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Late Pliocene lamproites from Bucak, Isparta (southwestern Turkey): Implications for mantle 'wedge' evolution during Africa-Anatolian plate convergence

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Abstract

Magmatism in the Kirka–Afyon–Isparta (KAI) region, southwestern Turkey, shows a temporal progression from calc-alkaline to ultrapotassic affinity. Magmatic activity is associated with the geodynamic evolution of the 'Isparta Angle' and is typical of a collision-affected convergent plate margin, most magmas being enriched in potassium and other large-ion lithophile elements (LILE) and depleted in high-field strength elements (HFSE) such as Ti, Zr, Nb, Ta, and Hf. However, Late Pliocene ultrapotassic lamproites in the south of 'Isparta Angle' show HFSE-rich incompatible element distributions, similar to those of 'non-orogenic' intraplate leucite basalts (ILB) and oceanic island basalts (OIB). Their association with HFSE-depleted 'orogenic' magmas suggests that ultrapotassic character reflects primarily crustal contamination of their mantle sources, rather than magma-wallrock reaction effects. Their relatively high content of Fe and Ti (for equivalent Mg content), and SiO_2-undersaturated character suggest that they segregated at relatively high pressures (>ca. 2.0 GPa) from fertile sources. In contrast, the older SiO_2-saturated, Afyon (orogenic) magmas which, for equivalent Mg content, show lower contents of Fe and Ti, are better explained as partial melts segregating at ca. 1.0–1.5 GPa from refractory (basalt-depleted) sources, similar to those of basalt-borne xenoliths tapping the lithospheric mantle. The notion of variably fertile contaminated mantle sources is compelling, but needs to be constrained in terms of the dynamic interaction between the lithosphere and asthenosphere and their respective contamination histories. Given the unlikelihood of *in situ* partial melting of the continental lithosphere mantle, we propose that both orogenic and non-orogenic magmas are generated at different pressures from sources within the convecting asthenosphere, contaminated by both lithospheric mantle and crustal components. This model rests on two testable conjectures: firstly, the interpretation that the continental lithospheric mantle is residual from partial melting at an earlier stage of its history and, secondly, that such material is incorporated into the asthenospheric flow field during and following subduction. The first of these is supported by the ambient compositions of continental basalt-borne xenoliths, while the second is contingent on the prediction that lithospheric mantle may be rheologically transformed during subduction-related hydration prior to its incorporation. The proximity of the Bucak lamproites to the Menderes Massif, a suspected Archean cratonic fragment, highlights the resemblance of these unusual rocks to intra-plate leucite-bearing lamproites elsewhere, whose genesis has been linked to mantle 'wedge convection' triggered beneath cratonic and circumcratonic lithosphere domain boundaries.

Keywords: SW Turkey; Lamproite; Shoshonite; Asthenosphere; Lithospheric mantle; Collision; Delamination; Subduction

1. Introduction

Neogene-Quaternary magmatism in Anatolia is closely associated with the assembly of microterranes during late stages of Tethyan closure and has been variously linked to the processes of subduction, microcontinental collision, and asthenospheric upwelling (e.g. Pearce et al., 1990; Yilmaz and Polat, 1998; Altunkaynak and Yilmaz, 1998; Aldanmaz et al., 2000, 2005; Yilmaz et al., 2001; Doglioni et al., 2003). During the tectonic evolution of western Anatolia, Early Tertiary calc-alkaline volcanism gave way to Lower-Mid

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Miocene potassic and ultrapotassic shoshonite activity and, more recently, to Late Neogene to Quaternary lamproites and alkali basalts (Fytikas et al., 1984; Savasçin and Gulec, 1990; Pe-Piper and Piper, 1992; Seyitoglu et al., 1997; Altunkaynak and Yilmaz, 1998; Genc, 1998; Francalanci et al., 2000; Innocenti et al., 2005; Tonarini et al., 2005). Similar magmatic associations have been recorded from central Anatolia (e.g. Besang et al., 1977), and northern, eastern, and southeastern Anatolia (e.g. Wilson et al., 1997; Aldanmaz et al., 2000; Yilmaz et al., 2001), and have in most cases been associated with short-lived, collision-related subduction-accretion episodes. While most of the associated magmatic products show depletion in high-field strength elements (HFSE) such as Ti, Nb, Ta, Nb, and Hf, usually ascribed to subduction-related or 'orogenic-type' processes, the relatively enriched rare earth element (REE) and HFSE distributions observed in 'non-orogenic' lamproites and associated sodic alkali basalts are closer in type to those of intraplate oceanic island basalts (OIB).

In the region of western Anatolia and the contiguous Aegean Sea, ultrapotassic magmatism is especially prominent in the so-called Isparta Angle, a triangular area between Kirka, Afyon, and Isparta (KAI) – which consists of a north-south-shifting structural complex linking Central Anatolia to the Isparta Angle (Besang et al., 1977; Yagmurlu et al., 1997; Francalanci et al., 2000) (Figs. 1 and 2). Magmatic activity in this region reflects two extensional stages that have been broadly recognized over much of western Anatolia (Kocyigit et al., 2000; Cihan et al., 2003; Coban et al., 2005). Calc-alkaline ignimbrite activity in the Kirka region (~21–18 Ma) (Besang et al., 1977), presumably marking the demise of north-vergent African plate subduction, gave way southward, initially, in the Mid- to Late Miocene (ca. 14.8–8.0 Ma) ultrapotassic to potassic lamproites and shoshonites, and their derivatives near Afyon (Besang et al., 1977; Keller, 1983; Savasçin and Oyman, 1998; Francalanci et al., 2000), and, secondly, south of Sandikli (Afyon) and in the Isparta region, north of Antalya, to Late Miocene to Pliocene shoshonites and lamproites (Francalanci et al., 1990, 2000; Aydar, 1998; Aydar et al., 1995; Alici et al., 2002; Coban et al., 2005).

Here, we report new data for Late Pliocene lamproites from Bucak, near Isparta (Figs. 1 and 2). These data, with other published data, are used to trace mantle compositional evolution beneath the evolving Africa-Anatolia plate boundary. The lamproites and their accompanying, potassic eruptions, are relatively enriched in HFSE compared to the previously-erupted orogenic magmas, suggesting they were unrelated to active subduction. However, the close spatial-temporal association of both 'orogenic' and 'non-orogenic' magma types suggests a need for petrogenetic models to reconcile their strongly contrasting magmatic sources with a common geodynamic setting.

Our understanding of geochemical mass balances affecting convecting mantle 'wedge' regions remains speculative, although clearly, post-collision potassic and ultrapotassic magmas, recorded from the Alpine, Appenine, Betic and Himalayan orogens – are closely linked in time and space to slab tearing and detachment (Davies and Von Blanckenburg, 1995; Wortel and Spakman, 1992, 2000). Targeted studies of seismic tomography,

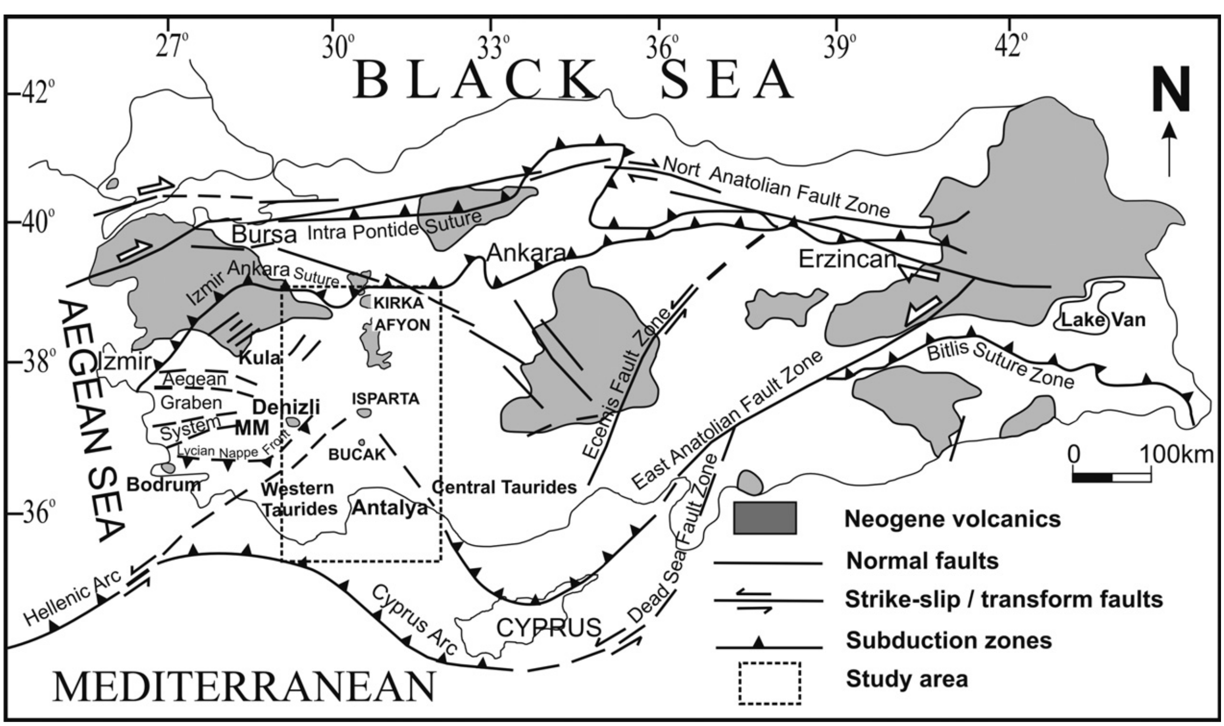

Fig. 1. Tectonic map, including Neogene volcanics (gray filled) of the eastern Mediterranean and Turkey (modified from Ercan et al., 1985; Pearce et al., 1990; Yagmurlu et al., 1997; Yilmaz and Polat, 1998; Aldanmaz et al., 2000; Kocyigit et al., 2000; Yilmaz et al., 2001). MM, Menderes Massif.

