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**Provenance and magmatic-tectonic setting of Campanian-aged
volcaniclastic sandstones of the Kannaviou Formation in western Cyprus:
evidence for a Southern Neotethyan continental margin volcanic arc**

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Abstract

Regionally developed late Cretaceous subduction-related magmatism in the eastern Mediterranean records progressive closure of the Southern Neotethys. In western Cyprus the late Cretaceous (c. 90 Ma) Troodos ophiolite is depositionally overlain by volcaniclastic sandstones (up to 750 m thick) that are dominated by redeposited pyroclastic fallout, interbedded with both non-calcareous and calcareous radiolarian-bearing mudstones. The sands were mainly deposited by channelised mass flow processes and to a lesser extent by turbidity currents in a deep-water forearc basin. The succession was folded and locally thrust-deformed related to latest Cretaceous emplacement of adjacent Mesozoic continental margin and oceanic units (Mamonia Complex). SIMS U-Pb analysis of euhedral to subhedral magmatic zircon crystals yielded a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 80.1 ± 1.1 Ma (Campanian). Sandstone petrography including compositional framework ($\text{Q}_{12}\text{F}_{21}\text{L}_{67}$) suggests sediment supply from a volcanic arc. Subordinate continentally derived detritus suggests

a continental margin arc setting for the mafic to felsic fallout ash. SIMS analysis of little-altered volcanic glass indicates high Th/Nb and Th/La ratios that suggest the involvement of continental crust and/or subducted terrigenous sediments in magma genesis. Whole-rock chondrite-normalised REE patterns have comparable trends to modern and ancient forearc basin volcanoclastic sands and sandstones (e.g., northwest Pacific region). Trace element chemistry, although scattered towards the continental island arc fields on some geochemical diagrams (e.g., Th-Sc-Zr), mainly suggests an oceanic island arc source for the Kannaviou Formation sandstones, with variable enrichments in V-Cr-Ni-Sc, depletion in Nb-Ta, and relatively low trace element ratios (e.g., La/Co, Th/Co, La/Sc, Th/Sc). An explanation for this apparent discrepancy is that the volcanoclastic sandstones were derived from a relatively primitive arc constructed on previously depleted, rifted (thinned) Neotethyan continental crust. The inferred continental margin arc developed during early-stage northward subduction of Southern Neotethys beneath a Tauride microcontinent to the north. Large volumes of volcanic ash were derived from continental margin arc volcanism, probably in the vicinity of the Kyrenia Range in northern Cyprus.

Keywords: late Cretaceous, volcanoclastic sediments, continental margin arc, provenance analysis, Southern Neotethys, Cyprus

1. Introduction

Volcanic edifices are commonly localised and may be concealed or eroded, whereas their volcanoclastic products are commonly widely disseminated and easily recognisable in the stratigraphic record (e.g., Fisher and Schmincke, 1984; Yamamoto, 2009; Carey and Schneider, 2011). Volcanoclastic sediments (either primary or reworked) preserve petrographic and geochemical evidence that is indicative of magma genesis and tectonic setting of eruption.

Here, we present and discuss the results of a comprehensive study on the late Cretaceous volcanoclastic sandstones of the Kannaviou Formation in western Cyprus. We utilise a combination of field-based sedimentology, structural geology, thin-section petrography, and geochemical studies (whole-rock composition, mineral and glass composition) together with U-Pb dating of zircon, in the light of the regional geological setting. The main driver, building on previous studies, is to provide geochemical data for the volcanoclastic sandstones and to date the timing of the related volcanism, both of which are essential to understand the regional tectonic-magmatic setting. For example, it was unknown whether the volcanism relates to either continental or oceanic arc volcanism, and whether or not the source volcanism and the background depositional ages are synchronous. Specifically our objectives are: (1) to understand the deposition of the volcanoclastic sediments, taking account of the stratigraphy, structure and field sedimentary features; (2) to determine the magmatic setting of the source rocks using a combination of petrographic, mineralogical and geochemical studies; (3) to determine the absolute age of the related arc volcanism using U-Pb zircon dating; (4) to compare the Kannaviou Formation with modern and ancient counterparts; and (5) to propose

a new magmatic-tectonic model for the late Cretaceous magmatism in the context of the Southern Neotethys in the eastern Mediterranean region. Taken together, the results allow a significant advance in understanding in the tectonic-magmatic development of the Southern Neotethys in the eastern Mediterranean, with implications for comparable volcanoclastic sediments and tectonic settings elsewhere.

The most noteworthy feature of the Southern Neotethys in the eastern Mediterranean region is the late Cretaceous Troodos ophiolite (92-90 Ma) (Mukasa and Ludden, 1987). This is generally accepted to have formed during initial stages of northward subduction of the Southern Neotethys (Gass, 1968; Moores and Vine, 1971; Robinson and Malpas, 1990; Robertson, 1990, 2002; Robertson and Xenophontos, 1993; Pearce and Robinson, 2010). The volcanoclastic sedimentary rocks of the Kannaviou Formation depositionally overlie the Troodos ophiolite in the west of the island (Lapierre, 1968, 1975; Robertson, 1977a) and are therefore critical to interpretation of the regional tectonic setting (Fig. 1). Previous studies of the Kannaviou Formation provided field observations, whole-rock analysis, some mineral chemical analysis (Lapierre, 1975; Robertson, 1977a; Gilbert and Robertson, 2013) and paleontological dating (Urquhart and Banner, 1994; Lord et al., 2000). However, this is the first study to include evidence of geochemistry and volcanic source age. The source area of the Kannaviou Formation was previously suggested to be in the Kyrenia Range of northern Cyprus (Robertson, 1977a; Gilbert and Robertson, 2013) (Fig. 1a), a hypothesis that is tested here using new U-Pb dating of detrital zircons.

2. Geological setting

Below, we outline key features of the geology of Cyprus that are specifically relevant to an understanding of the late Cretaceous volcanoclastic sediments.

2.1. Setting of the Kannaviou Formation relative to the Troodos ophiolite

The Troodos Massif in central and southern Cyprus comprises the late Cretaceous Troodos ophiolite and its *in situ* sedimentary cover including the Kannaviou Formation. Based on paleomagnetic evidence, the Troodos ophiolite and its sedimentary cover underwent anticlockwise rotation of up to c. 90°, between the late Campanian and early Eocene (Clube et al., 1985; Clube and Robertson, 1986; Morris, 1996; Inwood et al., 2009). The Kannaviou Formation has to be rotated by 90° clockwise when interpreting its regional tectonic-magmatic formation.

The ophiolite is overlain, first by Fe-Mn-rich metalliferous sediments (umbers) of the Campanian Perapedhi Formation (up to c. 10 m thick) (Robertson and Hudson, 1974; Robertson, 1975). In western Cyprus, the Perapedhi Formation begins with localised umbers, as elsewhere in Cyprus (Robertson, 1977a), and passes upwards first into radiolarian mudstones, and then into volcanoclastic sandstones that represent the base of the Kannaviou Formation (Robertson, 1977a; Urquhart and Banner, 1994; Gilbert and Robertson, 2013).

The Kannaviou Formation mainly crops out around the western periphery of the Troodos Massif, from near Kannaviou village, the type area, northwards to around Istinjo and Kinousa (Lapierre, 1968) (Fig. 1b). The formation comprises alternating bentonitic claystone, radiolarian mudstone and

volcaniclastic sandstone, up to 750 m thick (Robertson, 1977a; Gilbert and Robertson, 2013) (Fig. 2). A Campanian-early Maastrichtian age has been determined mainly using radiolarians (Urquhart and Banner, 1994; Bragina and Bragin, 1995; Lord et al., 2000). In its type area, the Kannaviou Formation passes upwards into submarine debris-flow deposits, known as the Kathikas Formation (Swarbrick and Naylor, 1980) (Fig. 2). Local interbeds of pelagic chalk are dated as late Maastrichtian (Swarbrick and Naylor, 1980; Urquhart and Banner, 1994; Lord et al., 2000). The Kannaviou Formation and/or the Kathikas Formation in different local areas are conformably overlain by deep-sea pelagic carbonates of the Lefkara Formation, ranging in age from Maastrichtian to Oligocene (Robertson, 1976; Lord et al., 2000) (Fig. 2). Lateral facies equivalents of the Kannaviou Formation along the southern margin of the Troodos Massif are restricted to bentonitic clays and radiolarian siltstone, c. 160 m thick (Moni Member) (Robertson, 1977b; Swarbrick and Robertson, 1980). In contrast, the formation is entirely absent along the northern and eastern margins of the Troodos Massif (Robertson and Hudson, 1974), which has a bearing on the regional tectonic setting.

