Tectonic reconstruction of Cyprus reveals Late Miocene continental collision of Africa and Anatolia

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Tectonic reconstruction of Cyprus reveals late Miocene continental collision of Africa and Anatolia

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Highlights:

17.5 km crustal shortening in N. Cyprus recorded 9-6 Ma Africa-Tauride collision

Kyrenia part of Africa from Late Cretaceous Troodos obduction until collision

Under-thrusting of Africa margin a viable cause of 0.5 Ma uplift in south Anatolia

Abstract

The extended northern continental margin of Africa is currently colliding with and under-thrusting northward below Cyprus. The age of onset of the collision is poorly constrained, but is critical if we are to quantify continental subduction and evaluate its role during spectacular recent uplift of southern Turkey from ~7 Ma to the present. Here, we revaluate the evolution of northern Cyprus and document the modern structure of the Kyrenia fold-thrust belt in a balanced cross section for the first time, to determine the timing and amount of shortening. The belt deformed an Upper Cretaceous to Miocene stratigraphy, which was deposited onto a metamorphic basement. The fold-thrust belt was previously proposed to have formed during two stages: in Eocene and late Miocene time, based on stratigraphic and limited structural evidence. We revaluate evidence for an Eocene phase of thrusting in Kyrenia and find that 1) repetitions of Eocene and older rocks are explained by Miocene thrusting, and newly documented olistoliths were misinterpreted as pre-Miocene thrust slices 2) uplift in Eocene time was not isolated to Kyrenia but also affected northern Arabia and results from regional dynamic topography or forebulge formation. Based on kinematic constraints from a plate reconstruction, we interpret that metamorphic rocks in Kyrenia were metamorphosed by Cretaceous burial below the Troodos ophiolite and were subsequently exhumed by upper plate extension during latest Cretaceous obduction onto the African margin. The latest Cretaceous-Miocene plate boundary between Africa and the Taurides was located to the north of Kyrenia. The Kyrenia fold-thrust belt accommodated a minimum of 17.5 km of shortening as a result of ~9-6 Ma thrusting as a result of initial Africa-Taurides collision. The northern edge of African continental crust arrived below the Taurides, where it is seismologically images, in late Pleistocene time and may have caused or contributed to recently recognised major Tauride uplift since ~0.5 Ma.

Keywords: Central Anatolian Plateau; Ophiolite; Subduction; Fold-thrust belt

1. Introduction

The eastern Mediterranean region contains the oldest in situ ocean floor on Earth today (Granot, 2016). That ocean floor may be Carboniferous in age, and is located in the Herodotus basin, to the west of the island of Cyprus. As part of the African plate, it is actively subducting below Eurasia (Figure 1). Farther to the west, another remnant of ancient oceanic crust with a probable Triassic age, is found in the Ionian basin, and is subducting below the Calabrian and Aegean regions (Speranza et al., 2012). Those two small oceanic basins were part of a once wider ‘Eastern Mediterranean Ocean’ basin, that, given the disparate age of the modern relics, formed by a complex opening history. In Mesozoic time the Eastern Mediterranean Ocean separated what has been interpreted as a large micro-continental domain known as the Adria-Turkey plate (Stampfli et al., 1991) or ‘Greater Adria’ (Gaina et al., 2013) from Africa. Greater Adria contained the continental crust that presently underlies the Adriatic Sea and the Apulian and Gargano peninsulas of Italy, as well as the continental crust that once underlay the Apennines, southern Alps, Dinarides, Hellenides; and the Taurides of Turkey (Gaina et al., 2013).

Much of the Eastern Mediterranean Ocean basin has been subducted. Kinematic reconstructions based on geological records of ophiolites found around the eastern Mediterranean Sea (Figure 1) suggest that at least two subduction systems were responsible for the demise of the Eastern Mediterranean Ocean. Cretaceous ophiolites (~90-94 Ma) with SSZ geochemistry were obducted northward onto the Tauride continental platform of southern Anatolia, and southward towards Africa and Arabia (Moores et al., 1984; Mukasa and Ludden, 1987; Parlak & Delaloye, 1996; Al-Riyami et al., 2002; Çelik et al., 2006; Bağci et al., 2008; Pearce & Robinson, 2010; Karaoğlan et al., 2013). The modern distribution of these ophiolites, and evidence from paleomagnetic data have been used to infer that in Late Cretaceous time, much of the Eastern Mediterranean Ocean was invaded from the east by an originally ~N-S striking and east-dipping subduction zone. That subduction zone radially rolled-back westwards, and consumed ancient Carboniferous and Triassic oceanic lithosphere, and replaced it with Cretaceous forearc and back-arc basin lithosphere (e.g., Maffione et al., 2017; Moix et al., 2008; Barrier & Vrielynck 2008). The Herodotus basin apparently escaped Cretaceous subduction (Granot, 2016).

Towards the east, the Eastern Mediterranean Ocean has entirely subducted between Arabia and the Tauride fold-thrust belt. This culminated in middle to late Miocene continent-

continent collision at the Bitlis Suture zone (Şengör et al., 2003; Hüsing et al., 2009; Okay et al., 2010). In western Turkey, oceanic subduction continues today. To the south of Cyprus, the extended continental margin of Africa is currently subducting, and continent-continent collision is thus underway and in its infancy. The down-going African margin contains the continental Eratosthenes seamount which is part of a wide (hyper-) extended African margin (e.g., Robertson, 1998b). The age of onset of continent-continent collision, and the history of Cyprus during the collision are poorly constrained but are important to reconstruct if we are to determine the role of continental collision in a much debated and enigmatic, major (>2 km) uplift of the southern margin of Turkey since ~7 Ma (e.g., Cosentino et al., 2012; Schildgen et al., 2012; Meijers et al., 2016; Abgarmi et al., 2017; Öğretmen et al., 2018; Delph et al., 2017).

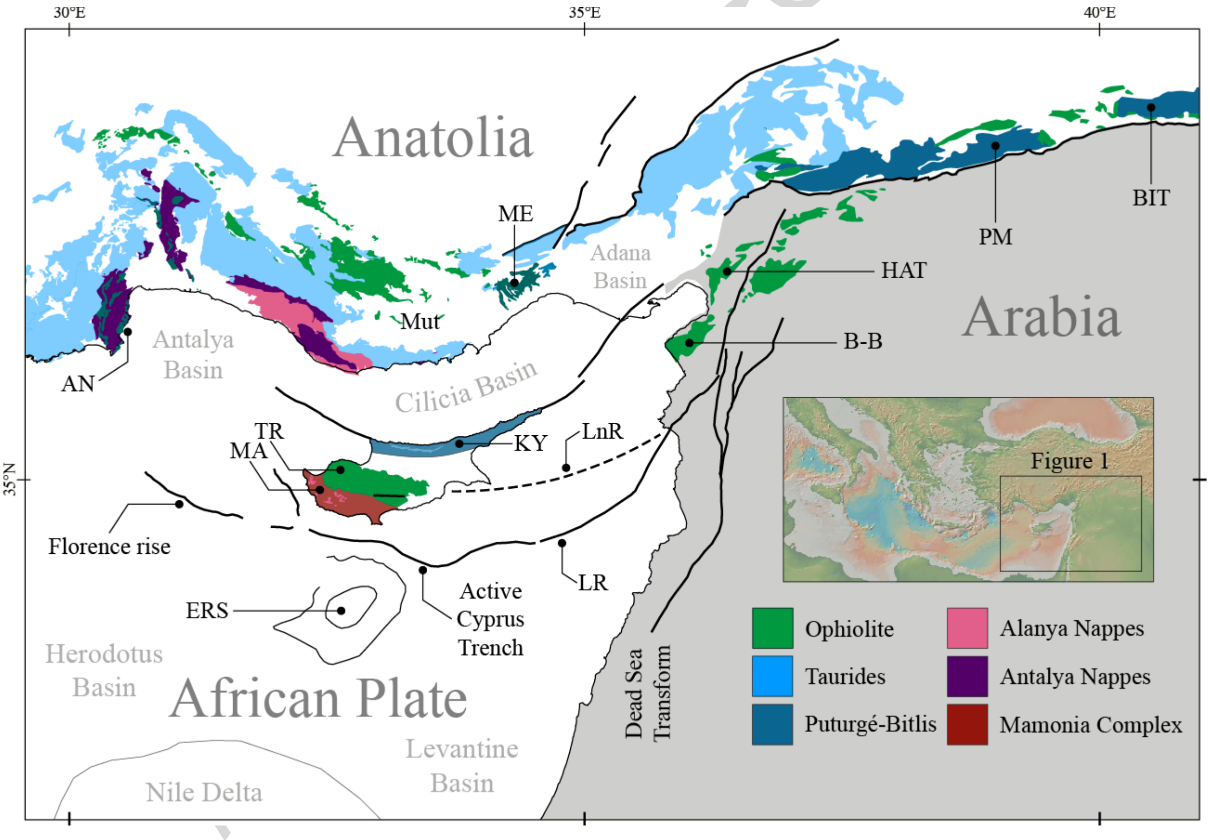


Figure 1: A map of the Eastern Mediterranean region, showing major tectonostratigraphic units, ophiolites, and major fault zones mentioned in this work. AN = Antalya Ophiolite; B-B = Baer-Bassit Ophiolite; BIT = Bitlis Massif; ERS = Eratosthenes Seamount; HAT = Hatay Ophiolite; KY = Kyrenia; LnR = Larnaka Ridge; LR = Latakia Ridge; MA = Mamonia Complex; ME = Mersin Ophiolite; PM = Puturgé Massif; TR = Troodos.

In this paper, we therefore study the stratigraphic, structural, and tectonic evolution of Cyprus since Cretaceous time to date the timing and location of closure of the Eastern Mediterranean Ocean. We provide the first balanced cross-section that viably restores and quantifies deformation on the island and we evaluate previous suggestions for the timing of shortening events. We will discuss our results in terms of the timing of Africa-Anatolia collision, and the implications for the contribution of this collision for uplift of the southern Anatolian margin.

2. Geological Setting

Evidence from GPS data demonstrate that at present, Cyprus forms the upper plate of a subduction zone that accommodates convergence between Africa and Eurasia (Reilinger et al., 2006). The modern plate boundary is marked by a 2.5 km deep seismically active trench, in which seismic reflection data show thrust faulting and thrust-related growth strata (Vidal et al., 2000; Hall et al., 2005b; Symeou et al., 2017). Seismic tomographic models show that the trench is associated with a north-dipping high-velocity anomaly interpreted as subducting African lithosphere, which may have undergone recent detachment to the north, and is still contiguous to the northwest of Cyprus (Biryol et al., 2011; van der Meer et al., 2018). There are few sub-crustal earthquakes associated with that anomaly below and to the north of Cyprus, meaning there is no active Benioff zone (Algermissen & Rogers, 2004; Cagnan & Tanircan, 2010), consistent with a broken slab.