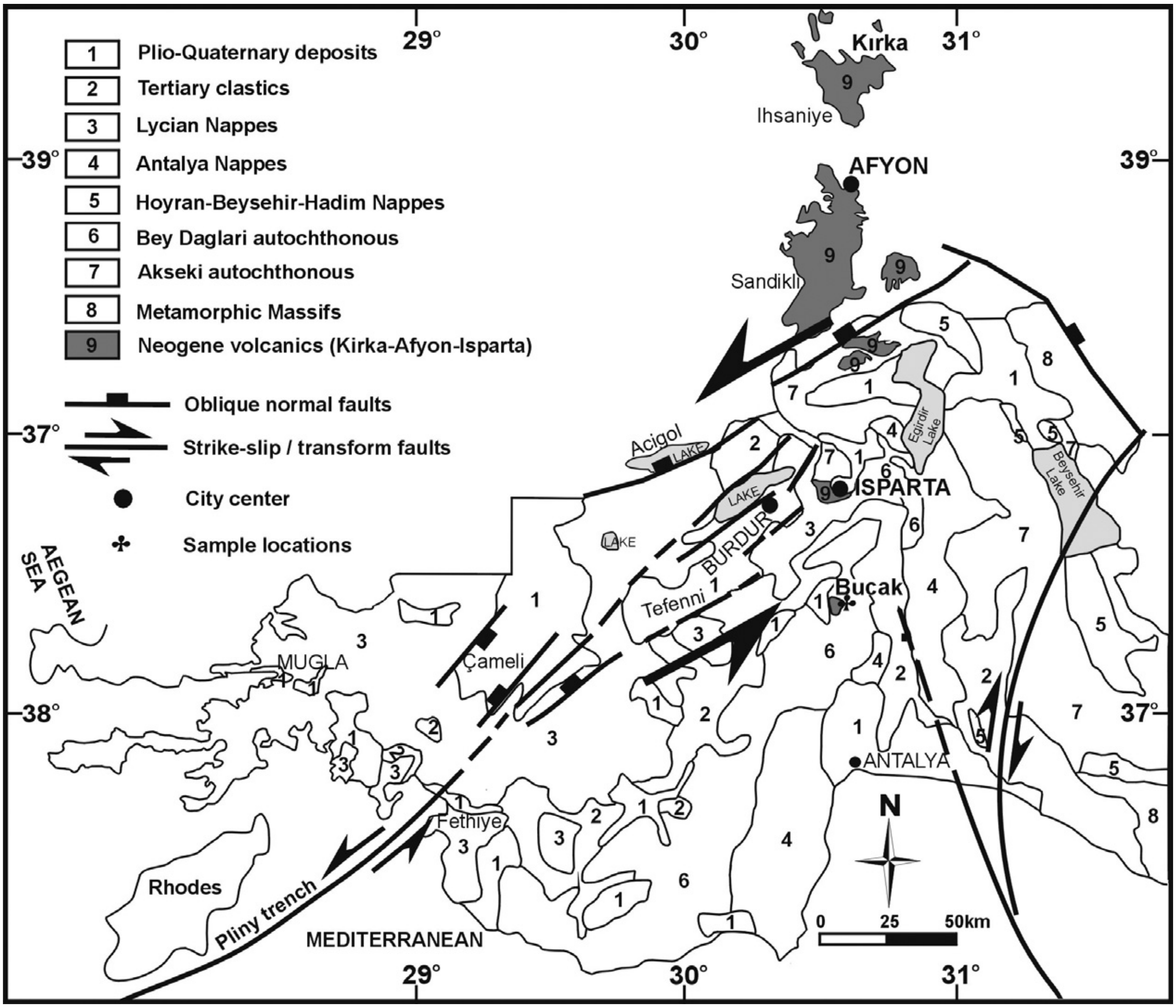

Fig. 2. Map of the Kırka-Afyon-Isparta region (southwest Turkey) showing the Isparta Angle and major structural elements (modified from Cemen et al., 1999; Alcicek et al., 2006, reference therein). Neogene volcanics modified from Yagmurlu et al. (1997) and Francalanci et al. (2000).

S-wave splitting and seismic attenuation offer the potential of 3D 'snap shots' of evolving plate configurations and thermal heterogeneities as boundary conditions for geochemical mass balance models of the type presented here. Based on the existing geochemical data base, we argue that: (1) strong LILE enrichments in both 'orogenic' shoshonites and 'non-orogenic' lamproites in the KAI province reflect primarily crustal contamination of their mantle sources. rather than the effects of magma-wallrock reactions, (2) models appealing to *in situ* partial melting of the lithospheric mantle, commonly proposed for post-collisional potassium-rich magmas, are fundamentally flawed, given the lack of a mechanism for melting cold, refractory peridotite, and the implicit requirement that contamination of such sources occurred before the collision. As an alternative, we propose that these magma types tap discreet regions of a common convecting mantle 'wedge', variably contaminated via delamination, of refractory (Fe-, HFSE-depleted), hydrated peridotite and lower crust, and subduction, of upper crustal regoliths and crust-derived sediment.

2. Magmatism and geodynamics in the Isparta angle

2.1. Geodynamic background

The accretion of continental Anatolia appears to have involved a succession of diachronous collisions between Gondwana microplates and accreting southern Eurasia. These were often accompanied by short-lived episodes of subduction-accretion, marked by rapid back-arc opening and forearc accretion, as represented by ophiolites (Sengör and Yilmaz, 1981; Sengör et al., 1984, 1993, 1987; Sengör and Natal'in, 1996) (Fig. 3). In the Late Miocene, this 'accordion-like' pattern was interrupted by sudden westward motion of the accreting Anatolian Block with concomitant opening of the Aegean Sea, interpreted by most workers as far-field effects of the Arabian-Eurasia collision (e.g. McKenzie, 1970, 1976; Yaltirak et al., 1998; Martinod et al., 2000; Bozkurt, 2001; Doglioni et al., 2003). The tectonic evolution of the Isparta Angle appears to have embodied both effects. This complex feature, stretching north from the Bay of Antalya (Fig. 4) marks

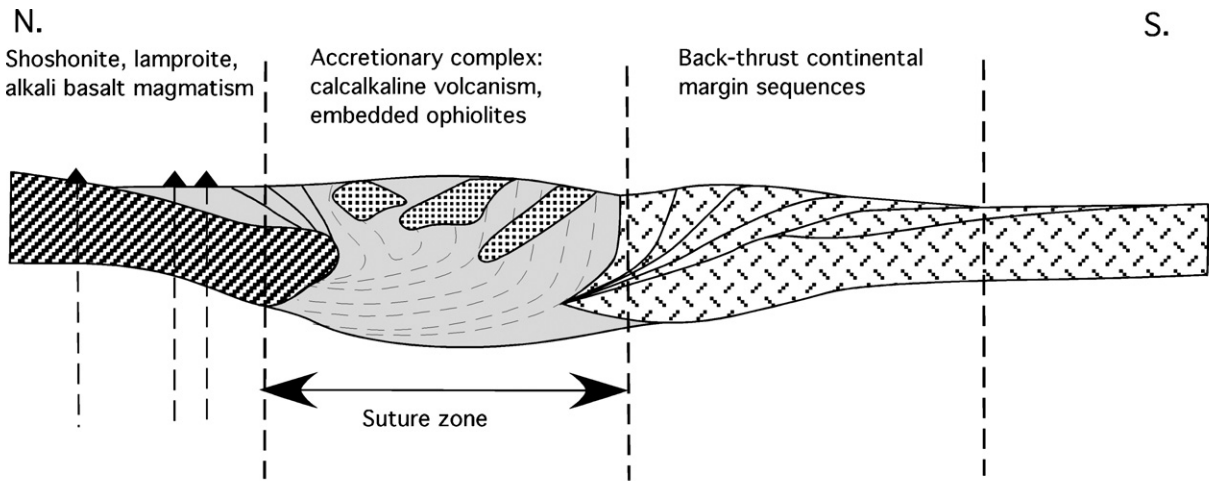

Fig. 3. 'Turkic-type' collisional orogenic belts characterized by large subduction-accretion complexes precluding apposition of pre-subduction continental pieces; after Sengör and Natal'in (1996).

