

Exhumation of high-pressure metamorphic rocks within an active convergent margin, Crete, Greece: A field guide

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Introduction

Subduction involves the horizontal convergence of two tectonic plates; however, researchers have long recognized that widespread extensional deformation is a characteristic of many subduction zones (e.g., Royden, 1993). The Hellenic subduction zone is one such region of overall plate convergence and widespread horizontal extension (Figure 1). As the lithosphere of the African plate dips to the north, deformation in the overriding European plate has thinned the continental crust, forming the Aegean Sea. Unlike in some subduction zones (e.g., Japan), extension in the overriding plate is not restricted to the backarc region; thinned continental crust underlies the Sea of Crete outboard of the volcanic arc. Although most of the attenuated continental crust is now submerged, the island of Crete provides a view of extensional deformation in the forearc region. Crete exposes a variety of variably metamorphosed sedimentary and volcanic units juxtaposed by thrust faulting during the Oligocene and later thinned by recent (and still active) normal faults. This tectonic thinning has exhumed high pressure-low temperature metamorphic rocks, which were only recently metamorphosed (~20 Ma) at about 35 km depth above the subducting slab. Thus, the island provides an excellent laboratory to study processes and consequences of syn-convergent extension.

The goal of this field excursion is to examine the tectonic evolution of Crete, including the deformational, metamorphic, thermal, and exhumational history of the rocks. The four day trip will begin in Psiloritis Mountains of central Crete, where we will introduce the major tectonostratigraphic units, and inspect several exposures of the Cretan detachment fault. On the second day, we will examine the thermal and deformational history of the high-pressure, low-temperature metamorphic rocks in Crete. The third day will focus on the development of sedimentary basins in western Crete in response to widespread

late Miocene extension. The fourth day will involve a hike through the deeply incised Samaria Gorge, which provides geomorphic evidence of recent rapid uplift of a large footwall block associated with active normal faulting along the south side of Crete.

Essential logistical information:

Crete is very accessible. The island is serviced by two airports, one in Irakleio and the other in Chania. Here, and in the other major cities in Crete, it is possible to rent a car or van. The localities described in the first three days of the field guide are road cuts that are easily reached by automobile. The fourth day of the trip is a hike through the Samaria Gorge in western Crete. To pass the gorge, it is best to take a bus to the village of Omalos. Public buses depart from Chania and Palaiochora. After you hike down the gorge, you will arrive in the village of Agia Roumeli. Here you can purchase a ticket for a boat that will transport you to Chora Sfakion, from where buses will be available for transportation back to Chania or elsewhere.

Field references:

Several relevant field guides have been published that describe the geology of central and western Crete. *Field Guide to the Geology of Crete* by Charalambos Fassoulas outlines 7 days of field stops throughout all of Crete. The book is intended for both the general public and geologists, and is well-illustrated with many color photographs. Another field guide has been written by Meulenkamp and others (1979), who describe a 4 day trip that focuses on the Miocene sediments of western and central Crete.

A good road map is essential for a field trip to Crete, as many of the smaller roads in the interior of the island are unmarked and can be difficult to find. We recommend the series of maps by produced by Road Editions.

The most comprehensive geologic map has been compiled by Creutzberg et al. (1977). A series of more detailed maps has been drafted by the Institute of Geological and Mining Exploration (IGME) of Greece.

Regional tectonic setting

Overview

The Mediterranean Sea represents a vestige of the Tethyan Seaway, an east-west trending ocean basin formed during the breakup of the Pangea. Throughout the Cenozoic, convergence between

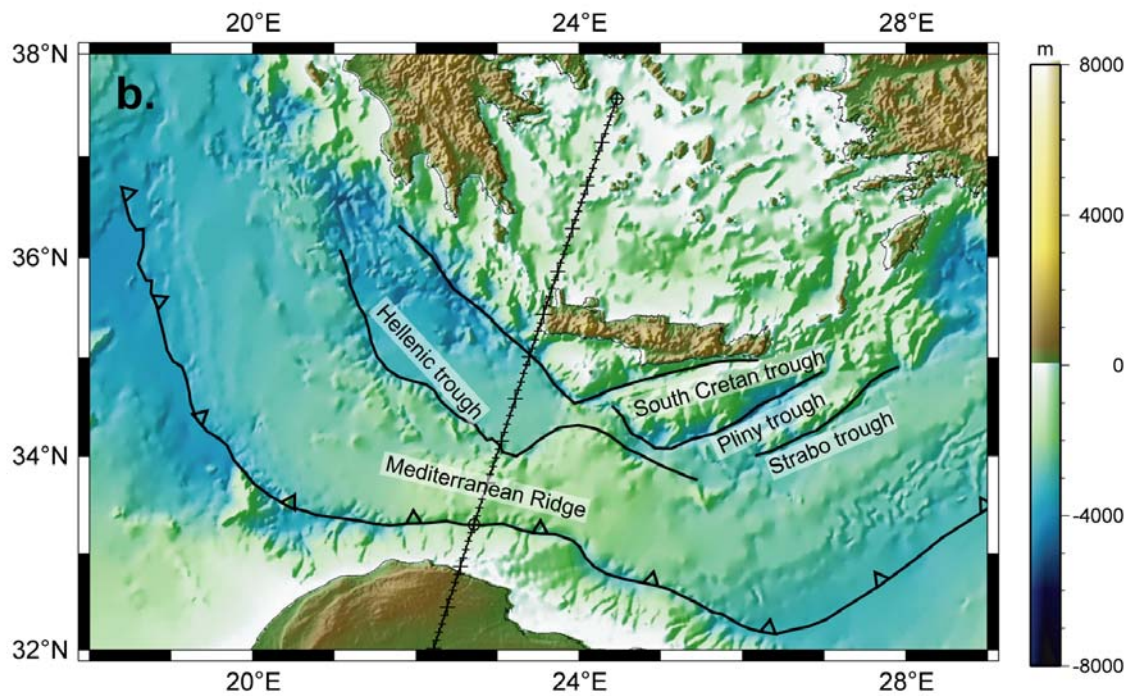
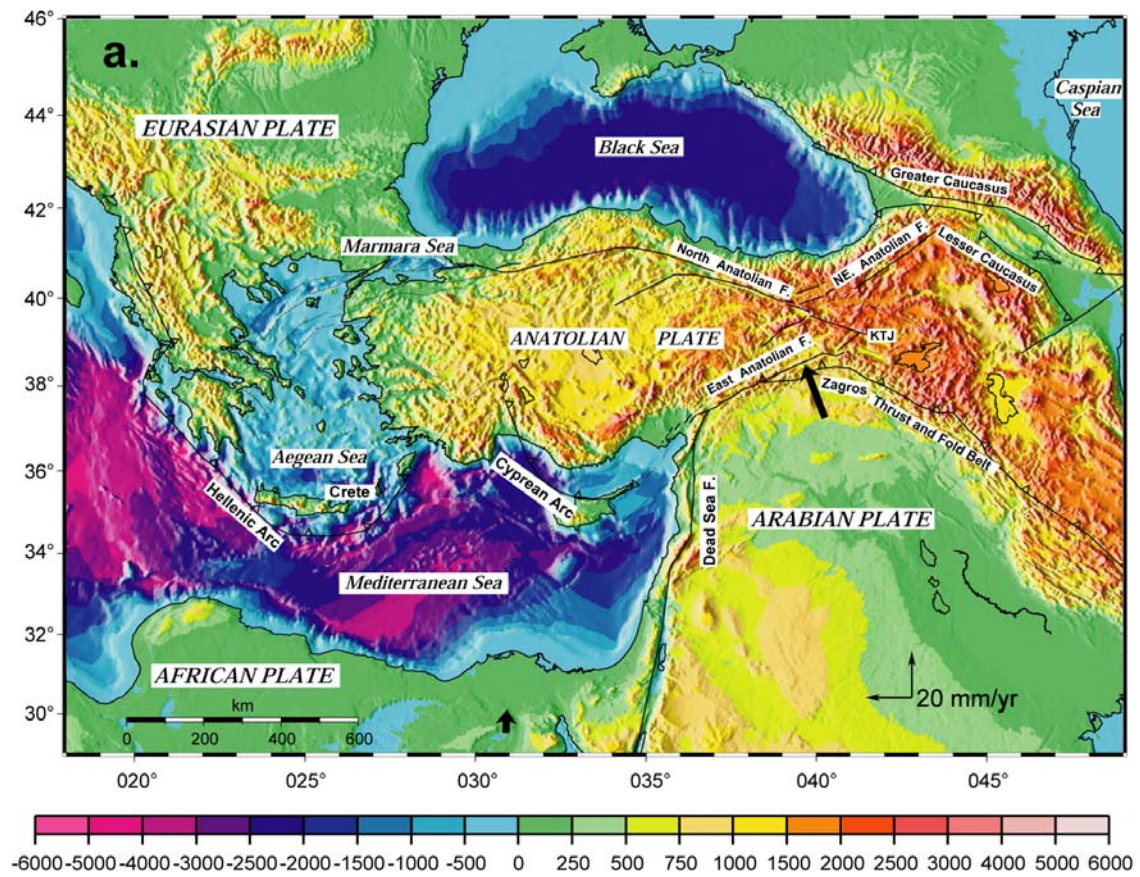


Figure 1 – a) Simplified tectonic map of the eastern Mediterranean region, from McCluskey et al. (2000). Structures are superimposed on topography and bathymetry. Heavy arrows show NUVEL-1A plate motions relative to Eurasia. KTJ – Karlova triple junction. b) Topography and bathymetry of the eastern Mediterranean. Line shows location of cross-section in Figure 2.

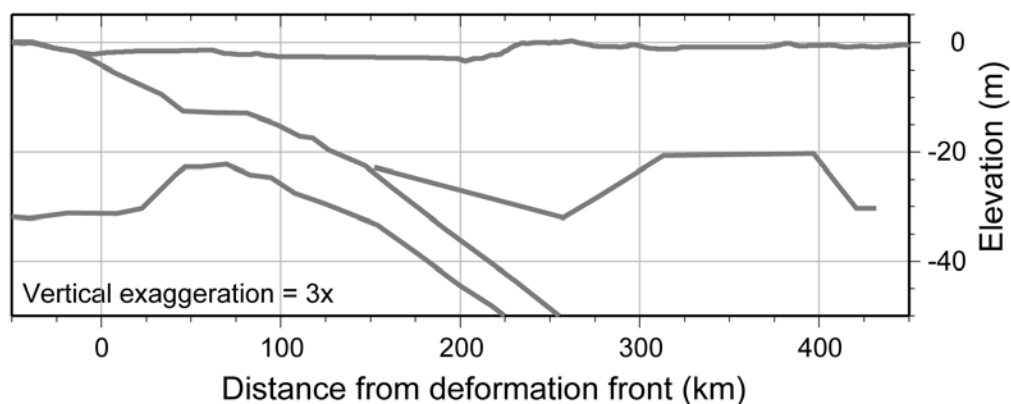
Eurasia and southern plates of Gondwana (including the African, Apulian, Arabian, and Indian plates) slowly consumed the Tethyan Seaway and led to the formation of the Alpine-Himalayan chain, an orogenic zone that extends for several thousand km from the Alps in southwestern Europe through the Middle East and into the Himalaya.

