

# The Thickness of the Mantle Lithosphere and Collision-Related Volcanism in the Lesser Caucasus

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## ABSTRACT

The Lesser Caucasus mountains sit on a transition within the Arabia–Eurasia collision zone between very thin lithosphere (<100 km) to the west, under Eastern Anatolia, and a very thick lithospheric root (up to 200 km) in the east, under western Iran. A transect of volcanic highlands running from NW to SE in the Lesser Caucasus allows us to look at the effects of lithosphere thickness variations on the geochemistry of volcanic rocks in this continental collision zone. Volcanic rocks from across the region show a wide compositional range from basanites to rhyolites, and have arc-like geochemical characteristics, typified by ubiquitous negative Nb–Ta anomalies. Magmatic rocks from the SE, where the lithosphere is thought to be thicker, are more enriched in incompatible trace elements, especially the light rare earth elements, Sr and P. They also have more radiogenic <sup>87</sup>Sr/<sup>86</sup>Sr, and less radiogenic <sup>143</sup>Nd/<sup>144</sup>Nd. Across the region, there is no correlation between SiO<sub>2</sub> content and Sr–Nd isotope ratios, revealing a lack of crustal contamination. Instead, ‘spiky’ mid-ocean ridge basalt normalized trace element patterns are the result of derivation from a subduction-modified mantle source, which probably inherited its subduction component from subduction of the Tethys Ocean prior to the onset of continent–continent collision in the late Miocene. In addition to the more isotopically enriched mantle source, modelling of non-modal batch melting suggests lower degrees of melting and the involvement of garnet as a residual phase in the SE. Melt thermobarometry calculations based on bulk-rock major elements confirm that melting in the SE must occur at greater depths in the mantle. Temperatures of melting below 1200°C, along with the subduction-modified source, suggest that melting occurred within the lithosphere. It is proposed that in the northern Lesser Caucasus this melting occurs close to the base of the very thin lithosphere (at a depth of ~45 km) as a result of small-scale delamination. A striking similarity between the conditions of melting in NW Iran and the southern Lesser Caucasus (two regions between which the difference in lithosphere thickness is ~100 km) suggests a common mechanism of melt generation in the mid-lithosphere (~75 km). The southern Lesser Caucasus magmas result from mixing between partial melts of deep lithosphere (~120 km in the south) and mid-lithosphere sources to give a composition intermediate between magmas from the northern Lesser Caucasus and NW Iran. The mid-lithosphere magma source has a distinct composition compared with the base of the lithosphere, which is argued to be the result of the increased retention of metasomatic components in phases such as apatite and amphibole, which are stabilized by lower temperatures prior to magma generation.

**Key words:** collision setting; Lesser Caucasus; magma petrogenesis; mantle lithosphere thickness; Sr–Nd isotopes

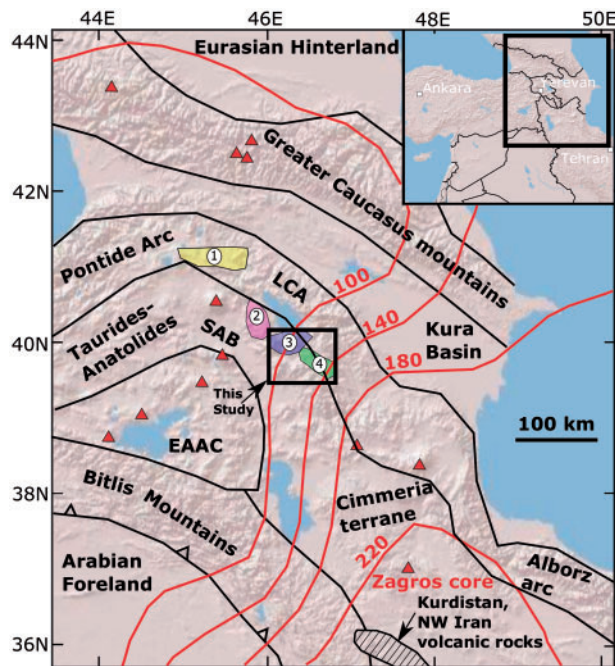
## INTRODUCTION

The Arabia–Eurasia collision zone is one of the very few places on Earth where it is possible to study active volcanism associated with a continent–continent collision event. The geodynamic processes that drive volcanism remain unresolved, with numerous and sometimes conflicting models for its origin (Pearce *et al.*, 1990; Keskin, 2003; Göğüş & Pysklywec, 2008; Faccenna *et al.*, 2013). Most of the volcanism in the region is located on the Anatolian–Armenian–Iranian Plateau, a broad uplifted region to the north and east of the Arabian foreland, with an average elevation of over 2 km (Sengür *et al.*, 2008; Priestley & McKenzie, 2013). The lithospheric structure of this plateau is considered to show a strong contrast between very thin mantle lithosphere in the west (below Eastern Anatolia), and very thick

mantle lithosphere in the Zagros Core to the SE (Fig. 1), below western Iran (Priestley *et al.*, 2012).

The Lesser Caucasus mountains sit close to the edge of the Zagros Core region, and therefore close to a transition from thick to thin lithosphere. The region thus provides an opportunity to look at the influence of lithospheric thickness on the geochemistry of collision-related magmas. Figure 1 shows the four volcanic highlands considered in this study, which form a NW–SE transect almost orthogonal to contours of lithospheric thickness from Priestley *et al.* (2012), which increase from NW to SE. It should be noted that the resolution on these lithospheric thickness estimates is limited by the 30–50 km vertical resolution of the seismic tomography data (McKenzie & Priestley, 2008). The Priestley *et al.* (2012) model is used because it is likely to give better local resolution in the Lesser Caucasus region than other global lithospheric thickness studies (Priestley & McKenzie, 2006, 2013).

Volcanic rocks from the SE of this transect, where the lithosphere is thought to be thicker, are known to be more potassic than volcanic rocks from the NW (Meliksetian, 2013). This study provides the first complete geochemical dataset for volcanic rocks from the SE of the Lesser Caucasus. This dataset includes a complete range of compositions from basanite to rhyolite, which are used to evaluate the extent to which crustal contamination is an important component of magma petrogenesis. We then compare the geochemistry of the more primitive mafic volcanic rocks between the NW and SE of the Lesser Caucasus, and the mechanisms by which thicker lithosphere in the SE might influence the geochemistry of volcanic rocks found there. To indicate how thicker mantle lithosphere might influence the composition of magmas, volcanic rocks from a region of very thick lithosphere (>200 km) in NW Iran (Fig. 1; Allen *et al.*, 2013) are used as an end-member comparison of melting in a thick lithosphere regime.

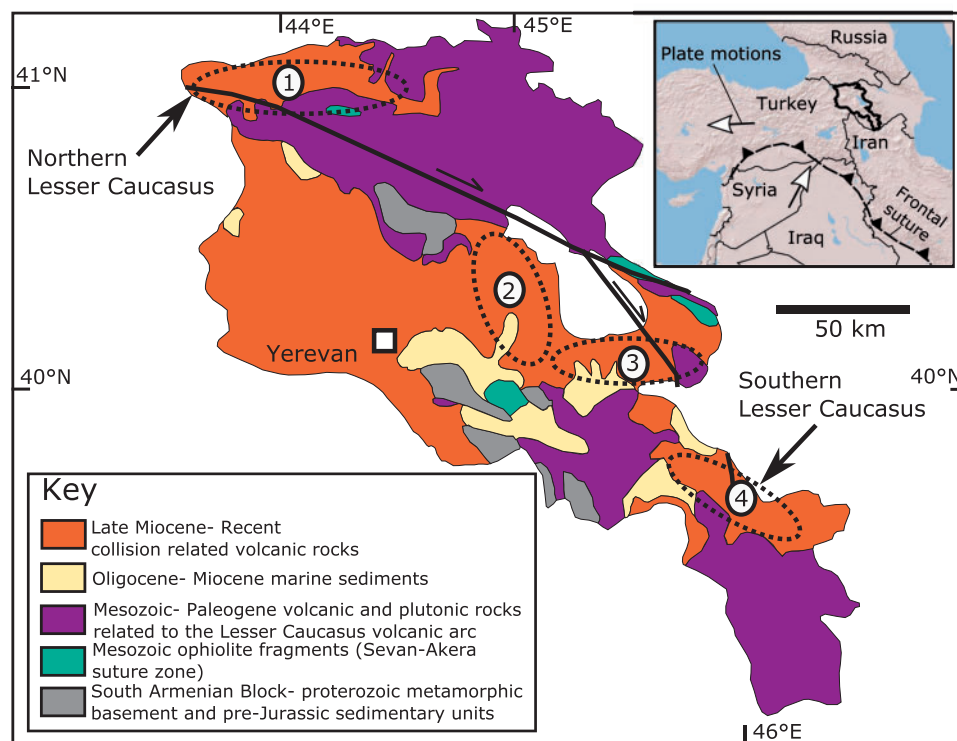


**Fig. 1.** Terrane map of the Caucasus region with inset showing the location of the Caucasus mountains. The major tectonic blocks are labelled, which from north to south are: Greater Caucasus, Kura Basin, Pontides–Lesser Caucasus (LCA)–Alborz Mesozoic–early Cenozoic arc, Taurides–Anatolides–South Armenian Block (SAB)–Cimmeria microcontinent terranes, East Anatolian Accretionary Complex (EAAC), Bitlis Mountains and the Arabian foreland. Terrane boundaries are after Neill *et al.* (2015). Red triangles denote the locations of major composite volcanoes. The locations of volcanic highlands that are used for the geochemical comparison that forms the focus of this study are shown by the coloured fields: 1, Shirak and Lori (Neill *et al.*, 2013, 2015); 2, Gegham (I. P. Savov, unpublished data); 3, Vardenis; 4, Syunik. New data for this study are from volcanic highlands 3 and 4. These volcanic highlands are close to parallel to major tectonic boundaries in the collision zone. Volcanic rocks from NW Iran referred to in the text are from the region labelled at the bottom of the map. Contours of lithospheric thickness from Priestley *et al.* (2012) are shown in red; numbers give lithospheric thickness in kilometres. The ‘Zagros Core’ refers to the region of maximum lithosphere thickness in the collision zone.

## GEOLOGICAL BACKGROUND

### Pre-collision geological history

The evolution of the Arabia–Eurasia collision zone, as a part of the Alpine–Himalayan orogenic belt, is a consequence of the closure of the Neotethys Ocean (Rolland, 2017, and references therein). The pre-Miocene geology of the territory of Armenia and the Lesser Caucasus (Fig. 2), like much of the interior of the Anatolian–Armenian–Iranian Plateau, is a complex amalgamation of a series of terranes (microplates), which accreted to the Eurasian continental margin during the closure of the Tethys Ocean (Hosseinpour *et al.*, 2016; Rolland, 2017). The north and east of Armenia include rocks associated with the Mesozoic to early Cenozoic volcanic arc of the Lesser Caucasus (Mederer *et al.*, 2013), an eastern continuation of the Pontides of Anatolia. The Pontide and Lesser Caucasus arcs together formed the active southern margin of the Eurasian Plate (Yilmaz *et al.*, 2000). Figure 3 illustrates the various stages of



**Fig. 2.** Geological map of the major geological units of the Lesser Caucasus mountains within the territory of Armenia. After Kharazyan (2005), Mederer *et al.* (2013) and Neill *et al.* (2015). The volcanic highlands that form the focus of this study are numbered: 1, Shirak and Lori; 2, Gegham; 3, Vardenis; 4, Syunik. The bold black line shows the Pambak–Sevan–Syunik strike-slip fault zone.

closure of the Neotethys Ocean. During the Mesozoic there were probably several subduction zones that contributed to the closure of the northern and southern Neotethys basins (Fig. 3, panels 1 and 2; Galoyan *et al.*, 2009; Rolland *et al.*, 2010, 2012; Sosson *et al.*, 2010; Karaođlan *et al.*, 2013, 2016; Mederer *et al.*, 2013; Topuz *et al.*, 2013a, 2013b, 2014; Hässig *et al.*, 2015).

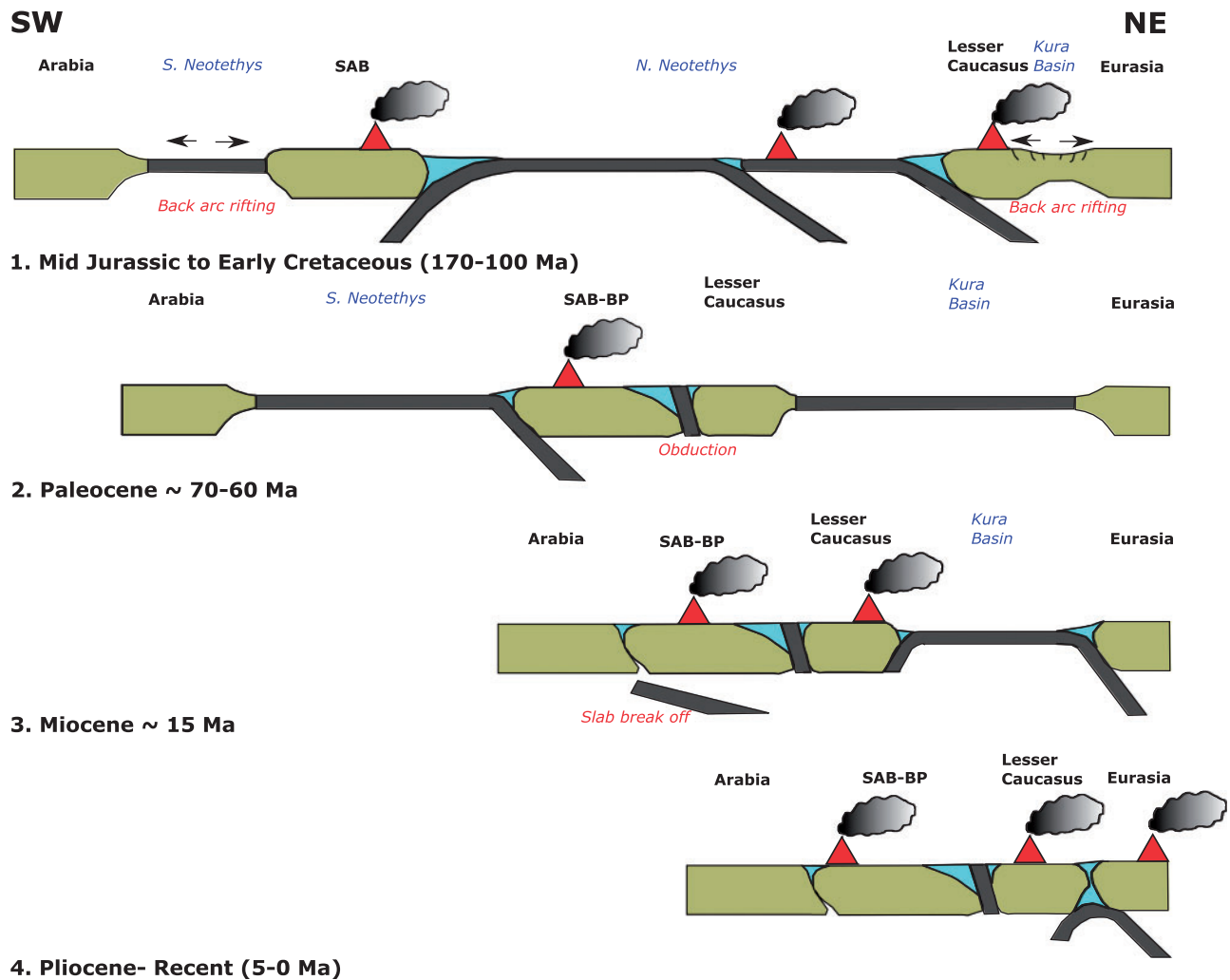
The Sevan–Akera suture zone is defined by several (probably North Neotethys-derived) ophiolite complexes (Fig. 2). The ophiolites were obducted onto the South Armenian block (SAB) at 88–83 Ma (Galoyan *et al.*, 2007; Rolland *et al.*, 2010; Sosson *et al.*, 2010). The SAB (Fig. 2) is a microcontinental fragment composed of Proterozoic metamorphic basement and its sedimentary cover (Şengör, 1984; Rolland *et al.*, 2009), which rifted from the Arabian margin in the early Mesozoic (Şengör, 1984; Jrbashian *et al.*, 1996) and is probably continuous with the Taurides–Anatolides of Anatolia (Rolland *et al.*, 2016) and the Cimmerian terrane of Iran (Stampfli *et al.*, 2013). Ophiolite obduction was followed by accretion of the SAB to the Pontide–Lesser Caucasus arc at 50–80 Ma (Rolland *et al.*, 2009; Sosson *et al.*, 2010), marking the closure of the North Neotethys. Northward subduction of the South Neotethys continued until ‘soft’ collision of Arabia with the Bitlis–Poturge terrane and SAB in the late Eocene–early Oligocene (Fig. 3, panel 3; Rolland *et al.*, 2012; Karaođlan *et al.*, 2013). Hard collision was delayed until

the Pliocene, following the Kura Basin closure to the north (Fig. 3, panel 4; Rolland, 2017).

The numerous subduction zones illustrated in Fig. 3 would have added slab-derived material to large parts of the mantle wedge beneath the present-day collision zone. It has been suggested that this signature of a subduction-modified mantle has been widely inherited by the post-collisional volcanic rocks in the Arabia–Eurasia collision zone (Pearce *et al.*, 1990; Keskin *et al.*, 1998; Keskin, 2003; Allen *et al.*, 2013; Oyan *et al.*, 2017).

### Post-Miocene volcanism and tectonics

The widespread volcanism occupies a wide zone across the collision zone (Fig. 1), arguing against any single subducting slab driving magma generation. Elevated topography of 1–2 km is observed across the collision zone; however, the deep lithospheric structure that might isostatically support this elevated topography shows strong contrasts between the east and west. In the west, a low seismic velocity anomaly in the uppermost mantle below the East Anatolian Plateau (Al-Lazki, 2003; Angus *et al.*, 2006; Zor, 2008; Gök *et al.*, 2011; Koulakov *et al.*, 2012; Skolbeltsyn *et al.*, 2014) has been used to suggest that the mantle lithosphere is very thin under Eastern Anatolia, such that the high topography is not isostatically supported. Based on the fairly uniform crustal thickness (~40 km; Angus *et al.*, 2006) and the very thin lithospheric mantle, both the current



**Fig. 3.** Illustration of the progressive closure of the oceanic domains that separated Arabia from Eurasia during the Mesozoic, after Rolland (2017). SAB, South Armenian Block; SAB-BP, South Armenian Block–Bitlis Poturge terrane. The movement of the Bitlis Poturge terrane from Arabia to the South Armenian Block is not shown in these figures but occurs between panels 1 and 2.

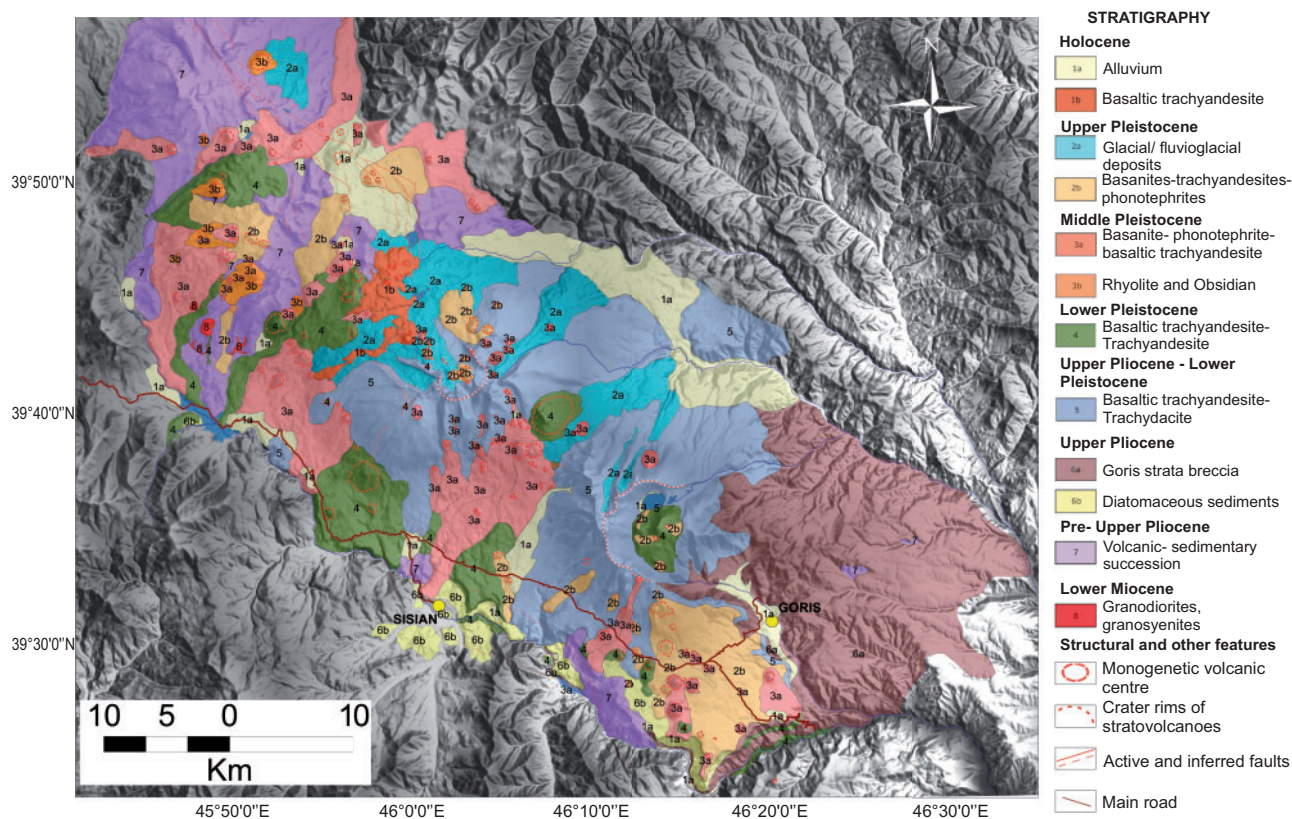
structure of this western part of the Plateau and the associated volcanism have been attributed to a slab break-off event (Keskin, 2003, 2007; Şengör *et al.*, 2003, 2008). Given the later onset of eruptions, and the more alkaline nature of magmas in the south (Pearce *et al.*, 1990; Keskin *et al.*, 1998), it has been postulated that the slab that broke off and thereby fuelled recent volcanism was a north-dipping South Neotethys slab below the Bitlis Suture (Keskin, 2003).

In contrast, relatively fast seismic velocities in the upper mantle under western Iran suggest that a thick lithospheric root isostatically compensates the high topography in the southeastern part of the collision zone, termed the ‘Zagros Core’ (Priestley & McKenzie, 2006, 2013; Priestley *et al.*, 2012). There is no evidence in NW Iran of the coupled spatio-temporal variability in volcanism seen in Eastern Anatolia, such that instead of slab break-off, it has been postulated that the volcanism is driven by small-scale lithospheric dripping or amphibole breakdown melting (Allen *et al.*, 2013; Kaislaniemi *et al.*, 2014; Lechmann *et al.*, 2018).

As can be seen in Fig. 1, the Lesser Caucasus is located close to the boundary between these two tectonic regimes. The transect of volcanic highlands thus provides a key opportunity to understand the effect of lithospheric thickness on the composition of mantle-derived magmas in a continental collision zone.