The Cyprus trench is indented by a small continental block known as the Eratosthenes seamount, which was penetrated by boreholes in IODP Leg 160 (Robertson, 1998b). The upper stratigraphy of the seamount contains an interpreted Messinian subaerial unconformity marked by palaeosols. These are overlain by bathyal pelagic sediments representing post-Messinian refilling of the Mediterranean basin, after which rapid tectonic subsidence occurred as the Eratosthenes seamount subsided into the Cyprus trench (Robertson, 1998a).

The Mamonia Complex is the southern-most unit exposed on Cyprus and consists of deformed Triassic to Lower Cretaceous sedimentary and volcanic rocks interpreted as remnants of a deep-marine passive continental margin (e.g., Swarbrick & Robertson, 1980). The Mamonia Complex is separated from the southern edge of the Troodos ophiolite by a steep serpentinite-filled fault zone (Bailey et al., 2000). The sense of motion on that fault zone has been difficult to constrain, and left-lateral or right-lateral oblique motion

(Swarbrick, 1993; Bailey et al., 2000), or dip-slip thrusting (Malpas et al., 1993) have been proposed. The age of juxtaposition of Troodos ophiolite against the Mamonia Complex is constrained by the youngest pre-deformational sediments that are of Campanian age (Urquhart & Banner, 1994), and undeformed Maastrichtian pelagic sediments, which seal the Mamonia fault zone (Swarbrick & Robertson, 1980).

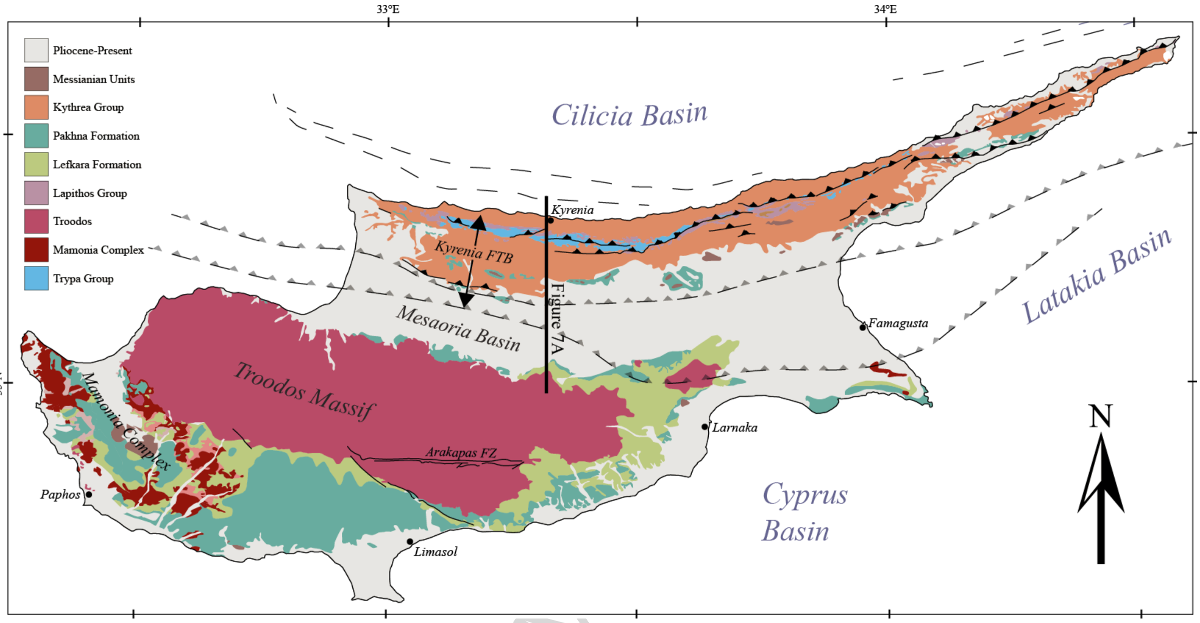


Figure 2: A simplified geological map based on the 1:250,000 scale 'Geological Map of Cyprus' published by the Geological Survey Department of Cyprus (Constantinou, 1995).

The Troodos ophiolite is one of the type localities of the Penrose pseudo-stratigraphy of oceanic lithosphere (Anonymous, 1972; Thy & Moores, 1988), and contains an ophiolitic crust that formed in Late Cretaceous time (92-90 Ma) (Mukasa & Ludden, 1987) in a supra-subduction zone (SSZ) setting (Moores et al., 1984; Pearce & Robinson, 2010). Paleomagnetic data collected from the ophiolitic crust and sedimentary cover of the Troodos ophiolite show that it was affected by a ~90° anti-clockwise vertical axis rotation after its formation (Clube & Robertson, 1986; Morris et al., 1990), of which ~65° was accommodated in pre-Maastrichtian times (Morris et al., 2006). Using a *net tectonic rotation analysis* (Allerton & Vine, 1987), the present ~N-S trending orientation of sheeted dykes of the Troodos ophiolite (e.g., Varga & Moores, 1985) were restored to a ~NE-SW orientation, and record ~NW-SE paleospreading (Morris & Maffione, 2016; Maffione et al., 2017). Major anti-clockwise rotations of 60-120° have also been recorded from the Hatay (Kızıldağ) and Baer-Bassit ophiolites that overlie the Arabian continental margin to the east of Cyprus, close to the Syria-Turkey national border (Morris et al., 2002; Inwood et al., 2009). Approximately

E-W to SE-NW spreading directions were also calculated from the sheeted dyke sections of those ophiolites (Maffione et al., 2017). The Hatay and Baer-Bassit ophiolites also carry a SSZ geochemical signature (Whitechurch et al., 1984; Lytwyn & Casey, 1993; Parlak et al., 2009), and U/Pb Zircon ages of ~92-95 Ma were derived from plagiogranite and gabbro of the Hatay ophiolite (Karaoğlan et al., 2013). These ophiolites were thrust onto the Arabian margin between Campanian and Maastrichtian times, based on the youngest age of shelf rocks incorporated into basal sheared serpentinite, and the oldest Arabian-derived rocks incorporated in the marine sedimentary cover of the ophiolite (Al-Riyami et al., 2002; Tinkler et al., 1981, Kaymakci et al., 2010, and references therein). The common spreading directions, age and origin of the crustal rocks, and sense and timing of rotations of the Baer-Bassit, Hatay and Troodos ophiolites suggest that they formed part of a common oceanic microplate that underwent a major rotation around the Arabian margin prior to Late Cretaceous obduction (e.g., Morris et al., 2006; Maffione et al., 2017).

The Troodos massif forms a gentle ~E-W trending dome structure, covered in the north by Campanian radiolarite (Follows & Robertson, 1990) and the Maastrichtian to Pliocene Mesaoria basin (Figure 2). At the modern southern edge of the basin, the stratigraphy thins and onlaps onto the *Upper Pillow Lavas* of the Troodos ophiolite (Cleintuar et al., 1977). Borehole data and evidence from magnetic surveys of Cyprus suggest that the Troodos ophiolite continues below the Mesaoria basin as far north as the Kyrenia range where it likely ends abruptly at a fault (Vine et al., 1973; Cleintuar et al., 1977). The Miocene and older stratigraphy in the southern part of the Mesaoria basin is largely covered by sub-horizontal Pliocene to Pleistocene rocks (e.g., Palamakumbura & Robertson, 2016), and the northern part of the basin is dominated by an E-W trending fold-thrust belt that deforms the upper Eocene to upper Miocene Kythrea Group. This is unconformably covered by undeformed Pliocene rocks (Palamakumbura & Robertson, 2016). The Kythrea Group consists of a ~2 km thick sequence of made up of interbedded marls and turbiditic sandstones and rare pelagic limestones, which have previously been subdivided into a series of sub-formations based on detailed sedimentological observations (McCay et al., 2013 and references therein). Paleoflow indicators within the Kythrea Group suggest that its sediments were derived from the east or northeast; lithic clasts within the sandstones were derived from metamorphic, igneous and sedimentary rocks (e.g., Weiler, 1970; McCay & Robertson, 2012). The E-W trending Ovgos Fault separates the Kythrea Group in the north from a Miocene-Pliocene sequence of marl, chalks, and neritic limestones of the Miocene Pakhna Formation and upper

Miocene-Pliocene Nicosia Formation (Harrison et al., 2008; Palamakumbura & Robertson, 2016). The deformation style in the fold-thrust belt will be described in detail in our results section, and has also been studied in part by Baroz (1979), and by Robertson and Kinnaird, (2015), and Harrison et al. (2004).

The Neogene evolution of Mediterranean basin to the east of Cyprus has been well studied with seismic reflection data (Vidal et al., 2000; Calon et al., 2005a; Calon et al., 2005b; Hall et al., 2005b; Hall et al., 2005a; Symeou et al., 2017). In general the basin there consists of multiple Miocene and younger piggy-back basins above the Kyrenia, Latakia, and Larnaka thrust culminations. The onset of thrusting on those structures is well defined by syn-kinematic packages in seismic stratigraphy of inferred mid to late Miocene age, although the age of that stratigraphy has never been confirmed with borehole data.

The northern extent of the Mesaoria basin is bound by the Kyrenia (or Kythrea, Baroz, 1979) thrust fault, which separates the basin sediments from the deeper stratigraphic units exposed in the Kyrenia range (Figure 2 & 3) (e.g., Baroz, 1979). The Kyrenia range is made up of a ridge of metamorphosed carbonates known as the Trypa Group, and an overlying uppermost Cretaceous (Maastrichtian) to Pleistocene non-metamorphic cover sequence of marine sedimentary and volcanic rocks (Robertson & Woodcock, 1986) (Figure 3). The metamorphosed rocks of the Trypa Group contain rare pelitic intervals with chlorite and stiplnomelane that indicate a greenschist facies metamorphic grade (Baroz, 1979). Fauna as young as Upper Jurassic have been reported within the metamorphosed limestones (Baroz, 1979), which constrains the maximum age of metamorphism. The Trypa Group is locally covered by Campanian sedimentary rocks of the Kiparisso Vouno Formation, and the lower Maastrichtian to Eocene Lapithos Group (Baroz, 1979; Robertson & Woodcock, 1986; Robertson et al., 2012; 2013).

