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# Evidence from Late Cretaceous-Paleogene volcanic rocks of the Kyrenia Range, northern Cyprus for the northern, active continental margin of the Southern Neotethys

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# LITHOS

## Evidence from Late Cretaceous-Paleogene volcanic rocks of the Kyrenia Range, northern Cyprus for the northern, active continental margin of the Southern Neotethys --Manuscript Draft--

<b>Manuscript Number:</b>	LITHOS9043R2
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<b>Corresponding Author:</b>	Guohui Chen Institute of Geology and Geophys, Chinese Academy of Sciences Beijing, CHINA
<b>First Author:</b>	Guohui Chen
<b>Order of Authors:</b>	Guohui Chen Alastair Robertson
<b>Abstract:</b>	<p>Late Cretaceous felsic and latest Cretaceous-Paleogene basaltic volcanic rocks are exposed throughout the Kyrenia Range, N Cyprus. Field mapping of the key, well-exposed western range indicates that the felsic volcanics are mainly crop out in the southerly, frontal part of the range, separated from the latest Cretaceous-Paleogene basaltic volcanics farther north by a thrust. New U-Pb zircon dating of the felsic volcanics indicates a primary age of c. 74.0 Ma (Late Campanian). The felsic volcanics are characterised by evolved compositions that were probably generated in an extensional subduction-related setting. Their relatively low Nb and Y but high Rb concentrations, together with characteristic Th/Ta ratios (6-20), suggest a mature continental arc setting. The latest Cretaceous-Paleogene basaltic volcanics have mainly within-plate chemical characteristics (in the east), although some have a subordinate subduction influence; e.g. negative Nb (in the west). Sr-Nd-Hf isotopic signatures (i.e. high positive <math>\epsilon\text{Nd}(t)</math> and <math>\epsilon\text{Hf}(t)</math> values) suggest derivation of the basalts from several OIB-like mantle sources, with probable involvement of a crustal (recycled) component (i.e. elevated Nb/Y ratios). Comparisons with SE Turkey, where Late Cretaceous arc-related granitic rocks are widely developed, suggest that the Kyrenia Range igneous rocks may have originated to the south of the Alanya metamorphic massif (N of Cyprus) and correlative continental units in SE Turkey. The latest Cretaceous-Paleogene basaltic volcanics in N Cyprus and SE Turkey (Maden Complex) are interpreted to represent incipient marginal basin formation, possibly in an oblique-convergent setting, prior to Miocene suturing with Arabia. In the light of alternatives, we infer genesis of the N Cyprus Late Cretaceous and Paleogene volcanic rocks related to stages in the development of the northerly active continental margin of the Southern Neotethys.</p>
<b>Suggested Reviewers:</b>	Fatih Karaođlan fkaraoglan@cu.edu.tr expert on arc magmatism in SE Turkey
	Osman Parlak parlak@cu.edu.tr expert on magmatism and tectonics especially in Turkey
	Pamela Kempton pkempton@ksu.edu
<b>Opposed Reviewers:</b>	

**Evidence from Late Cretaceous-Paleogene volcanic rocks of the Kyrenia Range,  
northern Cyprus for the northern, active continental margin of the Southern  
Neotethys**

Guohui Chen<sup>1\*</sup>, Alastair H. F. Robertson<sup>1</sup>

<sup>1</sup>School of GeoSciences, University of Edinburgh, West Mains Road, Edinburgh EH9  
3JW, UK

\*Current Address: State Key Laboratory of Lithospheric Evolution, Institute of Geology  
and Geophysics, Chinese Academy of Sciences, P.O. Box 9825, Beijing 100029,  
China

Corresponding author: Guohui Chen (Guohui.Chen@live.cn)

## Revision notes:

*Dear Professor Shellnutt,*

*Thank you for considering the further revision of our paper. We have acted on the comments made. We have marked the main edits using track changes. Please note that we not made any significant changes to the results or discussion, only refinements in the light of the review comments.*

*The main further changes of the paper are:*

- 1) The field descriptions are reorganized somewhat, with local description first, and then a summary;*
- 2) The section on the Late Cretaceous magmatic rocks in SE Turkey are revised beginning with units in the south, then the units in the north. More consideration is given to possible equivalents in the Kyrenia Range volcanics (see new section 5.4);*
- 3) More information is given on the HP-LT metamorphism and shoshonites related to Late Cretaceous subduction and possible collision in SE Turkey (lines 492-497) as this is relevant to the regional setting;*
- 4) The Eocene volcanics are discussed in slightly more detail, as suggested (lines 585-593, 653-664).*

*In summary, we hope that you will decide that our paper is now suitable for publication in Lithos.*

*Yours sincerely,  
Guohui Chen and Alastair Robertson*

## COMMENTS FROM EDITORS AND REVIEWERS

Editor: I requested the previous reviewers of your manuscript to evaluate the revised version however, only Reviewer #2 accepted the invitation. As you will see, Reviewer #2 is mostly satisfied with the improvements but, points out a few minor issues in the discussion that should be addressed. Furthermore, I think it is appropriate to move section 2 after section 3. It is common to have the methods after the 'geological background'. Please update the manuscript accordingly.

All the best.  
Greg Shellnutt  
Co-Editor-in-Chief

### Comments

Line 73. The methods should come after the geological background... or in this case the "Volcanism in the Kyrenia Range.

OK. It is reordered now.

Line 75. Delete 'initially'

OK. It is deleted now.

Line 100. Should it be "Whole rock procedural blanks"?

No. It refers the total procedural blanks. So it is correct;

Line 208. Should be either "... and a felsic debris-flow deposit are exposed." Or "... and felsic debris-flow deposits are exposed."

OK. It is revised to be "a felsic debris-flow deposit are exposed" now.

Lines 534-535. Should be, "... arc was explained by northward..."

Yes. Revised as suggested.

Line 565. Should be, "... work is needed..."

Line 670. Should be, "... alternatives are suggested; ..."

OK. They are revised now.

#### Reviewer #2: Dear Editor and authors,

I review the revised MS entitled "Evidence from Late Cretaceous-Paleogene volcanic rocks of the Kyrenia Range, Northern Cyprus for northern, active continental margin of the Southern Neotethys" submitted by Chen and Robertson.

First of all, I want to respond one of the authors' comments in "General Response (7)": "Also, it is normal for this research community to widely cite relevant literature (it is not uncommon in Turkey for aggrieved authors to take cases to university ethics committees if they feel their work has not been adequately cited)."

As a Turkish scientist, this sentence is directly addressing myself and looks trying to provoking and manipulating me to make political review rather than a scientific review. So, I strongly suggest the authors relying on scientific comments. Here all my comments rely on nothing but my scientific vision. The rest is not the authors' business.

I suggest using the subtitle "Results and Discussion" instead "Discussion". The authors present mostly their field and analytical data in sections 5.1 and 5.2.

*The Discussion section, especially sections 5.1 Genetic type/source characteristics and 5.2 Tectonic discrimination, provides further interpretation of the field and analytical data. So we think the subtitle "Discussion" here is OK and in line with use in similar papers.*

The introduction, Methodology and the presentation of the results of the work seem to be alright. However, the discussion is still debatable.

## Discussion

The authors still insisting on linking Kyrenia range to SE Anatolia. It is obvious that these units are continuation of each other in terms of paleogeography, however, their tectonic setting is quite different. The author should explain in detail, possibility of roll-back in a southerly located oceanic crust (Kızıldağ-Troodos oceanic crust) with an Eocene arc.

*More details of the paleogeographic continuation (agreed) are now given, to help clarify the setting of the various tectonic units in SE Tukey. In addition, more detailed comparisons (paleogeographically and geochemically) between the possible equivalents of the Kyrenia Range volcanics are included, together with literature interpretations. Several additional relevant recent references are included (section 5.4). However, we have kept the additions to a necessary minimum to contain the length of this section.*

If the Bitlis-Pütürge (BP) and Tauride platforms collided during late Cretaceous. The HP metamorphism of the BP and the shoshonitic magmatism during late Cretaceous located in Pertek (Elazığ) are addressing a late Cretaceous collision (see results of Oberhänsli et al, 2010, 2014; Sar et al., 2019). After this there should be no Berit Ocean in SE Anatolia and BP attached to Tauride platform acting the active margin of southerly located subduction, which produces the Kızıldağ and Troodos ophiolites.

*The genesis of oceanic crust, including Kızıldağ and Troodos ophiolites, by SSZ spreading is c. 90 Ma (e.g. Mukasa and Ludden, 1987). In contrast, the geochronology indicates a tectono-metamorphic age of Santonian-Campanian for the HP-LT metamorphism of the BP microcontinents (82-80 Ma; Oberhänsli et al., 2010, 2013; Çetinkaplan et al., 2016). Therefore, the younger magmatism is likely to result from an advanced stage of northward subduction of the oceanic plate, although a more southerly origin is possible for the Kyrenia volcanics, see Line 570-577. However, we agree that the Berit ocean probably sutured during the latest Cretaceous, while S Neotethys remained open to the south until the Miocene. We make this clearer.*

The Eocene arc magmatism should be a result of the southerly located subduction and the authors proposal of roll-back during Paleogene rejects the arc magmatism during Eocene.

*The Eocene arc magmatism could be caused by a southerly located subduction given the fact that the Berit meta-ophiolite is unconformably overlain by the Eocene Maden Complex N of the Malatya Metamorphic (Robertson et al., 2006). If the Meydan ophiolite and cross cutting granite are is the same unit (likely), then the Kyrenia Late Cretaceous volcanics should have a different origin,*

*maybe an arc further south. We indicated this before but now we are more explicit (lines 583-593).*

*The proposed model does not reject the arc magmatism during Eocene. The proposed model suggests the Eocene volcanics formed in an oblique-convergent setting, in other words, arc magmatism could possibly take place especially in local regions where remnant oceanic crust still exists. In addition, the Eocene Kyrenia rocks are mainly within-plate in affinity, rather than arc-related.*

If the authors proposal explains the existing of HP metamorphism and shoshonitic magmatism during late Cretaceous and arc magmatism during Eocene, I will be happy to read.

*More information about the HP-LT metamorphism is given in Line 487-497, as requested. The Late Cretaceous shoshonitic magmatism is now mentioned. However, literature review indicates that shoshonitic magmatism does not have a unique tectonic origin but may occur in a range of extensional to post-collisional subduction-related settings (see lines 551-561). Concerning the Eocene arc magmatism, there is geochemical evidence for this in places in SE Turkey but little supporting evidence for a regional-scale arc although as we note this could have subducted (see lines 583-593). However, in agreement with the reviewer, we infer an active margin rather than post-collisional setting for these volcanics in contrast to some other authors.*

Figures&tables look adequate and good in drawing.

My suggestion is accepting the MS with considering the comments above. The text should be spell checked.

Regards,  
Dr. Fatih KARAOĞLAN

Oberhänsli R, Koralay E, Candan O, Pourteau A, Bousquet R (2014) Late Cretaceous eclogitic high-pressure relics in the Bitlis Massif *Geodin Acta* 26:175-190 doi:10.1080/09853111.2013.858951

Oberhänsli R, Candan O, Bousquet R, Rimmele G, Okay A, Goff J (2010) Alpine high pressure evolution of the eastern Bitlis complex, SE Turkey. In: Sosson M, Kaymakci N, Stephenson RA, Bergerat F, Starostenko V (eds) *Sedimentary Basin Tectonics from the Black Sea and Caucasus to the Arabian Platform*, vol 340. vol 1. Geological Society, London, Special Publications, London, pp 461-483. doi:10.1144/sp340.20

Sar A, Ertürk MA, Rizeli ME (2019) Genesis of Late Cretaceous intra-oceanic arc intrusions in the Pertek area of Tunceli Province, eastern Turkey, and implications for the geodynamic evolution of the southern Neo-Tethys: Results of zircon U-Pb geochronology and geochemical and Sr-Nd isotopic analyses Lithos 350-351:105263 doi:10.1016/j.lithos.2019.105263

*These references were included.*



## Abstract

Late Cretaceous felsic and latest Cretaceous-Paleogene basaltic volcanic rocks are exposed throughout the Kyrenia Range, N Cyprus. Field mapping of the key, well-exposed western range indicates that the felsic volcanics are mainly crop out in the southerly, frontal part of the range, separated from the latest Cretaceous-Paleogene basaltic volcanics farther north by a thrust. New U-Pb zircon dating of the felsic volcanics indicates a primary age of c. 74.0 Ma (Late Campanian). The felsic volcanics are characterised by evolved compositions that were probably generated in an extensional subduction-related setting. Their relatively low Nb and Y but high Rb concentrations, together with characteristic Th/Ta ratios (6-20), suggest a mature continental arc setting. The latest Cretaceous-Paleogene basaltic volcanics have mainly within-plate chemical characteristics (in the east), although some have a subordinate subduction influence; e.g. negative Nb (in the west). Sr-Nd-Hf isotopic signatures (i.e. high positive  $\epsilon_{\text{Nd}}(t)$  and  $\epsilon_{\text{Hf}}(t)$  values) suggest derivation of the basalts from several OIB-like mantle sources, with probable involvement of a crustal (recycled) component (i.e. elevated Nb/Y ratios). Comparisons with SE Turkey, where Late Cretaceous arc-related granitic rocks are widely developed, suggest that the Kyrenia Range igneous rocks may have originated to the south of the Alanya metamorphic massif (N of Cyprus) and correlative continental units in SE Turkey. The latest Cretaceous-Paleogene basaltic volcanics in N Cyprus and SE Turkey (Maden Complex) are interpreted to represent incipient marginal basin formation, possibly in an oblique-convergent setting, prior to Miocene suturing with Arabia. In the light of alternatives, we infer genesis of the N Cyprus Late Cretaceous and Paleogene volcanic rocks related to stages in the development of the northerly active continental margin of the Southern Neotethys.

## Highlights

- Felsic volcanics erupted at c. 74 Ma in an active continental margin arc setting
- Some within-plate basalts (c. dated at 70-45 Ma) were affected by subduction
- The basalts are interpreted to represent incipient marginal basin formation
- The volcanism provides new evidence of subduction in the E Mediterranean region
- Northward subduction is supported by coeval arc magmatism in SE Turkey

1 **Evidence from Late Cretaceous-Paleogene volcanic rocks of the Kyrenia Range,**  
2 **northern Cyprus for the northern, active continental margin of the Southern**  
3 **Neotethys**

4 Guohui Chen<sup>1\*</sup>, Alastair H. F. Robertson<sup>1</sup>

5 <sup>1</sup>School of GeoSciences, University of Edinburgh, West Mains Road, Edinburgh EH9  
6 3JW, UK

7 \*Current Address: State Key Laboratory of Lithospheric Evolution, Institute of Geology  
8 and Geophysics, Chinese Academy of Sciences, P.O. Box 9825, Beijing 100029,  
9 China

10

11 Corresponding author: Guohui Chen (Guohui.Chen@live.cn)

12

13 Abstract

14 Late Cretaceous felsic and latest Cretaceous-Paleogene basaltic volcanic rocks are  
15 exposed throughout the Kyrenia Range, N Cyprus. Field mapping of the key, well-  
16 exposed western range indicates that the felsic volcanics are mainly crop out in the  
17 southerly, frontal part of the range, separated from the latest Cretaceous-Paleogene  
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22 concentrations, together with characteristic Th/Ta ratios (6-20), suggest a mature  
23 continental arc setting. The latest Cretaceous-Paleogene basaltic volcanics have  
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26 signatures (i.e. high positive  $\epsilon\text{Nd}(t)$  and  $\epsilon\text{Hf}(t)$  values) suggest derivation of the basalts  
27 from several OIB-like mantle sources, with probable involvement of a crustal (recycled)  
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29 Cretaceous arc-related granitic rocks are widely developed, suggest that the Kyrenia  
30 Range igneous rocks may have originated to the south of the Alanya metamorphic  
31 massif (N of Cyprus) and correlative continental units in SE Turkey. The latest  
32 Cretaceous-Paleogene basaltic volcanics in N Cyprus and SE Turkey (Maden  
33 Complex) are interpreted to represent incipient marginal basin formation, possibly in  
34 an oblique-convergent setting, prior to Miocene suturing with Arabia. In the light of  
35 alternatives, we infer genesis of the N Cyprus Late Cretaceous and Paleogene  
36 volcanic rocks related to stages in the development of the northerly active continental  
37 margin of the Southern Neotethys.

38 Keywords: Late Cretaceous-Paleogene; Volcanic rocks; Kyrenia Range; Arc  
39 magmatism; Marginal basin formation; Southern Neotethys

40

41 1. Introduction

42 Continental margin magmatism is a key feature of subduction at convergent plate  
43 margins (Dewey, 1969; Şengör and Natal'In, 1996). Information concerning the field  
44 relations, age, petrology and geochemistry are essential to understand the growth and  
45 demise of arcs, and are critical to reconstruct paleogeography and to test alternative  
46 tectonic hypotheses (e.g. Trehu et al., 1994; Tatsumi and Kogiso, 2003; Bowman et  
47 al., 2019).

48 The Eastern Mediterranean region, situated between the North African-Arabian  
49 and Anatolian (Eurasian) plates, includes tectonically emplaced remnants of Neotethys  
50 that developed from the preceding Paleotethys farther north (e.g. Le Pichon, 1982;  
51 Şengör et al., 1984; Robertson and Dixon, 1984; Stampfli and Borel, 2002). Much of  
52 the arc magmatism in the north, in the Pontides, of Jurassic-Eocene age (Fig. 1),  
53 relates to long-lived northward subduction of the Northern Neotethys (İzmir-Ankara-  
54 Erzincan ocean) (Adamia et al., 1977; Dercourt et al., 1993; Ustaömer and Robertson,  
55 1997; Rice et al., 2006). Subduction of the Southern Neotethys resulted in both  
56 continental margin and oceanic magmatism that ranges, overall, from Late Cretaceous  
57 to Paleogene (Fig. 1). Here, we focus on evidence of Late Cretaceous and Paleogene  
58 volcanism in the Kyrenia Range of northern Cyprus, representing the most westerly  
59 known occurrence of arc-type magmatism that can be related to subduction of the  
60 Southern Neotethys (Fig. 1) (Moore, 1960; Ducloz, 1972; Pearce, 1975; Baroz, 1979,  
61 1980; Robertson and Woodcock, 1986; Huang et al., 2007; Robertson et al., 2012b).

62 Our specific objectives here are: (1) to understand the tectono-stratigraphy of two,  
63 contrasting, felsic and basaltic volcanic units in relation to associated sedimentary  
64 units; (2) to determine the eruptive ages of both of the volcanic units directly, using  
65 geochronology. Previously, the volcanics were dated indirectly using microfossils  
66 within interbedded pelagic carbonates (Baroz, 1979; Robertson et al., 2012b); (3) to  
67 determine whether there is any petrological or geochemical variation (including

68 isotopic variation) along the Kyrenia Range (c. 100 km) and, if so, the implications; (4)  
69 to infer the magmatic-tectonic setting of eruption, compared with the evidence of arc  
70 volcanism, especially in southeast Turkey; (5) to test alternative tectono-magmatic  
71 hypotheses bearing in mind that arc magmatism plays a key role in plate tectonic  
72 interpretations.

73

## 74 2. Volcanism in the Kyrenia Range

75 Rifting of the Southern Neotethys during Late Permian and Early-Middle Triassic  
76 was followed by continental break-up during the Late Triassic-Early Jurassic  
77 (Robertson and Woodcock, 1979; Şengör and Yılmaz, 1981; Robertson and Dixon,  
78 1984; Garfunkel, 1998; Robertson et al., 2020). The Kyrenia Range then formed part  
79 of a carbonate platform that slowly subsided during Jurassic-Early Cretaceous in a  
80 passive margin setting (Robertson and Woodcock, 1986). The carbonate platform was  
81 deformed and metamorphosed under greenschist facies during the Late Cretaceous,  
82 extensionally exhumed, and then unconformably overlain by basic extrusive igneous  
83 rocks of latest Cretaceous-Paleogene age (Figs. 2-3) (Ducloz, 1972; Baroz, 1979;  
84 Robertson and Woodcock, 1986). In addition, felsic volcanics and tuffs are exposed  
85 as thrust slices at a low structural level in the south of the range (Figs. 2b, 3)  
86 (Robertson et al., 2012b). These volcanic rocks have no exposed base and are directly  
87 overlain by a much larger thrust sheet that includes the mainly basic extrusive igneous  
88 rocks.

89 Baroz (1979, 1980) mapped the volcanic rocks and carried out petrographic and  
90 whole-rock chemical analysis of major elements. He reported the presence of a lower  
91 stratigraphic sequence of basalt, dolerite, trachybasalt, trachyandesite, dacite and  
92 rhyolitic tuff, and interpreted this as a bimodal basic-acidic calc-alkaline suite, related  
93 to a Late Maastrichtian volcanic arc. In contrast, the Paleogene volcanic assemblage

94 was suggested to have erupted stratigraphically above in a post-collisional (intra-  
95 continental) strike-slip setting (Baroz, 1980). However, a thrust was later mapped  
96 between the mainly felsic and mainly basic suites (Robertson et al., 2012b),  
97 complicating this simple stratigraphy.

98 Chemical analysis of the Paleogene basaltic lavas, including immobile trace  
99 elements, initially suggested an alkaline, within-plate eruptive setting (Pearce, 1975).  
100 Subsequent chemical analysis, including some immobile elements, confirmed this  
101 mainly alkaline within-plate setting but also revealed some evidence of a subduction  
102 influence (e.g. negative Nb anomaly) (Robertson and Woodcock, 1986). The  
103 Paleogene lavas were inferred by these authors to have erupted in a transtensional  
104 setting along the northern, active margin of the 'Troodos ocean'. Huang et al. (2007)  
105 carried out additional chemical analysis of the Paleogene basaltic lavas, mainly from  
106 the eastern and western Kyrenia Range, with emphasis on immobile elements, and  
107 proposed a Late Cretaceous-Paleogene back-arc setting related to northward  
108 subduction.

109 The felsic rocks of the lowermost thrust sheet, known as the Fourkovouno  
110 (Selvilitepe)<sup>1</sup> Formation, up to 400 m thick, begin with marine water-lain felsic tuffs and  
111 subaqueous felsic debris-flow deposits and culminate in laterally discontinuous, thick-  
112 bedded to massive rhyolitic lava flows (Fig. 3) (Moore, 1960; Baroz, 1979; Huang et  
113 al., 2007; Robertson et al., 2012b). The felsic rocks are locally cut by small (meter-  
114 sized) basaltic/doleritic intrusions (not studied). A Late Cretaceous age (Late  
115 Campanian) was suggested for the felsic lavas, based on sparse planktic foraminifera  
116 (*Globotruncana* sp.) that were recognised in a chalky interbed near the top of the

---

<sup>1</sup> For simplicity, we use the traditional stratigraphy for the formation names (more recent Turkish equivalents are mentioned initially). However, we use current the Turkish names (e.g. for settlements); pre-existing names are mentioned initially.

117 succession (Baroz, 1979).

118 The Late Cretaceous and Paleogene basaltic igneous rocks have been divided  
119 into two formations, namely the Maastrichtian Melounda (Mallıdağ) Formation and the  
120 Paleocene-Middle Eocene Ayios Nikolaos (Yamaçköy) Formation (Fig. 2a).

121 The Melounda Formation, c. 300 m thick, unconformably overlies the exhumed  
122 Mesozoic platform carbonates of the Trypa (Tripa) Group or, in places,  
123 unmetamorphosed terrigenous turbidites of the Kiparisso Vouno (Alevkaya Tepe)  
124 Member of the Melounda Formation (Baroz, 1979; Robertson and Woodcock, 1986;  
125 Robertson et al., 2012b). The basaltic lavas of the Melounda Formation are commonly  
126 intercalated with pinkish pelagic limestones (Fig. 2a). The succession locally includes  
127 lenticular, lithoclastic breccias and finer-grained clastic facies that were reworked from  
128 the underlying meta-platform carbonates (Fig. 2a) (Robertson and Woodcock, 1986;  
129 Robertson et al., 2012b). Planktic foraminifera from pelagic carbonates between the  
130 lavas indicate a Late Maastrichtian age (Baroz, 1979; Robertson et al., 2012b).

131 The overlying Ayios Nikolaos Formation, 300-400 m thick, comprises basaltic  
132 volcanics that are variably interbedded with pelagic carbonates. Calciturbidites,  
133 calcareous debris-flow deposits and carbonate-rock breccias appear above the lavas,  
134 towards the top of the overall succession (Baroz, 1979; Robertson and Woodcock,  
135 1986; Robertson et al., 2012b, 2014) (Fig. 2a). Planktic foraminifera from pelagic  
136 carbonates that are interbedded with the lavas indicate a Late Paleocene to Mid-  
137 Eocene age range (Baroz, 1979; Robertson et al., 2012b).

138

### 139 3. Methods

140 The Kyrenia Range is a complex thrust belt and therefore it was essential to remap  
141 the western Kyrenia Range, where felsic and basaltic volcanic rocks are exposed  
142 together (Moore, 1960; Baroz, 1979; Robertson et al., 2012b). In addition, sedimentary



143 logs were measured and correlated to produce a composite stratigraphy of the  
144 volcanogenic successions that are exposed in two superimposed thrust sheets.

145 Optical microscopy was carried out on the samples collected. For the felsic  
146 lithologies, 13 samples were studied from the western range. For the basaltic  
147 lithologies, we used a combination of our samples (n=29) and also some that were  
148 previously collected for paleomagnetic study (n=26) throughout the Kyrenia Range  
149 (Morris et al., 2015).

150 Whole-rock major and trace element concentrations of the volcanic rocks were  
151 measured by X-ray fluorescence (XRF) on fused glass beads and pressed powder  
152 pellets at the School of GeoSciences, University of Edinburgh, using the well-known  
153 methods of Fitton et al. (1998) and Fitton and Godard (2004). Accuracy and precision  
154 are typically c. 5%. For representative samples, additional trace and rare earth element  
155 (REE) analysis was carried out by inductively coupled plasma-mass spectrometry  
156 (ICP-MS) at the ACME Analytical Laboratories, Vancouver. Major element contents  
157 were determined from a LiBO<sub>2</sub> fusion by ICP-ES, using 5 g of sample pulp. Trace  
158 element contents were determined from a LiBO<sub>2</sub> fusion by ICP-MS, again using 5 g of  
159 sample pulp. Detection limits range between 0.01 and 0.04 wt% for major oxides, 0.01  
160 and 0.1 ppm for trace and rare earth elements. The relative standard deviation for the  
161 REE is ~5% and for all other trace elements is up to 10%, with quality control using  
162 international geostandards (see <http://acmelab.com>).

163 In addition, Sr-Nd-Hf isotopic analysis was performed on a Neptune Plus multi-  
164 collector (MC)-ICP-MS at Wuhan SampleSolution Analytical Technology Co., Ltd.,  
165 China, as reported by Li et al. (2012). Whole procedural blanks were <100 pg for Sr,  
166 <50 pg for Nd and <50 pg for Hf. <sup>87</sup>Sr/<sup>86</sup>Sr, <sup>143</sup>Nd/<sup>144</sup>Nd and <sup>176</sup>Hf/<sup>177</sup>Hf ratios were  
167 normalised to <sup>86</sup>Sr/<sup>88</sup>Sr = 0.1194, <sup>146</sup>Nd/<sup>144</sup>Nd = 0.7219 and <sup>179</sup>Hf/<sup>177</sup>Hf = 0.7325, using  
168 the exponential law. Standard analysis yielded <sup>87</sup>Sr/<sup>86</sup>Sr = 0.710240 ± 11 (2SD, n = 4)  
169 for NBS987, <sup>143</sup>Nd/<sup>144</sup>Nd = 0.512440 ± 8 (2SD, n = 8) for GSB 04-3258-2015, and

170  $^{176}\text{Hf}/^{177}\text{Hf} = 0.282224 \pm 6$  (2SD,  $n = 6$ ) for Alfa Hf. In addition, USGS reference  
171 materials BCR-2 and RGM-2 were also analysed for Sr–Nd–Hf isotopes, and gave  
172 ratios of  $0.705039 \pm 8$  and  $0.704169 \pm 11$  for  $^{87}\text{Sr}/^{86}\text{Sr}$ ,  $0.512644 \pm 6$  and  $0.512808 \pm$   
173  $9$  for  $^{143}\text{Nd}/^{144}\text{Nd}$  and  $0.282864 \pm 6$  and  $0.283016 \pm 7$  for  $^{176}\text{Hf}/^{177}\text{Hf}$ , which is within  
174 error of recommended values (Thirlwall, 1991; Weis et al., 2006, 2007). The analytical  
175 data for the major, trace and rare earth elements/isotope ratio of the felsic and basaltic  
176 volcanics are listed in the Supplementary Table S1, S2, respectively.

177 In addition, two polished thin sections of basalt (no. 02 and 21) that contain  
178 relatively fresh feldspar and pyroxene crystals were selected for analysis of major  
179 elements using a Cameca SX100 electron microprobe at the School of GeoSciences,  
180 University of Edinburgh. Details of the analytical conditions and methods are given by  
181 Hayward (2011) and by Hartley and Thordarson (2013). The accuracy of major  
182 element determinations is  $<\pm 1\%$  of total value. Analytical data for feldspar and  
183 pyroxene are given in Supplementary Table S3, S4, respectively.

184 For dating, zircon crystals within 30-160  $\mu\text{m}$  fraction were separated from crushed  
185 felsic tuffaceous rock and rhyolitic lava (no. 14-18, 14-19 and 14-20; c. 5 kg each). A  
186 Wilfley table, Frantz Isodynamic magnetic separator and a high-density solution  
187 (lithium polytungstate; 2.85 g/ml) were used to aid separation. The zircon grains were  
188 randomly handpicked under a binocular microscope, mounted in epoxy resin and  
189 polished sufficiently to expose the centre of the grains. Internal structures were studied  
190 with a scanning electron microscope using cathodoluminescence (CL) at the School  
191 of GeoSciences, University of Edinburgh. U-Pb analysis was then performed using a  
192 Cameca IMS-1270 secondary ion mass spectrometer (SIMS) at the School of  
193 GeoSciences, University of Edinburgh, using the methods reported by Kelly et al.  
194 (2008) and Ustaömer et al. (2012). Errors of the reported ages are  $\pm 1\sigma$ . Related  
195 geochronological plots were produced using ISOPLOT (Ludwig, 2012). Analytical  
196 results are listed in the Supplementary Table S5.