### Geology and geochronology of collision-related volcanism in the Lesser Caucasus

Three major styles of volcanic activity can be observed in the Lesser Caucasus. The first of these is relatively low-volume eruptions from mostly small eruptive centres in monogenetic volcanic fields. In both the north and south of the Lesser Caucasus, this style of volcanism becomes increasingly dominant in the most recent eruptive products (Fig. 4; Connor *et al.*, 2011; Neill *et al.*, 2013). Second, large composite volcanoes are also found throughout the Lesser Caucasus. In the central Lesser Caucasus, such a volcano is Aragats (Connor *et al.*, 2011), and in the south the smaller stratovolcanoes Tskhouk and Ishkhanasar (Fig. 4) dominate



**Fig. 4.** Geological age map of the Syunik volcanic highland (field 4 in Figs 1 and 2), based on the work of Sergei Karapetyan, Ruben Jrbashyan and Gevorg Navasardyan of the Institute of Geological Sciences at the Armenian National Academy of Sciences. Red dashes, monogenetic volcanic centres; red and white dashes, crater rims of Ishkhanasar and Tskhouk stratovolcanoes; faint red lines, active faults; bold red line, main road. A PDF version of this map is provided as [Supplementary Data C](#).

the landscape. Such large volcanoes (4090 m altitude, ~70 km diameter in the case of Aragats), capable of generating many caldera collapse eruptions, are required to produce the ignimbrites that are widespread throughout the Lesser Caucasus. Finally, large fissure-fed ‘flood basalt’ style lava flows are found predominantly in some of the older (~3–2.05 Ma) volcanic successions (Sheth *et al.*, 2015).

Previous geochronological studies have constrained the ages of the volcanic deposits upon which this study focuses as being late Pliocene or younger. All of the samples analysed for geochemistry from the NW of the transect have ages of 2.5 Ma or younger, based on correlations of the oldest ‘valley series’ lavas of Neill *et al.* (2015) with sediments containing mammalian fossils (Kharazyan, 1983), and a K–Ar age of 2.5 Ma for one such ‘valley series’ lava in the Shirak province of NW Armenia in the northern Lesser Caucasus (Chernyshev *et al.*, 2002). Volcanic rocks from further south were all collected from Quaternary volcanic highlands, which form monogenetic volcanic fields and so have an even more restricted age range. A limited number of ages from the Syunik volcanic highland (number 4 in Figs 1 and 2) suggest that volcanism ranges from 1.3 to 0.11 Ma, based on two Ar–Ar ages of pumice layers in diatomaceous sediments (Joannin *et al.*, 2010), along

with K–Ar ages of local lava flows that overlie the sediments (Ollivier *et al.*, 2010). Archaeological evidence from  $^{14}\text{C}$  dating of petroglyphs and burial places around the youngest lavas in Syunik suggests that volcanism may extend to within the last 5 kyr (Karakhanian *et al.*, 2002). These ages are very similar to the estimated age span of the Gegham volcanic highland (highland 2 in Figs 1 and 2) of 1.2–0.02 Ma (Lebedev *et al.*, 2013). In summary, the complete age span of the volcanic rocks studied for geochemistry in this study is <3 Myr, and it is likely that the present-day structure of the lithosphere can be used as an interpretive framework for all samples.

## SAMPLING

Whereas recent work (Neill *et al.*, 2013, 2015) has provided important new geochemical data for collision-related volcanism in the northern Lesser Caucasus, our study presents the first comprehensive dataset for mafic volcanic rocks from the southern part of the Lesser Caucasus (Table 1). All samples are from the Vardenis and Syunik volcanic highlands (highlands 3 and 4 in Figs 1 and 2). Table 1 gives the coordinates for all the mafic samples in the Vardenis and Syunik volcanic highlands. Figure 4 shows the newly completed

**Table 1:** Major and trace element compositions of mafic volcanic rocks from the Vardenis and Syunik volcanic fields in the southern Lesser Caucasus

Volcanic field: Sample: Eruptive type: Latitude: Longitude: Elevation (m): Lab. analysis:*	Vardenis 6.27.08 Lava 40.0905 45.414 2257 RH	Vardenis 7.29.08 Lava 40.14205 45.6192 2101 RH	Vardenis 7.30.08 Bomb 40.0768 45.51587 2340 ACME	Vardenis 7.32.08 Lava 40.07712 45.5256 2280 RH	Vardenis 2.1.12 Lava 40.10283 45.39072 2188 ACME	Vardenis 1.1.13 Lava 40.02458 45.83628 2347 ACME	Vardenis 1.2.13 Lava 40.08497 45.83638 2375 ACME	Vardenis 3.2.13 Lava 40.05927 45.79923 2538 ACME	Vardenis 3.3.13 Lava 40.07163 45.78805 2539 ACME
SiO <sub>2</sub>	51.62	50.94	53.96	52.87	52.17	51.71	51.93	52.81	51.60
TiO <sub>2</sub>	1.20	1.34	1.01	1.23	1.11	1.39	1.38	1.29	1.37
Al <sub>2</sub> O <sub>3</sub>	16.31	16.37	16.47	16.93	17.40	17.31	17.39	16.42	17.23
Fe <sub>2</sub> O <sub>3</sub> (tot)†	9.61	9.69	6.82	9.49	7.77	8.34	8.39	7.87	8.25
MnO	0.13	0.14	0.12	0.14	0.13	0.13	0.13	0.14	0.13
MgO	6.23	6.36	4.08	4.40	4.60	4.15	4.12	4.03	4.12
CaO	8.54	8.43	7.97	7.93	8.27	7.34	7.32	7.49	7.34
Na <sub>2</sub> O	3.93	4.54	4.36	4.61	4.62	4.52	4.61	4.65	4.85
K <sub>2</sub> O	2.27	2.00	3.37	2.13	2.76	3.08	3.08	3.11	3.13
P <sub>2</sub> O <sub>5</sub>	0.52	0.72	0.61	0.77	0.95	1.23	1.23	0.97	1.21
Total	98.45	99.56	99.49	99.56	100.25	100.09	100.34	99.62	99.66
LOI			0.59		0.35	0.75	0.61	0.7	0.27
Mg#	56.2	56.5	54.2	47.9	54.0	49.6	49.3	50.4	49.7
Ni	77.6	86.8	18.8	45.9	32.8	12	32	7.1	23.1
La	39	57	52	58	84	64	96	21	79
Ce	76	99	91	102	148	112	180	35	151
Pr	9.0	10.8	9.9	11.0	16.3	12.1	20.6	3.7	16.8
Nd	37	41	34	42	57	41	73	13	59
Sm	6.6	7.3	5.5	7.6	8.0	6.4	10.0	2.5	8.9
Eu	1.9	2.1	1.5	2.1	2.2	1.7	2.6	0.8	2.4
Gd	5.0	5.8	4.5	5.9	6.1	5.0	7.4	2.5	6.7
Tb	0.8	1.0	0.6	1.0	0.7	0.7	0.8	0.4	0.8
Dy	3.6	4.3	3.7	4.7	3.8	3.5	4.5	2.1	4.5
Ho	0.7	0.8	0.7	0.9	0.7	0.6	0.7	0.4	0.8
Er	1.9	2.2	2.0	2.3	1.7	1.8	1.9	1.3	2.3
Tm	0.3	0.4	0.3	0.4	0.3	0.3	0.3	0.2	0.3
Yb	2.0	2.5	1.9	2.4	1.6	1.7	1.7	1.2	2.1
Lu	0.3	0.4	0.3	0.4	0.3	0.3	0.3	0.2	0.3
Ba	778	754	768	707	1098	881	1184	563	1162
Hf	3.4	4.1	4.4	4.1	4.3	4.5	4.6	2.8	4.4
Nb			18.9		28.3	22.2	31.6	8.2	27.6
Rb			67.7		41.1	56.9	35.4	57.3	42.5
Sr	1161	1217	845	1142	1890	1026	2084	495	1932
Pb	15.9	18.23		14.43					
Ta	1.1	1.1	0.9	1.0	1.2	1.2	1.5	0.6	1.2
Th	6.3	5.7	11.8	5.9	9.4	9.4	5.9	6.6	6.9
U	1.5	1.5	3.3	1.2	1.9	2.4	1.4	1.6	1.7
Zr	141	188	187	187	195	209	213	114	210
Y	19.4	24.9	19.9	26.8	18.9	18.5	21.2	12.5	23.6

(continued)

geological map of the Syunik volcanic highland; a PDF version is provided as [Supplementary Data C](#) ([supplementary data](#) are available for downloading at <http://www.petrology.oxfordjournals.org>). The map shows that even in this small area there is a wide range of volcanic structures and rock compositions, including large andesite–dacite stratovolcanoes, small basaltic scoria cones and rhyolitic monogenetic domes, often with obsidian flows. Sampling in the region covers the complete stratigraphic range of the map, but there was a bias towards the younger eruptive centres so as to collect the least altered samples for geochemical analysis. There is as yet no such similar geological map of Vardenis available; however, the sampling was similarly comprehensive in terms of rock compositions and stratigraphic range. The studied rock types range from

basanite to rhyolite, although [Table 1](#) includes data for only the most mafic samples, which form the main focus of this study. The complete dataset (including the more felsic samples) can be found in [Supplementary Data A](#).

## METHODS

### Major and trace elements

Sample preparation was done at the University of Leeds. Rock samples (60–100 g) were crushed in a TEMA agate mill. The agate was cleaned extensively between the crushing of each sample, including pre-contamination of the agate by the crushing of 50 g of sample, which was then discarded. Bulk-rock major and trace element analysis of samples from this study was

Table 1: Continued

Volcanic field:	Vardenis	Vardenis	Vardenis	Syunik	Syunik	Syunik	Syunik	Syunik	Syunik
Sample:	4.02.15	5.06.15	6.03.15	1.4A.08	2.6.08	2.7.08	2.8.08	2.9.08	2.10.08
Eruptive type:	Lava	Lava	Lava	Scoria	Bomb	Bomb	Lava	Bomb	Lava
Latitude:	40-17904	39-96908	39-98746	39-44133	39-48285	39-4587	39-4601	39-48452	39-46787
Longitude:	45-6186	45-68293	45-61326	46-08888	46-2693	46-24037	46-26768	46-2177	46-25762
Elevation (m):	1956	2872	3454	1876	1969	1849	1958	2107	1956
Lab. analysis:*	ACME	ACME	ACME	RH	RH	RH	RH	RH	RH
SiO <sub>2</sub>	51.51	51.47	53.73	50.52	53.23	50.44	51.74	51.58	47.08
TiO <sub>2</sub>	1.33	1.52	1.08	1.43	1.12	1.32	1.14	1.07	1.31
Al <sub>2</sub> O <sub>3</sub>	16.77	16.47	17.09	16.31	17.11	16.82	16.26	16.49	14.39
Fe <sub>2</sub> O <sub>3</sub> tott	8.57	8.60	7.60	9.97	8.65	9.31	8.75	8.46	11.45
MnO	0.14	0.14	0.13	0.14	0.13	0.14	0.13	0.13	0.15
MgO	6.15	4.56	4.06	5.20	4.15	4.44	4.77	4.71	11.65
CaO	8.33	8.59	7.24	8.74	6.87	8.16	8.61	7.03	9.83
Na <sub>2</sub> O	4.66	4.20	4.79	4.36	5.03	5.12	4.76	4.90	3.29
K <sub>2</sub> O	2.04	2.26	2.75	2.73	3.20	3.25	2.89	3.33	1.19
P <sub>2</sub> O <sub>5</sub>	0.73	0.93	0.90	1.02	0.96	1.12	1.00	0.93	0.53
Total	100.41	99.46	99.78	99.40	99.59	99.20	99.17	97.78	99.74
LOI	0.06	0.62	0.27						
Mg#	58.7	51.2	51.4	50.8	48.8	48.6	51.9	52.5	66.8
Ni	16.3	18.8	23.7	49.9	52.4	45.0	56.7	85.1	275.5
La	89	75	75	89	95	99	93	89	32
Ce	167	138	132	170	172	193	175	159	66
Pr	19.4	15.7	14.5	18.8	18.0	21.4	18.7	16.8	8.3
Nd	68	58	50	73	67	84	72	63	34
Sm	9.9	8.6	8.0	11.3	10.3	12.7	11.1	9.5	6.5
Eu	2.7	2.4	2.0	2.9	2.7	3.3	2.9	2.6	1.8
Gd	7.7	7.1	5.8	8.0	7.5	8.7	7.7	7.0	4.9
Tb	0.9	0.8	0.7	1.1	1.0	1.2	1.1	1.0	0.9
Dy	4.6	4.6	3.8	4.3	4.0	4.3	4.0	3.7	4.0
Ho	0.7	0.9	0.7	0.8	0.7	0.8	0.7	0.7	0.8
Er	2.1	2.4	2.0	2.0	2.0	2.0	1.9	1.8	2.1
Tm	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.4
Yb	2.0	2.1	1.8	2.0	2.0	2.0	1.9	1.9	2.2
Lu	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.4
Ba	1171	947	962	1109	1057	1220	1107	1192	430
Hf	5.5	4.5	4.3	4.5	5.1	4.9	5.4	4.7	3.0
Nb	28.3	20.0	25.8						
Rb	47.9	37.4	43.4						
Sr	1899	1537	1411	2003	1722	2531	1998	2050	1073
Pb				15.2	18.0	16.7	17.5	19.9	9.0
Ta	1.2	1.0	1.0	1.6	1.8	1.7	1.7	1.6	0.7
Th	7.1	6.9	7.1	6.5	8.2	6.8	8.4	7.8	2.6
U	1.6	1.5	1.4	1.4	1.7	1.5	1.5	1.7	0.7
Zr	245	202	215	199	234	224	246	220	126
Y	21.9	25.0	21.4	23.3	23.4	24.1	20.6	20.5	22.2

(continued)

done at ACME Labs by Bureau Veritas minerals, in Vancouver, Canada, and at Royal Holloway University, London.

For analysis at ACME Labs, samples were heated to 1000°C to determine loss on ignition (LOI), and then fused in a platinum–gold crucible with a lithium tetraborate flux. The resulting fusion beads were then analysed by X-ray fluorescence (XRF) for major elements. Two internal ACME standards reproduced expected values to better than 3% (for all oxides >1 wt %). Trace element concentrations were determined by inductively coupled plasma mass spectrometry (ICP-MS) on the fused beads after digestion in nitric acid. Analysis of Ni concentrations involved digestion in aqua regia at 95°C. Two internal ACME standards reproduced expected values to better than 10% for trace elements. Values for internal

and external standard data are shown in [Supplementary Data A](#).

Analysis at Royal Holloway was by inductively coupled plasma optical emission spectroscopy (ICP-OES) for major elements and some high-abundance trace elements (Sr, Zr, Ni) and ICP-MS for low-abundance trace elements. Major element analyses followed the methods described by [Walsh \*et al.\* \(1981\)](#) and [Garbe-Schönberg \(1993\)](#). The relative standard deviation on the external standards NIM-G, NIM-L, BHVO-1, RGM-1 and STM-1, as well as internal standards, was better than 5% (for all oxides >1 wt %). Trace elements were analysed after HNO<sub>3</sub>–HF–HClO<sub>4</sub> digestions. Prior to analysis, samples were spiked with 5 ng ml<sup>-1</sup> of indium (In) and rhenium (Re) for internal standardization. For analytical quality control, five international

Table 1: Continued

Volcanic field:	Syunik	Syunik	Syunik	Syunik	Syunik	Syunik	Syunik	Syunik	Syunik
Sample:	5.21.08	6.24.08	6.25.08	4.16.10	8.02.15	8.03.15	8.04.15	8.05.15	8.06.15
Eruptive type:	Lava	Scoria	Bomb	Lava	Bomb	Bomb	Lava	Bomb	Lava
Latitude:	39.75548	39.68745	39.68745	39.56577	39.42896	39.4141	39.49615	39.49431	39.50412
Longitude:	45.85745	45.91403	45.91403	46.21687	46.27307	46.28308	46.24583	46.24621	46.20926
Elevation (m):	2871	2514	2514	2731	1896	1603	2168	2185	2237
Lab. analysis:*	RH	RH	RH	RH	ACME	ACME	ACME	ACME	ACME
SiO <sub>2</sub>	53.46	48.97	48.91	53.53	49.16	51.58	52.01	45.80	50.11
TiO <sub>2</sub>	1.05	1.56	1.60	1.13	1.43	1.21	1.12	1.58	1.14
Al <sub>2</sub> O <sub>3</sub>	16.41	16.28	16.37	17.08	16.91	16.65	16.68	14.92	15.92
Fe <sub>2</sub> O <sub>3</sub> tott	8.62	10.06	10.14	8.48	9.03	7.66	7.26	9.56	7.28
MnO	0.13	0.14	0.14	0.12	0.14	0.13	0.13	0.16	0.13
MgO	4.84	4.77	4.66	4.02	5.37	4.29	4.54	7.77	4.44
CaO	7.12	8.63	8.81	7.71	8.66	8.01	7.23	10.02	8.60
Na <sub>2</sub> O	4.45	4.47	4.02	4.75	4.23	4.48	4.31	4.30	4.78
K <sub>2</sub> O	3.02	3.03	3.07	3.20	2.54	3.14	3.24	2.64	3.75
P <sub>2</sub> O <sub>5</sub>	0.92	1.30	1.36	0.93	1.24	1.06	0.98	1.27	1.18
Total	99.15	98.19	98.05	100.10	99.83	99.76	99.60	98.90	98.42
LOI					0.98	1.41	1.92	0.73	0.88
Mg#	52.7	48.4	47.6	48.4	54.1	52.6	55.3	61.7	54.7
Ni	66.9	38.3	38.6	37.5	66.3	25	66.3	117.1	19.9
La	85	97	98	82	101	94	94	75	103
Ce	150	194	198	152	197	175	177	158	190
Pr	15.9	21.8	22.1	16.0	23.1	20.0	19.6	19.6	21.3
Nd	60	86	87	60	81	71	68	75	76
Sm	9.4	13.2	13.5	9.0	11.7	9.7	9.5	11.0	10.4
Eu	2.4	3.3	3.4	2.4	3.1	2.6	2.5	3.0	2.8
Gd	7.0	8.9	9.3	6.5	8.0	6.8	6.6	7.6	7.4
Tb	1.1	1.2	1.3	0.8	0.9	0.8	0.7	0.9	0.9
Dy	4.1	4.5	4.7	3.9	4.6	4.0	3.8	4.7	4.3
Ho	0.8	0.8	0.8	0.8	0.7	0.7	0.7	0.8	0.7
Er	2.0	2.1	2.1	2.1	1.9	1.7	1.8	2.1	1.9
Tm	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3
Yb	2.1	1.9	2.0	1.6	1.7	1.7	1.6	1.9	1.7
Lu	0.4	0.3	0.3	0.2	0.3	0.2	0.3	0.3	0.3
Ba	1102	1176	1211	1152	1181	1180	1363	1093	1705
Hf	4.3	4.7	4.6	4.3	4.5	4.5	4.6	3.7	4.7
Nb				24.5	37.8	28.4	31.1	24.1	31.8
Rb				37.0	34.9	43.0	46.7	31.3	53.3
Sr	1585	2358	2576	1827	2150	2161	2198	2325	2812
Pb	15.7	17.2	23.2						
Ta	1.3	1.7	1.8	0.9	1.4	1.2	1.4	0.9	1.4
Th	8.3	6.1	6.0	5.5	6.6	6.6	7.9	3.6	7.8
U	1.4	1.4	1.3	1.0	1.4	1.2	1.5	0.8	1.9
Zr	192	198	198	201	204	217	244	166	236
Y	23.0	24.3	23.5	23.6	21.7	19.5	20.0	23.4	21.1

(continued)

reference materials were analysed: NIM-G, NIM-L, BHVO-1, RGM-1 and STM-1. The standards analysed as unknowns generally gave trace element concentrations that deviated by <15% from literature values; the results of these measurements are reported in [Supplementary Data A](#).

### Sr–Nd isotopes

Thermal ionization mass spectrometry (TIMS) was used for the analysis of  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$ . Strontium (Sr) and neodymium (Nd) were extracted from unspiked rock powders that were dissolved in an  $\text{HNO}_3$ :HF acid mixture (1:4), followed by conventional ion-exchange chromatographic techniques at the University of Leeds. Sr and Nd isotope ratios were measured on a ThermoScientific Triton multi-collector mass

spectrometer running in static mode. The normalization value for fractionation of  $^{87}\text{Sr}/^{86}\text{Sr}$  was  $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ ; that of  $^{143}\text{Nd}/^{144}\text{Nd}$  was  $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ . Instrument errors for determinations of  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$  are reported as the standard error,  $2\sigma$  (i.e. two standard deviations about the mean using 200–240 measurements). External precision ( $2\sigma$ ) for Sr and Nd isotopic ratios from successive replicate measurements of primary standards was better than 35 ppm for the NIST SRM-987 International Reference Standard ( $^{87}\text{Sr}/^{86}\text{Sr} = 0.710260$  for nine runs averaged, with a standard deviation of  $1.1 \times 10^{-5}$ ) and better than 25 ppm for the La Jolla Nd International Reference Standard ( $^{143}\text{Nd}/^{144}\text{Nd} = 0.511842$  for 11 runs averaged, with a standard deviation of  $2.5 \times 10^{-5}$ ). USGS standard BHVO-1 was also run as a validation material throughout the



Table 1: Continued

Volcanic field:	Syunik	Syunik	Syunik	Syunik	Syunik	Syunik	Syunik	Syunik	Syunik
Sample:	8.07.15	9.01.15	9.02.15	10.01.15	10.02.15	10.03.15	11.02.15	11.3.15	11.04.15
Eruptive type:	Bomb	Plug	Bomb	Bomb	Lava	Scoria	Lava	Scoria	Lava
Latitude:	39-50771	39-53042	39-51419	39-64401	39-64359	39-6413	39-62041	39-63605	39-63671
Longitude:	46-20948	46-22106	46-23795	46-102	46-10184	46-10138	46-02615	46-03841	46-04296
Elevation (m):	2314	2738	2608	2811	2693	2614	2321	2536	2542
Lab. analysis:*	ACME	ACME	ACME	ACME	ACME	ACME	ACME	ACME	ACME
SiO <sub>2</sub>	49.81	51.60	47.68	47.89	51.35	51.44	49.60	47.27	48.67
TiO <sub>2</sub>	1.37	1.22	1.78	1.26	1.48	1.46	1.25	1.80	1.73
Al <sub>2</sub> O <sub>3</sub>	16.80	17.40	15.80	15.58	16.93	16.91	16.87	15.49	15.86
Fe <sub>2</sub> O <sub>3</sub> tot†	8.15	8.20	10.07	8.14	8.96	8.92	8.49	10.07	9.77
MnO	0.13	0.13	0.15	0.14	0.14	0.14	0.14	0.15	0.14
MgO	4.55	4.42	6.28	4.45	4.78	4.79	5.09	5.97	5.65
CaO	8.29	7.51	9.54	9.64	8.23	8.07	8.48	9.87	9.69
Na <sub>2</sub> O	4.62	4.50	4.24	4.47	4.85	4.81	4.87	4.50	4.37
K <sub>2</sub> O	3.50	3.06	2.10	3.87	2.11	2.13	2.75	2.52	2.49
P <sub>2</sub> O <sub>5</sub>	1.17	1.26	1.14	1.54	0.81	0.82	1.27	1.17	1.12
Total	98.91	99.94	99.68	97.74	99.93	99.9	99.19	99.04	100.06
LOI	1.07	0.49	0.77	0.55	0.18	0.29	0.22	0.08	0.44
Mg#	52.8	51.6	55.3	52.0	51.4	51.5	54.3	54.0	53.4
Ni	29.6	41.6	60.8	12.1	34	35.8	49	41.7	35
La	98	104	75	102	61	64	95	88	86
Ce	193	189	151	195	115	114	175	187	175
Pr	22.7	21.2	18.0	22.9	12.9	13.0	20.3	22.3	20.9
Nd	82	72	68	81	48	47	71	82	77
Sm	11.7	10.1	10.4	11.6	8.0	7.6	10.0	11.7	10.7
Eu	3.0	2.7	2.9	3.1	2.3	2.4	2.7	3.1	2.9
Gd	7.6	7.7	8.1	8.6	6.9	7.1	7.1	8.6	7.8
Tb	0.9	0.9	0.9	0.9	0.8	0.9	0.8	0.9	0.9
Dy	4.6	4.8	4.9	4.8	5.0	4.9	4.4	4.8	4.6
Ho	0.7	0.7	0.8	0.8	0.9	0.9	0.7	0.7	0.8
Er	1.9	2.1	2.0	2.1	2.6	2.5	1.8	2.0	1.9
Tm	0.3	0.3	0.3	0.3	0.4	0.4	0.3	0.3	0.3
Yb	1.7	1.8	2.0	2.0	2.4	2.3	1.7	1.7	1.7
Lu	0.2	0.3	0.3	0.3	0.4	0.4	0.2	0.2	0.3
Ba	1446	1201	1022	1795	924	915	1301	1151	1105
Hf	4.6	4.3	4.0	4.5	5.1	4.8	3.9	3.9	4.0
Nb	28.8	31.3	21.9	28.7	19.4	20.5	23.2	33.4	32.6
Rb	44.2	41.5	26.9	51.2	34.0	33.7	33.2	33.8	34.1
Sr	2504	2032	2128	2889	1413	1363	2185	2348	2131
Pb									
Ta	1.2	1.2	0.9	1.1	0.8	0.8	1.0	1.3	1.2
Th	5.1	6.0	3.2	7.1	4.8	4.9	5.5	5.5	5.8
U	1.1	1.1	1.0	1.1	1.0	1.1	1.0	1.3	1.2
Zr	210	218	177	202	240	235	169	190	186
Y	20.3	22.3	24.0	25.0	26.6	26.3	21.6	22.0	22.3

\*Laboratory where samples were analysed: ACME, ACME labs, Bureau Veritas minerals, Vancouver, Canada; RH, Royal Holloway University, UK.

