Lithium in carbonatites – consequence of an enriched mantle source?

ALAN F. COOPER, LORRAINE A. PATERSON

Department of Geology, University of Otago, Dunedin, New Zealand

AND

DAVID L. REID

Department of Geological Sciences, University of Cape Town, Rondebosch 7700, South Africa

Abstract

The rare Li-mica taeniolite is described from the Dicker Willem carbonatite complex, Namibia, and from the Alpine carbonatitic lamprophyre dyke swarm at Haast River, New Zealand. At Haast River, taeniolite occurs in sodic and ultrasodic fenites derived from quartzo-feldspathic schists and rarely in metabasites, adjacent to dykes of tinguaite, trachyte and a spectrum of carbonatites ranging from Ca- to Fe- rich types. In Namibia, taeniolite is present in potassic fenites derived from quartz-feldspathic gneisses and granitoids at the margin of an early sövite phase of the complex and in a radial sövite dyke emanating from this centre.

The occurrence of taeniolite in these totally disparate carbonatite complexes, together with examples of lithian mica from other carbonatite complexes worldwide, raises the question of the status of Li as a 'carbonatitic element'. We argue that lithium is not a consequence of crustal assimilation or interaction, but reflects the geochemical character of the magmatic source. Li, an overlooked and little-analysed element, may be an integral part of metasomatic enrichment in the mantle, and of magmas derived by partial melting of such a source.

KEYWORDS: lithium, taeniolite, carbonatite, fenite, fenitization, mantle metasomatism.

Introduction

CARBONATITES are enriched in Ba, Sr, *REE*, Nb, Ti, Zr, Th, P and F (Gold, 1966; Henrich, 1966; Woolley and Kempe, 1989), elements normally present in only minor or trace amounts in other igneous rocks. However, this acknowledged geochemical fingerprint is somewhat blurred by the presence in other carbonatite complexes around the world of economic concentrations of Fe and Mn (Heinrich, 1966), V (Deans, 1981), while the Phalaborwa complex of north-eastern Transvaal, South Africa, is not only a major Cu deposit, but also yields by-products of Zr, P, U, Ni, Au, PGE, Ag, Se and Te (Palabora Mining Company Ltd, 1976; Ericksson, 1989). However, the economic separation of many of these elements is only possible with the large volumes of ore being processed at Phalaborwa.

Clearly the mechanisms of formation of carbonatites (either by direct partial melting of mantle, or separation from a silicate parent magma), coupled with the imperfectly understood processes of magmatic differentiation of carbonatite magmas and secondary enrichment processes, are capable of generating a variable geochemical signature.

This paper describes the occurrence of the rare Li-mica, taeniolite, in two carbonatite complexes of grossly dissimilar lithological association, geochemistry and tectonic environment. These and other occurrences worldwide raise two main questions:(1) should Li be included as a common (or characteristic?) element of carbonatite parageneses? and (2) does the enrichment form as a consequence of carbonatite fractionation, assimilation or contamination processes, or does it reflect the geochemical signature of the mantle source?

General geology

Dicker Willem, Namibia

The Dicker Willem carbonatite complex forms a circular, 3km diameter, inselberg rising 600m above the desert plain of the pre-Namib in south-west Namibia (Cooper, 1988). Carbonatites dominate this central-type complex, with an early, concentrically banded sövite containing xenoliths of ijolite, and syenitic, fenitised granite. Emplacement of the sövite is characterised by potassic fenitization resulting in metasomatic alteration of country rock for distances of up to several hundred metres from the intrusive contact. Granitic gneisses and granitoids are converted to K-feldspar – aegirine fenites. Dykes and plugs of K-feldspar trachyte intrude the fenites, and the entire sequence has been intensely brecciated during the subsequent intrusion of steeply dipping sheets of alvikite. Alvikites, commonly dolomite-phyric (Cooper and Reid, 1991), form the dominant rock type of the complex with remnants of the early sövite body represented by isolated xenoliths and screens. Alvikites are in turn crosscut by thin dykes of ferrocarbonatite with the intrusive history terminated by emplacement of high level breccia pipes and tuffisite dykes. K-Ar and Rb-Sr dating suggest an Eocene age of 49 ± 1 Ma (Reid *et al.*, 1990) for intrusion of the sövite phase of the complex.