197

## 198 4. Results

### 199 4.1. Volcanic successions

200 The felsic volcanics, mainly crop out in the western part of the Kyrenia Range in  
201 two thrust sheets (Figs. 3-4). The lower of these, which is relatively intact, extends from  
202 Geçitköy (Panagra) eastwards to Selvilitepe (Fourkovouno) (Fig. 5a-e). The second,  
203 structurally higher unit, comprises discontinuous exposures extending from north of  
204 Geçitköy, to southwest of Karsiyaka (Vasileia) and west of Alevkaya Tepe (Kiparisso  
205 Vouno) (Fig. 5f-j). Farther west, in the Kayalar (Orga) area (Fig. 3a), a small thrust slice  
206 exposes coarse-grained felsic tuff and a felsic debris-flow deposit. Small (tens of m-  
207 sized) outcrops of felsic volcanogenic rocks also occur farther east; e.g. northeast of  
208 Aşağı Dikmen (Kato Dikomou), between thrust sheets of Mesozoic meta-carbonate  
209 rocks (Baroz, 1979). Small exposures are also intersliced with Late Cretaceous pelagic  
210 carbonates of the Melounda Formation to the north of Ergenekon (Agios Khariton) (Fig.  
211 3a) (Robertson et al., 2012b).

212 The local successions are restored as a lower thrust sheet and an upper thrust  
213 sheet. The succession in the lower thrust sheet, c. 90 m thick, begins with weakly-  
214 bedded, white felsic tuff (beds <0.5 m thick) (Fig. 5a, c-e), followed by thick layers of  
215 matrix-supported felsic breccia-conglomerate (c. 10 m thick) that we interpret as  
216 debris-flow deposits (Fig. 5a, c, e). The succession passes upwards into massive,  
217 vitreous, rhyolitic lava flows, individually up to 10 m thick (Figs. 5a, Supplementary  
218 Figure S1). Localised normal grading; e.g. northeast of Geçitköy, indicates that the  
219 succession is the right way up, stratigraphically (Fig. 5a-b). In places, steeply dipping,  
220 irregularly shaped basaltic or doleritic intrusions, up to 2 m across, cut the felsic  
221 layering (Figs. 5a, Supplementary Figure S1). Adjacent to these intrusions, poorly  
222 consolidated tuffaceous sediments have undergone contact metamorphism to form

223 hard, dark grey, flint-like, recrystallised felsic rock (Robertson et al., 2012b). There is  
224 no evidence of similar intrusions in the structurally overlying basalts, which is in  
225 keeping with the mapped tectonic contact between the two units. The succession in  
226 the upper thrust sheet begins with greyish to greenish, tuffaceous matrix-supported  
227 conglomerates, including clasts of silicified rhyolite, chalk and basalt (Supplementary  
228 Figure S1). The estimated thickness ranges from c. 40 m north of Geçitköy (Fig. 5f),  
229 to c. 80 m southwest of Karsiyaka (Fig. 5g-h), to 30 m west of the Alevkaya Tepe (Fig.  
230 5i-j).

231         Contrasting field relations facilitate mapping of the Late Cretaceous versus the  
232 Paleogene formations (Baroz, 1979; Robertson and Woodcock, 1986; Hakyemez et  
233 al., 2000; Robertson et al., 2012b). In the far east of the Kyrenia Range, the Karpas  
234 Peninsula (Fig. 3a), both pillowed and massive basalt are widely exposed near Balalan  
235 (Platanissos). Interpillow carbonate and lenticular interbeds of pelagic carbonate (< 1  
236 m thick) (Figs. 3a, Supplementary Figure S2) contain Late Maastrichtian microfossils  
237 (Robertson et al., 2012b). In the eastern range, basaltic rocks of the Melounda  
238 Formation contain abundant pelagic carbonate and are overlain by massive basalts of  
239 the Ayios Nikolaos Formation; i.e. from Ağıllar (Mandres) to Mallıdağ (Melounda) (Fig.  
240 3a) (Baroz, 1979; Robertson and Woodcock, 1986; Hakyemez et al., 2000; Robertson  
241 et al., 2012b). In the central range, pillow lava with interbedded pelagic carbonate of  
242 the Melounda Formation occurs in two settings. The first is directly above the Upper  
243 Cretaceous basal unconformity, with the meta-carbonate platform rock beneath; e.g.  
244 near Tirmen (Trypimeni). The second setting is as one (or several), small thrust sheets  
245 along the southern flank of the range; e.g. Değirmenlik (Kythrea), Boğaz (Bogaz) and  
246 southwest of Alevkaya Tepe (Figs. 3a, Supplementary Figure S2). Basalt with some  
247 interbedded pelagic carbonate of the Ayios Nikolaos Formation is exposed higher in  
248 the succession (within a shear zone), near Tirmen, Arapköy (Klepini) and İncesu  
249 (Motides) (Baroz, 1979). In the western range, basaltic lavas interbedded with pelagic

250 carbonates of the Melounda Formation are exposed in a thrust sheet along the  
251 southern margin of the range; e.g. Geçitköy (Figs. 3, Supplementary Figure S2).  
252 Similar basaltic lavas with pelagic carbonates and local carbonate breccias are  
253 exposed along the northern margin of the range; e.g. near Karşiyaka and farther west  
254 (Baroz, 1979; Hakyemez et al., 2000; Robertson et al., 2012b). In addition, basaltic  
255 lava interbedded with pelagic chalk of the Ayios Nikolaos Formation occurs higher in  
256 the succession (southward-younging) along the eastern side of Geçitköy road (Huang  
257 et al., 2007; Robertson et al., 2012b).

258 In summary, the Late Maastrichtian basalts (Melounda Formation) are typically  
259 pillowed, forming repeated flows (individually up to c. 10 m thick), reaching a maximum  
260 of > 90 m in the east (near Balalan). The lavas include pink pelagic carbonate as  
261 interstitial sediment and discontinuous layers. In contrast, the Paleogene basalts  
262 (Ayios Nikolaos Formation) are commonly massive, especially in the eastern range,  
263 where individual lava flows reach c. 45 m thick, with relatively few pelagic intercalations  
264 (Supplementary Figure S2). In the central and western ranges, thinner lava flows  
265 (<10m) are intercalated with pelagic carbonates and reddish chert (formed by  
266 diagenetic replacement). During this work, we identified geochemical differences  
267 between the Late Cretaceous-Paleogene basaltic lavas of similar age along the far-  
268 eastern, eastern, central and western Kyrenia Range (see below).

269

## 270 4.2. Petrography

271 The felsic rocks are porphyritic, composed of quartz, plagioclase and sanidine,  
272 together with subordinate biotite and rare hornblende. Large sanidine crystals (0.3-1  
273 mm) show simple twinning, whereas plagioclase commonly has lamellar twinning  
274 (Supplementary Figure S3). Biotite laths (0.3-0.8 mm in length) are preferentially  
275 orientated parallel to flow layering. Rare quartz phenocrysts (<1%, 0.2-0.5 mm) are

276 commonly fragmented, probably related to quenching during eruption; i.e. explosive  
277 fragmentation (Supplementary Figure S3). The groundmass comprises  
278 microcrystalline quartz and feldspar, together with minor muscovite, biotite and rare  
279 hornblende (c. 60  $\mu\text{m}$ ). In some samples (e.g. nos. 14-19 and 14-20), the groundmass  
280 is cryptocrystalline, almost glassy (Supplementary Figure S3).

281 The basaltic rocks of both Late Cretaceous and Paleogene age generally fall into  
282 two groups in terms of mineral composition and texture. The first group (most common)  
283 is porphyritic with subhedral to euhedral clinopyroxene phenocrysts (30-40%), up to  
284 0.6 mm long, that occur interstitially or are intergrown with subhedral feldspar  
285 (Supplementary Figure S4, Table S3). Feldspar (45-60%) forms elongate, acicular  
286 laths, up to 1 mm long. Olivine is relatively rare (<5%). The second group is ophitic, as  
287 locally observed in the western Kyrenia Range with abundant anhedral to subhedral  
288 augite phenocrysts (up to 0.6 mm in size). Most plagioclase is acicular, enclosed or  
289 surrounded by augite (Supplementary Figure S4).

290 Most samples are slightly to moderately altered. Amygdales and veins are infilled  
291 with secondary minerals such as calcite and zeolite. Alteration is variable, for example,  
292 pyroxene and plagioclase are heavily altered to chlorite and clay minerals (e.g.  
293 smectite/sericite). The relatively high loss on ignition values result from secondary  
294 processes, which need to be taken into account prior to rock identification and tectonic  
295 discrimination.

#### 296 4.3. Whole-rock chemistry

297 Major-element oxides ( $\text{K}_2\text{O}$ ,  $\text{Na}_2\text{O}$  and  $\text{CaO}$ ) and trace elements (e.g. Rb, Sr and  
298 Ba) exhibit a relatively wide compositional range, consistent with the effects of  
299 alteration (see Supplementary Table S1-S2). Alteration can be inferred from variable  
300 LOI values that range between 0.3-5.4 wt% for the felsic volcanics and 4.09-14.57 wt%  
301 for the basaltic rocks. This is consistent with the presence of considerable amounts of

302 water- and/or carbonate dioxide-bearing minerals, for example chlorite and calcite, as  
303 observed petrographically (see Supplementary Figure S3, S4). Trace elements such  
304 as Ti, Zr, Y, Nb, Ta, V, Co, Th and REEs tend to be immobile during weathering and/or  
305 metamorphism below amphibole facies are, therefore, preferred for rock-type  
306 identification and tectonic discrimination (e.g. Pearce and Cann, 1973; Rollinson,  
307 1993). For basaltic rocks, samples with high crystal contents (cumulate composition or  
308 highly porphyritic) should be discounted from geochemical discrimination (Pearce,  
309 1996). In general, samples with immobile element  $\text{Al}_2\text{O}_3 > 20$  wt% (concentrated in  
310 feldspar), Sc  $> 50$  ppm (concentrated in clinopyroxene), or Ni  $> 200$  ppm (concentrated  
311 in olivine) are predicted to contain high amounts of cumulated minerals (Pearce, 1996).  
312 On this basis, basaltic rocks nos. 14-66, 14-50, 14-51, 14-58, 14-68, 08, 16, 25 and 28  
313 are discounted.

314 The more felsic assemblage is mainly rhyolitic in the classification of Winchester  
315 and Floyd (1977) but shows a trachytic affinity in the revised Zr/Ti vs. Nb/Y (Fig. 6a)  
316 diagram of Pearce (1996). Relatively high abundances of Th ( $>14$  ppm) but low Co  
317 concentrations ( $<12$  ppm) are consistent with high-K and shoshonitic affinities (Fig. 6b).  
318 Primitive mantle-normalised multi-element plots (Fig. 6c) show variable enrichments  
319 and depletions in large-ion lithophile elements (LILEs; e.g. Cs, Rb, Ba, K and Sr) that  
320 are influenced by alteration. The samples show negative anomalies of Nb, Zr and Ti,  
321 coupled with marked positive anomalies of Th/U, Pb, Nd and probably Sm (Fig. 6c).  
322 Total REE contents of the felsic assemblage (Fourkovouno Formation) vary from 115  
323 to 173 ppm. La ranges from 29-44 ppm,  $(\text{La}/\text{Yb})_{\text{N}}=12.63-18.98$ , suggestive of an alkalic  
324 affinity. The ratios  $(\text{La}/\text{Sm})_{\text{N}}=6.13-7.22$  and  $(\text{Gd}/\text{Yb})_{\text{N}}=1.26-1.68$  suggest a large LREE  
325 fractionation but low to moderate HREE fractionation (Fig. 6d). Pronounced negative  
326 Eu anomalies ( $\text{Eu}/\text{Eu}^*=0.32-0.48$ ) are consistent with plagioclase crystallization, as  
327 seen petrographically (see Supplementary Figure S3). In general, the samples are  
328 comparable to upper continental crust (UCC). The basaltic rocks plot in the fields of

329 basalt and alkali basalt, with intermediate degrees of fractionation ( $Zr/Ti=0.01-0.02$ )  
330 and moderate alkalinity ( $Nb/Y=0.1-1.8$ ) on the rock-type discrimination plot (Fig. 7a).  
331 The basalts have a calc-alkaline affinity, except for samples from Mallıdağ (no. 01),  
332 Tirmen (no. 19-51), Değirmenlik (no. 14-64) and İncesu (no. 19-67) that exhibit high-K  
333 and shoshonitic affinities (Fig. 7b). The trace element compositions of the basaltic  
334 rocks exhibit wide ranges of Sr, Ba, Rb, Th, Ta, Nb and Ce. However, Zr, Hf, Sm, Ti  
335 and Y generally comparable with enriched mid-ocean ridge basalt (E-MORB) and  
336 oceanic island basalt (OIB) (Fig. 8a-c).

337 The basaltic rocks are characterised by variable REE concentrations (67.2-239.4  
338 ppm) and light REE enrichments ( $(La/Yb)_N=3.18-19.35$ ) (Fig. 8d). Samples from the  
339 central range, at Değirmenlik and Tirmen, show much higher total REE contents and  
340 steeper REE patterns. Westwards in the Kyrenia Range, the patterns become  
341 smoother, with less pronounced enrichment in LREEs. In general, no marked Eu  
342 anomaly ( $\sim 1$ ) is observed.

#### 343 4.4. Whole-rock Sr-Nd-Hf isotopes

344 The selected basaltic samples have relatively uniform whole-rock Sr-Nd-Hf  
345 isotopic characteristics. Adopting a Late Cretaceous eruption age based on the  
346 paleontological dating (Robertson et al., 2012b), the calculated initial  $^{87}Sr/^{86}Sr$  ratios  
347 range from 0.705054 to 0.705701, with  $\epsilon Nd(t)$  values of 2.27 to 5.76. Initial  $^{176}Hf/^{177}Hf$   
348 and  $\epsilon Hf(t)$  values are 0.282934-0.283017, 7.36-10.14, respectively (see  
349 Supplementary Table S2).

#### 350 4.5. Zircon U-Pb geochronology

351 Zircon crystals separated from the felsic rocks are typically angular to sub-angular,  
352 30-160  $\mu m$  long. The grains have variable internal textures, including banded zoning  
353 (e.g. zircon 1), concentric zoning (nos. 2, 4, 5 and 12), and minor sector zoning (nos.



354 10, 11; Fig. 9a). Fractures (or depressions) and some inherited cores are affected by  
355 fluid-related chemical alteration (nos. 6 and 8) (Fig. 9a). The zircon grains have high  
356 Th/U ratios (0.19-1.24), suggestive of a magmatic origin (Rubatto, 2002). Twelve  
357 zircon grains were analysed with concordance levels ranging from 99-105%. The  
358 calculated  $^{206}\text{Pb}/^{238}\text{U}$  age with the generally lowest possible error was used for age  
359 determination. The much older zircon grain (no. 4) is subhedral with a patchy  
360 xenocrystic texture, suggestive of a recycled origin (Fig. 9a). The concordant or nearly  
361 concordant data (100-102%) with a young age distribution yielded weighted mean  
362  $^{206}\text{Pb}/^{238}\text{U}$  gave ages of  $74.0\pm 0.6$  Ma,  $74.0\pm 0.4$  Ma and  $71.7\pm 0.7$  Ma (Fig. 9b-d).

363

## 364 5. Discussion

### 365 5.1. Genetic type/source characteristics

366 For the felsic rocks, the high LREE/HREE (e.g.  $(\text{La}/\text{Yb})_{\text{N}}=12.63-18.98$ ) and  
367 negative Eu anomalies, resemble model UCC but with depletion in Eu, Sr and P (Fig.  
368 6c-d). The A/CNK (molar ratio of  $\text{Al}_2\text{O}_3/(\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O})$ ) ratio is useful to identify the  
369 source of felsic rocks (Chappell and White, 1992) but is questionable for altered rocks  
370 like those of the Kyrenia Range.  $\text{SiO}_2$ ,  $\text{P}_2\text{O}_5$  and Th can also be used as indices of the  
371 fractional crystallization of a felsic melt (Chappell and White, 1992), but again may be  
372 affected by alteration (specially  $\text{SiO}_2$ ). In our samples,  $\text{P}_2\text{O}_5$  decreases with increasing  
373  $\text{SiO}_2$  and Th (Fig. 10a-b), trends that are indicative of I (igneous)- or A (anorogenic)-  
374 type granites. The relatively low contents of Zr, Nb, Y, La, Ce, Zn and Ga further  
375 characterise these rocks as I-type (Fig. 10c-d). The abundances of Th and Co, in  
376 particular hint at a shoshonitic composition (Hastie et al., 2007).

377 The trace element patterns of the basaltic rocks show a wider range of variation  
378 (Fig. 8). E-MORB and OIB-type basalts were derived from relatively enriched sources,

379 specifically for the basalts of the Karpas Peninsula, eastern range and most of the  
380 central range. The lesser enrichment of Nb within the basalts of the central range at  
381 Ergenekon (no. 31), Arapköy (no. 14-62) and Alevkaya Tepe (no. 14-49), and the  
382 western range at Karsiyaka (no. 14) and Geçitköy (no. 10), can be explained by a  
383 relatively depleted mantle source, together with subduction modification. There is little  
384 evidence of the addition of radiogenic Sr from seawater, either to the source melt or  
385 during post-magmatic alteration. The Sr vs. Nd isotope correlation diagram (Fig. 11a)  
386 does not show a significant shift towards higher initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios as would be  
387 expected in both of these cases.

388 The basalts all have positive  $\epsilon\text{Nd}(t)$  values (+2.27 - +5.76), consistent with  
389 derivation from an OIB-like source mantle (Fig. 11a). The samples plot directly on the  
390 terrestrial array, overlapping with the Nd isotope compositions of Etna and the Aeolian  
391 Arc, but intermediate between these settings in Hf isotopes (Fig. 11b). The samples  
392 that show trace element evidence of a subduction-modified source, based on relatively  
393 flat MORB-normalised multi-element patterns and negative Nb anomalies (no. 10,  
394 Geçitköy), plot in the Aeolian Arc-type mantle field. Conversely, samples from the  
395 Karpas Peninsular, at Balalan, plot close to the field of Etna and exhibit less subduction  
396 modification (Fig. 11b). The trace element and isotopic data, therefore, yield consistent  
397 results.

398 The involvement of a crustal component is also suggested by incompatible  
399 element ratio plots, on which the basalts exhibit two distinct trends. The samples with  
400 relatively strong subduction signatures plot away from the MORB array (Fig. 12a),  
401 fingerprinting a subduction-related lithospheric mantle end member. A probable lower  
402 crust input is suggested by relatively higher La/Nb and Ba/Nb ratios (e.g. Wang et al.,  
403 2004). In contrast, an OIB-like mantle component dominated most of the other basalt  
404 samples, as indicated by relatively low La/Nb, Zr/Nb but high Nb/Y ratios (Fig. 12b).  
405 The ratios of highly incompatible elements with small distribution coefficients (e.g.

406 Zr/Nb, Nb/Th, Nb/La) are sensitive indicators of mantle source heterogeneity  
407 (Saunders et al., 1988; Condie, 2003; Pearce and Stern, 2006). The samples from the  
408 western locations fall in the fields of an enriched component and upper continental  
409 crust (Fig. 12c), which could represent mantle mixing with continental sediments  
410 (subducted) or crustal contamination. In contrast, the samples from the easterly  
411 locations resemble those of the OIB/recycled component, and cluster around a  
412 different batch-melting trajectory (10-20%) (Fig. 12c) that is representative of recycled  
413 oceanic lithosphere (Saunders et al., 1988). We, therefore, infer that the basalts are  
414 likely to have been derived from multiple sources (i.e. two or more source components  
415 were involved).

416

## 417 5.2. Tectonic discrimination

418 On the Nb vs. Y diagram (Fig. 13a), the felsic volcanics show similarities with  
419 volcanic arc granites and syn-collisional granites. One sample from Geçitköy (no. 16-  
420 08), with relatively high Nb and Y concentrations, plots near the field of within-plate  
421 granite. On the Th/Yb vs. Ta/Yb diagram (Fig. 13b), the volcanics mainly erupted on  
422 active continental margins. An active continental margin setting is therefore inferred  
423 for the felsic volcanics.

424 Basalts of the Kyrenia Range (e.g. easterly locations) mostly exhibit a within-plate  
425 affinity (Fig. 8). However, some samples, especially in the west, have a subduction-  
426 related signature (e.g. negative Nb anomalies) (Fig. 8b-c) akin to volcanic arc basalts.  
427 Most samples have a within-plate affinity on the Zr/Y vs. Zr (Fig. 13c) diagram, except  
428 for two samples, from Ergenekon and Arapköy (central range), that straddle the island  
429 arc and MORB fields. The basalts overlap the MORB and back-arc basin basalt fields  
430 on the Ti vs. V plot (Fig. 13d). Clinopyroxene compositions, as shown in the ternary  
431  $\text{TiO}_2\text{-MnO-Na}_2\text{O}$  plot (see Supplementary Figure S5), indicate a dominantly within-

432 plate affinity. The relatively high TiO<sub>2</sub> contents of the clinopyroxenes point to an  
433 enriched MORB-related source (see Supplementary Figure S5). Overall, the inferred  
434 subduction contribution decreases from west to east along the range.

435

### 436 5.3. Global comparison of the Kyrenia Range volcanic rocks

437 Global comparisons (see Supplementary Figure S6) lend support to our  
438 interpretation of the Kyrenia Range volcanic rocks as being subduction-related. In  
439 particular, within the Late Cretaceous felsic rocks, the enrichment of Rb, Th and Sm  
440 versus the depletion of Ti, Ba, Nb, P compared to primitive mantle, are characteristic  
441 of volcanic arc granites (Fig. 6c) (Pearce et al., 1984).

442 The Kyrenia Range Maastrichtian and Paleogene basalts of within-plate affinities  
443 (e.g. Balalan, eastern range) mainly have MORB-normalised spider patterns that are  
444 similar to the Pliocene-Quaternary OIB-type basalts of the Tyrrhenian Sea, W  
445 Mediterranean (Peccerillo, 2017). The MORB-normalised patterns of these basalts  
446 suggest a within-plate eruptive setting, with a small sub-crustal lithosphere  
447 metasomatic influence (Kastens et al. 1988; Peccerillo, 2017).

448 Back-arc basalts of the Early Cretaceous Sarmiento Complex, southern Chile  
449 (Saunders et al., 1979) and the Miocene Japan Sea (Tamaki et al., 1990) resemble a  
450 minority of the Kyrenia Range basalts (c. 10%; e.g. Geçitköy, western range), although  
451 with lower abundances of K, Rb, Ba, Th, Ta and Nb. The relative depletion of these  
452 elements is characteristic of eruption in a relatively wide, evolved back-arc basin (e.g.  
453 Bransfield Strait; Weaver et al., 1979). However, another possible setting is eruption  
454 in a back-arc basin that was influenced by oblique extension (e.g. Guaymas pull-apart  
455 basin; Brune et al., 2012). The relatively flat MORB-normalised patterns, akin to E-  
456 MORB, of both the Bransfield Strait and Guaymas back-arc basin basalts (e.g.  
457 Saunders et al., 1982; Keller et al., 2002) point to substantial involvement of depleted

458 MORB sources.

459

#### 460 5.4. Comparison with the adjacent Tethyan region

461 Late Cretaceous and Paleogene arc-related rocks; i.e. similar in age of those in  
462 the Kyrenia Range are critical to any regional magmatic-tectonic interpretation.

463 In western Cyprus, a Late Cretaceous arc-derived volcanogenic succession  
464 (Kannaviou Formation) overlies the Troodos ophiolite. The volcanoclastic sediments  
465 were derived from intermediate-felsic eruptions, c. 6 Ma before the felsic eruptions in  
466 the Kyrenia Range (c. 80 vs. 74 Ma). Grain size and textural evidence suggest that,  
467 depending on prevailing wind direction and strength, the eruptive centers are likely to  
468 have been <100 km away (Chen and Robertson, 2019). The centers for the Kannaviou  
469 Formation were located to the south of the Kyrenia Range or its lateral extension, areas  
470 that are not now exposed. The Kyrenia Range was affected by south-directed thrusting  
471 during Mid-Eocene and Late Miocene-earliest Pliocene (Baroz, 1979; Robertson et al.,  
472 2012b, 2014; McCay and Robertson, 2013; Robertson and Kinnaird, 2016), such that  
473 the distal (southerly) edge of the inferred over-riding active margin (see below) could  
474 have been subducted, structurally over-ridden or eroded. In other words, additional arc  
475 units could have existed to the south of their present location in the Kyrenia Range.

476 To the northeast (c. 200 km), the Misis-Andırın Range in southern Turkey is an  
477 extension of the Kyrenia lineament (Robertson et al., 2004). The outcrop, known as  
478 the Misis-Andırın Complex, includes massive basalts, pillow lavas, lava breccias and  
479 volcanoclastic debris-flows. These lithologies were previously inferred to be Miocene  
480 (Floyd et al., 1991, 1992). However, inter-lava sediments there have been dated as  
481 Campanian-Maastrichtian (Robertson et al., 2004), similar in age to the oldest basaltic  
482 rocks in the Kyrenia Range. The Miocene sediments are structurally intercalated. The  
483 basalts of the Misis-Andırın Complex are tholeiitic, with mildly enriched, subduction-

484 related characteristics (e.g. moderate La/Nb ratios), and are interpreted as of back-arc  
485 basin origin (Floyd et al., 1991, 1992), which is consistent with our interpretation of the  
486 Kyrenia Range basalts (Fig. 14a-b, Supplementary Figure S6).

487 The Kyrenia Range is backed to the north by the Alanya Massif (Fig. 1), which has  
488 been interpreted as a continental fragment, separate from the Tauride carbonate  
489 platform to the north (Robertson et al., 2012a; Çetinkaplan et al., 2016). The Alanya  
490 Massif is represented by the Late Precambrian-Mesozoic, mainly meta-sedimentary  
491 rocks. High pressure-low temperature (HP-LT) metamorphics document Late  
492 Cretaceous (82-80 Ma) subduction/exhumation (Oberhänsli et al., 2010, 2013). This is  
493 inferred to have taken place along the northern, active margin of the Southern  
494 Neotethys (Çetinkaplan et al., 2016; Robertson and Parlak, 2020). The felsic volcanics  
495 (Fourkovouno Formation) in the Kyrenia Range could have erupted along the southern  
496 (trailing) margin of the Alanya microcontinent where they could have escaped  
497 metamorphism. However, they could also represent a separate more southerly unit.

498 Late Cretaceous continental margin arc-type magmatic rocks also occur  
499 extensively farther east, in SE Turkey, as two separate belts.

500 The most southerly, known Late Cretaceous arc rocks in SE Turkey are exposed  
501 in the Helete area, part of a zone of frontal thrust sheets (Yıldırım, 2015; Nurlu et al.,  
502 2016) (Fig. 1). Two thrust assemblages are exposed, both cut by granites that have  
503 been radiometrically dated at 93-83 Ma (Cenomanian-Campanian) (Nurlu et al., 2016).  
504 The lower of the assemblages, the Helete volcanics, comprises arc-related basaltic  
505 andesite, andesite, dacite, rhyolite and common tuffaceous rocks. The upper of the  
506 two thrust assemblages is a dismembered Late Cretaceous supra-subduction zone-  
507 type ophiolite (Meydan ophiolite) (Yıldırım, 2015; Nurlu et al., 2016). The cross-cutting  
508 intrusive rocks sealed the thrust sheets during the Late Cretaceous. The granitic rocks  
509 are interpreted as a calc-alkaline, subduction-related suite that formed by mixing of  
510 mantle and crustal sources, as suggested by Sr-Nd-Pb isotopic data (Nurlu et al.,

511 2016). The granitic rocks exhibit relatively flat chondrite-normalised REE patterns,  
512 similar to E-MORB (Nurlu et al., 2016) (Supplementary Figure S6). In contrast, the  
513 chondrite-normalised REE patterns of the Kyrenia Range felsic rocks are similar to  
514 UCC (Fig. 6c), suggesting a more continentally-influenced setting.

515 The second belt of arc-related magmatism is located up to c. 100 km farther north  
516 and north-east, in the Keban (Malatya), Göksun (Kahramanmaraş) and Baskil (Elazığ)  
517 regions (Fig. 1) (Yazgan and Chessex, 1991; Parlak, 2006; Rızaoğlu et al., 2009;  
518 Karaoğlu et al., 2013, 2016). Arc-granitic rocks (Baskil intrusives) cut the Late  
519 Cretaceous Göksun ophiolite and equivalents (i.e. Berit meta-ophiolite (=North Berit  
520 ophiolite), Kömürhan ophiolite and İspendere ophiolite) that are inferred to have  
521 formed above a northward-dipping subduction zone (Parlak, 2006). Similar arc  
522 granites also cut the structurally over-riding Malatya metamorphic unit, which is  
523 interpreted as the southern part of the Tauride microcontinent. The Göksun ophiolite  
524 (and equivalents) and the Malatya metamorphic (Tauride) units were juxtaposed along  
525 an active continental margin when the arc-granitic rocks were intruded (Robertson et  
526 al., 2006; Rızaoğlu et al., 2009; Karaoğlu et al., 2013, 2016).

527 The extrusive rocks of the Göksun and related ophiolites can also be broadly  
528 correlated with widespread assemblages of basic igneous rocks and volcanogenic  
529 sediments in the Malatya-Elazığ region, known as the Yüksekova Complex (Perincek  
530 and Özkaya, 1981; Aktaş and Robertson, 1984). The basalts have a tholeiitic to  
531 tholeiitic-transitional character, variable Zr/Y (1.5-6) ratios and mantle-array-aligned  
532 Sr-Nd isotopes (Ural et al., 2015). These basic volcanics were derived from a depleted  
533 mantle source without a significant crustal contribution; i.e. from an intra-oceanic  
534 setting within the Southern Neotethys (Ural et al., 2015).

535 The arc rocks that cut both the Göksun (and related) ophiolites and the Malatya  
536 Metamorphics (Baskil intrusives) in the northern belt are mainly hornblende-biotite  
537 granodiorites and 'normal' granites, dated radiometrically at 88-82 Ma (Santonian-

538 Campanian) (Parlak, 2006; Rızaoğlu et al., 2009; Karaoğlan et al., 2016); i.e. up to c.  
539 5 Ma younger than Helete area granites in the south. These southerly granitic rocks  
540 have I-type, calc-alkaline arc affinities, with both mantle and crustal-derivation features  
541 (Nurlu et al., 2016). The Baskil intrusions of the northern belt are relatively enriched in  
542 LREE (Parlak, 2006; Rızaoğlu et al., 2009) compared to the Helete granites of the  
543 south, which suggests a more crustally influenced source during magma genesis for  
544 the former. Similarly, the Kyrenia Range basalts have relatively high Zr/Y (2.8-7.7) and  
545 Nb/Y (Figs. 12b, 13c), suggestive of a magmatic contribution from continental crust or  
546 subducted continentally derived sediments (Zindler and Hart, 1986; Pearce, 1996).