The Hellenic subduction zone is presently consuming the remnants of Tethyan seafloor, which is subducting northward beneath Crete as part the lithosphere of the African plate (Figures 1 and 2). Crete lies in the forearc of the Hellenic Subduction Zone. Active subduction is indicated by a north-dipping Wadati-Benioff seismic zone extending beneath Crete to a depth of about 200 km (Le Pichon and Angelier, 1979; Knapmeyer and Harjies, 2000). Tomographic studies show that this seismicity corresponds to a cold lithospheric slab that extends through the transition zone and into the lower mantle below Europe (Spakman et al., 1988). At the longitude of Crete, the subduction front in this system is located about 200 km south of the island, just north of the coast of Libya (Kastens, 1991; Kopf et al., 2003) (Figures 1 and 2). Sediments from the downgoing plate have accumulated in the Mediterranean Ridge complex, a broad accretionary wedge positioned between Africa and Crete. Kopf et al. (2003) estimate an accretionary flux of $> 17 \text{ km}^2/\text{myr}$. Between Crete and the crest of the Mediterranean ridge are a series of east- and

northeast-trending depressions or troughs (e.g., Hellenic, South Cretan, Pliny, Strabo in Figure 1b). Early papers considered that these depressions might be the active subduction zone, but subsequent work has shown that the front of the subduction wedge lies along the southern margin of the Mediterranean Ridge (Kastens, 1991). Some authors have suggested that these troughs might mark sinistral strike-slip faults associated with the tectonic escape of Turkey (e.g., Huguen et al., 2001), although direct evidence for the formation of these troughs is lacking.

Crete itself is positioned as an emergent high in the forearc of the subduction system. North of the island, the topography quickly drops off into the thinned continental crust of the Cretan Sea (Figure 2) (Makris and Stobbe, 1984). About 100 km north of Crete lies the volcanic arc of the Hellenic Subduction Zone, represented by the island of Santorini. Like in the Sea of Crete, the crust in the back-arc of the system is also attenuated continental crust (McKenzie, 1978; Le Pichon and Angelier, 1981). North of Santorini, the various islands of the Cyclades expose mid- to lower-crustal metamorphic and igneous rocks exhumed to the surface along low-angle detachment faults (Lister et al., 1984). The exposure of high-pressure metamorphic rocks in the Aegean Sea is one line of evidence that the crust there has been thinned through extensional faulting (Lister et al., 1984; Buick, 1991). This conclusion is confirmed by seismic reflection studies which demonstrate a crustal thickness of 20 to 25 km (e.g., Makris, 1978; Makris and Stobbe, 1984).

Figure 2 – Cross section through the Cretan wedge, showing variation in topography and depth to the Moho. (Moho from T. Meier et al., unpublished results, 2003).



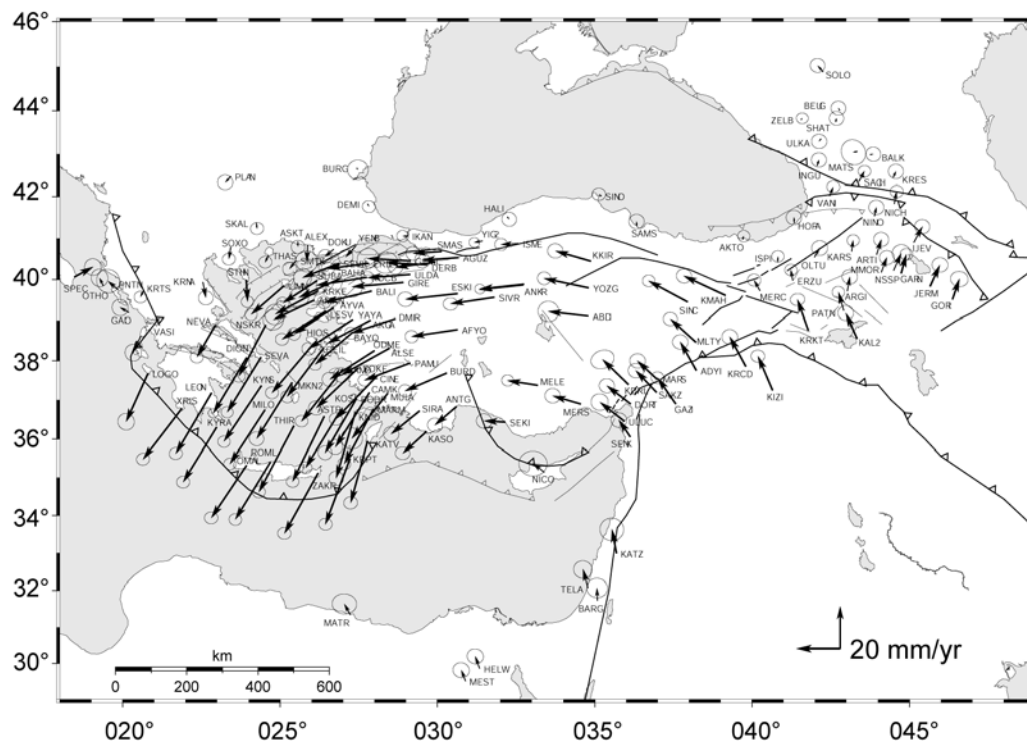


Figure 3 – GPS horizontal velocities and 95% confidence ellipses from McCluskey et al. (2000). Data are in an Eurasia-fixed reference frame for the period 1988-1987.

Global Positioning System (GPS) studies have provided a detailed view of the modern tectonic motions in the eastern Mediterranean (e.g., McCluskey et al., 2000) (Figure 3). In this region, one important feature shown by the velocity field is the westward migration of the Anatolian block towards the Aegean Sea along the right-lateral North Anatolian fault in northern Turkey. This movement is driven by the ongoing collision between the Arabian block and Eurasia, with Turkey “escaping” as a coherent block to the west (McKenzie, 1972). Escape is thought to start sometime between 12 Ma and 4 Ma (McCluskey et al., 2000). The extrusion of the Anatolian block can be modeled as a counterclockwise rotation (Le Pichon and Angelier, 1979; McCluskey et al., 2000), with both Turkey and the Aegean Sea moving with respect to Eurasia. Both the GPS data and seismic observations (Jackson, 1994) show that Crete and the southern Aegean are moving together at a coherent block. The divergent motion between the Aegean block and mainland Europe is accommodated in a zone of extension in the northern Aegean, with Crete and the Aegean Sea diverging from mainland Europe at a rate of about 30 mm/yr (McCluskey et al., 2000). Studies of global plate motions indicate northward motion of Africa relative to Europe at 10 mm/yr (e.g., Dewey et al.,

1989). Thus, the modern net convergence rate across the modern Hellenic subduction zone is about 40 mm/yr.

Southward migration of the Hellenic subduction zone

Subduction at the Hellenic Subduction Zone appears to have been operating continuously since at least 26 Ma, and likely back to 40 Ma (Spakman et al., 1988, Meulenkamp et al., 1988). Cretaceous units in the internal zones in the northern Aegean suggest that earlier subduction zones were probably operating further back in time (e.g., Kiliyas et al., 1999). Although the subduction process may have been continuous during most of the Cenozoic, it is clear that the significant changes in the geometry and kinematics of the zone have developed in recent geologic past. During the Oligocene, mainland Greece, Crete, and Turkey formed a roughly linear east-west trending belt. During the Oligocene, convergence led to the stacking of a series of sedimentary and volcanic units all along this margin (Le Pichon and Angelier, 1979; Bonneau, 1984).

These individual units represent different components of the Tethyan Seaway, including several carbonate platforms and adjacent basins (Robertson et al., 1991). The nappe stacking is commonly taken as evidence of continent-continent collision, but the continuity of the subducting slab beneath Crete suggests that the nappes probably represent the accretion of thin thrust sheets, as is typical of subduction zones, rather than the collision and accretion of lithospheric continental blocks.

Sometime since the Miocene, southward migration of the margin near Crete led to the development of the curvature of the Hellenic arc. Le Pichon and Angelier (1979) argue that the southward migration was related to the initiation of subduction in the region, which they believed occurred at 13 Ma because of the depth of active seismicity. However, the realization that subduction operated for a much longer time (Spakman et al., 1988) has forced a re-evaluation of the timing constraints for the southward migration of the Hellenic arc (see discussion in the Appendix of Kastens (1991)). Paleomagnetic results (Kissel and Laj, 1988; Duermeijer et al., 1998; Duermeijer et al., 2000) show that block rotations in the Hellenic arc were acquired mainly after 5 Ma, suggesting that the formation of the curvature of the Hellenic arc has developed since that time as well.

The southward migration of Crete appears to be related to the extension occurring above the Hellenic subduction zone. North of Crete, the crust beneath the Sea of Crete and the Aegean Sea is continental and typically 20-25 km thick (Makris and Stobbe, 1984). Prior to deformation, the crust of the Aegean Sea has been estimated to be 1.5 to 2 times thicker than the current thickness (McKenzie, 1978; Angelier et al., 1982).

How can widespread extension occur in regions of overall tectonic convergence? Several tectonic processes have been proposed to account for the southward migration of Crete and the formation of the Aegean Sea. One idea focuses on the "Anatolian push" associated with the westward escape of Turkey into the Aegean. The motion of the Anatolian block is proposed to be transmitted to the continental crust of the Aegean region which then spreads and extends southward, causing southward migration of the Hellenic subduction zone (Hatzfeld et al., 1997). This southward motion is sometimes described as overflow of the Aegean over the African plate. However, recent investigations suggest that the "Anatolian push" is probably of secondary importance in controlling the formation of the Aegean Sea. For example, using a finite element

model Meijer and Wortel (1997) found that the force applied to the Aegean by the Anatolian block had to be small to get a match to the modern velocity field. The GPS results of McCluskey et al. (2000) also suggest that the westward motion of Turkey does not strongly influence the southward motion of Crete. As described above, the divergence between Europe and the Aegean Sea is accommodated in a zone in the northern Aegean. Thus, extension across the southern Aegean and Crete appears to be related to a "pull" from the subduction zone rather than a "push" from Turkey.

A second idea is that the southward motion of Crete and the formation of the Aegean Sea is related to gravitational collapse of an orogenic topography (Dewey, 1988). The argument is that convergence during the Oligocene led to stacking of thrust nappes and thickening of continental crust. Subsequent thermal relaxation lead to gravitational spreading of the thicken zone. However, as noted by Thomson et al. (1999), there is no evidence for extremely thick continental crust in the Oligocene.