Haast River, New Zealand

Carbonatites of the Haast River area, south Westland, New Zealand, form a volumetrically minor (< 1% by volume) component of a lamprophyre-dominated dyke swarm, intrusive into country rock schists. Intrusion has been dated by U-Pb and Rb-Sr techniques at 21−25 Ma, coeval with the transtensional propagation of the nearby Alpine Fault plate boundary through the South Island, New Zealand (Cooper *et al.*, 1987). Although camptonite dykes are dominant, intrusives range from ouachitite, through sodalite tinguaite and trachyte to carbonatite. Carbonatites in turn range from calcite-rich types through ankerite- and siderite- dominated ferrocarbonatites, with extreme Ba enrichment manifest in one sill by a mineralogy dominated by norsethite (Cooper, 1971, 1986; Paterson, 1992). Fenitization of the country rock is a characteristic feature of all carbonatites, but is also developed, although less extensively, along the margins of trachytes, tinguaites and lamprophyres. Although fenitization is variable from one dyke type to another, the typical end-product of metasomatic alteration of quartzo-feldspathic schist is an aegirine–albite fenite (Cooper, 1971, 1986; Paterson, 1992). In contrast to the Dicker Willem potassic metasomatism, fenitization at Haast River is characteristically sodic. According to Woolley (1969, 1982) sodic fenitization is typically associated with deep seated carbonatitic activity, while potassic alteration is generally ascribed a shallow origin. These inferences are compatible with the subvolcanic quench textures observed at Dicker Willem (Cooper and Reid, 1991) and the major (> 20 km) uplift that has occurred along the Alpine Fault plate boundary in the past 5−10 Ma, bringing the Alpine lamprophyre dyke swarm from deeper crustal levels.

Field occurrence and petrography

Dicker Willem

Taeniolite occurs in a 5.5m wide, radial sövite dyke, intruding granitic country rock at the northern contact of the complex. The carbonatite is composed of strontian calcite, an aegirine-rich clinopyroxene, taeniolite, pyrochlore, apatite and magnetite, with the taeniolite typically occurring in grains up to 0.8 mm in length intergrown with aggregates of aegirine. Taeniolite is a colourless, micaceous mineral, that can be distinguished petrographically from muscovite only by its pseudo-uniaxial character and slightly lower refractive indices and birefringence. In this locality the country rock granitoid has been intensely brecciated, fenitized and pervasively veined by dykelets of carbonatite and trachyte. A sample of fenitized granitoid collected 7 m from the carbonatite contact also contains taeniolite, here in 0.2 mm grains in veinlets and intergranular films and patches of carbonate associated with turbid, metasomatically-recrystallised K-feldspar, aegirine and a pale blue-green, manganoan magnesio-arfvedsonite. Quartz and any mafic minerals of the parent granite have been replaced, the plagioclase recrystallized to K-feldspar, but original microcline, although fractured and net-veined, is still very fresh. Granitoids on the south-east margin of the complex also contain taeniolite, occurring with metasomatic K-feldspar, aegirine and calcite as a network of crosscutting veinlets, and intergranular patches pseudomorphing quartz and plagioclase of the parent granitoid.

Haast River

At Haast River the dominant country rock is a greywacke-derived, quartzo-feldspathic schist, ranging in metamorphic grade from greenschist to amphibolite facies. At these metamorphic grades the metagreywackes are typically segregated into quartz–albite and mica-rich laminae. During fenitization, biotite, chlorite and quartz are progressively replaced by blue amphibole, aegirine and albite, with the

foliation acting as channelways for metasomatic fluids. As a consequence of this structural control, fenites retain a banded appearance.

Tabular (40 × 80 μm) crystals of colourless taeniolite occur associated with metasomatic albite, and more commonly, intergrown with aegirine, in relict quartz-albite segregation bands of the parent schist. Taeniolite occurrence is not specific to dyke-type, occurring in fenites associated with tinguaite, as well as carbonatites which range from calcitic to sideritic varieties. Taeniolite occurs only rarely in metabasite-hosted fenites.