547 The northerly arc has been explained by northward subduction of Mesozoic  
548 oceanic basin (Berit ocean) that was located between two continental units; i.e. the  
549 Bitlis and Pütürge massifs in the south and the Tauride continent represented by the  
550 Malatya Metamorphics in the north (Robertson et al., 2012a; Karaoğlan et al., 2013,  
551 2016; Barrier et al., 2018). The Late Cretaceous arc-related intrusive magmatism of  
552 the northern belt in SE Turkey (Baskil intrusives) evolved towards collisional, including  
553 shoshonitic compositions (74-72 Ma), which are interpreted to indicate continental and  
554 post-collisional settings (Kuşcu et al., 2013; Erturk et al., 2018; Sar et al., 2019). The  
555 shoshonitic compositions of some of the granitic rocks, together with the evidence of  
556 HP/LT metamorphism, are consistent with the collision of the Tauride carbonate  
557 platform (i.e. Malatya Metamorphics) to the north with the Bitlis and Pütürge continental  
558 units to the south. However, the origins of shoshonites remain controversial because  
559 the necessary partial melting of mantle and interaction with subduction-related fluids  
560 can take place in a variety of pre-, syn- and post-collisional settings (e.g. Campbell et  
561 al., 2014).

562 Several authors have correlated the Late Cretaceous Göksun and related  
563 ophiolites in the north with the Meydan ophiolite (Berit meta-ophiolite of Yılmaz et al.,  
564 1993) in the south. If correct, the granitic intrusives in both belts originated as different



565 parts of a single arc lineament that straddled the northerly active continental margin of  
566 the oceanic basin. The southerly arc and ophiolitic rocks (Helete-Meydan) were  
567 emplaced southwards during the latest Cretaceous, although further southward  
568 thrusting took place, mainly during the Eocene and Miocene (Perinçek and Kozlu, 1984;  
569 Yılmaz et al., 1993).

570 In many reconstructions, oceanic crust still remained between the Bitlis-Pütürge  
571 continental units and Arabia until the Miocene (Yılmaz, 1993; Robertson et al., 2012a;  
572 Barrier et al., 2018; van Hinsbergen et al., 2020). The Bitlis and Pütürge massifs have  
573 been generally correlated with the Alanya metamorphic massif to the north of Cyprus  
574 (with or without a direct continuation) (Çetinkaplan et al., 2016). If correct, the Late  
575 Cretaceous Kyrenia Range felsic arc volcanism (and that of the Kannaviou Formation)  
576 could have been located to south of the Alanya continent, with a possible eastward  
577 continuation to south of the Bitlis and Pütürge continental units.

578 Paleogene magmatic rocks are also present in SE Turkey. Eocene volcanic rocks  
579 and associated minor intrusions, known as the Maden Complex, unconformably overlie  
580 and cut the Pütürge and Bitlis metamorphosed continental units (Fig. 1) (Hempton,  
581 1985; Yazgan and Chessex, 1991; Aktaş and Robertson, 1984, 1990; Yılmaz, 1993;  
582 Elmas and Yılmaz, 2003; Robertson et al., 2006; Erturk et al., 2018). The volcanics  
583 show enrichment in LILEs and relative depletion in Nb, Ta and Ti compared to MORB  
584 (Erturk et al., 2018), similar to a minority of the subduction-related volcanics in the  
585 Kyrenia Range (see Supplementary Figure S6). The Eocene Maden Complex  
586 magmatism is widely proposed to relate to back-arc rifting of the (Eurasian) active  
587 continental margin during the later stages of northward subduction of the Southern  
588 Neotethys, in keeping with the widespread view that collision with Arabia was delayed  
589 until the Miocene (Aktaş and Robertson, 1984, 1990; Yılmaz, 1993; Yılmaz et al., 1993;  
590 Yiğitbaş and Yılmaz, 1996; Robertson et al., 2006, 2007; van Hinsbergen et al., 2020).  
591 Oblique convergence, proposed by several authors (Aktaş and Robertson, 1990;

592 Elmas and Yılmaz, 2003), could have resulted in along-strike variation in the  
593 geochemical signatures of extension versus subduction in the Kyrenia Range,  
594 although more geochemical work is needed, regionally and globally to test this  
595 hypothesis.

596 Farther north, Eocene granitic rocks locally cut the continental Malatya  
597 metamorphic unit (Doğanşehir region) (Perinçek and Kozlu, 1984; Karaoğlan et al.,  
598 2013, 2016). However, no extrusive equivalents are known and, as noted above, the  
599 magmatic rocks in this lineament are unlikely to correlate directly with the Kyrenia  
600 Range.

601 During the Miocene, the active continental margin bordering the remnant Southern  
602 Neotethys collided with the Arabian passive continental margin to the south. In both  
603 the Kyrenia Range and SE Turkey, the Late Cretaceous and Paleogene magmatic  
604 rocks are exposed close the thrust front (Eurasia), implying that at least tens of kms of  
605 fore-arc crust have been lost (subducted or overridden), potentially removing additional  
606 frontal arc-related crust, both Late Cretaceous and Paleogene. Thus, the Kyrenia,  
607 Kannaviou (W Cyprus) and Maden Complex magmatism could all represent surviving  
608 products of a regional active continental margin.

609

## 610 5.5. Tectono-magmatic hypotheses

611 In the light of the regional setting, the felsic volcanics (Fourkovouno Formation, c.  
612 74 Ma), with continental margin arc affinities, are interpreted to have resulted from  
613 northward subduction beneath continental crust, possibly the Alanya continental unit  
614 to the north (Çetinkaplan et al., 2016; Robertson et al., 2020). The felsic volcanism is  
615 characteristic of relatively advanced subduction, suggesting that earlier eruptive  
616 centers existed along the active continental margin. The early volcanism is likely to be  
617 represented by the Campanian (c. 80 Ma) arc-derived volcanogenic sediments

618 (Kannaviou Formation) in western Cyprus (Robertson, 1977; Gilbert and Robertson,  
619 2013; Chen and Robertson, 2019). The underlying mantle wedge (sub-continental  
620 lithosphere) was metasomatically enriched during the Late Cretaceous subduction,  
621 with implications for the subsequent basaltic volcanism.

622 For the Maastrichtian to Late Eocene basaltic volcanics, there are three alternative  
623 tectono-magmatic models. The first model envisages a marginal basin setting related  
624 to northward subduction. Compared to typical back-arc basin basalts (Fig. 14a-b), the  
625 Kyrenia Range basalts have relatively high Th/Nb but low La/Nb ratios. The high  
626 abundances of Th are likely to reflect a continental crust influence (Floyd et al., 1991).  
627 The variation in La/Nb ratios of the basalts could reflect the influence of a within-plate  
628 mantle source (i.e. easterly basalts) and/or the arc maturity; i.e. the arc-like western  
629 basalts could have erupted closer to the distal (oceanward) edge of the convergent  
630 margin. However, a problem with this model is the absence of evidence for a  
631 Paleogene volcanic arc to the south. There is no evidence of arc-derived fallout tuff or  
632 volcanoclastic sediments in the Maastrichtian-Paleogene basaltic succession, in  
633 contrast with, for example, the SW Pacific Mariana and Tonga marginal basins (Clift,  
634 1994; Bryan et al., 2004).

635 The second model involves an extensional (or transtensional) setting. In this  
636 interpretation, the Late Maastrichtian and Paleogene 'enriched' within-plate volcanism  
637 was triggered by strike-slip (or transtension) along the active continental margin. One  
638 option is that the volcanism was linked to the anticlockwise paleorotation of the  
639 Troodos ophiolite to the south during Late Campanian-Early Eocene (Clube et al., 1985;  
640 Clube and Robertson, 1986; Morris et al., 2006, 2015; Hodgson et al., 2010). More  
641 generally, oblique convergence of the African plate with respect to Eurasia could have  
642 resulted in segmentation of the South Neotethyan active margin into subducting  
643 segments with arc magmatism, and also strike-slip segments without arc magmatism  
644 (Aktaş and Robertson, 1990). A modern example is the Andaman Sea where oblique

645 convergence between the Asian and the Australian plates resulted in segmented arc  
646 magmatism (e.g. Curray et al., 2005). However, there is limited information from  
647 different modern settings to test this hypothesis.

648 The third, preferred, scenario is that both the felsic and the basaltic volcanism are  
649 related to northward subduction during stages in the closure of the Mesozoic ocean  
650 basin to the south. The Late Cretaceous magmatism can then be explained by regional  
651 northward subduction (nearly orthogonal) resulting in continental margin arc volcanism.  
652 Slab-derived and residual slab components (fluid, melt) were released into the upper  
653 mantle lithosphere (e.g. Pearce, 1983; Harangi et al., 2007). The Late Maastrichtian  
654 and Paleogene basaltic volcanics then erupted in an extensional (or transtensional)  
655 setting, perhaps related to slab rollback and incipient marginal basin formation. Several  
656 different mantle sources are likely to have contributed along the Kyrenia Range. The  
657 variable subduction-related signatures were inherited from the previously modified  
658 mantle lithosphere in this interpretation. Examples of subduction-influenced settings  
659 that were affected by later extension-related melting include the Rio Grande Rift, west  
660 USA (e.g. Riecker, 1979) and the Neogene north Aegean-west Anatolia volcanic  
661 province (e.g. Şengör and Yılmaz, 1981; Ersoy and Palmer, 2013). In summary, the  
662 available evidence is consistent with Late Maastrichtian-Paleogene incipient marginal  
663 basin formation along the northern, active margin of an ocean basin (Southern  
664 Neotethys) that did not close until the Miocene.

#### 665 5.6. Testing alternative tectonic models

666 Finally, the Late Cretaceous-Paleogene volcanism in N Cyprus and SE Turkey  
667 can be used to test alternative plate tectonic models, including those by Robertson et  
668 al. (2012a), Karaoğlan et al. (2013, 2016), Maffione et al. (2017), Barrier et al. (2018),  
669 McPhee and van Hinsbergen (2019) and van Hinsbergen et al. (2020), as summarised  
670 in Figure 15.

671 In Reconstruction 1 (Fig. 15a-c) (Robertson et al., 2012a), the Southern Neotethys  
672 rifted in the Triassic, while Paleotethys evolved into the Northern Neotethys (İzmir-  
673 Ankara-Erzincan ocean). The Kyrenia Range is restored to the northern passive  
674 margin of the Southern Neotethys. Both intra-oceanic and continental margin  
675 subduction zones (northward-dipping) were active in the Southern Neotethys during  
676 the Late Cretaceous. The continental margin arc volcanism was constructed on, or to  
677 the south of one, or more, continental blocks (microcontinents), represented by the  
678 Kyrenia-Alanya-Pütürge-Bitlis lineament in the south and the Malatya-Keban  
679 lineament farther north. This reconstruction is compatible with our new evidence.

680 Reconstruction 2 (Fig. 15d-f) (Barrier et al., 2018) has many common features  
681 with (1), in particular the arc volcanism along the northern margin of the Southern  
682 Neotethys. However, the Kyrenia, Malatya and Pütürge-Bitlis massifs are restored as  
683 a single elongate continental unit rather than as microcontinents, as in (1). Also, little  
684 or no oceanic crust is inferred between the southerly continental unit and the Tauride  
685 carbonate platform to the north (i.e. Göksun and related ophiolites). This is difficult to  
686 reconcile with the Late Cretaceous and Eocene granitic magmatism cutting the  
687 Mesozoic Malatya (Keban) platform unless large-scale strike-slip and terrane  
688 duplication has taken place for which there is little supporting evidence.

689 In Reconstruction 3 (Fig. 15g-h) (Maffione et al., 2017; van Hinsbergen et al.,  
690 2020), the Kyrenia Range is restored to the southern passive margin of the Mesozoic  
691 Tethyan ocean. Late Cretaceous ophiolites that are presently located in several c. E-  
692 W, sub-parallel belts (i.e. N and S of the Bitlis and Pütürge massifs) originated above  
693 a single subduction zone that originated within Neotethys, dipping generally to the east  
694 or northeast (Stampfli and Borel, 2002; Moix et al., 2008). The subduction zone rolled  
695 back generally westwards until it collided with the Tauride and Arabian continents. In  
696 the south, the subduction zone broke through the oceanic gap between the African  
697 and Tauride continents, 'invaded' and replaced the Southern Neotethys with Late

698 Cretaceous supra-subduction zone fore-arc crust. Fore-arc ophiolites were then  
699 obducted generally northwards and southwards over the opposing continents (van  
700 Hinsbergen et al., 2020). Two alternatives are suggested; in one, the Bitlis and Pütürge  
701 continental units are correlated with Arabia (Maffione et al., 2017), whereas in the other  
702 these units are correlated with the southern margin of the Tauride continent (van  
703 Hinsbergen et al., 2020). The Kyrenia Range (Trypa Group) is placed on the N African  
704 continental margin in both alternatives. The metamorphism of the Mesozoic carbonate  
705 platform in the Kyrenia Range (Trypa Group) resulted from northeastward  
706 underthrusting/subduction beneath the obducted forearc Troodos ophiolite (van  
707 Hinsbergen et al., 2020). However, there are several problematic aspects: (1) seismic  
708 tomography in the Eastern Mediterranean supports generally northward subduction  
709 (e.g. Hafkenscheid et al., 2006), without evidence of a relict southward-dipping slab,  
710 as implied by the inferred northward emplacement of some Late Cretaceous ophiolites  
711 in S Turkey; (2) it does not allow for the tens to hundreds of km of subduction required  
712 to produce the Late Cretaceous arc magmatism in N Cyprus and SE Turkey. The  
713 Malatya metamorphic unit, cut by Late Cretaceous granites, is instead restored to the  
714 lower, downgoing plate (i.e. beneath the south margin of the Tauride continent)  
715 implying northward subduction; (3) the model (van Hinsbergen et al., 2020) does not  
716 allow for any Late Cretaceous continental arc-related magmatism in northern Cyprus,  
717 SE Turkey or Iran (e.g. Agard et al., 2011); (4) in this model, the proposed Late  
718 Eocene-Miocene (post 45 Ma) northward subduction is too young to explain the Late  
719 Cretaceous-Paleogene magmatic rocks in the Kyrenia Range and SE Turkey; (5) arc  
720 magmatism of Oligocene-Miocene age in S Turkey is absent, although this would be  
721 expected if significant northward subduction took place during this time, as in model 3.  
722

723 6. Conclusions

724 ● Latest Cretaceous mainly felsic volcanic rocks and latest Cretaceous-Paleogene  
725 mainly basaltic volcanic rocks are exposed throughout the Kyrenia Range,  
726 separated by a thrust.

727 ● Mapping and logging of the key, well-exposed western Kyrenia Range segment  
728 shows that the Late Cretaceous felsic arc volcanics occur as variably  
729 dismembered thrust sheets in the southerly, frontal part of thrust belt.

730 ● The primary eruptive age of the felsic arc volcanics (Fourkovouno Formation) is c.  
731 74.0 Ma (Late Campanian), based on U-Pb dating of zircons.

732 ● The latest Cretaceous-Paleogene basalts vary geochemically along the Kyrenia  
733 Range (E-W). The volcanics in the east resemble normal rift products. In contrast,  
734 the volcanics in the west have a subordinate subduction-related signature (e.g.  
735 negative Nb anomaly).

736 ● Sr-Nd-Hf isotopic data for the latest Cretaceous-Paleogene basalts suggest that  
737 they were derived from several OIB-like mantle sources, with the involvement of a  
738 crustally-derived (recycled) component.

739 ● The Kyrenia Range felsic volcanics show some geochemical similarities with the  
740 late Cretaceous arc-related rocks in S Turkey (Helete granite; 93-83 Ma). These  
741 felsic volcanics are interpreted to represent a fragment of a mature magmatic arc  
742 that was active along a regional-scale active continental margin. Regional  
743 reconstructions suggest that the Helete granites and related ophiolitic rocks were  
744 emplaced southwards from an oceanic basin (Berit ocean) to the north.

745 ● The latest Cretaceous-Paleogene volcanics in the Kyrenia Range have  
746 geochemical similarities with the Middle Eocene (c. 47 Ma) basaltic rocks in S  
747 Turkey (Maden Complex), suggesting an along-strike continuation of the same  
748 active continental margin.

749 ● The latest Cretaceous-Paleogene basaltic volcanics in the Kyrenia Range are  
750 interpreted as the products of an incipient marginal basin that developed during  
751 northward subduction, possibly related to oblique convergence, prior to collision  
752 with Arabia during the Miocene.

753 ● In the light of alternative regional tectonic models, the Late Cretaceous and  
754 Paleogene magmatic rocks of N Cyprus are interpreted to represent stages in the  
755 development of the northerly active continental margin of the Southern Neotethys.

756

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1198 Listing of Figures

1199 Fig. 1 Outline tectonic map of the Eastern Mediterranean region, including the main  
1200 volcanic units mentioned in the text. The tectonic framework is modified after  
1201 Robertson et al. (2012a). Note the location of the Kyrenia Range in northern Cyprus  
1202 (red box). Jurassic, Cretaceous and Eocene igneous rocks are shown in blue, red and  
1203 yellow, respectively (see the text for literature sources).

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1205 northern Cyprus. (a) Simplified log of the succession in the largest thrust sheet  
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1209 Robertson et al., 2012b and this study).

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1213 Ağıllar (Mandres), 3-Çınarlı (Platani), 4-Mallıdağ (Melounda); Central Range: 5-Tirmen  
1214 (Trypimeni), 6-Ergenekon (Agios Khariton), 7-Değirmenlik (Kythrea), 8-Arapköy  
1215 (Klepini), 9-Beylerbeyi (Bellapais), 10-Boğaz (Bogaz), 11-Pınarbaşı (Krini), 12-İncesu  
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1226 rocks.

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1236 crust (UCC) and lower continental crust (LCC) data are from Rudnick and Gao (2003),  
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1238 (1984).

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1240 Hastie et al., 2007) for the basaltic volcanics. Red symbols: Karpas Peninsula and  
1241 eastern range; green symbols: central range; blue symbols: western range. Paleogene  
1242 basalt samples are indicated by thick magenta-outlined symbols.

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1258 M (depleted mantle source)-type nature of the Late Cretaceous felsic volcanics.

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1261 magenta-outlined symbols. Comparative compositions are from Zindler and Hart  
1262 (1986); (b)  $\epsilon\text{Hf}(t)$  vs.  $\epsilon\text{Nd}(t)$  plot showing rocks for Kyrenia Range, northern Cyprus  
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1264 2018).

1265 Fig. 12 (a) Ba/Nb versus La/Nb diagram (after Zhang et al., 2013); (b) Nb/Y versus  
1266 Zr/Y diagram (after Condie, 2005), arrows in (b) indicate the effects of batch melting  
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1268 diagram (modified after Condie, 2003). Abbreviations: PM, primitive mantle; DM,  
1269 shallow depleted mantle; HIMU, high  $\mu$  (U/Pb) source; EMI and EMII, enriched  
1270 mantle sources; ARC, arc-related basalt; N-MORB, normal ocean ridge basalt; OIB,  
1271 oceanic island basalt; DEP, deep depleted mantle; EN, enriched component; REC,  
1272 recycled component; UCC, upper upper continental crust; LCC, lower continental crust;  
1273 BM, batch melting trajectory with percent melting noted. Numbers on mixing lines are  
1274 percents.

1275 Fig. 13 (a) Nb vs. Y diagram (Pearce et al., 1984) and (b) Th/Yb vs. Ta/Yb (Gorton and  
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1277 1979) and (d) V vs. Ti discrimination diagram (after Shervais, 1982) for the basaltic

1278 rocks. Abbreviations: WPG, within-plate granite; VAG, volcanic arc granite; syn-COLG,  
1279 syn-collisional granite; ORG, ocean ridge granite; WPB, within-plate basalt; MORB,  
1280 mid-ocean ridge basalt; BABB, back-arc basin basalt; ARC, arc-related basalt; OFB,  
1281 oceanic floor basalt.

1282 Fig. 14 (a) Th/Nb vs. La/Nb and (b) La/Nb versus Y diagrams (after Floyd et al., 1991).  
1283 Abbreviations: IAT, island arc tholeiite; BABB, back-arc basin basalt; OFB, oceanic  
1284 flood basalt; FAPB, fore-arc platform basalt; T/E-MORB, T-type/enriched-mid-ocean  
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1286 Fig. 15 Alternative tectonic models of Tethys in the Eastern Mediterranean region. (a-  
1287 c) Northward subduction with continental fragments rifted from Gondwana (Robertson  
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1289 2018); (g-h) Genesis of Late Cretaceous supra-subduction ophiolites at a single  
1290 subduction zone to the NE followed by roll-back of segments including the Southern  
1291 Neotethys (Maffione et al., 2017; McPhee and van Hinsbergen, 2019; van Hinsbergen  
1292 et al., 2020). Based on discussion (see text), model 1 is generally favoured.

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1294 Listing of Supplementary Figure

1295 Supplementary Figure S1 Field photographs of the felsic volcanics. (a) Exposure in  
1296 Geçitköy. Samples 14-18 and 14-19 are from this location; (b) Irregular-shaped  
1297 doleritic intrusion (hammer for scale); (c) Localised occurrence of greyish to greenish,  
1298 rhyolitic debris-flow unit, north of Geçitköy; (d) Greenish rhyolitic debris-flow unit with  
1299 angular rhyolitic clasts (pen for scale).

1300 Supplementary Figure S2 Field photographs of the basaltic volcanics. (a) Pillow lavas  
1301 and interstitial pink pelagic carbonate; roadcut c. 200 m north of Balalan; (b) basalt  
1302 interbedded with pelagic carbonate; thrust sheet above Miocene siliciclastic sediment;  
1303 c. 800 m northwest of Değirmenlik; (c) Basaltic volcanics intercalated with pelagic  
1304 carbonates and small slices of meta-platform carbonates of the Trypa Group along the  
1305 southern flank of the range, 800 m north of Boğaz; (d) Basalt-pelagic carbonate  
1306 intercalation, south of Karşıyaka.

1307 Supplementary Figure S3 Photomicrographs of felsic volcanics. (a) Subhedral  
1308 sanidine (Sa), plagioclase (Pl) and biotite (Bt), in a groundmass of devitrified glassy  
1309 volcanic shards (cross-polarised light); (b) felsic glass groundmass (Vg) with  
1310 microcrystalline quartz, irregular-shaped quartz (Q) and feldspar (Pl) (cross-polarised  
1311 light). Sample number is indicated in the bottom-left corner.

1312 Supplementary Figure S4 Photomicrographs of basalts. (a) Intersertal basalt with  
1313 euhedral granular augite (Px), elongate plagioclase laths (Pl) and opaque grains (Op)  
1314 (cross-polarised light); (b) Basalt with intersertal texture, in which randomly-oriented  
1315 plagioclase generally enclose pyroxenes (plane-polarised light); (c) Porphyritic basalt  
1316 with euhedral plagioclase and strongly altered olivine phenocrysts (cross-polarised  
1317 light); (d) Ophitic basalt with plagioclase phenocrysts, augite occurs as an inclusion  
1318 within individual plagioclase crystals (cross-polarised light). Sample number is  
1319 indicated in the bottom-left corner. Samples no. 21, 14-51 and 14-69 are from the  
1320 Melounda Formation, whereas sample no. 19-51 is from the Ayios Nikolaos Formation.

1321 Supplementary Figure S5 Tectonic discrimination plots. (a)  $\text{TiO}_2$ - $\text{MnO}$ - $\text{Na}_2\text{O}$  ternary  
1322 diagram (after Nisbet and Pearce, 1977) for clinopyroxenes in basalt; (b)  $\text{TiO}_2$ - $\text{SiO}_2$ -  
1323  $\text{Na}_2\text{O}$  diagram (after Beccaluva et al., 1989) for clinopyroxenes in basalt. Abbreviations:  
1324 VAB, volcanic arc basalt; OFB, ocean-floor basalt; WPA, within-plate alkali basalt;  
1325 WPT, within-plate tholeiitic basalt; BON, boninite; IAT, island-arc tholeiite; WPB,  
1326 within-plate basalt.

1327 Supplementary Figure S6 (a) Chondrite-normalised REE comparison patterns of the  
1328 Kyrenia Range felsic rocks and the Helete granites in SE Turkey (data after Nurlu et  
1329 al., 2016). Chondrite data after Sun and McDonough (1989); (b)-(c) MORB-normalised  
1330 multi-element spider diagrams for basalts. For comparison, the coloured fields show  
1331 the composition variations of basalts from the Kyrenia Range (this study), the Misis  
1332 Complex, S Turkey (Floyd et al., 1991), the Maden Complex, S Turkey (Erturk et al.,  
1333 2018), the Japan Sea (Chen et al., 2015), the Sarmiento Complex, Chile (Saunders et  
1334 al., 1979), the Bransfield Strait, Antarctic (Keller et al., 2002), the Guaymas Basin,  
1335 Caribbean (Saunders et al., 1982) and the Tyrrhenian Sea, Italy (Peccerillo, 2017).  
1336 MORB-normalised data are from Pearce et al. (1983). N-MORB, E-MORB and OIB  
1337 data are from Sun and McDonough (1989). Specifically, two representative basalts of  
1338 different affinities (within-plate vs. volcanic arc) from the Kyrenia Range are indicated.

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1348 Listing of Supplementary Tables

1349 Supplementary Table S1. Major element oxides, trace and rare earth elements for the  
1350 Late Cretaceous Fourkovouno (Selvilitepe) Formation.

1351 Supplementary Table S2. Major element oxides, trace and rare earth element analyses  
1352 for the latest Cretaceous-Paleogene basaltic volcanics.

1353 Supplementary Table S3. Electron microprobe analyses of feldspar in the latest  
1354 Cretaceous-Paleogene basaltic volcanics.

1355 Supplementary Table S4. Electron microprobe analyses of pyroxene in the latest  
1356 Cretaceous-Paleogene basaltic volcanics.

1357 Supplementary Table S5. SIMS zircon U-Pb analyses of the zircon grains separated  
1358 from the Late Cretaceous Fourkovouno (Selvilitepe) Formation.



1 **Evidence from Late Cretaceous-Paleogene volcanic rocks of the Kyrenia Range,**  
2 **northern Cyprus for the northern, active continental margin of the Southern**  
3 **Neotethys**

4 Guohui Chen<sup>1\*</sup>, Alastair H. F. Robertson<sup>1</sup>

5 <sup>1</sup>School of GeoSciences, University of Edinburgh, West Mains Road, Edinburgh EH9  
6 3JW, UK

7 \*Current Address: State Key Laboratory of Lithospheric Evolution, Institute of Geology  
8 and Geophysics, Chinese Academy of Sciences, P.O. Box 9825, Beijing 100029,  
9 China

10

11 Corresponding author: Guohui Chen (Guohui.Chen@live.cn)

12

13 Abstract

14 Late Cretaceous felsic and latest Cretaceous-Paleogene basaltic volcanic rocks are  
15 exposed throughout the Kyrenia Range, N Cyprus. Field mapping of the key, well-  
16 exposed western range indicates that the felsic volcanics are mainly ~~crop outexposed~~  
17 in the southerly, frontal part of the range, separated from the latest Cretaceous-  
18 Paleogene basaltic volcanics farther north by a thrust. New U-Pb zircon dating of the  
19 felsic volcanics indicates a primary age of c. 74.0 Ma (Late Campanian). The felsic  
20 volcanics are characterised by evolved ~~high potassium and shoshonitic~~ compositions  
21 ~~that were probably generated in an extensional subduction-related setting.~~ Their  
22 relatively low Nb ~~and~~ Y but high Rb concentrations, together with characteristic Th/Ta  
23 ratios (6-20), suggest a mature continental arc setting. The latest Cretaceous-  
24 Paleogene basaltic volcanics have mainly within-plate chemical characteristics (in the  
25 east), although some have a subordinate subduction influence; e.g. negative Nb (in

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26 the west). Sr-Nd-Hf isotopic signatures (i.e. high positive  $\epsilon\text{Nd}(t)$  and  $\epsilon\text{Hf}(t)$  values)  
27 suggest derivation of the basalts from several OIB-like mantle sources, with probable  
28 involvement of a crustal (recycled) component (i.e. elevated Nb/Y ratios). Comparisons  
29 with SE Turkey, where Late Cretaceous arc-related granitic rocks are widely developed,  
30 suggest that the Kyrenia Range igneous rocks may have originated to the south of the  
31 Alanya metamorphic massif (N of Cyprus) and correlative continental units in SE  
32 Turkey. T-The Kyrenia Range Late Cretaceous felsic volcanics are comparable with  
33 rare Late Cretaceous granitic rocks in the front of the over-riding Tauride thrust belt in  
34 SE Turkey (Holote unit). They are also comparable with Late Cretaceous arc-related  
35 magmatic rocks higher in the regional tectono-stratigraphy in SE Turkey, although  
36 these may relate to a different continental margin arc. We propose that the Late  
37 Cretaceous felsic volcanics in N Cyprus record mature arc magmatism, as in SE  
38 Turkey. In contrast, the latest Cretaceous-Paleogene basaltic volcanics in N Cyprus  
39 and SE Turkey (Maden Complex) are interpreted to represent incipient marginal basin  
40 formation, possibly in an oblique-convergent setting, prior to Miocene suturing with  
41 Arabia. In the light of alternatives, we infer genesis of the N Cyprus Late Cretaceous  
42 and Paleogene volcanic rocks related to stages in the development of the northerly  
43 active continental margin of the Southern Neotethys.

44 Keywords: Late Cretaceous-Paleogene; Volcanic rocks; Kyrenia Range; Arc  
45 magmatism; Marginal basin formation; Southern Neotethys

46

47 1. Introduction

48 Continental margin magmatism is a key feature of subduction at convergent plate  
49 margins (Dewey, 1969; Şengör and Natal'In, 1996). Information concerning the field  
50 relations, age, petrology and geochemistry are essential to understand the growth and  
51 demise of arcs, and are critical to reconstruct paleogeography and to test alternative  
52 tectonic hypotheses (e.g. Trehu et al., 1994; Tatsumi and Kogiso, 2003; Bowman et  
53 al., 2019).

54 The Eastern Mediterranean region, situated between the North African-Arabian  
55 and Anatolian (Eurasian) plates, includes tectonically emplaced remnants of Neotethys  
56 that developed from the preceding, ~~larger~~ Paleotethys farther north (e.g. Le Pichon,  
57 1982; Şengör et al., 1984; Robertson and Dixon, 1984; Stampfli and Borel, 2002).  
58 Much of the arc magmatism in the north, in the Pontides, of Jurassic-Eocene age (Fig.  
59 1), relates to long-lived northward subduction of the Northern Neotethys (İzmir-Ankara-  
60 Erzincan ocean) (Adamia et al., 1977; Dercourt et al., 1993; Ustaömer and Robertson,  
61 1997; Rice et al., 2006). Subduction of the Southern Neotethys ~~also took place during~~  
62 ~~the Late Cretaceous (95-90 Ma)~~, resulted in both continental margin and oceanic  
63 magmatism that ranges, overall, from Late Cretaceous to Paleogene (Fig. 1). Here, we  
64 focus on evidence of Late Cretaceous and Paleogene volcanism in the Kyrenia Range  
65 of northern Cyprus, representing the most westerly known occurrence of arc-type  
66 magmatism that can be related to subduction of the Southern Neotethys (Fig. 1)  
67 (Moore, 1960; Ducloz, 1972; Pearce, 1975; Baroz, 1979, 1980; Robertson and  
68 Woodcock, 1986; Huang et al., 2007; Robertson et al., 2012b).