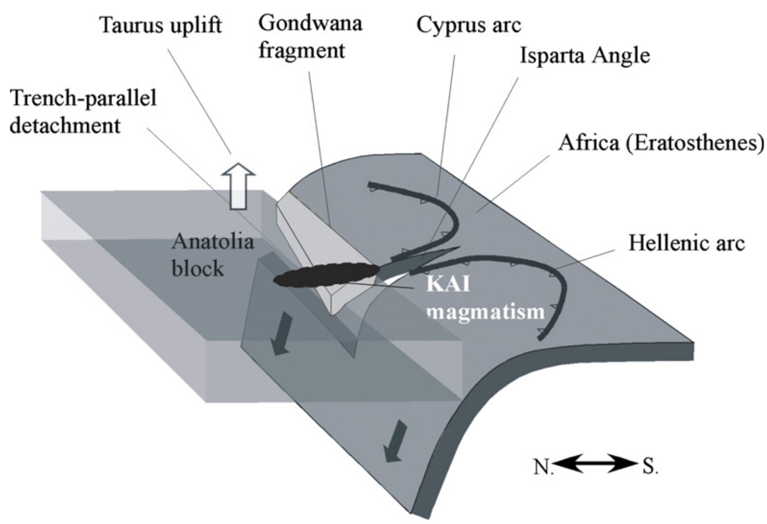

Fig. 4. Cartoon illustrating a hypothetical slab-tear developing beneath the Taurus Mountains in response to the collision of one or more Gondwana indentors (indicated but not shown) with Anatolia. Hypothetical geometry of subducting African lithosphere beneath the Cyprus and Hellenic arcs is also shown (modified after Barka and Reilinger, 1997; Sandvol et al., 2001, 2003; personal communication). KAI -Kırka-Afyon-Isparta volcanic field.

the conjunction of Hellenic and Cyprus arc systems and, moreover, may record the northward indentation of Anatolia by a detached Gondwana fragment (Sengör and Yilmaz, 1981; Sengör et al., 1984; Glover and Robertson, 1998b). Rollback of the Cyprus arc-trench system, initiated in the Late Cretaceous, appears to have terminated after the initiation of Anatolian escape, while rollback of the Hellenic arc to the west was able to continue unobstructed (e.g. Barka and Reilinger, 1997; Glover and Robertson, 1998b) (Fig. 4). Regional earthquake data and seismic tomographic models (e.g. Wortel and Spakman, 2000) clearly define both relict and actively subducting Cyprus and Hellenic slabs beneath the Isparta Angle, to the northwest and northeast respectively (Fig. 4), implying the presence of north-south-oriented slab-tearing with concomitant perturbations of the asthenospheric flow field and potential temperature.

Geodynamic evolution of this region is pivotal to understanding magmatic patterns and source evolution (see below). However, there are two competing model interpretations. The first contends that the Isparta Angle, a weak transtensional feature dominated by right-lateral strike-slip faulting, separates a western Anatolian-Aegean extensional domain from the eastern Anatolian Block (e.g. Glover and Robertson, 1998a,b). According to this view, the Isparta Angle expresses north-south slab-tearing associated with post-collision subduction (Figs. 2 and 4). An alternative model views the faulting as predominantly left-lateral, allowing for differential motion between the Hellenic and Cyprus arcs (e.g. Barka and Reilinger, 1997), and that crustal shortening, accommodated by the present-day Sultandag and Aksu thrust faults, occurred as a result of Anatolia overriding the Cyprus arc (Fig. 4). The left-lateral Burdur Fault is believed, moreover, to express tearing of a downgoing slab between the Hellenic and Cyprus subduction systems while faulting in the Antalya basin reflects transtension in the accommodation zone (Barka and Reilinger, 1997).

2.2. Igneous activity

The spatial-temporal pattern of KAI activity, Mid- to Late Miocene ultrapotassic and potassic lamproites and shoshonites of the northern Afyon Massif (ca. 14.8–8.0 Ma) (Besang et al., 1977), Pliocene shoshonites south of Sandikli-Afyon (ca. 4.5 Ma; Coban et al., 2005), and near-saturated shoshonites and latites (ca. 6–0.35 Ma, Lefevre et al., 1983; Özgür et al., 1990; Bilgin et al., 1990; Francalanci et al., 2000; Coban, 2005) and lamproites (ca. 4.2–4.7 Ma, Lefevre et al., 1983; Yagmurlu et al., 1997; Savaşçin and Oyman, 1998) near Isparta region, that followed the Kirka calc-alkaline activity (21–18 Ma), fundamentally constrains geodynamic models for the region (cf. Yilmaz and Polat, 1998; Aldanmaz et al., 2000; Pearce et al., 2000; Yilmaz et al., 2001). The youngest activity, occurring in Isparta (ca. 6.0–0.35 Ma; Lefevre et al., 1983; Kazanci, 1995; Coban, 2005), includes saturated to undersaturated potassic and ultrapotassic products (Aydar et al., 1996; Francalanci et al., 2000; Alici et al., 2002), in contrast to orogenic-type HFSE-depleted types that characterize most of the KAI activity. In contrast, Pliocene lamproites that appeared in Bucak, close to Isparta, are characterized by OIB-like HFSE-rich patterns (Francalanci et al., 2000; Alici et al., 2002) and relatively low Sr isotopic ratios (ca. 0.7038) (Francalanci et al., 2000).

The overall, two-stage southward migration of KAI activity is generally matched by increased potassic and ultrapotassic character and decreased SiO_2-saturation. The progression from SiO_2-saturated to SiO_2-undersaturated potassic, and ultimately sodic, magmas in the region, and the increased dominance of primitive melt products, is more clearly developed than elsewhere in the Anatolia-Aegean province. This progression is consistent with the evolution of magmatic activity associated with an active subduction system as progressive slab-tearing occurs in response to the eventual collision of continental plates (e.g. Serri et al., 1993). Such a slab-tearing model (discussed below) is also supported by tomographic evidence for anomalous low upper mantle Pn velocities and a sharp decrease in seismic attenuation, both indicative of decreased crustal thickness in the Isparta Angle (Sandvol et al., 2001, 2003; personal communication; Al-Lazki et al., 2004) (Fig. 4).

3. The Bucak lamproites

The Bucak lamproites were selected for study in view of their primitive, unfractionated character, as representatives of near-primary upper mantle melt.

3.1. Sampling and petrography

Phenocrysts, including Mg-rich olivine, phlogopite and diopsidic augite scattered in a matrix of K-richterite, Cr-spinel, leucite, sanidine, apatite and magnetite–ilmenite, show typical micro-texture of primitive ILB (Intra-plate Leucite Basalt). However, they lack phenocrysts of priderite, wadeite and plagioclase that appear to characterize evolved lamproites elsewhere, as in Afyon-Sandikli to the north (Akal and Helvaci, 2002; Coban and Flower, 2006). In general, the phenocryst components show high Mg# [$100 * Mg/(Mg + Fe^{2+})$; 74–81], although reaction morphologies and zoning indicate small, but significant, polybaric fractional crystallization effects. On the other hand, phenocryst-matrix relationships indicate that crustal wall-rock reaction was probably not a significant factor causing potassium-enrichment (Coban and Flower, 2006). The Bucak lamproites are distinguished from ultrapotassic shoshonites in Afyon by the presence of groundmass leucite ± sodalite (although leucite megacryts, nepheline and melilite have been found in the latter; Akal, 2003; Coban and Flower, 2006), reflecting strong SiO_2-undersaturation of their parent melts. The Afyon-Sandikli 'lamproites', in contrast, are dominated by assemblages of K-rich terite, olivine, phlogopite, clinopyroxene and sanidine.

4. Analytical methods and results

Major, trace element and REE analyses of selected rock samples were carried out using the X-ray fluorescent spectrometer, ICP and ICP-MS (inductively-coupled plasma mass spectrometry) methods at: (1) the University of Lausanne (Switzerland); (2) the Mineral Research and Exploration Laboratory (MTA) of Turkey; and (3) the Natural Resources Laboratory of Jordan. Average detection limit is 0.5 ppm for trace elements, and range in 0.007–0.012 ppm for REE. Accuracy and precision was evaluated analysing international reference samples as unknown samples (i.e. AGV-1, BHV-0, MFTH, QLO, GSP and STM). Errors for major and trace elements are better than 5%. Of the analyzed samples, major, trace and REE analyses of 15 representatives (showing the lowest 'loss-on-ignition' values) are given in Tables 1 and 2. In the figures, major elements are normalized to 100% volatile-free, and for comparison, all other published database have also been rearranged as dry-weight.

4.1. Geochemical variation

Published petrographic and whole-rock geochemical data for Afyon, Kirka, and Isparta (KAI) magmatic rocks (Besang et al., 1977; Keller, 1983; Cevikbas et al., 1988; Aydar et al., 1995, 1996, 2003; Savaşçin and Oyman, 1998; Alici et al., 1998; Francalanci et al., 1990, 2000; Akal and Helvaci, 2002; Akal, 2003) together with new data for the Bucak lamproites provide a basis for interpreting mantle source affinity and thermal state with a view to refining the regional geodynamic constraints.