†Royal Holloway major element data report Fe as FeO\*, which was recalculated to Fe<sub>2</sub>O<sub>3</sub> (tot) by dividing by a factor of 0.89981. Totals are the original totals prior to Fe<sub>2</sub>O<sub>3</sub> (tot) recalculation.

run period for both <sup>87</sup>Sr/<sup>86</sup>Sr and <sup>143</sup>Nd/<sup>144</sup>Nd. Once corrected to an NIST SRM 987 preferred value of 0.710246 using the accompanying primary standard measurement, all measurements of <sup>87</sup>Sr/<sup>86</sup>Sr for BHVO-1 reproduce a literature value of 0.703475 to within 2σ (Weis *et al.*, 2006). All measured <sup>87</sup>Sr/<sup>86</sup>Sr ratios were corrected to the literature NIST SRM 987 only. For Nd isotopes, all samples were corrected to the literature value of BHVO-1 (0.512986) giving an average correction of 0.00002. The total chemistry blanks for Sr and Nd were negligible during the period of measurements (0.4 ng for Sr and 0.3 ng for Nd).

## PETROGRAPHY

Southern Lesser Caucasus volcanic rocks are generally very fresh, with generally low LOI values (usually <1%) for the majority of mafic samples (Table 1). The groundmass can sometimes contain minor amounts of clay minerals, but in many cases it is very fresh, with small pockets of volcanic glass present in a minority of samples (Table 2).

Tephrites, trachybasalts and trachybasaltic andesites from Syunik always contain phenocrysts (>0.3 mm) of clinopyroxene, and may have olivine, amphibole and apatite phenocrysts. Clinopyroxene is

**Table 2:** Phenocryst and groundmass minerals for volcanic rocks from the southern Lesser Caucasus

Sample	Field	Volcano/place	Eruptive type	Rock name	Mg#	Phenocrysts	Groundmass
6.27.08	V	Khrbekner	Lava	TBA	56.2	ol, cpx, plag	plag, cpx, ox
4.1.12	V	Tsovak pyroclastic flow	Pumice	R	17.3	kspar, plag, qtz	glass
5.1.13	V	Smbatar	Lava	TA	50.9	cpx, plag, amph	plag, ox, cpx
5.1.15	V	Geghagar tuff	Igimbrite	R	44.5	kspar, plag, phlog, qtz	glass
5.3.15	V	Porak	Lava	TA	53.9	cpx, plag	plag, cpx, ox
5.4.15	V	Zhiligyol	Lava	TA	51.1	cpx, qtz (xen)	plag, glass, ox, cpx
5.5.15	V	Vent W of Zhiligyol	Plug	R	17.6	kspar, plag, bio, qtz	altered
5.9.15	V	Lake Al-Lich	Plug	Tr	29.5	plag, phlog, kspar, cpx	kspar, qtz, ox, cpx
6.3.15	V	Torgomayr	Lava	TBA	51.4	cpx	plag, bio, ox
7.2.15	V	Dome E of Trdatanist	Plug	R	25.4	kspar, qtz, plag, bio	altered
7.4.15	V	S slopes of V upland	Lava	TA	54.8	cpx, amph (pseud), plag	plag, cpx, ox, glass
7.5.15	V	S slopes of V upland	Lava	TBA	54.0	cpx, plag, amph	plag, cpx, ox, ol
1.4C.08	S	Garusar	Scoria	TBA	50.8	cpx, amph	plag, cpx, ox
2.5.08	S	Shinuayr	Bomb	TBA	50.5	cpx, amph, ol	glass, plag, ox, cpx
2.6.08	S	Morutumb	Bomb	TBA	48.8	cpx, amph, ol, ap	plag, glass, cpx, ox
2.7.08	S	Yerakov Blur	Bomb	PTe	48.6	amph, cpx	glass, plag, ox, cpx
2.9.08	S	Pokr Chobanasar	Bomb	TBA	52.5	cpx, amph, ol	plag, cpx, ox
2.10.08	S	Barurtumb	Lava	B	66.8	ol, cpx	plag, cpx, ox
5.21.08	S	Berd	Lava	TBA	52.7	ol, cpx	plag, ox
6.24.08	S	Tekblur	Scoria	PTe	48.4	cpx	glass, plag
3.10.10	S	Unit 7	Lava	Tr	42.5	plag, cpx, amph (pseud)	plag, ox, cpx
3.11.10	S	Unit 7	Lava	TA	46.5	cpx, amph (pseud)	plag, ox, cpx
4.19.10	S	Ishkhanasar	Lava	TA	44.5	plag, amph	plag
5.5.12	S	Tshkhouk	Lava	TA	50.0	cpx, plag, ox	plag, ox, cpx
8.2.15	S	Khozazblur	Bomb	TB	54.1	cpx, ol	plag, cpx, ox
8.3.15	S	Verjiblar	Bomb	TBA	52.6	cpx, amph	plag, cpx, ap
8.7.15	S	Marakhasar	Scoria	PTe	52.8	amph, cpx	glass, plag
9.1.15	S	Spiovblur	Plug	TBA	51.6	cpx, amph, plag, ol, ap	plag, cpx, ox
9.2.15	S	Chobanasar	Bomb	TB	55.3	cpx, ol, amph (pseud)	plag, cpx, ox
10.2.15	S	Kyorpasar	Lava	TBA	51.4	aphyric	plag, cpx, ol, ox
10.3.15	S	Mets Yerkvoryak	Scoria	TBA	51.5	plag, ol, cpx, ox	glass
10.6.15	S	Mets Yerkoryak	Scoria	TBA	51.6	cpx	glass, plag, cpx, ox
11.1.15	S	Quarry, Shaqi village	Lava	TBA	46.3	plag, cpx	plag, cpx, ox
11.3.15	S	Sherepasar	Scoria	Te	54.0	cpx, ol, amph	plag, glass, cpx, ox

Volcanic field abbreviations: V, Vardenis; S, Syunik. Mineral abbreviations: ol, olivine; cpx, clinopyroxene; plag, plagioclase; ox, Fe–Ti oxide; kspar, potassium feldspar; qtz, quartz; amph, amphibole; bio, biotite; ap, apatite. Mineral texture abbreviations: xen, xenocrystic; pseud, pseudomorph. Rock names: B, basalt; R, rhyolite; TB, trachybasalt; TBA, trachybasaltic andesite; TA, trachyandesite; Tr, trachyte; Te, tephrite–basanite; PTe, phonotephrite.

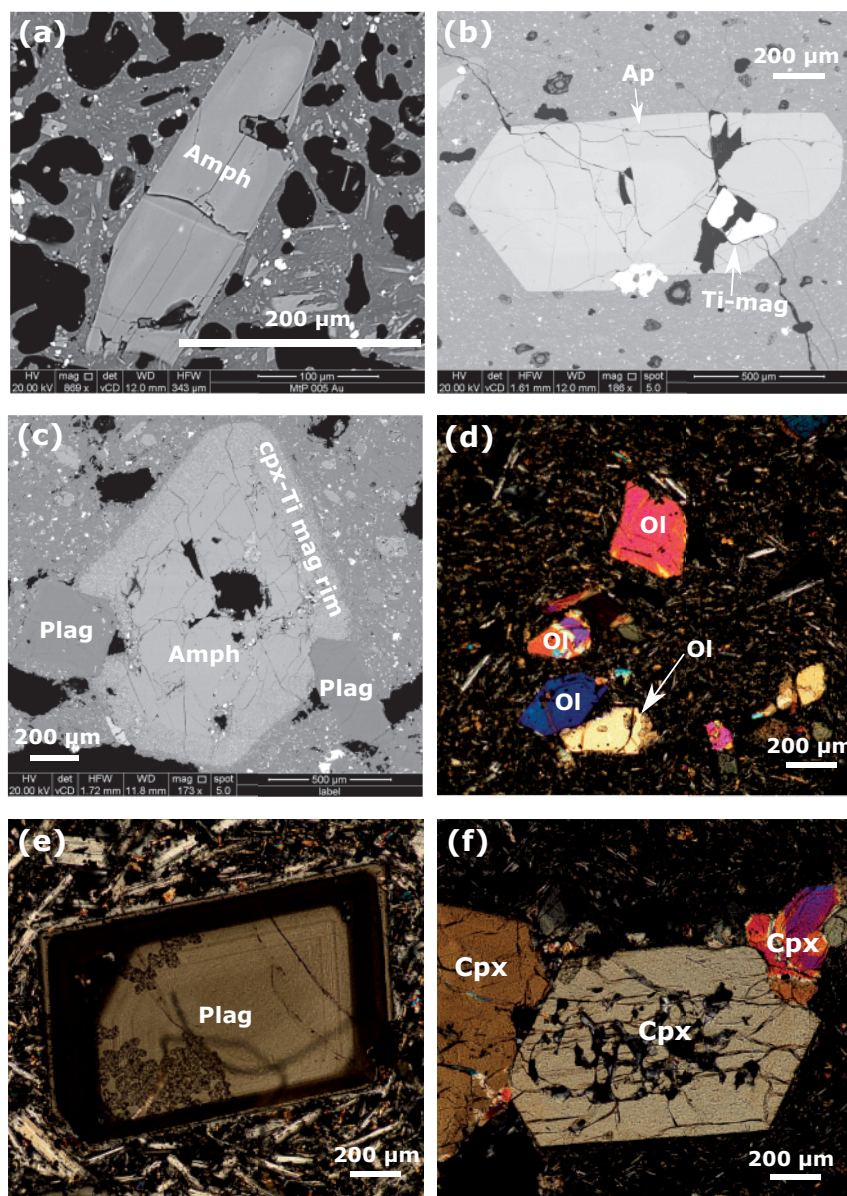
ubiquitous and normally occurs as glomerocrysts (Fig. 5f). Some samples have abundant olivine microphenocrysts (Fig. 5d) and phenocrysts (up to 600  $\mu\text{m}$ ), although olivine is generally restricted to the most mafic samples (low-silica trachybasaltic andesite or more mafic). Amphibole phenocrysts are common (up to 2 cm), and occur across the compositional range (Fig. 5a). In several samples amphibole is absent, or partially resorbed (Fig. 5c), owing to disequilibrium conditions prior to eruption. The apatite phenocrysts can reach up to 2 mm in size in some cases (Fig. 5b). The common occurrence of hydrous minerals, such as amphibole and apatite, is noteworthy in comparison with the volcanic rocks from the northern and central Lesser Caucasus, where such mineral phases are less common, certainly in the most mafic samples (Connor *et al.*, 2011; Neill *et al.*, 2013). Typical groundmass in mafic volcanic rocks from Syunik is composed of plagioclase, clinopyroxene, and Fe–Ti oxides, with volcanic glass in some scoria samples. It should be noted that plagioclase is present only in the groundmass and is not a phenocryst phase.

More evolved high-silica trachybasaltic andesites, trachyandesites and trachytes from Syunik show the

appearance of abundant plagioclase phenocrysts. In some cases, these plagioclase crystals (up to 2 mm) show evidence of multiple stages of crystallization with distinct core to rim zonation (Fig. 5e). Amphibole also becomes a more common phenocryst phase, whereas clinopyroxene is less dominant. The groundmass is increasingly dominated by plagioclase. Rhyolites are common in the northern part of Syunik, where they commonly form obsidian flows.

In Vardenis all samples are trachybasaltic andesites or more evolved, and plagioclase is always a phenocryst phase. In one trachybasaltic andesite (Table 2, sample 6.3.15), biotite is also a phenocryst phase. Rhyolites from Vardenis commonly have biotite, plagioclase, quartz and potassium feldspar as phenocrysts, and a groundmass composed of quartz and potassium feldspar.

From the occurrence of different minerals in the range of rock types we sampled across the Syunik and Vardenis volcanic highlands (34 samples; Table 2), it is possible to suggest a general order of crystallization for both areas, which would probably have been: olivine + Fe–Ti oxides  $\pm$  apatite; clinopyroxene + Fe–Ti oxides  $\pm$  olivine  $\pm$  amphibole; clinopyroxene +



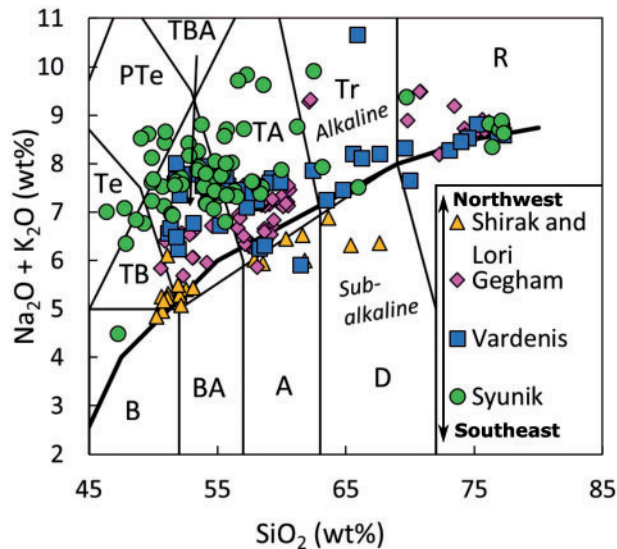
**Fig. 5.** Representative back-scattered electron images (a–c) and photomicrographs (d–f) of collision-related volcanic rocks in the southern Lesser Caucasus. (a) Unaltered amphibole (Amph) phenocryst in trachybasaltic andesite scoria from Syunik. Groundmass composed of clinopyroxene, plagioclase and oxides. (b) Large apatite (Ap) phenocryst > 1 mm in diameter, with oxide inclusions in trachybasaltic andesite lava, Syunik. (c) Large amphibole phenocryst with overgrowing plagioclase (Plag) grains in trachybasaltic andesite lava from Vardenis. The rim of the amphibole is broken down into oxides and clinopyroxene. (d) Crossed polars (XPL) image of olivine (Ol) phenocrysts in trachybasalt lava from Syunik. (e) XPL image of plagioclase phenocryst in trachybasaltic andesite lava with sharp compositional boundary between the core and rim (in extinction) of this crystal. (f) XPL image of clinopyroxene (Cpx) glomerocryst in trachyandesite sample from Vardenis.

plagioclase + Fe–Ti oxides ± amphibole ± phlogopite; plagioclase + potassium feldspar + Fe–Ti oxides ± phlogopite. The predominant focus of this study is on the mechanism of mantle melting, which requires the effect of fractional crystallization to be minimized. Only primitive samples with <54 wt % SiO<sub>2</sub> and >4 wt % MgO are used to investigate questions of magma petrogenesis. These samples will probably all have only fractionated mafic minerals, with no significant feldspar fractionation.

## RESULTS

### Major element characteristics

Across the Lesser Caucasus there is a great diversity in the compositions of the volcanic rocks within each volcanic highland, with most showing a complete compositional range from basalt to rhyolite (Fig. 6). Rhyolites are also present in the northern Lesser Caucasus (Karapetian *et al.*, 2001), although they were not sampled for the Neill *et al.* (2013) study. Southern Lesser Caucasus samples, with Syunik being the most



**Fig. 6.** Total alkalis vs silica classification diagram for volcanic rocks from the Lesser Caucasus mountain chain. Classification boundaries are from Cox *et al.* (1979), Le Bas *et al.* (1986) and Le Maitre *et al.* (1989). Data for Syunik and Vardenis are from this study, whereas Shirak and Lori data are from Neill *et al.* (2013, 2015), in this and all subsequent figures. Gegham data are from I. P. Savov *et al.* (unpublished data). Abbreviations for classification fields are as follows: B, basalt; BA, basaltic andesite; A, andesite; D, dacite; R, rhyolite; TB, trachybasalt; TBA, trachybasaltic andesite; TA, trachyandesite; Tr, trachyte; Te, tephrite/basanite; PTe, phonotephrite.

extreme, have more alkaline compositions compared with those from the north, with the largest  $\text{Na}_2\text{O}$  and  $\text{K}_2\text{O}$  variations between north and south in the most mafic samples (Fig. 6). Southern Lesser Caucasus samples are also more potassic when compared with their northern counterparts (shoshonitic versus calc-alkaline series; Fig. 7a). As well as being more alkaline, mafic southern Lesser Caucasus samples extend to lower  $\text{SiO}_2$  contents.

On some major element variation diagrams, such as  $\text{MgO}$  versus  $\text{SiO}_2$  (Fig. 7b), the trends are fairly similar for northern and southern Lesser Caucasus samples. However, other elements, notably P, display a significant geographical gradient in concentration for the most mafic samples, from 0.4 wt %  $\text{P}_2\text{O}_5$  in the north to as high as 1.6 wt % in the south (Fig. 7c).

### Trace element characteristics

Mid-ocean ridge basalt (MORB)-normalized trace element patterns all show profiles typical of subduction-related volcanic rocks, with positive anomalies for Ba, K, Pb and Sr and negative anomalies for the high field strength elements (HFSE) Nb, Ta and Ti (Fig. 8). Superimposed on this is an additional enrichment in incompatible trace elements, in particular the light rare earth elements (LREE; Fig. 9), Sr, Ba and P, which becomes increasingly pronounced to the south of the transect. Owing to the LREE enrichment, as well as a more moderate depletion in heavy rare earth elements (HREE), REE profiles become increasingly steep

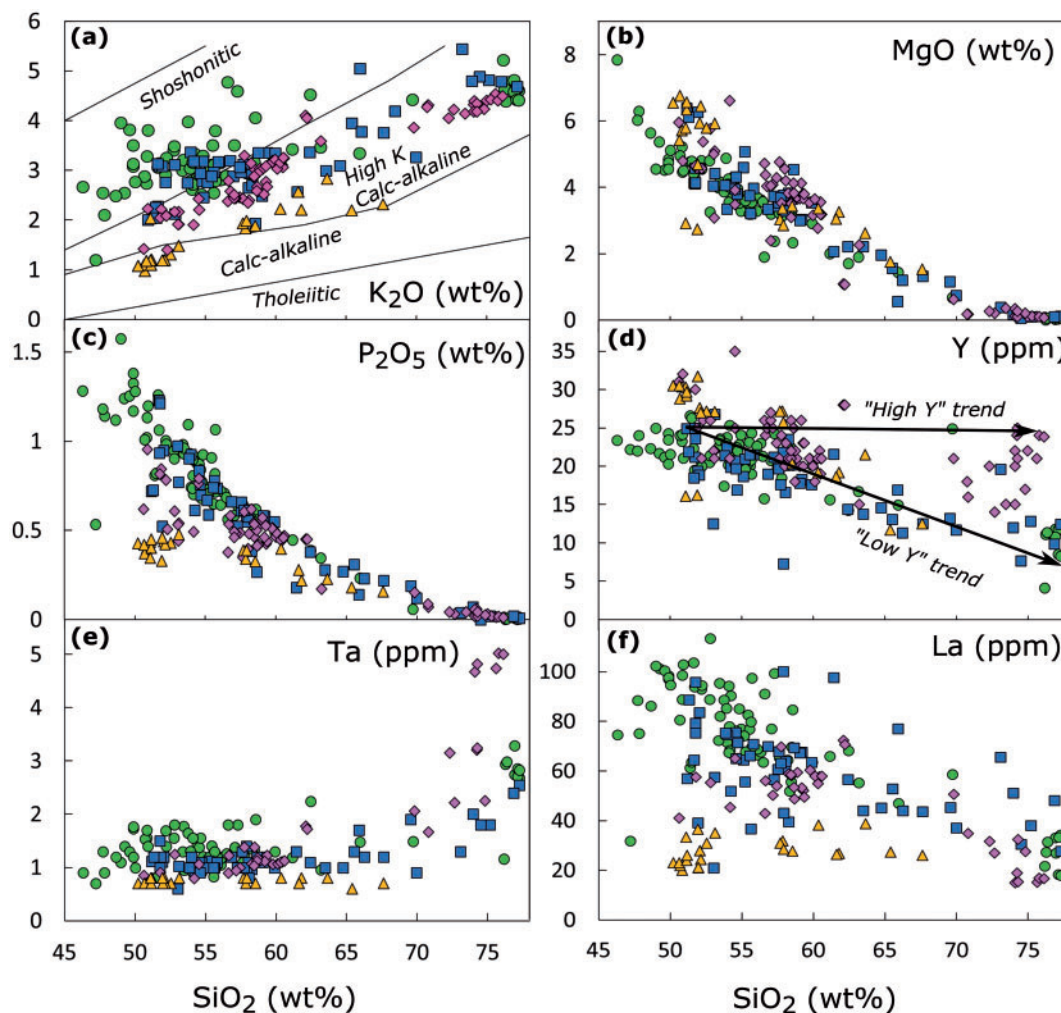
towards the south (Fig. 9), with Cl-normalized La/Yb ratios of five in the northern Lesser Caucasus compared with values as high as 40 in the south.