Mineralogy

Worldwide

Taeniolite is a rare lithian mica with an ideal formula $KLiMg_2Si_4O_{10}(F,OH)_2$. It was first described from nepheline syenite pegmatites of the Narsarsuk alkaline complex, southern Greenland by Flink (1901). As expected from its potassic, alumina-free end-member composition, or alumina-poor composition in complex taeniolite – muscovite – phlogopite solid solutions, taeniolite occurs in alkaline rocks, particularly in syenites, or in associated alkaline metasomatites (fenites). It has been reported at Khibina, Lovozero and Baikal (Gerasimovskii, 1965; Semenov, 1959; Vlasov et al., 1959), at Azov in the Ukraine (Eremenko and Valter, 1963), near, and at Magnet Cove, Arkansas (Stone and Milton, 1976; Smith, 1989), and at Mont Saint-Hilaire, Quebec (Mandarino and Anderson, 1989). Taeniolite occurs in a late-stage pegmatitic clot in a mafic, alkaline diatreme at Coyote Peak, California (Erd et al., 1983), and in metasomatites of the Karelia region (Rumyantseva et al., 1984). In addition to these and other, mainly Russian, occurrences in alkaline silicate igneous rocks, taeniolite or unspecified Li-mica has been reported from carbonatite at the Mbeya complex, Tanzania (Fick and van der Heyde, 1959), Sarfartoq, western Greenland (Secher and Larsen, 1980) and in dolomite of carbonatitic affinity from Bayan Obo, China (Le Bas et al., 1992), in carbonatite-related fenites at Magnet Cove (Miser and Stevens, 1938) and in albitites associated with carbonatites in the northern Baikal area (Gerasimovskii, 1965).

Dicker Willem and Haast River

In the Dicker Willem and Haast River fenites, taeniolite is generally so fine grained and of such low abundance that separation for optical and crystallographic characterisation and wet chemical analysis is impractical. Only two taeniolites were separated, AFC154, a carbonatite from Dicker Willem contains the coarsest grained taeniolite observed in either complex, while OU58764, a quartzo-feldspathic schist-derived fenite from Haast River, contains 5.7% modal taeniolite.

Taeniolite from both AFC 154 and OU58764 is colourless and pseudo-uniaxial, with 2V close to zero. Refractive indices were determined by oil-immersion techniques using monochromatic, Na_D light; values for AFC 154 are: $\alpha = 1.528$, $\beta = \gamma = 1.550$, and OU58764: $\alpha = 1.530$, $\beta = \gamma = 1.552$, all values ± 0.002. Debye-Scherrer powder patterns closely match spacings given by Miller and Johnson (1962) for synthetic, 1M polytype, fluormica. Cell dimensions for AFC 154 of: $a_0 = 5.213 \pm 0.002$, $b_0 = 9.040 \pm 0.003$, $c_0 = 10.134 \pm 0.003$ Å, $\beta = 99°$, $V = 470.79$ Å^3 and OU58764: $a_0 = 5.230 \pm 0.002$, $b_0 = 9.043 \pm 0.005$, $c_0 = 10.152 \pm 0.006$ Å, $\beta = 99°$, $V = 474.23$ Å^3 were determined by least-squares refinement of powder diffraction data using the programme of Benoit (1987).

The two mica separates (and whole rock powders) were dissolved in HF and $HClO_4$ and the Li contents determined by atomic absorption spectrometry using solutions prepared from analytical grade Li_2CO_3 as standards. The standard calibration was checked using international rock standard SY3, our analyses yielding an average Li content of 84 ppm, compared to the recommended value of 92 ppm (Abbey, 1983). Major element compositions of all taeniolites were determined by wavelength dispersive analytical techniques on a JEOL JXA-5A electron microprobe, using standard operating conditions (15 kV accelerating voltage, 20nA current on MgO standard), procedures (moving stage beneath electron beam to minimise sample damage) and data processing (ZAF). With the exception of OU58764 and AFC154, Li contents were calculated stoichiometrically, assuming 22 positively charged cations in the formula unit (equations in Paterson, 1992). Representative analyses of taeniolites are reported in Table 1, together with comparative analyses from similar parageneses worldwide.

Taeniolites from Dicker Willem contain only minor contents of the elements Ti, Al, Fe and Na that are not essential to the end member composition. However, analyses from Haast River and elsewhere around the world show considerably greater variation. Some of this variation could be attributed to physical interlayering of taeniolite with muscovite, or even biotite, especially in rocks like the Haast River fenites, where metasomatic taeniolite and relic, metamorphic muscovite and biotite occur together. However, in other cases the chemical variability is clearly the result of taeniolite–biotite solid solution, details of which are discussed by Paterson (1992).