The analyzed lamproites are near-peralkaline (P.I. > 0.6), ultrapotassic ($K_2O/Na_2O > 2.5$), and SiO_2-undersaturated, showing relatively low SiO_2 (45.72–49.19 wt%) and Al_2O_3

Table 1
Representative major (wt%) and trace (ppm) element compositions of Bucak lamproites

Sample No.	B1T	B5T	B6T	B7T	B3-TB	B6T-b	BT1-b	B6T-ba	A1	A2	B1a	B2a	B3	B4	B5
Rock type	lmp	lmp	lmp	lmp	lmp	lmp	lmp	lmp	lmp	lmp	lmp	lmp	lmp	lmp	lmp
SiO_2	47.61	49.19	47.62	47.98	48.66	48.26	47.65	48.24	47.5	46.8	47.56	45.72	47.6	46.7	47.1
TiO_2	1.59	1.77	1.74	1.64	1.76	1.66	1.71	1.66	2.01	1.7	1.5	1.7	1.69	1.78	1.95
Al_2O_3	9.37	10.82	9.48	11.00	10.43	9.53	9.09	9.73	9.5	10.4	10.43	9.9	9.91	9.3	8.89
Fe_2O_{3tot}	6.56	6.81	6.63	7.29	6.91	7.30	7.15	7.63	7.3	7.15	7.8	6.9	6.68	7.4	7.9
MnO	0.10	0.11	0.10	0.12	0.12	0.13	0.11	0.13	0.16	0.1	0.11	0.12	0.11	0.12	0.13
MgO	11.73	10.39	11.67	11.64	10.60	10.38	11.03	10.59	10.8	11.9	12.23	12.1	12.2	11.2	12.9
CaO	10.10	8.39	9.73	9.11	8.21	9.24	8.51	9.25	9.1	8.46	8.55	10.7	9.75	10.5	9.5
Na_2O	1.35	1.95	1.49	1.64	1.85	1.75	2.13	1.45	2.12	2.11	2.31	2.25	2.15	1.85	1.76
K_2O	7.36	6.32	7.35	4.81	5.04	7.51	6.56	8.51	6.9	6.5	5.72	6.8	6.3	6.78	6.3
P_2O_5	1.70	1.56	1.70	1.62	1.28	1.16	1.42	1.17	1.7	1.61	1.82	1.25	1.38	1.4	1.59
L.O.I.	1.55	1.68	1.44	2.11	3.85	3.65	2.58	2.45	2.6	2.51	1.8	1.72	1.51	2.05	1.9
Total	99.02	98.99	98.95	98.96	98.71	100.57	97.94	100.81	99.69	99.24	99.83	99.16	99.28	99.08	99.92
Mg#	80	76	78	79	78	74	78	79	77	79	82	80	80	77	78
K_2O/Na_2O	5.46	3.25	4.94	2.92	2.72	4.29	3.08	5.87	3.25	3.08	2.48	3.02	2.93	3.66	3.58
P.I.	0.93	0.76	0.93	0.59	0.66	0.97	0.96	1.02	0.95	0.83	0.77	0.91	0.85	0.93	0.91
Ba	3105	2284	3093	2528	2702	2113	2222	2435	1864	1343	2147	1589	1834	1453	1910
Sr	1651	1997	1664	2290	1485	2213	1582	1632	1839	2073	2282	1896	1347	1846	1371
Nb	178	173	184	146	203	176	203	190	164	147	104	162	172	188	216
Zr	929	990	955	882	878	674	1026	965	940	480	914	591	570	319	291
Y	20	24	19	26	25	27	18	16	28	21	23	22	19	29	23
U	4.7	5.7	5	6	<2	6.6	<2	<2	9.3	7.8	6.2	10	5.7	5.9	6.5
Rb	302	338	293	338	525	195	425	726	225	235	226	221	189	205	267
Th	17	16	16	16	10	15	10	15	18.2	14.6	19.5	19.8	22.4	17.4	20.8
Pb	10	14	12	15	11	46	14	11	42	15	11	19	26	43	52
Ni	408	469	414	510	569	421	471	597	534	476	372	480	457	474	463
Cr	623	566	585	647	618	623	583	675	787	473	602	615	657	696	792
Hf	20	19	19	18	17	14.8	16	17	16	17	18	20	17	15	15
Ta	10	10	10	6	5	4.9	5	7	4	8	10	6	5	5	4

Fe_2O_3 total as total Fe, Mg# = Mg/Mg + Fe^{2+}) ; (assuming wt% Fe_2O_3 = 1.1113 * (FeOt–FeO), Sack et al., 1980). lmp, lamproite; P.I., peralkalinite index molar ($K_2O + Na_2O$)/Al_2O_3.

Table 2
Representative REE (ppm) compositions of Bucak lamproites and Afyon-Sandıklı lamproites (L1*, L2*)

Sample No.	B1T	B5T	B6T	B7T	B3-TB	L1*	L2*
Rock type	lmp	lmp	lmp	lmp	lmp	lmp	lmp
La	178	146	176	147	188	45	46
Ce	339	272	336	277	321	90	93
Pr	24.12	29.54	20.87	21.55	21.91	10.76	11.13
Nd	89.88	111.13	82.5	82.41	81.91	46.3	48.1
Sm	13.18	16.42	12.23	14.31	12.71	9.01	9.3
Eu	3.31	3.94	3.72	3.58	3.66	2.06	2.17
Gd	9.81	11.24	9.95	10.32	10.76	7.18	7.37
Dy	5.55	5.49	4.24	4.22	4.26	5.26	5.18
Er	2.61	2.15	1.75	1.8	1.87	2.34	2.46
Yb	1.92	1.74	1.84	1.56	1.64	1.97	2.13
Lu	0.25	0.22	0.26	0.25	0.24	0.3	0.3
CeN/YbN	46	41	47	46	51	12	11
LaN/YbN	63	57	65	64	77	15	15

(8.89–11.00 wt%), and high contents of MgO (10.6–12.9 wt%), K_2O (5.04–8.51 wt%), CaO (8.39–10.7 wt%), TiO_2 (1.59–2.01 wt%), Na_2O (1.35–2.25 wt%), P_2O_5 (1.16–1.82 wt%), Ni (372–597 ppm) and Cr (473–792 ppm). These are matched by high contents of Ba (1343–3105 ppm), Sr (1347–2282 ppm), Zr (291–1026 ppm) and REE, with Mg# (100 * Mg/[Mg + Fe^{2+}]) ranging from 74 to 82. The Bucak lamproites are the most primitive reported in the KAI province. So-called lamproites from Afyon-Sandiki are highly evolved (Akal and Helvaci, 2002) and, like other ultrapotassics from Afyon, show HFSE depletions and negative Eu anomaly contrasting with the Bucak lavas (see also Fig. 8b).

Most classifications of ultrapotassic magma emphasize tectonic affinities on the basis of major element compositions (e.g. Peccerillo and Taylor, 1976; Foley et al., 1987; Conticelli and Peccerillo, 1992). Plots of K_2O versus SiO_2 and K_2O versus Na_2O (Fig. 5a and b) confirm the ultrapotassic affinity of the Bucak lamproites, while plots of CaO versus Al_2O_3 and CaO versus MgO (Fig. 6a and b) show they conform to the type II ('lamproite') group of Foley et al. (1987), in contrast to most Afyon and Kirka compositions, which mostly show type III ('orogenic') character (Fig. 6a and b). In Figs. 5 and 6, major element compositions of orogenic and non-orogenic magmas from the KAI province are also compared with intra-plate leucite basalt-ILB suites from the western Germany, the East African Rift, northeastern China, southern Brazil, and Central Spain. However, non-orogenic ILB compositions overlap with the latter in terms of major element criteria

(Figs. 5 and 6), although, fortunately, may be readily distinguished by their enriched HFSE patterns.

In common with most incompatible elements, the relative abundances of HFSE may be assumed to record mantle source character rather than magmatic fractionation effects (discussed below). Accordingly, plots of TiO_2 and K_2O/Na_2O versus Mg# (Fig. 7a and b) allow for the distinction of mantle 'fertility' and crustal contamination as factors determining primitive magma compositions. The concept of mantle fertility reflects the extent to which a fusible basalt component is present in peridotite as represented by compositions ranging between high-Ti, low Mg# fertile lherzolite and low-Ti, high Mg# refractory harzburgite. A fertile source has the potential to yield MORB and OIB-like suites while a refractory source-residual to previous basalt extraction, can yield (at significantly higher temperatures) calc-alkaline ankaramites and boninites (Nguyen and Flower, 2004).

In the context of the complete KAI database, the compositions of the Bucak lamproites and Afyon (Group B) shoshonites may be interpreted as products of a relatively fertile, Ti-rich source (Fig. 7a). In contrast, low-Ti series such as the Isparta shoshonites are inferred to have tapped relatively refractory sources, while the Kirka and Afyon series reflect sources of intermediate fertility (Fig. 7a). The variation of K_2O/Na_2O with Mg# (Fig. 7b) suggests that, except for negative covariances within discrete eruptive series, the overall variation of K_2O/Na_2O is decoupled from Ti and (presumably) other HFSE. Contamination of KAI magmatic sources thus occurred independently of the process or processes governing HFSE. If the latter reflect heterogeneities produced by prior basalt extraction, the K_2O/Na_2O variation almost certainly reflects subsequent contamination effects.