These additional enrichments do not resemble those of typical intra-plate ocean island basalts (OIB), which show uniform enrichments in all the most incompatible trace elements, rather than the larger enrichments seen in the LREE relative to the HFSE (Fig. 8e), as exemplified by Ta and La in Fig. 7e and f, respectively. These additional enrichments are seen most clearly in volcanic rocks from NW Iran (Fig. 8e), a region of very thick lithosphere—up to 220 km (Priestley *et al.*, 2012). In Fig. 8a–d we show that the variations in the composition of basalts within individual volcanic highlands are small, relative to the variations in composition between the highlands. Both the ubiquitous subduction ('arc') signatures and additional enrichment are highlighted in Fig. 10. All samples plot above the mantle array, typical for rocks from volcanic arcs. However, the samples from the southern Lesser Caucasus have higher Th/Yb and Ta/Yb relative to their northern counterparts (Fig. 10). Volcanic rocks from the thick lithosphere Zagros Core region of NW Iran again plot as a compositional end-member.

Given that amphiboles are relatively common in the southern Lesser Caucasus rocks, it should be noted that low Y (amphibole fractionation) trends dominate the Y versus  $\text{SiO}_2$  plot for these rocks (Fig. 7d). Samples from further north, in which anhydrous mineral assemblages are more common, show both high Y (anhydrous assemblage) and low Y (hydrous) fractionation trends.

### Sr–Nd isotope systematics

The Sr–Nd isotope compositions of the southern Lesser Caucasus samples are shown in Table 3.  $^{87}\text{Sr}/^{86}\text{Sr}$  varies from 0.7043 to 0.7047, whereas  $\epsilon\text{Nd}$  varies from +2.2 to +4.2. All mafic volcanic rocks from the Lesser Caucasus plot along the mantle array, with compositions more enriched than normal (N)-MORB but more depleted than Bulk Earth (Fig. 11). In general, those samples from the SE have higher (more radiogenic)  $^{87}\text{Sr}/^{86}\text{Sr}$ , and less radiogenic  $\epsilon\text{Nd}$  than those samples from the NW, defining a steep gradient on the Sr–Nd isotope diagram (Fig. 11). A few samples plot away from this trend to higher  $^{87}\text{Sr}/^{86}\text{Sr}$  (see Crustal contamination section of Discussion). Lesser Caucasus volcanic rocks are isotopically more depleted than volcanic rocks formed above very thick lithosphere in NW Iran (>200 km; Priestley *et al.*, 2012). They are also more depleted than volcanic rocks from the Lake Van area, which commonly display significantly more variable Sr–Nd isotope compositions (not shown), probably because those magmas interacted extensively with continental crust during ascent (Pearce *et al.*, 1990; Şen *et al.*, 2004; Özdemir *et al.*, 2006; Oyan *et al.*, 2017).



**Fig. 7.** Selected major and trace elements plotted versus  $\text{SiO}_2$ . (a)  $\text{K}_2\text{O}$ ; (b)  $\text{MgO}$ ; (c)  $\text{P}_2\text{O}_5$ ; (d) Y; (e) Ta; (f) La. All major elements are recalculated to 100% on an anhydrous basis. Symbols and data sources are as in Fig. 6.

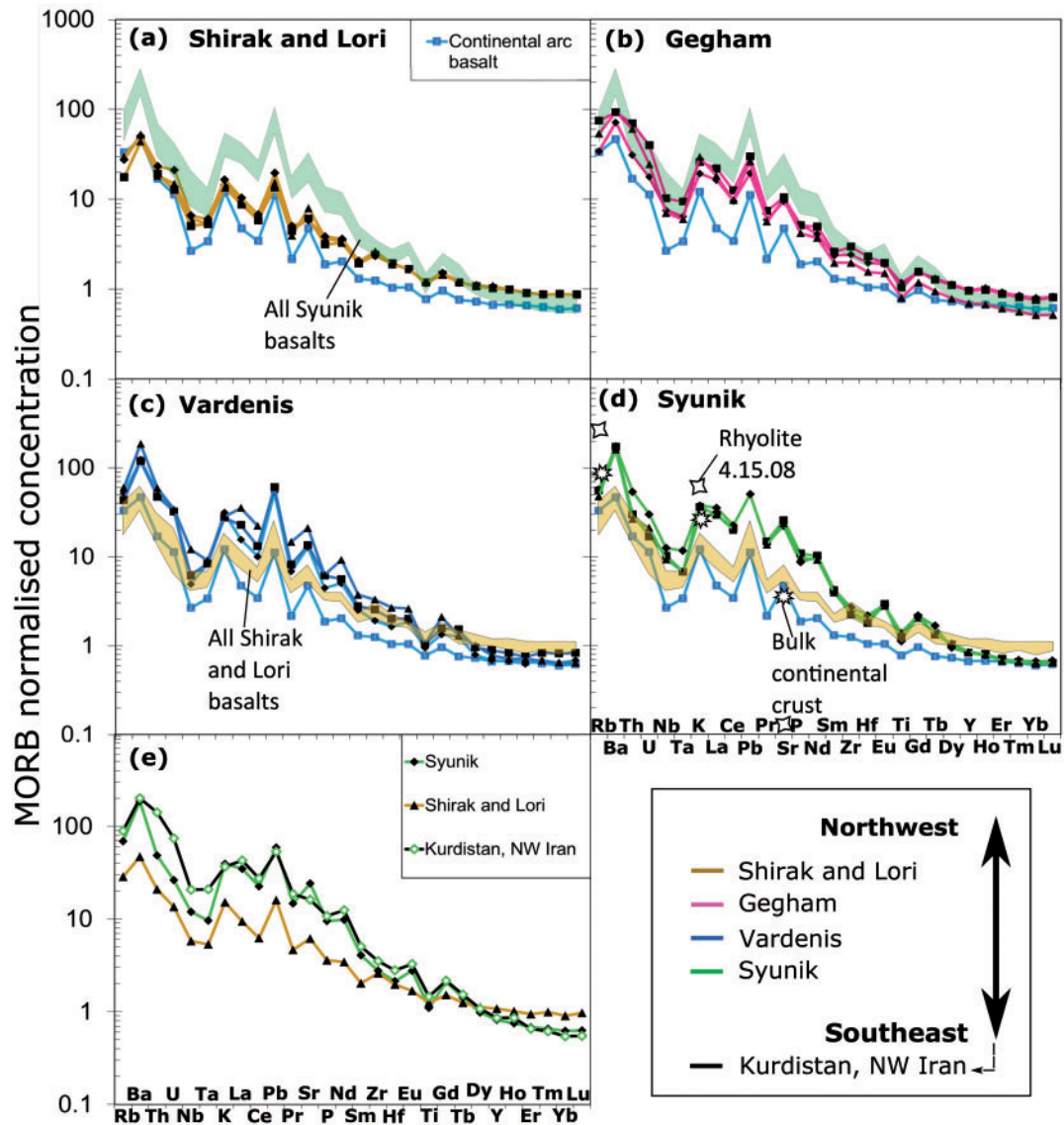
## DISCUSSION

### Crustal contamination

Volcanic rocks produced within the thicker lithosphere of the southern Lesser Caucasus have higher concentrations of incompatible trace elements, more radiogenic  $^{87}\text{Sr}/^{86}\text{Sr}$ , and lower  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios. All of these features could be produced by crustal contamination during magmatic evolution, as has been suggested for various parts of Eastern Anatolia, where assimilation of radiogenic ancient continental crust gives  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios of up to 0.7065 and marked variation of these ratios with  $\text{SiO}_2$  (Pearce *et al.*, 1990; Keskin *et al.*, 2006). All Sr–Nd isotope ratios for the Lesser Caucasus volcanic rocks show small variations compared with what would be expected if the magmas had been contaminated by continental crust. The crust in the Lesser Caucasus is composed of a mixture of felsic metamorphic basement (South Armenian Block), arc volcanic rocks and Mesozoic–Cenozoic sediments (Fig. 2). Only the arc-related volcanic rocks would have similar isotope compositions to the collision-related magmas, such that assimilation of any of the other lithologies

would alter, often dramatically, the isotope composition of the host magma. The majority of samples show insignificant variability in  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$  with  $\text{SiO}_2$  within individual volcanic highlands (Fig. 12), suggesting that isotope ratios are not being modified significantly during magma evolution and storage in the crust prior to eruption. The lack of isotopic variability with  $\text{SiO}_2$  content suggests that assimilation of South Armenian Block crust (Fig. 2), which with a basement  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio of  $\sim 0.7303$  (Baghdasaryan & Ghukasyan, 1985) should be easily identified, is unlikely. Similarly, assimilation of sedimentary material is also unlikely, given typical Tethyan (Mesozoic) flysch  $^{87}\text{Sr}/^{86}\text{Sr}$  of 0.7112 (Prelević *et al.*, 2008).

One sample from the Syunik volcanic highland (1-4A-08), which was sampled from a scoria cone containing large (centimetre-sized) felsic xenoliths entrained within the trachybasaltic andesite scoria, has trace element and isotopic characteristics indistinguishable from the other mafic samples. The Sr–Nd isotope composition of one of the xenoliths (sample 1-4B-08; Table 3) is also shown in Fig. 12, and is only slightly

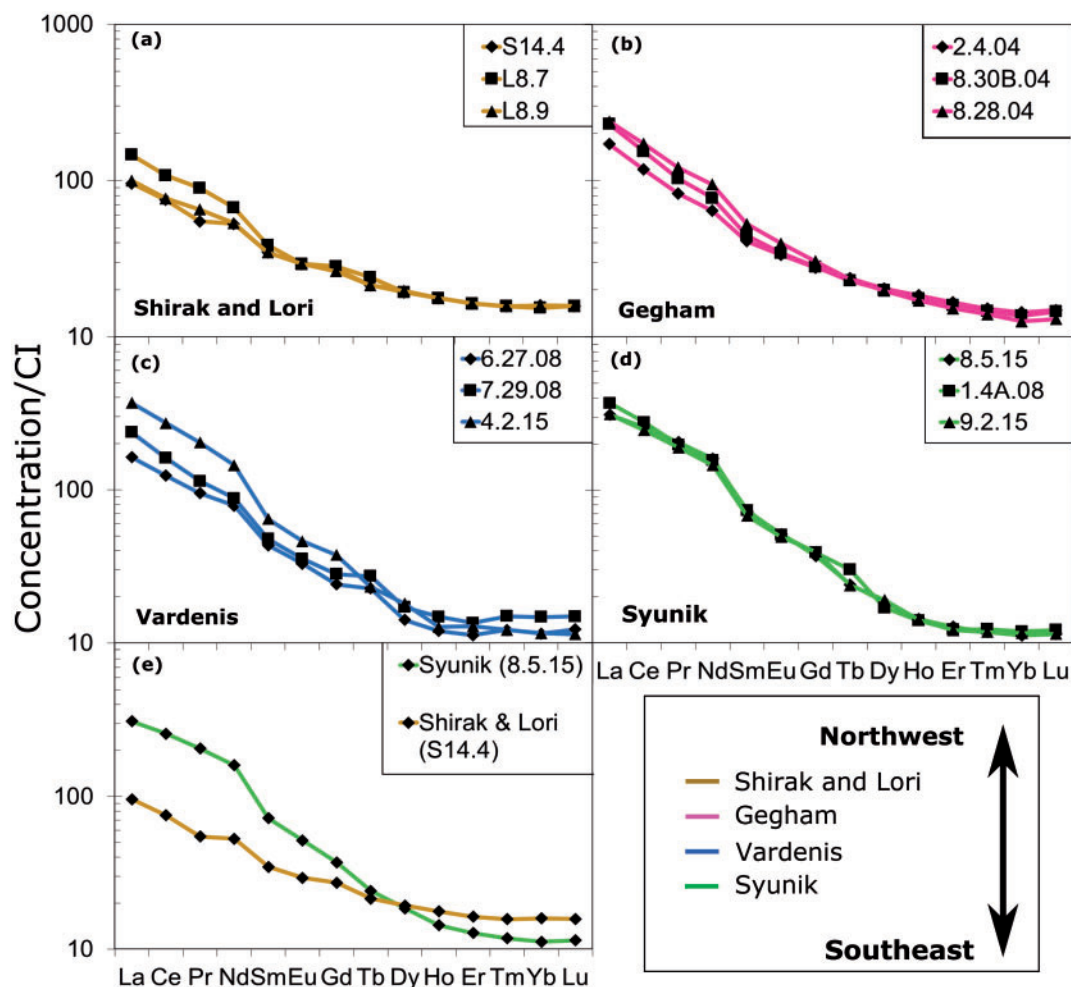


**Fig. 8.** MORB-normalized trace element patterns of mafic end-member samples for the Shirak and Lori (a), Gegham (b), Vardenis (c), and Syunik (d) volcanic highlands, which correspond to regions 1–4 in Fig. 1, respectively. The selected samples are the three samples with the highest MgO content (wt %) in each volcanic highland. Data sources are as in Fig. 6. For comparison, the total range of ‘basalts’ (<52% SiO<sub>2</sub>) from the Syunik volcanic highland is shown in (a) and (b) as the pale green field. The total range of Shirak and Lori basalts is shown as a gold field in (c) and (d). (d) also shows the composition of a Syunik rhyolite (diamonds) and bulk continental crust (stars) for Rb, K and Sr, to demonstrate whether these rhyolites are likely to be formed by assimilation and/or melting of continental crust. Composition of the continental crust is from Rudnick & Gao (2003). The total compositional variability is illustrated in (e), where averages from the geographical extremes (Shirak in the north and Syunik in the south) are compared. Also shown here is a sample from Kurdistan, NW Iran from Allen *et al.* (2013), formed from melting in a region of very thick lithosphere. Normalization factors are from Sun & McDonough (1989). The average composition of continental arc basalt is from Kelemen *et al.* (2003).

elevated in  $^{87}\text{Sr}/^{86}\text{Sr}$  (0.7049) above that of the Quaternary basalts, suggesting that an unrealistically high degree of assimilation would have to occur for the composition of the magma to be significantly affected. The  $^{143}\text{Nd}/^{144}\text{Nd}$  (0.5128) is indistinguishable from Syunik basalts. It is likely that the felsic xenolith is cogenetic with the trachybasaltic andesite scoria host. It provides evidence that basaltic magmas in the southern Lesser Caucasus are interacting with felsic igneous rocks in the crust, which have a similar origin, rather

than interacting with ancient crust, which would have strongly affected the Sr–Nd isotope ratios.

The curved patterns for major element variations versus SiO<sub>2</sub>, including MgO (Fig. 7b), P<sub>2</sub>O<sub>5</sub> (Fig. 7c) and Al<sub>2</sub>O<sub>3</sub> (not shown), suggest that the rhyolites are derived from extreme degrees of magmatic differentiation, rather than being crustal melts. These rhyolites also do not appear to be mixtures between primitive magmas and continental crust (Rudnick & Gao, 2003). The Rb and Sr contents of a rhyolite from Syunik are



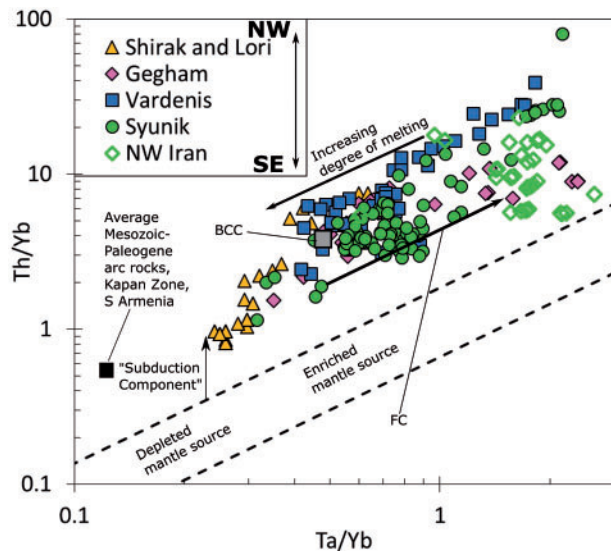
**Fig. 9.** Chondrite-normalized REE profiles for Plio-Pleistocene mafic volcanic rocks from the Lesser Caucasus for Shirak and Lori (a), Gegham (b), Vardenis (c) and Syunik (d) volcanic highlands. The total compositional variability is illustrated in (e), where samples from Syunik and Shirak are compared. Normalization factors are from Sun & McDonough (1989). Data sources are as in Fig. 6.

actually more extreme relative to the basalts than average continental crust (Rb more strongly enriched, Sr more strongly depleted in the rhyolite; Fig. 8d).  $K_2O$  is more enriched in the rhyolite than in the basalts, whereas continental crust has lower  $K_2O$  than continental crust (Fig. 8d). The lack of evidence for extensive assimilation of continental crust in the petrogenesis of the rhyolites suggests that these processes are likely to be unimportant in the petrogenesis of the more primitive magmas.

It is also possible that the magmas may have been contaminated by Mesozoic–Paleogene arc crust (Fig. 2) on their ascent, as has been suggested for the northern Lesser Caucasus in some isolated cases (Neill *et al.*, 2015). The similar Sr–Nd isotope compositions of these arc rocks (Mederer *et al.*, 2013) to the collision-related magmas mean that assimilation could be ‘cryptic’, without obvious modification of Sr–Nd isotope ratios. Given the arc-like incompatible trace element geochemistry of Lesser Caucasus volcanic rocks, arc–crust assimilation is unlikely to significantly alter trace element compositions, although it is possible that such assimilation

could explain some of the spread in the southern Lesser Caucasus data. The average composition of Mesozoic arc rocks from the southern Lesser Caucasus (Kapan Zone) is shown in Fig. 10 (Mederer *et al.*, 2013). The Th/Yb ratio of the arc rocks is elevated above the mantle array similar to the post-collisional Lesser Caucasus samples. However, the low Ta/Yb ratios of these arc rocks make them unlikely candidates to explain the more enriched composition of southern Lesser Caucasus magmas.

There are a few examples of elevated  $^{87}Sr/^{86}Sr$  ratios in evolved samples above the background range. In most cases, it seems that the Nd isotopes are unaffected. Some ancient arc samples from the Kapan zone, southern Lesser Caucasus, have very high initial  $^{87}Sr/^{86}Sr$ , but  $^{143}Nd/^{144}Nd$  does not vary significantly in the same samples (Mederer *et al.*, 2013). Extensive assimilation of such material would be required to explain the elevated  $^{87}Sr/^{86}Sr$  of these samples. Worthy of note, however, are the very high Rb/Sr ratios (7–23) in these rhyolites—sometimes 1000× greater than those of typical basalts (Table 1). Whereas decay of  $^{87}Rb$  over the



**Fig. 10.** Th/Yb vs Ta/Yb variations in the volcanic rocks of the Lesser Caucasus. Mantle source array and enrichment vectors are after Pearce (1983). Lesser Caucasus data sources are as in Fig. 6. NW Iran data are from Allen *et al.* (2013). Fractional crystallization (FC) vector: from basalt to andesite uses a 50% amphibole + 50% clinopyroxene assemblage and basaltic melt partition coefficients; from andesite to rhyolite uses a 50% plagioclase + 50% amphibole assemblage and andesite melt partition coefficients. The vector has a starting composition as the most mafic sample. Partition coefficients for Ta are: clinopyroxene–basalt = 0.017; amphibole–basalt = 0.05; amphibole–andesite = 0.21; plagioclase–andesite = 0.03. For Th: clinopyroxene–basalt = 0.007; amphibole–basalt = 0.05; amphibole–andesite = 0.16; plagioclase–andesite = 0.01. For Yb: clinopyroxene–basalt = 0.28; amphibole–basalt = 0.59; amphibole–andesite = 1.25; plagioclase–andesite = 1.25. All partition coefficients are from the GERM database ([earthref.org/GERM/](http://earthref.org/GERM/)). The average compositions of Mesozoic arc rocks from the Kapan zone arc rocks are from Mederer *et al.* (2013). The grey box labelled BCC is the composition of bulk continental crust from Rudnick & Gao (2003).

relatively short timeframe (<2 Myr) since cooling and crystallization of lavas will have negligible effects on the  $^{87}\text{Sr}/^{86}\text{Sr}$  of basaltic rocks, such high Rb/Sr ratios mean that post-crystallization decay will have a significant effect for rhyolites (Fig. 12a). Indeed, minimum ages for the rhyolites of between 0.6 and 1.4 Ma are sufficient to give initial  $^{87}\text{Sr}/^{86}\text{Sr}$  within the range observed for basalts. As such, even in these rhyolites, crustal contamination may have played an unimportant role in magma evolution. It should be noted that the lower Rb/Sr ratio of the felsic xenolith (0.12) means that its elevated  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio was probably present during emplacement of the host Quaternary mafic magma.

In summary, assimilation of crustal rocks with markedly distinct geochemistry is unlikely, and although assimilation of igneous rocks with similar petrogenetic origins cannot be precluded, such processes would be incapable of driving the enrichment in incompatible trace element concentrations and Sr–Nd isotope ratios observed from north to south in the volcanic rocks of the Lesser Caucasus. The lack of evidence for crustal contamination, along with the gradient in Sr–Nd isotope

ratios in Fig. 11, requires that there must be variation in the composition of the mantle source.

### Subduction modification of the mantle source in the Lesser Caucasus

All mafic samples show distinct negative Nb–Ta and Ti anomalies (Fig. 8), and positive spikes in Ba, K, Pb and Sr in normalized trace element patterns, such that the overall patterns, if not the absolute concentrations, are typical of arc magmas. In the absence of evidence for widespread crustal contamination, it is likely that these features are inherited from the mantle source. Collision between Arabia and Eurasia was preceded by subduction of various Tethyan ocean basins along several convergent margins (Fig. 3). These subducting slabs would probably have contributed slab material to the mantle below the collision zone.

Following the approaches of Hofmann (2003) and Turner *et al.* (2017), Fig. 13 illustrates how the ratios Ba/La, Ce/Pb, Sr/Nd and Nb/U in mafic samples (<54 wt %  $\text{SiO}_2$ , >4 wt % MgO) vary with latitude. These ratios show little variation within all MORB and OIB, but are much more variable in volcanic arc rocks owing to the subducted slab contribution, such that they can be used here as proxies. In all of the plots in Fig. 13 these ratios deviate from those of MORB or OIB, demonstrating the presence of a slab component in their mantle source. These ratios do not show consistent variability between the volcanic highlands (Fig. 13), although Sr/Nd ratios are slightly elevated to the south (Fig. 13c) and the Nb/U ratio reaches a minimum (highest slab contribution) in the central Lesser Caucasus. The lack of a consistent trend in these ratios between volcanic highlands means that the slab contribution is likely to be fairly uniform.

Although mixing between a depleted mantle source and the likely composition of subducted sediment (e.g. Tethyan flysch; Prelević *et al.*, 2008) comes close to explaining the Sr and Nd isotopic composition of the least enriched samples in Fig. 11, such mixing is unable to explain the trend towards the higher  $^{87}\text{Sr}/^{86}\text{Sr}$  and lower  $^{143}\text{Nd}/^{144}\text{Nd}$  seen in the SE. This enrichment in the southern Lesser Caucasus must therefore be driven by some other process. Pre-collision subduction events have probably imparted a subduction signature on the mantle source across the Lesser Caucasus. However, the geochemical gradient between volcanic rocks of the northern and southern Lesser Caucasus cannot be explained by differences in the composition or size of the slab contribution to the mantle source.