TABLE 1. Taeniolite analyses

	Dicker Willem			Haast River				Worldwide		
	1	2	3	4	5	6	7	8	9	10
SiO_2	57.44	56.11	58.03	59.61	56.75	58.04	59.43	53.50	58.82	52.63
TiO_2	0.01	0.23	0.04	0.03	0.07	0.07	0.07	1.06	0.11	n.a
Al_2O_3	0.07	0.12	0.35	0.48	2.67	8.26	0.45	3.00	1.29	4.60
Fe_2O_3	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a	0.40	0.55
FeO	0.43*	2.12*	0.20*	0.27*	1.25*	1.33*	0.66*	3.35	0.24	n.a
MgO	20.38	18.54	19.89	19.89	20.83	14.83	19.97	18.30	19.18	23.25
MnO	0.03	0.00	0.07	0.01	0.02	0.01	0.03	0.21	0.00	0.03
CaO	0.00	0.00	0.00	0.00	0.01	0.02	0.03	n.a	0.00	2.35
Na_2O	0.40	1.49	0.72	0.00	0.33	0.00	0.00	0.27	0.64	0.19
K_2O	11.41	11.16	11.39	11.12	10.63	10.05	11.57	11.30	10.44	8.90
Li_2O	3.06	(3.65)#	(3.65)#	2.81	(2.82)#	(2.89)#	(3.58)#	2.40	3.10	2.52
F	9.09	9.10	9.70	9.51	8.42	10.31	8.32	6.30	8.56	7.21
H_2O	n.a.	n.a.	n.a.	n.a.	(0.46)#	n.a.	(0.54)#	n.a.	0.59	n.a.
Total	102.32	102.51	103.96	103.73	104.26	105.81	104.65	99.69	103.37	102.23
Less O = F	3.83	3.83	4.09	4.00	3.55	4.34	3.50	2.65	3.60	3.02
Total	98.49	98.68	99.87	99.73	100.71	101.47	101.15	97.04	99.77	99.21

Cation proportions based on 22 positive charges per formula unit

Si	3.959	3.904	3.946	4.029	3.825	3.840	3.969	3.762	3.973	3.571
Al	0.006	0.010	0.028	0.000	0.175	0.160	0.031	0.238	0.027	0.368
Fe^{3+}	(0.006)#	(0.010)#	(0.028)#	0.000	0.000	0.000	0.000	–	0.000	0.028
Al	0.000	0.000	0.000	0.038	0.037	0.484	0.004	0.010	0.076	0.000
Ti	0.001	0.012	0.002	0.002	0.004	0.003	0.004	0.056	0.006	–
Fe^{3+}	–	–	–	–	–	–	–	–	0.020	0.000
Fe^{2+}	0.025	0.123	0.012	0.015	0.070	0.074	0.037	0.197	0.014	–
Mn	0.002	0.000	0.004	0.001	0.001	0.001	0.002	0.013	0.000	0.002
Mg	2.094	1.923	2.017	2.004	2.092	1.463	1.988	1.918	1.931	2.351
Li	0.848	1.020	0.974	0.764	0.765	0.768	0.961	0.679	0.842	0.688
Ca	0.000	0.000	0.000	0.000	0.001	0.001	0.002	–	0.000	0.171
Na	0.053	0.201	0.095	0.000	0.043	0.000	0.000	0.037	0.084	0.025
K	1.003	0.991	0.988	0.959	0.914	0.848	0.986	1.013	0.900	0.770
F	1.981	2.000	2.086	2.033	1.795	2.157	1.757	1.401	1.829	1.547
OH	–	–	–	–	(0.205)#	–	(0.243)#	–	0.266	–

1	AFC154 Sövite (Sr-calcite + aegirine + taeniolite).
2	AFC195B Fenitized granite
3	AFC152 Fenitized granite
4	OU58764 Quartzo-feldspathic fenite (dyke unknown)
5	OU58780 Quartzo-feldspathic fenite adjacent to tinguaite dyke
6	OU58673 Quartzo-feldspathic fenite adjacent to calcite carbonatite
7	OU58740 Quartzo-feldspathic fenite adjacent to siderite carbonatite
8	Coyote Peak, California (Erd *et al.* 1983)
9	Magnet Cove, Arkansas (Miser and Stevens, 1938)
10	Karelia, Sweden (Rumyantseva *et al.* 1984)
*	Total Fe as FeO
#	Li_2O, H_2O, Fe^{3+} estimated stoichiometrically (see text).
n.a.	not analysed

Geochemistry

Haast River

Lithium concentrations in fenites from the Haast River area range from <10 ppm to 819 ppm. Highest concentrations occur in fenitic aureoles of calcite carbonatites, but fenites associated with dolomitic and sideritic carbonatites are also enriched, reaching concentrations of 239 and 255 ppm respectively. The fenite with the highest Li content also has the highest modal taeniolite (11.5%), supporting the petrographic observation that taeniolite appears to be the dominant, or probably the only, Li-bearing phase.