2.2. Mamonia Complex

The Mamonia Complex of western Cyprus is relevant here because it represents a likely source of terrigenous sediment to the late Cretaceous Kannaviou Formation. The Mamonia Complex is dominated by two different tectono-stratigraphic units (Fig. 1b) of late Triassic to early Cretaceous age: (1) Sandstone gravity-flow deposits, pelagic and redeposited limestone, chert, mudstone and radiolarite of the sedimentary Ayios Photios Group (Lapierre, 1975; Robertson and Woodcock, 1979; Swarbrick and Robertson, 1980; Torley

and Robertson, 2018); and (2) basaltic volcanics, together with volumetrically subordinate shallow-water carbonate, hemipelagic calcilutite, chert and mudstone of the Dhiarizos Group (Lapierre, 1975; Robertson and Woodcock, 1979; Swarbrick and Robertson, 1980; Malpas et al., 1992; Lapierre et al., 2007). The two stratigraphic groups are separated by tectonic contacts, and the Mamonia Complex as a whole has been juxtaposed with the Troodos Massif by a combination of thrusting and strike-slip faulting along steep arcuate fault lineaments during the late Cretaceous (Robertson and Woodcock, 1979; Swarbrick, 1980; Robertson, 1990; Morris et al., 1990; Bailey et al., 2000). The Mamonia Complex is restored as a part of the earliest-formed oceanic crust (Dhiarizos Group) and adjacent passive margin (Ayios Photios Group) of the Southern Neotethys that rifted from the northern margin of Gondwana (Robertson and Woodcock, 1979; Torley and Robertson, 2018). The volcanic rocks are also relevant to the composition of the source mantle lithosphere that may have affected the composition of subsequent arc magmatism in the region.

2.3. Transform faulting

Oceanic transform faulting is relevant to the Kannaviou Formation because it is inferred to have created a sea-floor topography that greatly influenced the deposition of the Kannaviou Formation during the late Cretaceous. The Troodos ophiolite in central-south Cyprus is cut by the east-west-trending South Troodos Transform Fault Zone, up to c. 15 km wide (Moores and Vine, 1971; Murton, 1986; MacLeod, 1990; Morris et al., 1990; MacLeod and Murton, 1993) (Fig. 1a). Transform faulting was active during the late Cretaceous genesis of the Troodos ophiolite at a spreading axis (Murton and Gass, 1986; MacLeod and Murton, 1993). Critically, the transform fault extends westwards beneath

the late Cretaceous-Cenozoic sedimentary cover and links with the arcuate fault-bounded lineaments in western Cyprus (Robertson, 1977a; Malpas et al., 1992, 1993; Morris et al., 1998) (Fig. 1b). The lineaments entrain fragments of late Cretaceous ophiolitic rocks and their sedimentary cover including the Kannaviou Formation (Robertson, 1977a; Gilbert and Robertson, 2013), together with slivers of late Cretaceous amphibolite facies and greenschist facies metamorphic rocks (Lapierre, 1975; Swarbrick, 1980, 1993; Malpas et al., 1992). The Troodos Massif including the Kannaviou Formation was tectonically juxtaposed with the Mamonia Complex during the late Cretaceous (pre- to syn-Maastrichtian) (Robertson and Woodcock, 1979; Swarbrick, 1980, 1993; Malpas et al., 1992, 1993; Morris et al., 1998; Bailey et al., 2000).

2.4. Cenozoic-recent development

The post-depositional geological development needs to be taken into account to understand the sedimentary development of the Kannaviou Formation. In particular, western Cyprus was affected by extension along a c. north-south axis during late Miocene to recent, related to formation of the Polis graben (Robertson, 1977c; Payne and Robertson, 1995, 2000; Balmer et al., 2019). The main Kannaviou Formation outcrop lies within the eastern part of the graben. The formation together with the Troodos Massif as a whole was uplifted and exposed mainly during the Pleistocene (Robertson, 1977c; Poole and Robertson, 1998; Morag et al., 2016). The effects of the Cenozoic faulting (e.g., relative movement of fault blocks) and the Pleistocene uplift (e.g., landslipping) were taken account of when interpreting the Kannaviou Formation.

2.5. Kyrenia Range

The Kyrenia Range in northern Cyprus is mentioned in this paper mainly because it is a possible source of the volcanoclastic sands that supplied the Kannaviou Formation. If correct, the tectonic setting of the Kyrenia Range during the late Cretaceous has a direct bearing on the provenance and tectonic-magmatic setting of the volcanism represented by the Kannaviou Formation. The Kyrenia Range is dominated by Triassic-Cretaceous platform carbonates that were metamorphosed under greenschist facies conditions during the late Cretaceous (Baroz, 1979). Late Cretaceous felsic volcanics (Fourkovouno (Selvitepe) Formation) (Robertson and Woodcock, 1986; Huang et al., 2007; Robertson et al., 2012; Chen, 2018) have been suggested as the source of the Kannaviou Formation volcanoclastic sediments (Robertson, 1977a; Gilbert and Robertson, 2013). The succession continues with Maastrichtian-early Cenozoic basaltic volcanics and both pelagic and redeposited carbonates. The Kyrenia Range is interpreted as part of the northern active continental margin of the Southern Neotethys (Robertson and Woodcock, 1986; Robertson et al., 2012) that was juxtaposed with the Troodos Massif, possibly by the late Eocene and certainly by the late Miocene (Robertson et al., 2012, 2014). The Kyrenia Range escaped the major anticlockwise tectonic rotation of the Troodos Massif (Morris et al., 2015), and effectively forms the southern limit of Tauride continental crust (Fig. 1a).

3. Methods

3.1. Fieldwork and sampling

Previous geological maps (Lapierre, 1975; Robertson, 1977a) were compared and field checked. The type area of the Kannaviou Formation was remapped to illustrate the structure and sediment distribution (Fig. 3). More northerly exposures in the Sarama-Peristerona and Kinousa areas, and in the southern part of the Akamas Peninsula, were studied for comparison (see Supplementary material) (Robertson, 1977a). In addition, sedimentary logs of well-exposed local successions were measured to shed light on depositional processes and to aid sampling (Fig. 4). Fault distribution and kinematic studies were carried out to determine regional stress regimes during different time periods (see Supplementary material). During this study we used a combination of new and previously collected samples (n=20 and 18, respectively), and also new and existing laboratory data, as summarised in Table 1.

3.2. Petrography

Sandstones of medium to coarse grain size were studied using an optical microscope to determine their mineralogy and petrography. Scanning electron microscopy (SEM) was used to study grain fabric and texture. Fourteen medium-grained sandstones with well-preserved textures were selected for point counting (Table 1). At least 400 points was counted for each sample using a Swift Automatic point-counter, which gave a statistically meaningful result with 95% confidence limits (Van der Plas and Tobi, 1965). The Gazzi-Dickinson point-counting method was utilised to maximise source rock data while minimising the effects of grain size variation (Graham et al., 1976; Dickinson and Suczek, 1979; Ingersoll et al., 1984). The recorded components (with

explanatory notes) are listed in Table 2. The raw data (by percentage), and the calculated framework parameters are given in the Supplementary material.

3.3. Whole-rock major, trace, and rare earth element geochemistry

Whole-rock major and trace element concentrations of the sandstones (both new and previous samples) were analysed by X-ray fluorescence (XRF) at the School of GeoSciences, University of Edinburgh, using the methods of Fitton et al. (1998) and Fitton and Godard (2004) (see Table 1). Accuracy and precision estimates are comparable to those of these authors.

Ten newly collected samples (Table 1) were analysed at the ACME Laboratories, Vancouver for trace and rare-earth elements (REEs) by inductively coupled plasma mass spectrometer, following lithium metaborate/tetraborate fusion and nitric acid digestion. As an indication of precision, the major intra-lab standard (Reference Material STD SO-18/19) was analysed together with the samples, resulting in standard deviations ($\leq 6\%$) for all trace and rare earth elements. For further information on the procedures, geostandards and precision for the elements analysed see <http://acmelab.com>. All of the analytical data for the major, trace and rare earth elements are listed in the Supplementary material.

3.4. Glass and mineral chemistry

Two polished thin sections (nos. 1074 and 2164 of Robertson, 1977a) that contain abundant well-preserved undevitrified volcanic glass and feldspar were selected for analysis of major elements, using a Cameca SX100 electron microprobe (EMPA) at the School of GeoSciences, University of Edinburgh. Previous analyses of glass, feldspar and pyroxene used the same method (Gilbert and Robertson, 2013) (Table 1). Details of the analytical conditions and

methods are given by Hayward (2011) and by Hartley and Thordarson (2013). The accuracy of major element determinations is better than ± 1 % of total value. All analytical data for glass, feldspar and pyroxene are given in the Supplementary material.

Trace-element analyses of undevitrified volcanic glass were performed on four new polished thin sections (nos. 14-11, 14-14, 16-20, 16-22) (Table 1) using a Cameca IMS-4f ion probe at the School of GeoSciences, University of Edinburgh. The analytical data, standardization protocols and errors are given in the Supplementary material.

3.5. U-Th-Pb secondary ion microscopy (SIMS) geochronology

Zircon crystals were selected from the 50-125 μm fraction that was extracted from c. 20 kg of medium to coarse-grained sandstone (sample nos. 14-04, 14-11, 14-14) (Table 1). A Wilfley table, Frantz Isodynamic magnetic separator and a high density solution (lithium polytungstate; 2.85 g/ml) were used to aid separation. Zircon grains ($n=18$) were randomly picked under a binocular microscope and finally mounted in epoxy resin and polished down to expose the grain interior. Internal structures were studied with the SEM using cathodoluminescence (CL) at the School of GeoSciences, University of Edinburgh. U-Pb analysis was then performed on a Cameca ims-1270 SIMS at the School of GeoSciences, University of Edinburgh, using the methods detailed by Kelly et al. (2008) and Ustaömer et al. (2012). Errors on the reported ages are $\pm 1\sigma$. Geochronological plots were produced using the Excel Macro program ISOPLOT (Ludwig, 2012). Analytical results are listed in the Supplementary material.