Meter to kilometre scale blocks of the Trypa Group are included in the Lapithos Group, and were previously interpreted as either thrust slices (Baroz, 1979), or olistoliths (Ducloz, 1972 cited in Baroz, 1979). A >1 km block of ophiolitic mélange, with serpentinite, gabbro and dolerites with a SSZ geochemical signature, and red chert have also been reported from the Lapithos Group (Aldanmaz et al., in prep). The lower part of the Lapithos Group (known as the Melounda Formation, and Ayios Nikolaos Formation; Baroz, 1979) broadly consists of a basal conglomerate derived from the Trypa Group, and a lower Maastrichtian to Paleocene sequence of pelagic limestones and pillow lavas and basaltic breccia (Robertson et al., 2012), and above that, Paleocene to middle Eocene pelagic limestones contain debris flow conglomerates, calciturbidites, and pillow lavas and basaltic breccia (Robertson et al., 2012; 2013). Finally the upper part of the Lapithos Group (known as the Kalaograia-Ardana Formation; Baroz, 1979) consists of a middle Eocene sequence of interbedded mudstone, marl and sandstone, with debris flow conglomerates and olistoliths in the eastern part of the Kyrenia range (McCay et al., 2013). The upper formation notably contains exotic ‘Kantara limestones’, which are limestone clasts within the debris flow conglomerates and olistoliths up to several kilometres in size. Carboniferous to Permian ages have been proposed for some blocks; Triassic, Jurassic and Cretaceous ages have been confirmed, with the majority of blocks being Cretaceous in age (Baroz, 1979 and references therein; Robertson & Woodcock, 1986). The Lapithos Group and overlying Kythrea Group are separated by a regional unconformity. Along the northern flank of the Kyrenia range the Kythrea Group starts with a thin upper Eocene to lower Miocene basal conglomerate sequence and grades into a ~2 km thick middle Miocene sequence of interbedded mudstones and tabular turbidite sandstones (McCay & Robertson, 2012). Finally the Kythrea Group is unconformably overlain by terraces of marginal marine Plio-Pleistocene sediments (Palamakumbura et al., 2016).

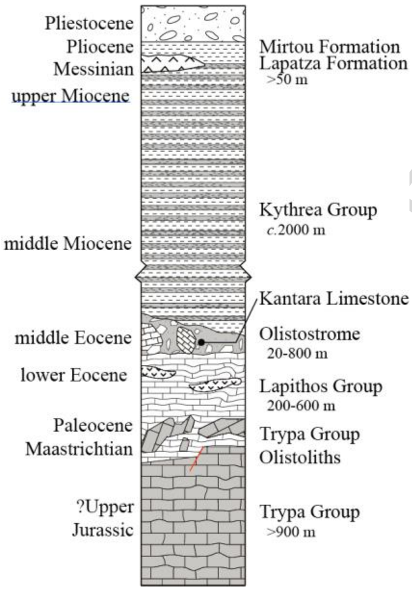


Figure 3: Generalised stratigraphy of the Kyrenia range. See text for descriptions of the units. Note that the thickness of the Kythrea Group has been condensed.

Northern Cyprus is separated from southern Turkey by the Cilicia basin, which contains a thick sequence of Miocene to Quaternary marine sedimentary rocks deposited in a sag basin that was affected by minor normal faulting and salt-related folding (Aksu et al., 2005; Walsh-Kennedy et al., 2014). The Pre-middle Miocene structure and stratigraphy underlying the basin is not imaged in published seismic data, and has not been penetrated by published borehole data, and so remains unconstrained.

To the north of Cyprus, in southern Turkey, the Tauride fold-thrust belt contains a record of Late Cretaceous to late middle Eocene accretion of upper crustal rock units derived from the Taurides platform (e.g., Gutnic et al., 1979; Özgül, 1984) (Figure 1). This forms part of an Anatolian orogen, which is in part time equivalent to the geological record we have described in Cyprus. The Anatolian orogen consists of a series of major continent-derived tectonostratigraphic units which contain an overall southward younging sequence of thrusting and metamorphism. These rocks accreted below 94-90 Ma ophiolites, and are interpreted to be a result of accretion above one or more north-dipping, intra-oceanic subduction zone(s) (Ricou et al., 1984; Moix et al., 2008; Robertson et al., 2009; Parlak et al., 2013; Gürer et al., 2016; Koç et al., 2016; Menant et al., 2016; McPhee et al., 2018). The Taurides non-metamorphic fold-thrust belt contains the latest and most southerly unit in that system, and accreted to Eurasia in late middle Eocene times (Sengör & Yilmaz, 1981; Özgül, 1984; van Hinsbergen et al., 2010; McPhee et al., 2018). After Tauride accretion in the Eocene, the subduction trench jumped southward and structurally below the Taurides. After that time, the position of the trench is enigmatic, and is one of the subjects of discussion in this paper.

Ophiolites overlying the northern flanks of the Taurides are widely accepted to have originated from the north, e.g., Beyşehir or Aladağ (Sengör & Yilmaz, 1981; Robertson et al., 2009; Pourteau et al., 2010; Parlak et al., 2013; van Hinsbergen et al., 2016; Menant et al., 2016; Plunder et al., 2016). Structural and stratigraphic evidence demonstrate, however, that the southern edge of the Taurides was obducted by a different set of ophiolites, from south to north (particularly those associated with the Antalya and Alanya nappes). This led to northward emplacement of the Antalya Nappes, which were derived from the southern Tauride passive margin, (Gutnic et al., 1979; Robertson & Woodcock, 1981; Özgül, 1984; Yilmaz & Maxwell, 1984), and emplacement of far-travelled, high pressure low temperature Alanya nappes with ~84 Ma peak metamorphism (Okay & Özgül, 1984; Çetinkaplan et al., 2016) , and overlying SSZ ophiolites and ophiolitic mélange with ages of ~90-95 Ma (Çelik et al., 2006). The frontal thrust faults at the base of the northward thrust Alanya-Antalya nappe-stack close to Alanya are sealed by Paleocene nummulitic limestone (Özgül, 1984), which implies that the northward thrusting of that nappe system was complete before the thrusting and accretion of the Taurides to Eurasia in the Eocene (McPhee et al., 2018). The southward-dipping subduction that led to the northward emplacement of the Antalya and Alanya Nappes onto the Tauride platform is interpreted to have been the same subduction zone that emplaced the Baer Bassit and Hatay ophiolites onto Arabia, and the Troodos ophiolite against Africa. That system is thought to have originated at a N-S trending trench segment in eastern Anatolia that radially rolled-back westwards, and emplaced ophiolites northward and southwards onto the margins of the Eastern Mediterranean Ocean (Maffione et al., 2017).

The Taurides, and the Alanya and Antalya Nappes today form a mountain range up to 3.5 km high that forms the southern margin of the Central Anatolian Plateau. This nappe stack is unconformably covered by mildly tilted but otherwise undeformed (Fernandez-Blanco, 2014) Miocene to Pleistocene marine sedimentary rocks of the Mut basin. The youngest Miocene marine sedimentary rocks of the Mut basin are ~7-8 Ma old and are currently found at up to 2.2 km elevation (Cosentino et al., 2012; Schildgen et al., 2012b), and are widely interpreted to reflect a late Miocene onset of uplift of the southern Taurides. Recently, Öğretmen et al. (2018) showed that marine uppermost Pliocene and Pleistocene sediments with ages as young as ~0.5 Ma unconformably overlie the marine upper Miocene sediments, at elevations up to 1.6 km. This shows that after a first, late Miocene phase of uplift and erosion, there was renewed Pliocene subsidence, followed by rapid late Pleistocene uplift. Recent geophysical studies suggested that subducted continental crust of the African plate reaches the southern Taurides (e.g., Abgarmi et al., 2017). The cause of the spectacular uplift of Anatolia remains hotly debated (Cosentino et al., 2012; Schildgen et al., 2012a; Bartol & Govers, 2014; Schildgen et al., 2014; Radeff et al., 2015), but it is generally thought that upper crustal processes alone cannot have been a driving mechanism in the uplift because the Mut basin is essentially undeformed.

3. Balanced cross section methods

We aimed to create a balanced and viable cross section across the Kyrenia range to investigate the structural style and magnitude of Cenozoic thrusting in the belt. We chose a N-S section line that is approximately perpendicular to the average strike of bedding in the Kythrea Group. We then surveyed a small-scale strip map along the section line. We collected dip measurements and samples for biostratigraphy, and compiled published subsurface data from deep boreholes as an additional constraint on the subsurface (see our supplementary file). We took care to check for way-up indicators in the Kythrea Group. Because previous subdivisions of the Kythrea Group cannot be straightforwardly mapped to the south of the Kyrenia range, we treated the Kythrea Group as a single unit. Our field observations were made north of the political border that divides Cyprus, and so we used data from a geological memoir (Bagnall, 1960) and field observations and borehole interpretations from the KL-1 and Tseri boreholes reported by Cleintuar et al., (1977) to reconstruct the southern margin of the Mesaoria basin. Pre-Pliocene rocks are sparsely exposed between the demilitarized zone that runs along the political border and the suburbs of Lefkosa and so we used a detailed small scale map of the bedrock published by the USGS (Harrison et al., 2008), which was created using surface exposures, and specially commissioned shallow boreholes and excavations.

In addition, we used evidence from published offshore seismic sections along strike of the Kyrenia range (Calon et al., 2005b) to inform our interpretation of the structural style. We used the seismic sections to estimate the regional dip of the belt (1-3°), to constrain the initial steepness and geometry of thrust faults, and to test whether the number of thrust faults we interpreted is reasonable. We did not reconstruct the internal structure of the Trypa Group, as we could not distinguish deformation caused by burial associated with Late Cretaceous, pre-Maastrichtian metamorphism from post-exhumation deformation. We did however reconstruct faults within the Trypa Group that deformed the overlying basin sequence.

We identified thrust faults at the surface where we documented stratigraphic repetition, or footwall cut-off geometries in bedded rocks. We used observed hanging wall deformation to predict the subsurface geometry of thrust faults (Berger & Johnson, 1980; Suppe, 1983), using an interactive forward model in the software package Move 2016. We changed the shape and displacement on each thrust fault until modelled hangingwall deformation recreated observed hangingwall deformation in our dataset. We used a fault-parallel-flow kinematic model (after Egan et al., 1997) in which material in the hangingwall moved along fault parallel paths during shortening. In appropriate places we modelled folds as fault propagation folds using a tri-shear algorithm in Move 2016 (after Erslev & Rogers, 1993).

Our cross section is a model, and during the construction of that model we made simplifying assumptions. First, we assumed plane strain. Second, we interpreted the simplest structural configuration that we saw to recreate the observed field data. In the absence of constraint, we assumed minimum shortening, by, for example, drawing hanging-wall cut-offs as close to the modern land surface as possible. We restored the deformation in our cross section to test whether our structural model was internally consistent (admissible) by unfolding the structures whilst maintaining the area of each thrust sheet. We verified the area balance by comparing the deformed and retro-deformed area of the section, and found that they were equal.

4. Results

We now describe our field observations, which will form the basis of our cross section, through the fold-thrust belt from south to north. We then present our balanced cross-section, with an explanation and justification for our structural model.