In general, these conclusions are consistent with the variation of incompatible trace elements. The Bucak samples show chondrite-normalized light rare earth element

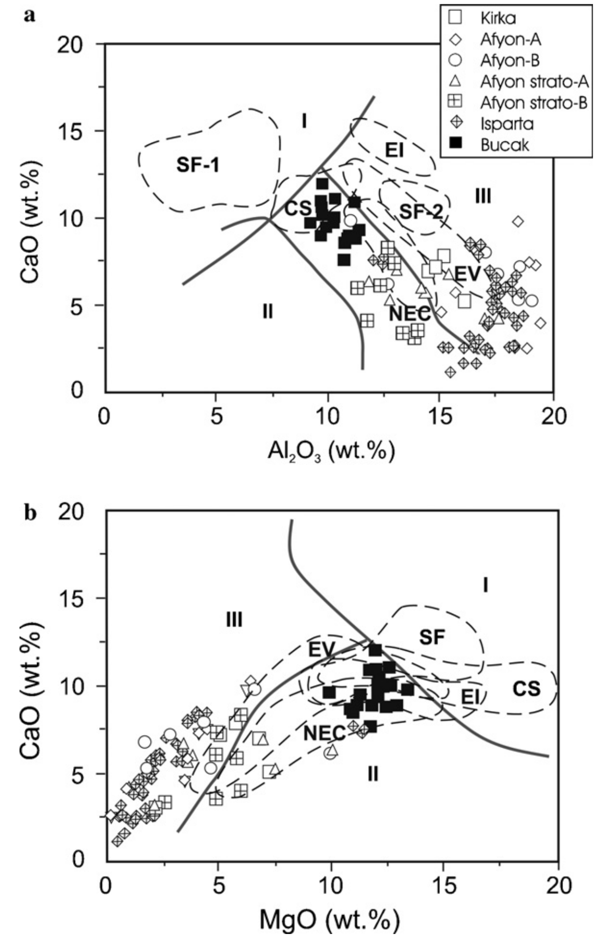

Fig. 6. Major element oxide plots: (a) CaO versus Al_2O_3 (wt%), and (b) CaO versus MgO (wt%)-classification of ultrapotassic rocks (I, Kamafugites; II, Lamproites; III, Roman Province Lavas; after Foley et al., 1987), for Kırka–Afyon–Isparta magmas in comparison with fields for representative ILB suites. Data sources as in Fig. 5.

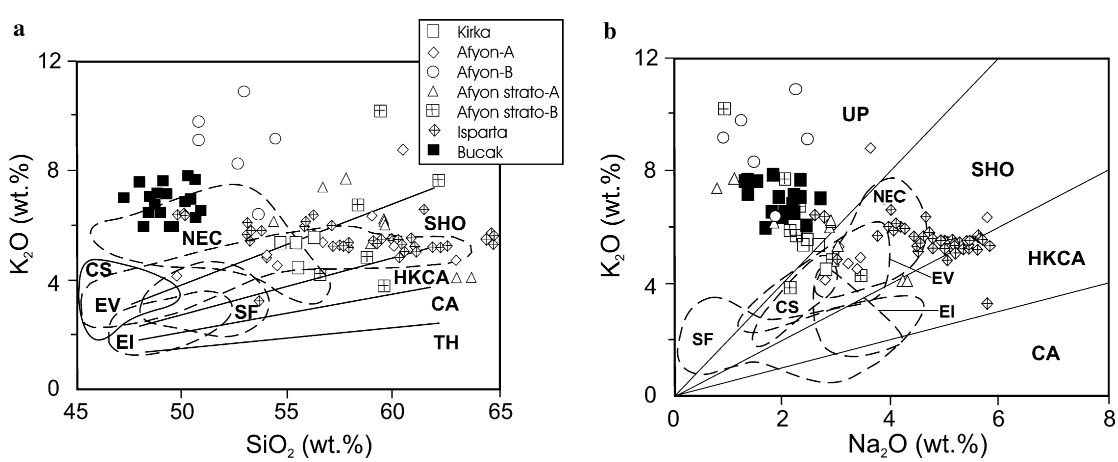

Fig. 5. Major element oxide plots of (a) K_2O versus SiO_2 (wt%): (b) K_2O versus Na_2O (wt%), for Kırka–Afyon–Isparta magmas (Alici et al., 1998; Francalanci et al., 2000; Aydar et al., 2003; and this work) in comparison with fields for representative intra-plate leucite basalt (ILB) suites; Eifel (EI), western Germany (Worner et al., 1986); Ciudad Real, central Spain (CS) (Ancochea and Ibarrola, 1982; Cebria and Lopez-Ruiz, 1995); Heilongjiang Province, northeastern China (NEC) (Zhang et al., 1995); San Francisco (SF), southern Brazil (Carlson et al., 1996; Thompson et al., 1998); and East Virunga (EV), Western Rift Valley, East Africa (Furman, 1988; Rogers et al., 1998).

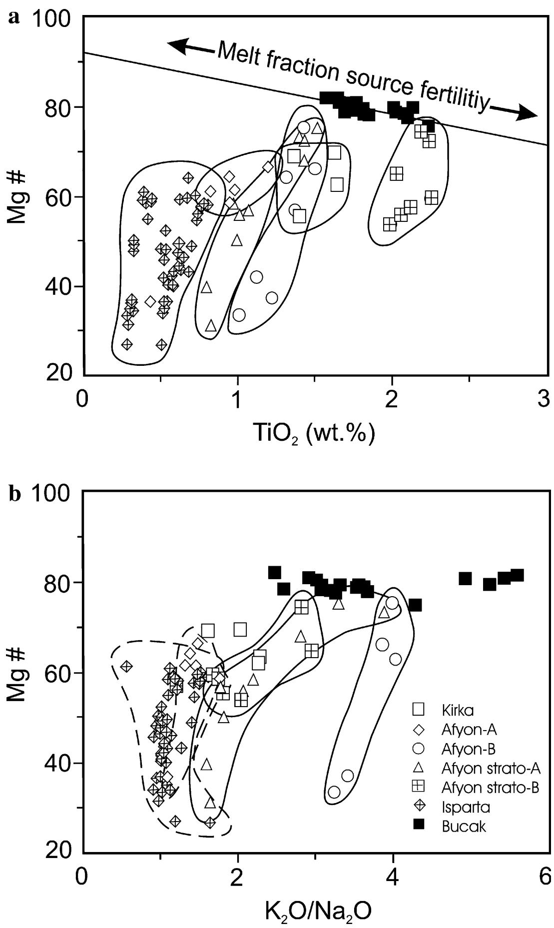

Fig. 7. Plots of: (a) Mg# 100× Mg/(Mg + Fe^{2+}) versus TiO_2 (wt%), (b) Mg# 100 × Mg/(Mg + Fe^{2+}) versus K_2O/Na_2O, for Kırka–Afyon–Isparta magmas, in comparison with fields for representative ILB suites. Data sources as in Fig. 5.

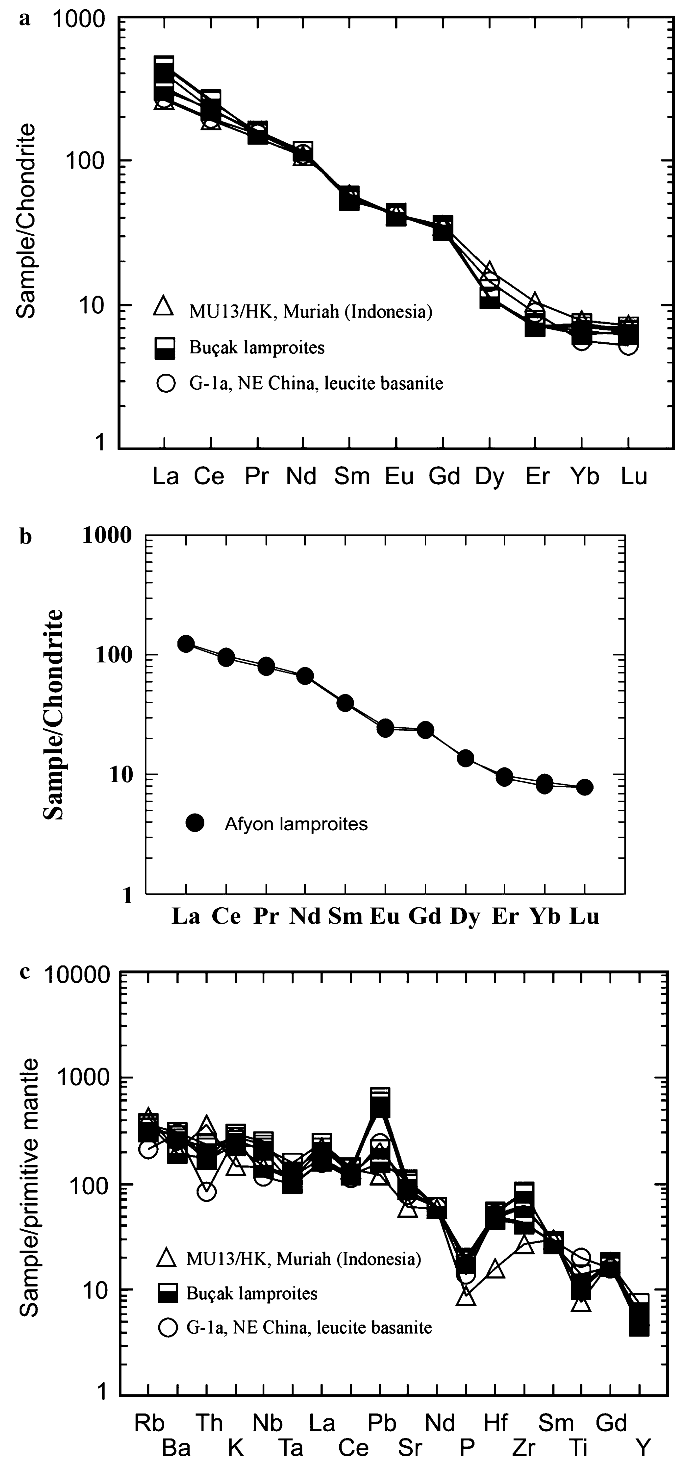

Fig. 8. (a and c) Rare earth and incompatible element distributions for Bucak lamproites and representative ILB samples from Muriah (Edwards et al., 1991) and northeast China (Zhang et al., 1995). (b) REE pattern of Afyon lamproites plotted for comparison. REE are normalized to chondrites, and (b) incompatible elements are normalized to primitive mantle. Normalizing values after Sun and McDonough (1989).