### Modelling the conditions of mantle melting

It is possible to investigate how a thicker lithosphere affects the conditions of melting by using the approach of Shaw (2005) to forward model the composition of the samples using a non-modal batch melting model (Table 4), with the equation



**Table 3:** Sr–Nd isotope compositions for Pleistocene lavas from the southern LC

	Rock type	$^{87}\text{Sr}/^{86}\text{Sr}$	$\pm 2\sigma (\times 10^{-6})$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\pm 2\sigma (\times 10^{-6})$	$\epsilon\text{Nd}$
<i>Vardenis</i>						
4.2.15*	TBA	0.704356	10	0.512802	12	3.20
5.3.15*		0.704461	8	0.512803	10	3.22
5.6.15	TBA	0.704505	12	0.512789	10	3.24
5.9.15	Tr	0.704506	12	0.512805	10	3.26
6.3.15		0.704429	8	0.512796	10	3.08
1.1.13*	TBA	0.704432	10	0.512808	10	3.32
1.2.13*	TBA	0.704420	8	0.512781	6	2.79
3.3.13*	TBA	0.704399	12	0.512798	10	3.12
5.1.13		0.704351	8	0.512805	6	3.26
5.2.13		0.704181	8	0.512836	10	3.86
6.2.13	TBA	0.704455	8	0.512815	8	3.45
6.27.08	TBA	0.704307	12	0.512783	12	2.83
7.29.08	TBA	0.70433	8	0.512799	14	3.51
<i>Syunik</i>						
8.3.15	TBA	0.704438	8	0.512785	8	2.87
8.4.15	TBA	0.704466	8	0.512794	9	3.04
8.5.15	Te	0.704615	8	0.512779	10	2.75
8.6.15	PTe	0.704553	10	0.512779	12	2.75
9.1.15	TBA	0.704439	10	0.512803	8	3.22
9.2.15	TB	0.704407	14	0.512801	8	3.18
10.1.15	PTe	0.704660	10	0.512761	10	2.40
10.2.15	TBA	0.704262	10	0.512856	10	4.25
10.3.15	TBA	0.704245	10	0.512821	10	3.57
11.1.15	TA	0.704378	10	0.512806	8	3.28
11.2.15	PTe	0.704595	12	0.512781	11	2.79
11.3.15	Te	0.704277	12	0.512795	5	3.06
5.5.12	TA	0.704432	10	0.512783	10	2.83
2.3.10	R	0.704425	8	0.512788	8	2.93
3.10.10	Tr	0.704419	8	0.512796	4	3.08
3.11.10	TA	0.704530	10	0.512783	4	2.83
4.19.10	TA	0.704330	8	0.512801	12	3.18
1.4A.08	TBA	0.704317	6	0.512792	10	3.00
1.4B.08 (xenolith)	G	0.704981	32	0.512797	12	3.10
2.7.08	PTe	0.704440	10	0.512797	7	3.10
2.8.08	TBA	0.704381	10	0.512799	10	3.14
2.10.08	B	0.704396	14	0.512827	12	3.69
4.15A.08	R	0.705207	6	0.512783	8	2.83
4.18A.08	R	0.704857	8			
5.20A.08	R	0.705482	14	0.512770	14	2.57
5.21.08	TBA	0.704475	9	0.512767	10	2.52
6.26.08	TA	0.704273	8			

Rock type abbreviations are as in Fig. 6 and Table 2. All Sr ratios are normalized to NIST SRM 987, and all Nd ratios are normalized to La Jolla corrected BHVO-1 (see Methods). Epsilon values are calculated for the present day using  $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} = 0.512638$  for Chondritic Earth (Jacobsen & Wasserburg, 1980).

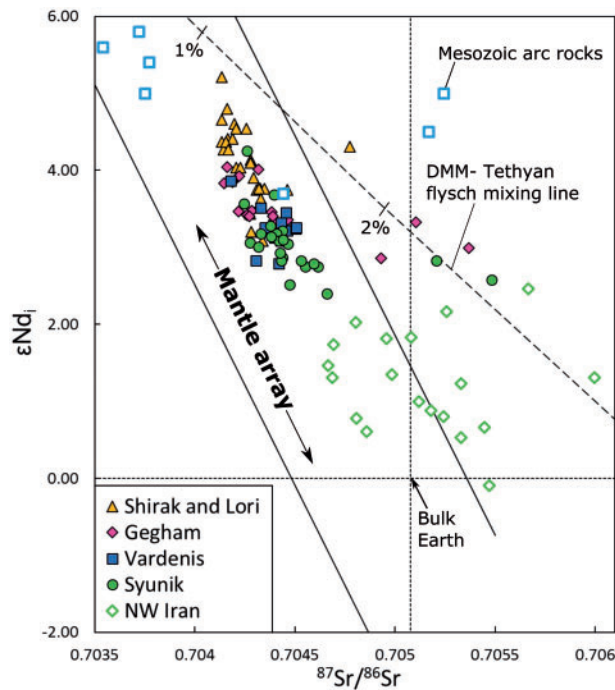
\*Samples from the east side of the Sevan–Akera suture (see text for discussion).

$$\frac{C_l}{C_0} = \frac{1}{D + F(1 - P)} \quad (1)$$

where  $C_l$  is the concentration of an element in the liquid,  $C_0$  is the concentration of that element in the source,  $D$  is the element's bulk partition coefficient,  $F$  is the fraction of melting, and  $P$  represents the partitioning of the element into the melt according to the proportion in which each mineral enters the melt. None of the samples in the region are in equilibrium with mantle olivine, so all must have undergone some fractional crystallization between last equilibration with the mantle and eruption. This is accounted for by assuming 8% fractional crystallization of an assemblage composed of 90% olivine and 10% spinel [following the approach of Shaw (2005)]. From equation (1) we can see that there are three parameters, each of which can be varied to

generate the trace element concentration of the most primitive southern Lesser Caucasus samples. First, the fraction of melting ( $F$ ) could vary, affecting the concentration of all incompatible trace elements. Second, if the modal mineralogy is changed then the partitioning of elements between source and melt [ $D$  and  $P$  in equation (1)] will change. Third, the concentration of elements in the source rock ( $C_0$ ) could be changed.

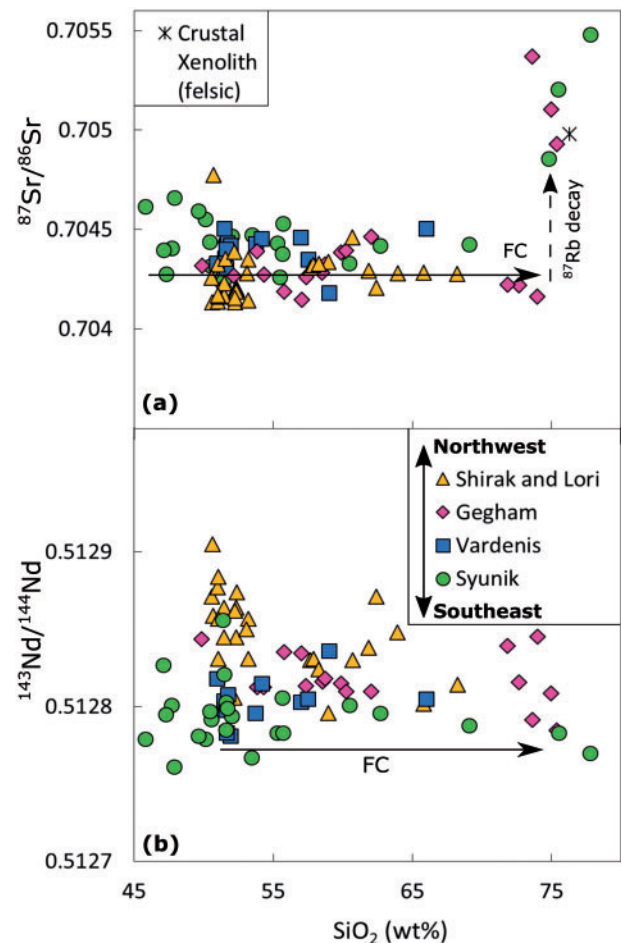
Our approach is to take the equivalent model of Neill *et al.* (2015) for the most geochemically depleted samples from the northern Lesser Caucasus and iteratively vary two of these parameters to attempt to reproduce the average composition of primitive basalts in the south. Variations in the mineralogy of the melt source are simplified to spinel versus garnet peridotite melting. Spinel peridotite modal mineralogy and melting proportions are from Shaw (2005) and Neill *et al.* (2015),



**Fig. 11.**  $\epsilon\text{Nd}$  vs  $^{87}\text{Sr}/^{86}\text{Sr}$  for volcanic rocks from the Lesser Caucasus. Bulk Earth value and mantle array are after Rollinson (1993). NW Iran data are from Allen *et al.* (2013); this region has very thick lithosphere, estimated to be >200 km by Priestley *et al.* (2012). Mixing line shown between depleted MORB mantle (DMM) and Tethyan flysch (Prelević *et al.*, 2008): DMM [Sr] = 21 ppm, [Nd] = 1.35 ppm,  $^{87}\text{Sr}/^{86}\text{Sr}$  = 0.7028,  $^{143}\text{Nd}/^{144}\text{Nd}$  = 0.5131; Tethyan flysch [Sr] = 349.2 ppm, [Nd] = 25.9 ppm,  $^{87}\text{Sr}/^{86}\text{Sr}$  = 0.7112,  $^{143}\text{Nd}/^{144}\text{Nd}$  = 0.512107.

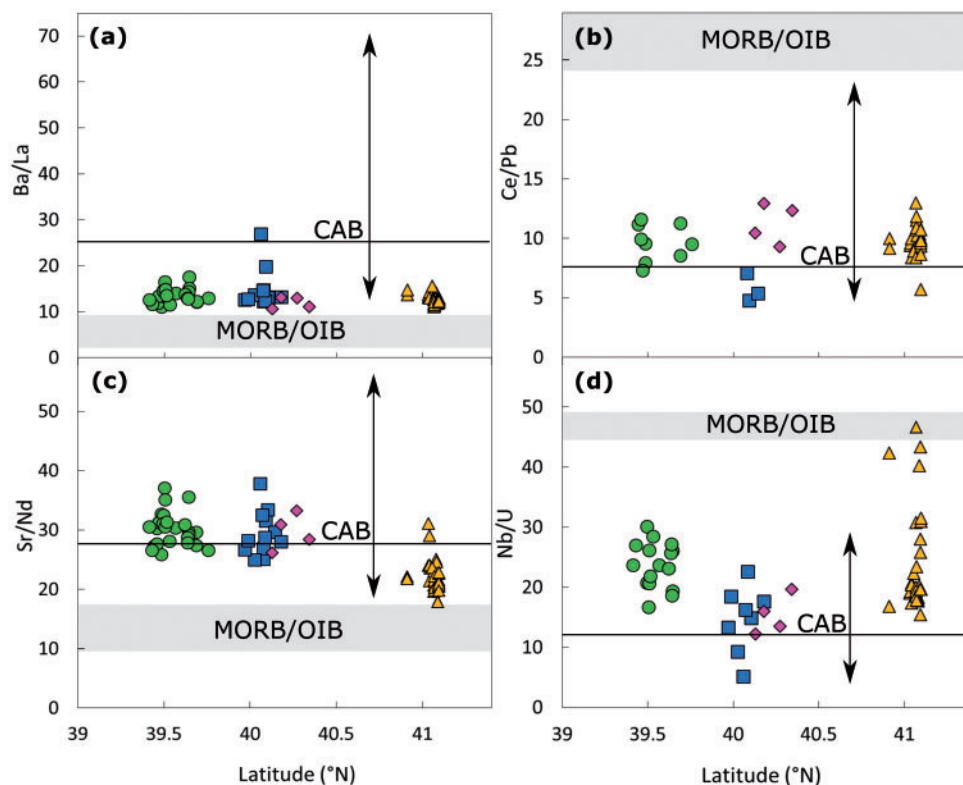
respectively, whereas for the garnet peridotite they are from Thirlwall *et al.* (1994) and Allen *et al.* (2013), respectively (Table 4). Varying the fraction of melting is self-explanatory. Remaining disparities between the model and observed trace element concentrations are likely to be due to differences in the composition of the mantle source. The different stages of melt modelling are shown in Table 4.

REE chemistry is used to obtain a qualitative understanding of the changing conditions of melting between the northern and southern Lesser Caucasus (Fig. 14). In Fig. 14a, two vectors are plotted, which show how the composition of the magmas should change with the degree of melting versus source mineralogy (presence or absence of garnet) and/or changes in source composition. Davidson *et al.* (2013) showed that REE partitioning between clinopyroxene and basaltic magma means that melting curves for peridotite will be very steep on a  $\text{Dy}/\text{Dy}^*$  versus  $\text{Dy}/\text{Yb}$  plot. The assumption that clinopyroxene is a residual phase is reasonable given the low degrees of melting (3% or less) previously estimated for the northern Lesser Caucasus (Neill *et al.*, 2015). Amphibole is also likely to be an important residual phase given the positive correlation between  $\text{Dy}/\text{Dy}^*$  and  $\text{Ti}/\text{Ti}^*$  (the size of the Ti anomaly in Fig. 8; not shown here). It is unclear from Fig. 14a whether garnet in the mantle source, or a LREE-enriched source is



**Fig. 12.** (a)  $^{87}\text{Sr}/^{86}\text{Sr}$  vs  $\text{SiO}_2$  (wt %) for all samples across the Lesser Caucasus. Fractional crystallization is denoted by a horizontal arrow labelled FC. The dashed line labelled  $^{87}\text{Rb}$  decay refers to the effect of post-crystallization radioactive decay in rocks with very high Rb/Sr ratios (see text for discussion). The crustal xenolith was collected in association with sample 1-4A-08 in the Syunik volcanic highland. It has a granitic texture. (b)  $^{143}\text{Nd}/^{144}\text{Nd}$  vs  $\text{SiO}_2$  (wt %).

responsible for the high Dy/Yb ratio in southern Lesser Caucasus samples. Both should give vectors close to horizontal in Fig. 14a, as both changes will steepen the REE profile rather than changing its curvature (Davidson *et al.*, 2013). The southern Lesser Caucasus samples have lower HREE abundances than samples from the north (Fig. 9). HREE all behave incompatibly during melting of spinel peridotite (Table 4), such that lower degrees of melting as suggested by the lower  $\text{Dy}/\text{Dy}^*$  ratio should only increase the concentration of HREE. Therefore, garnet as a residual phase during melting must explain at least some of the increase in  $\text{Dy}/\text{Yb}$  in Fig. 14a. In Fig. 14b, the northern Lesser Caucasus samples sit close to the spinel peridotite melting curve, but the southern Lesser Caucasus samples are intermediate between the spinel and garnet melting curves, suggesting melting of a mixed source involving both garnet and spinel peridotite. This could be a consequence of polybaric melting across the spinel–garnet



**Fig. 13.** The latitudinal variation in the ratios Ba/La (a), Ce/Pb (b), Sr/Nd (c) and Nb/U (d) of mafic volcanic rocks along the Lesser Caucasus mountain chain. These ratios are nearly invariant in MORB and OIB (shaded fields), but arc basalts show substantial variability in these ratios [CAB (continental arc basalt), with arrows showing the ranges across several arcs from Kelemen *et al.* (2003)]. As such, they are a proxy for slab input, and show a uniform input across the Lesser Caucasus. After Turner *et al.* (2017).

transition at ~75 km depth (Robinson & Wood, 1998), suggesting melting at greater depths in the south.

Returning to the partial melting model, the melt fraction was estimated on the basis that the Nb concentration (Table 4) is not affected by the mineralogy of the source rock, and assuming that its concentration in the source rock is constant along the transect. Utilizing this assumption gives 3% melting in the north, versus 1% melting in the south (Fig. 15). The proportions of garnet peridotite and spinel peridotite that contribute to the total mantle source are estimated from the Hf and Yb concentrations. Both elements are assumed to have constant concentrations in the mantle source because HFSE and HREE would be less affected by metasomatic events that could alter the source composition. The best fit to the Hf and Yb data is a magma source that is 65% garnet peridotite and 35% spinel peridotite (Fig. 15).

Across the Lesser Caucasus, as lithosphere thickness increases, the degree of melting decreases and melting occurs at greater depths, with garnet as a residual phase for a significant portion of the melting interval. However, the model is still unable to explain the high concentrations of several of the LREE (La, Gd and Dy; Fig. 15, black line), and requires changes in the composition of the mantle source with lithospheric thickness as well.

To understand how melting is occurring, it is important to understand where magma is forming with respect to the lithospheric structure. For this, we have used major element thermobarometry to estimate the conditions of last equilibration between magmas and the mantle.

### Pressure and temperature of melting

To calculate these intensive parameters, we use the parameterizations of Plank & Forsyth (2016), based on the major element chemistry of primitive magmas, building on the work of Lee *et al.* (2009). These calculations are for magmas produced from a peridotite mantle source at pressures below 3 GPa (Plank & Forsyth, 2016). The temperature ( $T$ ) dependence of Mg partitioning between olivine and melt (Roeder & Emslie, 1970) and the pressure ( $P$ ) dependence of silica activity in melts co-saturated in olivine and orthopyroxene (Carmichael *et al.*, 1970) are exploited to give the equations

$$T = 1264.5 + 7.85(\text{Mg}_4\text{Si}_2\text{O}_8) + \frac{8545}{\text{Si}_4\text{O}_8} - 5.96(\text{Al}_{163}\text{O}_8) - \Delta T_{\text{H}_2\text{O}} - \Delta T_{\text{CO}_2} \quad (2)$$

**Table 4:** Parameters used for non-modal batch melt modelling

Lithology		OL	OPX	CPX	AMPH	SP	GRNT	Sum
Spinel peridotite	Source mode	0.794	0.123	0.03	0.042	0.011	0	1
	Melt mode	0.15	0.15	0.22	0.42	0.06	0	1
Garnet peridotite	Source mode	0.569	0.212	0.077	0.026	0	0.116	1
	Melt mode	0.05	0.19	0.28	0.06	0	0.42	1
Starting composition		La	Gd	Dy	Yb	Zr	Hf	Nb
		0.5869	0.321	0.317	0.202	7	0.172	0.3635
Kd values		La	Gd	Dy	Yb	Zr	Hf	Nb
Olivine		0.0001	0.00076	0.0014	0.00364	0.004	0.006	0.0002
Orthopyroxene		0.0002	0.0128	0.0261	0.0986	0.005	0.01	0.0005
Clinopyroxene		0.054	0.4	0.442	0.427	0.13	0.2	0.0077
Amphibole		0.086	0.64	0.707	0.683	0.156	0.24	0.2
Spinel		0.0004	0.00042	0.0004	0.00053	0.005	0.01	0.001
Garnet		0.01	0.498	1.06	4.03	0.12	0.23	0.00054
Melting models		La	Gd	Dy	Yb	Zr	Hf	Nb
1. 3% melting spinel peridotite (Neill <i>et al.</i> , 2015)	Initial	17.3	5.3	4.9	2.7	169.0	3.6	10.0
	After 8% fractionation	18.8	5.8	5.3	2.9	183.6	3.9	10.9
2. 1% melting spinel peridotite	Initial	39.5	6.8	5.9	3.2	299.9	5.6	20.2
	After 8% fractionation	42.9	7.3	6.5	3.5	325.9	6.1	21.9
3. 1% melting garnet peridotite	Initial	34.0	2.8	1.7	0.4	175.3	2.8	23.1
	After 8% fractionation	37.0	3.1	1.9	0.4	190.5	3.1	25.1
4. 1% melting of 65% garnet peridotite 35% spinel peridotite	Initial	35.9	4.2	3.2	1.4	218.9	3.8	22.1
	After 8% fractionation	39.1	4.6	3.5	1.5	237.9	4.1	24.0
Most primitive Syunik sample for comparison		74.8	7.6	4.7	1.9	166.0	3.7	24.0

The source modes for spinel and garnet peridotite are from Shaw (2005) and Thirlwall *et al.* (1994) respectively. Melt modes are from Allen *et al.* (2013) and Neill *et al.* (2015). The starting composition of the mantle source is the same as that used by Neill *et al.* (2015). Partition coefficients are from Ionov *et al.* (2002), except for those for garnet, which come from McKenzie & O'Nions (1991) and Adam & Green (2006). The most primitive Syunik sample used for comparison is 8-5-15. OL, olivine; OPX, orthopyroxene; CPX, clinopyroxene; AMPH, amphibole; SP, spinel; GRNT, garnet.

$$P = \frac{\left( \ln(\text{Si}_4\text{O}_8) - 4.045 + 0.0114(\text{Fe}_4\text{Si}_2\text{O}_8) + 0.00052(\text{Ca}_4\text{Si}_2\text{O}_8)^2 + 0.0024(\text{Mg}_4\text{Si}_2\text{O}_8) \right)}{-336.3T^{-1} - 0.0007T^{1/2}} \quad (3)$$

All major element oxides are calculated in mol % as described by Lee *et al.* (2009), except that they are calculated on an anhydrous basis. The two terms  $\Delta T_{\text{H}_2\text{O}}$  and  $\Delta T_{\text{CO}_2}$  account for the lower temperature and higher pressure melting in the presence of volatiles as follows:

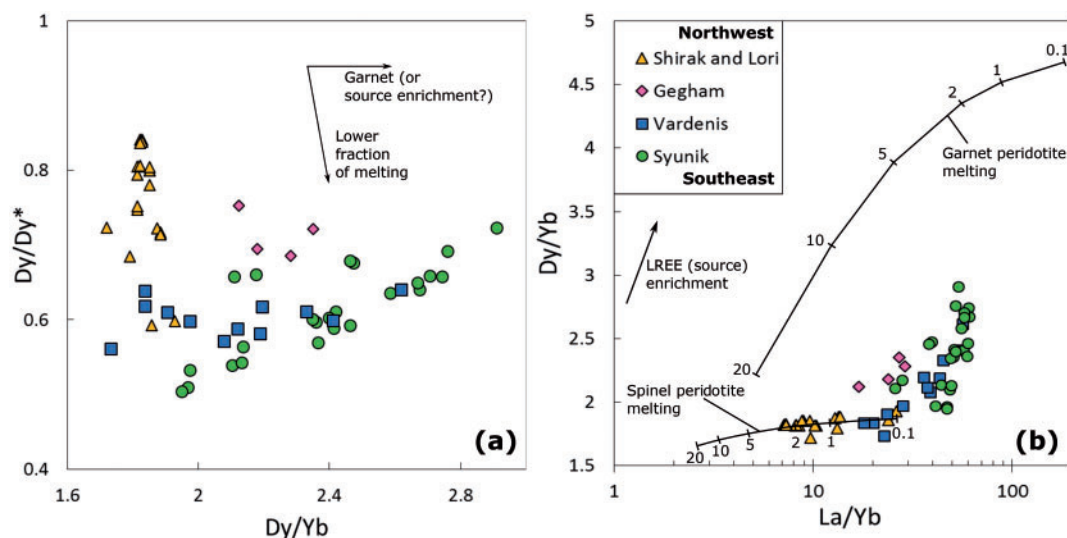
$$\Delta T_{\text{H}_2\text{O}} = 40.4(\text{H}_2\text{O}) - 2.97(\text{H}_2\text{O})^2 + 0.0761(\text{H}_2\text{O})^3 \quad (4)$$

$$\Delta T_{\text{CO}_2} = \frac{(\text{SiO}_2) - 50.3}{0.12(-1.067)} \quad (5)$$

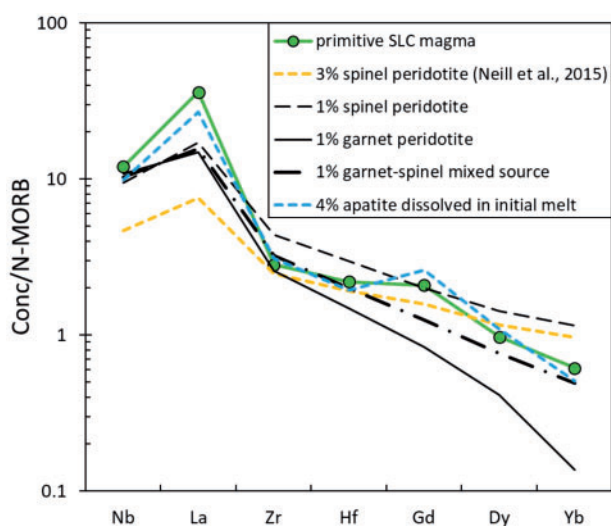
where  $\text{H}_2\text{O}$  and  $\text{SiO}_2$  are in wt %. To use these equations, the composition of the primary melt must be known. Given that no samples from the Lesser Caucasus have  $\text{Mg\#} > 67$ , it is clear that all the studied samples have undergone some degree of fractional crystallization. If only olivine has fractionated, then it is a simple process to add olivine incrementally to the

melt until it is in equilibrium with mantle olivine of  $\text{Fo}_{90}$  (Lee *et al.*, 2009). If, however, other phases, such as amphibole or clinopyroxene, have also fractionated, then this calculation becomes non-trivial. In a study on the collision-related volcanism of Anatolia, McNab *et al.* (2018) took 8.5 wt % bulk-rock MgO as a reasonable lower limit for magmas that have fractionated only olivine. Only two such samples exist in the Lesser Caucasus, and it is likely that they have both been affected by olivine accumulation rather than being truly primitive melts. Instead, the complete sample suite from the Shirak and Lori, and Syunik volcanic highlands (1 and 4 in Fig. 1) is used to project back to the likely composition at 8.5 wt % MgO of two end-member primitive magmas for the Lesser Caucasus. This correction is shown in Supplementary Data B.