A third idea is that extension in the Aegean Sea is related to rollback of the African plate (Le Pichon and Angelier, 1979). Meulenkamp et al. (1988) showed that subduction had been well-established prior to the formation of the curvature of the Hellenic Arc. They argue that abundant Miocene deformation in Crete is coincident with the initiation of the rollback process. Although tomographic results indicate that the slab is continuous to great depths (in excess of 600 km) at the longitude of Crete, a continuous slab has not been successfully imaged to the west beneath mainland Greece (Spakman et al., 1988; Wortel and Spakman, 2000). These observations have to the hypothesis that the descending slab is torn beneath Greece, forcing the still continuous segment beneath Crete to support the entire load of the downgoing slab. The concentration of the slab-pull force beneath Crete then drives rapid rollback as the African lithosphere sinks deeper into the mantle (Wortel and Spakman, 2000). The importance of rollback in controlling the tectonics of the eastern Mediterranean has been emphasized in finite-element simulations of the region, both for the Aegean Sea (Meijer and Wortel, 1997; ten Veen and Meijer, 1998) and also on the scale of the entire Mediterranean (Jiménez-Munt et al., 2003). In fact, Jiménez-Munt et al. (2003) conclude that it is not possible to match the observed geodetic measurements in the Mediterranean without deep subduction and rapid rollback in the Hellenic Subduction Zone.

Geology of Crete

Most of the geologic units exposed on Crete were initially assembled by thrust imbrication during the Oligocene (Creutzberg and Seidel, 1975; Bonneau, 1984; Hall et al., 1984). That initial sequence was thinned dramatically by normal faulting, starting at ~15 Ma and continuing to the present (Thomson et al., 1998, 1999). Even so, the original nappe sequence formed during Oligocene thrusting is still readily apparent, as summarized in Figure 4. The initial assembly of the nappes is widely thought to represent the collision of a microcontinent (known as Adria or Apulia) with the European margin (e.g., Robertson et al., 1991). We have noted above however that this accretion of continental rocks may have been largely thin-skinned in that there is no break observed in the slab beneath Crete. Bonneau (1984) demonstrated that many units in Crete are present to the west and east into mainland Greece and Turkey, respectively. In contrast to other subduction zones, the structural style recorded in Crete is well organized, consisting of imbrication and folding of stratally coherent units. The regional extent of the stratigraphic units indicates accretion of large tectonostratigraphic units, as emphasized in the collisional interpretation of Robertson et al. (1991).

For a start, it is best to view the tectonic sequence in Crete as composed of three parts, which in increasing structural position are: 1) the *sub-detachment nappes*, 2) the *supra-detachment nappes*, and 3) *syn-extensional sediments*. The *sub-detachment nappes* are defined by a common high pressure-low temperature (HP-LT) metamorphic history, imposed in Oligocene and early Miocene time, shortly after accretion at the Hellenic subduction zone. The stratigraphic units involved are the Plattenkalk, Trypali, and Phyllite-Quartzite (defined below), which consist mainly of carbonates, schists, and quartzites that span in age from Late Carboniferous to Oligocene. This metamorphosed tectonic assemblage is bounded at its top by a one or more low-angle normal faults, which are collectively referred to as the Cretan detachment. It is also important to note that the stratigraphic units that make up the HP-LT footwall of the Cretan detachment are never found above that structural level.

The *supra-detachment nappes* are pre-Miocene accretionary units that lie above the Cretan detachment. Like the *sub-detachment nappes*, this structural assemblage also shows evidence of Oligocene accretion and thrust imbrication, but it contrast it always remained structurally shallow after

accretion, as indicated by the absence of Cenozoic HP-LT metamorphism. The stratigraphic units within the *supra-detachment* nappes include the Tripolitza, Pindos, and Uppermost units. They are dominated by carbonates, ophiolitic rocks, and some siliciclastic turbidites, with ages spanning from Triassic to Eocene.

The *syn-extensional sediments* correspond to a series of high level sedimentary basins, ranging from middle Miocene to present in age. These basins are intimately associated with late Cenozoic normal faulting in Crete. They provide a detailed record of unroofing of the Cretan nappe sequence, and the emergence of the mountainous subaerial topography associated with modern Crete.

We provide a more detailed review below of Cretan geology, with the discussion moving from the structurally lowest rocks below the Cretan detachment upward to the Late Cenozoic extensional basins. We follow the practice of dividing the nappes in Crete according to their internal stratigraphy and referring to each nappe using the name of the dominant stratigraphic unit in the nappe.

The sub-detachment nappes

Plattenkalk nappe. The structurally lowest unit exposed on Crete is the Plattenkalk (PK) nappe (called Ida nappe by some), which is made up entirely of the Plattenkalk Group. The Plattenkalk Group is commonly overturned, and large recumbent folds are found in some areas. Nonetheless, the unit maintains a coherent well defined internal stratigraphy consisting of stromatolitic dolomite, carbonate breccia, and a distinctive sequence of platy, well-bedded carbonates with chert interbeds (Epting et al., 1972; Seidel et al., 1982; Hall and Audley-Charles, 1983; Krahel et al., 1988). The oldest Plattenkalk consists of shallow marine carbonates with Permian fusulinids (Bonneau, 1984). Younger parts of the section consist of carbonates and cherts that record a transition to deeper water conditions. The thickness of the Plattenkalk is debated, with estimates ranging from 1 and 2.5 km (compare Epting et al., 1972 and Hall and Audley-Charles, 1983). The Plattenkalk is devoid of siliciclastic input except for a thin, 10 – 30 m layer of flysch at the top of the section. In central Crete, foraminifera from this flysch have been dated by Bizon et al. (1976) as Oligocene (29.3 to 28.3 Ma). The Plattenkalk is thought to represent the sedimentary cover of the southern margin of the Adria continental block (sometimes referred to as the Apulia block), which presently underlies the modern Adriatic Sea. The Adria block is being overridden to west by the

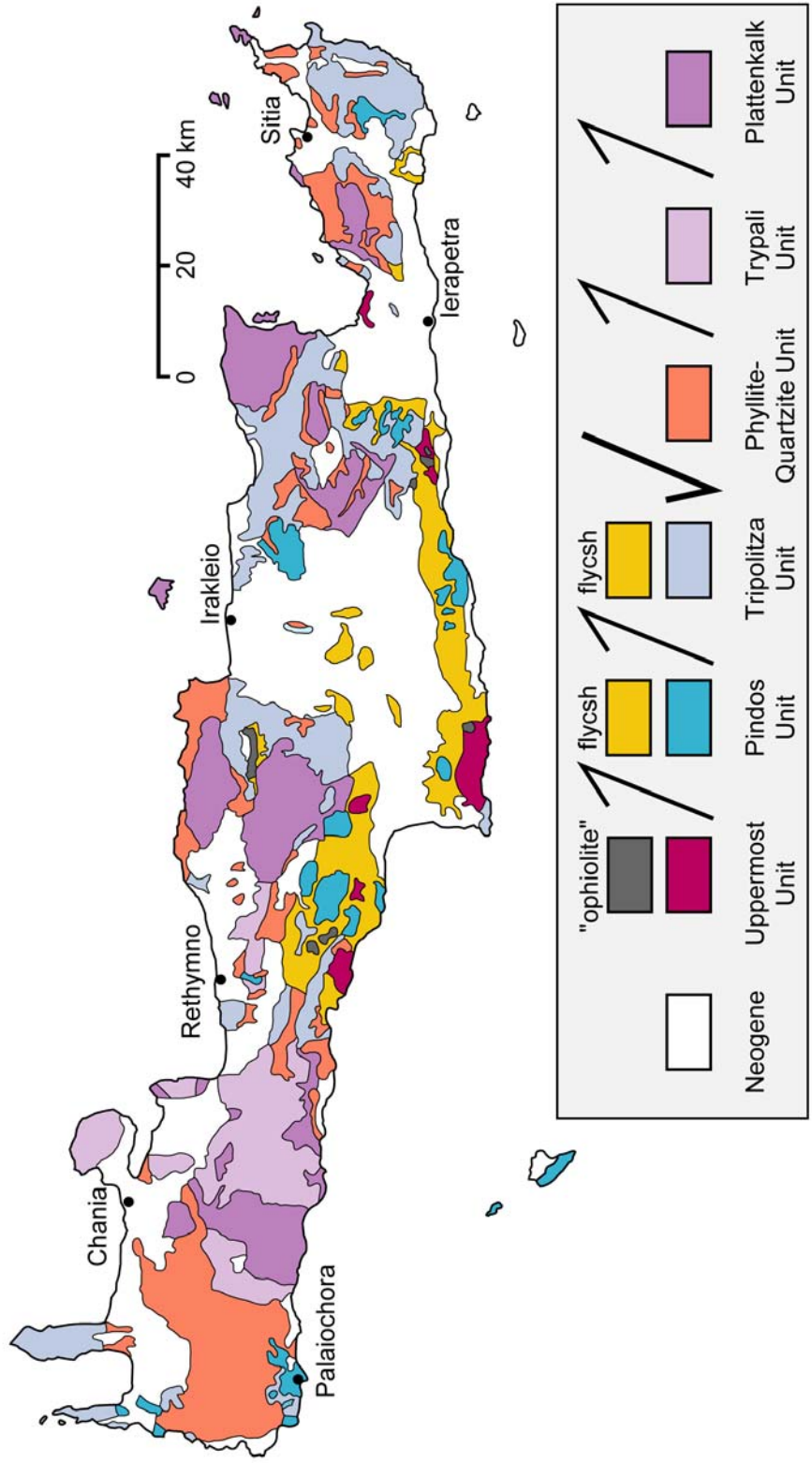


Figure 4 – Simplified geologic map of the basement units of Crete and their tectonostratigraphic relationships. After Creutzburg et al. (1977) and Thomson et al. (1999).

Apennine thrust belt and to the east by the Dinarides and Hellenides of the western Balkans and Greece. It also continues northward into the Alps, where it forms the highest structural unit in that Alpine collision zone.

Bonneau (1984) correlated the Plattenkalk Group with the carbonates of the Ionian Zone of western mainland Greece. The Plattenkalk is thought to be a fully allochthonous nappe. Although its base is not observed on Crete, the presence of an active subduction zone beneath Crete supports the interpretation that it is an accreted tectonic slice. Furthermore, to the east in Rhodes, the Plattenkalk is found overthrust on a more inboard carbonate platform, associated with the “pre-Apulian domain” from mainland Greece (Bonneau, 1984).