(LREE) enrichments ($Ce_N/Yb_N = 41–51$; $La_N/Yb_N = 64–77$) and small to negligible positive Eu anomalies (Eu/Eu* = 0.9–1.03) (Fig. 8a). These resemble those of ILB suites, rather than the upwardly-convex patterns and negative Eu anomalies (Fig. 8b) that characterize most orogenic ultrapotassic suites (e.g. Foley et al., 1987; Turner et al., 1999; Miller et al., 1999; Chung et al., 2001). Despite somewhat higher contents of LILE (e.g. Rb, Ba and Th) and Pb, and lower LREE contents, mantle-normalized HFSE-rich incompatible element distributions also conform to non-orogenic rather than orogenic patterns (Fig. 8c) (Hofmann et al., 1986).

Available Sr and Nd isotopic data for the Isparta potassic and ultrapotassic rocks (Alici et al., 1998; Francalanci et al., 2000; Coban, 2005) plot close to lower-left quadrant of the Sr–Nd diagram (Fig. 9), suggesting predominantly EM1 (Hart, 1984; Zindler and Hart, 1986) as a potential, if unexpected, source component. EM1-like isotopic compositions commonly characterize ILB eruptives in Eifel, western Germany (Worner et al., 1986), Muriah (Indonesia) (Edwards et al., 1991), NW Taiwan (Chung et al., 2001), San Francisco (SF), southern Brazil (Carlson et al., 1996; Thompson et al., 1998), northeastern China (Zhang et al., 1995) and east Virunga, the western limb of the East African Rift Valley (Furman, 1988; Bell and Peterson, 1991; Simonetti and Bell, 1994; Rogers et al., 1998), as well as cratonic mantle xenoliths (Menzies and Murthy, 1980a,b), and

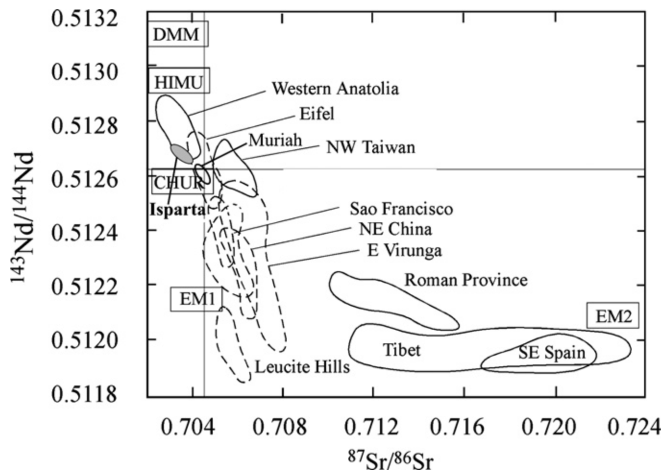

Fig. 9. Plots of $^{87}Sr/^{86}Sr$ versus $^{143}Nd/^{144}Nd$ isotopic ratios for lamproites from the Kırka–Afyon–Isparta magmatic province, in comparison with fields for representative ILB suites. Data sources as in Fig. 5, and additionally include, Western Anatolia (Ercan et al., 1985; Savaşçin and Gulec, 1990; Gülec, 1991; Aldanmaz et al., 2000; Francalanci et al., 2000; Yilmaz et al., 2001; Alici et al., 2002), Eifel (EI), western Germany (Worner et al., 1986); Muriah (Indonesia) (Edwards et al., 1991), NW Taiwan (Tu et al., 1992; Chung et al., 1994), Heilongjiang Province, northeastern China (NEC) (Zhang et al., 1995); San Francisco (SF), southern Brazil (Carlson et al., 1996; Thompson et al., 1998). For comparison Roman Province (Conticelli, 1998; Conticelli et al., 1992, 2002), Tibet (Miller et al., 1999), SE Spain (Turner et al., 1999; Toscani and Mariani, 2000), and Leucite Hills (Nelson et al., 1986; references therein) are also reported.

the exhumed peridotite from Ronda and Beni-Bousera (Reisberg and Zindler, 1985).

5. Discussion

The progression from calc-alkaline to high-K calc-alkaline, shoshonitic, lamproitic, and rarely alkali basaltic activity is commonly taken to record the transition from plate convergence to collision (e.g. Pe-Piper and Piper, 1992; Serri et al., 1993; Lavecchia and Stoppa, 1996; Turner et al., 1996, 1999; Conticelli, 1998; Peccerillo, 1999; De Astis et al., 2000). However, lamproitic ILB-type activity occurs mostly in non-orogenic settings and is lacking in most collision-related provinces. If we assume that ultrapotassic magmatism essentially reflects a contaminated mantle source character, rather than AFC-like wallrock interactions (Conticelli, 1998; Peccerillo, 1999), the significance of magmatic HFSE distributions can be reviewed on the basis of relevant experimental data before developing a comprehensive geodynamic model to explain shoshonitic and lamproitic magmas.

5.1. P–T conditions, source fertility

Except for their LILE contents, primitive shoshonitic magmas resemble boninites in several important respects, notably their SiO_2-saturated to oversaturated character (e.g. Tatsumi, 1982; Tatsumi and Ishizaka, 1982; Foley et al., 1987; Foley and Venturelli, 1989), the positive covariance of SiO_2 and Mg#, and their low contents of TiO_2 and FeO* (for equivalent MgO), cf. MORB and OIB (e.g. Van der Laan et al., 1989; Crawford et al., 1989; Bloomer et al., 1995) (see Fig. 7a and b). Given their geochemical similarities and the lack of experimental data for near-primary shoshonites (strictly speaking, 'absarokites'), experimental studies of boninites offer a basis for estimating pressure and temperature conditions at which shoshonitic magmas segregate from their mantle sources. For example, results from near-liquidus crystallisation studies on primitive boninites (e.g. Van der Laan et al., 1989; Umino and Kushiro, 1989; Falloon and Danyushevsky, 2000), and melting studies of refractory peridotite (e.g. Jaques and Green, 1980; Hirose and Kushiro, 1993; Hirose and Kawamoto, 1995; Hirose, 1997), suggest that boninite melts equilibrate with upper mantle olivine and orthopyroxene (±clinopyroxene) at temperatures between 1200 and 1350 °C and pressures of ? ca. 1.0 GPa. (cf. Van der Laan et al., 1989; Umino and Kushiro, 1989; Falloon and Danyushevsky, 2000). Such estimates appear to be consistent with experimental data for evolved shoshonitic compositions (Thompson, 1977; Nicholls and Whitford, 1983) and sanukitoids, a potassic variant of high-Mg andesite (Tatsumi, 1982; Tatsumi and Ishizaka, 1982), assuming that, prior to fractionation, these also equilibrated with harzburgitic or refractory lherzolitic (±phlogopite) residua.

In contrast, the relatively high content of FeO* and TiO_2 (and other HFSE), and results from near-liquidus experimental studies of primitive lamproite (e.g. Edgar and Vukadinovic, 1993; Foley, 1993), suggest these magmas tap relatively fertile sources, albeit similarly contaminated, at pressures of ca. 3.0 GPa or more. According to Foley (1993), lamproites represent the hybrid melt products of peridotitic pyroxene and garnet, and other interstitial vein minerals. However, others have argued on the basis of REE mass balances that lamproites typically equilibrate with garnet peridotite (e.g. Tainton and McKenzie, 1994; Sato, 1997; Inoue et al., 1998). A third view holds that lamproites are not primary at all but represent hybrids of evolved (high-K, low-Mg) melts and relict mantle xenocrysts (Mitchell, 1985; Mitchell and Bergman, 1991) but appears to be inconsistent with the whole-rock Mg# values of Bucak lamproites.

5.2. Implications of HFSE distributions

In support of the second of these models Sato (1997) proposed that lamproites tap the lowermost, garnet-bearing regions of the lithospheric mantle, metasomatized by potassic asthenospheric melts (Boyd et al., 1993; Pearson et al., 1995). The substantially lower equilibration pressures inferred for primitive shoshonites appear to support this view (e.g. Gallagher and Hawkesworth, 1992; Turner et al., 1996, 1999) although this need not be the case if substantial lithosphere thinning has occurred as a result of mantle temperature fluctuations or lithospheric stretching. Boundary layer models for colliding plate margins and their contingent stress fields are not well constrained, however, given the lack of information about mantle flow and the detachment and tearing of subducting lithospheric slabs. At least two questions need to be addressed: firstly whether shoshonite sources, if located within the lithospheric mantle, reflect earlier, perhaps ancient, metasomatic processes (e.g. Boyd et al., 1993; Pearson et al., 1995) tapped in response to the thinning of collision-thickened lithosphere (e.g. as proposed by Pe-Piper and Piper, 1992; Gallagher and Hawkesworth, 1992; Turner et al., 1996, 1999; following England and Houseman, 1989), or whether the shoshonite sources are generated during a collision episode, perhaps in response to crustal underplating, lithospheric mantle delamination, or sediment subduction, prior to post-collision magma formation (e.g. Varne, 1985; Serri et al., 1993; Flower et al., 1998).