To calculate the pressures and temperatures of melting, two magma composition parameters must be constrained:  $\text{Fe}^{3+}/\Sigma\text{Fe}$  and water content. The water content of the primary magma has a large effect on the temperature estimate through equation (4) ( $\sim 25^\circ\text{C}$  per wt %), but a smaller effect on the pressure estimate ( $\sim 30$  MPa per wt %). The presence of hydrous mineral



**Fig. 14.** (a)  $Dy/Dy^*$  vs  $Dy/Yb$  after Davidson *et al.* (2013). The two arrows depict the expected vectors from changing the fraction of melting or source mineralogy or composition. The vector for a lower fraction of melting is after Davidson *et al.* (2013, fig. 4), based on melting of an olivine–pyroxene–amphibole-dominated mantle source. The presence of garnet in the magma source should follow a horizontal vector after Davidson *et al.* (2013, fig. 5). (b)  $Dy/Yb$  vs  $La/Yb$ , melting curves based on the modal mineralogies, melting modes, and partition coefficients given in Table 5.



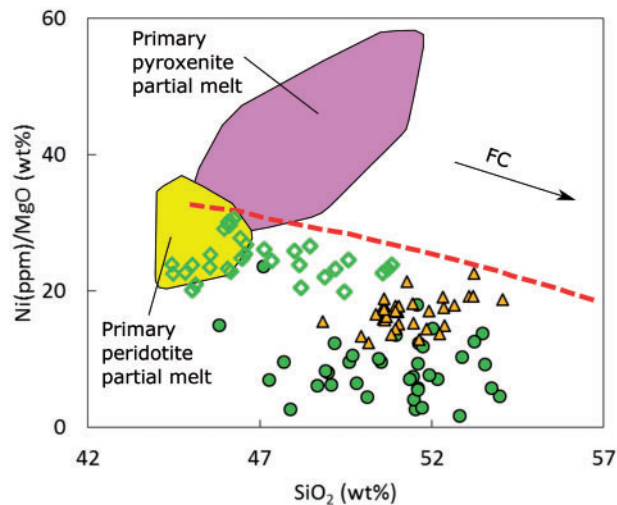
**Fig. 15.** N-MORB-normalized concentrations of Nb, La, Zr, Hf, Gd, Dy and Yb for the average of mafic samples from Syunik volcanic highland (green line) and several non-modal batch melting models that attempt to explain these trace element concentrations. Melting models: spinel peridotite at 3% melting (yellow dashed line), 1% melting of spinel peridotite (dashed black line), 1% melting of garnet peridotite (black continuous line) and 1% melting of a source composed of 65% garnet peridotite and 35% spinel peridotite (bold dash-dot line). The blue dashed line represents a magma derived from the same 65% garnet peridotite source, with 4% apatite dissolved in the melt. Apatite composition is for magmatic apatite from Western Turkey of Prelević *et al.* (2015).

phases in mafic rocks (as opposed to only in more evolved samples), only in the southern Lesser Caucasus (Fig. 5), suggests higher water contents in the south. Based on previous studies of amphibole peridotite xenoliths (Thirlwall *et al.*, 1994; Ionov & Hofmann,

1995), it is assumed that the mantle source contains between 2 and 10 modal % amphibole, which is assumed to contain 2 wt %  $H_2O$ , such that the mantle source could have up to 0.2 wt %  $H_2O$ . Using Ce partition coefficients for water during mantle melting (Thirlwall *et al.*, 1994; Dixon *et al.*, 2002; Ionov *et al.*, 2002), and the melting models shown in Table 4, gives 2–7 wt %  $H_2O$  for southern Lesser Caucasus primary magmas, and 1–4.6 wt % for the north.

As only ferrous iron substitutes in olivine, a higher  $Fe^{3+}/\Sigma Fe$  increases the apparent melt Mg#, and so reduces the amount of olivine addition required to produce the primary magma, and thus the calculated pressure and temperature. The  $Fe^{3+}/\Sigma Fe$  ratio varies from a minimum of 0.1 in MORB (Cottrell & Kelley, 2011) to as high as 0.3 in some arc basalts (Brounce *et al.*, 2014). Given the ubiquitous arc-type geochemical signatures in Lesser Caucasus volcanic rocks, we take 0.25 as a conservative estimate of the  $Fe^{3+}/\Sigma Fe$  ratio (Brounce *et al.*, 2014).

It should be noted that equations (2) and (3) will give meaningful estimates of the pressure and temperature of melting only if the mantle source is peridotitic, because the parameterizations assume the melt is saturated in both orthopyroxene and olivine (Plank & Forsyth, 2016). The use of these equations on magmas derived from a pyroxenite source would yield meaningless results. Figure 16 shows that for Lesser Caucasus and NW Iran samples, pyroxenite is not a major component of the mantle source. Pyroxenite partial melts have much higher Ni/MgO ratios. As magmas fractionate, those derived from a peridotite source should evolve along a trajectory below the dashed line in Fig. 16, whereas those from a pyroxenite source would plot above the line.

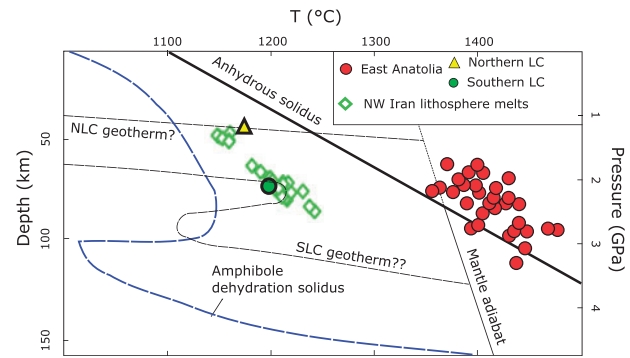


**Fig. 16.** Ni/MgO vs SiO<sub>2</sub> in primitive Lesser Caucasus and NW Iran bulk-rock samples (<54% SiO<sub>2</sub>, >4% MgO), after Allen *et al.* (2013). Pyroxenite melting should generate compositions above the bold red dashed line (Sobolev *et al.*, 2005). Primary melts of peridotite and pyroxenite (from Hawaii), after Sobolev *et al.* (2005). FC, fractional crystallization. Syunik, Shirak and Lori and NW Iran data are shown with the same symbols as in other figures.

Based on our water content estimates, southern Lesser Caucasus magmas last equilibrated with the mantle at between 1198 and 1292°C, whereas northern Lesser Caucasus magmas last equilibrated at between 1174 and 1267°C. Given the observation that amphibole is likely to be a residual phase in the mantle source (see previous section), it is likely that melting was occurring at temperatures close to the amphibole dehydration solidus, because amphibole is likely to completely break down within a few tens of degrees centigrade of crossing the dehydration solidus (Green & Falloon, 2005; Mandler & Grove, 2016). As such, it is likely that the minimum temperatures of these ranges are the more realistic. Using these temperatures gives pressure estimates of 2.1 GPa (~75 km) in the southern Lesser Caucasus, and 1.2 GPa (~45 km) in the north. These estimates are shown in Fig. 17, along with the position of the amphibole dehydration solidus. Southern Lesser Caucasus magmas are produced deeper, but at similar temperatures to those in the north.

The parameterizations first developed by Lee *et al.* (2009) were designed to expand the applicability of basalt melt geothermobarometry beyond mid-ocean ridge systems to any setting involving the melting of terrestrial peridotite. Both our modelled major element compositions and the pressure and temperature calculated fall within the experimental range of the dataset of Lee *et al.* (2009), and to this extent our approach is justified.

However, the presence of metasomatic phases such as amphibole in the mantle source means that the pressures and temperatures calculated here must be interpreted with caution. Alkaline volcanic rocks from monogenetic volcanic fields in Western Mexico,



**Fig. 17.** Depth vs temperature of melting for the northern Lesser Caucasus and southern Lesser Caucasus. Also shown are the anhydrous peridotite solidus (1100°C + 3.5°C km<sup>-1</sup>) after Plank & Forsyth (2016), the wet solidus (amphibole present) after Green & Falloon (2005), and samples from East Anatolia with low K/Nb (red circles) from McNab *et al.* (2018). The pressures and temperatures of melting of NW Iran magmas were calculated for this study assuming Fe<sup>3+</sup>/ΣFe of 0.25 and melt water contents of 7 wt %, as were used for the southern Lesser Caucasus estimate. (See text for discussion.)

geochemically similar to those from volcanic highlands in the Lesser Caucasus, have very high Fe<sup>3+</sup>/ΣFe ratios and whole-rock Mg# in excess of the ~72 value normally assumed to be in equilibrium with mantle peridotite (Carmichael *et al.*, 1996). The effects of a potential underestimation of both Fe<sup>3+</sup>/ΣFe and the amount of olivine addition required to produce the primary magma will tend to counteract each other, where the former would give an overestimation of temperature, and the latter an underestimate. In the case of Western Mexico, both of these observations are ascribed to the incongruent melting of phlogopite (Carmichael *et al.*, 1996), which does not seem to be a major metasomatic phase in the Lesser Caucasus mantle (see the section 'How does thicker mantle lithosphere influence the composition of the mantle source?' below).

Even if these issues are relevant to the Lesser Caucasus, the constraint provided by the position of the amphibole dehydration solidus shows the temperature estimates provided here to be reasonable, although some of the ~50°C disparity with the dehydration solidus could be explained by these issues. The sensitivities of the pressure estimate to Fe<sup>3+</sup>/ΣFe (-0.2 GPa per 0.1 increase), primary melt Mg# (+0.5 GPa per 10% increase) and estimated temperature (+0.1 GPa per 100°C) are all small. The pressure estimate is primarily based on silica activity in a system co-saturated in olivine and orthopyroxene. Given the indications from Fig. 16, and reasonably successful trace element modelling of partial melting of a peridotite mantle source, it seems likely that these pressure estimates are robust.

Further constraints on the thickness of the crust are required to interpret these results. Crustal thickness is estimated using the formulations of Hu *et al.* (2017), which link the Sr/Y and (La/Yb)<sub>N</sub> ratios of intermediate magmatic rocks from continental collision zones with the crustal thickness. The basis of this technique is the

polarizing effects different fractionating mineral assemblages have on the Sr/Y and (La/Yb)<sub>N</sub> ratios in shallow versus deep storage reservoirs. To the nearest 5 km, we estimate crustal thickness as 45 km in Shirak and Lori, 55 km in Gegham, and 60 km in both Vardenis and Syunik.

The depth of melting in the northern Lesser Caucasus is very similar to the ~45 km Moho depth, possibly suggesting the presence of a very thin mantle lithosphere. It is instructive to compare this result with recent estimates of melting conditions in neighbouring Eastern Anatolia from McNab *et al.* (2018). East Anatolian magmas can be split into high K/Nb (>500) and low K/Nb types. Low K/Nb magmas generally have OIB-like geochemistry, and are considered to be directly derived from melting of the convecting mantle, which is apparently anomalously hot in the region, plotting to the right of the ambient mantle adiabat (Fig. 17), at a much higher temperature than any temperatures modelled for the northern Lesser Caucasus. This can be interpreted as northern Lesser Caucasus magmas being derived from the lithosphere, and not the convecting mantle.

Deeper melting in the south is associated with thicker crust, but magma generation is occurring at much shallower depths than the >100 km lithosphere–asthenosphere boundary estimated by Priestley *et al.* (2012). The depth of melt equilibration is much deeper than the anhydrous solidus, confirming the need for a volatile-enriched lithospheric mantle root.

These *P–T* conditions of melting can be compared with estimates made for this study of the melting conditions of samples from NW Iran, based on the data of Allen *et al.* (2013). The *P–T* conditions of melting of NW Iran volcanic rocks were estimated using the same Fe<sup>3+</sup>/ΣFe and water contents as for southern Lesser Caucasus magmas (0.25 wt % and 7 wt %, respectively). This is on the basis that the geochemistry of NW Iran magmas is similar to that of southern Lesser Caucasus magmas (Fig. 8), and both formed in a thicker lithosphere regime. Most of the magmas in NW Iran formed under similar conditions to southern Lesser Caucasus magmas, with a small subsidiary group of samples that formed at shallower depths, similar to northern Lesser Caucasus magmas, possibly representing magmas that re-equilibrated with the mantle at the base of the crust during their ascent to the surface.

The melting depth in both the southern Lesser Caucasus and NW Iran is significantly shallower than the estimated lithospheric thicknesses. One possible explanation for what these melting conditions represent could be a thermal maximum in a back-bent non-linear geothermal gradient (Fig. 17). In recently thickened orogenic lithosphere, this type of kinked profile is probably more realistic than the linear geothermal gradient more typical of a cratonic region (Mather *et al.*, 2011).

It is worth noting that these differences in melting conditions probably reflect variations in the location of magma generation today, and are not the product of

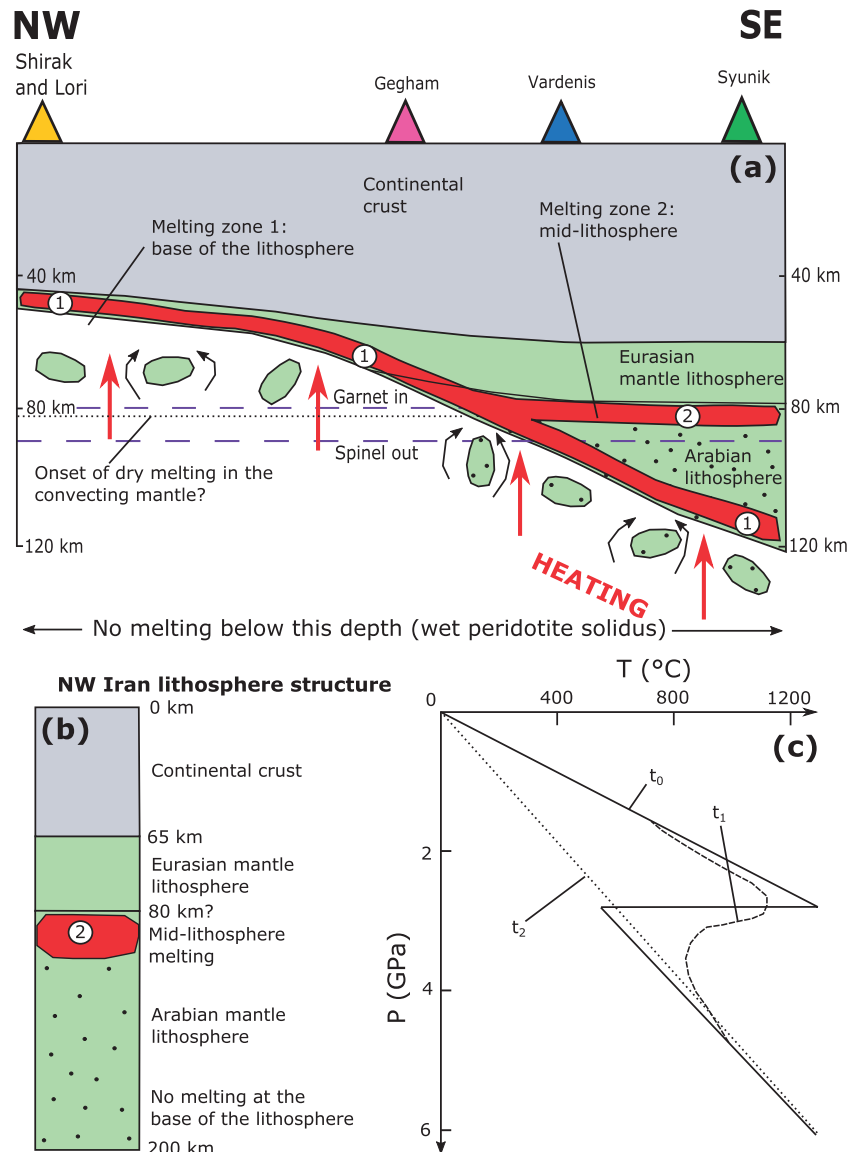
temporal changes in lithosphere structure. This is because, in Fig. 8, the trace element compositions of basalts from each volcanic highland are distinct, with each having a fairly narrow range of trace element compositions, despite all the volcanic highlands spanning an age range >1 Myr, and with the Gegham and Syunik highlands thought to have very similar timespans of activity (Joannin *et al.*, 2010; Lebedev *et al.*, 2013).

### Geodynamic implications

Using constraints from geochemistry and the estimated pressures and temperatures of melting, it is possible to develop a model of how melting occurs under the Lesser Caucasus region. It seems clear that a majority of melting is taking place in the lithosphere for two reasons. First, all samples display an arc-type geochemistry. Subduction ceased around 35 Ma (Rolland *et al.*, 2012), such that these arc signatures will probably be preserved only in the lithosphere. It is possible that the mantle lithosphere retains phases such as amphibole and rutile, which, if equilibrated with the magmas, are sufficient to impart the arc-type geochemical signature (Allen *et al.*, 2013). Second, both northern and southern Lesser Caucasus magmas appear to be produced at significantly lower temperatures than the low K/Nb magmas of Eastern Anatolia, which are thought to be derived from the convecting mantle (McNab *et al.*, 2018), such that the Lesser Caucasus magmas either have re-equilibrated in colder lithosphere or are entirely derived from the lithosphere.

Based on the smooth trend in <sup>87</sup>Sr/<sup>86</sup>Sr versus <sup>143</sup>Nd/<sup>144</sup>Nd in Fig. 11 and the gradational changes in trace element patterns (Fig. 8), it appears that magmas are produced from melting of two source types in varying proportions. If these two source types reflect geochemical end-members, one would be typified by the northern Lesser Caucasus volcanic rocks, whereas the other end-member would be most clearly seen in NW Iran.

Unfortunately, constraints on the lithospheric thickness are very limited for the northern Lesser Caucasus (Fig. 1), with the Priestley *et al.* (2012) model showing only that it is <100 km. Taking a typical conductive geothermal gradient for this lithosphere, melting would be expected to occur close to the lithosphere–asthenosphere boundary. Mantle–melt equilibration close to the base of the crust (Fig. 17) suggests that the lithosphere is very thin. If this were the case it is likely that there would be some melting of the convecting mantle based on elevated mantle *T<sub>p</sub>* in the region (McNab *et al.*, 2018). However, such a thin lithosphere would make it difficult to insulate the crust from heating, such that more crustal contamination would be expected. As was noted above, this mantle source type appears also to be present in the southern Lesser Caucasus, where the lithosphere is sufficiently thick to suppress melting of anhydrous convecting mantle, but where melting of a hydrous lithospheric source would still be possible. It



**Fig. 18.** (a) Geodynamic model of magmatism showing the lithosphere and upper mantle across the Lesser Caucasus. Crustal thickness is estimated on the basis of formulations relating the Sr/Y and  $(La/Yb)_N$  ratios in intermediate volcanic rocks to Moho depth (Hu *et al.*, 2017). Data filtering follows the approach outlined by Hu *et al.* (2017). Each volcanic highland has a range of Moho depth estimates of around 20 km. For both the Sr/Y and  $(La/Yb)_N$  the median value for each volcanic highland is taken and then the two values are averaged to give an estimate of crustal thickness. Lithosphere thickness estimates are from Priestley *et al.* (2012). It is noted that they only estimate thicknesses where the lithosphere is >100 km, so the lithospheric thickness in the NW is schematic after Neill *et al.* (2015). Melting at the wet peridotite solidus lower depth limit is on the basis of upwelling convecting mantle at 1300–1400°C  $T_p$ , which would lead to the wet solidus being crossed at ~140 km depth (Green & Falloon, 2005). Also shown are the two melting zones (labelled 1 and 2) that are discussed in the text. Melting zone 1 is along the base of the lithosphere, and is suggested to be in response to small-scale convective removal of the lithosphere. Melting zone 2 is in the mid-lithosphere. (b) Sketch of the lithospheric structure of NW Iran and the location of melting in the mid-lithosphere. (c) Schematic illustration of the thermal relaxation of a kinked geothermal gradient leading to melting in the mid-lithosphere from initial underplating at time  $t_0$ , through thermal evolution at time  $t_1$ , to establishing a cratonic geotherm by time  $t_2$ .

is likely that this magma type is derived from melting of the base of the lithosphere in response to small-scale convective removal, as suggested by Neill *et al.* (2015). This is shown in Fig. 18a as melt zone 1, with small portions of lithosphere being delaminated. This type of delamination is suggested to occur because a very small amount of water (a few hundred ppm) left over from previous subduction lowers the viscosity of the mantle sufficiently to allow more vigorous convection to render

the lithosphere–asthenosphere boundary unstable (Kaislaniemi *et al.*, 2014). Small-scale convective removal is preferred over catastrophic large-scale delamination because volcanism in the Lesser Caucasus is generally small-scale and sustained, with no evidence of crustal contamination, consistent with a continuous, less invasive process. Although this magma type is a mixing component in the southern Lesser Caucasus, it is not seen in NW Iran because the lithosphere is too



thick for wet or dry melting at its base (Green & Falloon, 2005; Priestley *et al.*, 2012). As the lithosphere is thickened, the degree of melting in response to convective removal will decrease as the wet peridotite solidus is approached with depth (Fig. 18a), which may help to explain the lower degrees of melting in the south.

The second melting zone is the sole magma source for volcanic rocks in NW Iran (Fig. 18b). Despite very different lithosphere thicknesses, both NW Iran and southern Lesser Caucasus magmas formed under similar conditions within the lithosphere (Fig. 17). It is suggested that continental collision could have led to the formation of a kinked geothermal gradient (Fig. 18c). A linear craton-style geothermal gradient is likely to be unrealistic for orogenic lithosphere (Allen *et al.*, 2013). It has been suggested that orogenesis may proceed by underthrusting of mantle lithosphere from the oncoming plate (e.g. Willett *et al.*, 1993). In this case underthrusting of Arabian mantle lithosphere, with its own pre-existing geothermal gradient, may lead to a kinked geothermal gradient with the kink forming along the subduction plane between overlying Eurasian lithosphere and underlying Arabian lithosphere. Upon initial collision, this kink will be very sharp ( $t_0$  in Fig. 18b), but over time the gradient will thermally relax ( $t_1$ ) towards a gradient that resembles the one characteristic for cratons ( $t_2$ ). This thermal relaxation would heat the lithosphere just under the kink, which would previously have been cool. Dewatering and melting of this horizon would follow, which is what we argue we see in our  $P$ - $T$  estimates of magma generation in Fig. 17. Given the lower degrees of melting in the southern Lesser Caucasus, it appears that this mechanism develops only low-degree melts.