Trypali Nappe. In western Crete, the Plattenkalk nappe is structurally overlain by a small nappe made up entirely of the Trypali Formation, which consists of well bedded dolomite, with peloidal mudstone and detrital carbonate layers (Creutzburg and Seidel, 1975). The age of the Trypali has been reported as upper Triassic (Karakitsios, 1987) or lower Jurassic (Kopp and Ott, 1977). These sediments were deposited primarily in peritidal depositional environments, including sabhkas, tidal flats, hypersaline lagoons (Pomoni-Papaioannou and Karakitsios, 2002). A distinctive feature of the Trypali is an abundance of brecciated horizons, interpreted to represent submarine fault scarp breccias formed either during an Early Jurassic collapse of the Plattenkalk carbonate platform (Hall et al., 1984) or flexure of the foreland shortly before subduction of the Plattenkalk rocks in the Eocene to

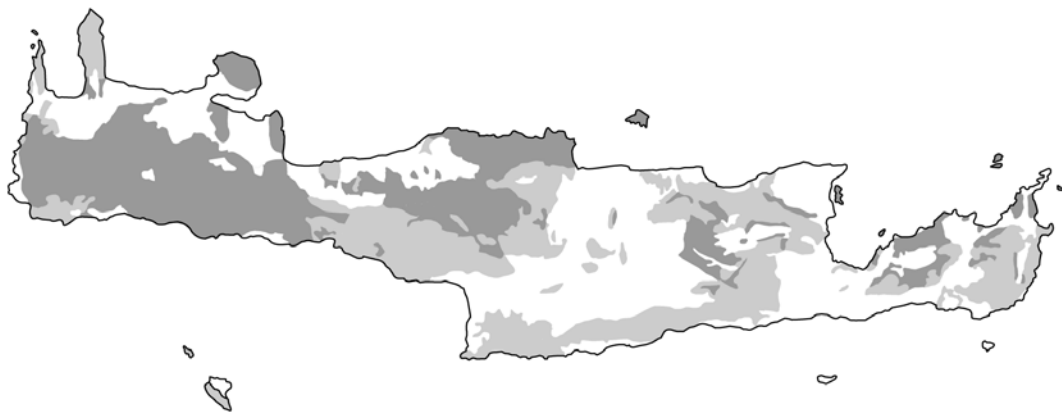
Oligocene (Thomson et al., 1999). Recently, Pomoni-Papaioannou and Karakitsios (2002) argue that the carbonate breccias of the Trypali are a recent feature formed by subaerial karst weathering during the late Quaternary.

Phyllite-Quartzite nappe. The next highest nappe is made up of the Phyllite-Quartzite (PQ) group. This nappe, which is widespread throughout Crete, is dominated by quartz-rich siliciclastic sediments, with minor limestone, gypsum, and volcanic rocks (Krahl et al., 1983). In eastern Crete, the sequence also includes slices of Hercynian basement (i.e. upper Paleozoic granitoid rocks) (Seidel et al., 1982). Conodonts define a Late Carboniferous to the Late Triassic age for the sedimentary rocks of the PQ (Krahl et al., 1983). The PQ is found throughout Crete and also in the southern Peloponnesus, where it occupies a similar structural position as on Crete (Theye and Seidel, 1991).

The supra-detachment nappes.

Tripolitza nappe. Above the Phyllite-Quartzite nappe are shallow-water carbonates of the Tripolitza nappe (also known as the Gavrovo nappe). The Tripolitza Group consists of shallow-water marine platform carbonates deposited between the Late Triassic and the middle Eocene. The Tripolitza and the overlying Pindos are correlated with the Triassic to upper Eocene carbonate platform sequence in mainland Greece (Creutzberg and

Figure 5 – Tectonic map showing the distribution of the sub-detachment nappes (in dark grey), which are characterized by Oligocene HP-LT metamorphism, and the supra-detachment nappes (in light grey), which were at shallow crustal levels during the Oligocene. The Neogene sediments are shown in white.



Seidel, 1975; Bonneau, 1984). At the base of the section are the Triassic Ravdoucha beds (Sannemann and Seidel, 1976), which are thought to be equivalent with the Tyros unit in mainland Greece. Higher up section, these shallow marine carbonates grade upwards into calciturbidites and eventually siliciclastic turbidites (Hall et al., 1984). From the Paleocene to Eocene, the Tripolitza is dominated by flysch deposits thought to be derived from the overriding Eurasian plate (Hall et al., 1984).

Pindos nappe. The Pindos nappe is the next in the structural sequence, overlying the Tripolitza in thrust contact. The Pindos is characterized by deep water sediments, including pelagic limestones, radiolarites, calciturbidites, and calc-breccias. These oldest parts of the unit are Late Triassic. Like the Plattenkalk and the Tripolitza, the Pindos unit is also capped by a Paleocene to Eocene turbidite sandstones and shales. Bonneau (1984) correlates the Pindos nappe on Crete with the Olonos-Pindos sequence exposed in mainland Greece. The Pindos is interpreted to be a deep basin that formed along the northern margin of the Adria microcontinent (Robertson et al., 1991).

Uppermost nappe. Tectonically above the rocks of the Pindos nappe are a diverse series of rocks that have been grouped into the Uppermost nappe. Although the rocks of the Uppermost nappe are older than the Jurassic, they contain no evidence for metamorphism in the Miocene (Seidel et al., 1976). A variety of rock types are present in the UM unit, including oceanic pillow basalts, gabbros, deep water marine sediments, amphibolites, schists, leucogranites, and an ophiolitic cap of mainly serpentinite (Seidel et al., 1976; Thomson et al., 1998). These rocks are thought to be derived from both the Jurassic to Cretaceous Pindos ocean, as well as the overriding European plate. Some authors (e.g., Bonneau, 1984; Creutzburg and Seidel, 1975) regard the various subunits of the Uppermost nappe as individual and mappable thrust nappes. In contrast, Hall et al. (1984) consider the various rocks (e.g., the Arvi, Miamou, Vatos, Asteroussia nappes) of the UM unit as large blocks in a “Blocky Flysch” that forms a major olistostrome in the upper part of the Pindos units.

Syn-Extensional Sediments

Sedimentation has occurred on Crete since the Miocene, likely beginning in the Serravallian. Meulenkamp et al. (1979) divided the sediments of Crete into six groups of formations, most of which are recognizable over all of Crete.

Prina Group. Dark limestone breccias and breccioconglomerates that generally have a well-cemented with a calcareous matrix. These sediments were deposited in non-marine to brackish or shallow marine environments.

Tefeli Group. “Non-consolidated” terrigenous clastic formations deposited on either the sediments of the Prina group or basement rocks. These sediments are predominantly conglomerates, sands, clays and were deposited in fresh-water, brackish, and marine environments

Vrysses Group. This group is composed of bioclastic, commonly reefal, limestones which constitute the lateral equivalent of alternations of laminated and homogenous, shallow-marine marls. In places, the marls contain gypsiferous intercalations. Vrysses group sediments are deposited on the Tefeli Group, pre-Neogene basement, and occasionally the Prina Group.

Hellenikon Group. Reddish, fluvio-lacustrine conglomerates, and occasional brackish and lagoonal deposits with gypsum. The Hellenikon conglomerates unconformably overlie the Vrysses Group, older Neogene strata, and locally pre-Neogene basement.

Finikia Group. Open marine marls and clays which commonly display laminated and locally siliceous interbeds. At many places the base of the Finikia Group is formed by a marl breccia. These sediments overlie the Hellenikon Group or the Vrysses Group.

Agia Galini Group. Coarse, generally reddish, non-marine conglomerates and sands that overlie (and are in part the lateral equivalent of) sediments of the Finikia Group. These represent the youngest Neogene rocks on Crete.

Undifferentiated Pleistocene. No formal subdivisions have been made. These marine terraces and continental deposits unconformably overlie Neogene or pre-Neogene rocks.

Metamorphic conditions

The sub-detachment nappes contain clear evidence of HP-LT metamorphism. The Plattenkalk nappe is composed mainly of lithologies unsuitable for thermobarometric work. Even so, a metabauxite horizon near the stratigraphic base of the Plattenkalk in central Crete contains lawsonite, magnesio-carpholite, pyrophyllite and diaspore. Theye et al. (1992) estimate maximum P-T conditions of about 1 GPa and 350°C. HP-LT metamorphic assemblages are more commonly found in the Phyllite-Quartzite nappe. Diagnostic minerals include carpholite, chloritoid, sudoite, phengite, aragonite, and jadeitic

pyroxene. Seidel et al. (1982) and Theye and Seidel (1991) estimate that exposed rocks in Crete show an increasing metamorphic grade from 0.6 GPa and 300°C in eastern Crete to 1.0 GPa and 400°C in western Crete.

This HP-LT metamorphism must be late Cenozoic in age since it involves Eocene and Oligocene turbidites in the Plattenkalk and Phyllite-Quartzite nappes. K-Ar and Ar-Ar dating of metamorphic white mica gives isotopic ages of 21 to 24 Ma (e.g., Jolivet et al., 1996). We interpret these ages as recording the time of growth of the metamorphic white mica, given that the maximum temperatures during metamorphism were probably less than the partial retention temperatures for white mica (~400°C).

The upper units lack evidence of Miocene HP/LT metamorphism. No high-pressure metamorphic minerals have been described from the Tripolitza and Pindos units, and the few metamorphic rocks in the UM unit are related to an older period of metamorphism and deformation (Seidel et al., 1976). Furthermore, Thomson et al. (1998) showed using apatite fission-track ages that most of the UM unit has been cooler than 120°C since 30 Ma.

The Cretan Detachment Fault and exhumation of the high pressure rocks

The juxtaposition of the HP-LT metamorphic rocks of the PQ unit with the unmetamorphosed rocks of the overlying Tripolitza unit has been taken as evidence for a significant normal fault, the Cretan Detachment (Fassoulas et al., 1994; Jolivet et al., 1996; Thomson et al., 1999). This fault, active in the Miocene, has cut out 20 to 25 km of structural section and is estimated to account for 85 to 90% of the exhumation experienced by the lower plate rocks (Thomson et al., 1999).

The geographic extent of the sub-detachment and supra-detachment units is shown in Figure 5. It is important to note that the original Cretan detachment has been extensively dismembered by younger normal faults. Nonetheless, the offsets on the younger faults are generally a kilometer or less, whereas the Cretan detachment seems to have cut out about 25 to 30 km of the Cretan nappe sequence. As a result, the map scale view shown in Figure 5 probably provides a reasonable view of the original distribution of the sub-detachment and supra-detachment units at the scale of Crete. Some authors have argued that the Cretan detachment formed with a consistent sense of motion, interpreted as top-S by Lister et al. (1984) or

top-N by Jolivet et al. (1996). With these interpretations in mind, note that the supra-detachment units are exposed both north and south of the island. Thus, there is no evidence on Crete of a top-N or top-S break-away zone. These features could be located offshore to the north or south. However, as an alternative, we suggest that the Cretan detachment might have formed by slip between a coaxially extending hangingwall (see Fassoulas et al., 1994; Fassoulas, 1999, for a similar coaxial interpretation). In this case, the sub-detachment nappes would have punched upward through the supra-detachment nappes, resulting in a variety of shear-sense directions on the detachment.