4. Results

4.1. Field relationships

4.1.1. Structure

Structural analysis was required to understand the stratigraphy of the Kannaviou Formation. The main part of the outcrop southeast of Kannaviou village is deformed into a large-scale syncline with a c. east-west-trending axis, as inferred by the presence of outcrops of ophiolitic basement both to the north and south (Fig. 3a-b). The core of the fold is made of variably folded mudstones and volcanoclastic sandstones, typically up to hundreds of metres in scale. Locally, the sandstone is structurally inverted, as indicated by overturned cross-bedding and normal grading. West of Kannaviou village, the succession is deformed into a large anticline (Fig. 3c). The core of this structure comprises thick-bedded, medium to coarse-grained volcanoclastic sandstone, siltstone and calcareous claystone. The Troodos ophiolitic lavas north of Kannaviou village (near Ezousa River) are affected by top-to-the northeast thrusting (Fig. 3). In the south, the main Kannaviou Formation outcrop is overthrust by arcuate lineaments of highly sheared ophiolitic serpentinite, as well exposed west of Statos (Fig. 3). Also, in the south of the Akamas Peninsula, ophiolitic rocks are thrust westwards over the Kannaviou Formation (Lapierre, 1975; Robertson, 1977a) (Fig. 1b).

The Kannaviou Formation is affected by two main generations of faulting. Overall, compression-related faulting relates to latest Cretaceous (Maastrichtian) emplacement of the Mesozoic continental margin and oceanic rocks of the Mamonia Complex (Robertson, 1977a; Swarbrick and Naylor, 1980). This was followed by mainly extensional faulting related to formation of

the Neogene Polis graben (Payne and Robertson, 1995, 2000). Mapping and kinematic analysis (see Supplementary material) show that faulting offsets parts of the succession but does not greatly disrupt the stratigraphy of the Kannaviou Formation.

4.1.2. Sedimentary succession

The lower part of the Kannaviou Formation (c. 100 m) is dominated by non-calcareous to variably calcareous claystone (Fig. 4a-b), interbedded with thin-bedded volcanoclastic mudstone (Figs. 4c, 5a). Several fine-grained, lenticular volcanoclastic sandstone packages, up to 10 m thick, are locally present (Figs. 4a-d, 5b). Bioturbation (<3 cm long) is commonly well developed (Fig. 5c). The mudstone and sandstone locally contain basalt, hyaloclastite breccia and/or palagonite clasts, 3-45 cm in diameter, with a matrix-supported fabric, as seen near the Troodos extrusive rock contact (Figs. 4b, 5d).

The middle part of the succession is characterised by thick-bedded, medium to coarse-grained volcanoclastic sandstone (Fig. 4a, e-h). The individual sandstone depositional unit is much thicker (up to 30 m), more massive and virtually structureless compared to those beneath. A prominent c. 25 m thick nearly homogeneous tabular sandstone northeast of Kannaviou village (Fig. 4a) exhibits occasional planar lamination, low-angle cross lamination and indistinct normal grading. Convolute lamination (up to 25 cm thick) and evidence of slumping occur locally (Figs. 4h, 5e). Cross-bedding measurements from small outcrops (e.g., near Istinjo) show a general preferred orientation of foreset lamination towards the southeast (Robertson, 1977a). This is consistent with the absence of the Kannaviou Formation farther east,

where pelagic chalks of the Lefkara Formation directly overlie the Perapedhi Formation or the Troodos ophiolitic lavas (Robertson, 1977a).

The upper part of the succession is dominated by claystone and volcaniclastic mudstone (Fig. 4e-g, i-j). Lenticular sandstone units (up to 20 m thick) are interbedded with structureless claystone and weakly-bedded volcaniclastic mudstone, for example around Sarama-Lasa (Figs. 4e-f, 5f). Sandstone, where present, is generally thin to medium-bedded, and forms lenses up to c. 300 m long. Individual sandstones are fine to medium-grained, well-laminated and normal-graded. Elsewhere, low-angle cross-bedding, planar cross-bedding and horizontal stratification are present. Mud rip-up clasts are locally present near the base of individual beds, and some beds are rich in plant material (i.e., branches, twigs and leaf debris). Dark epiclastic grains of basalt are commonly visible in hand specimen (<1 mm in size).

4.2. Petrography and detrital modes

Framework grains are characterised by sub-angular to sub-rounded morphologies and poorly to moderately-sorted fabrics. Quartz (2-18 modal %) and feldspar (6-21 modal %; up to 1200 μ m) are common. Abundant volcaniclastic material is present, mostly as transparent volcanic glass with only rare phenocrysts (see Section 4.2.2). In contrast, volcanic grains (mostly angular) range in composition from mafic to intermediate (dark coloured; <4%), to felsic (pale coloured; 2-15%) and commonly contain phenocrysts (see Section 4.2.2) (Fig. 6a). Grains of pyroxene (<3%, up to 500 μ m), biotite (<2%) and hornblende (<2%) occur variably. Clasts of radiolarian chert, mica-schist and foraminiferal pelagic limestone also occur locally (Fig. 6b-c). Microfossils are widespread, mostly radiolarians, planktic foraminifera and sponge spicules

(Fig. 6d). The sandstones are generally matrix-supported and carbonate-cemented. The matrix constitutes 6-48 modal %, and mainly comprises clay minerals, together with minor tiny detrital fragments of volcanic glass, feldspar and mica.

4.2.1. Compositional variation

The sandstones show considerable compositional variation from the base of the succession upwards. The lowermost sandstones (sample nos. 14-14, 19.1) are relatively fine-grained (c. 150-300 μm), generally non-calcareous, rich in quartz (including polycrystalline quartz; QFL%Q=17-19) but low in feldspar (QFL%F=13-15). Mica-schist is common and also locally chert. A medium to coarse-grained sandstone sample (no. 18.5) from near the Paleomylon River contains abundant feldspar (QFL%F=24) and lithic fragments (QFL%L=70) that are mostly of volcanic origin (97%). Where the Kannaviou Formation is exposed along the arcuate tectonic lineaments (Fig. 1b) only the base of the succession is exposed restricted to fine-grained, thin-bedded sandstone (Gilbert and Robertson, 2013). A typical sandstone (no. 14.5) from the lineament south of Episkopi is fine to medium-grained, carbonate-bearing litharenite with abundant polycrystalline quartz and metamorphic lithics (e.g., mica-schist).

In the middle part of the succession (sample nos. 14-12, 14-13, 21.2, 22.1), the grain size of the sandstone is typically c. 180-250 μm but reaches 400 μm locally (e.g., south of Kannaviou village; no. 22.1). Quartz (including polycrystalline quartz) decreases (QFL%Q=6-13), whereas feldspar increases (QFL%F=10-32). Volcanic lithics slightly increase and chert becomes common. Dolomite and carbonate cements become more abundant. Sandstones from the upper part of the succession (sample nos. 14-11, 23.1) are generally

coarse-grained (c. 300-500 μm) with more abundant volcanic lithics (LmLvLs%Lv=96-99). Polycrystalline quartz and mica-schist are sparse. Devitrified volcanic glass and recrystallised limestone are rarely present.

Four medium-grained, thick-bedded sandstones from the middle to upper part of the succession in the Sarama-Lasa area are characterised by marked compositional variation, with generally low textural maturity. Two sandstones (nos. 2124, 2158) have abundant lithic fragments (QFL%L=72-80) but few quartz grains (QFL%Q<5). Two other sandstones (nos. 2141, 2156) have a higher abundance of quartz, especially polycrystalline quartz (QFL%Q=15-20) and subhedral feldspar (QFL%F=26) but less lithic fragments (QFL%L=54-59). Volcanic lithics predominate (LmLvLs%Lv=92-96) in these two sandstones, with subordinate metamorphic (no. 2141; LmLvLs%Lm=6) and sedimentary (no. 2156; LmLvLs%Ls=4) fractions.

4.2.2. Glass morphology

The grain size is typically c. 100-700 μm but reaches 1200 μm locally (Fig. 7). Colourless glass is mainly felsic (Robertson, 1977a; Gilbert and Robertson, 2013), constituting over 75% of the grains analysed (see Section 4.3.1). Pinkish or brownish glass is of mafic to intermediate composition (Robertson, 1977a) (see Supplementary material). Taken together, the relative abundances of glass morphologies, glass colour (related to composition), textures and vesicle abundances allow four general glass groups to be recognised (Fig. 7): (1) transparent, highly vesicular pumiceous glass with a common fluid-like or rarely frothy texture; (2) colourless glass as blocky shards, cusped bubble wall, and rarely tubular or spindle-shaped glass, all of up to moderate vesicularity; (3)

pinkish glass composed of generally vesicular cusped shards and blocky shards; and (4) brownish glass with a microlitic texture.

4.2.3. Modal proportions

On the QFL diagram of McBride (1963) (Fig. 8a), sandstones from all sample localities are feldspathic litharenite. An undissected arc basin is indicated on the QFL diagram (e.g., Dickinson et al., 1983) (Fig. 8b) as the dominant provenance, especially for the Kannaviou and Paleomylon River area sandstones. Two samples from Sarama-Lasa (nos. 2141, 2156) and one from south of Episkopi (no. 14.5) plot in the transitional arc field. Plotted on the QFL and LmLvLs diagrams of Marsaglia and Ingersoll (1992) (Fig. 8c-d), the Kannaviou and Paleomylon River area samples plot in the continental arc field. The average Kannaviou Formation composition falls in the continental arc field ($Q_{12}F_{21}L_{67}$) (Fig. 8c) or the overlapping field of intra-oceanic and continental arcs ($Lm_2Lv_{94}Ls_4$) (Fig. 8d).