4.1 Field Observations

Mesaoria Basin

Our field observations come from well-exposed rocks to the north of Lefkosa. Between the suburb of Gönyeli and the Near East Technical University (NETU) campus on the outskirts of Lefkosa, Kythrea Group rocks mostly dip moderately northwards and are deformed by small scale (10’s of m) folding. North of NETU, the Kythrea Group rocks are well exposed, are typically steeply dipping, and form consistent dip domains (Figure 4A). Interestingly, these rocks are not affected by intense small-scale deformation, despite consisting of a large volume of weak mud rocks. The sandstone beds in the sequence form ridges in the landscape that we traced for many km on satellite imagery (e.g., Figure 4B). We collected 20 samples of claystone along a N-S transect through the Kythrea Group south of the Kyrenia range, 6 km east of our cross-section (Supplementary File 01). These samples were dated using calcareous nano fossils (carried out by Antonio Cascella at INGV Pisa). Six samples contained no diagnostic fauna, 13 samples were of Tortonian age, and one sample had a Serravallian age.

On the northern edge of the NETU campus we found a north-verging anticline and then a syncline with axes that plunge gently to the west. This fold-set is about 0.5 km in wavelength and is visible in satellite images (Figure 4B). The northern, south-dipping limb of the syncline has been obscured by the construction of a new road; we interpret the syncline as a footwall syncline that is cut by a thrust fault (Figure 5, NETU fault). North of that fold-set we found a domain of north-dipping bedding with very consistent dips (~58˚N) that extends northward for about 1 km. To the north of that domain, we found a cylindrical syncline about

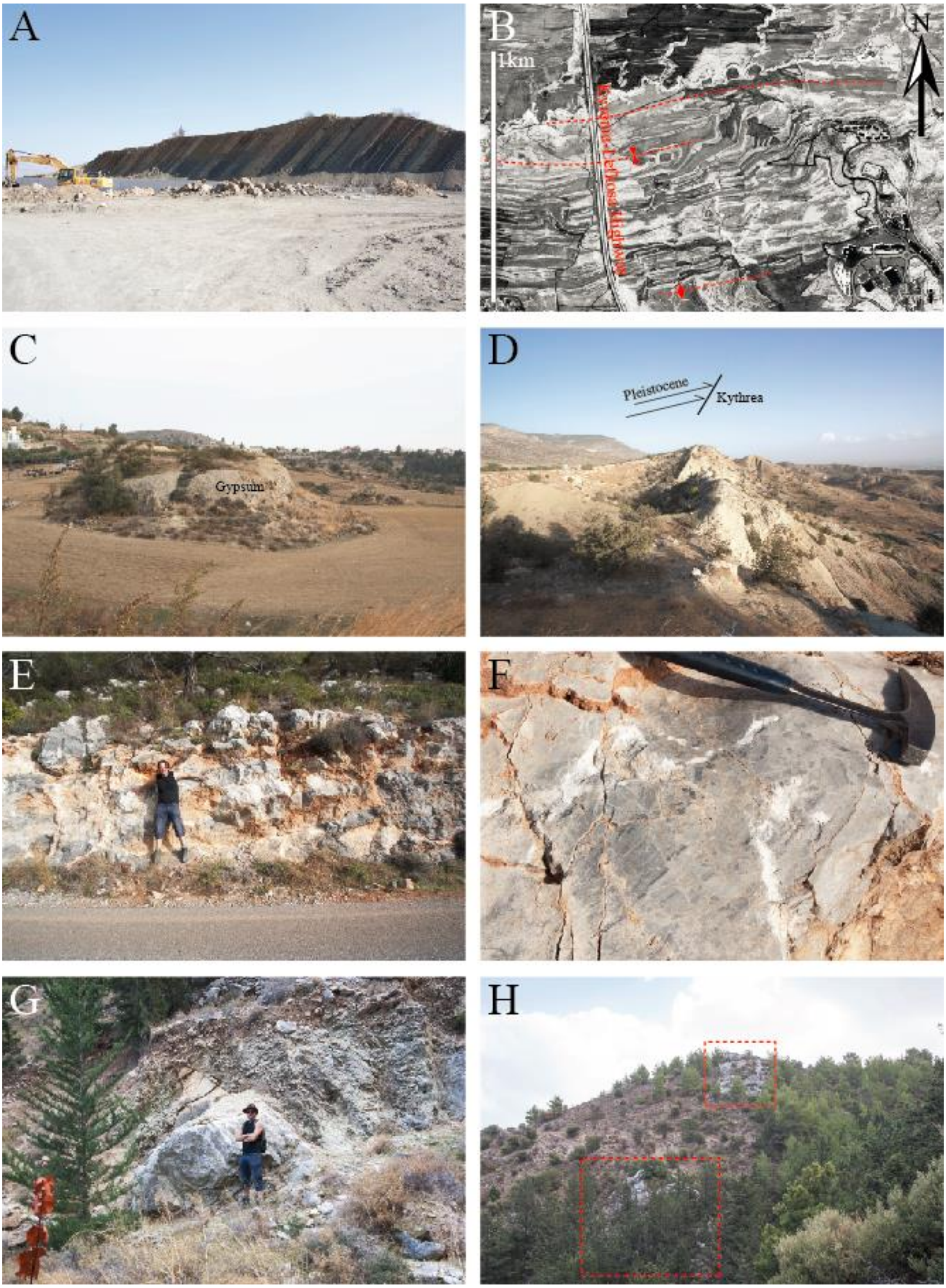
0.5 km wide, and then a north dipping domain (Figure 5, Taşkent fold). The syncline may be a footwall-syncline cut by a thrust fault, but the relationship is not exposed.

North of that fold-set, the bedding dips consistently ~50˚N until the village of Dikmen, just south of the Kyrenia range. The Kythrea Group is very poorly exposed and along the section strip map, a large military base covered much of the area, which limited access. We sampled an east-plunging anticline revealed on satellite images with two dip measurements to the west of the section line (Figure 5, Dikmen fold). To constrain the poorly exposed or inaccessible geology north of Dikmen, we made field observation ~5 km to the east of our section line at the village of Taşkent. We found that Kythrea rocks continue to dip steeply northward. We also found a large exposure of sub-horizontal Messinian gypsum (Figure 4C) that may unconformably cover steeply dipping rocks of the Kythrea Group, although the contact is not exposed.

Trypa and Lapithos Groups

North of Dikmen, Trypa Group marbles are juxtaposed with the Kythrea Group along a locally steep, north-dipping thrust contact (Figure 5, Kyrenia fault zone). Above that thrust, we found a 100 m thick marble unit and then a ~60 m thick bed of steeply dipping pelagic chalk that contains marble clasts and volcanic rocks. We traced that exposure of Lapithos Group rocks for a few kilometres eastward. The chalk is overlain by the main Trypa Group unit of the Kyrenia range. The contact represents either a thrust repetition or an overturned limb with the lowest marble unit being an olistolith.

Figure 4: A) Road cutting through Kythrea Group rocks just north of NETU, showing the steeply dipping but undeformed interbedded sandstones and mudstone which make up the Group. B) ORBIS satellite imagery of the area just north of NETU showing prominent sandstone beds which define folding in the Kythrea Group rocks and a footwall cut-off. C) A shallow dipping gypsum outlier unconformably overlying steeply dipping Kythrea Group rocks. D) Shallow north dipping siltstones of presumed Pleistocene age onlap onto steeply dipping Kythrea Group rocks on the Beş Parmaklar road. E) Bedding preserved in recrystallized limestone of the Trypa Group. F) Foliated Trypa Group marble cut by tiny faults, and brecciated. G) A ~3 m wide boulder of Trypa Group marble within the lowest 10 m of the Lapithos Group limestones. H) Two large (~30 m) Trypa Group olistoliths within the Lapithos Group. These are mapped in Figure 6.


A
B
1km
N
C
Gypsum
D
Pleistocene
Kythrea
E
F
G
H

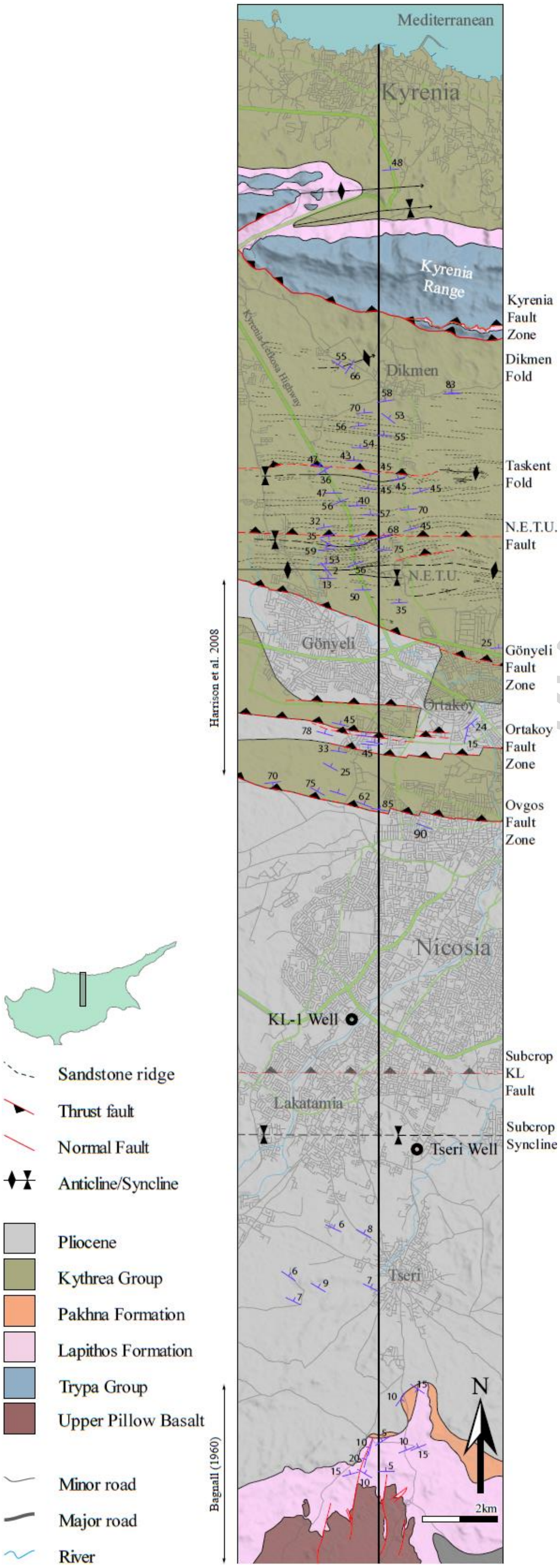

Mediterranean
Kyrenia
Kyrenia Range
Kyrenia Fault Zone
Dikmen
Dikmen Fold
Taskent Fold
N.E.T.U. Fault
N.E.T.U.
Gönyeli
Gönyeli Fault Zone
Ortakoy
Ortakoy Fault Zone
Ovgos Fault Zone
Harrison et al. 2008
Nicosia
KL-1 Well
Subcrop KL Fault
Lakatamia
Subcrop Syncline
Tseri Well
Tseri
N
2km
Bagnall (1960)
Sandstone ridge
Thrust fault
Normal Fault
Anticline/Syncline
Pliocene
Kythrea Group
Pakhna Formation
Lapithos Formation
Trypa Group
Upper Pillow Basalt
Minor road
Major road
River

Figure 5: Strip map along the line of the section showing representative dip data, lithostratigraphic boundaries, and major structures. We outlined continuous outcrops of sandstone in the Kythrea Group based on ORBIS satellite images. We included a road map from 'Open Street Map' to highlight the extent of the urban areas which obscured the geology. Note that areas of the map based on the USGS report of Harrison et al. (2008) and the Geological Survey of Cyprus memoir (Bagnall 1960) are demarcated on the left-hand margin. Map projection is in WGS84 UTM 36N.

The Trypa Group marble is largely made up of massive, medium to coarse-grained recrystallized limestone and dolomite. The marble locally contains relict bedding (Figure 4E), but in the upper part of the Trypa Group, along the northern side of the range, a well-developed mylonitic foliation has obliterated the bedding. We saw the foliation in many places. Towards the contact with the overlying Lapithos Group, we saw a mylonitic foliation overprinted by a cataclastic foliation that was parallel to the mylonitic foliation, and subparallel to the contact with the overlying Lapithos Group (Figure 4F). The upper surface of the marble is sharp and in places uneven. Locally, for example on the Beş Parmaklar road over the Kyrenia range, we found shallow northward-dipping siltstones that fill valleys in the marble (Figure 4D), and those are similar in appearance to the Pleistocene rocks we found elsewhere (Palamakumbura & Robertson, 2016).

Northern Kyrenia range

To the south of the town of Kyrenia, the Kyrenia range splits into two major ridges, separated by a valley that contains the Lefkosa-Kyrenia highway. The valley is defined by an east-plunging syncline, with Kythrea Group rocks in the core. The northern ridge is an east-plunging anticline that contains Trypa Group rocks in its core. Westward along the ridge, Trypa Group rocks are thrust over Lapithos Group rocks along a conspicuous contact, high above the highway (Figure 5, highway fault). We interpret the northern ridge as a hanging wall anticline above a thrust fault with an eastward decreasing displacement. The fault becomes a fault-propagation fold before it disappears eastwards.

On the northern side of the range, 1 km northwest of Malatya, we mapped irregular isolated blocks of Trypa Group marble up to ~0.5 km in size, floating in pelagic limestones of the Lapithos Group (Figure 6), and we interpret those as olistoliths. We also observed 2-3 m sized marble blocks and noted that they were not fault-bound (Figure 4 G & H). The large blocks were associated with a vertical ~N-S trending fault that dramatically affects the thickness of the Lapithos Group locally. The faults may be sealed by the upper stratigraphy of the Lapithos Group and are sealed by the Kythrea Group, which constrains the age of those faults as pre-late Eocene. The bedding there dips very steeply northward, and if we restore that bedding to horizontal, the steep faults become normal faults.

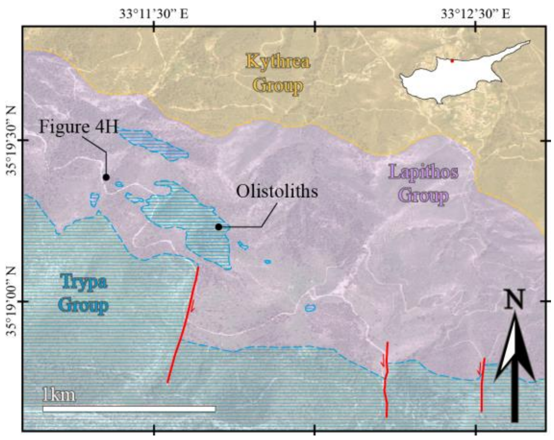


Figure 6: Small scale map of olistoliths of Trypa Group marble associated with growth faulting in the Lapithos Group, near Malatya in the eastern part of the Kyrenia range.

We found olistoliths like those in Malatya in many locations along the northern side of the Kyrenia range. They are characterised by irregular shapes, are discontinuous along the strike direction of the range, and are surrounded by pelagic carbonates of the Lapithos Group. In a few places, the marble ridges are more thin and elongate, and may be either be olistoliths or thrust fault repetitions, for example at St. Hilarion castle.

The pelagic limestone of the Lapithos Group is generally fractured and folded, contains localised shear fabrics, and is more deformed than the overlying Kythrea Group. We found no evidence, however, that post-Eocene rocks cover shear deformation in the Lapithos Group within the central part of the range. Instead, we observed that where the pelagic limestone was covered by competent conglomerates, shearing was parallel to the conglomerate-limestone contact, and conglomerate lenses were incorporated in the deformed chalks, suggesting that the shearing of the chalks is largely or entirely related to the Miocene deformation.

Above the olistolith-bearing Lapithos Group, we found a thick sequence of poorly exposed and steeply dipping Kythrea Group rocks, which included ophiolite-bearing conglomerates at the base. The Kythrea Group rocks are intensely folded, and/or slumped, and so we could not collect a representative dip measurement. The Kythrea Group rocks are covered by shallow dipping Pleistocene sedimentary rocks (Palamakumbura & Robertson, 2016).

4.2 Building a balanced cross section

We build our cross section from the southern edge of the Mesaoria basin where we use surface data from a geological map (Bagnall, 1960), and correlate that to the KL-1 and Tseri boreholes (Cleintuar et al., 1977). The top of the Troodos ophiolite around the southern basin edge must plunge relatively steeply towards the north. We find that the Pliocene and Messinian units must thicken significantly from the southern basin margin towards the Ovgos Fault as pointed out by Cleintuar et al. (1977).

To the north, in the central part of the Mesaoria basin, Harrison et al., (2008) reported stratigraphic repetition and folding (from an unpublished dip log) in the KL-1 borehole. We follow a basic interpretation of Cleintaur et al. (1977), which was based on correlation between well data, and infer a single shallow dipping thrust fault. The repetition of basal strata of the Lefkara Formation strongly suggests that the thrust fault penetrates the underlying Troodos ophiolite rather than shallowing out into the basin cover. Stratigraphic data from the KL-1 well show that Messinian and Pliocene rocks thin significantly over the hanging wall of the thrust fault, which partially constrains the onset of fault movement to Messinian or pre-Messinian times. Kythrea Group rocks must thicken northwards, to account for the large volumes of exposed rocks.

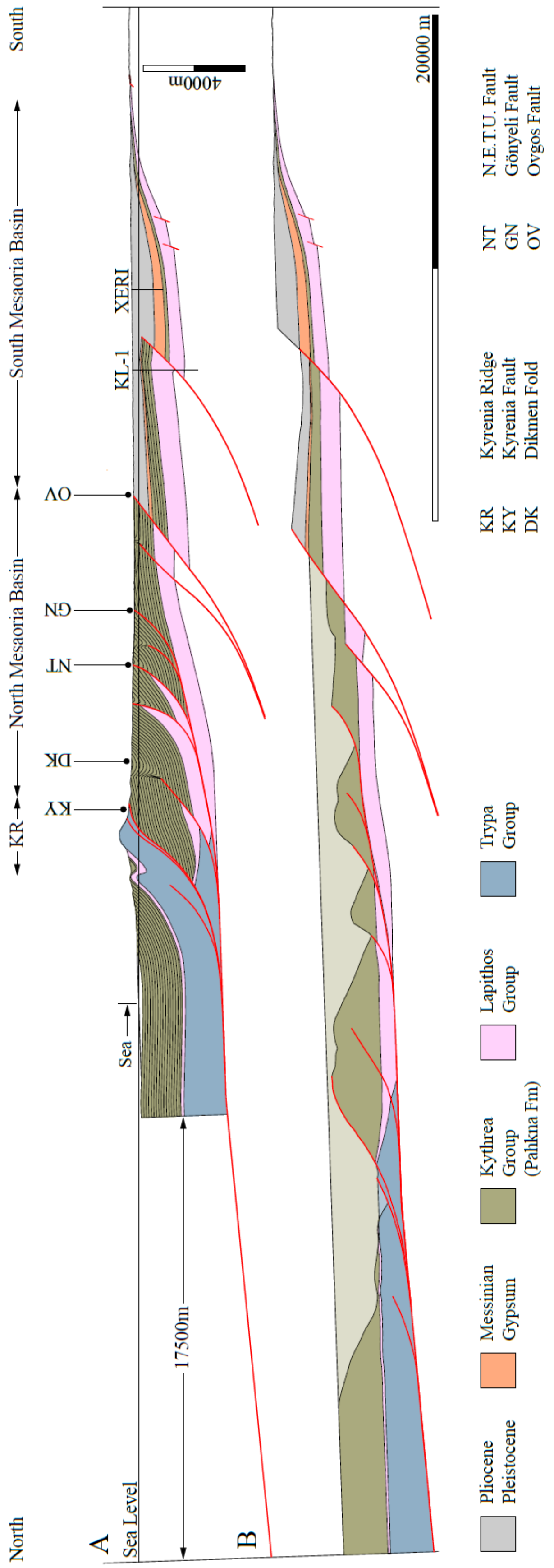
Moving northwards, we next reconstruct the Ovgos fault zone, which marks the southern boundary of the Kythrea Group. Harrison et al. (2008) found evidence for a series of thrust faults, which are associated with steep north-dipping to vertical bedding. The Ovgos fault zone was penetrated by the Lefkoniko borehole 35 km to the east. This penetrated ~2km of Kythrea Group rocks, which had been thrusted over a sequence of the Pakhna Formation and Lapithos Group and ophiolitic basement (Harrison et al., 2004). This relationship demonstrates reverse motion along a pre-existing normal fault or a steeply north-dipping basin margin. We treat the Ovgos fault as a relatively steep, reactivated listric normal fault, so as to explain the sudden and major change in the thickness of middle Miocene stratigraphy, and the appearance of the Kythrea Group. In our section, we interpret a fault splay and associated fault propagation fold within the hanging wall of the Ovgos fault. This interpretation is consistent with the steep splayed fault zone imaged in offshore seismic sections of the belt, in which a fault zone separates a thin and wedge shaped southern Mesaoria basin from a thick package of acoustically distinct Kythrea Group rocks (Calon et al., 2005a). The steep fault geometry we interpret for the Ovgos fault creates a steep north-dipping hanging wall, which folded the overlying rocks and partially accounts for the overall steep north-dips in the Kythrea Group. Our interpretation does not preclude oblique motion as suggested by Harrison et al. (2008) and Robertson and Kinnaird (2015).

North of the Ovgos fault zone, we have no constraint on the subsurface structure beneath the town of Gönyeli, and so we assume that there was no faulting there as the simplest case. The first exposures of the Kythrea Group north of Gönyeli are generally very steeply dipping, requiring a thrust fault with >1 km of displacement.

North of NETU, the Kythrea Group contains consistent domains of steeply dipping bedding that must be explained by thrust fault imbrication. We find that the steep dips are most straightforwardly recreated by ramp-flat fault shapes associated with thin-skinned thrusting within the upper ~3 km of stratigraphy. We choose an intermediate decollement along the Kythrea-Lapithos contact, as there are no older rocks incorporated in the thrust belt. We make a wedge-shaped Kythrea basin, which allows us to fill space toward the hinterland with fewer inferred thrust faults. Within that basin wedge we interpret a series of listric thrust faults, some of which reach the surface as we observed in the field, and some that we associate with fault propagation folding, to recreate large scale folding that we observed north of NETU.

Harrison et al. (2004) reported that ~100 m deep boreholes on the southern side of the Kyrenia ridge penetrated Kythrea Group rocks below the Trypa Group. This demonstrates that the Kyrenia thrust is at first relatively shallow dipping. We make a hanging wall cut-off in overlying Trypa Group units to minimize the amount of shortening on that fault, and assume that the modern surface of the Trypa Group represents its original top surface. We reconstruct folding of the Lapithos Group around the valley cutting the Kyrenia range, south of the town of Kyrenia. We and then reconstruct the southern ridge as a hanging wall anticline above the Kyrenia thrust, and the northern ridge as a fault propagation fold, that was steepened by tilting of the underlying thrust sheet.

Figure 7: A) Balanced and viable cross section of the Kyrenia fold-thrust belt. B) Retro-deformed cross section of the Kyrenia fold-thrust belt, at the same scale.



5. Discussion

5.1 Structural model evaluation

Our balanced cross section of northern Cyprus shows a minimum of 13.5 km of shortening in a thin-skinned belt plus a minimum of 4 km shortening on thick-skinned thrust faults. We successfully incorporate our field observations into a viable structural model. This model is the simplest structural model that accounts for all observations, but inevitably relies on some assumptions. We therefore first discuss how alternative assumptions may affect the style of deformation and magnitude of shortening in our model.

We infer both thick-skinned and thin-skinned thrusting because we found evidence for these structural styles in published seismic lines, well data, and our own field data. The first two major faults in the Mesaoria basin displace the Troodos ophiolite, and control the thickness of sediments in the Mesaoria basin. We interpret the KL and Ovgos fault as thick-skinned, as there is no obvious mechanical stratigraphy within the Troodos ophiolitic rocks in which a decollement should form. Our interpretation is consistent with the structural style in the more southerly thrust faults in the offshore part of the belt. If the Ovgos and KL faults were thin-skinned, they would most likely sole into the basal decollement beneath the Kythrea Group, and the incorporation of Troodos rocks would have to be the result of topography on the top Troodos surface, perhaps caused by earlier normal faulting. This would not significantly change our shortening estimate, but would require that there was no regional flexure in order to fill space in the hinterland.

We find that the steep bedding in the Kythrea Group is most simply reconstructed as thin-skinned thrusting above a moderately-dipping decollement (2°). That decollement likely reactivated the basement-basin interface in the deeper northern part of the basin, and then smoothly stepped up into the Kythrea Group towards the south. This was the simplest structural configuration to create the steep dips, whilst omitting deeper stratigraphy that we did not observe anywhere along the belt. If that deformation was instead created by thick-skinned thrusting, a smaller amount of shortening would be required, but on unrealistically rapidly steepening faults, within a much steeper regionally dipping basin. Our estimate of the depth of the intermediate decollement below the Kythrea Group is based on depth to magnetic basement calculations and we constrained the minimum thickness of the Kythrea Group in the Lefkoniko borehole (Cleintuar et al., 1977). Maintaining steep bedding with a thinner repeated sequence of Kythrea Group rocks would require many more thrust faults than we find evidence for. A thicker sequence than we reconstruct is possible, but larger displacements would be needed on each fault to recreate steeply dipping beds. Pelagic deposition dominated the basin evolution until middle Eocene times, meaning that Kythrea Group rocks likely filled relief in an older under-filled basin.

5.2 Timing of faulting, uplift and subsidence in the Miocene Kyrenia belt

The age of initial thrusting in the Kyrenia fold-thrust belt is constrained by the age of the upper-most formation in the Kythrea Group stratigraphy. This is the Lapatza Formation, which is thought to be Tortonian or lower Messinian (McCay et al., 2013). The wide domains of steeply dipping Kythrea Group rocks we mapped at the surface indicate significant erosion in the belt during or after a major phase of thrusting. We estimate up to ~1 km erosion based on the rotation and restoration of steep beds back to horizontal. This erosion may have corresponded to the Messinian regional unconformity (Calon et al., 2005b; Palamakumbura & Robertson, 2016): offshore seismic data show that this regional marker cross-cuts deformation in the belt, and has then been deformed by folding (e.g., Calon et al., 2005). The Pliocene Mirtou Formation reportedly covers the unconformity onshore (Baroz, 1979). We thus estimate a late Tortonian (~9 Ma) to pre-*Messinian Salinity Crisis* (~6) Ma age interval for major thrusting in the range.

We infer thickening of Pliocene rocks in the footwall of the KL fault, and syn-thrust packages have been interpreted in Pliocene and younger sedimentary rocks in the offshore Larnaka and Latakia ridges (e.g., Calon et al., 2005; Symeou et al., 2017). This implies that these thrust culminations remained active to some extent after Miocene times. Finally, Pleistocene and younger rocks are essentially undeformed by thrusting in the belt (e.g., Harrison et al., 2004; Palamakumbura & Robertson, 2016), and are only gently folded in offshore seismic lines, meaning that convergence is presently entirely accommodated to the south(west) of Cyprus.

5.3 Accommodating continuous Africa-Eurasia convergence through time

Most of the deformation we document affected the well-exposed Miocene and younger rocks of the fold-thrust belt, and was therefore Miocene or younger in age. However, a phase of Eocene shortening has also been interpreted in the Lapithos Group and Trypa Group, and has been used to infer Eocene collision of northern Cyprus with the Taurides (e.g., Robertson &

Woodcock, 1986; Robertson et al., 2014). Such an interpretation is important for the restoration of the location of the Africa-Europe plate boundary since the late Eocene.

Until the middle to late Eocene (~40-35 Ma), Africa-Eurasia convergence was accommodated by shortening and accretion of the Taurides (McPhee et al., 2018). Plate circuit reconstructions (e.g., Seton et al., 2012) in combination with kinematic restorations of Anatolia (Gürer & van Hinsbergen, 2018) predict that 300-400 km of the total 450-600 km post 35-40 Ma Africa-Europe convergence that occurred at the longitude of Cyprus, must have been accommodated by active subduction to the south of the Taurides (McPhee et al., 2018a; 2018b). An Eocene Tauride-Kyrenia collision would require that a subduction plate boundary that accommodated hundreds of kilometres of convergence was located to the south of Kyrenia, which is problematic.

First, there is no evidence for pre-Tortonian thrusting, let alone subduction, in the Mesaoria Basin. The Kythrea Group is continuous in terms of lithostratigraphy and biostratigraphy on either side of the Kyrenia Fault, when restored for Tortonian and younger deformation that we reconstruct. We found no evidence for deformation in the Mesaoria Basin that predates the formation of the Miocene Kyrenia fold-thrust belt. Our preferred structural model predicts an inverted normal-fault at the edge of the Kythrea Group depocentre, as in the interpretations of e.g., Calon et al. (2005), Robertson and McCallum, 1990, and Harrison et al. (2004), in which the lithostratigraphy of the Mesaoria basin has been interpreted as a southward transition from a shallow platform to deep marine basin.

Inferring that 300-400 km of plate convergence was accommodated by subduction to the south of the Troodos ophiolite is equally unlikely. There is no evidence for subduction south of the Troodos ophiolite in the period after the Mamonia fault zone was covered by Cretaceous sedimentary rocks (Urquhart & Banner, 1994). Seismic stratigraphy across major faults associated with the modern Cyprus trench, the development of flexural basins (e.g., Hall et al., 2005; Symeou et al., 2018), and structural and stratigraphic constraints on the onset of upper-plate deformation in southern Cyprus all suggest that the modern trench location formed only in latest Miocene or Pliocene time. In addition, the Troodos, Hatay, and Baer-Bassit ophiolites have been widely interpreted as a once contiguous microplate (e.g., Morris et al., 2006; Inwood et al., 2009). Subduction below the Hatay and Baer-Bassit ophiolites ended when the microplate thrusted onto Arabia around 70 Ma (Al Riyami & Robertson, 2002; Kaymakci et al., 2010). Troodos was likely obducted at the same time or just after the Hatay and Baer-Bassit ophiolites, but onto the African margin (Barrier & Vrielynck, 2008; Maffione et al., 2017). These Arabian ophiolites and the Troodos ophiolite currently form an E-W belt at the same latitude. Inferring a plate boundary between Troodos and Africa, but not between the Baer Bassit and Hatay ophiolites and Arabia, would restore the Troodos 300-400 km north of the Arabian ophiolites in the Eocene, which is highly unlikely in the view of their common rotation and formation history. It follows that if subduction below the Hatay and Baer-Bassit ophiolites stopped in Cretaceous times, subduction most likely also stopped below Troodos because of its obduction onto the stretched African margin, which was followed by slab break-off (the Arabia slabs of van der Meer et al., 2018). We infer that the Troodos ophiolite was a passively covering the north African margin until it collided with, and accreted to, the upper Anatolian plate in the Miocene.

5.4 Re-evaluating Eocene uplift and deformation

The only kinematically feasible place to accommodate hundreds of kilometres of continuous post-Eocene convergence is to the north of Kyrenia. We therefore briefly revaluate the four lines of argumentation for an Eocene phase of shortening in Kyrenia that would preclude this.

First, a regional unconformity between the Lapithos Group and Kythrea Group demonstrates uplift of northern Cyprus and surrounding areas in Eocene time. This uplift has been interpreted to reflect Eocene shortening (McCay & Robertson, 2012; Robertson et al., 2013). Uplift however, does not require shortening, and this phase of uplift is not unique to northern Cyprus. Evidence for contemporaneous uplift, emergence, or erosion has been reported in e.g., the Amanos Mountains of NW Arabia (Boulton, 2009; Duman et al., 2017) the Levant Margin (e.g., Brew et al., 2001; Gardosh & Druckman, 2006; Gardosh et al., 2006; Hawie et al., 2013b), and the Levant Basin (Bar et al., 2013), and was concluded from low-angle unconformities in the Arabian margin stratigraphy in the Zagros fold-thrust belt (Hessami et al., 2001).

The stratigraphy of northern Cyprus shows that open water deposition of the Lapithos Group was followed by uplift, and upon renewed subsidence, deposition of a sequence of foredeep deposits with upward -increasing deposition rates. This is typical of forebulge to foredeep transition (e.g., DeCelles et al., 1996) that we would predict with the approach of trench below the Taurides. Plate bending and forebulge formation need not be directly related to subduction below the Taurides, but may have been a response of the Arabian continent to the east during its approach towards Eurasia, prior to collision in Oligocene time (McQuarrie & van Hinsbergen, 2013). Upwelling in the Afar region, combined with a eustatic sea level fall (Gardosh et al., 2006) may have further accentuated the erosion phase.

Second, an interpreted repetition of Trypa Group and Lapithos Group rocks was previously used to argue for pre-Miocene thrusting (e.g., Robertson & Woodcock, 1986; Robertson et al., 2012; Robertson & Kinnaird, 2015). Repetition of older stratigraphy does not demonstrate old thrusting, unless it is sealed by younger sediments, and this has not been unambiguously observed for the Trypa Group (see for example the maps in Baroz, 1979). We consider it much more likely that these large blocks, mainly on the northern side of the Kyrenia range, are olistoliths within the Lapithos Group, because they are highly irregular in shape, discontinuous along strike of the Kyrenia ridge, and in several locations field relationships demonstrate that these blocks are surrounded on all sides by the Lapithos Group (Figure 4 G & H, Figure 6). These blocks likely detached from a high relief marble range in a submarine setting (Robertson et al., 2013). We thus reinterpret many thrust faults along the range as olistoliths. Otherwise, complex, laterally discontinuous and irregular faults with a combination of thrusting and normal displacement are required to explain those blocks.

In some places (e.g., at St. Hilarion Castle) we saw repetition of the Trypa Group and Lapithos Group, which could potentially be explained by thrust faulting. Those inferred faults however, are not sealed by the Kythrea Group, and are thus more simply explained as Miocene faults. Mesoscale thrust faults that involve both the Trypa Group and Lapithos Group have been observed in several places along the Kyrenia range (for example at *Sevili Tepe*, Robertson & Kinnaird, 2015). The omission of Miocene rocks in some of the Lapithos Group-Trypa Group thrust repetitions is simply explained by small thrust displacements.

It has been noted that the Lapithos Group was affected by more intense fracturing and shearing compared to the overlying Kythrea Group. We did not observe this difference in deformation in our study area, but it has been reported elsewhere in the range, and has been used to argue that the Lapithos Group was affected by thrusting prior to the deposition of the Kythrea Group (Robertson & Woodcock, 1984; Robertson & Kinnaird, 2015). Much of the deformation in the Lapithos Group may be attributed to Miocene thrusting, and differences in rheology of the Kythrea Group and Lapithos Group: they have very different textures, compositions and bulk rock properties. Shearing may also have been induced by emplacement of olistoliths and the deposition of debris flows onto weakly lithified and uncompacted sediments of the Lapithos Group. In the eastern parts of the range, the Kythrea Group reportedly covers sheared Lapithos Group rocks, and cleaved pelagic limestone pebbles believed to be from the Lapithos Group have been reported in the overlying olistostrome (Robertson & Woodcock, 1986). This suggests some active deformation locally and in the source region of the upper Eocene to Oligocene debris flows and olistostrome. Despite this, there is no evidence for a tectonic overburden in Eocene or Oligocene times that could have caused deformation in the Lapithos Group.

Finally, the non-metamorphic Kantara limestone clasts and olistoliths remain an enigmatic feature. Previously, it was proposed that the Kantara olistoliths were derived from southern Anatolia to the north of Kyrenia (Robertson & Woodcock, 1986), which would place the Kyrenia range next to southern Turkey during their deposition (McCay & Robertson, 2012). We see no problem invoking a more proximal source for those limestones: the modern subaerial extent of Cyprus developed in the Neogene and so olistoliths may rework sediments derived from submerged or underthrusted platform rocks to the northeast, but in the vicinity of Cyprus.

5.5 Cretaceous and younger geology of Cyprus: a record of two separate subduction zones

Metamorphism of the Trypa Group was previously postulated to result from Late Cretaceous under-thrusting below the Taurides (Robertson et al., 2013; Robertson & Kinnaird, 2015). This is very unlikely because there is evidence for Late Cretaceous northward emplacement of the Antalya-Alanya Nappes and Late Cretaceous SSZ ophiolites onto the Taurides (e.g., Özgül, 1976; McPhee et al., 2018), which implies that a southward -dipping subduction zone was active there. The burial and metamorphism of the Trypa Group occurred in the same time period as the emplacement of the Troodos ophiolite against the Mamonia Complex. The exhumation of the metamorphosed Trypa Group occurred just prior to the arrest of relative motion between the Troodos ophiolite and the Mamonia Complex (Urquhart & Banner, 1994). These points, and the occurrence of a large block of ophiolitic melange in the northeast Kyrenia range (close to the village of Kantara), lead us to interpret that metamorphism of the Trypa Group was caused by under-thrusting below the Troodos ophiolite in Late Cretaceous time. This places the Trypa Group at the northern margin of Africa from Late Cretaceous times.

The exhumation of Trypa Group rocks from greenschist-facies, equivalent to mid-crustal depths, must have occurred in late Campanian to Maastrichtian times by tectonic exhumation.

These metamorphic rocks are unconformably covered by open marine sediments with no intervening transgressive sequence (Robertson & Woodcock, 1986; Robertson et al., 2012), which requires extensional exhumation at the sea floor (Robertson & Woodcock, 1986; van Hinsbergen & Meulenkamp, 2006). We tentatively suggest that the top Trypa Group surface on the northern flank of the Kyrenia range, which contains contact-parallel mylonites and cataclasites, represents an extensional detachment that exhumed the Trypa Group rocks in the latest Cretaceous. Detailed future structural work is required to evaluate the kinematic and structural history of exhumation in more detail. Vine et al. (1973) based on magnetic survey data, suggested that the Troodos ophiolitic rocks extend in the subsurface as far north as the Kyrenia range and then stop abruptly. In our structural model, we propose that the abrupt change represents the southern break-away fault of a top-to-the-north detachment that exhumed the Trypa Group. This detachment must root to the north of the Kyrenia Range, in the Cilicia basin. This would make the Lapithos basin a post-obduction supra-detachment basin. A similar history of post-obduction exhumation is well-known for the Oman ophiolite, where exhumed metamorphosed Arabian crust is exposed to the north of the ophiolite along the Persian Gulf coast (Agard et al., 2010), and exhumation is thought to reflect post-slab break-off rebound (Duretz et al., 2016). We propose a similar history for Cyprus.

Burial of the Trypa Group below the Troodos ophiolite may have been coeval with the northward emplacement of the Antalya-Alanya Nappes onto the Taurides, and south-eastward emplacement of the Hatay and Baer-Bassit ophiolites onto Arabia (Morris et al., 2006; Inwood et al., 2009; Maffione et al., 2017). The multi-directional and coeval emplacement of those nappes is most simply explained by a model of westward invasion of an intra-oceanic subduction zone by a single east-dipping slab (Figures 8 and 9). This subduction zone consumed ancient oceanic lithosphere that separated the Taurides from Africa, and replaced it with Late Cretaceous oceanic lithosphere, which formed by SSZ spreading in the overriding plate. The ophiolites are remnants of that oceanic crust. The invasion model explains E-W spreading in the SSZ ophiolites shown by paleomagnetic data (Maffione et al., 2017; Morris et al., 2017), and vertical axis rotations of the ophiolites (e.g., Clube et al., 1985; Morris et al., 2002; 2006; Inwood et al., 2009). We interpret the Trypa Group as an allochthonous unit derived from the south-eastern margin of the Taurides (Figure 8). This would make it equivalent to the far-travelled Alanya nappes, which were likely derived from the eastern Tauride platform in the Cretaceous (Cetinkaplan et al., 2016). We interpret the Mamonia Complex as the para-autochthonous distal African margin, as well as rocks derived from the Eastern Mediterranean Ocean.

By late Eocene time, The Late Cretaceous SSZ oceanic lithosphere, which separated the Taurides from Africa, started to close by subduction into a north-dipping subduction zone. This subduction zone originated from the north, and by late Eocene time, had accreted the upper crust of the Tauride platform, and had fully subducted the lower crustal and mantle lithospheric underpinnings of the Tauride platform (van Hinsbergen et al., 2016; McPhee et al., 2018). The Trypa Group and its cover were part of the African plate after Late Cretaceous obduction of the Troodos ophiolite. This places the plate boundary between Africa and Eurasia (Taurides) the north of Kyrenia until the late Miocene onset of Kyrenia thrusting. The suture zone of the Eastern Mediterranean Ocean must thus be located beneath the Cilicia basin, between Kyrenia and southern Turkey. This suture is the along-strike equivalent of the Bitlis suture between Arabia and Anatolia. It cuts through an arcuate magnetic anomaly (Woodside, 1977) that stretches from the Hatay ophiolite, through Troodos, and on to the Antalya Ophiolite that represents the arc-shaped Cretaceous obduction front. The onset of thrusting in the Kyrenia Range around 9 Ma should then be considered as the start of Africa-Taurides continent-continent collision (Figures 8 & 9), a few million years later than the Arabia-Taurides collision, as a result of the paleogeographic shape of the margin. Ongoing Africa-Europe convergence after 9 Ma, totalling some 90 km, should thus have consumed the (stretched) continental crust, and the original continental crustal and lithospheric underpinnings of the Kyrenia fold-thrust belt. This continental crust should now be located deep in the subsurface, 90 km north of the Kyrenia range.

5.6 Implications for uplift of the southern margin of the Central Anatolian Plateau

We have argued so far that collision between Africa and the Taurides of southern Turkey occurred in late Miocene times, around 9 Ma, and therefore preceded uplift of the southern margin of the Central Anatolian Plateau. Evidence from river profiles, and dating of river terraces within the plateau drainage system (Cosentino et al., 2012; Schildgen et al., 2012a), as well as constraints on the timing of the development of an orographic rain shadow (Meijers et al., 2016) suggest that initial uplift (above sea level) occurred around 8-5 Ma, postdating the initial accretion of Kyrenia. We therefore do not conclude that collision of the distal African margin had an immediate response to Anatolian topography. We even consider it quite unlikely that the 8-5 Ma onset of uplift of southern Anatolia directly relates to collision with Africa. Such uplift would require shortening which, upon collision, would likely become distributed in the overriding plate (e.g., as in Tibet or Central Iran). There is no evidence for Miocene or younger shortening in the Mut basin (e.g., Fernandez-Blanco, 2014), and previously proposed dynamic topographic responses to mantle processes, such as slab break-off or delamination (Schildgen et al., 2014; Göğüş et al., 2017) may provide a better explanation for this uplift.