69 Our specific objectives here are: (1) to understand the tectono-stratigraphy of two,  
70 contrasting, felsic and basaltic volcanic units in relation to associated sedimentary  
71 units; (2) to determine ~~directly~~ the eruptive ages of both of the volcanic units directly,  
72 using geochronology. ~~Previously, the volcanics~~ ~~These~~ ~~were~~ ~~previously~~ dated indirectly  
73 using microfossils within interbedded pelagic carbonates (Baroz, 1979; Robertson et

74 al., 2012b); (3) to ~~investigate—determine whether there is any~~ petrological or  
75 geochemical variation ~~(including isotopic variation)~~ along the Kyrenia Range (c. 100  
76 km) ~~and, if so, the implications~~; (4) to infer the magmatic-tectonic setting of eruption,  
77 compared with the evidence of arc volcanism ~~globally, \_and~~ especially in southeast  
78 Turkey; (5) to test alternative tectono-magmatic hypotheses bearing in mind that arc  
79 magmatism ~~plays has~~ a key role in plate tectonic ~~interpretationss~~.

80

## 81 ~~2. Methods~~

82 ~~The Kyrenia Range is a complex thrust belt, and therefore it was essential initially,~~  
83 ~~to remap the western Kyrenia Range, where felsic and basaltic volcanic rocks are~~  
84 ~~exposed together (Moore, 1960; Baroz, 1979; Robertson et al., 2012b). In addition,~~  
85 ~~sedimentary logs were measured and correlated to produce a composite stratigraphy~~  
86 ~~of the volcanogenic successions that are exposed in two superimposed thrust sheets.~~

87 ~~Optical microscopy was carried out on the samples collected. For the felsic~~  
88 ~~lithologies, 13 samples were studied from the western range. For the basaltic~~  
89 ~~lithologies, we used a combination of new samples (n=29) and also samples that were~~  
90 ~~previously collected for paleomagnetic study (n=26) (Morris et al., 2015) throughout~~  
91 ~~the Kyrenia Range.~~

92 ~~Whole-rock major and trace element concentrations of the volcanic rocks were~~  
93 ~~measured by X-ray fluorescence (XRF) on fused glass beads and pressed powder~~  
94 ~~pellets at the School of GeoSciences, University of Edinburgh, using the well-known~~  
95 ~~methods of Fitton et al. (1998) and Fitton and Godard (2004). Accuracy and precision~~  
96 ~~are typically c. 5%. For representative samples, additional trace and rare earth element~~  
97 ~~(REE) analysis was carried out by inductively coupled plasma mass spectrometry~~  
98 ~~(ICP-MS) at the ACME Analytical Laboratories, Vancouver. Major element contents~~  
99 ~~were determined from a LiBO<sub>2</sub> fusion by ICP-ES by using 5 g of sample pulp. Trace~~

100 element contents were determined from a LiBO<sub>2</sub> fusion by ICP-MS by using 5 g of  
101 sample pulp. Detection limits range between 0.01 and 0.04 wt% for major oxides, 0.01  
102 and 0.1 ppm for trace and rare earth elements. The relative standard deviation for the  
103 REE is 5% and for all other trace elements is up to 10%, with quality control using  
104 international geostandards (see <http://acmelab.com>).

105 In addition, Sr-Nd-Hf isotopic analysis was performed on a Neptune Plus multi-collector  
106 (MC)-ICP-MS at Wuhan SampleSolution Analytical Technology Co., Ltd., China, as  
107 reported by Li et al. (2012). Whole procedural blanks were <100 pg for Sr, <50 pg for  
108 Nd and <50 pg for Hf. <sup>87</sup>Sr/<sup>86</sup>Sr, <sup>143</sup>Nd/<sup>144</sup>Nd and <sup>176</sup>Hf/<sup>177</sup>Hf ratios were normalised to  
109 <sup>86</sup>Sr/<sup>88</sup>Sr = 0.1194, <sup>146</sup>Nd/<sup>144</sup>Nd = 0.7219 and <sup>179</sup>Hf/<sup>177</sup>Hf = 0.7325, using the  
110 exponential law. Standard analysis yielded <sup>87</sup>Sr/<sup>86</sup>Sr = 0.710240 ± 11 (2SD, n = 4) for  
111 NBS987, <sup>143</sup>Nd/<sup>144</sup>Nd = 0.512440 ± 8 (2SD, n = 8) for GSB 04-3258-2015, and  
112 <sup>176</sup>Hf/<sup>177</sup>Hf = 0.282224 ± 6 (2SD, n = 6) for Alfa-Hf. In addition, USGS reference  
113 materials BCR-2 and RGM-2 were also analysed for Sr-Nd-Hf isotopes, and gave  
114 ratios of 0.705039 ± 8 and 0.704169 ± 11 for <sup>87</sup>Sr/<sup>86</sup>Sr, 0.512644 ± 6 and 0.512808 ±  
115 9 for <sup>143</sup>Nd/<sup>144</sup>Nd and 0.282864 ± 6 and 0.283016 ± 7 for <sup>176</sup>Hf/<sup>177</sup>Hf, which is within  
116 error of recommended values (Thirlwall, 1991; Weis et al., 2006, 2007). The analytical  
117 data for the major, trace and rare earth elements/isotope ratio of the felsic and basaltic  
118 volcanics are listed in the Supplementary Table S1, S2, respectively.

119 Furthermore, two polished thin sections of basalt (no. 02 and 21) that contain  
120 relatively fresh feldspar and pyroxene crystals were selected for analysis of major  
121 elements using a Cameca SX100 electron microprobe at the School of GeoSciences,  
122 University of Edinburgh. Details of the analytical conditions and methods are given by  
123 Hayward (2011) and by Hartley and Thordarson (2013). The accuracy of major  
124 element determinations is <± 1% of total value. Analytical data for feldspar and  
125 pyroxene are given in Supplementary Table S3, S4, respectively.

126 ~~Zircon crystals within 30-160  $\mu\text{m}$  fraction were separated from crushed felsic~~  
127 ~~tuffaceous rock and rhyolitic lava (no. 14-18, 14-19 and 14-20; c. 5 kg each). A Wilfley~~  
128 ~~table, Frantz Isodynamic magnetic separator and a high density solution (lithium~~  
129 ~~polytungstate; 2.85 g/ml) were used to aid separation. Zircon grains were randomly~~  
130 ~~handpicked under a binocular microscope, mounted in epoxy resin and polished~~  
131 ~~sufficiently to expose the center of the grains. Internal structures were studied with a~~  
132 ~~scanning electron microscope using cathodoluminescence (CL) at the School of~~  
133 ~~GeoSciences, University of Edinburgh. U-Pb analysis was then performed using a~~  
134 ~~Cameca IMS-1270 secondary ion mass spectrometer (SIMS) at the School of~~  
135 ~~GeoSciences, University of Edinburgh, using the methods reported by Kelly et al.~~  
136 ~~(2008) and Ustaömer et al. (2012). Errors of the reported ages are  $\pm 1\sigma$ . Related~~  
137 ~~geochronological plots were produced using ISOPLOT (Ludwig, 2012). Analytical~~  
138 ~~results are listed in the Supplementary Table S5.~~

#### 140 ~~32.~~ Volcanism in the Kyrenia Range

141 Rifting of the Southern Neotethys during Late Permian and Early-Middle Triassic  
142 was followed by continental break-up during the Late Triassic-Early Jurassic  
143 (Robertson and Woodcock, 1979; Şengör and Yılmaz, 1981; Robertson and Dixon,  
144 1984; Garfunkel, 1998; Robertson et al., 2020). The Kyrenia Range then formed part  
145 of a carbonate platform that slowly subsided during Jurassic ~~and~~ Early Cretaceous in  
146 a along a passive margin ~~of the Southern Neotethys setting~~ (Robertson and Woodcock,  
147 1986). The carbonate platform was deformed and metamorphosed under greenschist  
148 facies during the Late Cretaceous, extensionally exhumed, and then unconformably  
149 overlain by basic extrusive igneous rocks of latest Cretaceous ~~and~~ Paleogene age  
150 (Figs. 2-3) (Ducloz, 1972; Baroz, 1979; Robertson and Woodcock, 1986). In addition,  
151 felsic volcanics and tuffs are exposed as thrust slices at a low structural level in the  
152 south of the range (Figs. 2b, 3) (Robertson et al., 2012b). These volcanic rocks have

153 no exposed base and are directly overlain by ~~the a~~ much larger thrust sheet that  
154 includes the mainly basic extrusive igneous rocks.

155 Baroz (1979, 1980) mapped the volcanic rocks and carried out petrographic and  
156 whole-rock chemical analysis of major elements. He reported the presence of a lower  
157 stratigraphic sequence of basalt, dolerite, trachybasalt, trachyandesite, dacite and  
158 rhyolitic tuff, and interpreted this as a bimodal basic-acidic calc-alkaline suite, related  
159 to a Late Maastrichtian volcanic arc. In contrast, the Paleogene volcanic assemblage  
160 was suggested to have erupted stratigraphically above in a post-collisional (intra-  
161 continental) strike-slip setting (Baroz, 1980). However, a thrust was later mapped  
162 between the mainly felsic and mainly basic suites (Robertson et al., 2012b),  
163 complicating this simple stratigraphy.

164 Chemical analysis of the Paleogene basaltic lavas, including immobile trace  
165 elements, initially suggested an alkaline, within-plate eruptive setting (Pearce, 1975).  
166 Subsequent chemical analysis, including some immobile elements, confirmed ~~this a~~  
167 mainly alkaline within-plate setting but also revealed some evidence of a subduction  
168 influence (e.g. negative Nb anomaly) (Robertson and Woodcock, 1986). The  
169 Paleogene lavas were ~~at that time~~ inferred by these authors to have erupted in a  
170 transtensional setting along the northern, active margin of the 'Troodos ocean'. Huang  
171 et al. (2007) carried out additional chemical analysis of the Paleogene basaltic lavas,  
172 mainly from the eastern and western Kyrenia Range, with emphasis on immobile  
173 elements, and proposed a Late Cretaceous-Paleogene back-arc setting related to  
174 northward subduction.

175 The felsic rocks of the lower most thrust sheet, known as the Fourkovouno  
176 (Selvilitepe)<sup>1</sup> Formation, ~~(up to 400 m thick)~~, begin with marine water-lain felsic tuffs

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<sup>1</sup> For simplicity, wWe use the traditional stratigraphy for the formation names (more recent Turkish equivalents are mentioned initially). However, we use current the the-Turkish names

177 and subaqueous felsic debris-flow deposits and culminate in laterally discontinuous,  
178 thick-bedded to massive rhyolitic lava flows (Fig. 3) (Moore, 1960; Baroz, 1979; Huang  
179 et al., 2007; Robertson et al., 2012b). The felsic rocks are locally cut by small (meter-  
180 sized) basaltic/doleritic intrusions (~~that were~~ not studied). A Late Cretaceous age (Late  
181 Campanian) was ~~previously~~ suggested for the felsic lavas, based on sparse planktic  
182 foraminifera (*Globotruncana* sp.) that were recognised in a chalky interbed near the  
183 top of the succession (Baroz, 1979).

184 The Late Cretaceous and Paleogene basaltic igneous rocks have been divided  
185 into ~~three-two~~ formations, namely the Maastrichtian Melounda (Mallıdağ) Formation  
186 ~~and~~ the Paleocene-Middle Eocene Ayios Nikolaos (Yamaçköy) Formation (Fig. 2a).

187 The Melounda Formation, c. 300 m thick, unconformably overlies the exhumed  
188 Mesozoic platform carbonates of the Trypa (Tripa) Group or, in places,  
189 unmetamorphosed terrigenous turbidites of the Kiparisso Vouno (Alevkaya Tepe)  
190 Member of the Melounda Formation (Baroz, 1979; Robertson and Woodcock, 1986;  
191 Robertson et al., 2012b). The basaltic lavas of the Melounda Formation are commonly  
192 intercalated with pinkish pelagic limestones (Fig. 2a). The succession locally includes  
193 lenticular, lithoclastic breccias and finer-grained clastic facies that were reworked from  
194 the underlying meta-platform carbonates (Fig. 2a) (Robertson and Woodcock, 1986;  
195 Robertson et al., 2012b). Planktic foraminifera from ~~the~~ pelagic carbonates between  
196 the lavas indicate a Late Maastrichtian age (Baroz, 1979; Robertson et al., 2012b).

197 The overlying Ayios Nikolaos Formation, 300-400 m thick, comprises basaltic  
198 volcanics that are variably interbedded with pelagic carbonates. Calciturbidites,  
199 calcareous debris-flow deposits and carbonate-rock breccias appear above the lavas,  
200 towards the top of the overall succession (Baroz, 1979; Robertson and Woodcock,

---

(e.g. ~~for~~ that are currently in use (e.g. ~~for~~ settlements); pre-existing names are mentioned only initially).



201 1986; Robertson et al., 2012b, 2014) (Fig. 2a). Planktic foraminifera from pelagic  
202 carbonates that are interbedded with the lavas indicate a ~~maximum~~-Late Paleocene to  
203 Mid-Eocene age range (Baroz, 1979; Robertson et al., 2012b).

204

### 205 3. Methods

206 The Kyrenia Range is a complex thrust belt and therefore it was essential to remap  
207 the western Kyrenia Range, where felsic and basaltic volcanic rocks are exposed  
208 together (Moore, 1960; Baroz, 1979; Robertson et al., 2012b). In addition, sedimentary  
209 logs were measured and correlated to produce a composite stratigraphy of the  
210 volcanogenic successions that are exposed in two superimposed thrust sheets.

211 Optical microscopy was carried out on the samples collected. For the felsic  
212 lithologies, 13 samples were studied from the western range. For the basaltic  
213 lithologies, we used a combination of our samples (n=29) and also some that were  
214 previously collected for paleomagnetic study (n=26) throughout the Kyrenia Range  
215 (Morris et al., 2015).

216 Whole-rock major and trace element concentrations of the volcanic rocks were  
217 measured by X-ray fluorescence (XRF) on fused glass beads and pressed powder  
218 pellets at the School of GeoSciences, University of Edinburgh, using the well-known  
219 methods of Fitton et al. (1998) and Fitton and Godard (2004). Accuracy and precision  
220 are typically c. 5%. For representative samples, additional trace and rare earth element  
221 (REE) analysis was carried out by inductively coupled plasma-mass spectrometry  
222 (ICP-MS) at the ACME Analytical Laboratories, Vancouver. Major element contents  
223 were determined from a LiBO<sub>2</sub> fusion by ICP-ES, using 5 g of sample pulp. Trace  
224 element contents were determined from a LiBO<sub>2</sub> fusion by ICP-MS, again using 5 g of  
225 sample pulp. Detection limits range between 0.01 and 0.04 wt% for major oxides, 0.01  
226 and 0.1 ppm for trace and rare earth elements. The relative standard deviation for the

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227 REE is ~5% and for all other trace elements is up to 10%, with quality control using  
228 international geostandards (see <http://acmelab.com>).

229 In addition, Sr-Nd-Hf isotopic analysis was performed on a Neptune Plus multi-  
230 collector (MC)-ICP-MS at Wuhan SampleSolution Analytical Technology Co., Ltd.,  
231 China, as reported by Li et al. (2012). Whole procedural blanks were <100 pg for Sr,  
232 <50 pg for Nd and <50 pg for Hf.  $^{87}\text{Sr}/^{86}\text{Sr}$ ,  $^{143}\text{Nd}/^{144}\text{Nd}$  and  $^{176}\text{Hf}/^{177}\text{Hf}$  ratios were  
233 normalised to  $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ ,  $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$  and  $^{179}\text{Hf}/^{177}\text{Hf} = 0.7325$ , using  
234 the exponential law. Standard analysis yielded  $^{87}\text{Sr}/^{86}\text{Sr} = 0.710240 \pm 11$  (2SD, n = 4)  
235 for NBS987,  $^{143}\text{Nd}/^{144}\text{Nd} = 0.512440 \pm 8$  (2SD, n = 8) for GSB 04-3258-2015, and  
236  $^{176}\text{Hf}/^{177}\text{Hf} = 0.282224 \pm 6$  (2SD, n = 6) for Alfa Hf. In addition, USGS reference  
237 materials BCR-2 and RGM-2 were also analysed for Sr–Nd–Hf isotopes, and gave  
238 ratios of  $0.705039 \pm 8$  and  $0.704169 \pm 11$  for  $^{87}\text{Sr}/^{86}\text{Sr}$ ,  $0.512644 \pm 6$  and  $0.512808 \pm$   
239 9 for  $^{143}\text{Nd}/^{144}\text{Nd}$  and  $0.282864 \pm 6$  and  $0.283016 \pm 7$  for  $^{176}\text{Hf}/^{177}\text{Hf}$ , which is within  
240 error of recommended values (Thirlwall, 1991; Weis et al., 2006, 2007). The analytical  
241 data for the major, trace and rare earth elements/isotope ratio of the felsic and basaltic  
242 volcanics are listed in the Supplementary Table S1, S2, respectively.

243 In addition, two polished thin sections of basalt (no. 02 and 21) that contain  
244 relatively fresh feldspar and pyroxene crystals were selected for analysis of major  
245 elements using a Cameca SX100 electron microprobe at the School of GeoSciences,  
246 University of Edinburgh. Details of the analytical conditions and methods are given by  
247 Hayward (2011) and by Hartley and Thordarson (2013). The accuracy of major  
248 element determinations is  $<\pm 1\%$  of total value. Analytical data for feldspar and  
249 pyroxene are given in Supplementary Table S3, S4, respectively.

250 For dating, zircon crystals within 30-160  $\mu\text{m}$  fraction were separated from crushed  
251 felsic tuffaceous rock and rhyolitic lava (no. 14-18, 14-19 and 14-20; c. 5 kg each). A  
252 Wilfley table, Frantz Isodynamic magnetic separator and a high-density solution

253 (lithium polytungstate; 2.85 g/ml) were used to aid separation. The zircon grains were  
254 randomly handpicked under a binocular microscope, mounted in epoxy resin and  
255 polished sufficiently to expose the centre of the grains. Internal structures were studied  
256 with a scanning electron microscope using cathodoluminescence (CL) at the School  
257 of GeoSciences, University of Edinburgh. U-Pb analysis was then performed using a  
258 Cameca IMS-1270 secondary ion mass spectrometer (SIMS) at the School of  
259 GeoSciences, University of Edinburgh, using the methods reported by Kelly et al.  
260 (2008) and Ustaömer et al. (2012). Errors of the reported ages are  $\pm 1\sigma$ . Related  
261 geochronological plots were produced using ISOPLOT (Ludwig, 2012). Analytical  
262 results are listed in the Supplementary Table S5.

263

#### 264 4. Results

##### 265 4.1. ~~Restored~~ Volcanic successions

266 The felsic volcanics, mainly crop out in the western part of the Kyrenia Range in  
267 two thrust sheets (Figs. 3-4). The lower of these, which is relatively intact, extends from  
268 Geçitköy (Panagra) eastwards to Selvitepe (Fourkovouno) (Fig. 5a-e). The second,  
269 structurally higher unit, comprises discontinuous exposures extending from north of  
270 Geçitköy, to southwest of Karsiyaka (Vasileia) and west of Alevkaya Tepe (Kiparisso  
271 Vouno) (Fig. 5f-j). Farther west, in the Kayalar (Orga) area (Fig. 3a), a small thrust slice  
272 exposes ~~of~~ coarse-grained felsic tuff and a felsic debris-flow ~~deposits~~ ~~depositare~~  
273 ~~exposed~~. Small (tens of m-sized) outcrops of felsic volcanogenic rocks also occur  
274 farther east; e.g. northeast of Aşağı Dikmen (Kato Dikomon), between thrust sheets of  
275 ~~the~~ Mesozoic meta-carbonate rocks (Baroz, 1979). Small exposures are also  
276 intersliced with Late Cretaceous pelagic carbonates of the Melounda Formation to the  
277 north of Ergenekon (Agios Khariton) (Fig. 3a) (Robertson et al., 2012b).

278

279 The ~~restored local~~ successions are restored as a in the lower thrust sheet and an  
280 upper thrust sheet. The succession in the lower thrust sheet, c. 90 m thick, begins with  
281 weakly-bedded, white felsic tuff (beds <0.5 m thick) (Fig. 5a, c-e), followed by thick  
282 layers of matrix-supported felsic breccia-conglomerate (c. 10 m thick) that ~~weare~~  
283 ~~interpreted~~ as debris-flow deposits (Fig. 5a, c, e). The succession passes upwards into  
284 massive, vitreous, rhyolitic lava flows, individually up to 10 m thick (Figs. 5a,  
285 Supplementary Figure S1). Localised normal grading; e.g. northeast of Geçitköy,  
286 indicates that the succession is ~~,stratigraphically~~ the right way up, stratigraphically (Fig.  
287 5a-b). In places, steeply dipping, irregularly shaped basaltic or doleritic intrusions, up  
288 to 2 m ~~thick~~across, cut the felsic layering (Figs. 5a, Supplementary Figure S1). ~~In~~  
289 ~~contact with~~Adjacent to these intrusions, poorly consolidated tuffaceous sediments  
290 have undergone contact metamorphism to form hard, dark grey, flint-like, recrystallised  
291 felsic rock (Robertson et al., 2012b). There is no evidence of similar intrusions in the  
292 structurally overlying basalts, which is in keeping with the mapped tectonic contact  
293 between the two units.

294 The succession in the upper thrust sheet begins with greyish to greenish,  
295 tuffaceous matrix-supported conglomerates, including clasts of silicified rhyolite, chalk  
296 and basalt (Supplementary Figure S1). The estimated thickness ranges from c. 40 m  
297 north of Geçitköy (Fig. 5f), to c. 80 m southwest of Karsıyaka (Fig. 5g-h), to 30 m west  
298 of the Alevkaya Tepe (Fig. 5i-j).

299  
300 Contrasting field relations facilitate mapping of the Late Cretaceous versus the  
301 Paleogene formations (Baroz, 1979; Robertson and Woodcock, 1986; Hakyemez et  
302 al., 2000; Robertson et al., 2012b). ~~The Late Maastrichtian basalts (Melounda~~  
303 ~~Formation) are typically pillowed, forming repeated flows, up to c. 10 m thick. They~~  
304 ~~include pink pelagic carbonate as interstitial sediment and discontinuous layers (up to~~  
305 ~~40s cm thick), extending laterally for up to c. 10 m. In contrast, the Paleogene basalts~~

306 ~~(Ayios Nikolaos Formation) are commonly massive, especially in the eastern range,~~  
307 ~~where individual lava flows reach 10s of m thick, with relatively few pelagic~~  
308 ~~intercalations (Supplementary Figure S2). However, in the central and western ranges,~~  
309 ~~the lava flows are mainly thinner (<10m) and are intercalated with pelagic carbonates~~  
310 ~~that contain abundant reddish chert (formed by diagenetic replacement). The~~  
311 ~~contrasting features facilitate mapping of the two formations (Baroz, 1979; Robertson~~  
312 ~~and Woodcock, 1986; Hakyemez et al., 2000; Robertson et al., 2012b).~~

313 ~~During this work, it was found that there are geochemical differences between~~  
314 ~~lavas of similar age along the far eastern, eastern, central and western Kyrenia Range~~  
315 ~~(see below). We, therefore, give some more specific information on the lavas in these~~  
316 ~~areas.~~

317 In the far east of the Kyrenia Range, the Karpas Peninsula (Fig. 3a), both pillowed  
318 and massive basalt are widely exposed near Balalan (Platanissos). ~~The lavas include~~  
319 ~~interpillow carbonate and lenticular interbeds of pelagic carbonate (< 1 m thick) (Figs.~~  
320 ~~3a, Supplementary Figure S2) contain Late Maastrichtian microfossils (Robertson~~  
321 ~~et al., 2012b).~~

322 In the eastern range, basaltic rocks of the Melounda Formation contain abundant  
323 pelagic carbonate and are overlain by massive basalts of the Ayios Nikolaos Formation;  
324 i.e. from Ağıllar (Mandres) to Mallıdağ (Melounda) (Fig. 3a) (Baroz, 1979; Robertson  
325 and Woodcock, 1986; Hakyemez et al., 2000; Robertson et al., 2012b).

326 In the central range, pillow lava with interbedded pelagic carbonate of the  
327 Melounda Formation occurs in two settings. The first is directly above the Upper  
328 Cretaceous basal unconformity, with the ~~underlying~~ meta-carbonate platform rock  
329 beneath; e.g. near Tirmen (Trypimeni). The second setting is as one (or several), small  
330 thrust sheets along the southern flank of the range; e.g. Değirmenlik (Kythrea), Boğaz  
331 (Bogaz) and southwest of Alevkaya Tepe (Figs. 3a, Supplementary Figure S2). Basalt

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332 with some interbedded pelagic carbonate of the Ayios Nikolaos Formation is exposed  
333 higher in the succession (within a shear zone), near Tirmen, Arapköy (Klepini) and  
334 İncesu (Motides) (Baroz, 1979).

335 In the western range, basaltic lavas interbedded with pelagic carbonates of the  
336 Melounda Formation are exposed in a thrust sheet along the southern margin of the  
337 range; e.g. Geçitköy (Figs. 3, Supplementary Figure S2). Similar basaltic lavas with  
338 pelagic carbonates and local carbonate breccias, are exposed along the northern  
339 margin of the range; e.g. near Karşıyaka and farther west (Baroz, 1979; Hakyemez et  
340 al., 2000; Robertson et al., 2012b). In addition, basaltic lava interbedded with pelagic  
341 chalk of the Ayios Nikolaos Formation occurs higher in the succession (southward-  
342 younging) along the eastern side of Geçitköy road (Huang et al., 2007; Robertson et  
343 al., 2012b).

344 In summary, the Late Maastrichtian basalts (Melounda Formation) are typically  
345 pillowed, forming repeated flows (individually up to c. 10 m thick), reaching a maximum  
346 of > 90 m in the east (near Balalan). The lavas include pink pelagic carbonate as  
347 interstitial sediment and discontinuous layers. In contrast, the Paleogene basalts  
348 (Ayios Nikolaos Formation) are commonly massive, especially in the eastern range,  
349 where individual lava flows reach c. 45 m thick, with relatively few pelagic intercalations  
350 (Supplementary Figure S2). In the central and western ranges, thinner lava flows  
351 (<10m) are intercalated with pelagic carbonates and reddish chert (formed by  
352 diagenetic replacement). During this work, we identified geochemical differences  
353 between the Late Cretaceous-Paleogene basaltic lavas of similar age along the far-  
354 eastern, eastern, central and western Kyrenia Range (see below).

355

#### 356 4.2. Petrography

357 The felsic rocks are porphyritic, composed of quartz, plagioclase and sanidine,

358 together with subordinate biotite and rare hornblende. Large sanidine crystals (0.3-1  
359 mm) show simple twinning, whereas plagioclase commonly has lamellar twinning  
360 (Supplementary Figure S3). Biotite laths (0.3-0.8 mm in length) are preferentially  
361 orientated parallel to flow layering. Rare quartz phenocrysts (<1%, 0.2-0.5 mm) are  
362 commonly fragmented, probably related to quenching during eruption; i.e. explosive  
363 fragmentation (Supplementary Figure S3). The groundmass comprises  
364 microcrystalline quartz and feldspar, together with minor muscovite, biotite and rare  
365 hornblende (c. 60 µm). In some samples (e.g. nos. 14-19 and 14-20), the groundmass  
366 is cryptocrystalline, almost glassy (Supplementary Figure S3).

367 The basaltic rocks of both Late Cretaceous and Paleogene age generally fall into  
368 two groups in terms of mineral composition and texture. The first group (most common)  
369 is porphyritic with subhedral to euhedral clinopyroxene phenocrysts (30-40%), up to  
370 0.6 mm long, that occur interstitially or are intergrown with subhedral feldspar  
371 (Supplementary Figure S4, Table S3). Feldspar (45-60%) forms elongate, acicular  
372 laths, up to 1 mm long. Olivine is relatively rare (<5%). The second group is ophitic, as  
373 locally observed in the western Kyrenia Range with abundant anhedral to subhedral  
374 augite phenocrysts (up to 0.6 mm in size). Most plagioclase is acicular, enclosed or  
375 surrounded by augite (Supplementary Figure S4).

376 Most samples are slightly to moderately altered. Amygdales and veins are infilled  
377 with secondary minerals such as calcite and zeolite. Alteration is variable, for example,  
378 pyroxene and plagioclase are heavily altered to chlorite and clay minerals (e.g.  
379 smectite/sericite). The relatively high loss on ignition values result from secondary  
380 processes, which need to be taken into account prior to rock identification and tectonic  
381 discrimination.

#### 382 4.3. Whole-rock chemistry

383 Major-element oxides ( $K_2O$ ,  $Na_2O$  and  $CaO$ ) and trace elements (e.g. Rb, Sr and

384 Ba) exhibit a relatively wide compositional range, consistent with the effects of  
385 alteration (see Supplementary Table S1-S2). Alteration can be inferred from variable  
386 LOI values that range between 0.3-5.4 wt% for the felsic volcanics and 4.09-14.57 wt%  
387 for the basaltic rocks. This is consistent with the presence of considerable amounts of  
388 water- and/or carbonate dioxide-bearing minerals, for example chlorite and calcite, as  
389 observed petrographically (see Supplementary Figure S3, S4). Trace elements such  
390 as Ti, Zr, Y, Nb, Ta, V, Co, Th and REEs tend to be immobile during weathering and/or  
391 metamorphism below amphibole facies are, therefore, preferred for rock-type  
392 identification and tectonic discrimination (e.g. Pearce and Cann, 1973; Rollinson,  
393 1993). For basaltic rocks, samples with high crystal contents (cumulate composition or  
394 highly porphyritic) should be discounted from geochemical discrimination (Pearce,  
395 1996). In general, samples with immobile element  $Al_2O_3 > 20$  wt% (concentrated in  
396 feldspar), Sc  $> 50$  ppm (concentrated in clinopyroxene), or Ni  $> 200$  ppm (concentrated  
397 in olivine) are predicted to contain high amounts of cumulated minerals (Pearce, 1996).  
398 On this basis, basaltic rocks nos. 14-66, 14-50, 14-51, 14-58, 14-68, 08, 16, 25 and 28  
399 are discounted.