4.3. Glass and mineral chemistry

4.3.1. Volcanic glass

Volcanic glass of different morphology and from different localities shows considerable compositional variation (Fig. 9). The volcanic glass analysed (n=102) is divisible into two main chemical groups according to silica content: i.e., mafic/intermediate (<57 wt. % SiO_2), and felsic (>70 wt. % SiO_2). All of the glass samples have moderate Na_2O+K_2O contents (2.3-8.1 wt. %), typical of the subalkaline/tholeiitic series (Fig. 9). Glass from the Lempa (no. 4.2), Paleomylon (no. 18.5) and Sarama-Lasa (no. 2164) areas is felsic. Glass from the Kannaviou area (nos. 19.1, 21.1, 21.2, 23A.1) is largely felsic (91%), rarely

trending towards intermediate/mafic (9%). Glass from the Akamas area (no. 1074) is bimodal, mostly mafic (88%) and to a lesser extent felsic (12%).

Oxides such as TiO_2 , Al_2O_3 , CaO and FeO are tightly grouped and exhibit positive correlations with MgO (see Supplementary material). MgO contents cluster around 0-1 wt. % and 4-6 wt. %. FeO exhibits a wide range (0.6-13.7 wt. %), as does CaO (0.7-10.9 wt. %). Al_2O_3 content is relatively high (c. 10.1-19.5 wt. %), whereas TiO_2 is typically low (0.1-1.6 wt. %). Na_2O , K_2O and P_2O_5 are scattered.

The trace-element composition of the glass was also plotted on a normal mid-ocean ridge basalt (N-MORB)-normalised multi-element spider diagram (Fig. 10). Glass from near Kannaviou (nos. 14-11, 14-14) and from south of Lapithiou (nos. 16-20, 16-22) is compositionally similar, characterised by relative enrichment in Ba and Th (10-200×N-MORB), and also in Ta, Nb and La (1-10×N-MORB). Concentrations of Zr and Sm lie nearly parallel to N-MORB. Ti and Y are relatively depleted at 0.05-1×N-MORB concentrations.

4.3.2. Feldspar

Eighty-one feldspar grains were analysed from samples from five different localities (Table 1) (Fig. 11a). All of the feldspar grains have low orthoclase contents (<4.8%). Feldspar from the Akamas area (no. 1074; n=9) shows a wide range in anorthite content (55-91%; labradorite-bytownite to anorthite), whereas feldspar grains from the Kannaviou area (nos. 19.1, 21.1, 21.2, 23A.1; n=37) are more variable (40-95%; andesine-anorthite). A bimodal distribution of feldspar compositions exists in samples from the Lempa (no. 4.2; n=10),

Paleomylon (no. 18.5; n=12) and Sarama-Lasa (no. 2164; n=13) areas, with anorthite contents of mainly 30-50% (andesine) and 70-90% (bytownite).

4.3.3. Clinopyroxene

Forty-two clinopyroxene grains were analysed from the Kannaviou (nos. 19.1, 21.1, 21.2, 23A.1; n=24), Lempa (no. 4.2; n=8) and Paleomylon (no. 18.5; n=10) areas. Clinopyroxene grains show relatively uniform compositional proportions and are mainly augite (Morimoto, 1988) with one exception from the Lempa area (Fig. 11b).

4.3.4. Amphibole

Eight grains of amphibole from the Kannaviou (nos. 19.1, 21.1, 23A.1), Lempa (no. 4.2) and Paleomylon (no. 18.5) areas have relatively uniform magnesiohornblende compositions (Leake et al., 1997) (Fig. 11c).

4.4. Zircon U-Pb dating

Zircon grains are generally sparse and commonly poorly preserved. Most are euhedral and 50-100 μm long (Fig. 12). The crystals display variable internal structures including magmatic oscillatory zoning (zircon 1), inherited (xenocrystic) cores (zircon 5) and metamict structures (zircon 8). The zircon grains have high Th/U ratios (0.18-1.47), suggestive of a magmatic origin (Rubatto, 2002). The 18 analysed zircon grains have concordance levels ranging from 99-106%. The calculated $^{206}\text{Pb}/^{238}\text{U}$ age with the generally lowest possible error was used for age determination.

Zircon grains from the Armou area (no. 14-4; n=6) have a dominant age population of 81.9-85.7 Ma (n=5) (Fig. 13a), together with one grain of 578.9 Ma (zircon 6). Zircon grains from sandstones near Kannaviou village (nos. 14-

12, 14-14; n=12) have a significant zircon population of 78.1-83.6 Ma (n=11) (Fig. 13b). One grain was dated at 560.8 Ma (zircon 12). The only concordant or near concordant data (99-102%) within the dominant (relatively young) age distribution are represented by magmatic-textured euhedral zircon grains (nos. 4, 7, 9, 10, 13, 14, 16, 17) (Fig. 13c). These analyses yield a mean weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 80.1 ± 1.1 Ma (MSWD=1.7) (Fig. 13d), which is assumed to be the maximum age of the sandstone deposition, and also the probable age of the volcanic source.

Some of the zircon age data were excluded from the weighted age for the following reasons: The older zircon grains (nos. 6, 12) are rounded with a patchy xenocrystic texture that is suggestive of recycled origin (Fig. 12). Five zircon grains (nos. 1, 3, 8, 15, 18) show slightly reverse discordance, possibly due to slight loss of U and/or variable excess radiogenic Pb (e.g., Söderlund et al., 2010). Zircon grains 2 and 11 have relatively high common ^{204}Pb at 2.9 and 3.5 ppb, respectively. Zircon grain 5 that has an inherited core with a slightly older age (85.7 Ma) and zircon 7 with a relatively old age (83.6 Ma) were also excluded.

4.5. Whole-rock geochemistry

4.5.1. Major-element variation

The Kannaviou Formation sandstones show a wide compositional range (SiO_2 contents of 32-63 wt. %) (Fig. 14), characterised by low TiO_2 (<0.7 wt. %), Na_2O (0.9-2.9 wt. %), K_2O (0.4-1.8 wt. %) and Fe_2O_3 (2.8-5.4 wt. %). Al_2O_3 values are relatively high (typically 11-18 wt. %), as are CaO (5.0-26.6 wt. %) (see Supplementary material). Loss on ignition values are high (3-22 wt. %), as

expected given the common carbonate cement. As a result, we focus on trace element contents and inter-element ratios.

4.5.2. Trace-element variation

Selected trace element concentrations for the Kannaviou Formation sandstones are shown in multi-element diagrams (Fig. 15), normalised to upper continental crust (UCC). Patterns of sandstones from Armou, Lempa, Sarama-Lasa and south of Lapithiou are similar. Their trace element concentrations are well below UCC with the exceptions of Y, Tm and Yb. The Armou, Sarama-Lasa and south of Lapithiou area sandstones are mainly highly depleted in Th, U, Nb and La, whereas the Lempa samples (nos. 4.1, 4.2) are relatively enriched in these elements (with a more restricted range). In addition, one sample (no. 21.1) from the Kannaviou area is characterised by positive anomalies of Th, U, La and Ce but pronounced depletions in Ba and Sr, whereas others (e.g., nos. 14-12, 19.1, 23A.1) are similar to Lempa (no. 4.2) or Sarama-Lasa (no. 16-39). In contrast, the Paleomylon River area sandstones are highly depleted in all trace elements, with pronounced positive anomalies of Sr and wide ranges of U, K, Nb and La. However, Sr and probably also Y (no. 16-04) are slightly enriched. A sample from near Ezousa River (no. 16-30) is strongly depleted in U and has moderate negative anomalies of Ba, Sr and Zr.

Positive anomalies generally exist compared to UCC, especially for Sr, Ba and probably also for K, controlled by feldspar and/or K-bearing clay mineral content (e.g., illite, mica) (Nesbitt et al., 1980; McLennan et al., 1983; Feng and Kerrich, 1990). In contrast, negative anomalies of Zr, Hf, Th and U can be explained by a reduced detrital heavy mineral fraction (Taylor and McLennan,

1985) and/or relatively mafic provenance (Bauluz et al., 2000). The generally positive anomalies of Y and probably also of P, especially from the Sarama-Lasa area samples (nos. 16-33, 16-34), imply weathering enrichment of Y related to phosphatic clay and/or a detrital heavy minerals (Wronkiewicz and Condie, 1987).