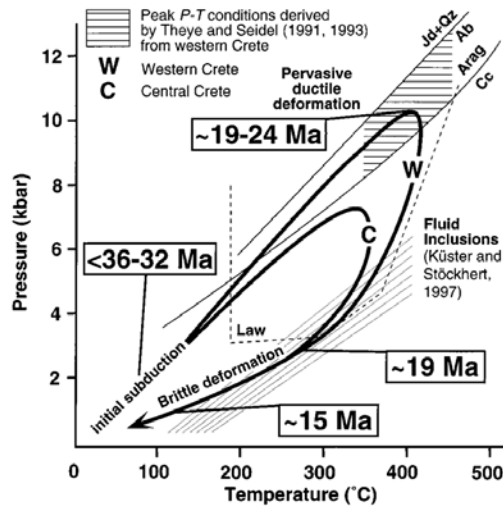
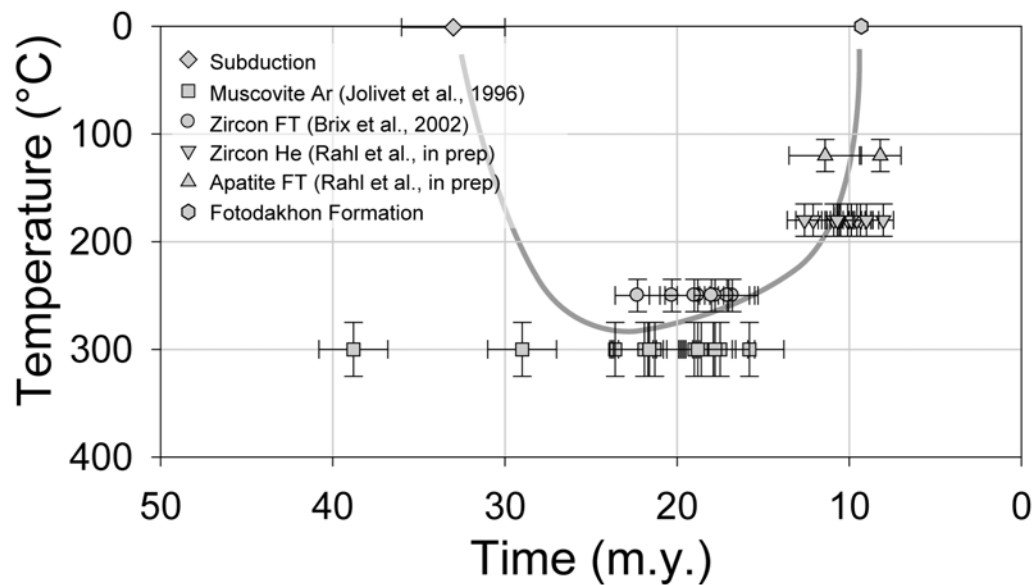


Figure 6 – Pressure-temperature-time (P-T-t) loop of the Phyllite-Quartzite unit of Crete, taken from Thomson et al. (1998). Ab–albite, Arag–argonite, Cc–calcite, Jd–jadeite, Qz–quartz, Law–lawsonite.

Thomson et al. (1999) provide a concise review of the P-T-t evolution of the sub-detachment rocks in Crete (Figure 6). The stratigraphic and thermochronologic constraints for the thermal evolution of these lower plate rocks are summarized in Figure 7. The Oligocene sediments in at the top of Plattenkalk Series demonstrate that the HP-LT tectonic units in Crete were subducted between 32 and 36 Ma. White mica Ar/Ar ages (Jolivet et al., 1996) record metamorphic growth and show peak P-T conditions were achieved between 21 and 24 Ma. Zircon fission-track data show cooling through 240°C at about 18 Ma (Brix et al., 2002). Zircon (U-Th)/He ages obtained by Peter Reiners (Rahl et al., in prep) show that the lower plate rocks cooled below 210°C around 10 Ma. Conglomeratic deposits that



contain clasts of PQ indicate the lower plate rocks were exposed at the surface by about 9 Ma (calcareous nannoplankton biozone NN 10) (Frydas and Keupp, 1996).

Extension in the Miocene led to the development of many sedimentary basins throughout Crete (e.g., Meulenkamp et al., 1979; Fortuin and Peters, 1983; ten Veen and Postma, 1999a). The oldest of these basins are thought to be Serravallian, but these units are typically unfossiliferous and difficult to date (Meulenkamp et al., 1979). After a period of subsidence during the Messinian, renewed block uplift led to the formation of widespread basins throughout the island (Meulenkamp et al., 1994).

Active tectonics of Crete

Abundant evidence exists that deformation is still ongoing in Crete. Modern seismicity reveals that Crete is experiencing extension, either radially (Angelier et al., 1982) or oriented east-west (e.g., Armijo et al., 1992). In addition to the continuing sedimentation in fault-bounded basins, many faults have been documented to offset young (Pliocene) sediments (ten Veen and Postma, 1999b; Fassoulas, 2001).

The geomorphology of Crete also indicates a young, actively uplifting landscape. For instance, Crete is famous for several large gorges, such as the Ha gorge in eastern Crete and the Samaria Gorge in western Crete. Messinian deposits are found throughout the island, sometimes at elevations in

Figure 7 – Time-temperature evolution of the Phyllite-Quartzite unit in central and western Crete. As noted by Thomson et al. (1999), paleogeographic constraints suggest that the Phyllite-quartzite unit entered the subduction zone 4-7 m.y. before deposition of the youngest Plattenkalk sediments. Seidel et al. (1982) and Jolivet et al. (1996) report K-Ar and Ar-Ar dates from white mica which likely record sub-closure temperature growth of mica during metamorphism at depth. Brix et al. (2002) report zircon fission track ages from central and western Crete, indicating cooling through about 240°C at about 18 Ma. New (U-Th)/He data (obtained in collaboration with Peter Reiners) indicate that cooling through about 180°C at about 10 Ma. Apatite fission-track ages (measured by Ray Donnelick) indicate cooling through 120°C also occurred at around 10 Ma. Finally, the observation of cobbles derived from the Phyllite-Quartzite in the sediments of the Fotokadhon Formation indicates that exhumation was complete by 9 Ma.

excess of 1000 m, indicating substantial uplift in the past 5 m.y. (Meulenkamp et al., 1994). In a detailed biostratigraphic study, Meulenkamp et al. (1994) demonstrated that maximum uplift in Pliocene to Recent times in central Crete was on the order of 2000 m. A ^{14}C study of wavecut terraces in western Crete demonstrated that uplift along the south coast is occurring at rates of up to 6 m/yr (Pirazzoli et al., 1982).

What is driving the modern uplift of Crete? Some workers, such as Meulenkamp et al. (1988), emphasize the role that recent reverse faults have played in Crete. In this model, Crete is thickening and uplifting due to horizontal contraction associated with the opening of the Aegean Sea. One attractive

alternative, first proposed by Le Pichon and Angelier (1979), is the underplating of subducted sediments beneath Crete, driving tectonic uplift. In addition to explaining the modern uplift, underplating is also consistent with seismic observations. In contrast to the Aegean Sea, the crust beneath Crete is typically 35 to 40 km thick, which is odd given that the late Cenozoic normal faulting should have substantially thinned the crust. Knapmeyer and Harjies (2000) identify low velocity rocks beneath Crete which they interpret as recently underplated sediments. Platt (1986) has argued that underplating can cause extension within a subduction wedge. We suspect that at least some of the extensional faulting observed in Crete is related to underplating, as was also proposed by Fassoulas et al. (1994).

We note that this view contrasts with models that regard Crete as a rigid, non-deforming backstop to the sediments of the Mediterranean Ridge (e.g., Kopf et al., 2003; Le Pichon et al., 2002). Although the rocks of Crete have higher seismic velocities than the unlithified sediments that compose the Mediterranean Ridge, the abundant evidence for active deformation on Crete make it clear that the island is not behaving in a rigid fashion, as required for the backstop interpretation. Thus, we prefer to view Crete as part of doubly vergent wedge (e.g., Willett, 1999), with the north side of Crete corresponding to the back side of this subduction wedge system.

Field itinerary

DAY 1: PSILORITIS MOUNTAINS, CENTRAL CRETE

We begin our trip in the Psiloritis (or Ida) Mountains of central Crete. This is an ideal area for a start because nearly all of important tectonostratigraphic units are accessible here. Additionally, the area contains excellent exposures of the Cretan detachment fault. The trip will depart from the village of Anogia, heading southward on the road up to the Nida Plateau.

Drive southward on the road from Anogia towards the Nida Plateau. Pull over at about 3.6 km from the village of Anogia along curve in road.

Day 1, Stop 1: Folded Plattenkalk Group

GPS: N 35.26891
E 024.88907

Elevation: 960 m

The deepest tectonostratigraphic unit exposed on Crete is the Plattenkalk Group, a Permian through Oligocene sequence of marbles, dolomites, and platy limestones that represents the sedimentary cover of the southern continental margin of the Adria microcontinent. The Group gets its name from a succession of platy limestones interbedded with chert horizons that make up the bulk of the section. These characteristic beds are can be seen at this locality, along the road from Anogia to the Psiloritis Mountains. Sparse undeformed fossils are present here, indicating that this part of the section was deposited in the Eocene (Bizon et al., 1976).

The Plattenkalk is characterized by tight to isoclinal mesoscale folds. The folds generally have east-west trending axes and are asymmetric with southward vergence. Here, the folded layers of the Plattenkalk generally maintain their thickness, indicating that the folds have developed primarily by flexural shear. Note that the cleavage-bedding intersection is the same on both the upright and overturned limbs, with a geometry consistent with the top-S vergence of the folds. The deformation in the limbs was apparently dominated by an outcrop-scale top-S shearing. Note that this deformation could be due to shear in a limb of larger fold, especially given the fact that folds with amplitudes of 500 m and greater are recognized in the Plattenkalk. Folding must have been younger than deposition of the youngest Plattenkalk (about 29 Ma) and the geometry of the folding is inconsistent with the extensional deformation associated with the Cretan detachment. Thus, this folding is commonly attributed to thrusting during late Oligocene accretion at the Hellenic subduction zone.

Generally the Plattenkalk rocks have a fine grained texture. However, recrystallization and an increase in grain size are commonly observed around cracks and at bedding contacts. These features may mark pathways for fluids that promoted metamorphic recrystallization. The bulk of the Plattenkalk must have been very dry, because there is little evidence for metamorphism in the platy limestone horizons despite the fact that these rocks reached maximum P-T conditions estimated at 0.8 GPa and 350°C (Theye and Seidel, 1991).

Continue to drive southward. As we climb in elevation, we will pass through the detachment and can be seen in the hillslopes to the east and west. The Plattenkalk is platy and more subdued, whereas the Tripolitza, which lies above the detachment, is exposed in rockier outcrops. Pull over along the side of the road after about 11.7 km from Stop 1.