400 The more felsic assemblage is mainly rhyolitic in the classification of Winchester  
401 and Floyd (1977) but shows a trachytic affinity in the revised Zr/Ti vs. Nb/Y (Fig. 6a)  
402 diagram of Pearce (1996). Relatively high abundances of Th ( $>14$  ppm) but low Co  
403 concentrations ( $<12$  ppm) are consistent with high-K and shoshonitic affinities (Fig. 6b).  
404 Primitive mantle-normalised multi-element plots (Fig. 6c) show variable enrichments  
405 and depletions in large-ion lithophile elements (LILEs; e.g. Cs, Rb, Ba, K and Sr) that  
406 are influenced by alteration. The samples show negative anomalies of Nb, Zr and Ti,  
407 coupled with marked positive anomalies of Th/U, Pb, Nd and probably Sm (Fig. 6c).  
408 Total REE contents of the felsic assemblage (Fourkovouno Formation) vary from 115  
409 to 173 ppm. La ranges from 29-44 ppm,  $(La/Yb)_N=12.63-18.98$ , suggestive of an alkalic  
410 affinity. The ratios  $(La/Sm)_N=6.13-7.22$  and  $(Gd/Yb)_N=1.26-1.68$  suggest a large LREE



411 fractionation but low to moderate HREE fractionation (Fig. 6d). Pronounced negative  
412 Eu anomalies ( $\text{Eu}/\text{Eu}^*=0.32\text{-}0.48$ ) are consistent with plagioclase crystallization, as  
413 seen petrographically (see Supplementary Figure S3). In general, the samples are  
414 comparable to upper continental crust (UCC). The basaltic rocks plot in the fields of  
415 basalt and alkali basalt, with intermediate degrees of fractionation ( $\text{Zr}/\text{Ti}=0.01\text{-}0.02$ )  
416 and moderate alkalinity ( $\text{Nb}/\text{Y}=0.1\text{-}1.8$ ) on the rock-type discrimination plot (Fig. 7a).  
417 The basalts have a calc-alkaline affinity, except for samples from Mallıdağ (no. 01),  
418 Tirmen (no. 19-51), Değirmenlik (no. 14-64) and İncesu (no. 19-67) that exhibit high-K  
419 and shoshonitic affinities (Fig. 7b). The trace element compositions of the basaltic  
420 rocks exhibit wide ranges of Sr, Ba, Rb, Th, Ta, Nb and Ce. However, Zr, Hf, Sm, Ti  
421 and Y generally comparable with enriched mid-ocean ridge basalt (E-MORB) and  
422 oceanic island basalt (OIB) (Fig. 8a-c).

423 The basaltic rocks are characterised by variable REE concentrations (67.2-239.4  
424 ppm) and light REE enrichments ( $(\text{La}/\text{Yb})_N=3.18\text{-}19.35$ ) (Fig. 8d). Samples from the  
425 central range, at Değirmenlik and Tirmen, show much higher total REE contents and  
426 steeper REE patterns. Westwards in the Kyrenia Range, the patterns become  
427 smoother, with less pronounced enrichment in LREEs. In general, no marked Eu  
428 anomaly ( $\sim 1$ ) is observed.

#### 429 4.4. Whole-rock Sr-Nd-Hf isotopes

430 The selected basaltic samples have relatively uniform whole-rock Sr-Nd-Hf  
431 isotopic characteristics. Adopting a Late Cretaceous eruption age based on the  
432 paleontological dating (Robertson et al., 2012b), the calculated initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios  
433 range from 0.705054 to 0.705701, with  $\epsilon\text{Nd}(t)$  values of 2.27 to 5.76. Initial  $^{176}\text{Hf}/^{177}\text{Hf}$   
434 and  $\epsilon\text{Hf}(t)$  values are 0.282934-0.283017, 7.36-10.14, respectively (see  
435 Supplementary Table S2).

#### 436 4.5. Zircon U-Pb geochronology

437 Zircon crystals separated from the felsic rocks are typically angular to sub-angular,  
438 30-160  $\mu\text{m}$  long. The grains have variable internal textures, including banded zoning  
439 (e.g. zircon 1), concentric zoning (nos. 2, 4, 5 and 12), and minor sector zoning (nos.  
440 10, 11; Fig. 9a). Fractures (or depressions) and some inherited cores are affected by  
441 fluid-related chemical alteration (nos. 6 and 8) (Fig. 9a). The zircon grains have high  
442 Th/U ratios (0.19-1.24), suggestive of a magmatic origin (Rubatto, 2002). Twelve  
443 zircon grains were analysed with concordance levels ranging from 99-105%. The  
444 calculated  $^{206}\text{Pb}/^{238}\text{U}$  age with the generally lowest possible error was used for age  
445 determination. The much older zircon grain (no. 4) is subhedral with a patchy  
446 xenocrystic texture, suggestive of a recycled origin (Fig. 9a). The concordant or nearly  
447 concordant data (100-102%) with a young age distribution yielded weighted mean  
448  $^{206}\text{Pb}/^{238}\text{U}$  gave ages of  $74.0\pm 0.6$  Ma,  $74.0\pm 0.4$  Ma and  $71.7\pm 0.7$  Ma (Fig. 9b-d).

449

#### 450 5. Discussion

##### 451 5.1. Genetic type/source characteristics

452 For the felsic rocks, the high LREE/HREE (e.g.  $(\text{La}/\text{Yb})_{\text{N}}=12.63-18.98$ ) and  
453 negative Eu anomalies, resemble model UCC but with depletion in Eu, Sr and P (Fig.  
454 6c-d). The A/CNK (molar ratio of  $\text{Al}_2\text{O}_3/(\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O})$ ) ratio is useful to identify the  
455 source of felsic rocks (Chappell and White, 1992) but is questionable for altered rocks  
456 like those of the Kyrenia Range.  $\text{SiO}_2$ ,  $\text{P}_2\text{O}_5$  and Th can also be used as indices of the  
457 fractional crystallization of a felsic melt (Chappell and White, 1992), but again may be  
458 affected by alteration (specially  $\text{SiO}_2$ ). In our samples,  $\text{P}_2\text{O}_5$  decreases with increasing  
459  $\text{SiO}_2$  and Th (Fig. 10a-b), trends that are indicative of I (igneous)- or A (anorogenic)-  
460 type granites. The relatively low contents of Zr, Nb, Y, La, Ce, Zn and Ga further

461 characterise these rocks as I-type (Fig. 10c-d). [The abundances of Th and Co, in](#)  
462 [particular hint at a shoshonitic composition \(Hastie et al., 2007\).](#)

463 The trace element patterns of the basaltic rocks show a wider range of variation  
464 (Fig. 8). E-MORB and OIB-type basalts were derived from relatively enriched sources,  
465 specifically for the basalts of the Karpas Peninsula, eastern range and most of the  
466 central range. The lesser enrichment of Nb within the basalts of the central range at  
467 Ergenekon (no. 31), Arapköy (no. 14-62) and Alevkaya Tepe (no. 14-49), and the  
468 western range at Karsıyaka (no. 14) and Geçitköy (no. 10), can be explained by a  
469 relatively depleted mantle source, together with subduction modification. There is little  
470 evidence of the addition of radiogenic Sr from seawater, either to the source melt or  
471 during post-magmatic alteration. The Sr vs. Nd isotope correlation diagram (Fig. 11a)  
472 does not show a significant shift towards higher initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios as would be  
473 expected in both of these cases.

474 The basalts all have positive  $\epsilon\text{Nd}(t)$  values (+2.27 - +5.76), consistent with  
475 derivation from an OIB-like source mantle (Fig. 11a). The samples plot directly on the  
476 terrestrial array, overlapping with the Nd isotope compositions of Etna and the Aeolian  
477 Arc, but intermediate between these settings in Hf isotopes (Fig. 11b). The samples  
478 that show trace element evidence of a subduction-modified source, based on relatively  
479 flat MORB-normalised multi-element patterns and negative Nb anomalies (no. 10,  
480 Geçitköy), plot in the Aeolian Arc-type mantle [field](#). Conversely, samples from the  
481 Karpas Peninsular, at Balalan, plot close to the field of Etna and exhibit less subduction  
482 modification (Fig. 11b). The trace element and isotopic data, therefore, yield consistent  
483 results.

484 The involvement of a crustal component is also suggested by incompatible  
485 element ratio plots, on which the basalts exhibit two distinct trends. The samples with  
486 relatively strong subduction signatures plot away from the MORB array (Fig. 12a),  
487 fingerprinting a subduction-related lithospheric mantle end member. A probable lower

488 crust input is suggested by relatively higher La/Nb and Ba/Nb ratios (e.g. Wang et al.,  
489 2004). In contrast, an OIB-like mantle component dominated most of the other basalt  
490 samples, as indicated by relatively low La/Nb, Zr/Nb but high Nb/Y ratios (Fig. 12b).  
491 The ratios of highly incompatible elements with small distribution coefficients (e.g.  
492 Zr/Nb, Nb/Th, Nb/La) are sensitive indicators of mantle source heterogeneity  
493 (Saunders et al., 1988; Condie, 2003; Pearce and Stern, 2006). The samples from the  
494 western locations fall in the fields of an enriched component and upper continental  
495 crust (Fig. 12c), which could represent mantle mixing with continental sediments  
496 (subducted) or crustal contamination. In contrast, the samples from the easterly  
497 locations resemble those of the OIB/recycled component, and cluster around a  
498 different batch-melting trajectory (10-20%) (Fig. 12c) that is representative of recycled  
499 oceanic lithosphere (Saunders et al., 1988). We, therefore, infer that the basalts are  
500 likely to have been derived from multiple sources (i.e. two or more source components  
501 were involved).

502

## 503 5.2. Tectonic discrimination

504 On the Nb vs. Y diagram (Fig. 13a), the felsic volcanics show similarities with  
505 volcanic arc granites and syn-collisional granites. One sample from Geçitköy (no. 16-  
506 08), with relatively high Nb and Y concentrations, plots near the field of within-plate  
507 granite. On the Th/Yb vs. Ta/Yb diagram (Fig. 13b), the volcanics mainly erupted on  
508 active continental margins. An active continental margin setting is therefore inferred  
509 for the felsic volcanics.

510 Basalts of the Kyrenia Range (e.g. easterly locations) mostly exhibit a within-plate  
511 affinity (Fig. 8). However, some samples, especially in the west, have a subduction-  
512 related signature (e.g. negative Nb anomalies) (Fig. 8b-c) akin to volcanic arc basalts.  
513 Most samples have a within-plate affinity on the Zr/Y vs. Zr (Fig. 13c) diagram, except

514 for two samples, from Ergenekon and Arapköy (central range), that straddle the island  
515 arc and MORB fields. The basalts overlap the MORB and back-arc basin basalt fields  
516 on the Ti vs. V plot (Fig. 13d). Clinopyroxene compositions, as shown in the ternary  
517 TiO<sub>2</sub>-MnO-Na<sub>2</sub>O plot (see Supplementary Figure S5), indicate a dominantly within-  
518 plate affinity. The relatively high TiO<sub>2</sub> contents of the clinopyroxenes point to an  
519 enriched MORB-related source (see Supplementary Figure S5). Overall, the inferred  
520 subduction contribution decreases from west to east along the range.

521

### 522 5.3. Global comparison of the Kyrenia Range volcanic rocks

523 Global ~~comparison (see Supplementary Figure S6) lend~~ comparisons (see  
524 Supplementary Figure S6) lend -support to our interpretation of the Kyrenia Range  
525 volcanic rocks as being subduction-related. In particular,

526 ~~W~~ within the Late Cretaceous felsic rocks ~~of the Kyrenia Range~~, the enrichment of  
527 Rb, Th and Sm ~~and versus~~ the depletion of Ti, Ba, Nb, P compared to primitive mantle,  
528 are characteristic of volcanic arc granites (Fig. 6c) (Pearce et al., 1984).

529 The Kyrenia Range Maastrichtian and Paleogene basalts of within-plate affinities  
530 (e.g. Balalan, eastern range) mainly ~~r~~ have MORB-normalised spider patterns that are  
531 similar to the Pliocene-Quaternary OIB-type basalts of the Tyrrhenian Sea, W  
532 Mediterranean (Peccerillo, 2017). The MORB-normalised patterns of these basalts  
533 suggest a within-plate eruptive setting, with a small sub-crustal lithosphere  
534 metasomatic influence (Kastens et al. 1988; Peccerillo, 2017).

535 Back-arc basalts of the Early Cretaceous Sarmiento Complex, southern Chile  
536 (Saunders et al., 1979) and the Miocene Japan Sea (Tamaki et al., 1990) resemble a  
537 minority of the Kyrenia Range basalts (c. 10%; e.g. Geçitköy, western range), ~~but~~  
538 although with lower abundances of K, Rb, Ba, Th, Ta and Nb. The relative ~~depletions~~

539 ~~depletion of these elements is characteristic of~~ ~~are likely to be representative of~~  
540 eruption in a relatively wide, evolved back-arc basin (e.g. Bransfield Strait; Weaver et  
541 al., 1979). However, another possible setting is eruption in a back-arc basin that was  
542 influenced by oblique extension (e.g. Guaymas pull-apart basin; Brune et al., 2012).  
543 The relatively flat MORB-normalised patterns, akin to E-MORB, of both the Bransfield  
544 Strait and Guaymas back-arc basin basalts (e.g. Saunders et al., 1982; Keller et al.,  
545 2002) point to substantial involvement of depleted MORB sources.

546

#### 547 5.4. Comparison with the adjacent Tethyan region

548 Late Cretaceous and Paleogene arc-related rocks; i.e. similar in age of those in  
549 the Kyrenia Range are critical to any regional magmatic-tectonic interpretation.

550 In western Cyprus, ~~there is a~~ Late Cretaceous arc-derived volcanogenic  
551 succession (Kannaviou Formation) ~~that~~ overlies the Troodos ophiolite. The  
552 volcanoclastic sediments ~~there~~ were derived from intermediate-felsic ~~composition~~  
553 eruptions, c. 6 Ma before the felsic eruptions in the Kyrenia Range (c. 80 vs. 74 Ma).  
554 Grain size and textural evidence suggest that, depending on prevailing wind direction  
555 and strength, the eruptive centers are likely to have been <100 km away (Chen and  
556 Robertson, 2019). The centers ~~for the Kannaviou Formation were~~ ~~could have been~~  
557 located to the south of the Kyrenia Range or its lateral extension, areas that are not  
558 now exposed. The Kyrenia Range was affected by south-directed thrusting during Mid-  
559 Eocene and Late Miocene-earliest Pliocene (Baroz, 1979; Robertson et al., 2012b,  
560 2014; McCay and Robertson, 2013; Robertson and Kinnaird, 2016), such that the distal  
561 (southerly) edge of the inferred over-riding active margin (see below) could have been  
562 subducted, structurally over-ridden or eroded. In other words, additional arc units could  
563 have existed to the south of their present location in the Kyrenia Range.

564 To the northeast (c. 200 km), the Misis-Andirın Range in southern Turkey is an

565 extension of the Kyrenia lineament (Robertson et al., 2004). The outcrop, known as  
566 the Misis-Andırın Complex, includes massive basalts, pillow lavas, lava breccias and  
567 volcanoclastic debris-flows. These lithologies were previously inferred to be Miocene  
568 (Floyd et al., 1991, 1992). However, inter-lava sediments there have been dated as  
569 Campanian-Maastrichtian (Robertson et al., 2004), similar in age to the oldest basaltic  
570 rocks in the Kyrenia Range. The Miocene sediments are structurally intercalated. The  
571 basalts of the Misis-Andırın Complex are tholeiitic, with mildly enriched, subduction-  
572 related characteristics (e.g. moderate La/Nb ratios), and are interpreted as of back-arc  
573 basin origin (Floyd et al., 1991, 1992), which is consistent with our interpretation of the  
574 Kyrenia Range basalts ~~(Floyd et al., 1991, 1992)~~ (Fig. 14a-b, Supplementary Figure  
575 S6).

576 The Kyrenia Range is backed to the north by the Alanya Massif (Fig. 1), which has  
577 been interpreted as a continental fragment, separate from the Tauride carbonate  
578 platform to the north (Robertson et al., 2012a; Çetinkaplan et al., 2016). The Alanya  
579 Massif is represented by the Late Precambrian-Mesozoic, mainly meta-sedimentary  
580 rocks. High pressure-low temperature (HP-LT) metamorphics document Late  
581 Cretaceous (82-80 Ma) subduction/exhumation (Oberhänsli et al., 2010, 2013). This is  
582 inferred to have taken place along the northern, active margin of the Southern  
583 Neotethys (Çetinkaplan et al., 2016; Robertson and Parlak, 2020). The felsic volcanics  
584 (Fourkovouno Formation) in the Kyrenia Range could have erupted along the southern  
585 (trailing) margin of the Alanya microcontinent where they could have escaped  
586 metamorphism. However, they could also represent a separate more southerly unit.

587 Late Cretaceous continental margin arc-type magmatic rocks also occur  
588 extensively farther east, in SE Turkey, as two separate belts. ~~The Kyrenia Range~~  
589 ~~volcanics are most comparable with volcanogenic units that occur within an extension~~  
590 ~~of the same (or similar) tectonic lineament; i.e. close to the leading edge of the over-~~  
591 ~~riding Tauride (Eurasian) continent.~~

592 The most southerly, known Late Cretaceous arc rocks in SE Turkey are exposed  
593 in the Helete area, part of a zone of frontal thrust sheets (Yıldırım, 2015; Nurlu et al.,  
594 2016) (Fig. 1). Two thrust assemblages are exposed, both cut by granites that have  
595 been Arc-related volcanogenic units there comprise basaltic andesite, andesite, dacite,  
596 rhyolite and common tuffaceous rocks. Cross-cutting granites have been  
597 radiometrically dated at 93-83 Ma (Cenomanian-Campanian) (Nurlu et al., 2016). The  
598 lower of the assemblages, the Helete volcanics, comprises arc-related basaltic  
599 andesite, andesite, dacite, rhyolite and common tuffaceous rocks. The upper of the  
600 two thrust assemblages is a dismembered Late Cretaceous supra-subduction zone-  
601 type ophiolite (Meydan ophiolite) (Yıldırım, 2015; Nurlu et al., 2016). The cross-cutting  
602 intrusive rocks sealed the thrust sheets during the Late Cretaceous. The granitic rocks  
603 The intrusive rocks are interpreted as a calc-alkaline, subduction-related suite that  
604 formed by mixing of mantle and crustal sources, as suggested by Sr-Nd-Pb isotopic  
605 data (Nurlu et al., 2016). The granitic rocks exhibit relatively flat chondrite-normalised  
606 REE patterns, similar to E-MORB (Nurlu et al., 2016) (Supplementary Figure S6). In  
607 contrast, the chondrite-normalised REE patterns of the Kyrenia Range felsic rocks are  
608 similar to UCC (Fig. 6c), suggesting a more continentally-influenced proximal (northerly)  
609 setting.

610 The second belt of arc-related magmatism is located up to c. 100 km ~~located~~  
611 farther north and north-east, in the Keban (Malatya), Göksun (Kahramanmaraş) and  
612 Baskil (Elazığ) regions (Fig. 1) (Yazgan and Chessex, 1991; Parlak, 2006; Rızaoğlu et  
613 al., 2009; Karaoğlu et al., 2013, 2016) (c. 100 km), where a Arc-granitic rocks (Baskil  
614 intrusives) cut ~~both~~ the Late Cretaceous Göksun ophiolite and equivalents (i.e. Berit  
615 meta-ophiolite (=North Berit ophiolite), Kömürhan ophiolite and İspendere ophiolite)  
616 that are inferred to have formed above a northward-dipping subduction zone (Parlak,  
617 2006). Similar the arc granites also cut the structurally over-riding Malatya metamorphic  
618 unit, which ~~that~~ is interpreted as the southern part of the Tauride microcontinent. The



619 Göksun ophiolite (and equivalents) and the Malatya metamorphic (Tauride) units were  
620 juxtaposed along an active continental margin when the arc-granitic rocks were  
621 intruded (Robertson et al., 2006; Rızaođlu et al., 2009; Karaođlan et al., 2013, 2016).

622 The extrusive rocks of the Göksun and related ophiolites can also be broadly  
623 correlated with widespread assemblages of basic igneous rocks and volcanogenic  
624 sediments in the Malatya-Elazığ region, known as the Yüksekova Complex (Perincek  
625 and Özkaya, 1981; Aktaş and Robertson, 1984). The basalts have a tholeiitic to  
626 tholeiitic-transitional character, variable Zr/Y (1.5-6) ratios and mantle-array-aligned  
627 Sr-Nd isotopes (Ural et al., 2015). These basic volcanics were derived from a depleted  
628 mantle source without a significant crustal contribution; i.e. from an intra-oceanic  
629 setting within the Southern Neotethys (Ural et al., 2015).

630 The arc rocks that cut both the Göksun (and related) ophiolites and the Malatya  
631 Metamorphics (Baskil intrusives) in the northern belt are mainly hornblende-biotite  
632 granodiorites and 'normal' granites, dated radiometrically at 88-82 Ma (Santonian-  
633 Campanian) (Parlak, 2006; Rızaođlu et al., 2009; Karaođlan et al., 2016); i.e. up to c.  
634 5 Ma younger than Helete area granites in the south. These southerly granitic rocks  
635 have I-type, calc-alkaline arc affinities, with both mantle and crustal-derivation features  
636 (Nurlu et al., 2016). The Baskil intrusions of the northern belt are relatively enriched in  
637 LREE (Parlak, 2006; Rızaođlu et al., 2009) compared to the Helete granites of the  
638 south, which suggests a more crustally influenced source during magma genesis for  
639 the former. Similarly, the Kyrenia Range basalts have relatively high Zr/Y (2.8-7.7) and  
640 Nb/Y (Figs. 12b, 13c), suggestive of a magmatic contribution from continental crust or  
641 subducted continentally derived sediments (Zindler and Hart, 1986; Pearce, 1996).

642 The northerly arc has been explained by northward subduction of Mesozoic  
643 oceanic basin (Berit ocean) that was located between two continental units; i.e. the  
644 Bitlis and Pütürge massifs in the south and the Tauride continent represented by the  
645 Malatya Metamorphics in the north (Robertson et al., 2012a; Karaođlan et al., 2013,

646 2016; Barrier et al., 2018). The Late Cretaceous arc-related intrusive magmatism of  
647 the northern belt in SE Turkey (Baskil intrusives) evolved towards collisional, including  
648 shoshonitic compositions (74-72 Ma), which are interpreted to indicate continental and  
649 post-collisional settings (Kuşcu et al., 2013; Erturk et al., 2018; Sar et al., 2019). The  
650 shoshonitic compositions of some of the granitic rocks, together with the evidence of  
651 HP/LT metamorphism, are consistent with the collision of the Tauride carbonate  
652 platform (i.e. Malatya Metamorphics) to the north with the Bitlis and Pütürge continental  
653 units to the south. However, the origins of shoshonites remain controversial because  
654 the necessary partial melting of mantle and interaction with subduction-related fluids  
655 can take place in a variety of pre-, syn- and post-collisional settings (e.g. Campbell et  
656 al., 2014).

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657 Several authors have correlated the Late Cretaceous Göksun and related  
658 ophiolites in the north with the Meydan ophiolite (Berit meta-ophiolite of Yılmaz et al.,  
659 1993) in the south. If correct, the granitic intrusives in both belts originated as different  
660 parts of a single arc lineament that straddled the northerly active continental margin of  
661 the oceanic basin. The southerly arc and ophiolitic rocks (Helete-Meydan) were  
662 emplaced southwards during the latest Cretaceous, although further southward  
663 thrusting took place, mainly during the Eocene and Miocene (Perinçek and Kozlu, 1984;  
664 Yılmaz et al., 1993).

665 In many reconstructions, oceanic crust still remained between the Bitlis-Pütürge  
666 continental units and Arabia until the Miocene (Yılmaz, 1993; Robertson et al., 2012a;  
667 Barrier et al., 2018; van Hinsbergen et al., 2020). The Bitlis and Pütürge massifs have  
668 been generally correlated with the Alanya metamorphic massif to the north of Cyprus  
669 (with or without a direct continuation) (Çetinkaplan et al., 2016). If correct, the Late  
670 Cretaceous Kyrenia Range felsic arc volcanism (and that of the Kannaviou Formation)  
671 could have been located to south of the Alanya continent, with a possible eastward  
672 continuation to south of the Bitlis and Pütürge continental units. (Parlak, 2006;

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673 ~~Robertson et al., 2006). Examples occur in the Koban (Malatya), Göksun~~  
674 ~~(Kahramanmaraç) and Baskil (Elazığ) regions (Fig. 1) (Yazgan and Chescek, 1994;~~  
675 ~~Parlak, 2006; Rızaoğlu et al., 2009; Karaoğlu et al., 2013, 2016). The arc rocks are~~  
676 ~~mainly hornblende biotite granodiorites and 'normal' granites that are radiometrically~~  
677 ~~dated as 88–82 Ma (Santonian–Campanian) (Rızaoğlu et al., 2009; Karaoğlu et al.,~~  
678 ~~2016). The granitic rocks have I-type, calc-alkaline arc affinities, with both mantle and~~  
679 ~~crustal derivation features (Parlak, 2006; Rızaoğlu et al., 2009; Karaoğlu et al., 2013,~~  
680 ~~2016). The intrusions are relatively enriched in LREE (Parlak, 2006; Rızaoğlu et al.,~~  
681 ~~2009) compared to the Hellele granites, which suggests a more crustally influenced~~  
682 ~~source during magma genesis. The above granitic rocks are located in a structurally~~  
683 ~~higher position than the Kyronia and Hellele arc rocks (i.e. above the Pütürge~~  
684 ~~metamorphic massif) and are likely to have a different region. The northerly arc have~~  
685 ~~been explained by northward subduction of a small oceanic basin (Borit ocean) that~~  
686 ~~was located between two continental units; i.e. Bitlis–Pütürge in the south and the~~  
687 ~~Tauride continent (Malatya) in the north (Robertson et al., 2012a; Karaoğlu et al.,~~  
688 ~~2013, 2016; Barrier et al., 2018).~~

689 All of the above extrusive and intrusive felsic arc rocks differ from the Late  
690 Cretaceous ophiolitic and oceanic arc rocks in the region. These include a widely  
691 exposed assemblage (Malatya–Elazığ region) of basic igneous rocks and volcanogenic  
692 sediments, known as the Yüksekova Complex (Perincek and Özkaya, 1981; Aktaş and  
693 Robertson, 1984). The basalts have a tholeiitic to tholeiitic transitional character,  
694 variable Zr/Y (1.5–6) ratios and mantle array aligned Sr–Nd isotopes (Ural et al., 2015).  
695 These basic volcanics were derived from a depleted mantle source without a significant  
696 crustal contribution; i.e. from an intra-oceanic setting within the Southern Neotethys  
697 (Ural et al., 2015). In contrast, the Kyronia Range basalts have higher Zr/Y (2.8–7.7)  
698 and Nb/Y (Figs. 12b, 13c), suggestive of a magmatic contribution from continental crust  
699 or subducted continentally derived sediments (Zindler and Hart, 1986; Pearce, 1996).

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700 Paleogene magmatic rocks are also present in SE Turkey. Eocene volcanic rocks  
701 and associated minor intrusions, known as the Maden Complex, unconformably overlie  
702 and cut the Pütürge and Bitlis ~~continental metamorphic units~~metamorphosed  
703 continental units (Fig. 1) (Hempton, 1985; Yazgan and Chessex, 1991; Aktaş and  
704 Robertson, 1984, 1990; Yılmaz, 1993; Elmas and Yılmaz, 2003; Robertson et al., 2006;  
705 Erturk et al., 2018). The volcanics show enrichment in LILEs and relative depletion in  
706 Nb, Ta and Ti compared to MORB (Erturk et al., 2018), similar to a minority of the  
707 subduction-related volcanics in the Kyrenia Range (see Supplementary Figure S6).  
708 The Eocene Maden Complex magmatism is widely proposed to relate to back-arc  
709 rifting of the (Eurasian) active continental margin during the later stages of northward  
710 subduction of the Southern Neotethys, in keeping with the widespread view that  
711 collision with Arabia was delayed until the Miocene (Aktaş and Robertson, 1984, 1990;  
712 Yılmaz, 1993; Yılmaz et al., 1993; -Yiğitbaş and Yılmaz, 1996; Robertson et al., 2006,  
713 2007; van Hinsbergen et al., 2020). Oblique convergence, -as-proposed by several  
714 authors (Aktaş and Robertson, 1990; Elmas and Yılmaz, 2003), could have resulted in  
715 along-strike variation in the geochemical signatures of extension versus subduction in  
716 the Kyrenia Range, although more geochemical work is-is needed, regionally and  
717 globally to test this possibility hypothesis.

718 Farther north, Eocene granitic rocks locally cut the continental Malatya  
719 metamorphic unit (Doğanşehir region) (Perinçek and Kozlu, 1984; Karaoğlan et al.,  
720 2013, 2016). However, no extrusive equivalents are known and, as noted above, the  
721 magmatic rocks in this lineament are unlikely to correlate directly with the Kyrenia  
722 Range.

723 During the Miocene, the active continental margin bordering the remnant Southern  
724 Neotethys collided with the Arabian passive continental margin to the south. In both  
725 the Kyrenia Range and SE Turkey, the Late Cretaceous and Paleogene magmatic  
726 rocks are exposed close the thrust front (Eurasia), implying that at least tens of kms of

727 fore-arc crust have been lost (subducted or overridden), potentially removing additional  
728 frontal arc-related crust, both Late Cretaceous and Paleogene. Thus, the Kyrenia,  
729 Kannaviou (W Cyprus) and ~~Helete-Maden Complex~~ magmatism could all represent  
730 surviving products of a regional ~~Late Cretaceous~~ active continental margin ~~are~~.

731

#### 732 5.5. Tectono-magmatic hypotheses

733 In the light of the regional setting, ~~t~~The felsic volcanics (Fourkovouno Formation,  
734 c. 74 Ma) with continental margin arc affinities, are interpreted to have resulted from  
735 northward subduction beneath continental crust, possibly the Alanya continental unit  
736 to the north (Çetinkaplan et al., 2016; Robertson et al., 2020), of the Taurides. ~~The~~  
737 felsic volcanism is characteristic of relatively advanced subduction, suggesting that  
738 earlier eruptive centers existed along the active continental margin. The early  
739 volcanism is likely to be represented by the Campanian (c. 80 Ma) arc-derived  
740 volcanogenic sediments (Kannaviou Formation) in western Cyprus (Robertson, 1977;  
741 Gilbert and Robertson, 2013; Chen and Robertson, 2019). The underlying mantle  
742 wedge (sub-continental lithosphere) was metasomatically enriched during the Late  
743 Cretaceous subduction, with implications for the subsequent basaltic volcanism.