4.5.3. Rare earth element variation

Chondrite-normalised rare earth element (REE) patterns of selected sandstones from the Kannaviou Formation and reference data from UCC (Rudnick and Gao, 2003; Hu and Gao, 2008) and deep-sea sands from various settings (McLennan et al., 1990) are shown in Figure 16. Sandstones from the Kannaviou area (nos. 14-11, 14-12, 14-14) show moderately fractionated REE patterns with $(La/Yb)_N=1.83-3.93$ and $(Gd/Yb)_N=0.99-1.16$, generally negative Eu anomalies (0.73-1.0) and relatively low total REE contents (45.1-67.9 ppm). Sandstones from the Paleomylon River area (nos. 16-04, 16-13, 16-14) are comparatively enriched in LREE, have negative Eu anomalies (0.81-0.83) and flatten out towards HREEs. Sandstone from Sarama-Lasa (no. 16-34) has a fractionated REE pattern with a moderate $(La/Yb)_N=2.01$, a negative Eu anomaly (0.71) and relative HREE enrichment. Sandstones from south of Lapithiou (nos. 16-22, 16-23, 16-29) are similar to the Kannaviou area, but with considerably more pronounced REE fractionation and lower HREE concentrations. In contrast, the reference data (McLennan et al., 1990) show much less variation. In general, the average Kannaviou Formation sandstone parallels average forearc basin sand composition.

5. Discussion

5.1. Local geological influences

The thickness and facies variation of the sandstone, especially in the lower part of the Kannaviou Formation, were strongly influenced by the contemporaneous sea-floor topography. The northeasterly limit of the Kannaviou Formation coincides with the tectonically controlled relief that was created by the westward extension of the c. east-west-trending South Troodos Transform Fault (Robertson, 1977a) (Fig. 17). The transform fault zone formed a bathymetric depression with a strongly rising fault-controlled topography to the north and northeast of the type area around Kannaviou. Fault scarp-derived (epiclastic) lava breccia and relatively coarse-grained, thick-bedded sandstone and minor conglomerate accumulated near the northern limit of the transform, as seen in the Paleomylon River valley (Fig. 4b). After its accumulation during the Campanian (c. 80 Ma) the Kannaviou Formation was affected by north to northeast-directed compression in its type area around Kannaviou. The compression relates to docking of the relatively allochthonous Triassic-Cretaceous continental margin and oceanic rocks of the Mamonia Complex (Robertson and Woodcock, 1979; Malpas et al., 1992; Bailey et al., 2000) (Fig. 4d). During the Paleogene, the Kannaviou Formation was covered by the pelagic chalks of the Lefkara Formation (Figs. 3, 4e). The formation was later transected by c. northwest-southeast-trending steep extensional faults related to the Polis graben during late Miocene-Pliocene (Payne and Robertson, 1995, 2000; Balmer et al., 2019) (see Supplementary material). The present-day exposure is locally influenced by landsliding, triggered by the Pleistocene uplift of Cyprus and related earthquakes (Mercier et al., 1973; Robertson, 1977c;

Poole and Robertson, 1998; Morag et al., 2016). It was necessary to take account of all of these features to understand the primary deposition of the volcaniclastic sediments.

5.2. Deposition of the volcaniclastic sandstone

The non-calcareous claystone and radiolarian mudstone in the lower part of the succession (e.g., Fig. 4a-b) accumulated relatively slowly, near or beneath the carbonate compensation depth (CCD) based mainly on the carbonate and microfossil content of the interbedded fine-grained sediments (Robertson, 1977a). The near-basal matrix-supported basaltic clasts and occasional lenticular, amalgamated sandstone packages with composite bedding (Figs. 4b, 5b) are indicative of relatively rapid, channelised emplacement by mass flows. The petrographic evidence of basaltic epiclasts near the base of the succession (no. 18.5) (Fig. 4b), partially degraded hyaloclastite (Gilbert and Robertson, 2013), and the high abundance of fine-grained matrix (see Supplementary material), indicate derivation from the adjacent, then submerged Troodos ophiolite.

Upwards, the increased occurrence of calcareous claystone, mudstone and volcaniclastic sandstone, with local evidence of convolute lamination, record relatively high sedimentation rates and common tectonic instability (see Section 4.1.2) (e.g., Figs. 4h, 5e). Sporadic, indistinct planar and low-angle cross-lamination are indicative of subordinate turbidity current deposition in parts of the succession (Robertson, 1977a) (Fig. 4a). The relatively coarse grain size, the increase in subhedral feldspar grains, and the occurrence of subangular chert and mica-schist grains (see Section 4.2.1), together suggest that source lithologies were being rapidly degraded and reworked.

The relatively thick (up to 20 m thick by c. 300 m long), lenticular volcanoclastic sandstones, locally rich in mud rip-up clasts (see Section 4.1.2) are indicative of rapid, pulsed, channelised accumulation (Robertson, 1977a), similar to that reported from comparable sediments elsewhere (e.g., Nishimura et al., 1991). The appearance of locally abundant although fragmentary plant material indicates input from land (see Section 4.1.2). The upward presence of calcareous mudstone interbeds (e.g., Fig. 4e-g, i-j) with calcareous microfossils indicates a change to deposition above the CCD.

The vast amount of vesicular pyroclastic material (Fig. 7) is interpreted to have originated from explosive volcanic eruptions (Robertson, 1977a; Gilbert and Robertson, 2013), similar to many volcanoclastic deposits elsewhere (Schindlbeck et al., 2013; Kutterolf et al., 2014). Excellent examples were recently documented from the northwest Pacific region (Busby et al., 2017; Kutterolf et al., 2018; Robertson et al., 2018). The restriction of the pyroclastic material to sand-sized or smaller (see Section 4.2.2) points to initial explosive eruption, followed by dominantly aeolian ash transport. The eruptions are likely to have been subaerial but could also have been in shallow water (< c. 300 m) where hydrostatic pressure was insufficient to prevent explosive release of ash into the atmosphere. Such phreatic eruptions are for example known from the 1965 eruption of Taal Volcano, Philippines (Moore et al., 1966), and the 1975-1977 activity of La Soufrière de Guadeloupe (Lesser Antilles) (Boichu et al., 2011). After initial ash fallout, and probable local current reworking, the volcanoclastic sediment represented by the Kannaviou Formation was carried downslope by mass flows in submarine channels, individually up to 25 m deep by up to 300 m wide (Figs. 4b, 5f). Sequences of repeated turbidites are not

developed in contrast to many deep-sea successions in modern and ancient forearc basins (Dickinson, 1995; Pickering and Hiscott, 2016). Also, undisturbed primary ash fallout deposits are effectively absent. The lack of conglomeratic or pyroclastic debris-flows deposits within the Kannaviou Formation hints at long-distance aeolian transport that left such coarse-grained material behind. Alternatively, a topographic barrier (e.g., ocean floor fault zone) (Fig. 17) could have isolated the depositional area from coarser-grained supply. The dominance of thick, channelised mass-flow (Fig. 4), relative to classical turbidites points, points to relatively proximal deposition in an upper to mid-fan setting (Howell and Normark, 1982). It is therefore unlikely that the sands were transported very long distances (hundreds to thousands of km) by axial flow as in some modern trench settings, such as the Peru-Chile trench (Schweller et al., 1981) or the Aleutian trench (Piper et al., 1973). However, initial long-distance sediment transport by wind and currents is not precluded, as discussed further below.

5.3. Factors affecting sandstone composition

5.3.1. Alteration of mineral and glass grains

In thin section, calcium is relatively abundant in the form of foraminifera, calcite spar (cement), clay minerals and feldspar (up to 14.5 modal %). Relatively high values of CaO, together with positive correlations of CaO and loss on ignition (LOI) values, are consistent with the carbonate content (Fig. 18a). Dissolved feldspar grains are relatively common. Some detrital pyroxene grains show superficial alteration to chlorite. In places, secondary porosity including small late-stage cracks is infilled by clay minerals or calcite spar. Only rarely are glass

shards partly devitrified, in which case they typically have thin clay or carbonate cement coatings.

The relatively unaltered nature of the glass (c. 86% of the grains) is supported by the analytical totals of typically >92 wt. % (see Supplementary material). A few glass grains, mainly from Sarama (no. 2164), gave totals of 89-91 wt. %. The typically high average totals of c. 95 wt. % are similar to those reported from a range of Pleistocene deep-sea sediments, including tephra from the Mediterranean Sea (90-99 wt. %) (Clift and Blusztajn, 1999), Central America (c. 93 wt. %) (Kutterolf et al., 2008a, b), New Zealand (92-99 wt. %) (Lowe et al., 2008) and also Neogene tephra from the northwest Pacific (90-100 wt. %) (Bryant et al., 2003; Kutterolf et al., 2018). A possible explanation of the relatively unaltered nature of the Kannaviou Formation sandstones, especially the volcanic glass, is because of the protective effects of diagenetic carbonate cement (e.g., Ulmer-Scholle et al., 2015).

A few glass grains (especially two from the Kannaviou area) have unusually low analytical totals (89-90 wt. %) and high silica contents (61-71 wt. %); this could have several explanations: (1) loss of Na and K during analysis (Nielsen and Sigurdsson, 1981; Clift and Fitton, 1998); (2) relatively high eruptive volatile contents, following crystal fractionation (Burnham and Jahns, 1962); and (3) secondary hydration (Clift and Dixon, 1994). The majority of the glass grains have high silica contents similar, for example, to Mediterranean tephra (Clift and Blusztajn, 1999). The relationship between measured volatile-loss (100%-Total) and silica content (Clift and Dixon, 1994) (Fig. 18b) suggests that little secondary hydration of the Kannaviou Formation glass has generally taken place.

