, 1982), which would contain 'femic' minerals like orthopyroxene and others listed above, together with anorthite-rich plagioclase. However, in order to attain Li-mica granite composition, magma derived from such a residuum would have to undergo marked fractionation of femic-rich phases, alkali feldspar and anorthite, and become markedly enriched in Na.

(5) *Metasomatism associated with volatile input.* Addition of a volatile phase containing the trace alkali elements, F, H_2O, and possibly Na, Nb and Sn, could efficiently transform biotite granite as presently exposed or, more likely, residual crustal

material that provided an earlier biotite granite or, indeed, more basic granitoid with a composition between these two extremes into Li-mica granite compositions without the need to invoke fractionation of biotite granite magma, partial melting of biotite granite or even partial melting of residual lower crust.

Metasomatism would involve the transformation of biotite to Li-mica, a change actually observed in xenoliths and at contacts with the Tregonning granite (Stone, 1984; Stone *et al.*, 1988), and Na enrichment by removal of Ca from and addition of Na to plagioclase (Manning and Exley, 1984) and, perhaps, some Na–K exchange in alkali feldspar. Such initial transformation could generate compositions that correspond with a real (low-temperature) minimum in the QZ–AB–OR plane in the F-rich system QZ–AB–OR–H_2O–F (Manning, 1981) that would enhance fusion and lead to the (simultaneous) production of a Li-mica granite melt. As bulk compositions of Li-mica granites are close to minima in this F-rich system, only relatively small increases in temperature are needed to produce a large amount of low-viscosity melt that could rise rapidly to high crustal levels.

Conclusions

The discussions above indicate that Li-mica granites are unlikely to have been derived from the biotite granites by differentiation or partial fusion, or from partial fusion of lower crustal residual biotite-free granulite. Partial fusion of biotite-enriched lower residual crust could generate the trace alkalis and F, but would be accompanied by marked Fe-Mg and Ca enrichment and deficiency in Na. The simplest petrogenetic model for the Tregonning granite is that of a more basic granitoid (or residual crustal rock) transformed through mainly alkali and F metasomatism into Li-mica granite composition, either at a high crustal level, but below the present erosion level, or at a deeper level.

The source of trace alkalis and F required for transformation could have been the volatiles left after the crystallisation of biotite granite magma, implying high-level generation of Li-mica granite (as formerly indicated by Stone, 1975), or be mantle derived. However, evolution of biotite granite leads to microgranites that show relatively little or no increase in F or trace alkalis (Table 2), so that a source outside this system is more feasible. A subcrustal source (mantle or crust/mantle interface) is more likely on account of its higher energy environment for generating volatiles in and metasomatism above an active mantle, and the close temporal relationship between the alkali-rich Permian volcanic rocks and the granites (Leat *et al.*, 1987; Thorpe, 1987).

This contrasts with the generation of the biotite granites, which are believed to represent products of fractionated magmas largely derived initially by direct partial melting of lower crustal semi-pelite or metagreywacke, though again, perhaps with some addition from the mantle (Exley *et al.*, 1983; Stone and Exley, 1986; Thorpe, 1987). All the petrogeneses considered here imply the generation of Li-mica granite magma after the biotite granites. Field relations support this, although isotopic ages (Darbyshire and Shepherd, 1985, 1987) do not clearly discriminate between the biotite granites and the Tregonning granite.

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