Day 1, Stop 2: Oligocene “metaflysch” at the top of the Plattenkalk

GPS: N 35.22268
E 024.87861
Elevation: 1433 m

Following the road southward towards the Nida Plateau, we have moved stratigraphically upsection in the Plattenkalk. Here, we have reached the youngest rocks of the unit, and a facies change occurs from the well-bedded platy limestones with chert interbeds to an approximately 10 m-30 m thick package of metamorphosed turbidite sandstone, or metaflysch as it is locally called.

This metaflysch interval is the youngest part of the Plattenkalk, and is thought to represent an influx of siliciclastic continental material as the Plattenkalk carbonate platform approached the subduction zone. It also provides a useful age datum: Bizon et al. (1976) describe foraminifera (*Globigerina ampliapertura* zone) from this flysch that give an age of 29.3 to 28.3 Ma (Thomson et al., 1999).

The flysch has a weak, gently dipping bedding. A penetrative cleavage dips towards the north, as well as a spaced pressure-solution of cleavage that dips to the south. Minor boudinage in the carbonate layers indicate that the rocks have experience some degree of layer parallel extension.

Continue driving southward. After about roughly 2 km from stop 2, turn left on a small access road that leads to the church of Agios Fanourios. After driving a short distance you will see a driveway on the left side of the road which leads up to the church. Park in the area behind the church.

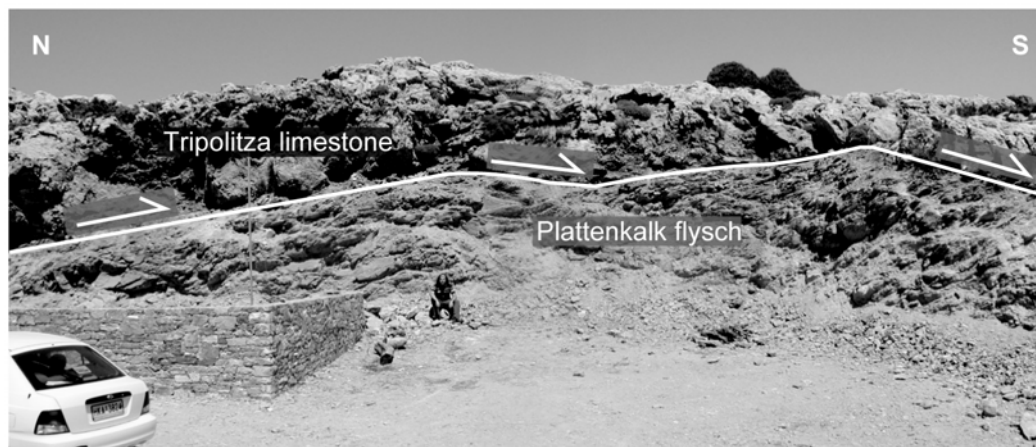
Day 1, Stop 3: The Cretan detachment at the church of Agios Fanourios

GPS: N 35.21425
E 024.87474
Elevation: 1395 m

In central Crete, the HP-LT metamorphosed Plattenkalk rocks are tectonically covered by unmetamorphosed limestones of the Tripolitza Group. These units are separated by the Cretan detachment, a significant Miocene normal fault, which has cut out approximately 20 km of structural section. An excellent exposure of the detachment can be found behind the church of Agios Fanourios (Figure 8). The fault zone is nearly horizontal at this location. The footwall is the Oligocene metaflysch of the Plattenkalk (same unit as the last stop), and the hangingwall, Tripolitza limestone.

The relatively weak metaflysch preserves brittle structures that indicate top-S motion at this location on the detachment fault. Brittle, low-angle shear zones dip into the detachment zone. A schistosity has developed parallel to the detachment, and a steep, northward dipping foliation is present in the fault gouge.

Figure 8 – Panorama of the Cretan detachment fault at the church of Agios Fanourios (Day 1, Stop 3). Here, the detachment fault is nearly horizontal. Limestones of the Tripolitza Group in the hanging wall rest upon the metamorphosed Plattenkalk flysch. Brittle shear-sense indicators developed in the flysch indicate top-S motion along this part of the detachment.



Return from the church to the main road. Continue southward towards the Nida Plateau. After about 2.3 km, pull over on the left side of the road at the GeoParks sign which provides a description of the local geology.

Day 1, Stop 4: Scenic view of the Cretan detachment and Nida's Plateau

GPS: N 35.21531
E 024.85879
Elevation: 1400 m

This location provides a scenic west-facing view of the Cretan detachment cutting through the landscape (Figure 9). The tallest peaks of the Psiloritis Mountains are visible to the west. Towards the southwest (left side of view), one can see the sharp low-angle contact between the Tripolitza nappe forming the mountainous peaks and the whitish, folded rocks of the Plattenkalk nappe below. This view is cut by a moderately dipping E-side-down normal fault which offsets the nappes and the detachment. Thus, the continuation of the detachment to the north (right side of view) can be found in a lower position near the base of the mountain.

This view also provides a nice view of a large recumbent fold in the Plattenkalk nappe. Although the upper limb of the unit is coherent along most of the top of the ridge, towards the northern end of the view it is folded into a large, asymmetric, south-verging fold. These folds are larger versions of what was observed at stop 1 today. They are also attributed to Oligocene accretion of the nappes.

In the foreground, one can see the Nida Plateau, which formed due to karstic weathering. Several sinkholes are recognized at the northern end of the plateau. The weathering has created an extensive cave system. In the nearby Tafkoura cave, east of the Nida Plateau, the tunnels reach a depth of 950 m below the surface. Furthermore, the Idaion Andro Cave is visible at the the scarp of the normal

fault on the northern side of the view. According to legend, this cave is where Zeus was raised (see below).

Continue along the road towards the Nida Plateau for 4.6 km. Park near the taverna at the end of the road.

Day 1, Stop 5: Well preserved top-S shear sense indicators (Optional Stop)

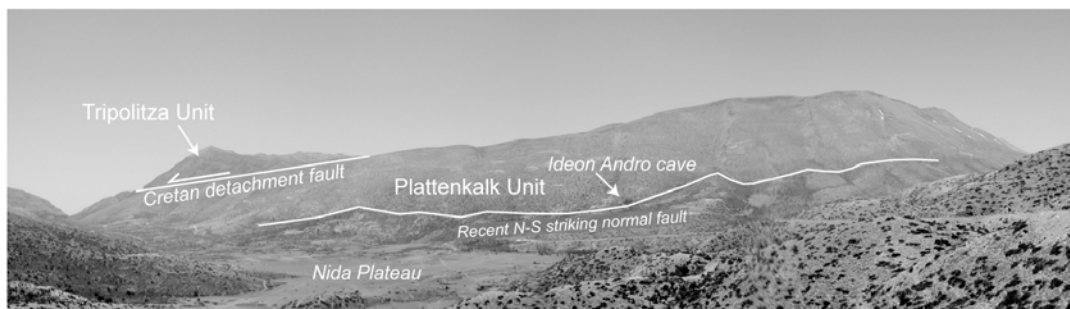
GPS: N 35.20684
E 024.83489
Elevation: 1301 m

Just to the west of the taverna at the end of the road is another excellent exposure of the detachment. Fault zone structures are well developed in a 1 to 2 m of zone directly below the stratigraphic break that marks the detachment surface. Again, the structures, including kink folds and S-C structures, are consistent with a top-S motion on the detachment fault.

Cultural note: the Idaion Andro cave

According to legend, Zeus, the most powerful god in the Greek pantheon, was raised here in the Idaion Andro cave in the Psiloritis Mountains. An oracle had prophesized to Zeus' father, Cronus, that he would one day be overthrown by one of his children. In an attempt to avoid this fate, Cronus ate each of his children as they were born so that they would not be able to harm him. His wife, Rhea, grew tired of this, and when she gave birth to her youngest son, Zeus, she secretly hid him at the Idaion Andro cave on Crete. She wrapped a stone in swaddling clothes and presented it to Cronus, who devoured it thinking that it was his son Zeus. Rhea entrusted

Figure 9 – A panorama of the Psiloritis Mountains above Nida's plateau in central Crete. See text (Day 1, Stop 4) for details.



Zeus to two Nymphs, who cared for the child. Armed youths known as the Kourites performed war dances around the young Zeus' cradle, to drown out the young child's cries. After he matured, Zeus was aided by Gaia (in some versions of the story Metis) and they were able to force Cronus to regurgitate Zeus' five brothers and sisters. Aided by his siblings, Zeus led a revolt that brought the end to the reign of the Titans.

Because of this myth, the Idaion Andro appears to have been the most important religious site in Crete during Minoan and Classical times. Archaeological excavations have revealed abundant, gold, silver, and stone artifacts, today displayed at the Irakleio Archaeological museum.

Return to Anogia, retracing the path that we took up into the mountains. Turn right to leave town, heading towards the east. After about 7 km you should reach the village of Gonies. A road on the right heads uphill out of town, exposing a nearly continuous section about rock. Pull over and stop after continuing up the road for another 1.5 km.

Day 1, Stop 6: The supra-detachment nappe sequence

GPS: N 35.28664

E 024.92609

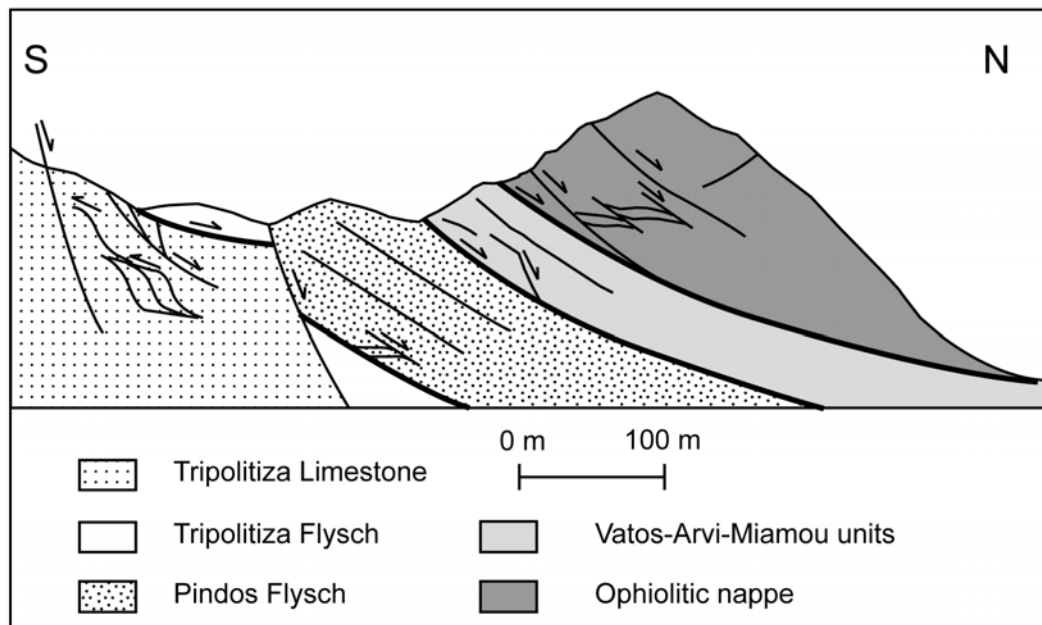
Elevation: 700 m

This road provides an opportunity to view a nearly continuous section that exposes all of the supra-detachment nappes. The traverse, walking downhill along the road, will move upsection through this structural sequence (Figure 10). Note that the section is cut by a number of minor normal faults, which have obscured original contact relations and thicknesses of the nappe units.