5.3.2. Grain-size, weathering, sorting and recycling

Several factors can affect sandstone chemical composition, including grain-size variation, weathering, sorting and recycling (e.g., Wronkiewicz and Condie, 1987; Cullers, 1995). Some elemental concentrations and ratios can be selectively affected by grain-size variation. However, MnO, CaO, K₂O and Sr concentrations, and also La/Co, Th/Co, La/Ni and Th/Cr ratios are normally not much affected (Cullers, 1995). Finer-grained fractions typically comprise relatively homogeneous detritus and therefore generally have relatively consistent elemental concentrations and ratios (Cullers, 1995, 2000). These fractions are typically characterised by relatively high Al₂O₃, Fe₂O₃, MgO, TiO₂, Th, Sc, Nb, Y and REE, low SiO₂, Na₂O and Ba, and relatively low Eu/Eu*, La/Sc and Th/Sc (Taylor and McLennan, 1985; Cullers, 1995). In the Kannaviou Formation sandstones Na₂O (correlation coefficient $r=0.83$), TiO₂ ($r=0.67$), Sc ($r=0.68$), Be ($r=0.61$), Ga ($r=0.92$), Hf ($r=0.70$), V ($r=0.47$), Zr ($r=0.48$), Eu ($r=0.90$) and HREE ($r=0.4-0.6$) all correlate with Al₂O₃ (see Supplementary material), which suggests a clay mineral influence (e.g., clayey matrix, illite, kaolinite) on chemical composition (e.g., Roser and Korsch, 1986; Feng and Kerrich, 1990). To overcome these effects, trace element ratios, namely La/Co, Th/Co, Th/Cr, Eu/Eu*, La/Sc, Th/Sc, Zr/Sc and La/Th were used to characterise provenance and tectonic setting (Taylor and McLennan, 1985; Bhatia and Crook, 1986; Condie and Wronkiewicz, 1990; Cullers, 1995, 2000). However, caution is required because some plots (e.g., La-Th-Sc, Th-Sc-Zr) can be influenced by local sorting, and by mafic/heavy mineral variation (Bhatia and Crook, 1986; Floyd et al., 1991).

Th/U ratios can indicate source area weathering and sedimentary recycling (e.g., Taylor and McLennan, 1985; McLennan et al., 1993). The majority of the Kannaviou Formation sandstones have Th/U ratios of 1.83-4.0, generally similar to active margin-derived turbidites (McLennan et al., 1990). The low values suggest derivation from little weathered magmatic arc-related source rocks (McLennan et al., 1993). In contrast, relatively elevated Th/U ratios (5-28) in some samples (nos. 16-16, 16-20, 16-25, 16-39, 21.1) point to enhanced subaerial weathering.

The absence of significant zircon enrichment (37-99 ppm) precludes marked sediment recycling or sorting effects (McLennan et al., 1993). The depositional age from microfossils (Campanian-early Maastrichtian; 84-72 Ma) (Urquhart and Banner, 1994; Bragina and Bragin, 1995) is broadly coeval with the related volcanism as indicated by zircon geochronology (c. 80 Ma). This evidence rules out chemical differentiation resulting from long-term sedimentary recycling.

5.4. Source rock characteristics

5.4.1. Petrographic constraints

The QFL diagram (Fig. 8b) illustrates entire grain populations and emphasises grain stability, weathering, transport mechanisms and source rocks. All of the samples plot near the lithic apex with the main contribution being from volcanic glass, consistent with a magmatic arc source (e.g., Dickinson et al., 1983). Sandstones from south of Episkopi (no. 14.5) and near Sarama-Lasa (nos. 2141, 2156) cluster towards the recycled field, with relatively high quartz abundances (especially polycrystalline quartz at 2-3 modal %). These clastic materials were probably supplied from petrographically similar quartzose

sandstones in the adjacent Mamonia Complex, especially those of late Triassic age (Robertson, 1977a). The mean composition of the Kannaviou Formation plots in the area of overlapping fields (Fig. 8d) in the LmLvLs diagram is explained by the high volcanic glass content (15-47 modal %). Despite this, the relatively high average quartz content ($QFL\%Q=12$) in the Kannaviou Formation sandstones supports input from a continental margin arc (Marsaglia and Ingersoll, 1992; Marsaglia et al., 2016).

5.4.2. Constraints from glass composition

The volcanic glass has relatively low abundances of Na_2O (1.9-4.4 wt. %), K_2O (0.2-4.6 wt. %), TiO_2 (0.1-1.6 wt. %) and P_2O_5 (<0.2 wt. %), typical of incipient subduction-related volcanism (Woodhead et al., 1993; Straub, 2003). Enrichment in Ba and Th, together with depletion in Ti and Nb, support an arc origin (e.g., Pearce, 1983). The glass compositions plot close to the tholeiitic trend on the ternary $Na_2O+K_2O-FeO-MgO$ (AFM) diagram (Fig. 19a). The trend towards the alkaline apex likely represents arc magmatic maturity (Kuno, 1966; Geringer et al., 1986), and is consistent with differentiation of a single parental basaltic magma (Gilbert and Robertson, 2013).

Volcanic glass composition is critical to evaluate continental versus oceanic sediment provenance (Bryant et al., 2003; Kutterolf et al., 2014). Elevated U/Th, Ba/La and Sm/La ratios indicate pelagic sediment and/or fluid influence on magmatic composition (Patino et al., 2000; Straub and Layne, 2003; Plank, 2005; Carr et al., 2007). Elevated ratios of Th/La, Rb/Hf and Th/Nb suggest continental crust and/or subducted terrigenous sediment influence (Hannah et al., 2002; Bryant et al., 2003; Stern et al., 2003). The relatively elevated Th/Nb and Th/La ratios in the Kannaviou Formation glass (Fig. 19b-c)

therefore support the involvement of continental crust and/or subducted terrigenous sediment in magma genesis.

5.4.3. Constraints from mineral analysis

The dominance of augite, with one rare diopside (Fig. 11b), suggests a mafic igneous source (Deer et al., 2013). The clinopyroxene compositions, as shown in the ternary $\text{TiO}_2\text{-MnO-Na}_2\text{O}$ discriminant plot (after Nisbet and Pearce, 1977) (Fig. 19d), indicate a volcanic arc affinity for all of the grains analysed. In addition, the two Ediacaran zircon grains (560 and 578 Ma) originated from older continental crust. The Triassic-Cretaceous sedimentary succession of the adjacent Mamonia Complex is a likely source for the grains of these ages based on detrital zircon geochronology (Chen, 2018).

5.4.4. Constraints from whole-rock geochemistry

The relatively uniform ratios of La/Co, Th/Co and Th/Cr, together with La/Sc and Th/Sc in the Kananviou Formation sandstones, are consistent with derivation from predominantly mafic source rocks, with a subordinate variable input from felsic rocks (Cullers, 1995, 2000) (Table 3). Sandstones with unusual La/Sc or Th/Sc represent locally increased detrital felsic input (e.g., feldspar, muscovite) or reduced mafic contribution (Taylor and McLennan, 1985; Condie and Wronkiewicz, 1990; Cullers, 2000). Th/Sc vs. Zr/Sc and La/Th vs. Hf diagrams indicate similar compositional variations (Fig. 20a-b).

Moderately fractionated REE patterns, low ΣREE abundances (45-79 ppm) and low $(\text{La/Yb})_N$ (1.8-6.2), together indicate dominant derivation from mafic to intermediate-composition rocks. Slight HREE enrichment in one sample (no. 16-34) from directly beneath the Kathikas Formation in the Sarama-Lasa area (Fig. 4f) probably resulted from heavy mineral enrichment (e.g., zircon) and/or

input of relatively degraded fine-grained material (Taylor and McLennan, 1985), both of which could have been derived from the Mamonia Complex. Negative Eu anomalies of 0.71-1.00 are slightly higher than those in craton-derived post-Archean Australia Shale (PAAS) (0.66; Taylor and McLennan, 1985) and UCC (0.70; Rudnick and Gao, 2003). These ratios are comparable to those of sandstones derived from mafic sources (0.71-0.95; Cullers, 2000) and/or magmatic arcs (0.79-1.04; Bhatia, 1985) associated with active margin settings (McLennan et al., 1990). Overall, the linear array of sandstone data (Fig. 20c) is consistent with a mixture of compositionally variable young arc-derived material and upper crust-sourced material (e.g., McLennan and Taylor, 1984) that was probably derived from the adjacent Mamonia Complex.

5.5. Provenance

The trace element ratios of the volcanic glass, coupled with the presence of some terrigenous sediment, support a continental arc affinity (Fig. 19b-c). Despite this, the trace-element ratios of the sandstones generally fingerprint an oceanic island arc setting. Specifically, the ratios of La/Sc (0.58 ± 0.18), Th/Sc (0.17 ± 0.06) and La/Th (4.08 ± 0.77) are generally comparable to those of sandstones derived from an oceanic island arc setting (0.55 ± 0.22 , 0.15 ± 0.08 , 4.26 ± 1.2). However, several ratios (e.g., Eu/Eu*) suggest a continental contribution. The scattering towards the continental island arc field is consistent with locally reduced mafic contribution or increased felsic input (Bhatia and Crook, 1986; Cullers, 1995, 2000; Cullers and Podkovyrov, 2000). On the ternary Th-Sc-Zr diagram (Bhatia and Crook, 1986) (Fig. 21a) all of the sandstones plot in or near the oceanic island arc field.