744 For the Maastrichtian to Late Eocene basaltic volcanics, there are three alternative  
745 tectono-magmatic models. The first model envisages a marginal basin setting related  
746 to northward subduction. ~~The Southern Neotethys remained partly open between the~~  
747 ~~Taurides and the Troodos ophiolite during the Late Cretaceous to Mid-Eocene.~~  
748 ~~Subduction of the remnant ocean gave rise to the basaltic volcanism, including the~~  
749 ~~Misis-Andırın Range.~~ Compared to typical back-arc basin basalts (Fig. 14a-b), the  
750 ~~basalts of the~~ Kyrenia Range basalts (and the Misis-Andırın Range) have relatively  
751 high Th/Nb but low La/Nb ratios. The high abundances of Th are likely to reflect a  
752 continental crust ~~influence~~ input (Floyd et al., 1991). The variation in La/Nb ratios of the

753 basalts could reflect the influence of a within-plate mantle source (i.e. easterly basalts)  
754 and/or the arc maturity; (i.e. the arc-like western basalts could have erupted closer to  
755 the distal (oceanward) edge of the convergent margin). However, a problem with this  
756 model is the absence of evidence for a Paleogene volcanic arc to the south. There is  
757 no evidence of arc-derived fallout tuff or volcanoclastic sediments in the Maastrichtian-  
758 Paleogene basaltic succession, in contrast with, for example, the SW Pacific Mariana  
759 and Tonga marginal basins (Clift, 1994; Bryan et al., 2004).

760 The second model involves an extensional (or transtensional) setting. In this  
761 interpretation, the Late Maastrichtian and Paleogene 'enriched' within-plate volcanism  
762 was triggered by strike-slip (or transtension) along the ~~South Neotethyan~~ active  
763 continental margin. One option is that the volcanism was linked to the anticlockwise  
764 paleorotation of the Troodos ophiolite to the south during Late Campanian-Early  
765 Eocene (Clube et al., 1985; Clube and Robertson, 1986; Morris et al., 2006, 2015;  
766 Hodgson et al., 2010). More generally, oblique convergence of the African plate with  
767 respect to Eurasia could have resulted in segmentation of the South Neotethyan active  
768 margin into subducting segments with arc magmatism, and also strike-slip segments  
769 without arc magmatism (Aktaş and Robertson, 1990). A modern example is the  
770 Andaman Sea where oblique convergence between the Asian and the Australian  
771 plates resulted in segmented arc magmatism (e.g. Curray et al., 2005). However, there  
772 is limited information from different modern settings to test this hypothesis.

773 The third, preferred, scenario is that both the felsic and the basaltic volcanism are  
774 related to northward subduction during stages in the closure of the ~~Southern~~  
775 ~~Neotethys-Mesozoic ocean basin to the south~~. The Late Cretaceous magmatism can  
776 then be explained by regional northward subduction (nearly orthogonal) resulting in  
777 continental margin arc volcanism. Slab-derived and residual slab components (fluid,  
778 melt) were released into the upper mantle lithosphere (e.g. Pearce, 1983; Harangi et  
779 al., 2007). The Late Maastrichtian and Paleogene basaltic volcanics then erupted in

780 an extensional (or transtensional) setting, perhaps related to slab rollback and incipient  
781 marginal basin formation. Several different mantle sources are likely to have  
782 contributed along the Kyrenia Range. The variable subduction-related signatures were  
783 inherited from the previously modified mantle lithosphere in this interpretation.  
784 Examples of subduction-influenced settings that were affected by later extension-  
785 related melting include the Rio Grande Rift, west USA (e.g. Riecker, 1979) and the  
786 Neogene north Aegean-west Anatolia volcanic province (e.g. Şengör and Yılmaz, 1981;  
787 Ersoy and Palmer, 2013). In summary, the available evidence is consistent with Late  
788 Maastrichtian-Paleogene incipient marginal basin formation along the northern, active  
789 margin of ~~the Southern Neotethys~~ an ocean basin (Southern Neotethys) that did not  
790 close until the Miocene.

#### 791 5.6. Testing alternative tectonic models

792 Finally, the Late Cretaceous-Paleogene volcanism in N Cyprus and SE Turkey  
793 can be used to test alternative plate tectonic models, including those by Robertson et  
794 al. (2012a), Karaoğlan et al. (2013, 2016), Maffione et al. (2017), Barrier et al. (2018),  
795 McPhee and van Hinsbergen (2019) and van Hinsbergen et al. (2020), as summarised  
796 in Figure 15.

797 In Reconstruction 1 (Fig. 15a-c) (Robertson et al., 2012a), the Southern Neotethys  
798 rifted in the Triassic, while Paleotethys evolved into the Northern Neotethys (İzmir-  
799 Ankara-Erzincan ocean). The Kyrenia Range is restored to the northern passive  
800 margin of the Southern Neotethys. Both intra-oceanic and continental margin  
801 subduction zones (northward-dipping) were active in the Southern Neotethys during  
802 the Late Cretaceous. The continental margin arc volcanism was constructed on, or to  
803 the south of one, or more, continental blocks (microcontinents), represented by the  
804 Kyrenia-Alanya-Pütürge-Bitlis lineament in the south and the Malatya-Keban  
805 lineament farther north. This reconstruction is compatible with our new evidence, ~~while~~

806 ~~prompting questions concerning the implications of two subparallel arc magmatic belts~~  
807 ~~(outside the present scope).~~

808 Reconstruction 2 (Fig. 15d-f) (Barrier et al., 2018) has many common features  
809 with (1), in particular the arc volcanism along the northern margin of the Southern  
810 Neotethys. However, the Kyrenia, Malatya and Pütürge-Bitlis massifs are restored as  
811 a single elongate continental unit rather than as microcontinents, as in (1). Also, little  
812 or no oceanic crust is inferred between the southerly continental unit and the Tauride  
813 carbonate platform to the north (i.e. Göksun and related ophiolites). This is difficult to  
814 reconcile with the Late Cretaceous and Eocene granitic magmatism cutting the  
815 Mesozoic Malatya (Keban) platform unless large-scale strike-slip and terrane  
816 duplication has taken place for which there is little supporting evidence.

817 In Reconstruction 3 (Fig. 15g-h) (Maffione et al., 2017; van Hinsbergen et al.,  
818 2020), the Kyrenia Range is restored to the southern passive margin of the Mesozoic  
819 Tethyan ocean. Late Cretaceous ophiolites that are presently located in several c. E-  
820 W, sub-parallel belts (i.e. N and S of the Bitlis and Pütürge massifs) originated above  
821 a single subduction zone that originated within Neotethys, dipping generally to the east  
822 or northeast (Stampfli and Borel, 2002; Moix et al., 2008). The subduction zone rolled  
823 back generally westwards until it collided with the Tauride and Arabian continents. In  
824 the south, the subduction zone broke through the oceanic gap between the African  
825 and Tauride continents, 'invaded' and replaced the Southern Neotethys with Late  
826 Cretaceous supra-subduction zone fore-arc crust. Fore-arc ophiolites were then  
827 obducted generally northwards and southwards over the opposing continents (van  
828 Hinsbergen et al., 2020). Two alternatives ~~have been~~are suggested; in one, the Bitlis  
829 and Pütürge continental units are correlated with Arabia (Maffione et al., 2017),  
830 whereas in the other these units are correlated with the southern margin of the Tauride  
831 ~~continents-continent~~ (van Hinsbergen et al., 2020). The Kyrenia Range (Trypa Group)  
832 is placed on the N African continental margin in both alternatives. The metamorphism



833 of the Mesozoic carbonate platform in the Kyrenia Range (Trypa Group) resulted from  
834 northeastward underthrusting/subduction beneath the obducted forearc Troodos  
835 ophiolite (van Hinsbergen et al., 2020). However, there are several problematic  
836 aspects: (1) seismic tomography in the Eastern Mediterranean supports generally  
837 northward subduction (e.g. Hafkenscheid et al., 2006), without evidence of a relict  
838 southward-dipping slab, as implied by the inferred northward emplacement of some  
839 Late Cretaceous ophiolites in S Turkey (~~it is however unclear for how long such a trace~~  
840 ~~would be preserved~~); (2) it does not allow for the tens to hundreds of km of subduction  
841 required to produce the Late Cretaceous arc magmatism in N Cyprus and SE Turkey.  
842 ~~Specifically, the magmatically crosscut~~ Malatya metamorphic unit, ~~cut by Late~~  
843 ~~Cretaceous granites, is instead~~ restored to the lower, downgoing plate (i.e. ~~beneath~~  
844 ~~the south~~ margin of ~~the~~ Tauride continents) ~~implying-suggesting-~~ northward  
845 subduction; (3) the model (van Hinsbergen et al., 2020) does not allow for any Late  
846 Cretaceous continental arc-related magmatism in ~~northern~~ Cyprus, SE Turkey or Iran  
847 (e.g. Agard et al., 2011); (4) in this model, the ~~proposed~~ Late Eocene-Miocene (post  
848 45 Ma) northward subduction is ~~also-~~ too young to explain the Late Cretaceous-  
849 Paleogene magmatic rocks in the Kyrenia Range and SE Turkey; (5) arc magmatism  
850 of Oligocene-Miocene age in S Turkey is absent, although this would be expected if  
851 significant northward subduction took place during this time, ~~as in model 3.~~

852

## 853 6. Conclusions

- 854 ● Latest Cretaceous mainly felsic volcanic rocks and latest Cretaceous-Paleogene  
855 mainly basaltic volcanic rocks are exposed throughout the Kyrenia Range,  
856 separated by a thrust.
- 857 ● Mapping and logging of the key, well-exposed ~~segment of the~~ western Kyrenia  
858 Range ~~segment, N Cyprus-~~ shows that the Late Cretaceous felsic arc volcanics

859 occur as variably dismembered thrust sheets in the southerly, frontal part of thrust  
860 belt.

861 ● The primary eruptive age of the felsic arc volcanics (Fourkovouno Formation) is  
862 ~~determined as~~ c. 74.0 Ma (Late Campanian), based on U-Pb dating of zircons.

863 ● The latest Cretaceous-Paleogene basalts vary geochemically along the Kyrenia  
864 Range (E-W). The volcanics in the east resemble normal rift products. In contrast,  
865 the volcanics in the west have a subordinate subduction-related signature (e.g.  
866 negative Nb anomaly).

867 ● Sr-Nd-Hf isotopic data for the latest Cretaceous-Paleogene basalts suggest that  
868 they were derived from several OIB-like mantle sources, ~~probably~~ with the  
869 involvement of a crustally-derived (recycled) component.

870 ● The Kyrenia Range felsic volcanics show some geochemical similarities with the  
871 late Cretaceous (~~93-83 Ma~~) arc-related rocks in S Turkey (Helete granite; 93-83  
872 Ma). These felsic volcanics are interpreted to represent a fragment of a mature  
873 magmatic arc that was active along a regional-scale active continental margin.  
874 Regional reconstructions suggest that the Helete granites and related ophiolitic  
875 rocks were emplaced southwards from an oceanic basin (Berit ocean) to the north.

876 ● The latest Cretaceous-Paleogene volcanics in the Kyrenia Range have  
877 geochemical similarities with the Middle Eocene (c. 47 Ma) basaltic rocks in S  
878 Turkey (Maden Complex), suggesting an along-strike continuation of the same  
879 active continental margin.-

880 ● The latest Cretaceous-Paleogene basaltic volcanics in the Kyrenia Range are  
881 interpreted as the products of an incipient marginal basin that developed during  
882 northward subduction, possibly related to oblique convergence, prior to collision  
883 with Arabia during the Miocene.-

884 ● ~~In the light of alternative regional~~ ~~Having taken account of a range of~~ tectonic  
885 models, the Late Cretaceous ~~and~~ Paleogene magmatic rocks ~~in Kyrenia Range,~~  
886 ~~of~~ N Cyprus are interpreted to represent stages in the development of the northerly  
887 active continental margin of the Southern Neotethys.

888

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1331 Listing of Figures

1332 Fig. 1 Outline tectonic map of the Eastern Mediterranean region, including the main  
1333 volcanic units mentioned in the text. The tectonic framework is modified after  
1334 Robertson et al. (2012a). Note the location of the Kyrenia Range in northern Cyprus  
1335 (red box). Jurassic, Cretaceous and Eocene igneous rocks are shown in blue, red and  
1336 yellow, respectively (see the text for literature sources).

1337 Fig. 2 Successions shown in two main thrust sheets exposed in the Kyrenia Range,  
1338 northern Cyprus. (a) Simplified log of the succession in the largest thrust sheet  
1339 including the latest Cretaceous-Paleogene volcanics (Melounda Formation); (b)  
1340 Restored log of the small, dismembered frontal thrust sheets that expose the Late  
1341 Cretaceous felsic volcanics (Fourkovouno Formation) (data from Baroz, 1979;  
1342 Robertson et al., 2012b and this study).

1343 Fig. 3 (a) Outline geological map of the Kyrenia Range (modified from Robertson et al.,  
1344 2012b). Locations from which basalts were collected are numbered on the map, with  
1345 small yellow boxes. Karpas Peninsula: 1-Balalan (Platanissos); Eastern Range: 2-  
1346 Ağıllar (Mandres), 3-Çınarlı (Platani), 4-Mallıdağ (Melounda); Central Range: 5-Tirmen  
1347 (Trypimeni), 6-Ergenekon (Agios Khariton), 7-Değirmenlik (Kythrea), 8-Arapköy  
1348 (Klepini), 9-Beylerbeyi (Bellapais), 10-Boğaz (Bogaz), 11-Pınarbaşı (Krini), 12-İncesu  
1349 (Motides), 13-Alevkaya Tepe (Kiparisso Vouno); Western Range, 14-Karşıyaka  
1350 (Vasileia), 15-Geçitköy (Panagra); (b) Geological map of the western Kyrenia Range,  
1351 northern Cyprus made during this study (based on mapping by Baroz, 1979). Locations  
1352 of cross sections (AA', BB' and CC') are indicated. Note that the two contrasting, mainly  
1353 felsic and mainly basaltic volcanic units are separated by a thrust fault (Robertson et  
1354 al., 2012b; this study; see Fig. 2).

1355 Fig. 4 (a) A-A' cross section of Geçitköy that is dominated by a south-verging,  
1356 recumbent anticline (Late Miocene-earliest Pliocene); (b) B-B' cross-section of  
1357 Selvilitepe showing southward imbrication of basalt/pelagic chalk and felsic

1358 volcanogenic rocks; (c) C-C' cross-section showing imbricated felsic volcanogenic  
1359 rocks.

1360 Fig. 5 Measured stratigraphic logs of partial successions in the felsic volcanics and  
1361 structurally associated units. The structural lower unit (logs a-e) is relatively intact,  
1362 whereas the higher unit (logs f-j) is characterised by the thrust intercalations with the  
1363 Trypa Group. Fig. 6 (a) Zr/Ti vs. Nb/Y diagram (after Pearce, 1996) and (b) Th vs. Co  
1364 diagram (after Hastie et al., 2007) for the Kyrenia Range felsic rocks; (c) Primitive  
1365 mantle-normalised multi-element spider diagram and (d) chondrite-normalised rare  
1366 earth element (REE) patterns of the felsic volcanics with reference data for upper  
1367 continental crust (UCC) and lower continental crust (LCC). Normalising values of  
1368 primitive mantle and chondrite after Sun and McDonough (1989), upper continental  
1369 crust (UCC) and lower continental crust (LCC) data are from Rudnick and Gao (2003),  
1370 and Hu and Gao (2008). The Chile volcanic arc granite data are after Pearce et al.  
1371 (1984).

1372 Fig. 7 (a) Zr/Ti vs. Nb/Y diagram (after Pearce, 1996) and (b) Th vs. Co diagram (after  
1373 Hastie et al., 2007) for the basaltic volcanics. Red symbols: Karpas Peninsula and  
1374 eastern range; green symbols: central range; blue symbols: western range. Paleogene  
1375 basalt samples are indicated by thick magenta-outlined symbols.

1376 Fig. 8 Mid-ocean ridge basalt (MORB)-normalised multi-element spider diagram for  
1377 basalts from the Karpas Peninsula and the eastern (a), central (b) and western (c)  
1378 Kyrenia Range. MORB-normalised data are from Pearce et al. (1983); (d) Chondrite-  
1379 normalised REE patterns of selected basalts. Chondrite, OIB and E-MORB data are  
1380 from Sun and McDonough (1989). Paleogene basalt samples are indicated by solid  
1381 black lines and magenta-outlined symbols.

1382 Fig. 9 (a) Cathodoluminescence images of zircon grains separated from the samples  
1383 of Geçitköy. Red circles mark inconsistent analyses that were omitted from ~~the~~ the

1384 weighted mean  $^{206}\text{Pb}/^{238}\text{U}$  age calculation. Locations of the ion probe analysis spots  
1385 and the corresponding ages ( $^{206}\text{Pb}/^{238}\text{U} \pm 1\sigma$ ) are indicated. Scale bar = 20  $\mu\text{m}$ ; (b-d)  
1386 Wetherill Concordia diagram for the zircons analyses from sample 14-18 (b), 14-19 (c)  
1387 and 14-20 (d).

1388 Fig. 10 (a-b) Plots of  $\text{P}_2\text{O}_5$  vs.  $\text{SiO}_2$  and Th, respectively, show that all of the samples  
1389 follow the I-type granite trend; (c) Zr vs.  $10000 \times \text{Ga}/\text{Al}$ , and (d) Nb vs.  $10000 \times \text{Ga}/\text{Al}$   
1390 discrimination diagrams of Whalen et al. (1987), showing the I-, S (sedimentary)- and  
1391 M (depleted mantle source)-type nature of the Late Cretaceous felsic volcanics.

1392 Fig. 11 (a)  $\epsilon\text{Nd}(t)$  vs.  $\epsilon\text{Sr}(t)$  diagram for the basaltic rocks analysed. The samples are  
1393 mainly Late Cretaceous ( $n=4$ ); Paleogene basalt samples ( $n=2$ ) are indicated by  
1394 magenta-outlined symbols. Comparative compositions are from Zindler and Hart  
1395 (1986); (b)  $\epsilon\text{Hf}(t)$  vs.  $\epsilon\text{Nd}(t)$  plot showing rocks for Kyrenia Range, northern Cyprus  
1396 relative to other volcanic rocks of the western Mediterranean (after Kempton et al.,  
1397 2018).

1398 Fig. 12 (a) Ba/Nb versus La/Nb diagram (after Zhang et al., 2013); (b) Nb/Y versus  
1399 Zr/Y diagram (after Condie, 2005), arrows in (b) indicate the effects of batch melting  
1400 (F) and the probable effect of fluids derived from subduction (SUB); (c) Zr/Nb vs. Nb/Th  
1401 diagram (modified after Condie, 2003). Abbreviations: PM, primitive mantle; DM,  
1402 shallow depleted mantle; HIMU, high  $m\mu$  (U/Pb) source; EMI and EMII, enriched  
1403 mantle sources; ARC, arc-related basalt; N-MORB, normal ocean ridge basalt; OIB,  
1404 oceanic island basalt; DEP, deep depleted mantle; EN, enriched component; REC,  
1405 recycled component; UCC, upper upper continental crust; LCC, lower continental crust;  
1406 BM, batch melting trajectory with percent melting noted. Numbers on mixing lines are  
1407 percents.

1408 Fig. 13 (a) Nb vs. Y diagram (Pearce et al., 1984) and (b) Th/Yb vs. Ta/Yb (Gorton and  
1409 Schandl, 2000) for the felsic volcanics; (c) Zr/Y vs. Zr diagram (after Pearce and Norry,  
1410 1979) and (d) V vs. Ti discrimination diagram (after Shervais, 1982) for the basaltic

1411 rocks. Abbreviations: WPG, within-plate granite; VAG, volcanic arc granite; syn-COLG,  
1412 syn-collisional granite; ORG, ocean ridge granite; WPB, within-plate basalt; MORB,  
1413 mid-ocean ridge basalt; BABB, back-arc basin basalt; ARC, arc-related basalt; OFB,  
1414 oceanic floor basalt.

1415 Fig. 14 (a) Th/Nb vs. La/Nb and (b) La/Nb versus Y diagrams (after Floyd et al., 1991).  
1416 Abbreviations: IAT, island arc tholeiite; BABB, back-arc basin basalt; OFB, oceanic  
1417 flood basalt; FAPB, fore-arc platform basalt; T/E-MORB, T-type/enriched-mid-ocean  
1418 ridge basalt.

1419 Fig. 15 Alternative tectonic models of Tethys in the Eastern Mediterranean region. (a-  
1420 c) Northward subduction with continental fragments rifted from Gondwana (Robertson  
1421 et al., 2012a); (d-f) Northward subduction and marginal basin formation (Barrier et al.,  
1422 2018); (g-h) Genesis of Late Cretaceous supra-subduction ophiolites at a single  
1423 subduction zone to the NE followed by roll-back of segments including the Southern  
1424 Neotethys (Maffione et al., 2017; McPhee and van Hinsbergen, 2019; van Hinsbergen  
1425 et al., 2020). Based on discussion (see text), model [a-1](#) is generally favoured.

1426

1427 Listing of Supplementary Figure

1428 Supplementary Figure S1 Field photographs of the felsic volcanics. (a) Exposure in  
1429 Geçitköy. Samples 14-18 and 14-19 are from this location; (b) Irregular-shaped  
1430 doleritic intrusion (hammer for scale); (c) Localised occurrence of greyish to greenish,  
1431 rhyolitic debris-flow unit, north of Geçitköy; (d) Greenish rhyolitic debris-flow unit with  
1432 angular rhyolitic clasts (pen for scale).

1433 Supplementary Figure S2 Field photographs of the basaltic volcanics. (a) Pillow lavas  
1434 and interstitial pink pelagic carbonate; roadcut c. 200 m north of Balalan; (b) basalt  
1435 interbedded with pelagic carbonate; thrust sheet above Miocene siliciclastic sediment;  
1436 c. 800 m northwest of Değirmenlik; (c) Basaltic volcanics intercalated with pelagic  
1437 carbonates and small slices of meta-platform carbonates of the Trypa Group along the  
1438 southern flank of the range, 800 m north of Boğaz; (d) Basalt-pelagic carbonate  
1439 intercalation, south of Karşiyaka.

1440 Supplementary Figure S3 Photomicrographs of felsic volcanics. (a) Subhedral  
1441 sanidine (Sa), plagioclase (Pl) and biotite (Bt), in a groundmass of devitrified glassy  
1442 volcanic shards (cross-polarised light); (b) felsic glass groundmass (Vg) with  
1443 microcrystalline quartz, irregular-shaped quartz (Q) and feldspar (Pl) (cross-polarised  
1444 light). Sample number is indicated in the bottom-left corner.

1445 Supplementary Figure S4 Photomicrographs of basalts. (a) Intersertal basalt with  
1446 euhedral granular augite (Px), elongate plagioclase laths (Pl) and opaque grains (Op)  
1447 (cross-polarised light); (b) Basalt with intersertal texture, in which randomly-oriented  
1448 plagioclase generally enclose pyroxenes (plane-polarised light); (c) Porphyritic basalt  
1449 with euhedral plagioclase and strongly altered olivine phenocrysts (cross-polarised  
1450 light); (d) Ophitic basalt with plagioclase phenocrysts, augite occurs as an inclusion  
1451 within individual plagioclase crystals (cross-polarised light). Sample number is  
1452 indicated in the bottom-left corner. Samples no. 21, 14-51 and 14-69 are from the  
1453 Melounda Formation, whereas sample no. 19-51 is from the Ayios Nikolaos Formation.



1454 Supplementary Figure S5 Tectonic discrimination plots. (a)  $\text{TiO}_2$ - $\text{MnO}$ - $\text{Na}_2\text{O}$  ternary  
1455 diagram (after Nisbet and Pearce, 1977) for clinopyroxenes in basalt; (b)  $\text{TiO}_2$ - $\text{SiO}_2$ -  
1456  $\text{Na}_2\text{O}$  diagram (after Beccaluva et al., 1989) for clinopyroxenes in basalt. Abbreviations:  
1457 VAB, volcanic arc basalt; OFB, ocean-floor basalt; WPA, within-plate alkali basalt;  
1458 WPT, within-plate tholeiitic basalt; BON, boninite; IAT, island-arc tholeiite; WPB,  
1459 within-plate basalt.

1460 Supplementary Figure S6 (a) Chondrite-normalised REE comparison patterns of the  
1461 Kyrenia Range felsic rocks and the Helete granites in SE Turkey (data after Nurlu et  
1462 al., 2016). Chondrite data after Sun and McDonough (1989); (b)-(c) MORB-normalised  
1463 multi-element spider diagrams for basalts. For comparison, the coloured fields show  
1464 the composition variations of basalts from the Kyrenia Range (this study), the Misis  
1465 Complex, S Turkey (Floyd et al., 1991), the Maden Complex, S Turkey (Erturk et al.,  
1466 2018), the Japan Sea (Chen et al., 2015), the Sarmiento Complex, Chile (Saunders et  
1467 al., 1979), the Bransfield Strait, Antarctic (Keller et al., 2002), the Guaymas Basin,  
1468 Caribbean (Saunders et al., 1982) and the Tyrrhenian Sea, Italy (Peccerillo, 2017).  
1469 MORB-normalised data are from Pearce et al. (1983). N-MORB, E-MORB and OIB  
1470 data are from Sun and McDonough (1989). Specifically, two representative basalts of  
1471 different affinities (within-plate vs. volcanic arc) from the Kyrenia Range are indicated.

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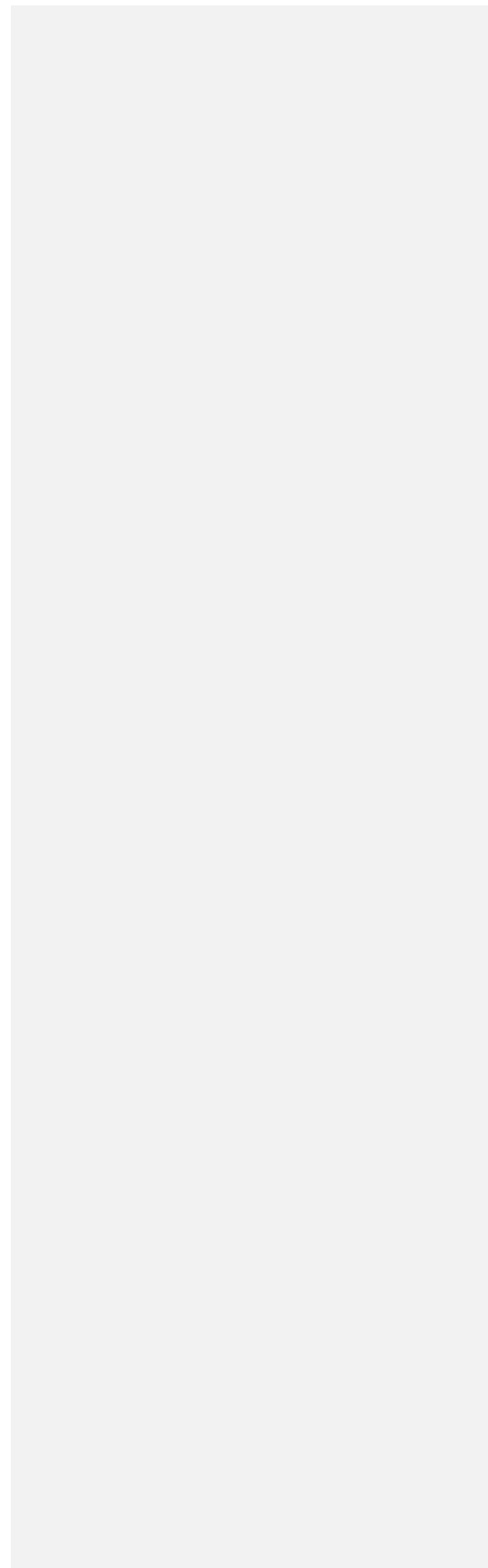
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1481 Listing of Supplementary Tables

1482 Supplementary Table S1. Major element oxides, trace and rare earth elements for the  
1483 Late Cretaceous Fourkovouno (Selvilitepe) Formation.

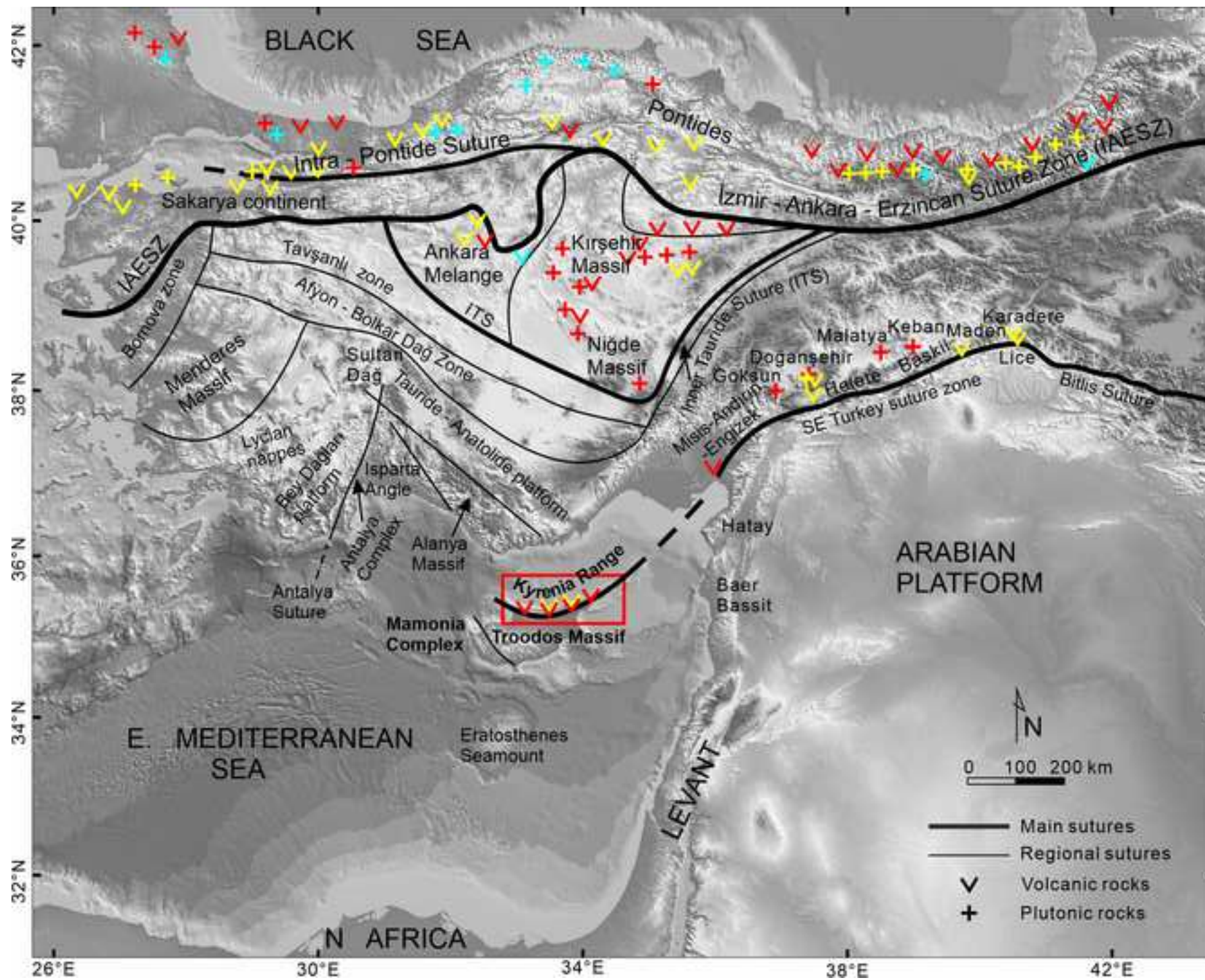
1484 Supplementary Table S2. Major element oxides, trace and rare earth element analyses  
1485 for the latest Cretaceous-Paleogene basaltic volcanics.