Several workers have argued that the persistence of HFSE depletions in post-collision magmas bears fundamentally on these questions (Conticelli, 1998; Peccerillo, 1999; De Astis et al., 2000). If the observed HFSE depletions reflect selective slab-derived enrichment in LILE and LREE (e.g. Morris and Hart, 1983; Davidson et al., 1990; Conticelli and Peccerillo, 1992; Hawkesworth et al., 1993; Pearce et al., 1999), partial melting that leads to magma formation in the mantle wedge involves either significantly larger melt fractions than those producing MORB or OIB, or peridotite sources that have been previously depleted. If, on the other hand, HFSE depletions in mantle wedge magmas represent the buffering effects of HFSE-rich accessory phases (e.g. Green and Pearson, 1986; Stolz et al., 1988, 1990; Reagan and Gill, 1989; Foley and Wheller, 1990; Thirlwall et al., 1994; Elliott et al., 1997), their phase equilibrium relationships, with respect to partial melt fraction and bulk composition, need to be reconciled with the spatial-temporal distribution of HFSE-depleted and non-depleted magmas.

The possibility that HFSE distributions are an instrinsic attribute of a heterogeneous mantle wedge (e.g. Woodhead et al., 1993, 2001) has strong appeal for a number of reasons. An important clue is the observation that HFSE abundances in subduction-related and post-collision magmas, while decoupled from those of LILE and LREE, show a close correlation with 'basaltic' elements such Fe, Al, Ca, and Na (e.g. Nguyen and Flower, 2004). This observation has been corroborated by mass balance models developed for mantle compositions of distinct fertility (as expressed by Mg# and basaltic element contents) using the MELTS algorithm (Nguyen and Flower, 2004), and supports the contention that source fertility is a major, if not the dominant, control on HFSE abundances in mantle wedge regions (e.g. Duncan and Green, 1987; McCulloch and Gamble, 1991; Gamble et al., 1993; Woodhead et al., 1993, 2001).

The notion that magmatic HFSE abundances reflect source fertility, rather than slab-derived contamination or relict phase effects, is consistent with both of the scenarios described above (e.g. Thompson, 1977; Nicholls and Whitford, 1983; Pe-Piper and Piper, 1992; Turner et al., 1996, 1999; cf. Serri et al., 1993; Flower et al., 1998). Each appeals to refractory peridotite sources for orogenic (HFSE-depleted) ultrapotassic magmas – the first requiring in situ partial melting of the continental lithospheric mantle as a response to collision-induced extension (e.g. Pe-Piper and Piper, 1992; Turner et al., 1996, 1999), the second implying a compositionally similar source within the underlying asthenosphere, with contamination of the latter by refractory peridotite, in addition to slab-derived components (Kay and Kay, 1993; Serri et al., 1993; Lavecchia and Stoppa, 1996; Flower et al., 1998). In both scenarios, partial melting occurs in response to plate collision. However, the first of these implies that contamination of the source may be an unrelated phenomenon, reflecting an earlier, perhaps ancient, fractionation history (e.g. Pe-Piper and Piper, 1992; Turner et al., 1996, 1999). On the other hand, if magmas tap enriched sources in the convecting wedge it is implicit that both contamination and melting are related to a collision in progress (Serri et al., 1993; Flower et al., 1998). The first case is consistent with the known melt-derivative compositions of the ambient lithospheric mantle, as sampled by basalt-borne xenoliths, and is considered to be partly residual to earlier episodes of partial melting. On the other hand, near-contemporaneous enrichment of the HFSE-depleted ultrapotassic sources is supported by their near-universal correspondence to plate collisions and by the lack of such enrichment in most basalt-borne mantle xenolith populations (Nguyen and Flower, 2004).

5.3. Significance of lamproite

In Fig. 10, incompatible element compositions of orogenic and non-orogenic magmas from the KAI province are compared with ILB suites from the Eifel-western Germany (Worner et al., 1986), the East African Rift, northeastern China, southern Brazil, and Central Spain. In Fig. 10a and b, the Bucak lamproites and ILB suites conform to a diffuse mixing curve that includes (non-enriched) N-MORB and averages for 'bulk continental crust' (BCC) (Rudnick and Fountain, 1995) and 'global subducting sediments' (GLOSS) (Plank and Langmuir, 1998), distinct from the relatively Nb- and Ta-depleted orogenic compositions. In contrast, the range of Ce/Pb in lamproite and ILB (Fig. 10c) are greater than in orogenic KAI magmas, and also significantly higher than that of BCC (Rudnick and Fountain, 1995), GLOSS (Plank and Langmuir, 1998) and slab-derived fluids (Keppler, 1996; Ayers, 1998). Variation of Nb/La–Ba/La (Fig. 10d) shows similar relationships, suggesting that fluid and crustal contamination was superimposed on an existing heterogeneous array. This interpretation is supported by a series of simple mixing models involving hypothetical compositional end-members: 'average N-MORB', 'average OIB', 'EM' (enriched mantle), and 'uncontaminated refractory mantle' (the average of 'non-enriched' Tonga arc lava compositions, summarized in Fig. 11 in terms of Ba/Nb–Y/Nb).

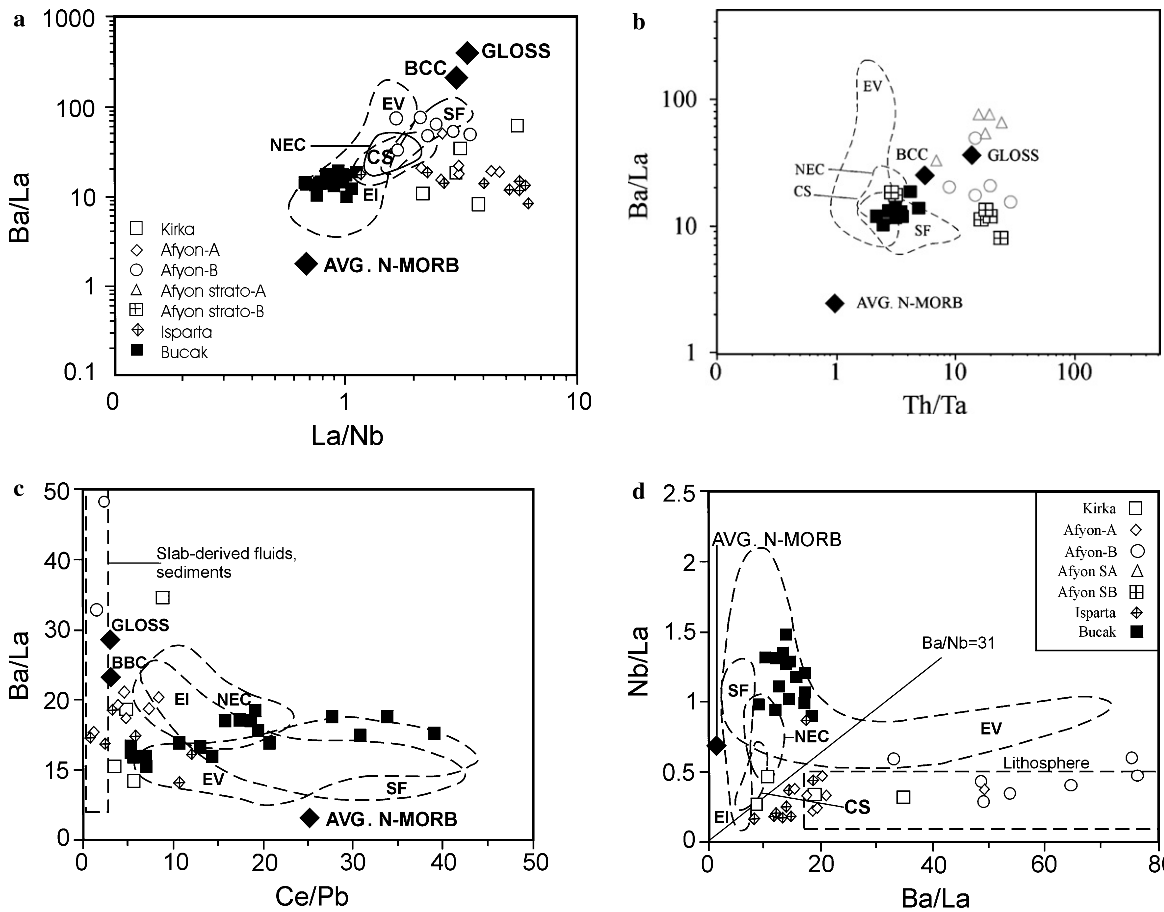

Fig. 10. Variation of trace element ratios for Kırka–Afyon–Isparta magmas, in comparison with fields for representative ILB suites: (a) Ba/La–La/Nb, (b) Ba/La–Th/Ta, (c) Ba/La–Ce/Pb, and (d) Nb/La–Ba/La. Data sources as in Fig. 5. N-MORB and averages for 'bulk continental crust' (BCC) after Rudnick and Fountain (1995). Global subducting sediments (GLOSS), after Plank and Langmuir (1998), and slab-derived fluids after Keppler (1996), Ayers (1998), Ben Othman et al. (1989). Lithosphere field is adapted from Menzies and Kyle (1991), and reference therein.