To summarize, in the northern Lesser Caucasus magma is generated by small-scale delamination events heating the base of the lithosphere. This process also occurs in the southern Lesser Caucasus, but at greater depths. Magmas generated by this mechanism mix with a second type of magma produced in the mid-lithosphere from thermal relaxation of a kinked geotherm. Further south, in NW Iran, melting of the base of the lithosphere is suppressed owing to the depth being too great even for wet melting. However, melting in the mid-lithosphere continues.

### How does thicker mantle lithosphere influence the composition of the mantle source?

The gradient in Sr-Nd isotope ratios in Fig. 11, along with a lack of evidence for crustal contamination, requires that there be some variation in the composition of the mantle source. One possibility is that magmas are tapping different lithosphere domains. Northern Lesser Caucasus magmas are exclusively found on Mesozoic arc lithosphere (Fig. 2), whereas southern Lesser Caucasus magmas are on South Armenian Block lithosphere, or else are very close to the suture. However, crossing such lithospheric sutures

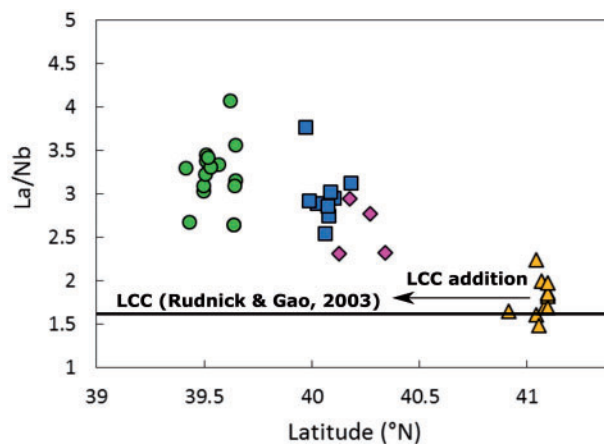
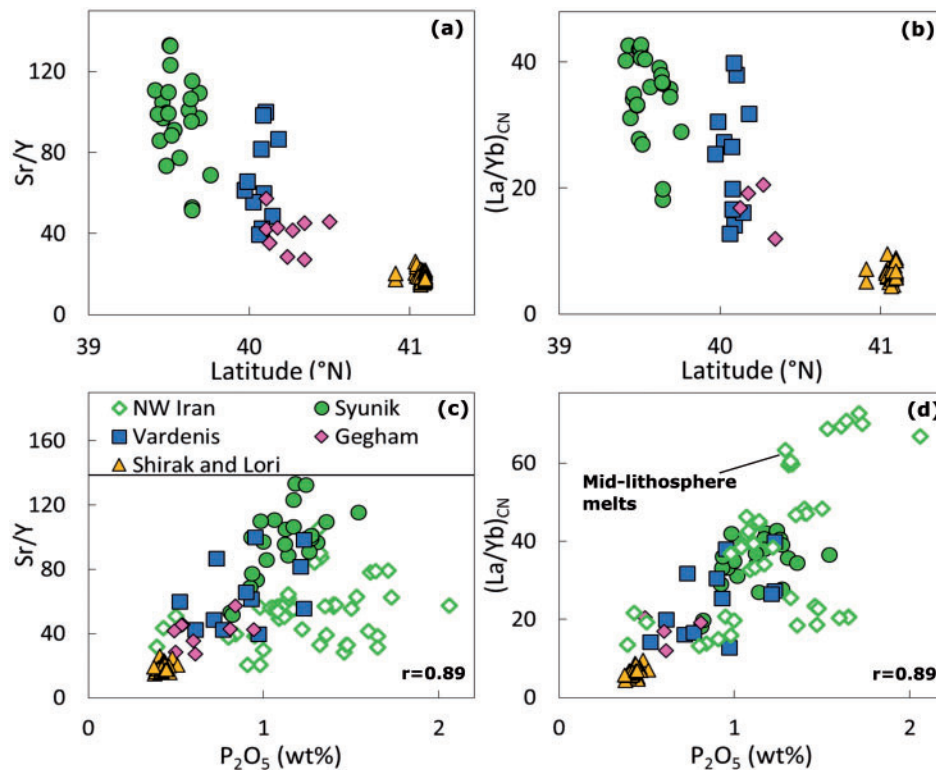


Fig. 19. La/Nb vs latitude for primitive volcanic rocks in the Lesser Caucasus. The least enriched Shirak and Lori samples have a La/Nb ratio of  $\sim 1.6$ . Fraction of melting and garnet vs spinel peridotite melting have a limited effect on this ratio. Lower continental crust (LCC) has a La/Nb ratio with a similar value of around 1.6 (Rudnick & Gao, 2003). Therefore if LCC is enriching the melt source, the La/Nb ratio should be close to invariant.

would probably result in a step-change in isotope ratios, rather than the smooth gradation observed in the Lesser Caucasus. A minority of samples from the Vardenis volcanic highland (Figs 1 and 2) are thought to be on the east (Mesozoic arc) side of the suture; these are shown by an asterisk in Table 3. As Table 3 shows, these samples have Sr-Nd isotope compositions indistinguishable from those of other Vardenis samples, suggesting that the suture zone is not a major dividing line in isotope composition. This suggests that melting zone 2 in the mid-lithosphere has a different composition from melting zone 1 at the base of the lithosphere. In this discussion on the nature of the mantle sources, only the most mafic samples ( $>4$  wt % MgO,  $<54$  % SiO<sub>2</sub>) are used to try and minimize the effects of fractional crystallization on trace element contents.

The lower crust is thought to behave as a weak layer during continental collision (Bürgmann & Dresen, 2008). This could lead to some lower crust being incorporated into the mantle lithosphere perhaps during underthrusting of Arabian lithosphere, as has been suggested for numerical models of other collision zones such as the Himalayas (e.g. Toussaint *et al.*, 2004; Li *et al.*, 2011). This could lead to the addition of lower continental crust (LCC) to melting zone 2 in the middle of the lithosphere, and hence enrichment of the mantle source of the southern Lesser Caucasus. Incorporation of LCC into the mantle source could result in significant enrichments in all of the most incompatible trace elements, as in Fig. 8e, given the much higher concentrations of most incompatible trace elements in the lower crust relative to the primitive mantle (Sun & McDonough, 1989; Rudnick & Gao, 2003).

The La/Nb ratios of the least enriched samples from the northern Lesser Caucasus and the same ratio estimated for the LCC (Rudnick & Gao, 2003) provide a



**Fig. 20.** Sr/Y and La/Yb show strong variations between the northern and southern Lesser Caucasus as shown in (a) and (b). Both of these ratios show strong correlations with  $P_2O_5$  (wt %) (c, d);  $r = 0.89$  in both cases. NW Iran data in (c) and (d) represent melting in a region of very thick lithosphere [data from Allen *et al.* (2013)].

serendipitous coincidence, with both around 1.6 (Fig. 19). Given that La/Nb is unaffected by melting and crystallization (i.e. both elements have similar bulk partition coefficients with respect to the melt) regardless of source mineralogy (Ionov *et al.*, 2002; McKenzie & O’Nions, 1991), any variations in the La/Nb ratio should reflect variations in the source La/Nb. If the lower crust was responsible for the source enrichment, La/Nb should be near-constant along the transect. Samples from the southern Lesser Caucasus have much higher La/Nb ranging from 2.5 to 5 (Fig. 19), which cannot be explained by the addition of average LCC.

However, it is worth noting that several estimates of the composition of individual lower crust sections do give higher La/Nb of up to 4.9 (Weaver & Tarney, 1984; Villaseca *et al.*, 1999; Jagoutz & Schmidt, 2013), such that the involvement of lower crust in the southern Lesser Caucasus magma source cannot be precluded.

Lithospheric mantle that is significantly shallower than the lithosphere–asthenosphere boundary (melting zone 2) is expected to be colder than the deep lithosphere (melting zone 1) prior to any heating, even if the lithosphere does have a kinked geothermal gradient (Fig. 17). These lower temperatures should stabilize minerals such as amphibole and phlogopite, which can retain chemical components derived from mantle metasomatism (Luth, 2003; Frost, 2006; and references therein). In the deep lithosphere, these components would

be subject to upward mobilization by fluid release following dehydration at higher pre-melting temperatures. Underthrusting of Arabian lithosphere, as suggested in Fig. 18, would also introduce a new lithospheric domain, which could have a different composition (including isotopically) from the Eurasian lithosphere.

If this metasomatic material is responsible for the enrichments in the southern Lesser Caucasus, then it must have a composition capable of producing those most enriched melts. La/Yb and Sr/Y are two ratios that increase most dramatically with latitude, as shown in Fig. 20a and b. Both of these ratios show excellent correlations with  $P_2O_5$  content (Fig. 20c and d) within all Lesser Caucasus samples. The La/Yb ratios and  $P_2O_5$  contents of the NW Iran samples (interpreted to be pure melting zone 2 magmas) appear as end-members on the Lesser Caucasus mixing line (Fig. 20d), suggesting that apatite may be a metasomatic phase in the host rock of melting zone 2 (Fig. 18). This is consistent with the presence of large apatite crystals in some southern Lesser Caucasus samples (up to 2 mm across; Fig. 5b). It is also consistent with previous suggestions that metasomatic apatite may be an important phase in the Iranian subcontinental lithospheric mantle (Pang *et al.*, 2013). As was shown in Fig. 15, it is difficult to produce such high LREE concentrations (shown by La in Fig. 15) simply by lower degrees of melting or more garnet in the mantle source. Apatite is soluble in melts with a low

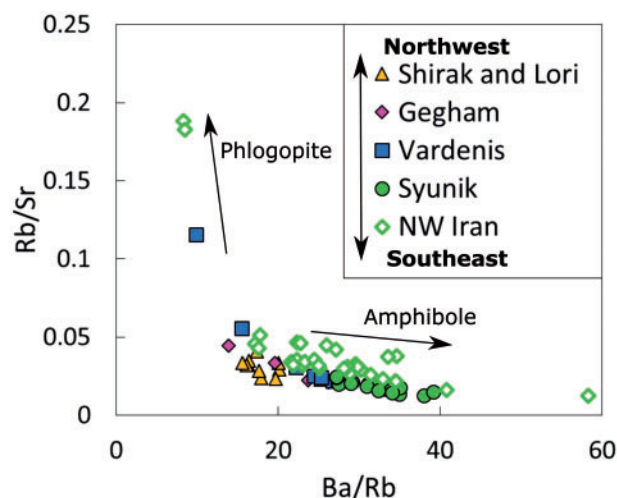


Fig. 21. Rb/Sr vs Ba/Rb for Lesser Caucasus and NW Iran samples, after Furman & Graham (1999). (See text for discussion.)

SiO<sub>2</sub> content (Watson, 1980), such that apatite may simply be added to the initial modelled melt composition, as shown by the blue dashed line in Fig. 15, producing a much improved agreement with natural southern Lesser Caucasus samples.

Although apatite can explain the enrichment in LREE, it cannot explain the large enrichment in other elements such as Ba, which probably requires wider metasomatic reworking. Two minerals that could explain this enrichment in Ba are amphibole and phlogopite. As Fig. 21 shows, the high Ba/Rb ratio favours amphibole as the major metasomatic phase in the source of both the Lesser Caucasus and NW Iran magmas. However, in both the southern Lesser Caucasus and NW Iran there are a minority of samples for which phlogopite could be an important metasomatic phase. It is noteworthy that the presence of phlogopite in the mantle source is indicated in a minority of cases for Vardenis, but not Syunik (Fig. 21), given that biotite is found as a phenocryst phase only in mafic samples from Vardenis (see Petrography section). The presence of these metasomatic signatures in the southern Lesser Caucasus and NW Iran is consistent with melting zone 2 being host to apatite, amphibole, and occasionally phlogopite prior to melting.

Sr/Y is a ratio that does not follow a simple mixing pattern. It increases across the Lesser Caucasus, but is actually lower in NW Iran compared with the southern Lesser Caucasus. This ratio probably reflects the involvement of garnet in the mantle source (Defant & Drummond, 1990), which should become dominant at greater depths. In NW Iran, all melting is occurring in the mid-lithosphere; however, in the southern Lesser Caucasus magmatism is driven by melting in both the mid-lithosphere and at the base of the lithosphere (Fig. 18). This means that in the southern Lesser Caucasus, the average depth of melting may actually be greater than in NW Iran, despite the thinner lithosphere, leading to the mantle source being more dominated by

garnet, increasing the Sr/Y ratio of the resulting magmas.

High Sr/Y ratios have been associated with adakites and the melting of oceanic slabs (Defant & Drummond, 1990). Adakitic signatures have been seen in some volcanic suites from NW Iran (e.g. Ghalamghash *et al.*, 2016), suggesting that the Iranian enriched signature could somehow derive from slab melting. These adakites are generally andesites and dacites. The more primitive NW Iran lavas of Allen *et al.* (2013) and the southern Lesser Caucasus lavas have higher Y contents than true adakites, and it is likely that the adakite-like compositions of andesites and dacites are derived from fractional crystallization processes (e.g. Chiaradia *et al.*, 2009), rather than slab melting.

Melting zone 2 does appear to have a more enriched composition than melting zone 1, as demonstrated by signatures of the metasomatic minerals apatite and amphibole in magmas derived from the southern Lesser Caucasus and NW Iran. However, variations between the chemistry of the northern and southern Lesser Caucasus volcanic rocks do not just reflect mixing of source reservoirs, but also changes in the degree of melting (lower degree of melting will increase La/Yb and Sr/Y ratios), and the mineralogy of the melt source (garnet in the melt source of the southern Lesser Caucasus also increases Sr/Y and La/Yb). The multiple parameters controlling the trace element composition of Lesser Caucasus volcanic rocks mean that variations between the northern Lesser Caucasus, southern Lesser Caucasus and NW Iran are often non-linear.

## SUMMARY AND CONCLUSIONS

Magmas generated in the thicker lithosphere of the southern Lesser Caucasus have higher incompatible trace element concentrations, higher <sup>87</sup>Sr/<sup>86</sup>Sr ratios and lower <sup>143</sup>Nd/<sup>144</sup>Nd ratios than volcanic rocks from the northern Lesser Caucasus. A lack of consistent variation between the isotope compositions of basalts and rhyolites and SiO<sub>2</sub> suggests that crustal contamination is unimportant in generating the enriched geochemistry. The negative Nb–Ta anomalies, and enrichments in large ion lithophile elements and LREE are instead likely to be produced by partial melting of a subduction-modified mantle source. This subduction component is uniform across the Lesser Caucasus and is probably inherited from Mesozoic (Tethyan) slab subduction prior to continental collision. The more enriched geochemistry of southern Lesser Caucasus rocks is the result of lower degrees of melting, an increased proportion of garnet in the mantle source, and also a distinct mantle source composition.

The temperatures of melt formation in the mantle are all less than 1200°C, which when compared with the much higher temperatures for magmas formed in the asthenosphere below nearby Eastern Anatolia suggests that magma generation occurs in the lithosphere, which is also consistent with the ubiquitous subduction

signature. Very similar conditions of melt generation in NW Iran and the southern Lesser Caucasus, as well as several similarities in geochemistry, suggest a common magma generation mechanism in the mid-lithosphere, despite very different lithospheric thicknesses. This magma type appears to mix in the southern Lesser Caucasus with magma from a second source, which probably originates at the base of the lithosphere. This latter magma source is the sole site of magma generation in the northern Lesser Caucasus, where melting occurs at the base of an ~50 km thick lithosphere. In the southern Lesser Caucasus melting occurs at ~75 km depth, significantly shallower than the estimated 120 km thickness of the lithosphere. Melting in the mid-lithosphere occurs as a result of relaxation of a kinked geothermal gradient, whereas melting at the base of the lithosphere is the result of small-scale delamination events. This latter melting mechanism proceeds only until the point where the lithosphere becomes too thick to melt at its base, even if the mantle peridotite is hydrated. The enriched composition of the mid-lithospheric mantle source could be derived from the incorporation of weak lower crust during collision. However, several signatures of the metasomatic minerals amphibole, apatite and occasionally phlogopite suggest that the enriched nature of the mantle source in the mid-lithosphere is derived from the increased retention of metasomatic components in hydrous minerals prior to the post-collisional magmatism.

Interestingly, it appears that a melt source exists at somewhat less than 100 km depth regardless of the lithospheric thickness across the Anatolian–Armenian–Iranian plateau. This is consistent with the geophysical observations of Maggi & Priestley (2005), which show a low shear wave velocity at around 100 km depth below the entire plateau. Further work on understanding the interplay between lithospheric thickness and melt generation in continental collision zones would benefit from detailed tomographic work in the critical region of the Lesser Caucasus to help us better understand how the thickness of the lithosphere varies along this mountain range. Investigations of the petrogenesis of primitive magmatic rocks from NW Iran could elucidate whether thermal relaxation of a kinked geothermal gradient is a viable mechanism to generate magma in the mid-lithosphere. Studies of stable fluid-sensitive isotope systems such as O and B would help decipher the nature and role of inherited subduction components in the generation of collision-related magmas.

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## SUPPLEMENTARY DATA

Supplementary data for this paper are available at *Journal of Petrology* online.

## REFERENCES

- Adam, J. & Green, T. (2006). Trace element partitioning between mica- and amphibole-bearing garnet lherzolite and hydrous basanitic melt: 1. Experimental results and the investigation of controls on partitioning behavior. *Contributions to Mineralogy and Petrology* **152**, 1–17.
- Al-Lazki, A. I. (2003). Tomographic Pn velocity and anisotropy structure beneath the Anatolian plateau (eastern Turkey) and the surrounding regions. *Geophysical Research Letters* **30**, 4–7.
- Allen, M. B., Kheirkhah, M., Neill, I., Emami, M. H. & McLeod, C. L. (2013). Generation of arc and within-plate chemical signatures in collision zone magmatism: Quaternary lavas from Kurdistan province, Iran. *Journal of Petrology* **54**, 887–911.
- Angus, D. A., Wilson, D. C., Sandvol, E. & Ni, J. F. (2006). Lithospheric structure of the Arabian and Eurasian collision zone in eastern Turkey from S-wave receiver functions. *Geophysical Journal International* **166**, 1335–1346.
- Baghdasaryan, G. P. & Ghukasyan, R. K. (1985). Geochronology of magmatic, metamorphic and ore formations of Armenian SSR. Yerevan, Armenia: Academy of Sciences of Armenian Soviet Socialist Republic (in Russian), 291 pp.
- Brounce, M. N., Kelley, K. A. & Cottrell, E. (2014). Variations in  $Fe^{3+}/\Sigma Fe$  of Mariana Arc basalts and mantle wedge  $fO_2$ . *Journal of Petrology* **55**, 2513–2536.
- Bürgmann, R. & Dresen, G. (2008). Rheology of the lower crust and upper mantle: evidence from rock mechanics, geodesy,

- and field observations. *Annual Review of Earth and Planetary Sciences* **36**, 531–567.
- Carmichael, I. S. E., Nicholls, J. T. & Smith, A. L. (1970). Silica activity in igneous rocks. *Contributions to Mineralogy and Petrology* **55**, 246–263.
- Carmichael, I. S. E., Lange, R. A. & Luhr, J. F. (1996). Quaternary minettes and associated volcanic rocks of Mascota, western Mexico: a consequence of plate extension above a subduction modified mantle wedge. *Contributions to Mineralogy and Petrology* **124**, 302–333.
- Chernyshev, I. V., Lebedev, V. A., Arakleyants, M. M., Jrbashyan, R. T. & Ghukasyan, Y. G. (2002). Geochronology of the Aragats volcanic centre, Armenia: evidence from K/Ar dating. *Doklady Earth Sciences* **384**, 393–398 (in Russian).
- Chiaradia, M., Müntener, O., Beate, B. & Fontignie, D. (2009). Adakite-like volcanism of Ecuador: lower crust magmatic evolution and recycling. *Contributions to Mineralogy and Petrology* **158**, 563–588.
- Connor, C., Connor, L., Halama, R., Meliksetian, K. & Savov, I. (2011). Volcanic Hazard Assessment of the Armenia Nuclear Power Plant Site: Final Report, Tampa (USA), Leeds (UK) and Yerevan (Armenia): University of South Florida, Leeds University and GeoRisk, pp. 1–274.
- Cottrell, E. & Kelley, K. A. (2011). The oxidation state of Fe in MORB glasses and the oxygen fugacity of the upper mantle. *Earth and Planetary Science Letters* **305**, 270–282.
- Cox, K. G., Bell, J. D. & Pankhurst, R. J. (1979). *The Classification of Igneous Rocks*. London: Allen & Unwin.
- Davidson, J., Turner, S. & Plank, T. (2013). Dy/Dy\*: variations arising from mantle sources and petrogenetic processes. *Journal of Petrology* **54**, 525–537.
- Defant, M. J. & Drummond, M. S. (1990). Derivation of some modern arc magmas by melting of young subducted lithosphere. *Nature* **347**, 662–665.
- Dixon, J. E., Leist, L., Langmuir, C. & Schilling, J. G. (2002). Recycled dehydrated lithosphere observed in plume-influenced mid-ocean-ridge basalt. *Nature* **420**, 385–389.
- Faccenna, C., Becker, T. W., Jolivet, L. & Keskin, M. (2013). Mantle convection in the Middle East: Reconciling Afar upwelling, Arabia indentation and Aegean trench rollback. *Earth and Planetary Science Letters* **375**, 254–269.
- Frost, D. J. (2006). The stability of hydrous mantle phases. In: Keppler, H. & Smyth, J. R. (eds) *Water in Nominally Anhydrous Minerals*. Mineralogical Society of America and Geochemical Society, *Reviews in Mineralogy and Geochemistry*, **62**, 243–271.
- Furman, T. & Graham, D. (1999). Erosion of lithospheric mantle beneath the East African Rift system: geochemical evidence from the Kivu volcanic province. *Lithos* **48**, 237–262.
- Galoyan, G., Rolland, Y., Sosson, M., Corsini, M. & Melkonyan, R. (2007). Evidence for superposed MORB, oceanic plateau and volcanic arc series in the Lesser Caucasus (Stepanavan, Armenia). *Comptes Rendus Géoscience* **339**, 482–492.
- Galoyan, G., Rolland, Y., Sosson, M., Corsini, M., Billo, S., Verati, C. & Melkonyan, R. (2009). Geology, geochemistry and Ar/Ar dating of Sevan ophiolites (Lesser Caucasus, Armenia): evidence for Jurassic back-arc opening and hot spot event between the South Armenian Block and Eurasia. *Journal of Asian Earth Sciences* **34**, 135–153.
- Garbe-Schönberg, C. D. (1993). Simultaneous determination of thirty-seven trace elements in twenty-eight international rock standards by ICP-MS. *Geostandards and Geoanalytical Research* **17**, 81–97.
- Ghalamghash, J., Mousavi, S. Z., Hassanzadeh, J. & Schmitt, A. K. (2016). Geology, zircon geochronology, and petrogenesis of Sabalan volcano (northwestern Iran). *Journal of Volcanology and Geothermal Research* **327**, 192–207.
- Göğüş, O. H. & Pysklywec, R. N. (2008). Mantle lithosphere delamination driving plateau uplift and synconvergent extension in eastern Anatolia. *Geology* **36**, 723–726.
- Gök, R., Mellors, R. J., Sandvol, E., Pasyanos, M., Hauk, T., Takedatsu, R. & Javakishviri, Z. (2011). Lithospheric velocity structure of the Anatolian plateau–Caucasus–Caspian region. *Journal of Geophysical Research: Solid Earth* **116**, 1–14.
- Green, D. H. & Falloon, T. J. (2005). Primary magmas at mid-ocean ridges, ‘hotspots’, and other intraplate settings: constraints on mantle potential temperature. In: Foulger, G. R., Natland, J. H., Presnall, D. C. & Anderson, D. L. (eds) *Plates, Plumes and Paradigms*. Geological Society of America, *Special Papers* **388**, 217–248.
- Hässig, M., Rolland, Y., Sahakyan, L., Sosson, M., Galoyan, G., Avagyan, A., Bosch, D. & Müller, C. (2015). Multi-stage metamorphism in the South Armenian Block during the Late Jurassic to Early Cretaceous: tectonics over south-dipping subduction of Paleotethys. *Journal of Asian Earth Science* **102**, 4–23.
- Hofmann, A. W. (2003). Sampling mantle heterogeneity through oceanic basalts: isotopes and trace elements. In: Carlson, R. W. (ed.) *The Mantle and Core*. Treatise on Geochemistry, Vol. 2. Amsterdam: Elsevier, pp. 61–102.
- Hosseinpour, M., Williams, S., Seton, M., Barnett-Moore, N. & Müller, R. D. (2016). Tectonic evolution of Western Tethys from Jurassic to present day: coupling geological and geophysical data with seismic tomography models. *International Geology Review* **58**, 1616–1645.
- Hu, F., Ducea, M. N., Liu, S. & Chapman, J. B. (2017). Quantifying crustal thickness in continental collisional belts: global perspective and a geologic application. *Scientific Reports* **7**, 7058.
- Ionov, D. A. & Hofmann, A. W. (1995). Nb–Ta-rich mantle amphiboles and micas: Implications for subduction-related metasomatic trace element fractionations. *Earth and Planetary Science Letters* **131**, 341–356.
- Ionov, D. A., Bodinier, J. L., Mukasa, S. B. & Zanetti, A. (2002). Mechanisms and sources of mantle metasomatism: major and trace element compositions of peridotite xenoliths from Spitsbergen in the context of numerical modelling. *Journal of Petrology* **43**, 2219–2259.
- Jacobsen, S. B. & Wasserburg, G. J. (1980). Sm–Nd isotopic evolution of chondrites. *Earth and Planetary Science Letters* **50**, 139–155.
- Jagoutz, O. & Schmidt, M. W. (2013). The composition of the foundered complement to the continental crust and a re-evaluation of fluxes in arcs. *Earth and Planetary Science Letters* **371**, 177–190.
- Joannin, S., Cornée, J.-J., Münch, P., Fornari, M., Vasiliev, I., Krijgsman, W., Nahapetyan, S., Gabrielyan, I., Ollivier, V., Roiron, P. & Chataigner, C. (2010). Early Pleistocene climate cycles in continental deposits of the Lesser Caucasus of Armenia inferred from palynology, magnetostratigraphy, and <sup>40</sup>Ar/<sup>39</sup>Ar dating. *Earth and Planetary Science Letters* **291**, 149–158.
- Jrbashyan, R. T., Kazarian, G. A., Karapetian, S. G., Meliksetian, K. B., Mnatsakanian, A. & Shirinian, K. G. (1996). Meso-Cenozoic basaltic volcanism in the northeastern part of Armenian Highland. *Letters to the Armenian Academy of Sciences: Earth Sciences* **49**, 19–32 (in Russian).
- Kaislaniemi, L., Van Hunen, J., Allen, M. B. & Neill, I. (2014). Sublithospheric small-scale convection—a mechanism for collision zone magmatism. *Geology* **42**, 291–294.