1486 Supplementary Table S3. Electron microprobe analyses of feldspar in the latest  
1487 Cretaceous-Paleogene basaltic volcanics.

1488 Supplementary Table S4. Electron microprobe analyses of pyroxene in the latest  
1489 Cretaceous-Paleogene basaltic volcanics.

1490 Supplementary Table S5. SIMS zircon U-Pb analyses of the zircon grains separated  
1491 from the Late Cretaceous Fourkovouno (Selvilitepe) Formation.

Figure 1



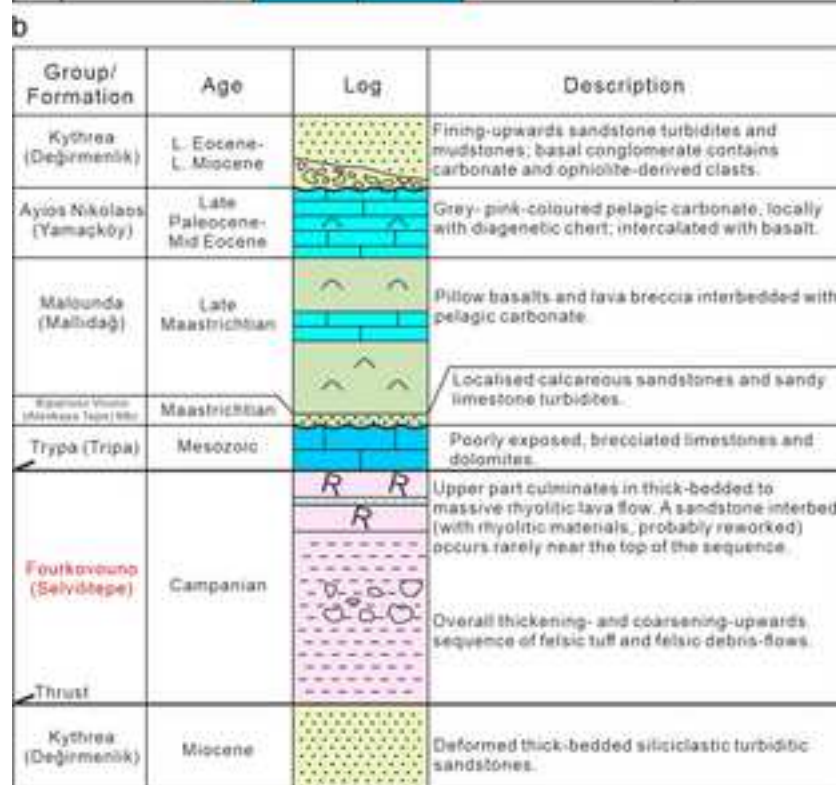
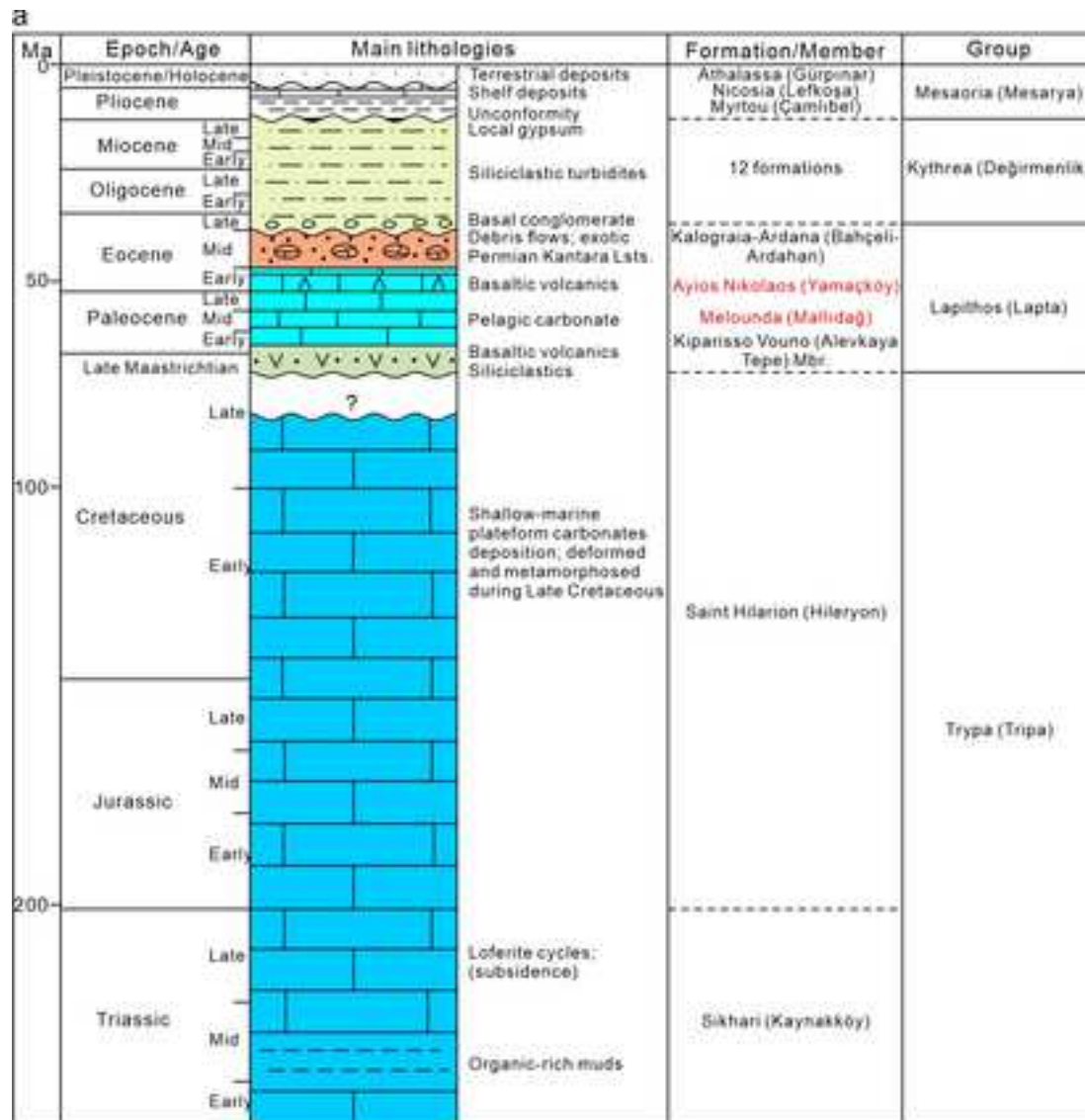
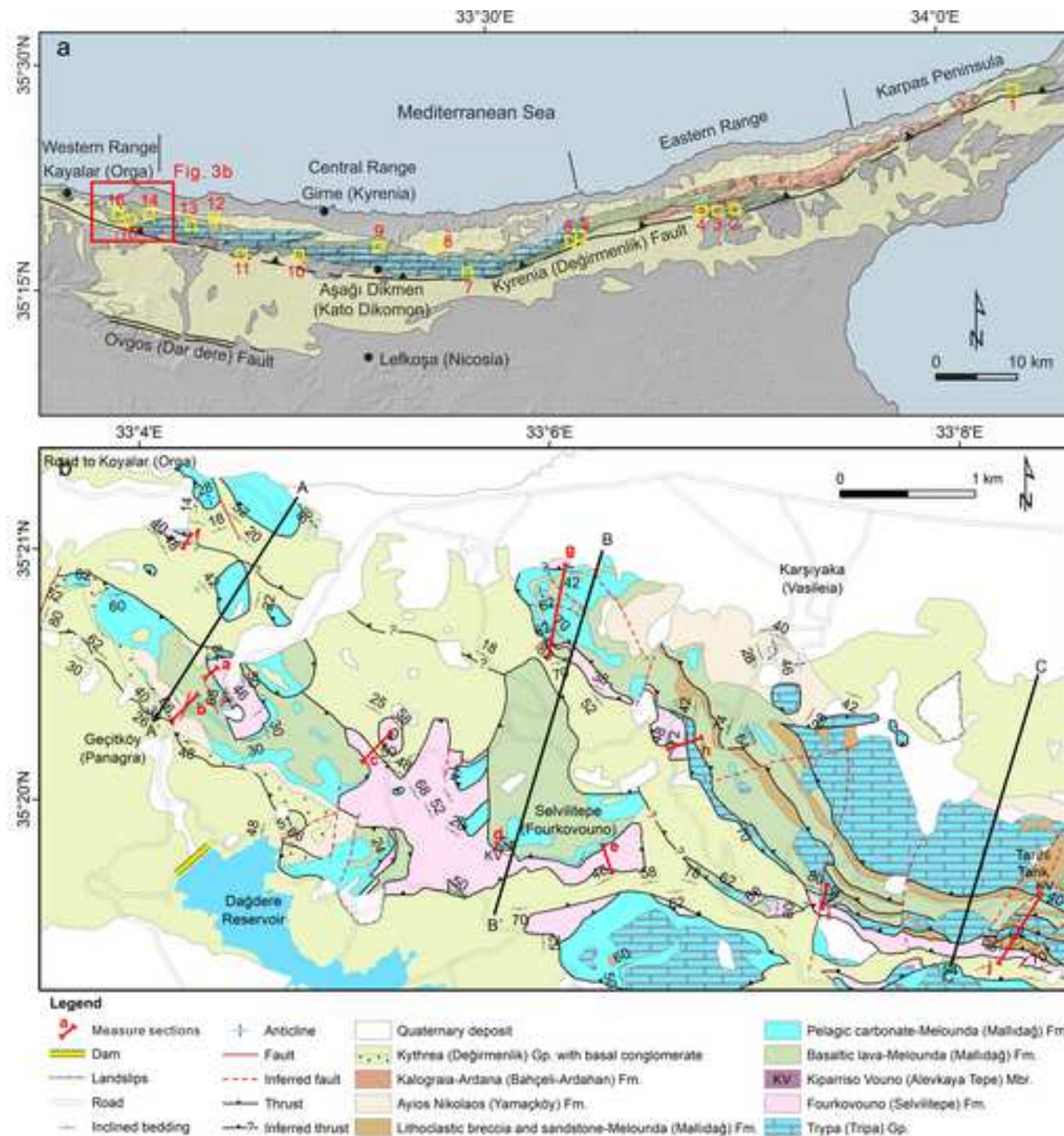


Figure 3

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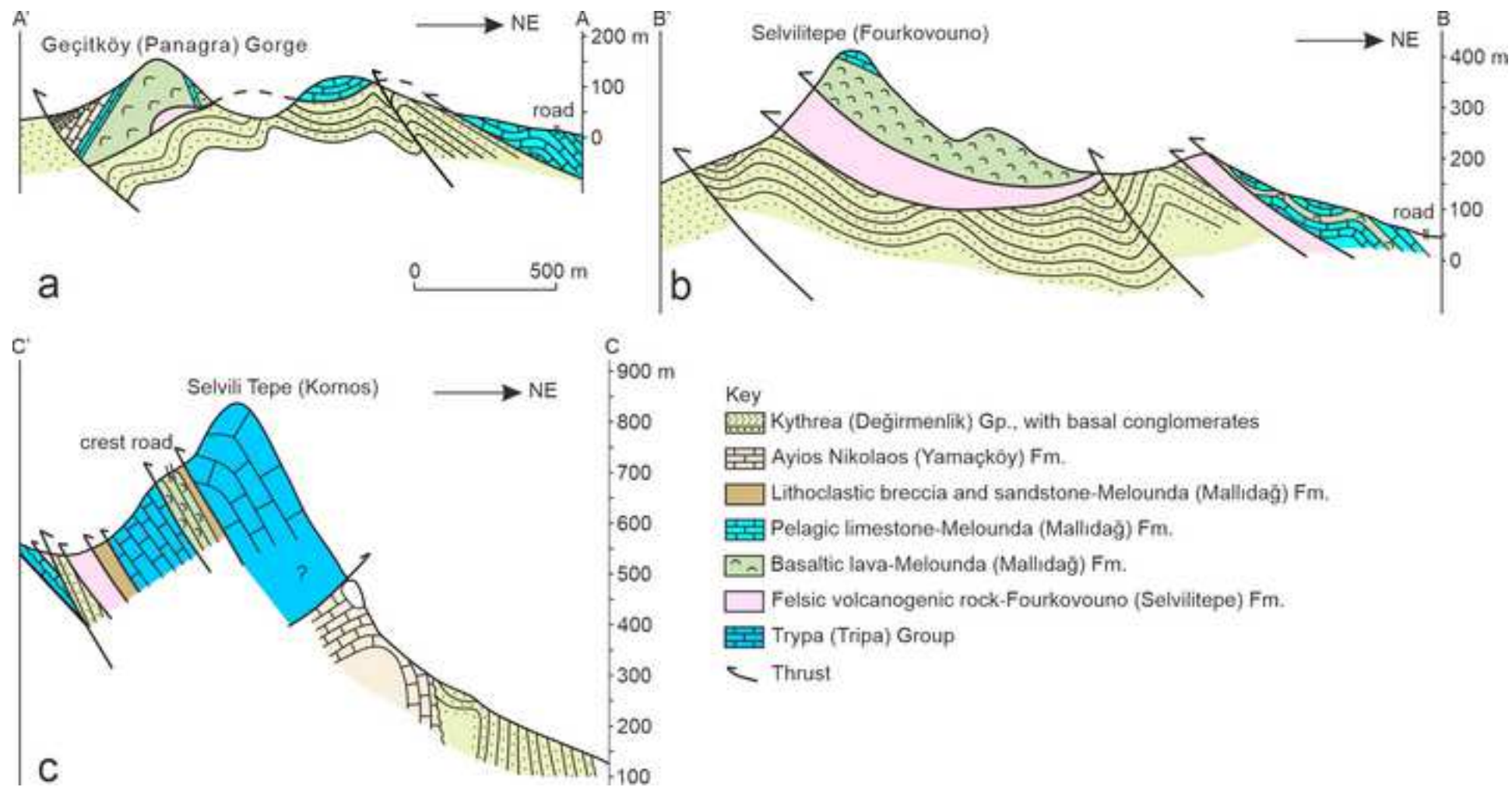
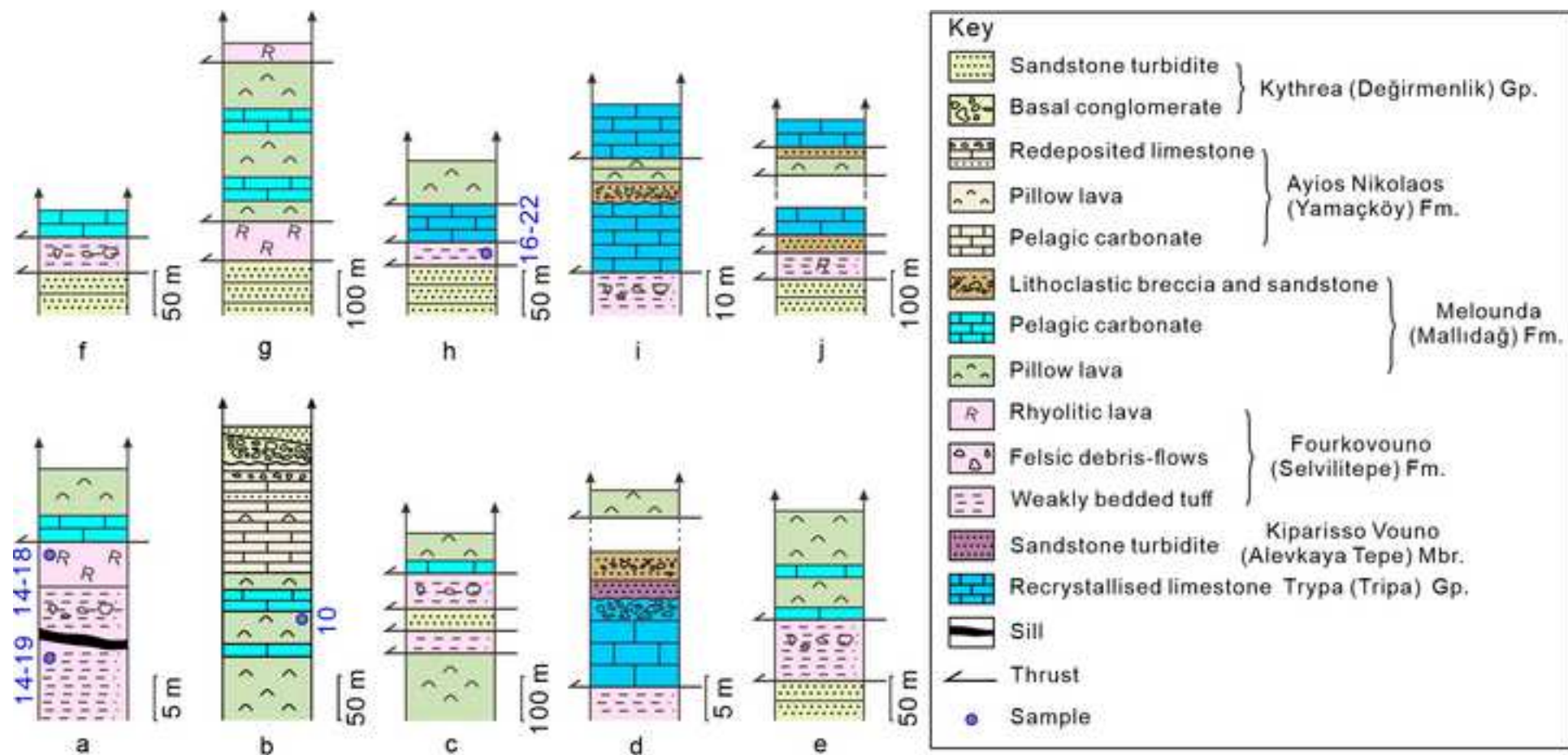
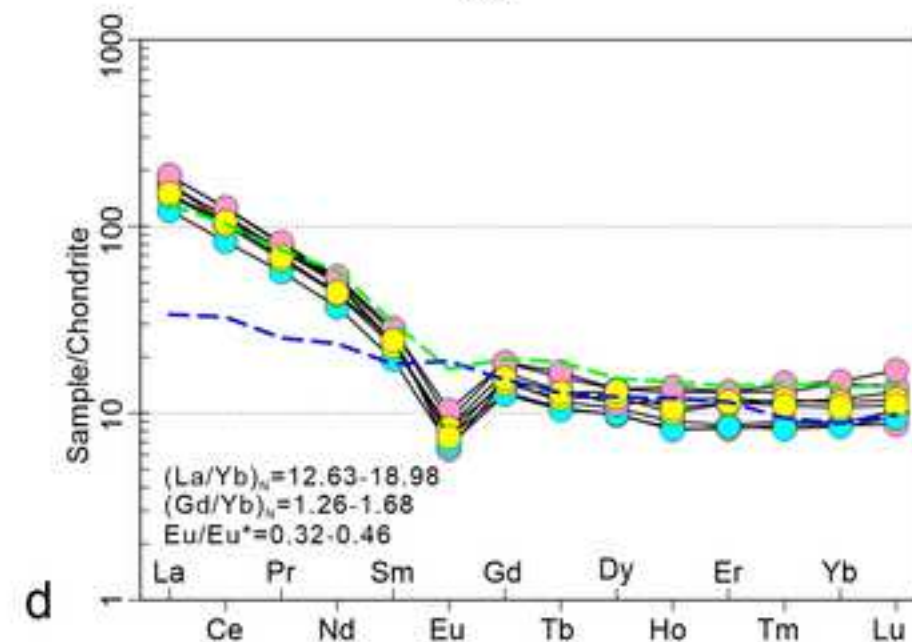
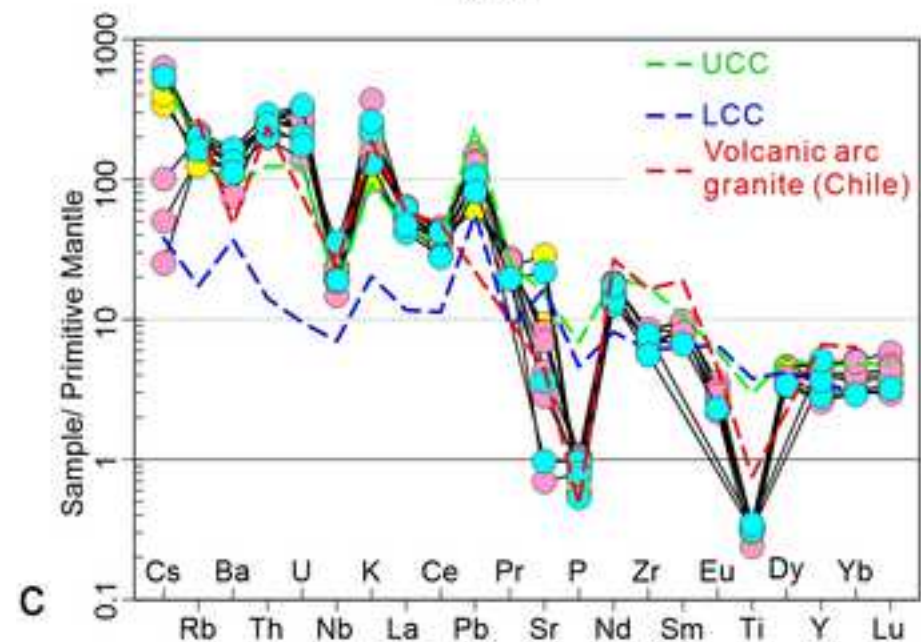
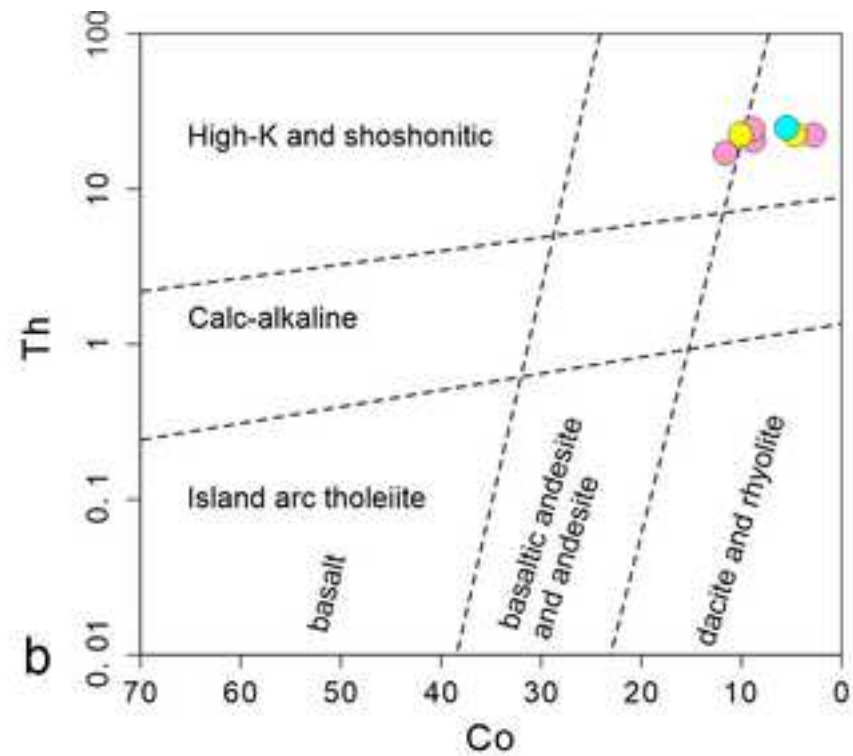
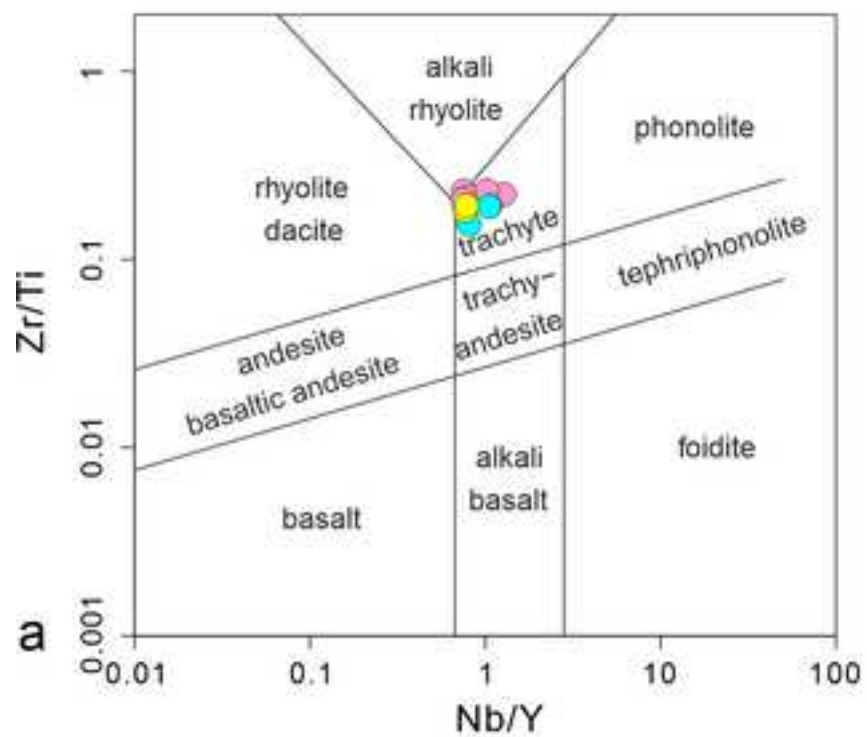


Figure 5







● Geçitköy (Panagra) ● Selvilitepe (Fourkovouno) ● Karşıyaka (Vasileia)

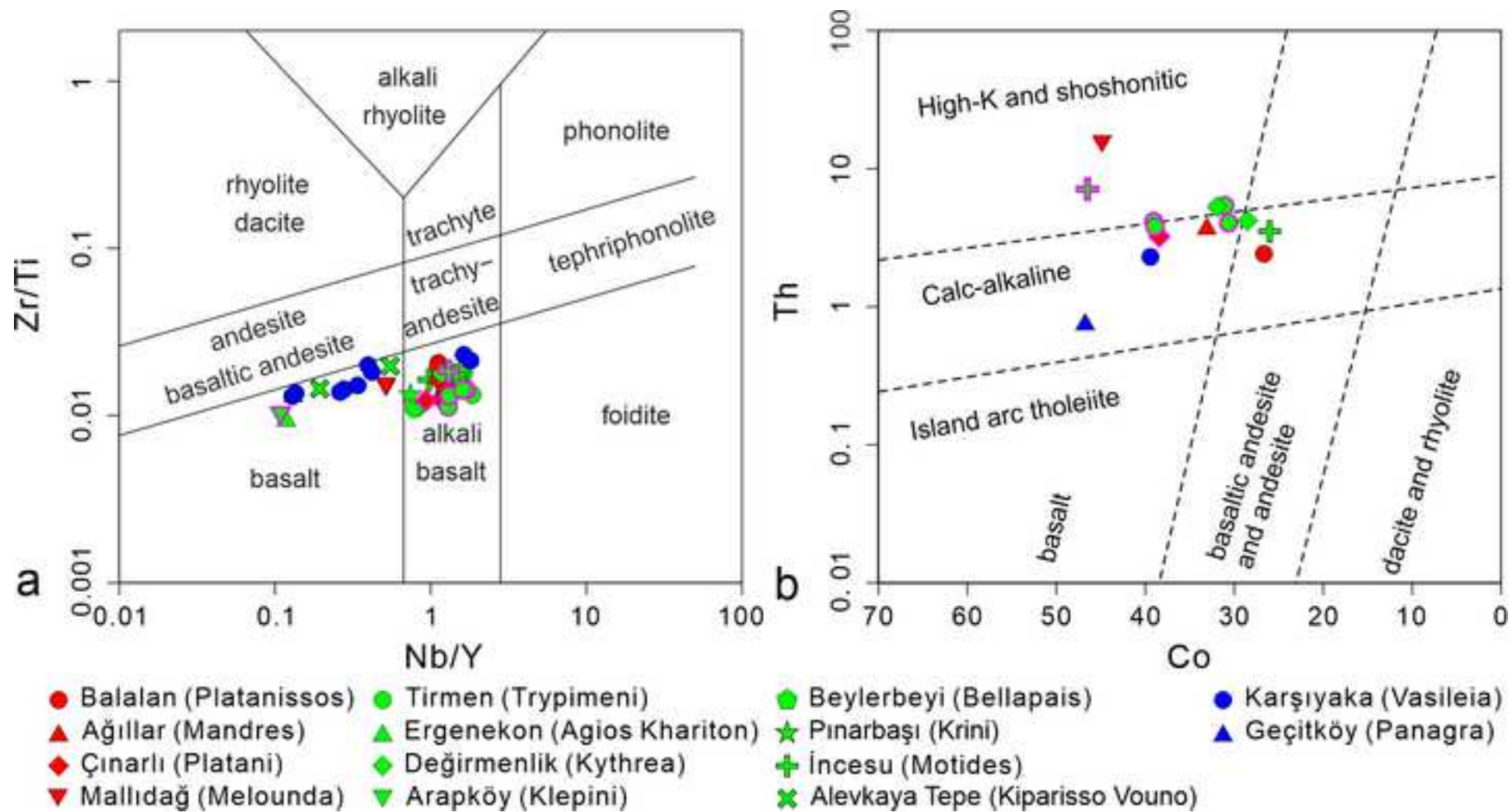
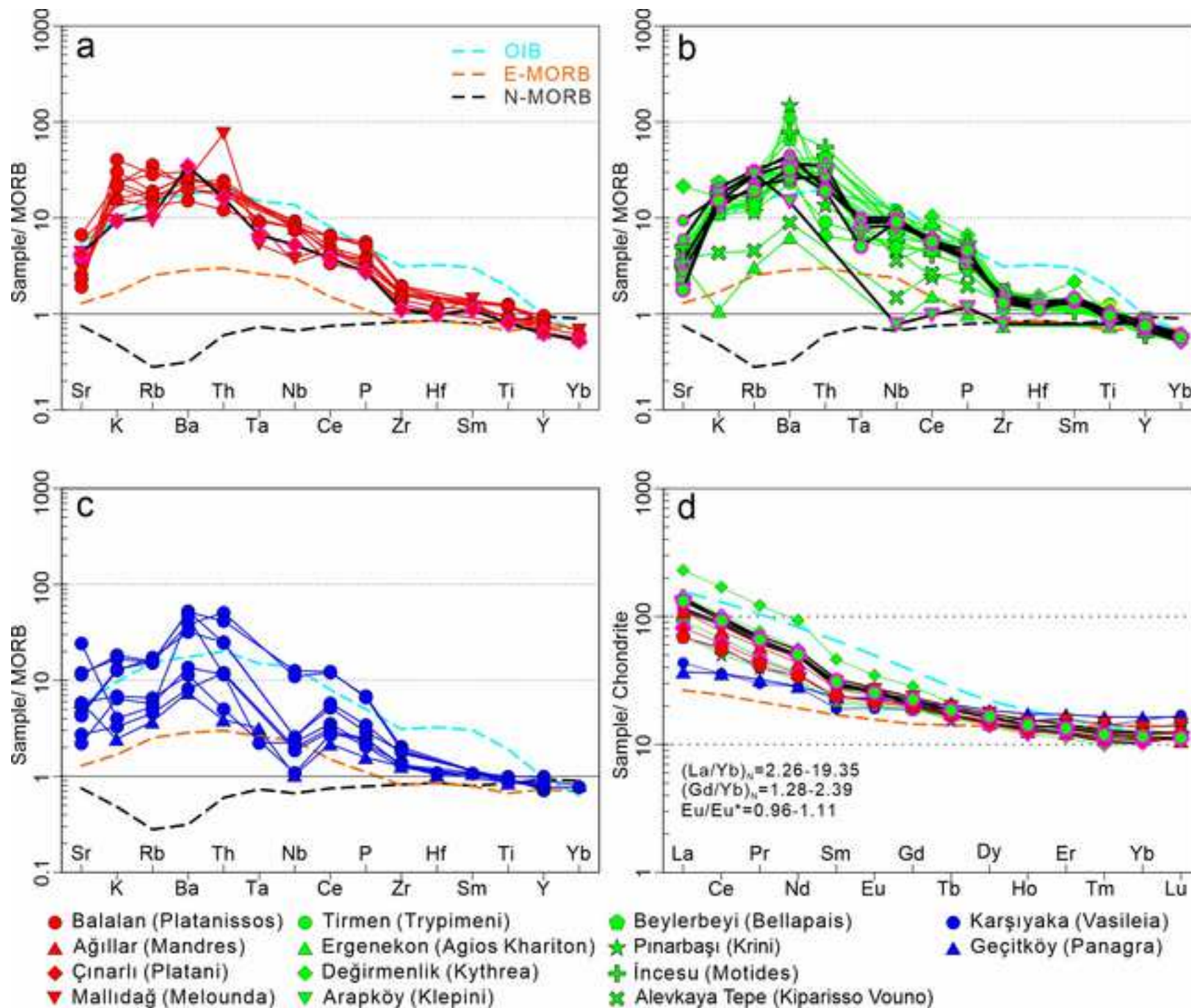
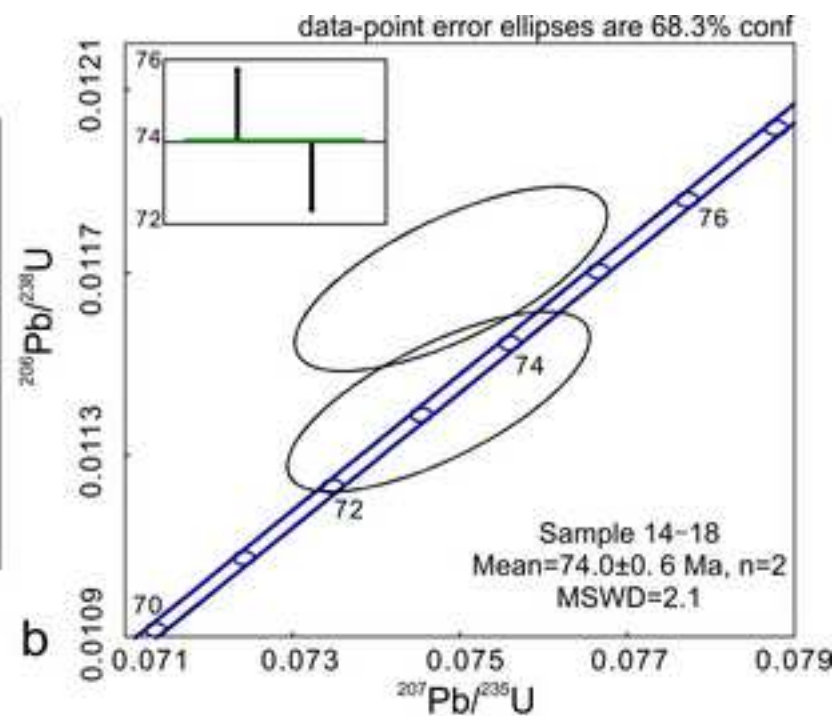