Variation of KAI data appears to confirm the inferences based on the major element variation (Fig. 7a) that parental lamproite melts and Group B Afyon ultrapotassic lavas reflect partial melting of relatively fertile, Ti-rich sources, while the low-Ti shoshonitic parent magmas tapped relatively refractory sources, Kirka and Afyon potassic series mostly reflecting sources of intermediate fertility. Despite the broad range of magmatic segregation pressures suggested by experimental studies of shoshonites and lamproites, the case for asthenospheric rather than lithospheric sources is supported by the similarity of LILE, LREE patterns in the KAI magmas (implying similar mantle contaminants) and the degree of covariation between high-field strength and 'basaltic' element abundances (e.g. Fig. 7). Their shared geodynamic setting and spatial-temporal associations suggest that the pattern of magmatic activity represents a progressive response to thermal and compositional changes in the mantle wedge. We conclude, on the basis of both major and trace element variation (Figs. 7, 10 and 11), that the Bucak lamproites, in common with most ILB suites, tapped strongly contaminated, relatively fertile parts of an evolving mantle wedge. The large pressure differential of lamproites and shoshonites probably reflects several factors, including the significantly greater buoyancy of garnet-free refractory mantle, its greater susceptibility to slab-derived hydration, assuming it

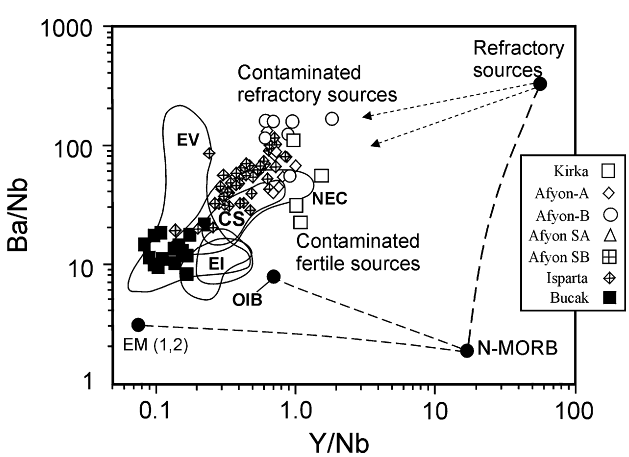

Fig. 11. Variation of Ba/Nb versus Y/Nb for Kırka–Afyon–Isparta magmatic compositions, in comparison with fields for representative ILB suites. Calculated model mixing curves are shown between endmember compositions: average N-MORB, average OIB, 'EM' (enriched mantle); and 'uncontaminated refractory mantle' (the average of 'non-enriched' Tonga arc lava compositions). Data sources as in Fig. 5.

to have been delaminated from the overriding plate, and the likelihood that in addition to phlogopite, amphibole breakdown played a significant role in triggering adiabatic decompression of shoshonitic sources.

6. Geodynamic synthesis

Geophysical studies suggest that this process is accompanied by detachment of the subducting oceanic slab or, if collision was diachronous, lateral slab-tearing (Wortel and Spakman, 2000; Sandvol et al., 2001, 2003; personal communication), contamination of a variably fertile mantle wedge, and a progressive increase in the depth of partial melting. In the Isparta Angle region a sharp decrease in seismic attenuation is consistent with a significant decrease in crustal thickness while slab-tearing appears to be indicated by the presence of anomalous low upper mantle Pn velocities (Sandvol et al., 2001, 2003; personal communication; Al-Lazki et al., 2004). Our preferred geodynamic model, developed from earlier hypotheses (Venturelli et al., 1984; Robert et al., 1992; Flower et al., 1998), assumes that incipient subduction and/or delamination of continental material (Molnar and Gray, 1979; Lallemand, 1995) are common features of colliding plate margins. We propose that hydrated, crust-enriched, refractory lithospheric mantle is removed from the overriding plate by tectonic erosion, due to the combined effects of orogenic thickening (Sacks and Secor, 1990; Platt and England, 1994) and asthenosphere flow, resulting from slab detachment (Davies and Von Blanckenburg, 1995).

The effects of mantle hydration in determining mantle flow during and following a plate collision are shown schematically in Fig. 12a–e (after Davies and Von Blanckenburg, 1995), assuming post-collision slab breakoff is accompanied by continuing slab dehydration. Assuming that dehydration occurs mostly at <ca. 1.0 GPa pressure, refractory lithospheric mantle of the overriding plate (rather than the convecting mantle wedge) is likely to be incorporated by the ductile asthenospheric, despite the greater buoyancy of the former. The progression from low-K calc-alkaline sources to high-K refractory to fertile sources can be considered accordingly (Fig. 12a–e). As subduction of African oceanic lithosphere beneath accreting Eurasia gives way to collision (Fig. 12a) refractory continental lithospheric mantle incorporated by ductile 'corner flow' is dragged down to ca.100 km depth where amphibole dehydration allows for incipient melting and a viscosity 'crisis'. Calc-alkaline magmas are produced by a combination of rapid gravitational rebound of buoyant, low-viscosity mantle and 'back-arc' extension. Slab detachment, orogen thickening, and continental crust underplating (Fig. 12b) are accompanied by: (i) continued delamination of refractory lithospheric mantle and lower continental crust by asthenospheric 'corner flow'; (ii) influx of fertile asthenosphere; and (iii) mixing of the latter with the contaminated mantle wedge. High-K calc-alkaline melts result from the combined effects of amphibole and (where

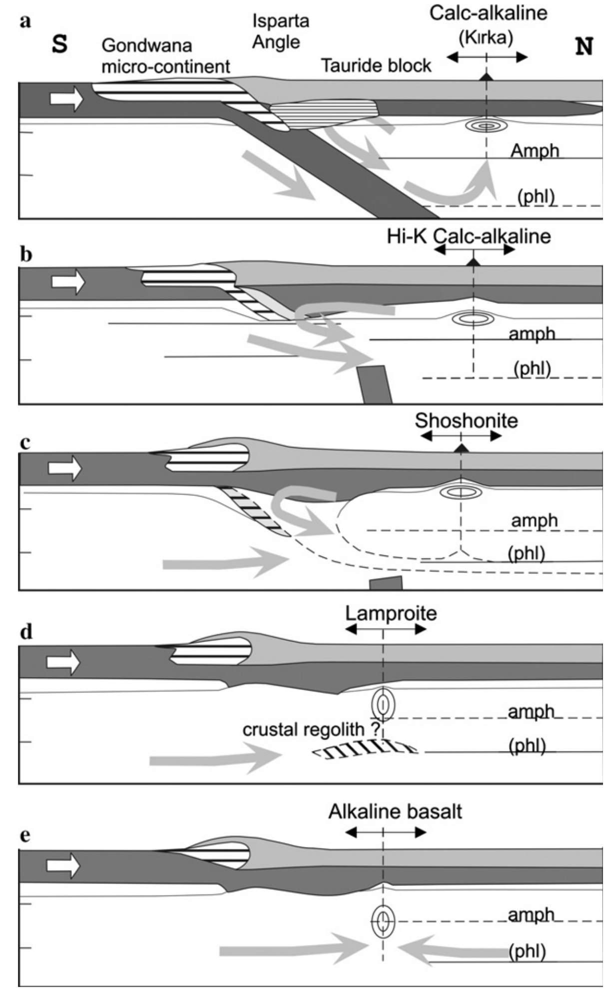

Fig. 12. Geodynamic model showing slab breakoff with formation of asthenospheric 'window', dehydration of 'subducted' continental crust and hydration of overlying refractory lithospheric mantle, thickening of hydrated mantle in response to local viscosity decrease and coupled asthenosphere flow convergence, initiation of supra-slab extension. Lithosphere structure and mantle flow dynamics associated with syn- and post-collision slab breakoff. Adapted from Davies and Von Blanckenburg (1995). (a) Pre-syn-collision subduction of African oceanic lithosphere beneath accreting Eurasia, showing mantle wedge 'corner flow', hydration and consequent delamination of refractory continental lithospheric mantle; dehydration of downward-driven, delaminated CLM (ca. 100 km depth), incipient melting. Primitive calcalkaline magma are produced by rapid decompression in response to the gravitational rebound of low-viscosity, refractory mantle and 'back-arc' extension (b) Slab detachment, orogen thickening, and continental crust underplating are accompanied by (i) continued delamination of refractory lithospheric mantle and lower continental crust by asthenospheric 'corner flow', (ii) influx of fertile asthenosphere, and (iii) mixing of the latter with the contaminated mantle wedge. High-K calcalkaline melts result from the combined effects of amphibole and (where present) phlogopite dehydration (at ca. 100 and 200 km, respectively) and 'back-arc' extension. (c) Phlogopite-enrichment of subcontinental asthenosphere follows continued incorporation of delaminated lithospheric mantle and crust, increasing reservoir of refractory, crust-contaminated magma sources. (d) Continued development of ultrapotassic mantle sources as with mixing of fertile asthenosphere with relict crustal components. (e) Re-establishment of crustal contaminant-free asthenospheric magma sources tapped in response to post-collision thermal anomalies and collision-related lithosphere extension.

present) phlogopite dehydration (at ca.100 and 200 km, respectively) and 'back-arc' extension. As phlogopite-enrichment (from relict crustal components) persists, refractory, crust-contaminated magma sources accumulate while becoming progressively mixed with newly-influxed asthenosphere (Fig. 12c–d). The eventual re-establishment of contaminant-free asthenosphere is then tapped in response to post-collision thermal perturbations and collision-related lithosphere extension (Fig. 12e). This model is preferred to others (e.g. Turner et al., 1996) which imply that partly hydrated lithospheric mantle reflects long term enrichment by melt fractions percolating from trenchward-flowing asthenosphere or from an earlier convergent margins (e.g. McKenzie, 1989), and is favored here in view of its ability to transmit collision-enhanced ('shoshonitic') sediment signatures from a sub-orogenic region to the site of shoshonite melt segregation and the opportunity for decompression melting offered by foreland stretching and possible circulation of hot asthenosphere cells (Smith, 1992; Platt and England, 1994; Wu and Levshin, 1994; Lave et al., 1996).

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