- Karakhanian, A., Djrbashian, R., Trifonov, V., Philip, H., Arakelian, S. & Avagian, A. (2002). Holocene-historical volcanism and active faults as natural risk factors for Armenia and adjacent countries. *Journal of Volcanology and Geothermal Research* **113**, 319–344.
- Karaođlan, F., Parlak, O., Robertson, A., Thöni, M., Klötzli, U., Koller, F. & Okay, A. İ. (2013). Evidence of Eocene high-temperature/high-pressure metamorphism of ophiolitic rocks and granitoid intrusion related to Neotethyan subduction processes (Dođanşehir area, SE Anatolia). In: Robertson, A. H. F., Parlak, O. & Ünlüođenç, U. C. (eds) *Geological Development of Anatolia and the Easternmost Mediterranean Region*. Geological Society, London, Special Publications **372**, 249–272.
- Karaođlan, F., Parlak, O., Hejl, E., Neubauer, F. & Klötzli, U. (2016). The temporal evolution of the active margin along the South-east Anatolian Orogenic Belt (SE Turkey): evidence from U–Pb, Ar–Ar and fission track chronology. *Gondwana Research* **33**, 190–208.
- Karapetian, S. G., Jrbashian, R. T. & Mnatsakanian, A. K. (2001). Late collision rhyolitic volcanism in the north-eastern part of the Armenian highland. *Journal of Volcanology and Geothermal Research* **112**, 189–220.
- Kelemen, P. B., Hanghøj, K. & Greene, A. R. (2003). One view of the geochemistry of subduction-related magmatic arcs, with an emphasis on primitive andesite and lower crust. In: Holland, H. D. & Turekian, K. K. (eds) *The Crust. Treatise on Geochemistry, Vol. 3*. Amsterdam: Elsevier, pp. 593–658.
- Keskin, M. (2003). Magma generation by slab steepening and breakoff beneath a subduction–accretion complex: an alternative model for collision-related volcanism in Eastern Anatolia, Turkey. *Geophysical Research Letters* **30**, 7–10.
- Keskin, M. (2007). Eastern Anatolia: a hot spot in a collision zone without a mantle plume. In: Foulger, G. R. & Jurdy, D. M. (eds) *Plates, plumes and planetary processes. Geological Society of America Special Paper* **430**, pp. 693–722.
- Keskin, M., Pearce, J. A. & Mitchell, J. G. (1998). Volcano-stratigraphy and geochemistry of collision-related volcanism on the Erzurum–Kars Plateau, northeastern Turkey. *Journal of Volcanology and Geothermal Research* **85**, 355–404.
- Keskin, M., Pearce, J. A., Kempton, P. D. & Greenwood, P. (2006). Magma–crust interactions and magma plumbing in a postcollisional setting: geochemical evidence from the Erzurum–Kars volcanic plateau, eastern Turkey. In: Dilek, Y. & Pavlides, S. (eds) *Postcollisional Tectonics and Magmatism in the Mediterranean Region and Asia. Geological Society of America, Special Papers* **409**, 475–505.
- Kharazyan, E. K. (1983). Geology of recent volcanism of north-west part of Armenian SSR (basins of rivers Dzoraget and Akhuryan). PhD thesis, *ArmGeologia*, Yerevan, 55 pp.
- Kharazyan, E. K. (2005). *Geological Map of Armenia*. Yerevan (Armenia): Ministry of Nature Protection of Republic of Armenia (1 sheet).
- Koulakov, I., Zabelina, I., Amanatashvili, I. & Meskhia, V. (2012). Nature of orogenesis and volcanism in the Caucasus region based on results of regional tomography. *Solid Earth* **3**, 327–337.
- Le Bas, M., Le Maitre, R., Streckeisen, A. & Zanettin, B. (1986). A chemical classification of volcanic rocks based on the total alkali–silica diagram. *Journal of Petrology* **27**, 745–750.
- Lebedev, V. A., Chernyshev, I. V., Shatagin, K. N., Bubnov, S. N. & Yakushev, A. I. (2013). The Quaternary volcanic rocks of the Geghama highland, Lesser Caucasus, Armenia: geochronology, isotopic Sr–Nd characteristics, and origin. *Journal of Volcanology and Seismology* **7**, 204–229.
- Lechmann, A., Burg, J. P., Ulmer, P., Guillong, M. & Faridi, M. (2018). Metasomatized mantle as the source of Mid-Miocene–Quaternary volcanism in NW-Iranian Azerbaijan: geochronological and geochemical evidence. *Lithos* **304**, 311–328.
- Lee, C. T. A., Luffi, P., Plank, T., Dalton, H. & Leeman, W. P. (2009). Constraints on the depths and temperatures of basaltic magma generation on Earth and other terrestrial planets using new thermobarometers for mafic magmas. *Earth and Planetary Science Letters* **279**, 20–33.
- Le Maitre, R. W., Bateman, P., Dudek, A., Keller, J., Lameyre, J., Le Bas, M., Sabine, P., Schmid, R., Sorensen, H. & Streckeisen, A. (1989). *A Classification of Igneous Rocks and Glossary of Terms: Recommendations of the International Union of Geological Sciences Subcommittee on the Systematics of Igneous Rocks*. Oxford: Blackwell Scientific.
- Li, Z. H., Xu, Z. Q. & Gerya, T. V. (2011). Flat versus steep subduction: contrasting modes for the formation and exhumation of high- to ultrahigh-pressure rocks in continental collision zones. *Earth and Planetary Science Letters* **301**, 65–77.
- Luth, R. W. (2003). Mantle volatiles—distribution and consequences. In: Carlson, R. W. (ed.) *The Mantle and Core. Treatise in Geochemistry, Vol. 2*. Amsterdam: Elsevier–Pergamon, pp. 319–361.
- Maggi, A. & Priestley, K. (2005). Surface waveform tomography of the Turkish–Iranian plateau. *Geophysical Journal International* **160**, 1068–1080.
- Mandler, B. E. & Grove, T. L. (2016). Controls on the stability and composition of amphibole in the Earth’s mantle. *Contributions to Mineralogy and Petrology* **171**, 68.
- Mather, K. A., Pearson, D. G., McKenzie, D., Kjarsgaard, B. A. & Priestley, K. (2011). Constraints on the depth and thermal history of cratonic lithosphere from peridotite xenoliths, xenocrysts and seismology. *Lithos* **125**, 729–742.
- McKenzie, D. & O’Nions, R. K. (1991). Partial melt distributions from inversion of rare earth element concentrations. *Journal of Petrology* **32**, 1021–1091.
- McKenzie, D. & Priestley, K. (2008). The influence of lithospheric thickness variations on continental evolution. *Lithos* **102**, 1–11.
- McNab, F., Ball, P. W., Hoggard, M. J. & White, N. J. (2018). Neogene uplift and magmatism of Anatolia: Insights from drainage analysis and basaltic geochemistry. *Geochemistry, Geophysics, Geosystems* **19**, 1–39.
- Mederer, J., Moritz, R., Ulianov, A. & Chiaradia, M. (2013). Middle Jurassic to Cenozoic evolution of arc magmatism during Neotethys subduction and arc–continent collision in the Kapan Zone, southern Armenia. *Lithos* **177**, 61–78.
- Meliksetian, K. (2013). Pliocene–Quaternary volcanism of the Syunik upland. *Archäologie in Armenien* **2**, 247–258.
- Neill, I., Meliksetian, K., Allen, M. B., Navarsardyan, G. & Karapetyan, S. (2013). Pliocene–Quaternary volcanic rocks of NW Armenia: magmatism and lithospheric dynamics within an active orogenic plateau. *Lithos* **180–181**, 200–215.
- Neill, I., Meliksetian, K., Allen, M. B., Navarsardyan, G. & Kuiper, K. (2015). Petrogenesis of mafic collision zone magmatism: the Armenian sector of the Turkish–Iranian Plateau. *Chemical Geology* **403**, 24–41.
- Ollivier, V., Nahapetyan, S., Roiron, P., Gabrielyan, I., Gasparyan, B., Chataigner, C. & Munch, P. (2010). Quaternary volcano-lacustrine patterns and palaeobotanical data in southern Armenia. *Quaternary International* **223**, 312–326.
- Oyan, V., Keskin, M., Lebedev, V. A., Chugaev, A. V., Sharkov, E. V. & Ünal, E. (2017). Petrology and geochemistry of the Quaternary mafic volcanism to the NE of Lake Van, Eastern

- Anatolian Collision Zone, Turkey. *Journal of Petrology* **58**, 1701–1728.
- Ozdemir, Y., Karaoğlu, O., Tolloğlu, A. U. & Güleç, N. (2006). Volcanostratigraphy and petrogenesis of the Nemrut strato-volcano (East Anatolian High Plateau): the most recent post-collisional volcanism in Turkey. *Chemical Geology* **226**, 189–211.
- Pang, K. N., Chung, S. L., Zarrinkoub, M. H., Lin, Y. C., Lee, H. Y., Lo, C. H. & Khatib, M. M. (2013). Iranian ultrapotassic volcanism at ~11 Ma signifies the initiation of post-collisional magmatism in the Arabia–Eurasia collision zone. *Terra Nova* **25**, 405–413.
- Pearce, J. A. (1983). Role of the sub-continental lithosphere in magma genesis at active continental margins. In: Hawkesworth, C. J. & Norry, M. J. (eds) *Continental Basalts and Mantle Xenoliths*. Nantwich: Shiva Press, pp. 230–249.
- Pearce, J. A., Bender, J. F., De Long, S. E., Kidd, W. S. F., Low, P. J., Güner, Y., Saroglu, F., Yilmaz, Y., Moorbath, S. & Mitchell, J. G. (1990). Genesis of collision volcanism in Eastern Anatolia, Turkey. *Journal of Volcanology and Geothermal Research* **44**, 189–229.
- Plank, T. & Forsyth, D. W. (2016). Thermal structure and melting conditions in the mantle beneath the Basin and Range province from seismology and petrology. *Geochemistry, Geophysics, Geosystems* **17**, 1312–1338.
- Prelević, D., Foley, S. F., Romer, R. & Conticelli, S. (2008). Mediterranean Tertiary lamproites derived from multiple source components in postcollisional dynamics. *Geochimica et Cosmochimica Acta* **72**, 2125–2156.
- Prelević, D., Akal, C., Romer, R. L., Mertz-Kraus, R. & Helvacı, C. (2015). Magmatic response to slab tearing: constraints from the Afyon Alkaline Volcanic Complex, Western Turkey. *Journal of Petrology* **56**, 527–562.
- Priestley, K. & McKenzie, D. (2006). The thermal structure of the lithosphere from shear wave velocities. *Earth and Planetary Science Letters* **244**, 285–301.
- Priestley, K. & McKenzie, D. (2013). The relationship between shear wave velocity, temperature, attenuation and viscosity in the shallow part of the mantle. *Earth and Planetary Science Letters* **381**, 78–91.
- Priestley, K., McKenzie, D., Barron, J., Tatar, M. & Debayle, E. (2012). The Zagros Core: deformation of the continental lithospheric mantle. *Geochemistry, Geophysics, Geosystems* **13**, 1–21.
- Robinson, J. A. C. & Wood, B. J. (1998). The depth of the spinel to garnet transition at the peridotite solidus. *Earth and Planetary Science Letters* **164**, 277–284.
- Roeder, P. L. & Emslie, R. F. (1970). Olivine–liquid equilibrium. *Contributions to Mineralogy and Petrology* **29**, 275–289.
- Rolland, Y. (2017). Caucasus collisional history: review of data from East Anatolia to West Iran. *Gondwana Research* **49**, 130–146.
- Rolland, Y., Billo, S., Corsini, M., Sosson, M. & Galoyan, G. (2009). Blueschists of the Amassia–Stepanavan Suture Zone (Armenia): linking Tethys subduction history from E-Turkey to W-Iran. *International Journal of Earth Sciences* **98**, 533–550.
- Rolland, Y., Galoyan, G., Sosson, M., Melkonyan, R. & Avagyan, A. (2010). The Armenian ophiolites: insights for Jurassic back-arc formation, Lower Cretaceous hot-spot magmatism, and Upper Cretaceous obduction over the South Armenian Block. In: Sosson, M., Kaymakci, N., Stephenson, R. A., Bergerat, F. & Starostenko, V. (eds) *Sedimentary Basin Tectonics from the Black Sea and Caucasus to the Arabian Platform*. Geological Society, London, Special Publications **340**, 353–382.
- Rolland, Y., Perincek, D., Kaymakci, N., Sosson, M., Barrier, E. & Avagyan, A. (2012). Evidence for ~80–75 Ma subduction jump during Anatolide–Tauride–Armenian block accretion and ~48 Ma Arabia–Eurasia collision in Lesser Caucasus–East Anatolia. *Journal of Geodynamics* **56–57**, 76–85.
- Rolland, Y., Hässig, M., Bosch, D., Meijers, M. J. M., Sosson, M., Bruguier, O., Adamia, S. & Sadradze, N. (2016). A review of the plate convergence history of the East Anatolia–Transcaucasus region during the Variscan: insights from the Georgian basement and its connection to the Eastern Pontides. *Journal of Geodynamics* **96**, 131–145.
- Rollinson, H. R. (1993). *Using Geochemical Data*. London: Routledge.
- Rudnick, R. L. & Gao, S. (2003). Composition of the continental crust. In: Holland, H. D. & Turekian, K. K. (eds) *The Crust. Treatise on Geochemistry, Vol. 3*. Amsterdam: Elsevier, pp. 659–723.
- Şen, P. A., Temel, A. & Gourgaud, A. (2004). Petrogenetic modelling of Quaternary post-collisional volcanism: a case study of central and eastern Anatolia. *Geological Magazine* **141**, 81–98.
- Sengör, A. M. C. (1984). The Cimmeride orogenic system and the tectonics of Eurasia. In: Geological Society of America, Special Papers **195**, 181–241.
- Sengör, A. M. C., Özeren, S., Genç, T. & Zor, E. (2003). East Anatolian high plateau as a mantle-supported, north–south shortened domal structure. *Geophysical Research Letters* **30**, 8.1–8.4.
- Sengör, A. M. C., Sinan, M., Keskin, M., Sak, M., De, A. & Kayan, İ. (2008). Eastern Turkish high plateau as a small Turkic-type orogen: Implications for post-collisional crust-forming processes in Turkic-type orogens. *Earth-Science Reviews* **90**, 1–48.
- Shaw, D. M. (2005). *Trace Elements in Magmas, a Theoretical Treatment*. Cambridge: Cambridge University Press.
- Sheth, H., Meliksetian, K., Gevorgyan, H., Israyelyan, A. & Navasardyan, G. (2015). Intracanyon basalt lavas of the Debed River (northern Armenia), part of a Pliocene–Pleistocene continental flood basalt province in the South Caucasus. *Journal of Volcanology and Geothermal Research* **295**, 1–15.
- Skolbel'syn, G., Mellors, R., Gök, R., Turkelli, N., Yetirmishli, G. & Sandvol, E. (2014). Upper mantle S wave velocity structure of the East Anatolian–Caucasus region. *Tectonics* **33**, 207–221.
- Sobolev, A. V., Hofmann, A. W., Sobolev, S. V. & Nikogosian, I. K. (2005). An olivine-free mantle source of Hawaiian shield basalts. *Nature* **434**, 590–597.
- Sosson, M., Rolland, Y., Danelian, T., Muller, C., Melkonyan, R., Adamia, S., Kangarli, T., Avagyan, A. & Galoyan, G. (2010). Subductions, obduction and collision in the Lesser Caucasus (Armenia, Azerbaijan, Georgia), new insights. In: Sosson, M. (ed.) *Sedimentary Basin Tectonics from the Black Sea and Caucasus to the Arabian Platform*. Geological Society, London, Special Publications **340**, 329–352.
- Stampfli, G. M., Hochard, C., Vérard, C., Wilhem, C. & vonRaumer, J. (2013). The formation of Pangea. *Tectonophysics* **593**, 1–19.
- Sun, S. S. & McDonough, W. F. (1989). Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and processes. In: Saunders, A. D. & Norry, M. J. (eds) *Magmatism in the Ocean Basins*. Geological Society, London, Special Publications **42**, 313–345.
- Thirlwall, M. F., Upton, B. G. J. & Jenkins, C. (1994). Interaction between continental lithosphere and the Iceland plume:

- Sr–Nd–Pb isotope geochemistry of Tertiary basalts, NE Greenland. *Journal of Petrology* **35**, 839–879.
- Topuz, G., Celik, O. F., Sengor, A. M. C., Altintas, I. E., Zack, T., Rolland, Y. & Barth, M. (2013a). Jurassic ophiolite formation and emplacement as backstop to a subduction–accretion complex in northeast Turkey, the Refahiye ophiolite, and relation to the Balkan ophiolites. *American Journal of Science* **313**, 1054–1087.
- Topuz, G., Göçmengil, G., Rolland, Y., Çelik, Ö. F., Zack, T. & Schmitt, A. K. (2013b). Jurassic accretionary complex and ophiolite from northeast Turkey: no evidence for the Cimmerian continent. *Geology* **41**, 255–258.
- Topuz, G., Okay, A. I., Altherr, R., Schwarz, W. H., Sunal, G. & Altinkaynak, L. (2014). Triassic warm subduction in northeast Turkey: evidence from the Agvanis metamorphic rocks. *Island Arc* **23**, 181–205.
- Toussaint, G., Burov, E. & Avouac, J. P. (2004). Tectonic evolution of a continental collision zone: a thermomechanical numerical model. *Tectonics* **23**, 1–24.
- Turner, S. J., Langmuir, C. H., Dungan, M. A. & Escrig, S. (2017). The importance of mantle wedge heterogeneity to subduction zone magmatism and the origin of EM1. *Earth and Planetary Science Letters* **472**, 216–228.
- Villaseca, C., Downes, H., Pin, C. & Barbero, L. (1999). Nature and composition of the lower continental crust in central Spain and the granulite–granite linkage: inferences from granulitic xenoliths. *Journal of Petrology* **40**, 1465–1496.
- Walsh, J. N., Buckley, F. & Barker, J. (1981). The simultaneous determination of the rare-earth elements in rocks using inductively coupled plasma source spectrometry. *Chemical Geology* **33**, 141–153.
- Watson, E. B. (1980). Apatite and phosphorus in mantle source regions: an experimental study of apatite/melt equilibria at pressures to 25 kbar. *Earth and Planetary Science Letters* **51**, 322–335.
- Weaver, B. L. & Tarney, J. (1984). Empirical approach to estimating the composition of the continental crust. *Nature* **310**, 575–577.
- Weis, D., Kieffer, B., Maerschalk, C., Barling, J., D., Jong, J., Williams, G. A., Hanano, D., Pretorius, W., Mattielli, N., Scoates, J. S., Goolaerts, A., Friedman, R. M. & Mahoney, J. B. (2006). High-precision isotopic characterization of USGS reference materials by TIMS and MC-ICP-MS. *Geochemistry, Geophysics, Geosystems* **7**, 1–30.
- Willett, S., Beaumont, C. & Fullsack, P. (1993). Mechanical model for the tectonics of doubly vergent compressional orogens. *Geology* **21**, 371–374.
- Yilmaz, A., Adamia, S., Chabukiani, A., Chkhotua, T., Erdoğan, K., Tuzcu, S. & Karabiyikoğlu, M. (2000). Structural correlation of the Southern Transcaucasus (Georgia)–Eastern Pontides (Turkey). In: Bozkurt, E., Winchester, J. A. & Piper, J. D. A. (eds) *Tectonics and Magmatism in Turkey and the Surrounding Area*. Geological Society, London, Special Publications **173**, 171–182.
- Zor, E. (2008). Tomographic evidence of slab detachment beneath eastern Turkey and the Caucasus. *Geophysical Journal International* **175**, 1273–1282.