Walking down the road, the first outcrops are in limestones of the Tripolitza, which grade upward into Eocene sandstone and minor conglomerate of the Tripolitza flysch. These siliciclastic sediments are thought to represent continental-derived trench-fill deposits, which would have accumulated as the Tripolitza platform approached the Hellenic subduction zone (e.g., Hall et al., 1984). The Tripolitza flysch is much thicker than the flysch at the top of the Plattenkalk, and is typically on the order of 100 to 200 m. The entire thickness of the Tripolitza is about 700 m thick.

Continuing downhill (and upsection), we pass into the Jurassic red ribbon cherts of the Pindos nappe. These rocks contain radiolarites and chert with shale interbeds, and are exposed for about 150 m along the road. At this point, the ribbon cherts grade into the Pindos flysch, which here consists of about 75 m of siliciclastic sandstone, shale, and

Figure 10 – Schematic cross-section illustrating the structure of the upper-plate units in the area of the village of Gonies (after Fassoulas et al., 1994).



minor ribbon chert. This unit continues to the break in exposure near the windmill on the west side of the road.

After the windmill is a lithologically heterogeneous assemblage of rocks characteristic of the Uppermost nappe. The first exposures are calcareous-siliciclastic sediments of the Vatos member. These are followed by ultra-mafic rocks and the serpentinites. Continue walking towards the left at the split in the road. Exposures of serpentized ultramafic rocks and subordinate mafic rocks continue for the next several hundred meters where we reach the end of the traverse at the intersection with the main country road.

You will see evidence of several normal faults within this interval. Brittle shear-sense indicators show top-N motion on these faults. They are attributed to widespread late Cenozoic extensional faulting that followed Oligocene accretion of the nappes (Fassoulas, 1994).

Reset the odometer on your vehicle. From the downhill end of this section of road, begin to drive back towards Anogia. A little more than 1 km out of there will be an intersection with a sign pointing towards Aidonochori. Turn right, towards Aidonochori. Past the village of Aidonochori, turn left at the fork in the road onto a gravel road. On your left you should see Miocene sediments. Follow this road to the town of Chonos and take the left road out of town towards Drosia. The outcrops in this area belong to the Plattenkalk Group. After reaching Drosia, turn left out of town, towards the village of Doxarou. Within this village, turn right towards the Monastery of Timios Stabros. Drive up this road until the odometer reaches about 16.5 km and park along side the road.

Day 1, Stop 7: Ductile folds in the Plattenkalk

GPS: N 35.35980

E 024.84586

Elevation: 370 m

Along this north-south road, the Plattenkalk is folded into a large antiform. In its core are the lower parts of the Plattenkalk Group. These strata lack the distinctive platy limestones with chert interbeds that we saw at previous stops. As we walk southward down the slope, we walk upsection and back into the characteristic platy horizons. These strata host spectacular asymmetric, south-vergent folds in the Plattenkalk unit. These folds have roughly east-west trending hinge lines, plunging gently towards the east.

The rocks here apparently experienced higher temperatures than those exposed in the Psiloritis Mountains (Day 1, Stop 1). The hinges of the folds are not as sharp as at other localities. Additionally, changes in layer thickness indicate that the layers have deformed internally.

Drive back to Drosia and continue straight through the town. Turn left on the road that heads to the village of Aloides. Pass through the village. Continue until you have reached roughly about 9 km from the previous stop. Pull over along the side of the road.

Day 1, Stop 8: Stratigraphy of the Plattenkalk

GPS: N 35.37925

E 024.88744

Elevation: 400 m

Travelling north along the road towards the northern coast, we pass through a long, continuous section of the Plattenkalk. This section provides an opportunity to view much of the stratigraphy of the Plattenkalk Group (Figure 11). As we drive towards the north, we are actually driving downsection.

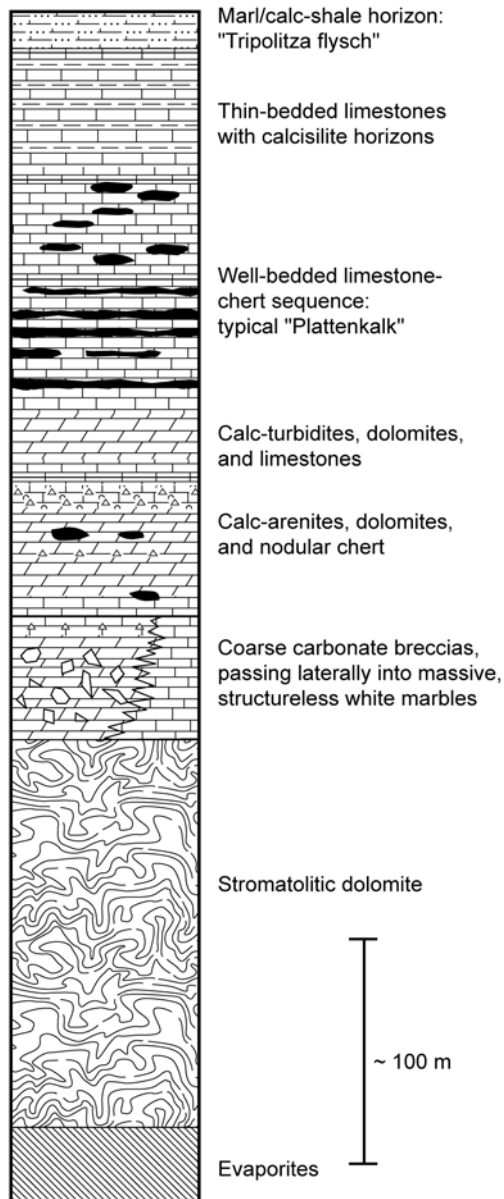
According to Epting et al. (1972), we are driving through the overturned limb of a large, asymmetric, south-vergent recumbent fold. This allows us to move through the entire stratigraphy of the Plattenkalk unit. We begin in the characteristic platy limestones, interbedded with chert deposits. As we move down section we encounter massive, white marbles thought to be lower Jurassic. Finally, lower down we move into the Triassic stromatolitic dolomite. Epting et al. (1972) estimate the entire thickness of the Plattenkalk to be about 2600 m.

An alternative has been proposed by Hall and Audley-Charles (1983), who argue that rather than consisting of a single recumbent fold, the structure of the Plattenkalk in this unit is dominated by many tight isoclinal folds that repeat many parts of the section. They also infer the existence of several faults that repeat parts of the section. They infer a maximum thickness for the Plattenkalk Group of about 1000 m.

The Plattenkalk Group sediments record the deepening of a sedimentary basin. The oldest part of the section is composed of shales and massive stromatolitic dolomite, indicating intertidal sedimentation. Upsection, these rocks give way to carbonate breccias and channelled and graded calcarenites. These are interpreted as recording rapid subsidence of the platform following the Triassic

(Hall and Audley-Charles, 1983). Farther upsection, the calcarenites grade into finer grained more basinal deposits of the characteristic limestone-chert sequence. Similar deposition is inferred to continue more or less continuously through the Mesozoic and early Cenozoic until the addition of the siliciclastic material of the Oligocene metaflysch at the top of the section.

Figure 11 – Stratigraphic column of the Plattenkalk Group in central Crete, simplified from Hall and Audley-Charles (1983).



At the field guide stop, we will see the beautiful Triassic stromatolites. Here, these structures clearly show that the sedimentary section is overturned.

Continue down the road. As we near a bend in the road at about 1 km from the last stop, pull over. The metabauxite horizon is not obvious and can be difficult to find in the outcrop. The GPS coordinates provide the best means of finding the location of the metabauxite.

Day 1, Stop 8: Metabauxite horizon within the Plattenkalk (Optional stop)

Elevation: 277 m
N 35.38589
E 024.89451

Continuing down the road, we encounter rocks from deeper in the stratigraphic section. At the base of the stromatolitic dolomite is a metabauxite horizon, significant because it provides the only thermobarometric data from within the Plattenkalk unit (Seidel et al., 1982; Theye et al., 1992; Theye and Seidel, 2001). The assemblage within the metabauxite contains magnesio-carpholite coexisting with pyrophyllite, sudoite, and disapore, leading to an estimate of metamorphic conditions at 0.8 GPa and 350°C.

Continue all the way down until you reach the National Road. Turn right (towards the east) and drive for 4 km. The road will have a large hairpin turn. When you are through this turn and begin travelling on the straight part of the road, immediately pull over on the right side of the road. Fossils of the "Fodele beds" can be seen in the outcrop along the side of the road.

Day 1, Stop 9: The "Fodele beds" of the Plattenkalk

GPS: N 35.38406
E 024.91861
Elevation: 115 m

Fossils of the "Fodele beds" of the Plattenkalk Group can be seen in the outcrop along the side of the road. These represent the oldest part of the Plattenkalk Group (Epting et al., 1972). Visible with the naked eye are abundant corals, bryozoa, and brachiopods (Kuss, 1980). Although these rocks experienced high-pressure metamorphic conditions, the fossils show no evidence for deformation or recrystallization. In-filling structures

in both corals and gastropods indicate the beds here are overturned.

Turn the car around and head westward. The rest of the stops take place in western Crete, and you will want to drive as far westward as possible before the next day. Both Rethymno and Chania are nice cities to visit for an evening.

DAY 2: PHYLLITE-QUARTZITE NAPPE IN WESTERN CRETE

Day two is focused on the Phyllite-Quartzite nappe, the structurally highest of the sub-detachment nappes. This unit has been intensely studied, including its metamorphic, thermochronologic, and deformational history. We will examine the Phyllite-Quartzite Group, discuss both the ductile and brittle deformation in the unit, and consider the processes responsible for its exhumation.

Drive westward along the National Road. As you drive westward past Chania, you will note many exposures of Miocene sediments in large roadcuts. Most of these sediments were deposited in the Pliocene. We will discuss the significance of the Miocene sediments in greater detail tomorrow

Eventually, you will pass over onto the western side of the Rodopos peninsula and to the north the Bay of Kissamos will be visible to the north. Look for the road that leads to Nopigia. Turn right, and follow the road to the coast. When it approaches the coast, this road turns towards the east. Follow it along the western side of the peninsula. The paved road will give way to gravel, and farther along there will be a white church. Drive past the church and over the small crest in the road. Park in the pulloff adjacent to several rocky outcrops

will be along the coast. This outcrop is about 2.4 km from the intersection with the National Road.