Figure 8

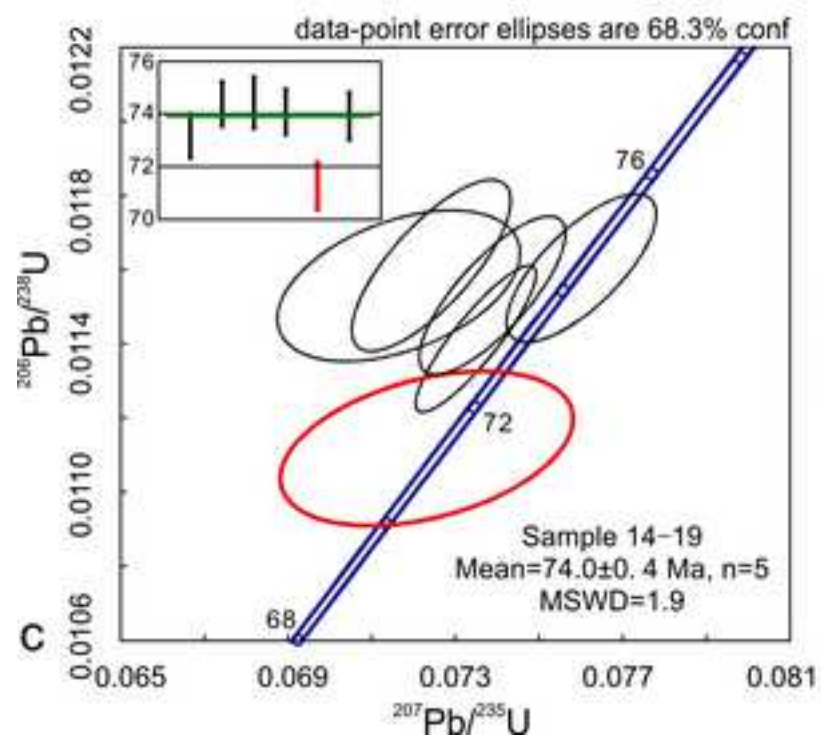




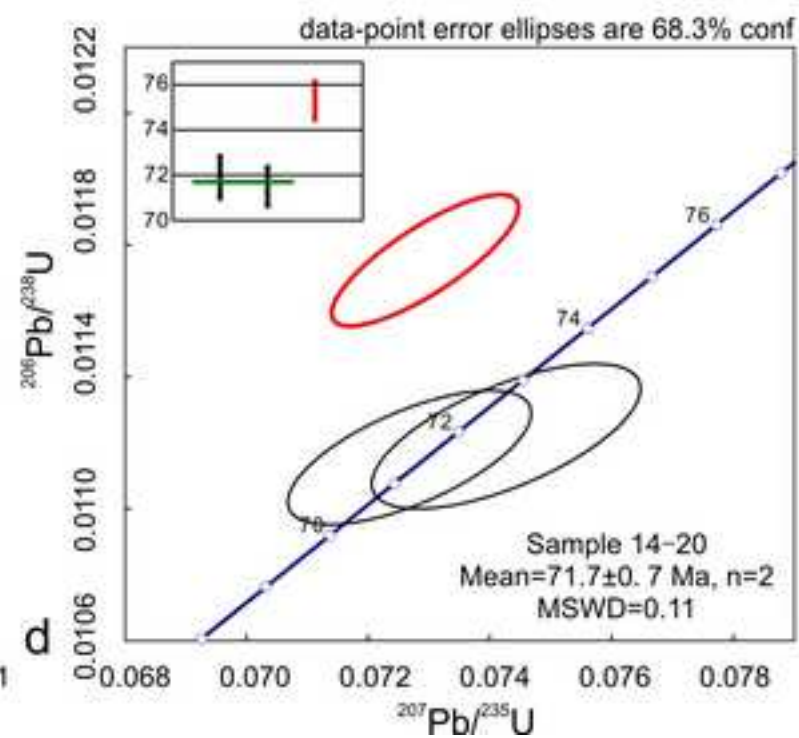
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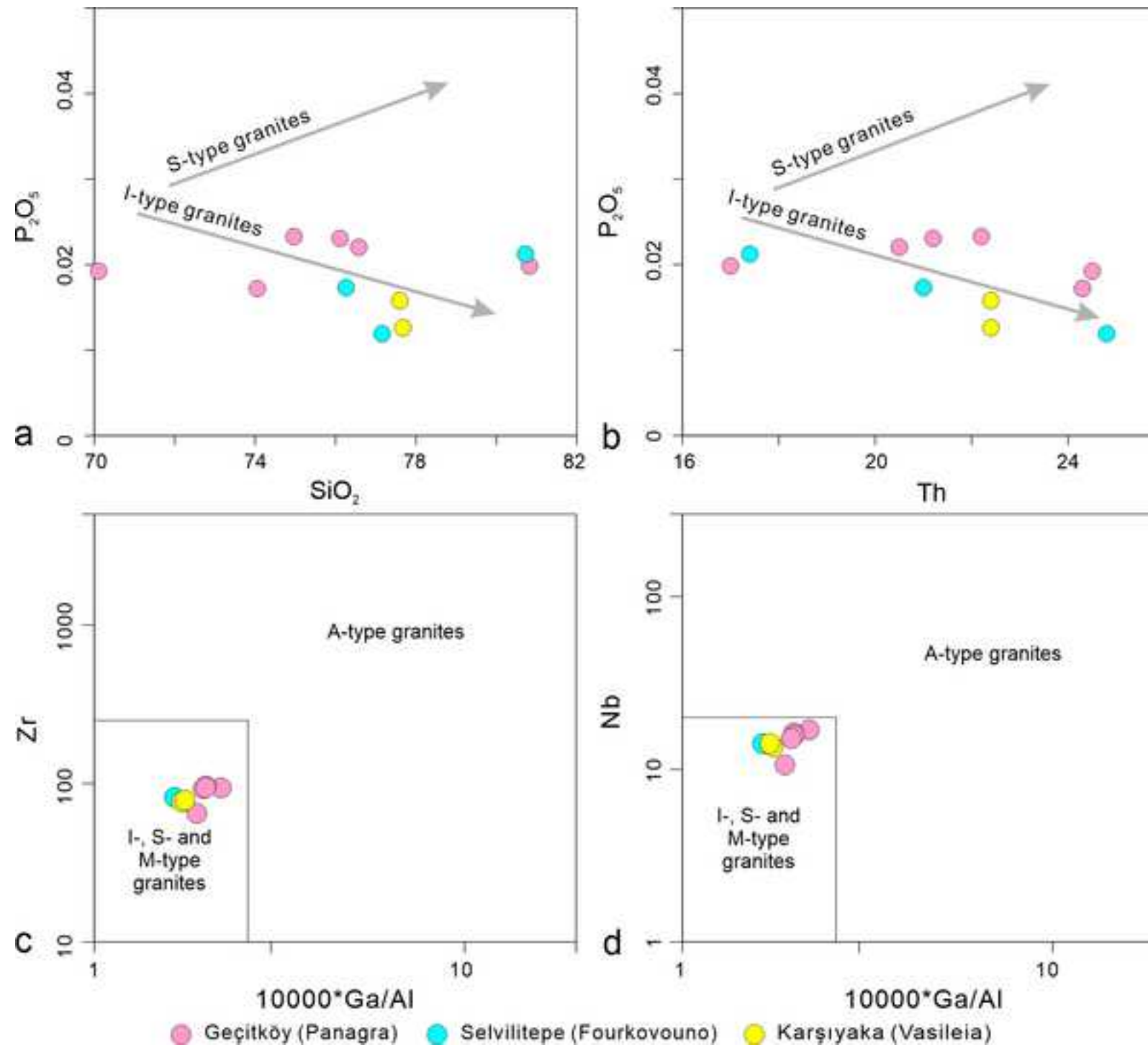
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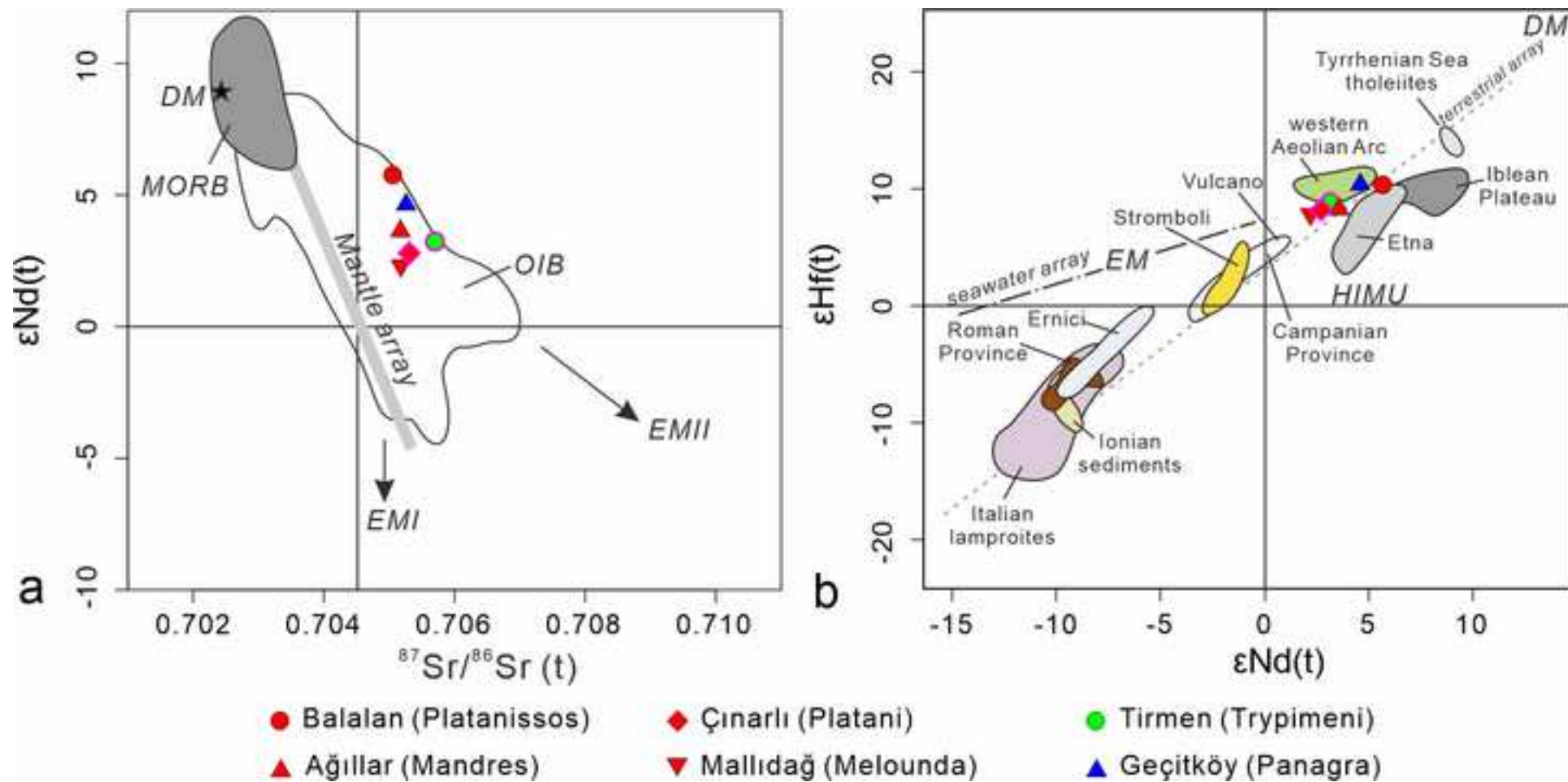


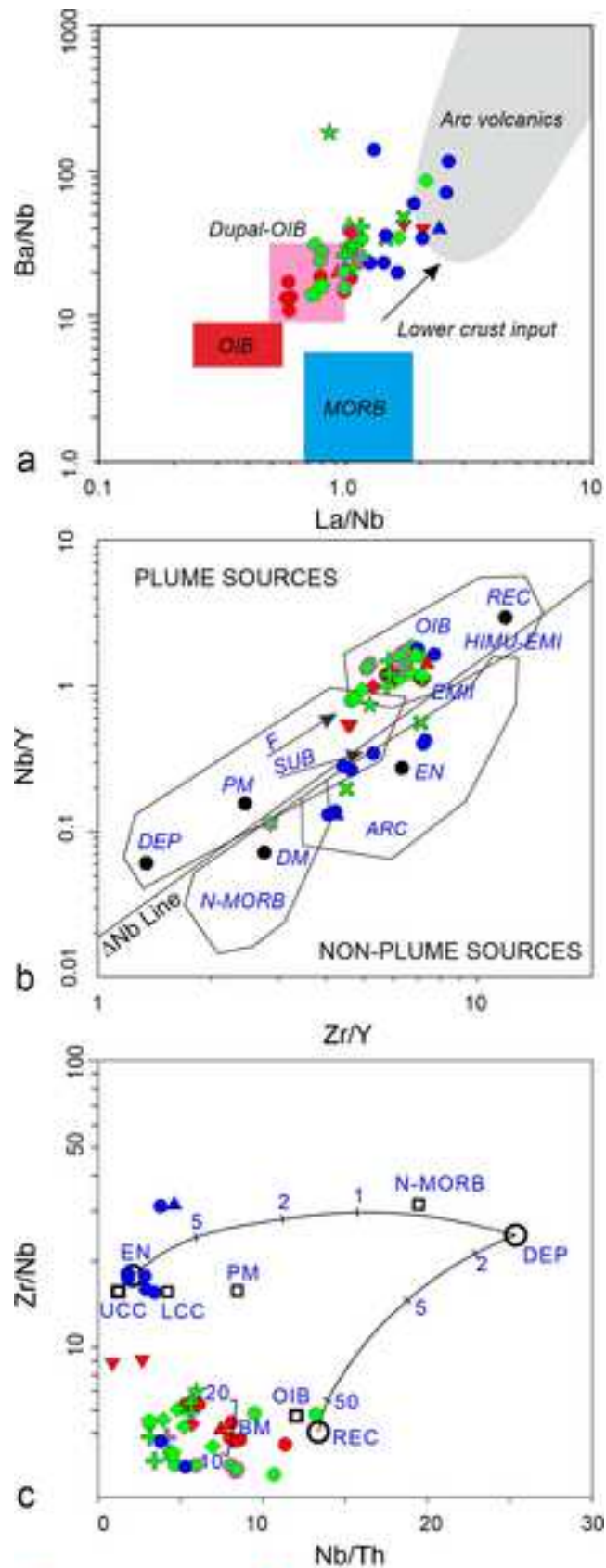
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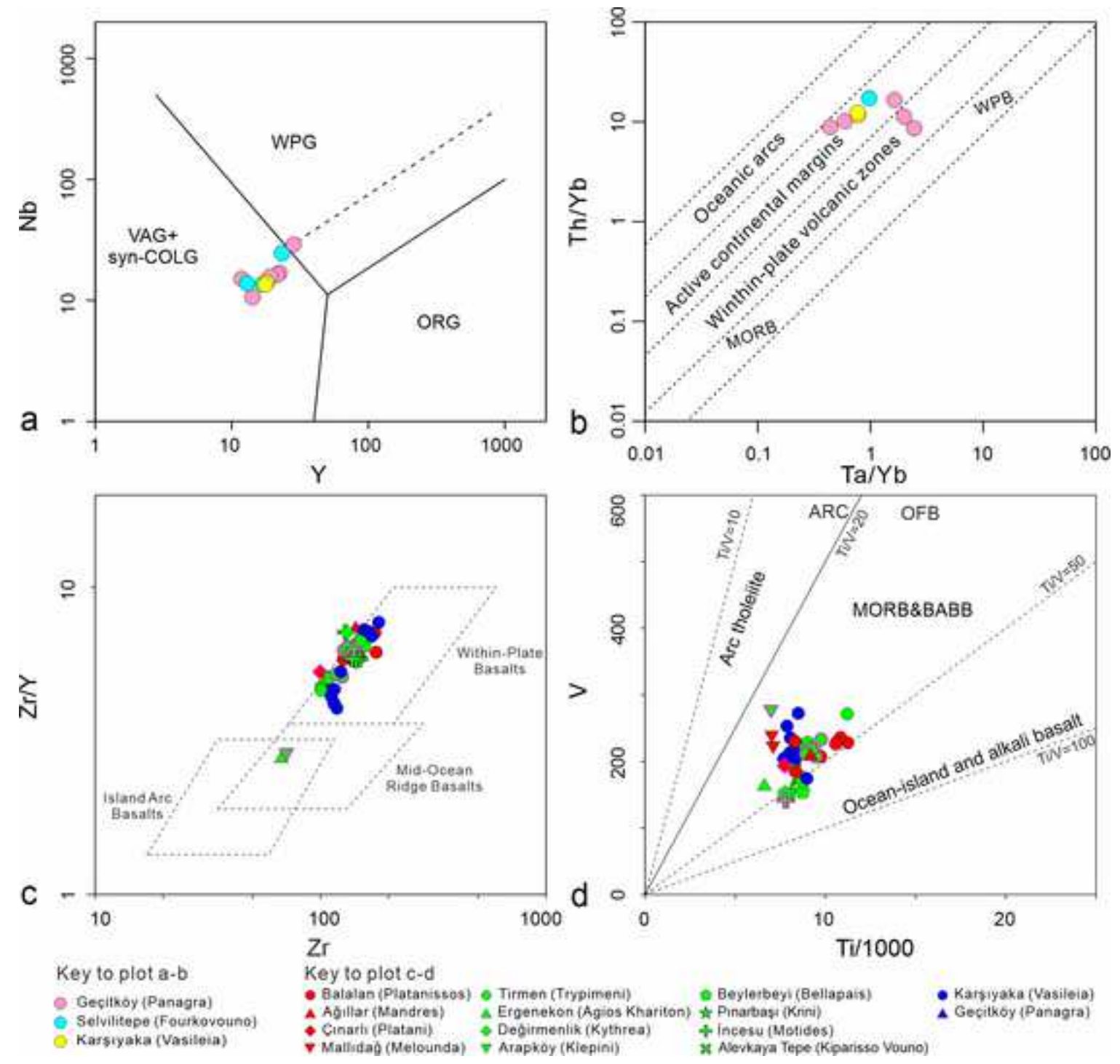
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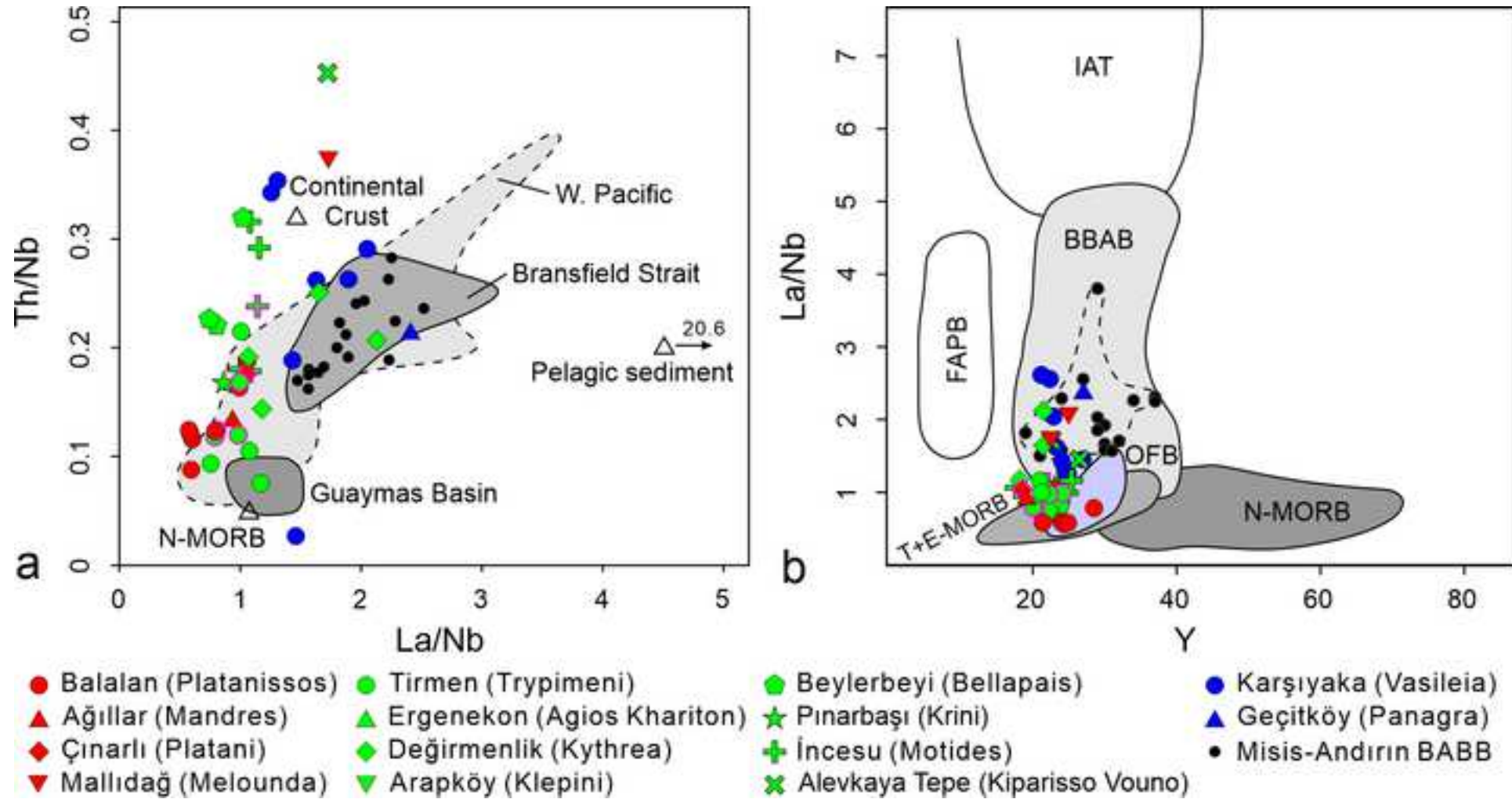


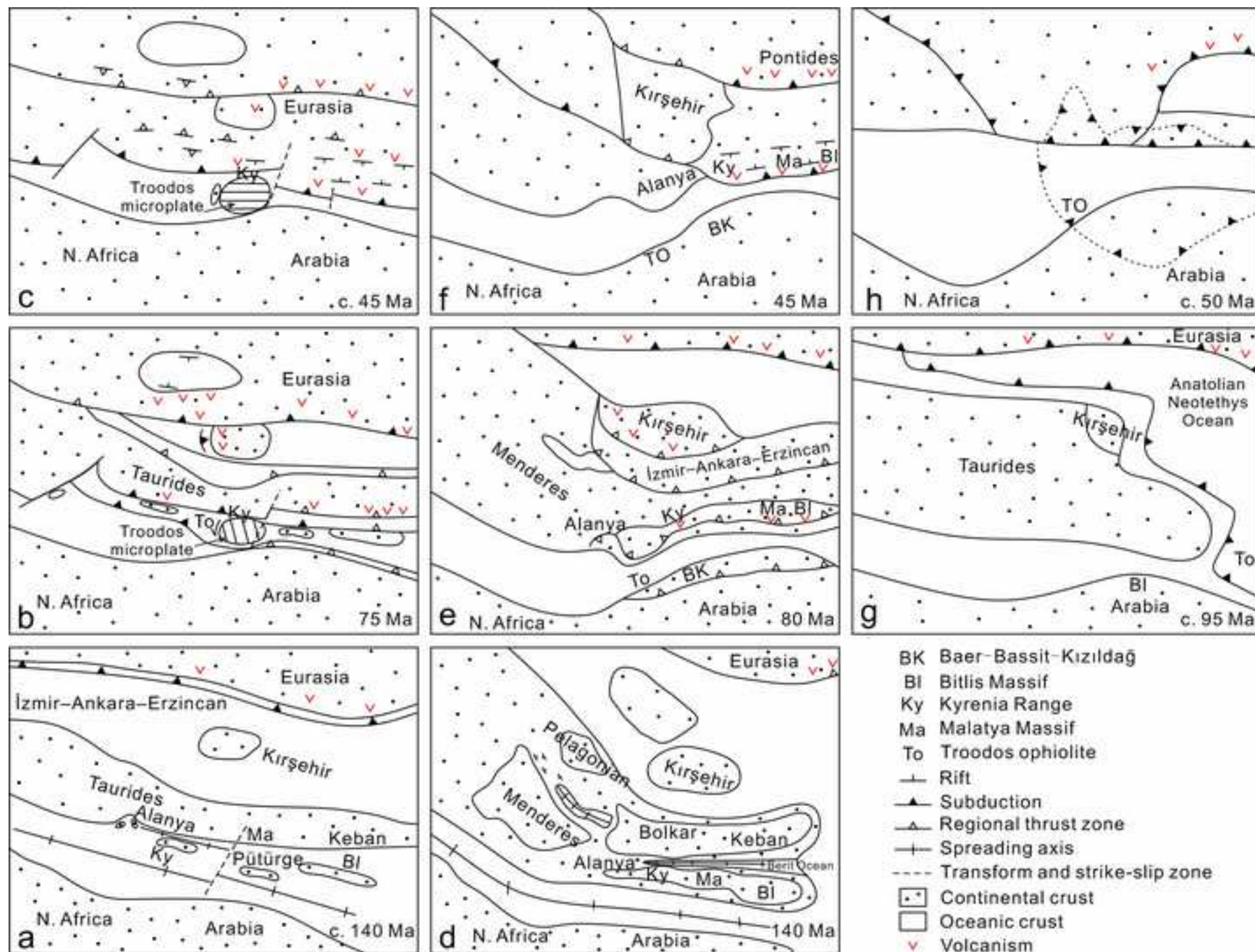


- Balalan (Platanissos)
- ▲ Ağıllar (Mandres)
- ◆ Çınarlı (Platani)
- ▼ Mallıdağ (Melounda)
- Tirmen (Trypimeni)
- ▲ Ergenekon (Agios Khariton)
- ◆ Değirmenlik (Kythrea)
- ▼ Arapköy (Klepini)
- Beylerbeyi (Bellapais)
- ★ Pınarbaşı (Krini)
- ◆ İncesu (Motides)
- × Alevkaya Tepe (Kiparisso Vouno)
- Karşıyaka (Vasileia)
- ▲ Geçitköy (Panagra)











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The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: