

Is there a Cretan Supradetachment Basin? Insights From Detailed Mapping on Northwestern Crete (Greece)

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Abstract The island of Crete in the South Aegean forearc exposes a fragmented and dismembered non-metamorphic nappe stack that is separated from underlying, exhumed high-pressure, low-temperature metamorphic rocks by an extensional detachment system. Exhumation and nappe thinning is thought to result from large-scale extension, which occurred mostly between ~20 and 13 Ma according to cooling ages. Such major extension normally forms surface depressions, but surprisingly, sedimentary basin sediments on Crete post-date the bulk of exhumation and are younger than ~11 Ma. Because the oldest sedimentary rocks do not rework metamorphic rocks, they were interpreted as a (late-stage) supra-detachment basin, although the tectonic reconstruction of the oldest sediments is challenging. Here, we provide a new detailed geological map of northwestern Crete where the oldest sediments are best exposed. We show that the stratigraphy contains several hiatuses because of tectonic reorganizations that separate superimposed basin systems that were bounded by different major faults accommodating first N-S and later E-W extension. We find that even the oldest of these fault systems, starting at 10.9 Ma and governing the oldest sedimentary units of the Topolia conglomerates, must already have cut through the Cretan Detachment. The sedimentary basins of northwestern Crete thus entirely post-date activity of this detachment. Final exhumation of Crete's HP-LT complex was instead likely related to erosion in uplifted footwalls of normal faults. Our results highlight a paradox that during Crete's crustal thinning and HP-LT rock exhumation, it maintained a high topography, and that all basin formation occurred during later fore-arc extension.

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1 Introduction

The Aegean region is well-known for its widespread extension and the exhumation of metamorphic complexes along extensional detachment systems (*Brun and Sokoutis, 2007; Jolivet et al., 1996; Jolivet and Brun, 2010; Lister et al., 1984; Ring et al., 2010*). These complexes expose previously buried nappes that formed by Cenozoic upper crustal accretion derived from African Plate lithosphere that subducted along the Hellenic trench (*van Hinsbergen et al., 2005a*) and that typically metamorphosed at high pressure and low temperature (HP-LT) (*Matthews and Schliestedt, 1984; Seidel et al., 1982*). Extension, which occurred mostly in Neogene time and is thought to relate to roll-back of the subducted African lithospheric slab (*Jolivet et al., 2013; Pichon and Angelier, 1981; Meulenkamp et al., 1988*), subsequently accommodated hundreds of kilometers of extension (*Gautier et al., 1999; van Hinsbergen and Schmid, 2012*), strongly thinning the original forearc crust (*Jolivet and Brun, 2010*). This thinning not only led to metamorphic rock exhumation along extensional detachments, but also to development of associated,

synchronous supra-detachment sedimentary basins (*Oner and Dilek, 2013; Sánchez-Gómez et al., 2002*). In that context, it is puzzling that the region with arguably the strongest crustal thinning, in the south Aegean region on the island of Crete, basin sedimentation only overlaps with the very final stages of metamorphic complex exhumation (*Seidel et al., 2007; van Hinsbergen and Meulenkamp, 2006; Zachariasse et al., 2011*).

The South Aegean region exposes a high-pressure metamorphic complex that reached peak metamorphic conditions in the earliest Miocene (~24-20 Ma, *Jolivet et al. 1996*) and that exhumed mostly before 15-12 Ma as suggested by low-temperature thermochronological data (*Marsellos et al., 2010; Thomson et al., 1998*). This metamorphic complex is separated by the N-S extensional Cretan Detachment from overlying, highly dismembered anchi-metamorphic nappes and ophiolites that collectively in western Greece, where extension is minimal, may reach a thickness of 20 km or more (*Jacobshagen et al., 1986*), but that on Crete are together not thicker than ~2.5 km (≤ 1 km without ophiolites *Bonneau 1984; Creutzburg and Seidel 1975*). This strong thinning may result from the combined effects of trench-normal extension resulting

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from roll-back, and trench-parallel extension related to oroclinal bending (*van Hinsbergen and Schmid, 2012*). Surprisingly, there is no sedimentary record that formed during the exhumation of the Cretan high-pressure metamorphic record: the widespread marine and non-marine sedimentary record does not get older than ~11 Ma (*Zachariasse et al., 2011*). Because the oldest sediments formed prior to the final exhumation of the HP-LT metamorphic rock units, it was speculated to invoke a supradetachment basin that formed only in the final stages of detachment-related exhumation (*van Hinsbergen and Meulenkamp, 2006*). However, detailed mapping and stratigraphic correlation in Central Crete limited that possible correlation to less than 1 Myr (*Zachariasse et al., 2011*).

Northwestern Crete exposes the best and most complete stratigraphy of Crete. Among the most impressive geological features are rugged and deep gorged mountain ridges of cemented conglomerates that are the oldest basin sediments in northwestern Crete: the Topolia Formation (Fm) that is devoid of HP-LT metamorphic rock debris (*Kopp and Richter, 1983*). Current models speculate about the tectonic setting in which these formed. They were proposed to form in an extensional setting where deposition of Topolia conglomerates occurs during final exhumation of the HP-LT rocks along the N-dipping Cretan Detachment (*Jolivet et al., 1996; Seidel et al., 2007; van Hinsbergen and Meulenkamp, 2006*). Alternatively, *Ring and Yngwe (2018)* infer that the exhumation occurred in a compressional setting where the Topolia Fm was deposited in depressions that resulted from erosion of uplifted, thrustured Tripolitza and Pindos nappes, whereby *Jolivet et al. (1996)*'s Cretan Detachment was instead interpreted as a top-to-the-SSW thrust fault. In their model, extensional basin formation occurred later (*Ring and Yngwe, 2018*). But so far, a systematic and detailed structural and stratigraphic mapping of the Topolia Fm and the subsequent (non-) marine sedimentary sequences up to the Quaternary to resolve the tectonic and sedimentary history, has not been performed.

To resolve the question about structural setting and position of the Topolia conglomerates relative to final exhumation, the first author conducted a detailed field study on the oldest basin units in northwestern Crete and mapped their contacts. This study gradually has extended to the entire Neogene succession. In this article, we summarize the architecture of the nappe stack of western Crete and present a detailed lithostratigraphy for the Neogene deposits of northwestern Crete. We present a detailed geological map and structural cross sections, which we will use to infer the main fault structures. Finally, we discuss the basin history and evaluate the relationship with exhumation of the HP-LT metamorphic rocks on northwestern Crete in the context of widespread Aegean forearc extension.

2 Geological Setting

2.1 Aegean Orogenic Architecture

Crete is a horst in the outer Hellenic arc that has been emerged in its modern configuration only since the Pleistocene (*van Hinsbergen and Meulenkamp, 2006; Zachariasse et al., 2011*). The island exposes an intensely deformed stack of Alpine nappes (*Bonneau, 1984; Creutzburg and Seidel, 1975*). The nappe stack of Crete extends southward to the so-called continental backstop which marks the boundary between Aegean lithosphere and an accretionary prism known as the Mediterranean Ridge that consists of Upper Cenozoic deep marine sediments that were scraped off from Eastern Mediterranean oceanic crust (*Huguen et al., 2006; Lallemant et al., 1994; Robertson and Kopf, 1998*) (Figure 1). To the north, Crete borders the deep-marine South Aegean Basin that formed in the Late Miocene to the south of the present volcanic arc (*Hsü, 1977*) (Figure 1).

The nappe stack exposed on Crete formed during ~100 Ma continuous NE to NNE-wards subduction of mostly thinned continental crust and overlying platform and deep basin sediments derived from a microcontinental domain ('Greater Adria', *van Hinsbergen et al. 2020*) that was separated since Triassic-Jurassic time from both Africa and Eurasia by ocean basins (*Jolivet and Brun, 2010; van Hinsbergen et al., 2005a*).

The superposition of the nappes is inversely related to their timing of subduction due to the process of foreland propagating accretion, whereby the lowermost nappe is the youngest and most external and the highest one the oldest and most internal part of the subducted Greater Adria lithosphere (*Jolivet and Brun, 2010; Schmid et al., 2020; van Hinsbergen et al., 2005a, 2020*).

An overview of stacking order, lithology, age, and metamorphism of nappes on Crete is given in *Fassoulas et al. (1994)*. From base to top, Crete exposes two nappes (Plattenkalk and overlying Phyllite-Quartzite) that experienced HP-LT metamorphism during the Late Oligocene to Early Miocene (*Jolivet et al., 1996; Seidel et al., 1982*). These are overlain by two essentially non-metamorphic nappes (Tripolitza and overlying Pindos nappes) and a composite Uppermost unit that includes ophiolites and associated intermediate and high-grade metamorphic rocks of Mesozoic age (*Seidel et al., 1981*) that likely relates to the subduction history prior to the arrival of Greater Adriatic lithosphere at the trench (*van Hinsbergen et al., 2020*). Figure 2 shows the distribution of the Alpine nappes on western Crete after *Creutzburg (1977)*.

2.2 The Nappe Stack of Western Crete

The lowermost nappe is the HP-LT metamorphic Plattenkalk unit that has a lower part of Permian to Upper Triassic platform carbonates and intercalated phyllites, a hiatus spanning the Middle Triassic, and an upper part of Jurassic to Eocene deep marine (turbiditic) meta-limestones with chert nodules and

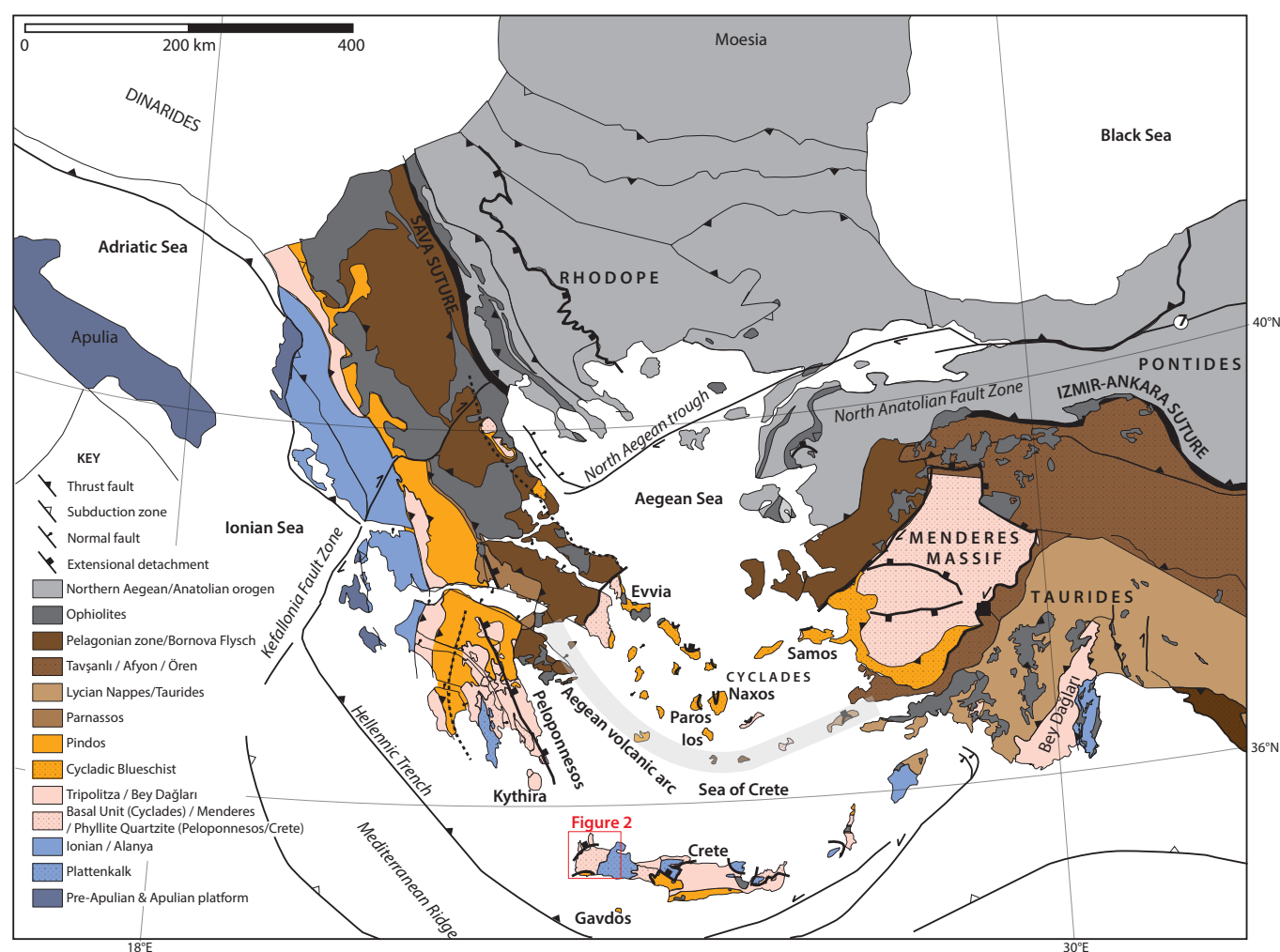


Figure 1 – Tectonic map of the Aegean region (simplified from *van Hinsbergen and Schmid* 2012).

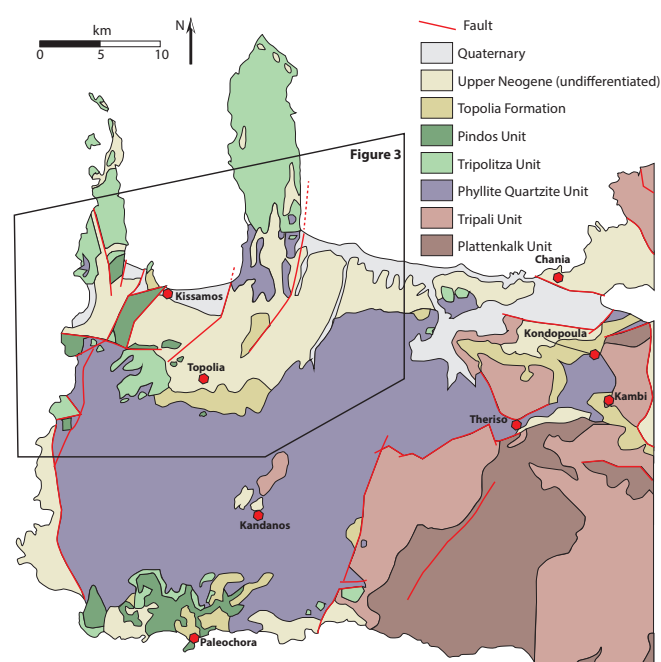


Figure 2 – Geological map of western Crete (after *Creutzburg* 1977). Box shows the studied area.

locally thin phyllites (*Bonneau, 1984; Creutzburg and Seidel, 1975; Epting et al., 1972; Hall et al., 1984;*

König and Se, 1980). The Plattenkalk Unit is in western Crete locally overlain by a separately mapped Trypali unit (*Creutzburg, 1977; EAGME, 1969*) (Figure 2) of unstratified shallow marine crystalline carbonates and local limestone breccia, and rauhwackes that is poorly dated as Late Triassic-Early Jurassic (*Kopp and Ott, 1977; Pomoni-Papaioannou and Karakitsios, 2002*); it is thought to be a separate facies of the lower Plattenkalk unit (*Hall and Audley-Charles, 1983*). Remnants of Oligocene siliciclastic meta-sediments (flysch) conformably overlying the Plattenkalk are known from central (*Bonneau, 1973; Bonneau et al., 1977*) and eastern Crete (*Fytrolakis, 1972*). Given the similarities in lithology and depositional age as well as in subduction age, the Plattenkalk is correlated to the Ionian zone exposed in the western Aegean foreland (*Schmid et al., 2020; Thiébaud, 1979*). The Plattenkalk was metamorphosed to ≥ 7 kbar and 380°C on Crete (*Seidel, 1978*).

The HP-LT metamorphic Phyllite-Quartzite nappe is widespread in western Crete (*Creutzburg, 1977; EAGME, 2002; Seidel et al., 1982; Jolivet et al., 1996*) (Figure 2) and consists of an Upper Carboniferous-Upper Triassic series of slates, quartzites (including meta-conglomerates), phyllites, coarse crystalline limestones, and locally gypsum/anhydrite

(*EAGME*, 1969, 2002). Conodont fossils show that the Phyllite-Quartzite unit is deep marine (*EAGME*, 2002; *Krahl et al.*, 1983, 1986). Massive gypsum accumulations are Late Triassic in age (*EAGME*, 2002) and in eastern Crete thought to have been deposited in shallow marine lagoons (*Dornsiepen et al.*, 2001; *Krahl et al.*, 1983, 1986). In eastern Crete, the Phyllite-Quartzite unit includes Upper Carboniferous-Lower Permian high-grade metamorphic rocks that are interpreted as a pre-Alpine crystalline basement that stratigraphically underlies the metasedimentary series (*Dornsiepen and Manutsoğlu*, 1994; *Romano et al.*, 2006; *Seidel et al.*, 1982).

The Tripolitza and Pindos nappes occur north and south of the present central Phyllite-Quartzite horst on western Crete (*Creutzburg*, 1977; *EAGME*, 2002) (Figure 2). The Tripolitza nappe is characterized by Upper Triassic to Middle Eocene shallow marine, mostly unstratified platform carbonates topped by Upper Eocene to Oligocene foreland basin deposits (flysch) (*Bonneau et al.*, 1977; *Creutzburg and Seidel*, 1975; *Bonneau et al.*, 1977).

Contacts between the Phyllite-Quartzite and Tripolitza units are tectonic, but the two units likely formed originally one coherent stratigraphic sequence, whereby the Phyllite-Quartzite once likely formed the stratigraphic base of the Tripolitza unit (*Dornsiepen et al.*, 2001; *Sannemann and Seidel*, 1976). The two units were disconnected, the Phyllite-Quartzite Unit underthrust to greater depth and was subsequently exhumed. The modern contact is a fault that is widely interpreted as the Cretan extensional Detachment (*Fassoulas et al.*, 1994; *Jolivet et al.*, 1996). However, certainly not all exhumation occurred along the Cretan Detachment because *Rahl et al.* (2005) showed that the Tripolitza limestones reached ~250 °C which - with typically low geothermal gradients in subduction settings - could represent a depth of perhaps 20 km. On par with their strongly dismemberment, thinning of the nappes above the Cretan Detachment must have considerably contributed to exhumation of the Phyllite-Quartzite and Plattenkalk units.

The Pindos nappe was thrust over the Tripolitza nappe, and on Crete composed mainly of deep marine Upper Triassic to Cretaceous limestones (partly crystalline) and associated cherts and overlying upper Paleocene and mostly Eocene shales and silt/sandstones interpreted as flysch (*Bonneau et al.*, 1977; *Bonneau and Fleury*, 1971; *Creutzburg and Seidel*, 1975; *EAGME*, 2002; *Seidel*, 1971).

Rocks of the heterogeneous Uppermost Unit, consisting of ophiolites and associated metamorphic rocks, are absent in western Crete and therefore this unit is not further discussed here. The total maximum thickness of the scattered remains of the upper nappes above the Cretan Detachment on western Crete is < 1 km (*EAGME*, 2002). The fission track data from western and central Crete suggest that the Phyllite-Quartzite unit was exhumed to 2-3 km depth around 15-12 Ma (*Marsellos et al.*, 2010; *Thomson et al.*, 1998, 1999)

showing that the thinning of the upper nappes was achieved prior to the onset of the Neogene sedimentation (*van Hinsbergen and Meulenkamp*, 2006; *Zachariasse et al.*, 2011).

3 Methods

The construction of the geological map for northwestern Crete is based on over four months of fieldwork by WJZ in the period 2019-2023. The original map was drawn on Google Earth (imagery date 9-1-2018) and based on hundreds of GPS fixed field observations and then projected on OpenStreetMap topographical data using MaPublisher software (<https://www.avenza.com/mapublisher>) (Figure 3). For reasons of readability, most geographic information had to be omitted in Figure 3. The names of the villages and gorges that we refer to in the text are, however, given in the full PDF version of the geological map in Supporting Information (SI) Item 1A. This SI Item also shows relevant bedding orientations. The new geological map on Google Earth is available in SI Item 1B. Geological information for each location on which the map is based is available in SI Item 2 including legend for acronyms used. When SI Items 1B and 2 are opened, all data are projected on Google Earth. The geological map is stripped of dunes, beach sands/gravels, beach rocks, and alluvial deposits.

4 Results

4.1 Additional Observations on the Nappe Stack of Northwestern Crete

In northwestern Crete, the dark-colored and bedded crystalline limestones in the Phyllite-Quartzite nappe are often altered tectonically into large yellow beige and cavernous weathering lumps of unstratified brecciated coarse crystalline limestones.

The dark grey limestones of the Tripolitza nappe vary from mud to grainstones (often rich in *Nummulites* and very occasionally occurring together with *Alveolina*) but in some places they are slightly crystalline and in others they have been converted into black dolomites. The light to dark grey and occasionally violet colored thinly bedded dolostones along the coast and higher up the slope north of ancient Falasarna are interpreted to represent the local base of the Tripolitza carbonates. On the Rhodope peninsula of northwestern Crete, the local base of the Tripolitza unit includes Triassic volcanic rocks (the Ravdoucha beds) (*Papanikolaou and Vassilakis*, 2010). The Tripolitza unit on western Crete becomes a few hundred meters thick, but in some places on northwestern Crete, it is missing, and the Pindos unit directly overlies the Phyllite-Quartzite unit (*EAGME*, 2002) (Figure 3).

The true stratigraphic succession of the Pindos nappe in northwestern Crete is difficult to determine in the field due to intense folding and subsequent dismemberment along faults. Three main lithologies could be distinguished: 1) beige bedded limestones (mostly mudstones) with cherts; 2) varicolored alternation of

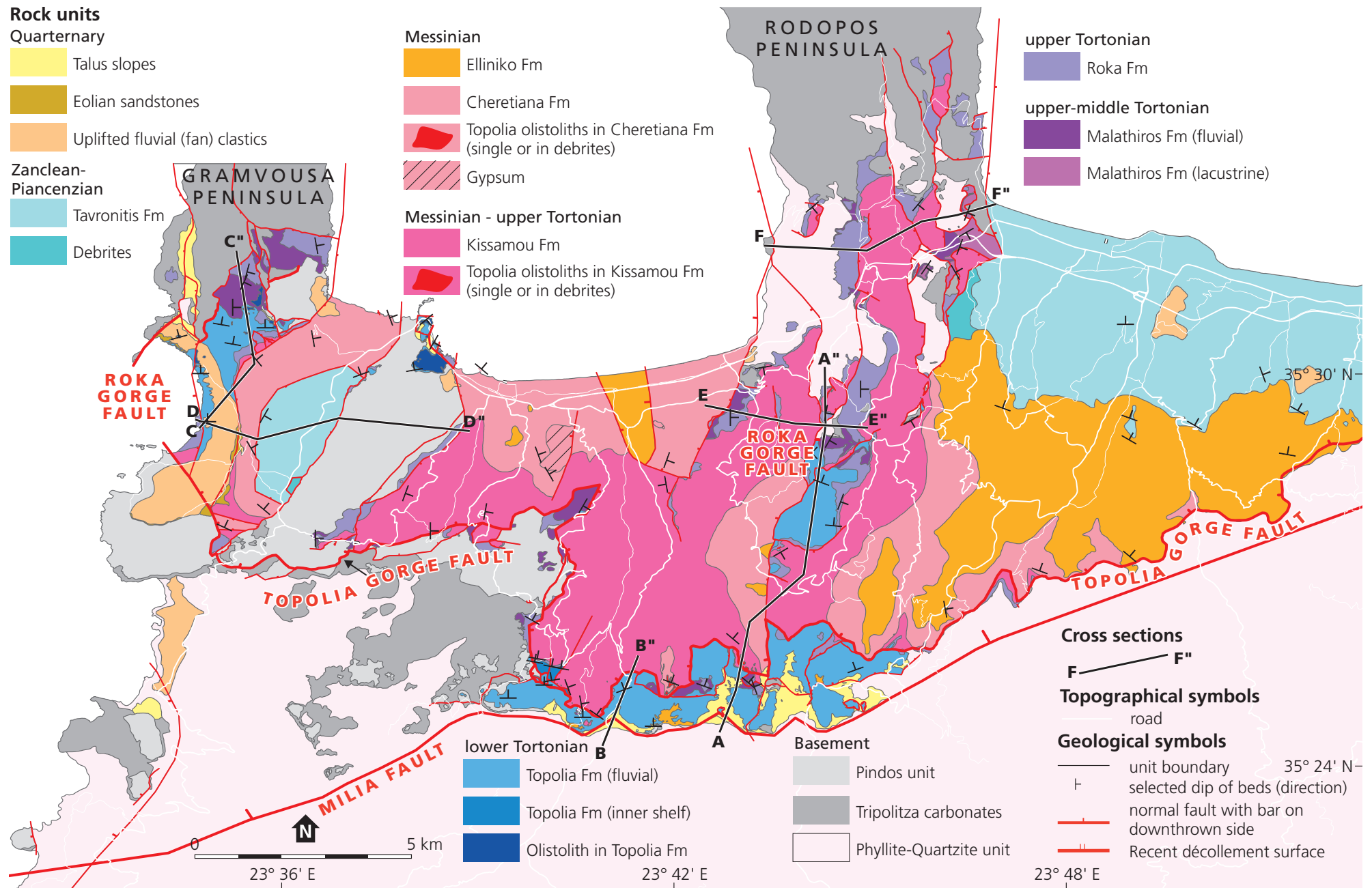


Figure 3 – New detailed geological map of northwestern Crete. For a detailed, larger version of this map, see Supplementary Information Item 1B.

shales, siltstones and limestones with chert (some being ooid grainstones); 3) alternation of greenish grey shales, greenish brown turbiditic sandstones and subordinate beige limestones of the flysch. In several places in northwestern Crete, the flysch is foliated with abundant calcite veins, and may include olistoliths of wackestones with *Assilina*, *Nummulites*, and miliolids. If we accept that *Assilina* is an Early-Middle Eocene foraminiferal species (*Al Menoufy and Boukhary, 2021*) then this type of flysch (foliated and with olistoliths) is Pindos flysch where the olistoliths must have been supplied from the original Tripolitza carbonate platform in the downgoing plate adjacent to the foredeep in the south.

The distribution of the three main lithologies mentioned above lack lateral continuity and stratigraphic coherence due to intense deformation. The three distinguished lithologies are therefore mapped as one single unit in Figure 3, viz. the Pindos unit. Because any occurrences of Tripolitza flysch are indistinguishable from the Pindos flysch, and the regional tectonostratigraphic coherence is often lost due to the strong dismemberment of the nappes, it is possible that some occurrences mapped as Pindos (flysch) belong to the Tripolitza nappe. If so, this does not influence our interpretation of the Neogene basin evolution or its relationship to the Cretan Detachment.

4.2 Neogene Lithostratigraphy, Depositional Conditions, and Ages

The Neogene sediments of northwestern Crete were first mapped, described, and formally subdivided, into five lithostratigraphic units, by *Freudenthal (1969)*. *Meulenkamp et al. (1977)* updated the lithostratigraphy for the western part of this area, including the same units under partly different names of which we adopt here only the Hellenikon Fm (current spelling Elliniko Fm). *Kopp and Richter (1983)* identified a deepest stratigraphic unit consisting of dolomite/limestone conglomerates (the Topolia Fm), which was previously assumed to belong to the pre-Neogene basement by *Freudenthal (1969)*. We have updated the Neogene lithostratigraphy in northwestern Crete and re-evaluated their chronostratigraphic position, depositional conditions, and ages (Figure 4).

The deep marine sediments in this study are dated by correlating presence/absence patterns and coiling of age diagnostic plankton foraminiferal species to the succession of fourteen Miocene age-calibrated biozones in *Zachariasse et al. (2011)* and *Zachariasse and Lourens (2021)*. Full discussions about the age of various Miocene formations are lengthy because based on a large, partially published, data base. Therefore, we have chosen to keep those discussions short in the text by referring to all details in SI Items 4, 5, 6 and 8. For the discussion on Pliocene ages we refer to the ten Pliocene age-calibrated biozones based on multi-source published biochronologic data in SI Item 3. Depth estimates of deep marine sediments are based on the presence of benthic foraminiferal depth markers using selected taxa

and their depth ranges given in *van Hinsbergen et al. (2005b)*.

4.2.1 Topolia Formation

Lithology: The Topolia Fm comprises a clastic series dominated by greyish to locally reddish conglomerates of exclusively Tripolitza and Pindos debris and few interbeds of greyish (pebbly) sandstones/siltstones and minor reddish colored fines (*Kopp and Richter, 1983; Seidel et al., 2007*). Conglomerates are strongly cemented between Kakopetros and Sirikari (Figure 5a, b) and at Roka (Figure 5c,d) where they form rugged mountain ridges (Figure 5e). They are more poorly cemented on Gramvousa (Figure 5f, g). Conglomerates are mostly dominated by Tripolitza debris, but they may alternate stratigraphically or geographically with Pindos-rich conglomerates. The mostly subangular debris is poorly sorted to unsorted and mostly clast supported. Bedding is often obscured by strong cementation, karstification, and travertine coatings especially in weathered cliff faces where a few erosion cavities show only glimpses of the actual series (Figure 5h, i). Average clast size measured at a scattered number of locations is 10-50 cm. Exotic blocks of Tripolitza carbonates, which we interpret as olistoliths, with sizes ranging from a few square meters to 0.3 km² occur at Trachilos (Figure 6a-b), southern Gramvousa and Roka Gorge. Mappable ones are shown in Figure 3.

A few small outcrops in the poorly exposed area north of the Keramariano Gorge (NE of Milia) (Figure 3) show sediments found nowhere else in northwestern Crete. They include grey clays and brownish sandstones with mollusks such as oyster, *Turritella*, Cerithiidae gastropods, Cardiidae bivalves, and once with solitary corals. Another outcrop off the road 1 km south from Ano Kalathenes exposes a thin layer of silty clays with foraminifers between poorly sorted conglomerates of upper nappes debris. All these sediments dip 40-50° to the north conform the dips in the Topolia conglomerates to the south and are overlain by more gently N-dipping marine sediments belonging to younger formations (Roka and Kissamou Fms, see below) (Figure 4). Therefore, we consider these marine sediments to belong to the uppermost part of the Topolia Fm, but we assigned them a distinct color on the geological map (Figure 3).

Thickness and reference sections: The Topolia Fm is well exposed in the Roka and Topolia Gorges, along the Falasarna coast and on the ridge above Falasarna on southern Gramvousa where we made most observations.

Kopp and Richter (1983) and *Seidel et al. (2007)* estimated thicknesses between 300m and 500m. We measured a minimum thickness of 200m between dips in top and basal part of the Topolia Fm in the northern Roka Gorge. Another minimum thickness of 250m is measured along the ridge above Falasarna. Therefore, our best estimate is 250-300m. The actual thickness of the marine uppermost part of the Topolia Fm is unknown.

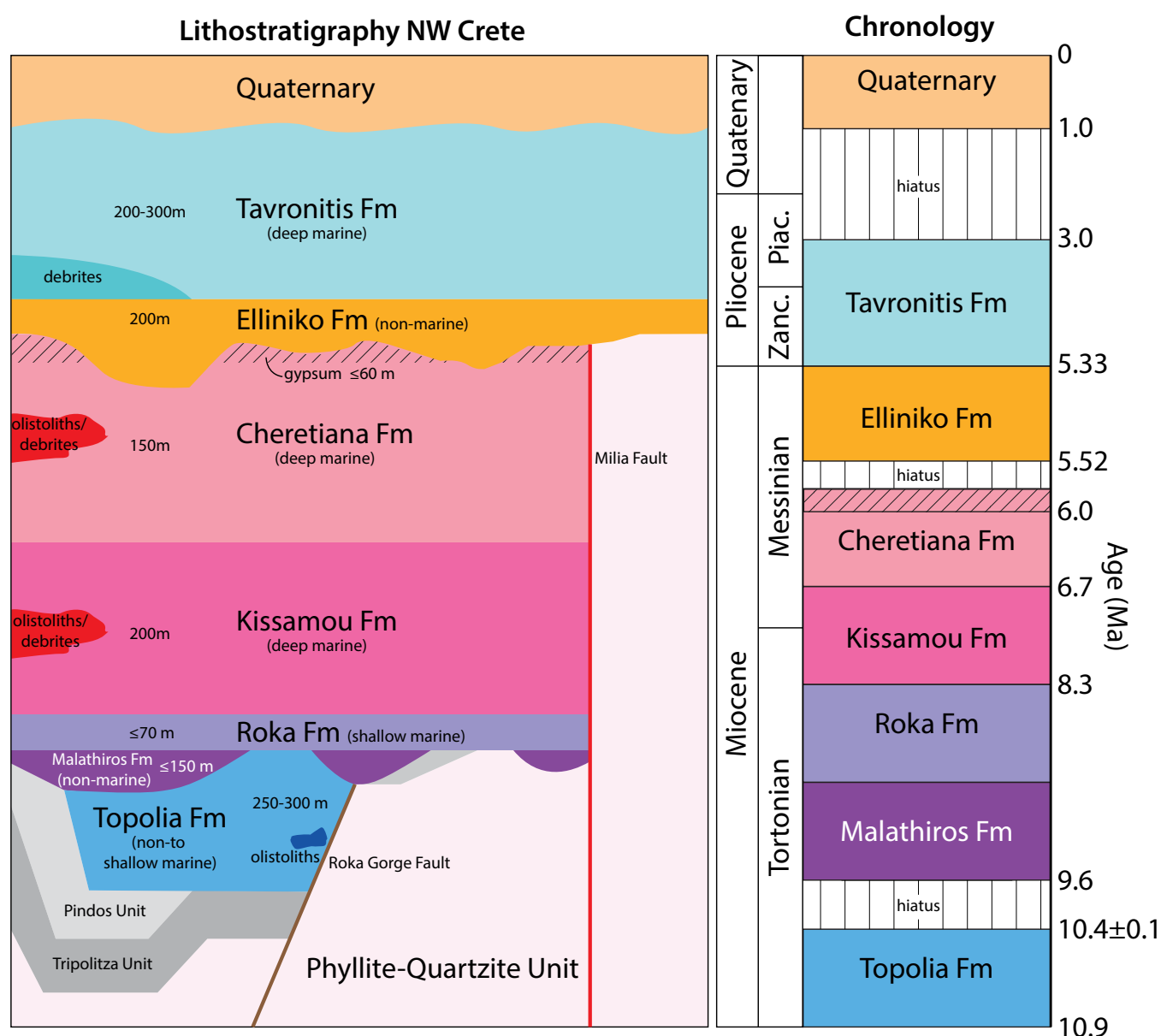


Figure 4 – Neogene lithostratigraphy for northwestern Crete along with chronostratigraphic position and depositional environment (**left**) and ages for lithostratigraphic units (**right**). Age for base Elliniko Fm is equated with age for base Lago Mare in *Krijgsman et al.* (2024).

Contacts with underlying basement units: The Topolia Fm stratigraphically overlies the Pindos unit except in the area of the Deliana Gorge (south of the village Deliana) where it overlies Tripolitza carbonates (Figure 3). Contacts are rarely exposed. At Sirikari, cemented Topolia conglomerates overlie a weathered surface of purple siltstones and grey limestones of the Pindos unit (Figure 6c). Other contacts are exposed on the west and east edges of the Deliana Gorge where Topolia conglomerates stratigraphically overlie a smooth and sub-horizontal surface of grey crystalline Tripolitza limestones. At the easternmost occurrence of the Topolia Fm, a few meters of smeared varicolored Pindos siliciclastics are sandwiched between Topolia conglomerates and Phyllite-Quartzite (see SI Item 1A). This contact is disturbed by recent landsliding but demonstrates that the Topolia Fm overlies the Pindos unit. *Kopp and Richter* (1983) reported (poor) stratigraphic contacts with Pindos and Tripolitza

basement from a few other locations but they are no longer exposed.

Geographic distribution: Occurrences of the Topolia Fm are limited to the western part of the studied area, viz. in the areas around Roka, southern Gramvousa, and Trachilos, and between Kakopetros and Sirikari with minor occurrences at Charchaliana and Episkopi (Figure 3). *Creutzburg* (1977) report Topolia conglomerates overlying HP-LT metamorphic rocks from the area south of Chania, to the east of our mapping area (Figure 2). We studied these occurrences at three different locations (at the villages Kondopoulo and Kambi, and N-entrance Theriso Gorge; Figure 2) and find that they are not Topolia conglomerates but brecciated (and sometimes karstified and cavernous), mostly grey, non-bedded, coarse crystalline Trypali limestones.



Figure 5 – Cemented Topolia conglomerates at (A) Kakopetros; (B) in Topolia Gorge and (C-D) Roka Gorge; (E) Topolia conglomerates viewed from a vantage point slightly NW of Mouri to the S; less cemented Topolia conglomerates at (F) southern Gramvousa and (G) Falasarna coast; travertine coated Topolia conglomerates in (H) Topolia Gorge and (I) Roka Gorge.



Figure 6 – (A-B) Tripolitza olistolith in Topolia conglomerates at Trachilos; (C) stratigraphic contact between Pindos rocks and Topolia conglomerates ~0.8 km NW of Sirikari; (D) Topolia conglomerates with imbricated pebbles in the Roka Gorge (drainage to the S (right)); (E) stack fluvial conglomerates in Malathiros Fm; dominant fine-grained Malathiros Fm (F) along National Road (NR) 1 km W of turnoff to Kalidonia and (G) above N-entrance Roka Gorge; (H) pebbly sandstones and fines of Malathiros Fm filling paleorelief in Topolia conglomerates along new road to Vlatos ~1 km ENE of Milia; (I) bioturbated sandstone in Roka Fm 2 km NNW of Azogiras.

Depositional environment: The clast-supported, poorly sorted to unsorted conglomerates of mostly subangular Pindos and Tripolitza debris point to a proximal fluvial environment, in agreement with *Kopp and Richter* (1983) and *Seidel et al.* (2007). The fluvial deposits may have been part of a complex of alluvial fans with gravel bars in braided streams while finer grained sediments accumulated in flood basins or under drier climatic conditions, but we have too little paleocurrent data to show any fan-shaped pattern(s). Imbricated pebbles at four different locations (Topolia Gorge *van Hinsbergen et al.* 2008, Roka Gorge [Figure 6d] and southern Gramvousa) indicate S-directed drainage and three roughly N-S running channel fills on southern Gramvousa allow for southbound transport. Samples from the marine uppermost part contain shallow epifaunal benthic foraminiferal species (*Ammonia parkinsoniana*, *Textularia agglutinans*, miliolids and few *Gyroidina umbonata*) and rare planktonic foraminifers, typical for a muddy inner shelf environment in agreement with the macrofossils.

Age: The Topolia Fm itself cannot be dated, and ages in the literature therefore vary considerably. *Seidel et al.* (2007) related basin formation and deposition of the Topolia Fm to a 20-15 Ma old phase of intense brittle deformation of the Phyllite-Quartzite which they assume affected also the upper nappes by forming half-grabens. *van Hinsbergen and Meulenkaamp* (2006) synchronized the formation of the large E-W trending basin accommodating the westward-draining Males-Viannos River system on eastern and central Crete with that of a deep marine basin on Gavdos in the late Serravallian (11.6-12.7 Ma, *Hilgen et al.* 2012). The fluvial Males-Viannos Basin extended from the eastern tip of Crete to ~55 km east of the studied area (a total distance of 220 km) (*Zachariasse et al.*, 2011). Sediments of this Males-Viannos River system have also been reported from southwestern Crete where they overlie Topolia conglomerates making the Topolia Fm older than the late Serravallian Males-Viannos River sediments (*van Hinsbergen et al.*, 2008). A similar stratigraphic position for both units has been published in *van Hinsbergen and Meulenkaamp* (2006) and *Zachariasse et al.* (2011) but conflicts with the interfingering of both units in eastern Crete (*Fortuin*, 1977; *Peters*, 1985). In this study, the Topolia Fm is considered to be time-equivalent with the Males-Viannos River sediments and by equating the formation of both fluvial basins on Crete with marine subsidence on Gavdos, the age for base Topolia Fm can be determined. Basin formation on Gavdos (located on the outer Cretan margin, see Figure 1) is characterized by rapid subsidence to 500m below sea level. Calibration of selected planktonic foraminiferal species patterns in the deep marine sediments of Gavdos to astronomically-tuned patterns in the Tortonian of Sicily (Italy) yields an updated age of 10.9 Ma for basin formation on Gavdos and therefore for base Topolia Fm (see SI Item 4 for a full account of the Neogene stratigraphy and vertical motion history of Gavdos). The Topolia Fm is unconformably overlain by undatable non-marine sediments of the Malathiros Fm. However,

Topolia equivalents in southwestern Crete are overlain by deep marine sediments that have an age range of 10.4 ± 0.1 Ma using plankton foraminiferal biostratigraphy (work in progress). An age range of 10.9 to 10.4 ± 0.1 Ma for the Topolia Fm indicates an early Tortonian age for this unit (Figure 4).

4.2.2 Malathiros Formation (new name)

In the original definition, the Roka Fm comprises non-marine and shallow marine deposits (*Freudenthal*, 1969). In this study, the non-marine deposits are separately mapped and included in the new Malathiros Fm.

Lithology: The non-marine sediments of the Malathiros Fm form locally thick sequences and appear to fill depressions in both the Topolia Fm and Alpine basement (Figure 3). They are made up of variably cemented conglomerates, (pebbly) sandstones and reddish to orange yellowish fines. Conglomerates can be up to 30 m thick (Figure 6e) and consist mostly of poorly sorted, clast supported, subangular to rounded debris originating from the upper nappes as well as from the Phyllite-Quartzite unit. Sometimes Phyllite-Quartzite debris is dominant. Most exposures are dominated by fine grained sediments with thin conglomerates (Figure 6f-g). The few limestones are pisolithic.

The N-dipping Topolia conglomerates on southern Gramvousa are overlain by E-dipping conglomerates and fines. Neither in composition and cementation, nor in rounding, sorting and arrangement of the pebbles do these E-dipping conglomerates differ from the N-dipping ones except that the E-dipping series has more interbedded cemented sandstones (some with planar laminations) and fines and is conformably overlain by the shallow marine Roka Fm (see below). Based on the angular unconformity between the N and E-dipping series, we attribute the non-marine E-dipping series to the Malathiros Fm (Figure 3).

In the area around and southwest of Kolymbari, fluvial conglomerates and minor reddish fines are overlain by grey (silty) clays with thinly bedded, poorly cemented sandstones and beige marls with small brownish burrows. Dominant candonid ostracods in several samples are characteristic of an (oligohaline) lacustrine environment (pers. comm. Marius Stoica, August 2023). The many small sized burrows in the beige marls are remarkable but not unusual in lacustrine sediments (*Uchman and Álvaro*, 2000; *Uchman et al.*, 2007). These lacustrine sediments are included in the Malathiros Fm but have been assigned a distinct color on the geological map (see Figure 3).

Thickness and type locality: *Freudenthal* (1969) mentions a maximum thickness of 60m for the non-marine part of his Roka Fm (here assigned to the Malathiros Fm) but the fluvial sequences to the north of Roka and south of Malathiros are up to some 150m thick. The thickness of the lacustrine part is at least 30m. The type locality is the series of outcrops along the road

Malathiros-Sasalos, ~ 2 km south of Malathiros (Figure 3).

Contacts with underlying units: The formation unconformably overlies the often karstified surface of the Topolia Fm (Figure 6h) as well as the upper nappes and Phyllite-Quartzite (Figure 3).

Depositional environment: The poorly sorted and clast supported conglomerates point to deposition from braided streams with the sandstones and reddish fines being deposited in flood basins and/or during drier climatic conditions. Unsorted conglomerates of more angular debris probably represent flash flood deposits. The same composition of the Malathiros and Topolia conglomerates in the area above Falasarna indicates that the debris in the first mentioned conglomerates is reworked from the second. The lacustrine sediments with dominant candonid ostracods are exposed over an area of ~ 5 km² (see Figure 3) indicating the existence of a fresh to oligohaline lake of at least the same size at the time of deposition of the Malathiros Fm.

Age: The base of the shallow marine Roka Fm, which conformably overlies the Malathiros Fm, is undatable but probably not much older than the top Roka Fm at 8.3 Ma (see discussion under Roka Fm). The age for base of the Malathiros Fm is set at ~ 9.6 Ma based on age constraints derived elsewhere from Crete, viz. mammal faunas, magnetic polarity data, and lithostratigraphic correlations (see SI Item 5 for all details). According to these age constraints, the Malathiros Fm belongs to the middle – upper Tortonian (Figure 4).

4.2.3 Roka Formation (emended)

In the original definition, the Roka Fm comprises non-marine and shallow marine deposits (*Freudenthal*, 1969). In this study, the Roka Fm is limited to the shallow marine sediments.

Lithology: The emended Roka Fm consists of variably cemented (pebbly) bioturbated sandstones (Figure 6i) (sometimes conglomeratic) and bioclastic limestones with typical shallow marine fossils such as *Heterostegina*, *Clypeaster*, *Porites* and *Tarbellastrea* corals, oysters, *Lithothamnion*, and *Pecten latissima* (*Freudenthal*, 1969). Worth a mention is the presence of the eurybathymetric brachiopod species *Megerlia truncata* in the shallow marine sediments to the N of Kaliviana (determination by Alfred Dulai in March 2023) and described earlier from the Tortonian of Crete by *Georgiades-Dikeoulia* (1974).

Thickness: The Roka Fm is well exposed and 20-30 m thick in the gorge immediately west of Kamara and north of Kolymbari. On southern Gramvousa, the Roka Fm is also well exposed and reaches its maximum (constructed) thickness of 75m.

Contacts with underlying units: Contacts with the underlying Malathiros Fm are conformable and unconformable where the Malathiros Fm overlies the Topolia Fm and all deeper basement units (Figure 3).

Depositional environment: The fossil association points to warm, inner shelf conditions with sandy to hard substrates (< 100m)

Age: The top of the Roka Fm is dated by the age for base of the overlying deep marine Kissamou Fm. The basal part of the Kissamou Fm in three of the *Freudenthal* (1969) sections (SI Item 6.1-6.3) all belong to upper plankton foraminiferal biozone 8 (for definition and age, see Appendix S2 and Table S1 in *Zachariasse et al.* 2011 and Appendix 6 in *Zachariasse and Lourens* 2021). Age for top biozone 8 is 7.9 Ma, so the base Kissamou Fm in these sections is > 7.9 Ma (the true base is not exposed or faulted - for details of the 3 sections, see SI Item 6.1-6.3).

The magnetobiostratigraphy of section Skouloudiana (*Krijgsman et al.*, 1994; *Langereis et al.*, 1984) (SI Item 6.4; for location of the village, see SI Item 1B) shows that the sampled base probably belongs to Chron 4r.1n (age bottom is 8.3 Ma in *Hilgen et al.* 2012). The true base Kissamou Fm should be older since the Roka Fm is not exposed in or near the section (see Figure 3).

In this study we set the age for base Kissamou Fm at 8.3 Ma consistent with the age for the well-exposed base of the equivalent deep marine Apostoli Fm near the nominate village in the Rethymnon area (see SI Item 6.5). The base of the Roka Fm is undatable but probably not much older given the small thickness of the unit and the rapid subsidence to deep marine conditions at the top.

4.2.4 Kissamou Formation

Lithology: This unit is made up of deep marine greyish marls with minor mostly medium cemented sandstones (*Freudenthal*, 1969). Sedimentary cycles are manifest by more and less cemented marls (Figure 7a-b) and only rarely by homogeneous-laminated marl couplets. In several areas along the current southern basin margin, marls are siltier and sandstones more abundant, and often coarser grained to pebbly and even conglomeratic (separately mapped as Koukounares Fm by *Freudenthal* 1969 but here included in the Kissamou Fm). Sandstones in these basin margin areas often show planar laminations and occasionally normal grading and scour marks, and often include bioclasts of *Lithothamnion*, *Heterostegina* and bivalves and may form stacks up to 6m thick. They are classified as turbiditic sandstones just as their more distal and finer equivalents to the north. Boulders/blocks of Topolia conglomerates (up to 10m in size), singly or in debrites of up to 10m thick, occur in the deep marine marls along the current southern basin margin (Figure 3). Some of these debrites are coarse calcarenitic with (apart from Topolia blocks) slumped marls and basement debris. Few debrites consist of algal balls only. A 6 m thick debrite along the road Voukolies-Kandanos, ~ 0.7 km south of Dromonero (dated as late Tortonian) is made up of unsorted angular Phyllite-Quartzite debris and turbiditic sandstones. Northwards slumped packages of strata are exposed along the new road to Vlatos above the western edge of the Topolia Gorge (Figure 7c-d).

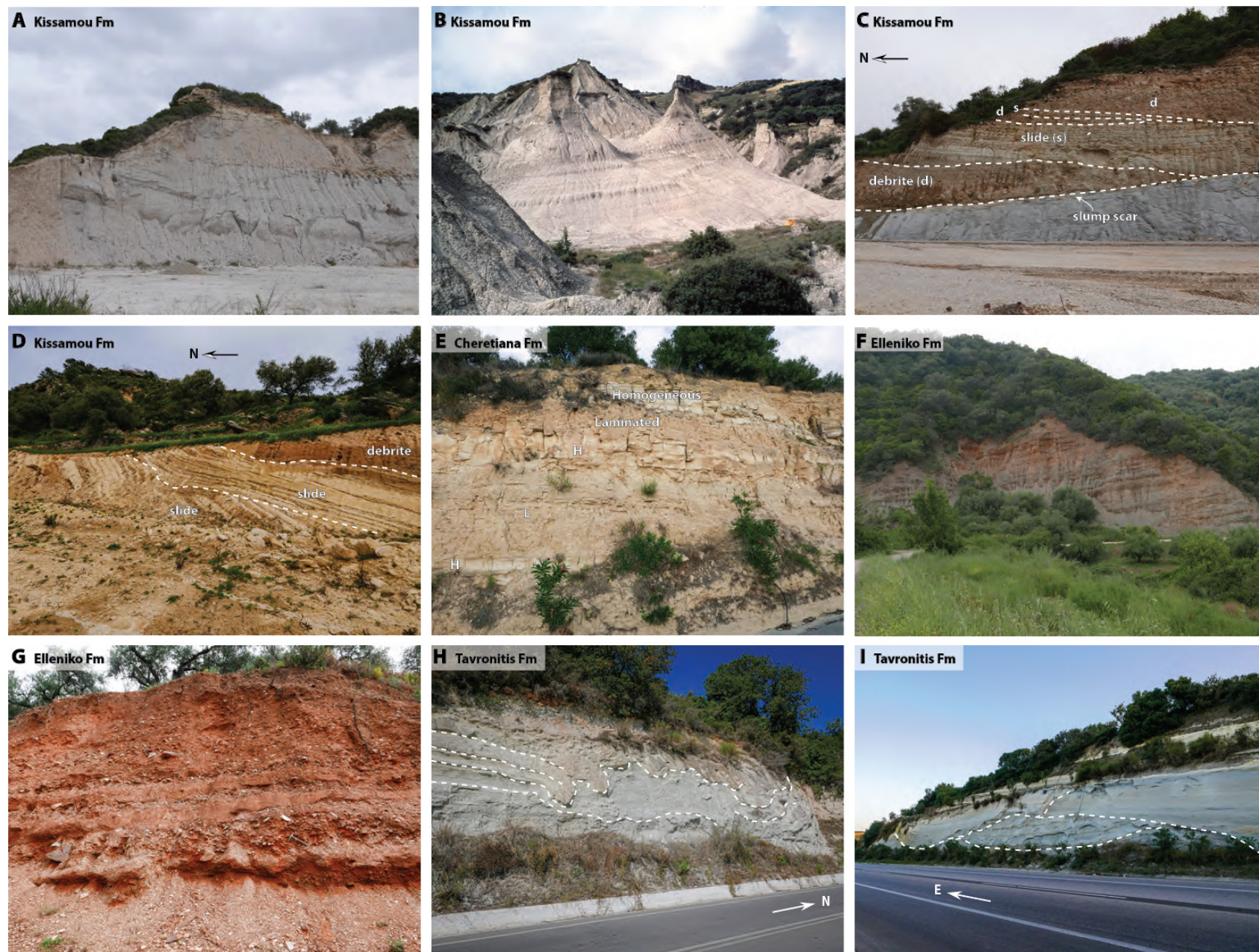


Figure 7 – Deep marine marls with vague cyclicity in Kissamou Fm (**A**) along NR 0.7 km E of turnoff to Kalidonia and (**B**) type section at Potamida; (**C-D**) slides and debrites in Kissamou Fm along new road to Vlatos 1 km SW of Topolia; (**E**) precession-driven cycles of homogenous and laminated marls in top type section Cheretiana Fm; (**F**) bedded fluvial conglomerates in Elleniko Fm 1.6 km WSW of Limni; (**G**) poorly bedded conglomerates of angular Phyllite-Quartzite debris and fines in Elleniko Fm 0.8 km NE of Psathogiannos; (**H**) slumped marls in Tavronitis Fm 1.5 km NNE of Neratzia; (**I**) 10 m-scale filled slump scars in Tavronitis Fm along NR 4.6 km E from turnoff Tavronitis.

Thickness and reference sections: Thickness is estimated at ~ 200 m (*Freudenthal, 1969*). The badlands at the village of Potamida (Figure 7b) provide one of the best exposures in the distal facies (assigned as type section by *Freudenthal 1969*). The more proximal sandy facies is best exposed along the new road to Vlatos above the Topolia Gorge.

Contacts with underlying units: The Kissamou Fm conformably overlies the Roka Fm but contacts are rarely exposed.

Depositional environment: Benthic foraminiferal depth markers in some 20 samples from all over the area point to a deep marine (500-750 m) environment (for details, see SI Item 7).

Age: Base of the Kissamou Fm is dated at 8.3 Ma (see under Roka Fm) and the top at 6.7 Ma (see under Cheretiana Fm) indicating that this unit belongs to the upper Tortonian-Messinian (Figure 4).

4.2.5 Cheretiana Formation

Lithology: Characteristic for this unit is the alternation of beige (more calcareous) homogeneous and brownish laminated marls (sapropels) with cemented sandstones/calcarenes (*Freudenthal, 1969*) (Figure 7e). Laminated marls are up to 6 m thick and often contain siliceous sponge spicules. Sandstones often show planar laminations and marl flakes/pebbles and less often normal grading, scour marks and bedding parallel burrows. Coarse sandstones may have bioclasts (often *Lithothamnion*) or are made up of bio-skeletal material. Here, too, the sandstones are turbiditic and form stacks up to 2m thick. Debrites with Topolia boulders/blocks, whether mixed with Phyllite-Quartzite debris or coarse calcarenites and slumped marls occur along the current southern basin margin west of Zimbragos (Figure 3). Eastwards, debrites (many and up to 8 m thick) consist of angular Phyllite-Quartzite debris (not mapped separately). Gypsum occurrences at Dafni and north of Cheretiana belong to the uppermost part of the Cheretiana Fm and are mapped as a separate subunit (Figure 3). However, the transition from marls to gypsum is nowhere exposed. The gypsum varies from laminated (balatino gypsum) to selenitic and alternates with rarely exposed calcareous marls. This evaporite unit accumulated during the early (marine) stage of the Messinian Salinity Crisis (MSC).

Thickness and reference sections: The 100 m mentioned by *Freudenthal (1969)* is at the low side because the constructed thickness in the scarp south of Platanos is already ~ 150 m while the evaporite unit is at least 60-70 m thick. The distal facies is well exposed at the village of Cheretiana (type section, Figure 7e) while the proximal facies with Phyllite-Quartzite-rich debrites are, for example, reasonably well exposed along road to Fotokadon from the main road Voukolies-Kakopetros and the serpentine bends in the road Zimbragos-Kakopetros, 0.6 km south of Zimbragos.

Contacts with underlying units: The Cheretiana Fm conformably overlies Kissamou Fm. Both units, however, are poorly exposed in the heavily cultivated hillside landscape and the boundary on the map is therefore a rough approximation (Figure 3).

Depositional environment: Benthic foraminiferal depth markers are absent in our 17 samples. *van Hinsbergen and Meulen Kamp (2006)* reported a depositional depth of 460 ± 300 m for the type section using percentages of planktonic foraminifers and the paleodepth equation of *van der Zwaan et al. (1990)*. Rare occurrences of benthic foraminifers belonging to *Oridorsalis* in new samples from the upper part of type section point at a paleodepth of > 300 m (SI Item 8) in agreement with the presence of sapropels because their formation in the open eastern Mediterranean (EMED) occurred below 300 m for both Quaternary (*Rohling and Gieskes, 1989*) and Messinian ones (*Zachariasse et al., 2021*).

Age: The ~ 6.7 Ma for base is based on the assumption that the change from the more clayey Kissamou Fm to the more calcareous sediments of the Cheretiana Fm is time-equivalent with a similar change in central Crete dated at 6.7 Ma (*Zachariasse et al., 2021*). The gypsum is younger than the age of 6.00 Ma for onset of the MSC in the EMED (*Zachariasse and Lourens, 2021*) and older than today's consensus age of 5.60 Ma for the drawdown of the Mediterranean and associated erosion in marginal basins (*Andreotto et al., 2021; Roveri et al., 2014*) and so, the Cheretiana Fm belongs to the Messinian (Figure 4).

Note to the claimed 6.05 Ma old hominin-like footprints at (1km NW of) Trachilos: Recently, *Zachariasse and Lourens (2022)* challenged the published age of 6.05 Ma for the Trachilos site (*Kirscher et al., 2021*) from which *Gierliński et al. (2017)* described hominin-like footprints and argued that the actual age for this site is ~ 3 Ma. Follow-up research by WJZ has shown that the turbiditic bioclast-rich sandstones with greyish brown marl flakes of the Trachilos footprint site also occur in a 2.5 m high cliff on a small island close off the coast of Kissamos (see Figure 3). Unlike the Trachilos site samples, the sample from (partly laminated) greyish brown marl flakes in the lowermost coarse skeletal sandstone on this small island does have age diagnostic planktonic foraminifers (for a full account of the foraminiferal faunas, see SI Item 8). Their age range of 6.34-6.00 Ma makes these sediments as old as the uppermost part of the type section of the Cheretiana Fm (for a detailed argumentation of age and depositional depth for top Cheretianan type section, see SI Item 8). Moreover, this age range fits the Chron 3A.1n assignment for the Trachilos site by *Kirscher et al. (2021)*. So, both the sediments of the Trachilos footprint site and the small island belong to the deep marine Cheretiana Fm. The ostracods from this small island sample and another sample from the Trachilos footprint site are Late Miocene shallow marine ostracods (personal comm. Marius Stoica, August 2023) and must be ex situ just as the shallow marine benthic foraminifers. Even at

this age, no hominins could have walked around at the Trachilos site given the deep marine environment for the Cheretiana Fm (see above).

Assigning the sediments of the Trachilos footprint site to the Cheretiana Fm is further confirmed by homogeneous-laminated marls with bioclast-rich sandstones ~ 1.7 km SSW of the Trachilos site (at Youphoria Villas) interpreted previously as Pliocene. Age diagnostic planktonic foraminifers in two samples, however, point to an age range of 6.7-6.4 Ma meaning that the Pliocene hanging wall block closes south of the Trachilos footprint site (Figure 3).

4.2.6 Elliniko Formation

Lithology: This unit is dominated by thick series of reddish conglomerates, (pebbly) sandstones and fines (Meulenkamp *et al.*, 1977). Conglomerates are bedded and consist of poorly sorted to unsorted, clast to matrix supported Phyllite-Quartzite debris (Figure 7f-g). Debris varies from well-rounded to angular with clast sizes of up to 4 m³. Some intervals within this reddish clastic series are made up of messy crystalline, cavernous whitish limestones (N of Psathogiannos) and may alternate with chalks with gyrogonites of charaphytes and brackish ostracods (Marius Stoica, pers. com. August 2023), (lignitic) clays, and sands (e.g. ~0.7 km southeast of Moulameriana). In the area of Neratzia, the uppermost part of this unit consists of azoic clays, silts and fine sandstones (sometimes varved) with few fine conglomerates. Greyish sands and beige varved clays are also exposed south of the village of Vouves. Several isolated occurrences of reddish poorly cemented, unsorted to poorly sorted and mostly clast supported conglomerates of mostly angular Phyllite-Quartzite debris and fines in between Kakopetros and Sirikari are assigned to the Elliniko Fm (Figure 3).

Thickness and reference sections: The unit is up to 200 m thick (Meulenkamp *et al.*, 1977). Series of good outcrops are to be seen along the roads from Voukolies to Kakopetros and from Limni to Manoliopoulo.

Contacts with underlying units: The contact with the underlying Cheretiana Fm is an erosional unconformity that equates with the Messinian Erosional Surface in marginal basins [MES] in Roveri *et al.* (2014). The Elliniko Fm overlies the erosional surface of the Topolia Fm and the Phyllite-Quartzite to the south of the current basin margin (Figure 3).

Depositional environment: The unsorted to poorly sorted, clast to matrix supported conglomerates are interpreted as braided stream and flash flood deposits where interbedded fines were deposited in flood basins or under drier climatic conditions. Sedimentation may have taken place on large alluvial fans. The fine grained and partly varved sediments and minor chalks, with brackish ostracods in one sample are fresh to brackish lacustrine deposits.

Age: The erosional unconformity that truncates the Cheretiana Fm is dated at 5.60 Ma being the age for

the MES (Roveri *et al.*, 2014). Deposition resumed at 5.52 Ma being the age for the base of Elliniko-type of sediments elsewhere in the Mediterranean (commonly termed Lago Mare sediments) (Krijgsman *et al.*, 2024). The deep marine sediments overlying the Elliniko Fm belong to the Pliocene whose base has been dated at 5.33 Ma (Lourens *et al.*, 2004). The Elliniko Fm thus belongs to the uppermost Messinian (also referred to as Lago Mare stage) (Figure 4).

4.2.7 Tavronitis Formation

Lithology: In the area of Platanos, this unit comprises whitish beige, often bioturbated, homogeneous marls and poorly cemented to uncemented, coarse (pebbly) calcarenites and finer brownish sandstones, and minor poorly sorted, clast supported conglomerates. In some outcrops, homogeneous marls alternate with brownish laminated marls (sapropels), one rich in siliceous sponge spicules. Calcarenites/sandstones are often bioturbated and irregularly bedded with rare planar laminations. They may form stacks of up to 3 m thick. Coarse calcarenites are mostly rich in bioclasts, especially bivalves, but also the benthic foraminifer *Amphistegina*, bryozoans, and echinoid spines are characteristic constituents.

In the area east of Spilia, a more complete series is exposed with a lower part being composed of only beige whitish homogeneous marls (Loutraki) or with few interbeds of Phyllite-Quartzite-rich conglomerates and uncemented brownish sandstones (Neratzia) or with many poorly cemented brownish sandstones (Elliniko) (Figure 3). The beige whitish homogeneous (Trubi) marls with vague cyclicity change upwards into more greyish marls with incidental sapropels and slumps (Figure 7h), slump-scars (Figure 7i), and poorly cemented to uncemented sandstones (with stacks up to 5m). Both the greyish marls and sandstones often have large burrows (up to 10 cm long). Sandstones may show normal grading, planar lamination, and convex-up geometry, typically of a turbiditic origin. Few flute casts indicate transport from the SE. Debris flows of coarse sand, marl balls, Phyllite-Quartzite debris, some with bivalves, occur throughout the unit but are more frequent in the exposed uppermost part. The Tavronitis Fm east of Spilia thus shows an upward coarsening trend towards more sandstones and debrites as well as slumps. Biostratigraphy shows that the coarser upper part in the area east of Spilia is time-equivalent with the Travonitis Fm in the Platanos area (see Item 9). It is in this part of the Tavronitis Fm that sapropels occur. Sandstones in the Platanos area are generally coarser and richer in bioclasts, and conglomerates more frequent suggestive of a position closer to the basin margin than the Tavronitis Fm in the area east of Spilia.

Occurrences of unstratified rubble and slides of whitish, partly cavernous/brecciated crystalline limestones (partly bedded) with or without calcareous and partly laminated marls at Spilia and Platanos are mapped as a separate subunit (Figure 3). They are interpreted as debrites with lithologies derived from a

younger part of the Cheretiana Fm. Similar deposits are widespread on Crete and often mixed with Lower Pliocene marls (Zachariasse *et al.*, 2008).

Thickness and reference sections: A thickness of 80 m is mentioned in *Freudenthal* (1969) but that is certainly an underestimate. The complete series is relatively well exposed along both banks of the Tavronitis River. Clear-cut faults cannot be seen but are probably present given the strong faulting of the studied area. The constructed thickness of ~400m without faults is therefore probably an overestimate. The actual stratigraphic thickness is probably closer to 200-300m.

Contacts with underlying units: The formation stratigraphically overlies the Elliniko Fm. Contacts are well exposed in the scarp directly north of Neratzia (see SI Item 2). The contacts to the south of Elliniko and Kaloudiana, described by *Spaak* (1982), are no longer exposed (at Kaloudiana there is now no Pliocene visible at all).

Depositional environment: Benthic foraminiferal depth markers in some 15 samples from the areas Platanos and east of Spilia point to a deep marine (500-750 m) environment (for details, see SI Item 7).

Age: A Zanclean-Piacenzian age for the Tavronitis Fm was concluded by *Frydas and Keupp* (1996) based on calcareous nannofossil biostratigraphy. Age diagnostic planktonic foraminiferal species in 15 samples from the areas Platanos and east of Spilia indicate that the Tavronitis Fm extends from Zanclean biozone 1 up to midway Piacenzian biozone 8 with an age range of 5.33 to ~3.00 Ma (SI Item 9; for definition and ages of Pliocene biozones, see SI Item 3).

4.2.8 Quaternary Deposits

Quaternary deposits are divided in three different units. The oldest unit (Figure 3) consists of uplifted (sub)horizontally bedded reddish brown fines and poorly to unsorted, clast-supported fluvial (fan) conglomerates up to 50m thick in the Sfinari area. They form distinct terraces in the landscape (Figures 9f,g). The debris in the conglomerates is of local provenance being angular Phyllite-Quartzite debris at Sfinari, subangular-subrounded upper nappes debris in the southern Falasarna coastal area to subangular Pindos debris immediately west of Kissamos. The (sub)rounded quartzites NW of Loutraki are probably reworked from the Elliniko Fm. The uplifted fluvial conglomerates in the southern Falasarna coastal area seem to overlie a few meters of horizontally bedded cemented coarse (pebbly) sandstones with bioclasts as exposed in three different locations elsewhere in the Falasarna coastal area (see SI Item 2). A younger unit (Figure 3) is made up of variably dipping laminated skeletal (no micrite) sandstones. These beige to sometimes reddish and once mottled cross-bedded sandstones are up to 15 m thick. These sandstones are interpreted as eolian and probably deposited during the last glacial period. The youngest unit (Figure 3) consists of talus slopes that are

especially impressive in the gorges through the Topolia conglomerates.

4.3 The Main Fault Systems

Figures 3 and 8 together provide an overview of the main faults and their relationship with the Neogene stratigraphy of northwestern Crete (for locations of the cross-sections, see Figure 3). A glimpse of the oldest fault system is shown in the northern part of Figure 8A where a steeply S-dipping Roka Gorge Fault separates Topolia Fm in the Roka Gorge from Phyllite-Quartzite to the north (Figure 9a). The fault is covered by the Malathiros Fm and that same unit also unconformably overlies the Topolia Fm (Figure 9a). Thus, this (northern) part of cross-section in Figure 8A reflects a history of faulting that pre-dates the Malathiros Fm and post-dates the deposition of the Topolia Fm where the latter unit pre-dates the exposure of the Phyllite-Quartzite unit. The northern bounding fault of the Topolia Fm farther West is mapped in the southern Gramvousa peninsula where it separates the Topolia Fm from Tripolitza basement to the north (Figure 3). This fault, which we interpret as the western equivalent of the Roka Gorge Fault is partly covered by the Malathiros Fm and partly re-used by post-Roka Fm faults (Figures 3; 8C and 9E).

We infer that the northern bounding fault of the Topolia Fm on southern Gramvousa steps southward along a N-S fault to continue further eastwards into the Roka Gorge (Figure 10). East of the line Episkopi-Zimbragos, the Topolia Fm is absent (Figure 3). Here, the Roka Fm overlies the Phyllite-Quartzite unit in the southern part (Figure 3). This configuration suggests another southbound step in the northern bounding fault of the Topolia Fm. The absence of the Topolia Fm to the south of the Milia Fault should be the result of erosion on the later formed Phyllite-Quartzite horst (Figure 3).

The two, kms-long southbound steps in the northern bounding fault of the Topolia Fm are interpreted as transfer faults (Figure 10). In this configuration, southbound drainage is expected to be accompanied by westward drainage in the areas adjacent to the breached relay ramps, but finding such directions is an illusion given the few paleoflow directions that can be measured.

Most (48 out of 55) dips in the Topolia Fm are to the north (except 7 in the Trachilos area) (see stereoplots in Figure 10) and have average dip angles of 43° (n=13) in the Roka Gorge, 25° (n=11) on southern Gramvousa and 45° (n=13) in the southern part (Figure 9b; SI Item 2). Constructed stratigraphic thicknesses assuming no faults would yield 1200, 770, and 1000 m for the Topolia Fm in the Roka Gorge (Figure 8A) and in the cross-sections of Figures 8B-C, which is unrealistically large. We therefore presume that the Topolia Fm in these areas is distributed over north tilted blocks bounded by normal faults that root in a common S-dipping décollement whose surface outcrop must be located north of the series tilted blocks in Figure 8A (dashed in black). Tilting of the Topolia Fm pre-dates the Malathiros Fm. It is important to add here that the inferred normal faults in the Topolia Fm of Figure 8 are invisible in the field due to the uniform

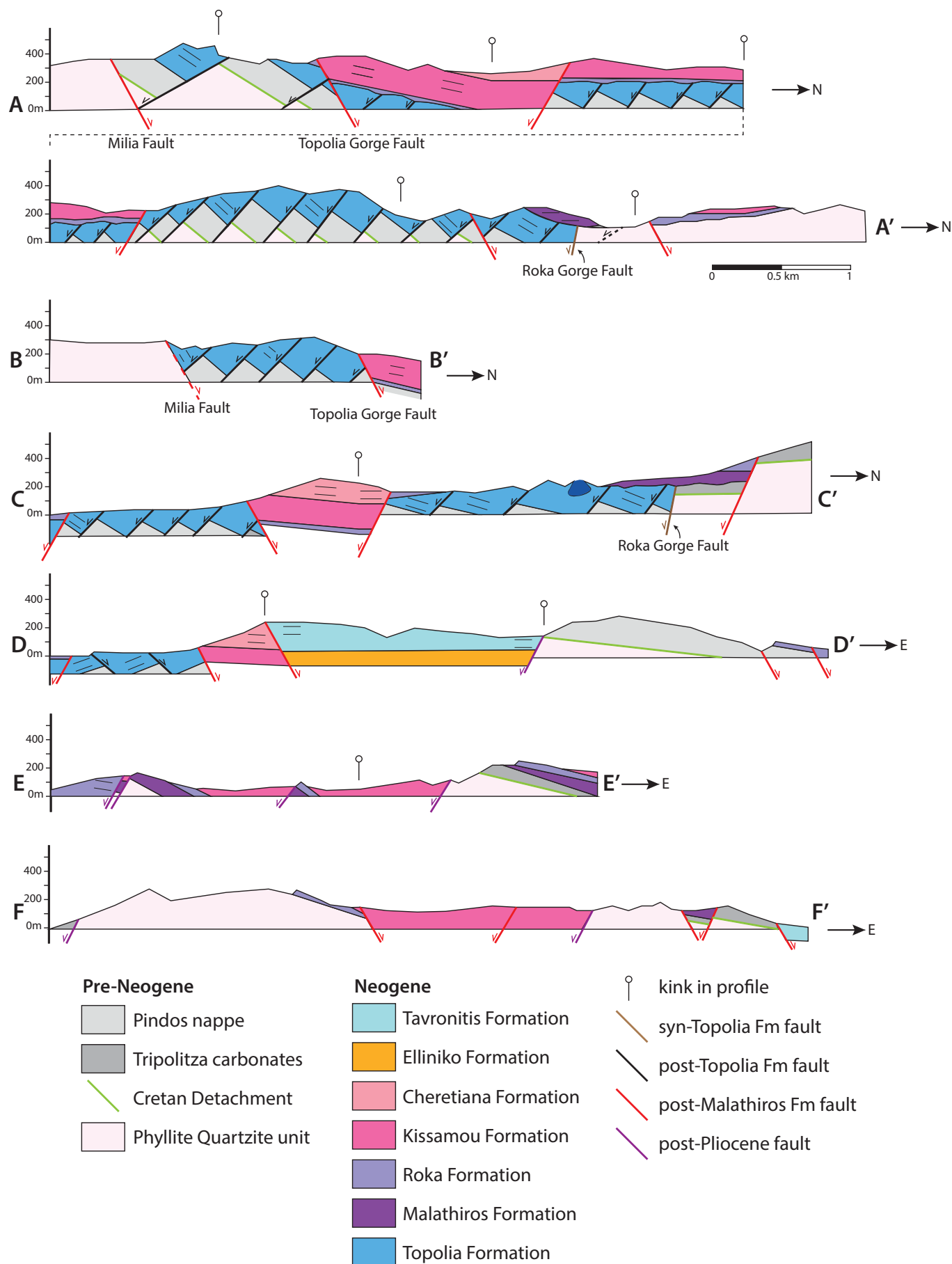


Figure 8 – Geological cross sections. Nearest dips were used to calculate reduced dips in cross sections via $\text{tg } \alpha_R = \text{tg } \alpha \cdot \sin [90 - \gamma]$ with α_R being reduced dip and γ being angle between azimuth cross-section and dip direction.

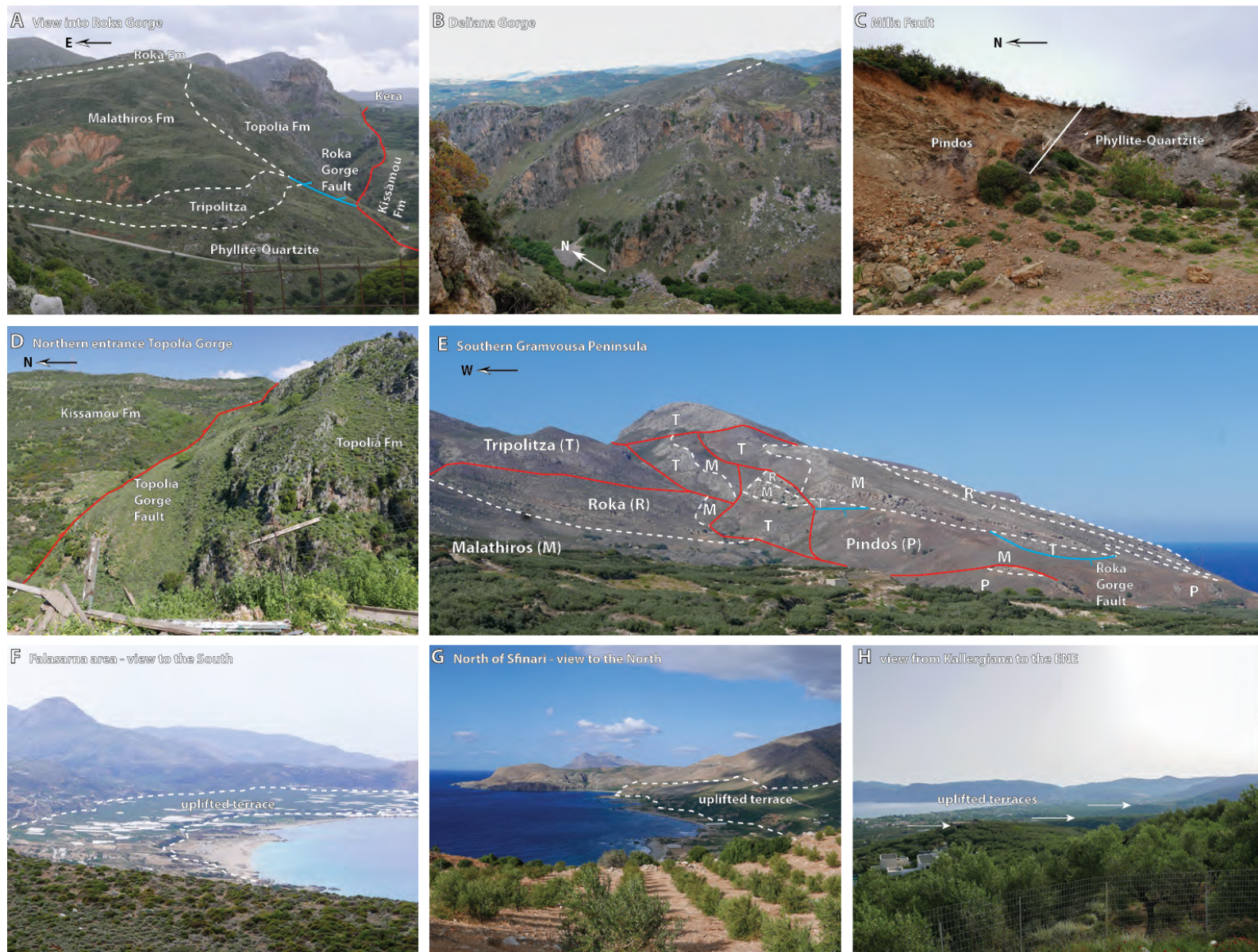


Figure 9 – (A) view into the Roka Gorge from the north showing deeper part of the Roka Gorge Fault (in blue) that separates Phyllite-Quartzite from Topolia conglomerates and younger fault between Topolia and Kissamou Fms; (B) northward dipping Topolia conglomerates (view of E-side of Deliana Gorge); (C) 50°N dipping Milia Fault between Pindos and Phyllite-Quartzite rocks along old road Topolia-Vlatos (0.9 km ESE of Milia but no longer visible due to construction of new road); (D) Topolia Gorge Fault separating Topolia conglomerates from Kissamou Fm to the north (view of E-side of north entrance of Topolia Gorge); (E) eastward tilted blocks on southern Gramvousa dissected by post-Roka faults (view from the S). In blue: Roka Gorge Fault; (F) uplifted terrace in Falasarna area (view to the S); (G) uplifted terrace north of Sfinari; (H) uplifted and incised terrace (arrows) formed in Messinian sediments landward of present north coast (view from Kallergiana to the ENE).

lithology, strong karstification and the travertine coated and vegetated gorge walls (Figures 5h,i and 9b). Number and size of the blocks in Figure 8 are determined by the thickness of the Topolia Fm, dips and the fact that the Topolia and Roka Gorges are floored by the Topolia Fm.

Most other normal faults in the N-S trending cross-sections of Figures 8A-C clearly post-date the series of north-tilted Topolia conglomerates, but not all are of the same age. The north-dipping Milia Fault in Figures 8A-B is the northern boundary of the present-day central Cretan Phyllite-Quartzite-cored horst. The coarse marginal facies of the Kissamou Fm including Topolia olistoliths (single or in debrites) directly north of this horst (Figure 3) indicates that the Milia Fault bounded the deep marine basin in times of deposition of the Kissamou Fm, with a hinterland to the south (Figure 9c). We suspect that the Milia Fault continues west- and eastwards between remnants of Tripolitza and Pindos nappes in the north and the Phyllite-Quartzite-dominated region to the south (Figure 3). Faults north of the Milia Fault in Figures 8A-B are younger because they offset the Kissamou as well as the Cheretiana Fms. The southernmost of these younger faults, the Topolia Gorge Fault in Figures 8A-B, is north-dipping and part of a fault system that forms the current boundary that separates the hilly landscape of post-Topolia Fm sediments to the north from the rugged mountains of Topolia conglomerates or basement to the south (Figure 9d) and steps northwards west of Topolia and continues via Polyrrinia to the west coast (Figure 3).

Finally, the E-W trending cross-sections in Figures 8D-F show several W-dipping faults (marked in purple) bounding E-tilted blocks (Figure 9e). Tilting probably occurred after emergence of western Crete because of Piacenzian sediments in the hangingwall of one these W-dipping faults (Figure 8D).

5 Discussion

5.1 The Basin History of Northwestern Crete

5.1.1 Formation of the Topolia Basin

Reconstructing the geometry, limits, and bounding faults of the sedimentary basin in which the Topolia Fm was deposited relies on interpretations of the scattered remains of the Topolia Fm and the formation's content. The Topolia Fm is bounded from all younger units by unconformities, and this first stage of sedimentation thus likely occurred in a basin that was significantly different from the younger depocenters.

From the newly collected paleocurrent data from southern Gramvousa and the Roka Gorge, as well as previous observations in the Topolia Gorge (*van Hinsbergen et al.*, 2008) that show systematically S-directed drainage, it follows that the source of the Topolia Fm was located to the north. The Topolia conglomerates occur up to the modern west coast (Figures 3 and 6) indicating that the mostly non-marine

Topolia Basin extended farther westwards than the modern west-coast, which thus post-dates the Topolia Fm.

To the east, we find the Topolia Fm up to the line Zimbragos – Episkopi (Figures 3,10). The formation is found on southwestern Crete as well (*Creutzburg*, 1977; *Kopp and Richter*, 1983; *Seidel et al.*, 2007; *van Hinsbergen et al.*, 2008) indicating that the Topolia Basin extended all the way to the south coast as argued by *Kopp and Richter* (1983) or that it may have been part of a larger system of half-grabens (*Seidel et al.*, 2007).

Pindos and Tripolitza debris was supplied by headwater rivers from a topographic relief to the north into the Topolia Basin where it deposited as fluvial (fan) conglomerates. The olistoliths of Tripolitza carbonates (up to 0.3 km² in size) in the southern Gramvousa-Trachilos area and a small one in the Roka Gorge show that the relief must have been considerable and steep.

The tectonic nature of the Topolia Basin is poorly constrained. Nonetheless, the paleocurrents suggest a southward tilt of the basin floor. This, and the south-dipping Roka Gorge Fault that must have formed during the deposition of the Topolia Fm is consistent with dominantly N-S extension as driver for basin formation on Crete, in line with the earliest, E-W trending Males-Viannos Basin of eastern and central Crete (*Zachariasse et al.*, 2011). On northwestern Crete, the Roka Gorge Fault is the local northern edge of the basin. Since no Topolia Fm conglomerates are known to the north of this fault, the currently exposed fault may have been the deeper (Roka Gorge) and shallower (Gramvousa) portions of the original northern basin-bounding fault (Figure 10). When the Topolia Fm was deposited, the Phyllite-Quartzite unit was not at the surface yet, but it must have been in the shallow subsurface such that the Topolia Fm at the northern entrance of the Roka Gorge became juxtaposed to the Phyllite-Quartzite unit shortly after deposition. This suggests that the Topolia Basin formed by activity of steep faults that cut through the Cretan Detachment.

On northwestern Crete, the Roka Gorge Fault may represent the northern edge of the basin, but the overall paleo-flow suggests that the larger subsidence occurred along faults in the south of the basin, well outside of the mapped area, likely to the south of western Crete.

The Topolia Basin submerged to < 100 m below sea level as evidenced by inner shelf sediments in a small area South of Ano Kalathenes that have similar dips as the fluvial Topolia conglomerates to the S and are unconformably covered by the Roka Fm (Figure 3).

5.1.2 Break up of the Topolia Basin; Transition to the Northwest Cretan Basin

The breakup of the Topolia Basin is illustrated schematically in Figure 11. The Topolia Basin was at some point uplifted and subjected to denudation. The formation was then tilted, and we hypothesized

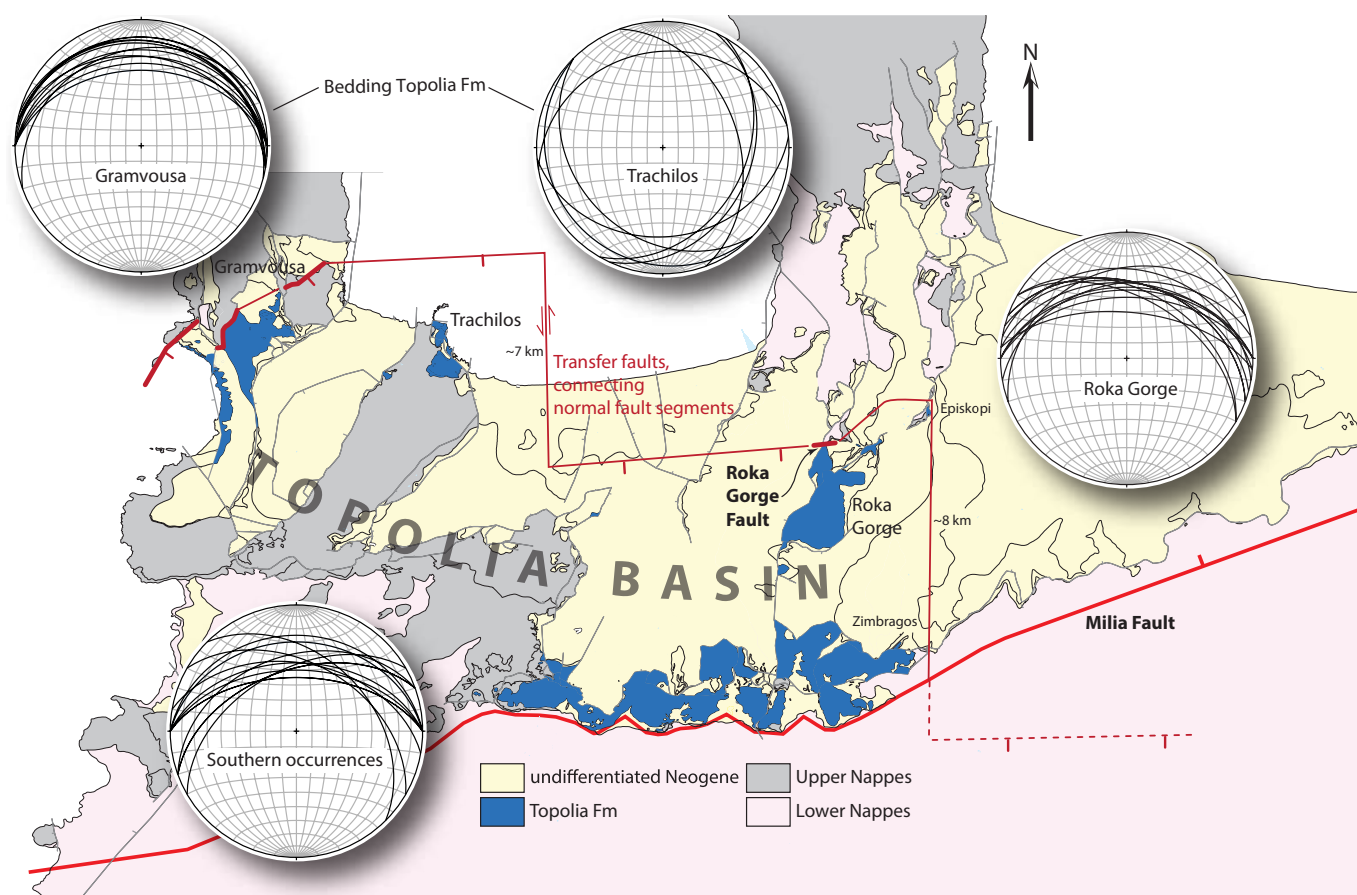


Figure 10 – Distribution of the Topolia Fm in northwestern Crete. The Topolia Basin-bounding fault is interpreted to step southward along N-S trending transfer faults. Stereoplots present bedding orientations in the Topolia Fm.

above that this may have occurred in a series of northward-tilted blocks separated by normal (listric) faults that we speculate root in a S-dipping decollement. We infer that the northern break-away point of this fault was located to the north of the Roka Gorge Fault because even the exposed deeper portion of this fault is very steep (Figures 8A and 11; Figure 9a).

Some 500 kyr after the formation of the Topolia Basin and before the deposition of the Malathiros Fm, the Phyllite-Quartzite unit became first exposed in the break-away fault of the decollement and later also on the footwall to the north (Figures 8A and 11). Deposition of the fluvial Malathiros Fm reflects renewed subsidence related to ongoing extension.

Other than the Topolia Fm that occurs also on southern Crete, rocks of the Malathiros Fm and the next-younger, and more widespread shallow marine Roka Fm, are restricted to northwestern Crete. They are not found to the south of the Milia Fault, and we infer that this fault started to localize a depression in northwestern Crete. Its deposits, starting with the Malathiros Fm, unconformably overlie the Topolia Fm, and we consider this a separate, younger basin that became superimposed on the remains of the Topolia Basin: the Northwest Cretan Basin. The highly variable thickness of the Malathiros Fm indicates that sedimentation in the Northwest Cretan Basin began with the filling of depressions while the widespread shallow

marine sediments of the next younger Roka Fm reflect submergence to below sea level (Figures 8A and 11).

The timing of the breakup of the Topolia Basin is constrained by the oldest deep marine sediments overlying the Topolia Fm on southwestern Crete, with an age of 10.4 ± 0.1 Ma (see section 5.1) and the onset of the deposition of the Malathiros Fm around 9.6 Ma (see SI Item 5).

5.1.3 Late Neogene Evolution of the Northwest Cretan Basin

From 9.6 Ma onwards the Northwest Cretan Basin subsided to deep marine conditions, as reflected by deposition of the fluvial-lacustrine Malathiros Fm, conformably followed by the shallow marine Roka Fm and the deep marine Kissamou and Cheretiana Fms (Figure 4). Sediments of this basin are restricted to northwestern Crete, from the west coast to Chania in the east, and from the north tip of the Gramvousa Peninsula (*Frydas and Keupp, 1996*) to the central Phyllite-Quartzite basement horst in the south (*Freudenthal, 1969, Figure 3*). Contrary to the older, Topolia Basin, the hinterland for the Northwest Cretan Basin was in the south, which is particularly evident from debris flow deposits and many coarse-grained turbiditic sandstones in the Kissamou and Cheretiana Fms directly north of the Milia Fault, while finer equivalents with more marls in between occur further north.

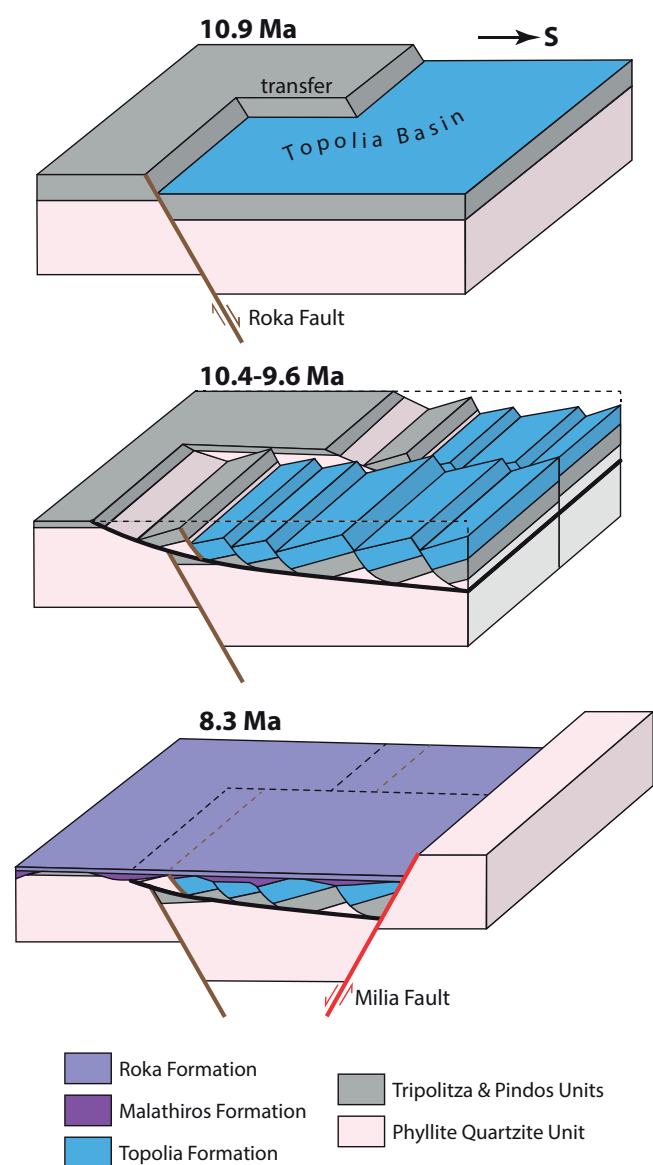


Figure 11 – Schematic block diagrams showing the interpreted breakup history of the Topolia Basin.

Subsidence in the basin started slowly, associated with depression-filling fluvial sediments (Malathiros Fm) containing for the first time Phyllite-Quartzite debris originating from the newly formed footwall of the Northwest Cretan Basin. The subsequent shallow marine Roka Fm and overlying deep marine Kissamou Fm reflect rapid subsidence to 500-750 m below sea level (*van Hinsbergen and Meulenkamp, 2006*; SI Item 7).

The uniform depositional depth range for the Kissamou Fm (SI Item 7), and probably also for the Cheretiana Fm, suggests that sedimentation kept roughly pace with tectonic subsidence. The amount of tectonic subsidence from end Roka to end Cheretiana Fm (without Messinian gypsum; 8.3-6.0 Ma) is ~350 m, yielding an average subsidence/sedimentation rate of ~15 cm/kyr). Total subsidence of the Northwest Cretan Basin between onset submergence to below sea level and drawdown of the Mediterranean at 5.60 Ma then amounts to 900-1250 m (400-500 m sediment thickness and basin floor 500-750 m below sea level).

The change from the greyish clays of the Kissamou Fm to the beige, more calcareous marls of the Cheretiana Fm is documented all over Crete and is attributed to diminished clastic supply (*van Hinsbergen and Meulenkamp, 2006*). On northwestern Crete, the above change is accompanied by an increase in number and thickness of sapropels possibly due to the increase in Eastern Mediterranean salinities to hypersaline values around the same time (i.e. 6.7 Ma), thought to reflect restrictions in the Mediterranean-Atlantic seawater exchange (*Zachariasse and Lourens, 2021*).

Likewise, the change from the deep-marine Cheretiana to the non-marine conglomerates of the Elliniko Fm, intervened by a period of erosion, is related to the Messinian Salinity Crisis. The terminal Messinian is the time interval defined by an increasingly obstructed Mediterranean-Atlantic exchange, with a major marine evaporitic stage between 6.00 and 5.60 Ma followed by drawdown of the Mediterranean Sea level at 5.60 Ma and restoration of this exchange at 5.33 Ma (*Krijgsman et al., 1999; Roveri et al., 2014*); for the latest overviews of the MSC, see *Andreotto et al. (2021); Krijgsman et al. (2024)*). On northwestern Crete, drawdown is manifested by an erosional unconformity (MES in *Roveri et al. 2014*) which cuts into the gypsum and deeper layers of the Cheretiana Fm.

The 200 m thick Elliniko Fm of fluvial and lacustrine clastic sediments with debris still supplied from a southern hinterland represents the Lago Mare stage of the MSC. This stage follows drawdown and requires a substantial rise in the base level of the South Aegean forearc Basin. This is likely caused by the inflow of oligohaline water from the Paratethys lakes of southern Eurasia witness the occasional presence of brackish ostracods in the Elliniko Fm (see also *Krijgsman et al. 2024*).

The sharp transition from the non-marine Elliniko Fm to the deep marine Tavronitis Fm marks the restoration of the Mediterranean-Atlantic exchange.

From the distribution of the deep-marine sediments in the Northwest Cretan Basin, we infer that the southern basin-bounding Miliaria Fault stepped northwards during the terminal Messinian to form the Topolia Gorge Fault (Figure 12). We infer this for instance from the lack of a (coarse) basin margin facies of the Cheretiana Fm near Limni (Figures 3 and 12). There, homogeneous and laminated Cheretiana marls, which reflect distal, deep-water sedimentation, are located immediately adjacent to the Topolia Gorge Fault that separates the Northwest Cretan Basin sediments from the basement. The Elliniko Fm, however, is located both to the S and N of this fault, suggesting that prior to draw-down, the Topolia Gorge Fault must have cut the Cheretiana basin floor and dropped the north down relative to the south (Figure 12). Another argument is that the southernmost remnants Roka, Kissamou and Elliniko Fms overlying Topolia conglomerates between Zimbragos and Topolia (Figure 3) require erosion of Roka and Kissamou Fms prior to the deposition of the Elliniko Fm.

Renewed Mediterranean-Atlantic exchange at 5.33 Ma caused the sea level to rise ≥ 500 m on northwestern Crete. It is therefore likely that the Tavronitis Fm was also present on the new footwall given the current elevation differences of < 200 m between basal Pliocene sediments to the north and remnants Elliniko Fm to the south of the Topolia Gorge Fault (Figure 12). The Topolia Gorge Fault thus accommodated 600-700 m (400-500 m Roka, Kissamou and Cheretiana and 200 m Elliniko sediments) (Figure 12). Such a large amount of tectonic subsidence in the terminal Messinian is not unusual for Crete (Zachariasse *et al.*, 2008).