The average Kannaviou Formation sandstone has negative Nb-Ta anomalies compared to the UCC-normalised multi-element patterns of average greywacke from several different tectonic environments (Floyd et al., 1991) (Fig. 21b). The negative anomalies are comparable to those of oceanic island arc-sourced greywackes, suggestive of sediment sources involving subduction-related magmatic rocks (Floyd et al., 1991). The variable enrichments in V-Cr-Ni-Sc and the slightly positive Ti anomalies are also consistent with arc provenance (either continental or oceanic). In addition, the pronounced depletion in Zr and Hf precludes significant enrichment in heavy minerals (e.g., zircon). The patterns are generally consistent with a source composed predominantly of subduction-related rocks of oceanic island arc affinity, plus a minor continental input (McLennan et al., 1990, 1993; Floyd et al., 1991).

5.6. Tectonic-magmatic setting

5.6.1. Continental arc magmatism on thinned, depleted continental crust?

The most likely explanation of the apparent discrepancy between the continental margin arc setting inferred from the glass trace element chemistry and the mainly oceanic arc origin suggested by the trace and REE data is that the arc magmas represented by the Kannaviou Formation were not extracted from fertile asthenosphere above normal-thickness continental crust. Arc-type signatures could instead result if magma was extracted from crust that was relatively thin and/or depleted by previous melt extraction (Fitton et al., 1997; Condie, 2003). In the case of the Kannaviou Formation, both of these situations are likely to have occurred because the region was affected by Triassic rift-related magmatism to form the Southern Neotethys. Early spreading of this

ocean basin is represented by the late Triassic MORB (Dhiarizos Group) of the Mamonia Complex (Robertson and Woodcock, 1979; Malpas et al., 1992, 1993; Chan et al., 2008). Active continental margins in several other regions are known to have relatively primitive oceanic arc-type signatures, as indicated by the chemistry of derived volcanoclastic sediments and related tuffaceous deposits, for example the Brook Street Terrane in southern New Zealand (Robertson et al., in press).

5.6.2. Arc-volcanoclastic sediment comparisons

Despite their inferred continental margin arc provenance, as mainly inferred from the glass composition, the Kannaviou Formation sandstones accumulated on oceanic crust. Examples of volcanic arc-sourced successions overlying oceanic crust are well represented in the southwest and north Pacific regions (e.g., Horn et al., 1969; Marsaglia, 1992; Clift et al., 1998; Kutterolf et al., 2018; Robertson et al., 2018). Ancient examples of oceanic arc-derived volcanoclastic sediments that accumulated on oceanic crust include the late Cretaceous Nindam Formation in northwest India (Robertson and Degnan, 1994; Clift et al., 2000, 2002), the Jurassic Talkeetna Volcanic Formation in Alaska (Plafker et al., 1989; Clift et al., 2005; Greene et al., 2006), the Permian Takitimu Group in southern New Zealand (Houghton and Landis, 1989; Spandler et al., 2005; Robertson and Palamakumbura, in press; Robertson et al., in press), the Jurassic Gran Cañon Formation (Critelli et al., 2002; Busby, 2004) and the Cretaceous Alisitos Group (Busby et al., 2006; Marsaglia et al., 2016) in Mexico. In addition, there are many instances of arc-derived successions overlying continental crust (Dickinson, 1995), to name one well-documented example,

the late Cenozoic Central American Volcanic Arc (Carr, 1984; Kutterolf et al., 2008a, b; Schindlbeck et al., 2016).

A key requirement for the genesis of the Kannaviou Formation sandstones is for supra-subduction zone oceanic lithosphere, represented by the Troodos ophiolite, to collide with an adjacent continental margin arc, following subduction of intervening Neotethyan oceanic crust. The tectonic-magmatic setting of the Kannaviou Formation can be broadly compared to the intersection of the Mariana-Bonin forearc (counterpart of the Troodos ophiolite) with the Japanese continental island arc, as exposed in the Miura and Boso Peninsulas (Ogawa et al., 1985, 2008). However, no exact counterpart is known elsewhere for the complex microplate setting of Cyprus during the latest Cretaceous.

5.6.3. Implications for late Cretaceous arc magmatism in the eastern Mediterranean

The previously suggested source (Robertson, 1977a; Gilbert and Robertson, 2013) of the Kannaviou Formation is the Fourkovouno (Selvilitepe) Formation of the Kyrenia Range. This is characterised by a c. 400 m thick late Cretaceous subduction-related volcanogenic succession that encompasses felsic tuffaceous rocks, felsic debris-flow deposits and rhyolitic lava flows (Moore, 1960; Poisson, 1977; Baroz, 1979, 1980; Robertson and Woodcock, 1986; Huang et al., 2007; Robertson et al., 2012; Chen, 2018). However, there are differences between the Kannaviou Formation source volcanism and that of the Kyrenia Range continental margin arc magmatism: (1) the Kannaviou Formation sandstones were derived from an incipient continental margin arc, whereas the Kyrenia Range volcanic rocks represent a more evolved Andean-type active continental margin; and (2) the Kannaviou Formation volcanism is

significantly older (80.1 ± 1.1 Ma) than that of the Kyrenia Range ($^{206}\text{Pb}/^{238}\text{U}$ age of 72.9 ± 1.0 Ma) (Chen, 2018). A possible explanation for these differences is that older, immature arc volcanism originally existed in the Kyrenia Range area but has been concealed by subsequent subduction and/or continental collision, consistent with the evidence of southward thrusting during the Middle Eocene and Late Miocene-earliest Pliocene (Baroz, 1979; Robertson et al., 2012, 2014; McCay and Robertson, 2013; Robertson and Kinnaird, 2016).

The northerly active continental margin of the Southern Neotethys, represented in north Cyprus by the Kyrenia Range (Robertson et al., 2012), extended eastwards through the Misis-Andırın lineament of southern Turkey into southeast Turkey (e.g., Robertson, 2002) (Fig. 1a). Late Cretaceous continental margin arc magmatism in southeast Turkey is also interpreted to relate to northward subduction of the Southern Neotethys beneath a northerly-located Tauride continental unit (e.g., Robertson, 2002; Parlak, 2006; Rızaoğlu et al., 2009; Robertson et al., 2006, 2007, 2012). There are two potential sources of felsic ash in southeast Turkey: (1) tholeiitic oceanic arc magmatism dated at 83-75 Ma (Aktaş and Robertson, 1984, 1990; Robertson et al., 2007; Karaoğlu et al., 2013); and (2) I-type, calc-alkaline Andean-type intrusive rocks, dated at 88-82 Ma (Santonian-Campanian) (Parlak, 2006; Robertson et al., 2006; Rızaoğlu et al., 2009; Karaoğlu et al., 2016). Unfortunately, any extrusive equivalents of the continental arc magmatism were eroded or are not exposed. Of the two occurrences of arc magmatism, based on the geochemical evidence, only the continental margin arc volcanism could be a source for the Kannaviou Formation volcanic ash, assuming long-distance transport by winds and/or currents. Prevailing easterly winds and currents have been modelled for

the late Cretaceous (e.g., Hay, 2011), such that a source well to the east of Cyprus should be considered.

5.6.4. Pyroclastic fallout dispersal efficiency

Pyroclastic fallout components such as glass shards, feldspar crystals, mafic crystals, lithic grains and pumice can be dispersed substantial distances by wind (Carey and Sigurdsson, 1982, 2000; Fisher and Schmincke, 1984; Pyle, 1989). In general, the dispersal distance is mainly controlled by a combination of grain size, prevailing wind strength and eruptive volume (i.e., column height) (Pyle, 1989; Carey, 1996). Relatively recent (<0.5 Ma) ash of grain size 32-500 μm that was derived from the continental margin Japan arc can be correlated with tuffaceous sediments of the oceanic Izu-Bonin forearc, up to 1000 km away (Kutterolf et al., 2018). Maximum grain size versus distance from source relationships are, for example, known for ash layers that were produced by the Toba eruption in Sumatra (75 ka; Ninkovich et al., 1978), the Campanian ignimbrite eruption in Italy (33 ka; Cornell et al., 1983) and the Minoan eruption of Santorini (1620 B.C.; Watkins et al., 1978). The maximum sizes of the feldspar/glass crystals and the mafic crystals (e.g., pyroxene) of the Kannaviou Formation sandstones are up to 1200 μm and c. 500 μm , respectively. Assuming the crystals were sourced from a common eruptive centre, these sizes are compatible with eruptions on the approximate scale of the Minoan eruption of Santorini (28 km³; Watkins et al., 1978), with a likely distance of c. 100 km from the source arc (Fig. 22). This distance is significantly less than that of the known continental margin arc volcanism in southeast Turkey (c. 300 km). Also, to achieve the up to 750 m thickness of the Kannaviou Formation, repeated Santorini-scale eruptions would be needed over a relatively short time

period (c. 0.3 Ma), followed by channelised sediment transport of up to c. 300 km (e.g., Mason et al., 2004; Deligne et al., 2011). The volcanoclastic sandstones are found only in western Cyprus and not overlying the Troodos ophiolite as a whole, which points to a relatively local source, followed by bathymetrically controlled (channelised) deposition. The most likely scenario is that the Kannaviou Formation source eruptions were located close to the Kyrenia Range but have since been concealed tectonically, as a result of the crustal contraction that took place in response to subduction or collision.