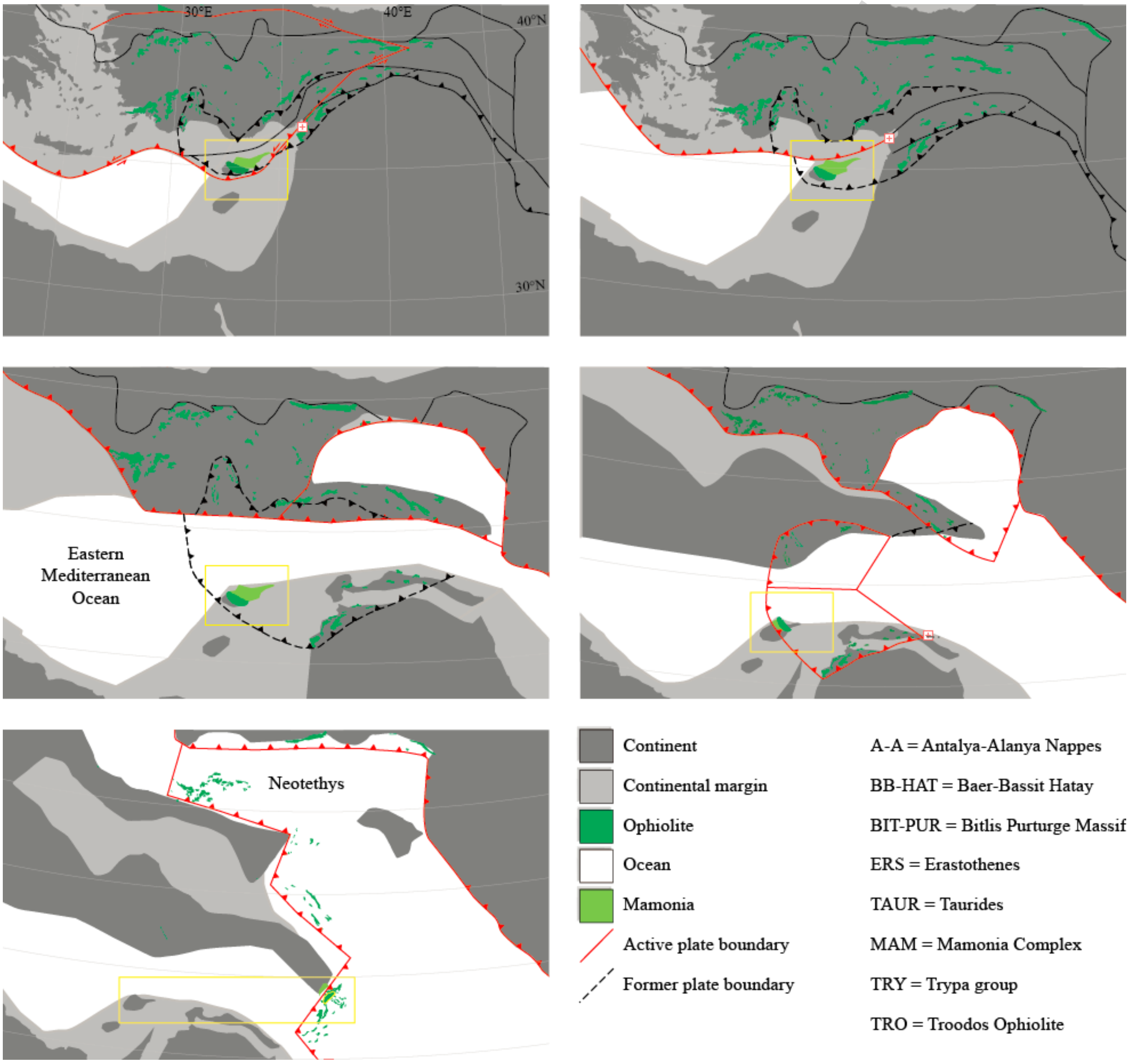


Figure 8: Evolution of Cyprus cast in the paleogeographic and plate kinematic evolution of the eastern Mediterranean region. Evolution of the Aegean and west Anatolian region follows van Hinsbergen and Schmid (2012); Central and Eastern Anatolia follows van Hinsbergen et al. (2016) and Gürer and van Hinsbergen (2018); intra-oceanic subduction evolution follows Maffione et al. (2017) and Maffione and van Hinsbergen (2018); Arabia-Eurasia collision

zone evolution follows McQuarrie and van Hinsbergen (2013); Caucasus reconstruction according to van der Boon et al. (2018).The Yellow box outlines the area of interest.

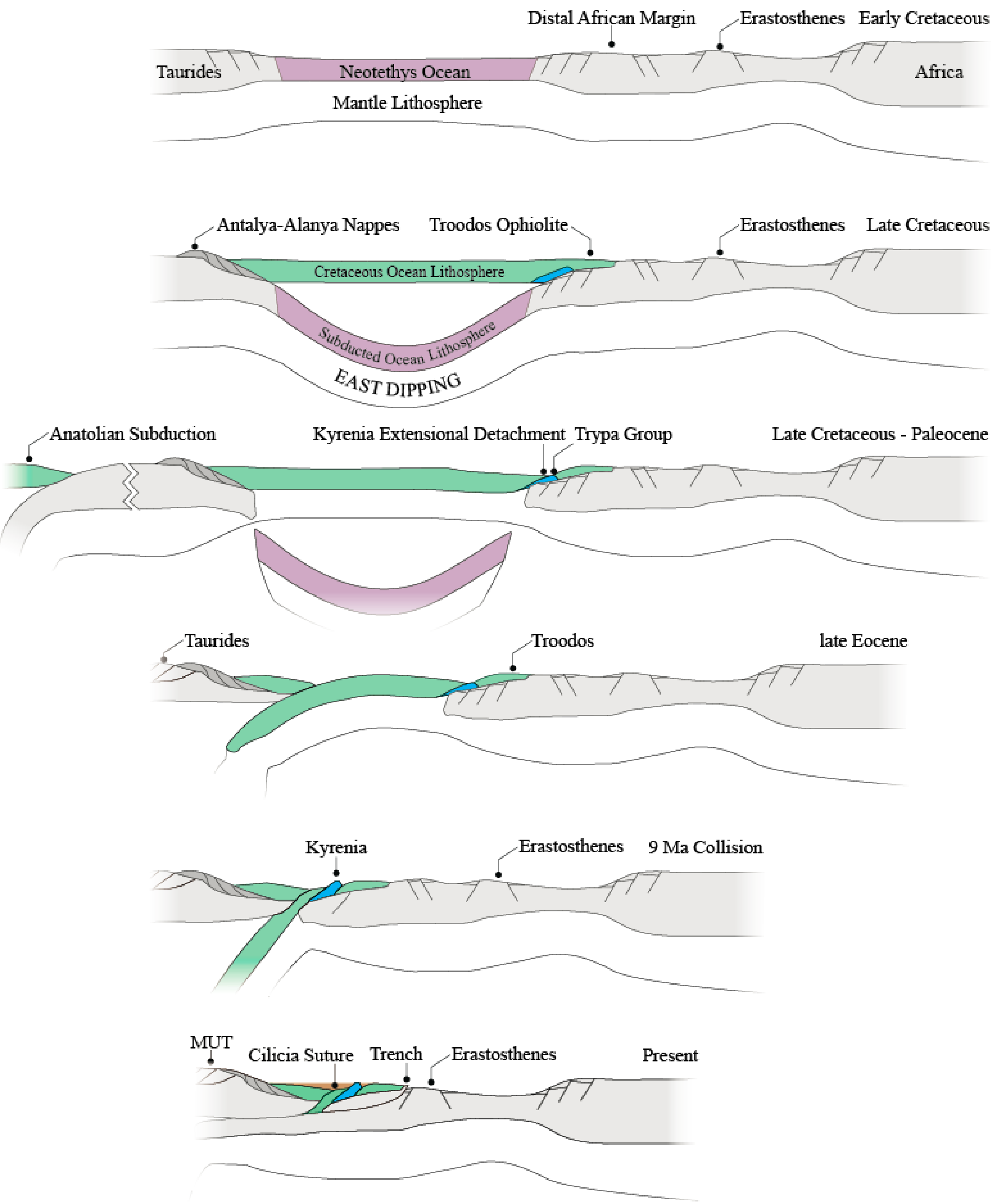


Figure 9: Sketch cross sections describing key phases in the evolution of subduction and accretion on Cyprus. A = Initial basins between the Taurides and Africa. B = Subduction invasion emplaced ophiolite onto Kyrenia in the South, and the Taurides in the North, and deformed the underlying passive margin. Note that the slab is east dipping in this section. C =

The east-dipping slab rolled back westward, and extension in the overriding plate exhumed metamorphosed Kyrenia range rocks. D = The southward migrating trench of the Anatolian subduction zone reached the ocean basin that separated Cyprus and the Taurides, and oceanic subduction continued on a north dipping slab. E = The ocean basin that separated Cyprus and the Taurides was consumed by subduction, and Cyprus was affected by thrusting as the passive margin rocks entered the trench. F = African extended passive margin underthrusts the Taurides.

The under-thrusting of the African continental margin below the Tauride fold-thrust belt would replace the pre-collisional oceanic foreland of the fold-thrust belt with a continental one, which would cause uplift. Geophysical analyses have suggested that at present, the northernmost African continental crust must be underlying the Tauride fold-thrust belt, to explain the high crustal thicknesses there (Abgarmi et al., 2017). Our kinematic model now allows us to calculate that this continental margin started to underthrust ~9 Ma ago, off-scraping the Kyrenia range fold-thrust belt at the front, and under-thrusting its underpinnings northward. Given the slow Africa-Europe convergence in the Neogene (Seton et al., 2012), this continental crustal front only arrived below the southern Taurides in late Pleistocene time.

Middle Pleistocene marine sediments reported on the southern edge of the Taurides showed that the southern Anatolian margin experienced very rapid uplift (~3 mm/yr) since only ~0.5 Ma (Öğretmen et al., 2018), causing up to 1600 m of uplift. We propose that the arrival of the subducting African continental margin may explain this recent, rapid uplift.

6. Conclusions

In this paper, we study the structural and tectonic history of Cyprus to evaluate when and where continent-continent collision occurred in the Eastern Mediterranean region and to evaluate possible dynamic effects. Our conclusions are summarized as follows:

- We predict that the Cilicia basin overlies a Miocene suture equivalent to the Bitlis suture to the east, which marks the closure of the former Eastern Mediterranean Ocean. The ocean basin was consumed by a northward-dipping subduction zone below the Taurides since middle or late Eocene time.
- Metamorphism of the Trypa Group of the Kyrenia Range of Northern Cyprus is best explained by southward obduction below the Troodos Ophiolite in

Cretaceous times. This obduction was a result of the westward radial invasion of a rolling-back trench that carried far-travelled supra-subduction zone ophiolites and accretionary prisms, including parts of the Mamonia Complex, towards circum-Eastern Mediterranean Ocean continents (Taurides, Arabia, Africa).

- The onset of the collision between the Taurides and Cyprus was marked by the formation of the Kyrenia fold thrust-belt, which was affected by 17.5 km of shortening largely between Tortonian and Pliocene times (~9-6 Ma).
- Closure of the ocean basin to the north of Kyrenia, and the onset of continental collision occurred too far south to be a driving force in the initial uplift of the Central Anatolian Plateau at 8-5 Ma. A lag period between collision and late, major uplift of the plateau margin at ~0.5 Ma corresponds to the time needed for Africa-Eurasia convergence to deliver African lithosphere ~70 km northward below the plateau margin, where it is currently imaged in geophysical data. This young, spectacular uplift is thus likely the response to Africa-Taurides continental collision.

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Graphical abstract