Also in the Pliocene, subsidence continued along the Topolia Gorge Fault as evidenced by northerly directed flute casts in the upper Tavronitis Fm. The many sandstones with small-scale crossbedding and convex up geometries in the basal part during the first 400 kyr of the Pliocene (biozone 1, see SI Item 3) are interpreted as being part of a submarine fan that developed in front of a source area in the south. Paleodepth estimates of 500-750 m from base to top and from west to east (SI Item 7) are like those of before the Messinian Salinity Crisis during deposition of the Kissamou and Cheretiana Fms, again suggesting that sedimentation kept roughly pace with tectonic subsidence with rates of ~ 10 cm/kyr. The coarsening upwards in the Tavronitis Fm is likely reflecting the onset of wholesale uplift of Crete since the early Pliocene (Meulenkamp *et al.*, 1994; van Hinsbergen and Meulenkamp, 2006; Zachariasse *et al.*, 2008), even though our paleodepth estimates of 500-750 m (SI Item 7) show that most uplift occurred after the deposition of the youngest Tavronitis sediments in the Northwest

Cretan Basin. These youngest sediments belong to biozone 8 with ages for bottom and top of 3.31 and 2.72 Ma (SI Items 3, 9). These sediments are now exposed at elevations of 240 m in the Platanos area and 50 m at Tavronitis requiring uplift rates on the order of 20 to 30 cm/kyr. With these uplift rates, the sea floor surfaced sometime between 1 and 0.2 Ma, somewhat later than the timing for the northern (1.7-1.1 Ma) and southern Heraklion Basin (1.9-1.4 Ma) on central Crete (for calculations of the ages for the surfacing of the sea floor, see SI Item 10). These uplift rates are lower than estimates for the late Quaternary based on raised terraces, which are on the order of 50-80 cm/kyr (Robertson *et al.*, 2023; Strobl *et al.*, 2014), suggesting that uplift rates accelerated since the late Pliocene.

Once above sea level, the Northwest Cretan Basin broke up into horsts and grabens with the downthrow along the N-S trending normal faults being greatest on the west side of the fault blocks (Figures 8D-F) suggesting tilting along W-dipping normal fault(s) showing that E-W extension on northwestern Crete (re)started in the Late Pliocene-Pleistocene after an initial phase of arc-parallel extension preceding basin formation (van Hinsbergen and Schmid, 2012).

Finally, the interplay between ongoing uplift of northwestern Crete and (inter)glacial sea level changes is manifested by uplifted terraces at different elevations. Also, the ~ 180 m high cliff on the landward side of the Falasarna coastal area should have been shaped by tectonic uplift and sea level changes. The inner neritic sediments of the wide uplifted terrace of the Falasarna coastal area (Figure 9f) deposited during an interglacial sea level high stand (e.g. Eemian) whereas the fluvial (fan) conglomerates on top belong to a glacial low stand phase. The uplifted fluvial (fan) conglomerates at Sfinari (Figure 9g) belong to ≥ 1 glacial low stands and overlies an old, abraded surface of Phyllite-Quartzite (see also Figure 3). Terraces along the north coast are present in the form of uplifted fluvial (fan) conglomerates and a steplike (terraced) interruption in the northern slope of the hilly Neogene landscape (Figure 9h). The present coastline of northwest Crete was formed during the 365 AD earthquake which lifted the coast from 7 m at Falasarna (Kelletat, 1991) through 6.6 m at the Roman harbor of Kissamos (Stiros and Papageorgiou, 2001) to 2 m at Chania in the east (Kelletat, 1991).

5.2 Relationship Between Basin Evolution and the Final Exhumation of HP-LT Rocks on Northwestern Crete

The question that initiated this study was whether the formation of the oldest sedimentary basins on Crete was governed by the same processes and structures that drove the long-term exhumation of HP-LT metamorphic rocks on Crete from depths of tens of kilometers to the near-surface, as often inferred, or reflect an unrelated and younger tectonic event. In any case, it is clear that when sedimentation started, the Phyllite-Quartzite unit must have already been exhumed to close to

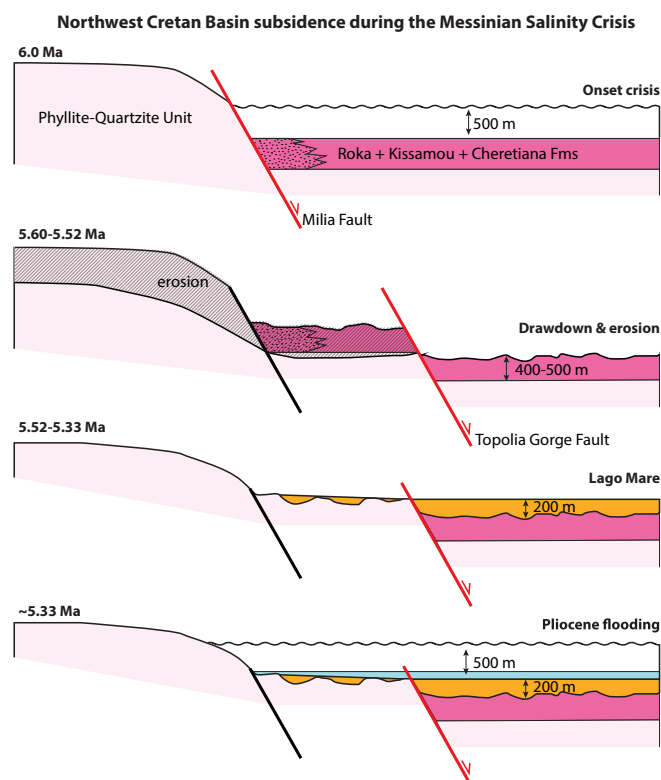


Figure 12 – Schematic N-S cross-section at Limni visualizing the subsidence history of the Northwest Cretan Basin during the Messinian Salinity Crisis (6.0-5.33 Ma).

the surface, demonstrating that the main exhumation mechanism did not generate topographic depressions prone to sedimentation. Despite the absence of HP-LT metamorphic rock debris in the Topolia Fm, existing interpretations link the deposition of the Topolia Fm to exhumation along the Cretan Detachment. If the Topolia Fm formed in an extensional setting then this opened possibilities that the Topolia Fm deposited in an extensional basin whose faults rooted in the Cretan Detachment (Jolivet *et al.*, 1996; Seidel *et al.*, 2007; van Hinsbergen and Meulen Kamp, 2006), or alternatively, that the Topolia Fm was deposited in valleys eroded into a compressional setting preceding regional extension and basin formation (Ring and Yngwe, 2018).

Our analysis shows that shortly after the deposition, the Topolia Fm disintegrated into a series of northward tilted blocks above S-dipping normal (listric) faults that became inactive during the formation of the Milia Fault being the southern boundary of the Northwest Cretan Basin. We showed that the Roka Gorge Fault locally juxtaposed the northernmost Topolia Fm against Phyllite-Quartzite and that this fault is unconformably covered by the ≤ 9.6 Ma old Malathiros Fm (Figures 8A and 11). So, prior to the formation of the Northwest Cretan Basin at ~ 9.6 Ma, the Phyllite-Quartzite was at the surface, and the contact between the Phyllite-Quartzite Unit with the Tripolitza and overlying units, interpreted as the Cretan Detachment (Jolivet *et al.*, 1996) was already cut by high-angle normal faults. This confirms findings from central Crete, where block faulting along high-angle normal faults that must cut this contact also occurred since at least 9.6 Ma (Zachariasse *et al.*, 2011).

This limits the potential interval in which sedimentation occurred coevally with exhumation by extension along the Cretan Detachment (or thrusting, in the scenario of Ring and Yngwe 2018) to only the deposition of the Topolia Fm. But we find also for the formation of the Topolia Basin no evidence for a relationship with the Cretan Detachment. Instead, we consider it more likely that the northern boundary of the basin formed by the Roka Gorge Fault already cut through the detachment. The last stages of exhumation, which brought the Phyllite-Quartzite unit to the surface, most likely occurred via subaerial erosion between the incipient breakup of the Topolia Basin (upper age limit of 10.4 ± 0.1 Ma) and 9.6 Ma.

Our findings are thus in agreement with the models of Jolivet *et al.* (1996) and van Hinsbergen and Meulen Kamp (2006): final exhumation took place in an extensional setting and post-dates Topolia deposition. However, in the present study exhumation occurred in the footwall of a high-angle normal fault that cut through, rather than rooted in the top-to-the-North Cretan Detachment of Jolivet *et al.* (1996), as suggested by van Hinsbergen and Meulen Kamp (2006) and Seidel *et al.* (2007). Instead, our detailed study of the structure and stratigraphy of northwest Crete yielded no evidence that supports the interpretation that the Topolia Basin was a supra-detachment basin as in the models of Jolivet

et al. (1996), van Hinsbergen and Meulen Kamp (2006), and Seidel *et al.* (2007). We also find no evidence that the Topolia Basin formed in a setting of shortening, thrusting, or folding as suggested by Ring and Yngwe (2018).

Instead, we infer that the Cretan Detachment must have become inactive before basin formation at 10.9 Ma. The Topolia Basin that extended from northwestern Crete all the way to the south coast (Creutzburg, 1977; Kopp and Richter, 1983; van Hinsbergen *et al.*, 2008), and the subsequent localization into the Northwest Cretan Basin, occurred under N-S extension, like the extension direction along the Cretan Detachment.

However, there must have been a distinct subdivision between the activity of structures that accommodated the main and final exhumation of the Phyllite-Quartzite and underlying Plattenkalk and Tripali units. Despite the major thinning of the upper nappes above the detachment, and the main exhumation of the underlying units from great depth, Crete must have remained topographically high. van Hinsbergen and Schmid (2012) tentatively ascribed the stretching of the Tripolitza and higher nappes, which had all accreted to the Aegean upper plate around 25 Ma, to trench-parallel extension in the forearc during kinematically restored and paleomagnetically documented oroclinal bending. They explained the lack of subsidence during this extreme thinning to the buoyant rise of the Phyllite Quartzite and deeper nappes that had previously been suggested to account for their exhumation (Thomson *et al.*, 1998, 1999). Because this rise occurs along the slab, the exhuming HP wedge may have remained uncoupled to the upper plate until late in the exhumation (van Hinsbergen and Schmid, 2012); their Figure 13). The earliest subsidence and extensional basin formation around 10.9 Ma that we document in this paper may mark the onset of coupling between the exhumed HP wedge and the remains of the Tripolitza and higher nappes, stretching in unison above the retreating Aegean slab. We speculate that the extreme thinning of the upper nappes led to a loss of structural coherence, such that further extension could no longer be accommodated along the existing detachment. The upper nappes became passive on the exhumed Phyllite-Quartzite unit, and ongoing extension became accommodated along new fault systems that crosscut the older ones, and that we document in this paper. Ongoing rise of subducted buoyant rock at depth may have continued to play a role in the vertical motions that ensued after the onset of basin formation and could have played a role in the youngest uplift (e.g., Gallen *et al.*, 2014) and perhaps underpin the transition from the Topolia to the Northwest Cretan Basin. A key unknown remains the detailed structure of the highly dismembered upper nappes, which may contain further clues to the dynamics that governed exhumation in an extensional setting, but without subsidence.

6 Conclusions

The stratigraphic analysis and detailed mapping of the Neogene of northwestern Crete has yielded the following conclusions:

1. The Neogene on northwestern Crete deposited in two superimposed basin systems each bounded by different sets of normal faults and under ongoing N-S extension.
2. The oldest Topolia Basin (10.9-10.4 Ma) dates from before the final exhumation of the HP-LT nappes and extended across most of present-day western Crete. Its northern boundary is a S-dipping normal fault which cuts through the Cretan Detachment and jumps in two, kms-long, steps southwards along N-S trending transfer faults.
3. The break-up of the Topolia Basin between 10.4-9.6 Ma is marked by uplift and the formation of N-dipping blocks above S-dipping normal (listric) faults that we infer root in a decollement and is concluded by the formation of the Northwest Cretan Basin whose southern N-dipping boundary fault cuts through the decollement.
4. Phyllite-Quartzite debris in the oldest (≤ 9.6 Ma) fluvial unit of the Northwest Cretan Basin reflects exhumation and erosion on the newly formed footwall of the Northwest Cretan Basin. Earlier exhumation, between 10.4 and 9.6 Ma, must have occurred through erosion on the former footwall of the Topolia Basin but this period lacks sediments.
5. The Northwest Cretan Basin subsided to 500-750 m below sea level at 8.3 Ma and remained at that depth after the southern boundary stepped northwards during the end-Messinian drawdown thereby accommodating 800-1000m subsidence between 5.52 and ~ 3.00 Ma on top of the 900-1250 m subsidence between 8.3 and 5.60 Ma along the original southern boundary.
6. Surfacing of the Northwest Cretan Basin floor occurred at the earliest 1 Ma ago being slightly later than the earliest ages of 1.7 and 1.9 Ma for the northern and southern Heraklion Basin on central Crete. Surfacing is followed by a Pleistocene phase of E-W extension under ongoing uplift of Crete.

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Author contributions

WJ Zachariasse performed fieldwork and research, analysed samples, made figures, and drafted the manuscript. **DJJ** van Hinsbergen improved the manuscript by restructuring and placing it in a broader context and converted the raw into the final figures.

Data availability

Supporting information, Google Earth maps, and data are available on Figshare: <https://doi.org/10.6084/m9.figshare.27637413.v1>. They include the following:

- Item 1: Geological map of northwest Crete. (A) Full PDF version of the geological map of northwestern Crete including geographical place names and bedding orientations. (B) Original map drawn on Google Earth.
- Item 2: Locations visited in this study along with lithological and any other information.
- Item 3: Pliocene planktonic foraminiferal biozones and ages.
- Item 4: The Neogene stratigraphy and vertical motion history of Gavdos Island.
- Item 5: Age constraints for the base of the Malathiros Fm.
- Item 6: Age estimates for base Kissamou Fm in northwestern Crete and base of the time-equivalent Apostoli Fm in the Rethymnon area.
- Item 7: Depositional depth estimates for upper Miocene and Pliocene locations.
- Item 8: The age of sediments on the small island off the coast of the town of Kissamos and implications for age and depth of the Trachilos 'hominin-like footprint' site.
- Item 9: Ages for Pliocene samples.
- Item 10: Details for calculating seafloor surfacing on northwest and central Crete.

Competing interests

The authors declare no competing interests.

Peer review

This publication was peer-reviewed by Armel Ménant and Alessandro Petrocchia. The full peer-review report can be found here: Review Report.

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