5.6.5. Tectonic-magmatic model

Our proposed model (Fig. 23) builds on a previously proposed double subduction scenario (e.g., Robertson, 2002; Parlak, 2006; Karaoğlu et al., 2016), in which the Troodos ophiolite formed above an intra-oceanic subduction zone (c. 90 Ma) (Mukasa and Ludden, 1987), and then moved northwards related to subduction beneath Tauride microcontinent crust. Resulting early-stage (incipient) continental margin arc volcanism supplied pyroclastic fallout deposits to a deep-marine forearc basin to the south, coupled with gravity flow reworking. In addition, terrigenous sediment, mostly siliciclastic and recrystallised carbonate, was supplied from the adjacent continental margin, represented by the Triassic-Cretaceous terrigenous-derived sedimentary rocks of the Mamonia Complex (Robertson and Woodcock, 1979; Torley and Robertson, 2018).

6. Conclusions

In this study we have used a combination field-based sedimentology and structural geology, optical petrography, whole-rock, glass and mineral chemistry, and also U-Pb zircon dating to infer that the late Cretaceous (80.1±1.0 Ma) arc volcanism of the Kannaviou Formation in western Cyprus was derived from an incipient continental margin arc. A similar integrated multi-technique approach has the potential to shed new light on a range of modern and ancient arc-related sediments elsewhere. The presence of terrigenous material within some of the volcanoclastic sandstones initially suggested a continental margin arc setting. The SIMS trace element analysis of relatively unaltered volcanic glass confirmed a continental influence. Despite this, several conventional trace element tectonic discriminant diagrams point to an oceanic arc source. The likely solution of this apparent discrepancy is that the continental margin arc was constructed on magmatically depleted lithosphere that was thinned during regional late Permian-early Jurassic rifting of the Southern Neotethys. Effective interpretation of comparable arc-derived volcanoclastic sandstones elsewhere therefore depends on a multi-technique approach, especially glass trace element analysis (where possible) and should not rely on interpretation of whole-rock chemical data alone.

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Fig. 1. (a) Simplified map showing Cyprus in the eastern Mediterranean. The locations of the late Cretaceous volcanogenic sediments (FF; blue box) in the Kyrenia Range, northern Cyprus and the late Cretaceous arc-related rocks in southeast Turkey (green box) are indicated; (b) Outline geological map of western Cyprus, showing the main outcrops of the late Cretaceous Kannaviou Formation, and their relationship to the Troodos ophiolite and Mamonia Complex (modified after Robertson, 1977a; Robertson and Woodcock, 1979; Swarbrick, 1980). Abbreviations: STTFZ, South Troodos Transform Fault Zone; FF, Fourkovouno (Selvilitepe) Formation.

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ophiolite forms by incipient spreading above a separate subduction zone farther
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Fig. 1

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Table 1
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Sample	Geographic location	Geographic coordinates	QFL	XRF ^a	ICP-MS	EMPA ^a	IMS-4f	U-Pb	Note ^b
14-04	Armou	34°47'40"N, 32°28'19"E		X				X	
14-11	Kannaviou	34°54'48"N, 32°32'59"E	X	X	X		X		
14-12	Kannaviou	34°54'53"N, 32°33'29"E	X	X	X			X	
14-13	Kannaviou	34°54'53"N, 32°34'16"E	X	X					
14-14	Kannaviou	34°55'20"N, 32°34'34"E	X	X	X		X	X	
16-04	Paleomylon	34°55'08"N, 32°35'01"E		X	X				
16-13	Paleomylon	34°55'06"N, 32°35'18"E		X	X				
16-14	Paleomylon	34°55'02"N, 32°35'29"E		X	X				
16-15	Paleomylon	34°54'44"N, 32°35'44"E		X					
16-16	Paleomylon	34°55'03"N, 32°36'03"E		X					
16-20	south of Lapithiou	34°53'55"N, 32°36'25"E		X			X		
16-22	south of Lapithiou	34°53'44"N, 32°35'36"E		X	X		X		
16-23	south of Lapithiou	34°53'50"N, 32°35'07"E		X	X				
16-25	south of Lapithiou	34°53'54"N, 32°34'51"E		X					
16-29	south of Lapithiou	34°54'09"N, 32°35'19"E		X	X				
16-30	Ezousa River	34°55'36"N, 32°35'00"E		X					
16-31	Kannaviou-Kri. Marottou	34°55'50"N, 32°33'53"E		X					
16-33	Sarama	34°57'25"N, 32°32'28"E		X					
16-34	Sarama	34°56'52"N, 32°32'02"E		X	X				
16-39	Lasa	34°55'47"N, 32°31'51"E		X					
4.1	Lempa	34°48'52"N, 32°24'08"E		X ^a					
4.2	Lempa	34°48'52"N, 32°24'08"E		X ^a		X ^a			
6.2	Armou	34°47'43"N, 32°28'22"E		X ^a					
14.5	south of Episkopi	34°46'08"N, 32°32'28"E	X						
18.4	Paleomylon	34°55'02"N, 32°36'00"E		X ^a					
18.5	Paleomylon	34°55'02"N, 32°36'00"E	X	X ^a		X ^a			
19.1	Kannaviou	34°55'07"N, 32°34'33"E	X	X ^a		X ^a			
21.1	Kannaviou	34°54'53"N, 32°34'10"E		X ^a		X ^a			
21.2	Kannaviou	34°54'48"N, 32°34'20"E	X	X ^a		X ^a			
22.1	Kannaviou	34°54'48"N, 32°34'20"E	X						
23.1	Kannaviou	34°55'14"N, 32°33'09"E	X	X ^a					
23A.1	Kannaviou	34°55'03"N, 32°33'05"E		X ^a		X ^a			
1074	Akamas	34°59'31"N, 32°19'56"E				X			
2124	Lasa	NA	X						
2141	Sarama	NA	X						
2156	Sarama	NA	X						
2158	Sarama	NA	X						
2164	Sarama	NA				X			

QFL: petrographic thin section study for point counting;
XRF: major and trace element analysed by X-ray fluorescence spectrometry;
ICP-MS: trace and rare earth element analysed by inductively coupled plasma mass spectrometry;
EMPA: mineral chemistry analysed by electron microprobe;
IMS-4f: glass trace element analysed by ion microprobe;
U-Pb: zircon U-Pb geochronology analysed by secondary-ion mass spectrometry;

^a Data from Gilbert and Robertson (2013);

^b Thin section, polished section source.

Table 2
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Main parameter		Calculated parameter	
Qm	Monocrystalline quartz	Q	Total quartz grains (=Qm+Qp)
Qp	Polycrystalline quartz	F	Total feldspar grains (=Fg)
Fg ^a	Feldspar grains	Lv	Volcanic lithics (=Vg+Lvm+Lv ^f)
Fl	Feldspar in lithics	Lsm	Sedimentary and metamorphic lithics (=Ls+Lm)
Vg ^b	Volcanic glass	L	Total unstable lithic fragments (=Lv+Lsm+Lo+Fl)
Lvm ^c	Intermediate to mafic volcanic lithics	Lt	Total lithic fragments (=L+Qp)
Lv ^f ^d	Felsic volcanic lithics		
Ls	Sedimentary lithics		
Lm	Metamorphic lithics		
Lo	Other lithics		
Px	Detrital pyroxenes		
Am	Detrital amphiboles		
Bt	Detrital biotite grains		
Ms	Detrital muscovite flakes		
Cal	Carbonate grains		
Bio	Bioclastic detritus		
M	Matrix		
C	Cement		

^a no differentiation of feldspar;
^b volcanic glass (relatively pure, with very minor phenocrysts);
^c dark coloured volcanic lithics, with common pyroxene and opaque crystals;
^d pale coloured volcanic lithics with common quartz and feldspar laths.

Table 3
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	Kannaviou Formation							Sediment from felsic sources ^a		Sediment from mafic sources ^a	
	Armou	Ezousa	Kanaviou	Lempa	Paleomylon	Sarama-Lasa	S Lapithiou	Coarse fractions	Fine fractions	Coarse fractions	Fine fractions
La/Co	-	-	0.24-0.84	-	0.69-1.13	0.59	0.67	1.8-13.8	1.4-22.4	0.14-0.38	-
Th/Co	-	-	0.05-0.31	-	0.16-0.44	0.12	0.21	0.67-19.4	0.3-7.5	0.04-1.4	-
Th/Cr	0.03	0.01	0.03-0.71	0.08	0.03-0.12	0.02-0.05	0.05	0.13-2.7	0.067-4.0	0.018-0.046	0.002-0.045
La/Sc	0.27-1.58	-	0.08-1.52	0.70-2.30	0.04-0.77	0.04-0.45	0.47	2.5-16.3	0.7-27.7	0.43-0.86	0.4-1.1
Th/Sc	0.17	-	0.69	0.26-0.39	0-0.32	0.02-0.10	0.14	0.84-20.5	0.64-18.1	0.05-0.22	0.05-0.4
Eu/Eu*	-	-	0.73-1.00	-	0.81-0.83	0.71	0.81-0.86	0.40-0.94	0.32-0.83	0.71-0.95	0.7-1.02

^a data from Cullers (1995, 2000)

Supplementary Appendix A

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Supplementary Appendix B

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Supplementary Figure 1

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Supplementary Figure 2

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Supplementary Figure 3

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Supplementary Figure 4

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