In fenite aureoles where several specimens have been analysed, the highest Li concentrations generally occur adjacent to the contact with carbonatite and decrease systematically towards low grade fenites at the distal edge of fenitization. For example, in samples taken at distances of 10, 15 and 100 cm from a thin siderite dyke (Li content 3 ppm), the Li concentration decreases from 255 ppm (in taeniolite-bearing sample OU 58740, Table 1) to 114 ppm in a subjacent fenite, before falling to 14 ppm in petrographically unaltered schist, 1m from the carbonatite contact. Only rarely is there a departure from this pattern of Li distribution, although a similar siderite dyke, spatially associated with norsethite-rich carbonatites, shows the reverse of the trend, with Li concentrations increasing towards low-grade fenites. While this variation may reflect redistribution of Li by later stage processes (Bailey and Gwozdz, 1994), primary differences in the Li content of the host dyke are more likely to influence the composition of the coexisting fenitizing fluid. Lithium concentrations in carbonatites range from 1 to 24 ppm (avg. 4 ppm, $n = 35$), and three trachyte/tinguaite dykes range from 65 to 80 ppm. The generally low concentrations of Li in the parent carbonatite or tinguaite compared with associated fenites suggest that Li is very strongly partitioned into the fenitizing fluid and fixed by Mg as taeniolite in quartzo-feldspathic schist.

Dicker Willem

Taeniolite-bearing carbonatite dyke AFC154 contains the association of strontian calcite with aegirine, biotite, magnetite and pyrochlore and on mineralogical grounds is clearly related to the early sövite intrusion (Cooper, 1988). However, the aegirine-rich composition of the dyke pyroxene, and the incompatible element-rich, whole-rock composition relative to the central sövite (Cooper and Reid, unpublished data) suggest that the dyke represents a batch of chemically evolved magma. A whole rock analysis of the dyke yields a Li concentration of 188 ppm, but this is likely to represent an extreme value, with taeniolite-free carbonatites having lower Li contents. An incipiently fenitized, taeniolite-bearing granitoid from the southeast margin of the complex, AFC195, contains 54 ppm Li.

Discussion

Gold (1966), in the first compilation of carbonatite analyses quotes a 'typical' abundance figure, based on 46 analyses from 11 complexes, of 15 ppm, which can be compared with an average abundance of Li for silicate igneous rocks, cited in the same paper, of 50 ppm. Gerasimovsky (1978) drew a similar conclusion to Gold, that Li is not a typical element of carbonatites, by presenting seven analyses from both the eastern and western rifts of East Africa. Most analyses range from 11 to 29 ppm.

The most recent compilation of carbonatite analyses by Woolley and Kempe (1989) uses rigorous criteria to assess the quality of the analysis, accepting only those analyses which are sufficiently complete to enable the host rock type to be classified. Only one calcio-carbonatite is listed, with a concentration of 0.1 ppm Li, with a single ferro-carbonatite cited of 10 ppm. However, there are published reports of enrichment of Li in carbonatites. Sövites from Tundulu, Malawi (Garson, 1963; reported in Le Bas, 1981) have an average Li content of 55 ppm. McLemore (1987) reports Li concentrations of up to 84 ppm in primary silico-carbonatites from New Mexico, while volcanic calcite carbonatites from the Kontozero area, Kola Peninsula, contain up to 200 ppm (Pyatenko and Osokin, 1988). Kogarko (1993, Table 2) reports average Li contents of 4.4 ppm (range 1.0 to 18.3 ppm) in sövites, 11.75 ppm (1.0 to 72 ppm (but 133 ppm in Table 1)) in dolomitic carbonatite, 14.7 ppm (0.6 to 50.08 ppm) in ferro-carbonatite, and 22.3 ppm (4.1 to 499 ppm) in silicate-carbonate rocks from the Cape Verde Islands. High contents of Li are reported for the natrocarbonatite of Oldoinyo Lengai, with Bowden (1962) quoting concentrations of 100 ppm Li for the 1960 lavas, while 189 ppm Li is cited by Gerasimovsky (1978) for lavas of unspecified age. Associated silicate lavas at Oldoinyo Lengai have Li contents of between 17 and 66 ppm (Donaldson et al., 1987). The Oldoinyo Lengai natrocarbonatite does not contain Li-bearing micas, but carbonatites in which Li-micas have been identified petrographically, namely Mbeya, Sarfartoq and Bayan Obo, have unfortunately not been analysed for Li.

Although the database is limited, and Li concentrations even within a single carbonatite complex are highly variable, enrichment of Li in some carbonatites is real and warrants an explanation.

At Haast River, where it has been shown that carbonatite-related fenites are Li-enriched, radiogenic isotope systematics suggest that the magmas have been derived from a mantle that has previously been metasomatically enriched (Barreiro and Cooper, 1987). Crustal interaction is limited and assimilation does not appear to be petrogenetically important. A similar conclusion applies to Dicker Willem also, based on the low initial Sr isotope ratios (Reid *et al.*, 1990; unpublished data), suggesting that in at least two complexes crustal contamination is inappropriate to explain the Li enrichment. Similarly, the carbonatites at Cape Verde Islands are in an oceanic setting (Kogarko, 1993), where, apart from the possibility of minor continent-derived sediment at the base of the volcanic edifice, the scope for crustal contamination or assimilation is negligible.

As is increasingly recognised by carbonatite researchers (e.g. Jones, 1989), the chemical characteristics of carbonatites are inherited from their mantle source, so the cause of the Li-enrichment in carbonatites might best be sought in processes which fertilise parental mantle peridotites in carbonate and other incompatible elements. Similar conclusions have been reached for Li anomalies in alkaline basic silicate magmas and in MORB, where enrichment or depletion has been ascribed to parental characteristics of the mantle source area (Vasell and Conticelli, 1990; Liotard *et al.*, 1991; Bailey and Gwozdz, 1978).

Unfortunately, the database for Li concentrations in mantle is, like that for carbonatites, similarly meagre. An average Li content for mantle peridotites is cited as 1.9 ± 0.2 ppm (Ryan and Langmuir, 1987), and 2.5 ± 1.4 ppm for 36 peridotite xenoliths (Hartman and Wedepohl, 1990). However, contents of Li measured by ion microprobe analysis on olivine, orthopyroxene, clinopyroxene and garnet of typically depleted mantle nodules (Hervig *et al.*, 1986) would tend to suggest somewhat higher average values for the mantle. On the basis of a comparative geochemical study of different types of peridotite nodules, Hervig *et al.* (1986) conclude that in the mantle, at temperatures higher than the stability field of mica (another likely Li repository), "metasomatic fluids passing through lherzolite would enrich it in Ti, Na, Li, P, Fe^{2+} and Fe^{3+} but remove Cr, V and Sc". Perhaps the most comprehensive account of the geochemistry of metasomatic enrichment of peridotite is that of Hartmann and Wedepohl (1990) who studied xenoliths in basic volcanics of the Hessian Depression, Germany. They used the accumulation of K (and correlated incompatible trace elements) as an index of alteration and showed that Li, amongst many other elements, increased progressively with the intensity of metasomatism.

Conclusions

We have shown that fenitization, varying in character from potassic, as at Dicker Willem, Namibia, to sodic or ultrasodic, as at Haast River, New Zealand, leads to Li enrichment of the country rock adjacent to carbonatite and alkaline silicate intrusions. Zoning profiles in fenite aureoles suggest that Li is preferentially partitioned into fenitizing fluids; only rarely, as in the Dicker Willem sövite occurrence, do the intruding magmas retain their lithian characteristics through the processes of crystallization and separation of a fluid phase. We propose that the lithium-rich nature of the magmas has an ultimate origin in metasomatic alteration of mantle peridotite, in a process remarkably similar to low pressure, crustal fenitization (Haggerty, 1989; Meen *et al.*, 1989). Subsequent melting of such chemically modified mantle yields alkaline or carbonatitic magmas enriched in Li and other incompatible elements.

Acknowledgements

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