Day 2, Stop 1: The Phyllite-Quartzite unit

GPS: N 35.51943

E 023.72288

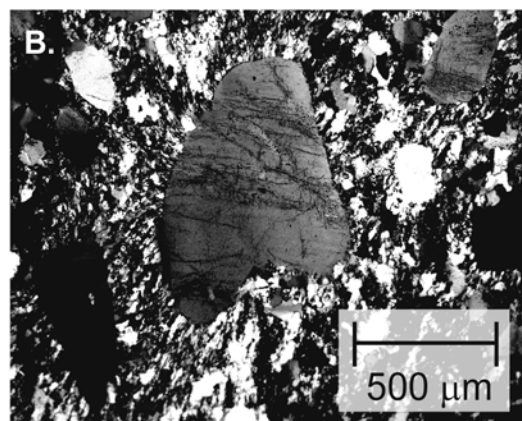
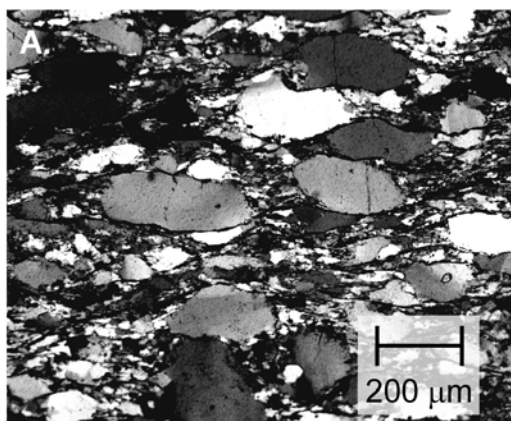
Elevation: 3 m

The excellent coastal exposures near the village of Nopigia provide a representative view of the Phyllite-Quartzite Group. The section consists of well bedded quartzites and phyllites, with beds ranging from a few cm up to about 1 m thick. Several carbonate-rich layers are also locally present.

Conodont fossils indicate that the Phyllite-Quartzite Group was deposited in a marine setting. The unit has yielded numerous conodont ages ranging from Carboniferous to Triassic (Krahl et al., 1983). The older parts of the section are dominated by siliciclastic sediments, although some carbonates, metavolcanics, and basic intrusives are also present. Carbonate beds become increasingly common in the younger part of the section, as we will see later in the field trip (Day 2, Stop 7).

Figure 12 – Photomicrographs illustrating the ductile microstructures common in the Phyllite-Quartzite nappe.

X, Y and Z indicate the principal strain directions for maximum extension, intermediate, and maximum shortening, respectively. a) A quartzite thin section illustrating the pressure-solution fabric, with Z oriented in the vertical, and X in the horizontal. The individual quartz grains show little evidence for internal deformation, such as recrystallization or undulose extinction. However, the grains have been shortened in the Z direction by dissolution and extended in the X direction by precipitation of fibrous overgrowths. b) The same sample as in (a) but here observed parallel to cleavage (X-Y section). Note there that the fibrous overgrowths are oriented in all directions indicating flattening deformation.



The Nopigia section shows consistent bedding attitudes throughout indicating that folds were not developed here. We will see mesoscale folds elsewhere in the Phyllite-Quartzite, but such folds are generally not as prevalent as in the Plattenkalk. Microstructural investigations show that these rocks have deformed by pressure solution processes (Figure 12; Schwarz and Stöckhert, 1996; Stöckhert et al., 1999), with material removed from surfaces of detrital quartz grains perpendicular to the shortening direction and redeposited as fibrous overgrowths in the extension direction within the foliation plane. With a hand lens, it is possible to see quartz grains truncated by the cleavage and “beard-shaped” overgrowths oriented parallel to cleavage. Our absolute strain measurements indicate a mass-loss (anisochoric) flattening deformation. Mass-loss is indicated by the fact that the shortening perpendicular to the cleavage is significantly greater than the total extension, which occurs in all directions parallel to the cleavage (Figure 13).

Thin quartz veins are common in this locality and are particularly abundant in the more competent quartzite beds. Schwarz and Stöckhert (1996) argue that these veins formed perpendicular to bedding during diagenesis. The oblique orientation of the views seen at present in the outcrop is attributed to shearing during the pressure-solution deformation.

A brittle shear zone can be seen in the outcrops at the base of the hillslope to the east of the shoreline. The outcrop is composed of thinly-bedded quartzite and phyllite which have been strongly distributed by brittle faulting. Fault zone structures are fairly symmetric, but there seems to be a predominance of top-S kinematic indicators.

Farther north along the coast, we can see Miocene carbonates capping the ridge. Farther up, a roughly N-S trending fault drops down rocks to the west. We can see the Miocene sediments near sea level, illustrating the significant offset along this fault. We will return to these north-south trending faults when we discuss the neotectonics of the island.

The Phyllite-Quartzite rocks are useful for studying the metamorphic, deformation, and exhumation history of the HP-LT rocks in the footwall of the Cretan detachment. Unlike the Plattenkalk Group, which is dominated by carbonates, the siliciclastic lithologies of the Phyllite-Quartzite contain assemblages amenable to metamorphic and thermochronologic study. The peak metamorphic conditions for the Phyllite-Quartzite throughout Crete are loosely bracketed to between 0.4-1.0 GPa, and 300-400°C by the co-

existence of pyrophyllite, low albite, and lawsonite, plus local aragonite and jadeite (Seidel et al., 1982; Theye and Seidel, 1991; Theye et al., 1992; Brix et al., 2002). Thermobarometry using phengite-muscovite solid solution and exchange between carpholite and chloritoid suggests that metamorphic grade increases systematically from eastern to western Crete, from 0.6 GPa and 300°C in the east to 1.0 GPa and 400°C in west. However, we note that these estimates assume an H_2O activity of unity. If the activity of water deviated from one, the maximum temperatures experienced by the Phyllite-Quartzite could be much less. The preservation of aragonite also suggests lower peak temperatures, since aragonite would have quickly inverted to calcite if the rocks entered the calcite stability field at temperatures significantly greater than about 235°C (Liu and Yund, 1993). K/Ar and Ar/Ar dates from syn-metamorphic white micas in the Phyllite-Quartzite suggest that metamorphism occurred between 21 and 24 Ma (Seidel et al., 1982; Jolivet et al., 1996).

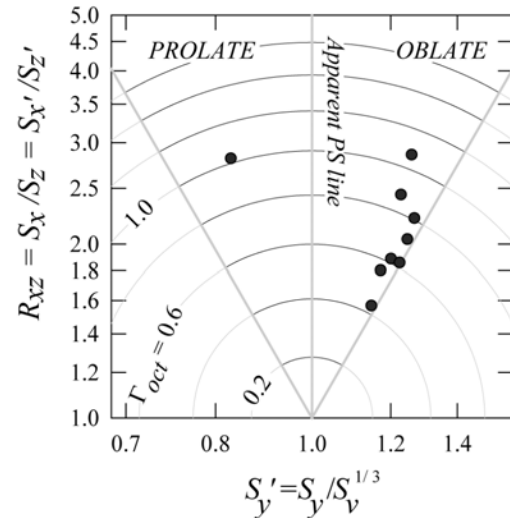


Figure 13 – Nadai plot showing absolute strain data from the Phyllite-Quartzite Unit in western Crete. All samples except one show a flattening strain. Uniaxial shortening was compensated by extension in all directions within the foliation plane.

Return to the National Road and travel westward towards Kissamos. Continue through to the town of Platanos. Follow the western coast road towards the village Sfinari. Pull over in a turn off approximately 20.4 km from the intersection of the National Road and the road to Nopigia.

Day 2, Stop 2: Brittle deformation and extension

GPS: N 35.44627

E 023.57976

Elevation: 228 m

Park at the pulloff just past one of the tight turns at the north end of the West Coast road. Numerous fresh cuts along this road provide an excellent view of both brittle and ductile structures in the Phyllite-Quartzite nappe. The pressure solution fabric is similar to that observed at the Nopigia locality (Day 2, Stop 1). Anisochoric flattening strains have been measured here as well. Quartz C-axis measurements show no development of a lattice preferred orientation, once again indicating that pressure solution was the dominant ductile mechanism (Schwarz and Stöckhert, 1996). Mesoscale isoclinal folds are locally present at several locations along this traverse.



Figure 14 – Photo of a well-developed normal-sense fault zone in western Crete (Day 2, Stop 2).

Throughout Crete, the Phyllite-Quartzite nappe is commonly cut by numerous brittle faults (Figure 14). This relationship is well displayed along the West Coast road. These fault zones generally lie at a high-angle to the more gently dipping pressure solution cleavage in the Phyllite-Quartzite. These faults typically have gouge zones that are several cm

to several m thick. Brittle kinematic indicators are abundant, usually as Reidel composite structures, with R shears and P foliations most common. We have used these indicators to assess the brittle strain in the Phyllite-Quartzite nappe. In western Crete, almost all of the measured faults show normal-sense offset, with a nearly equal distribution of top-N and top-S indicators (Figure 15). Thus, at the regional scale, the brittle deformation appears to involve N-S coaxial extension within the Phyllite-Quartzite nappe. This brittle deformation overprints the pressure-solution cleavage and likely post-dates the higher temperatures associated with metamorphism. Thus, brittle faulting may be synchronous with late Cenozoic development of the Cretan detachment.

Along this stretch of road, the Phyllite-Quartzite is everywhere overlain by a low-angle fault that separates the Phyllite-Quartzite unit from an unmetamorphosed Miocene(?) conglomerate composed of carbonate clasts (Figure 16). We interpret this low-angle fault as a high-level part of the Cretan detachment (figure 12 in Jolivet et al., 1996 shows a similar interpretation for this structure). At the next ridge down near the next turn in the road is a klippen of this Miocene(?) unit. The conglomerate there is matrix supported and consists mainly of rounded limestone cobbles, which look as if they may have been derived from Tripolitza Group.

A trail goes down from the main road, allowing a view of the fault zone beneath the klippen. Here, in the lowermost part of the fault zone is a breccia that contains angular clasts of Phyllite-Quartzite material. Farther up in the fault zone, the Phyllite-Quartzite clasts disappear, and only limestone clasts are present. We will see a similar Miocene(?) conglomerate later in the trip (the Topolia conglomerate, Day 3, Stop 2). An emerging idea is that these conglomeratic units represent syn-extensional basins that were deposited in the hangingwall of Cretan detachment (e.g., Seidel, 2003). The evidence at Topolia is more convincing, so we will wait until then to more fully develop this idea.

Continue following the western coast road southward to the village of Sfinari. From Sfinari, the next stop is 9.8 km, around a turn just past the village of Kampos.

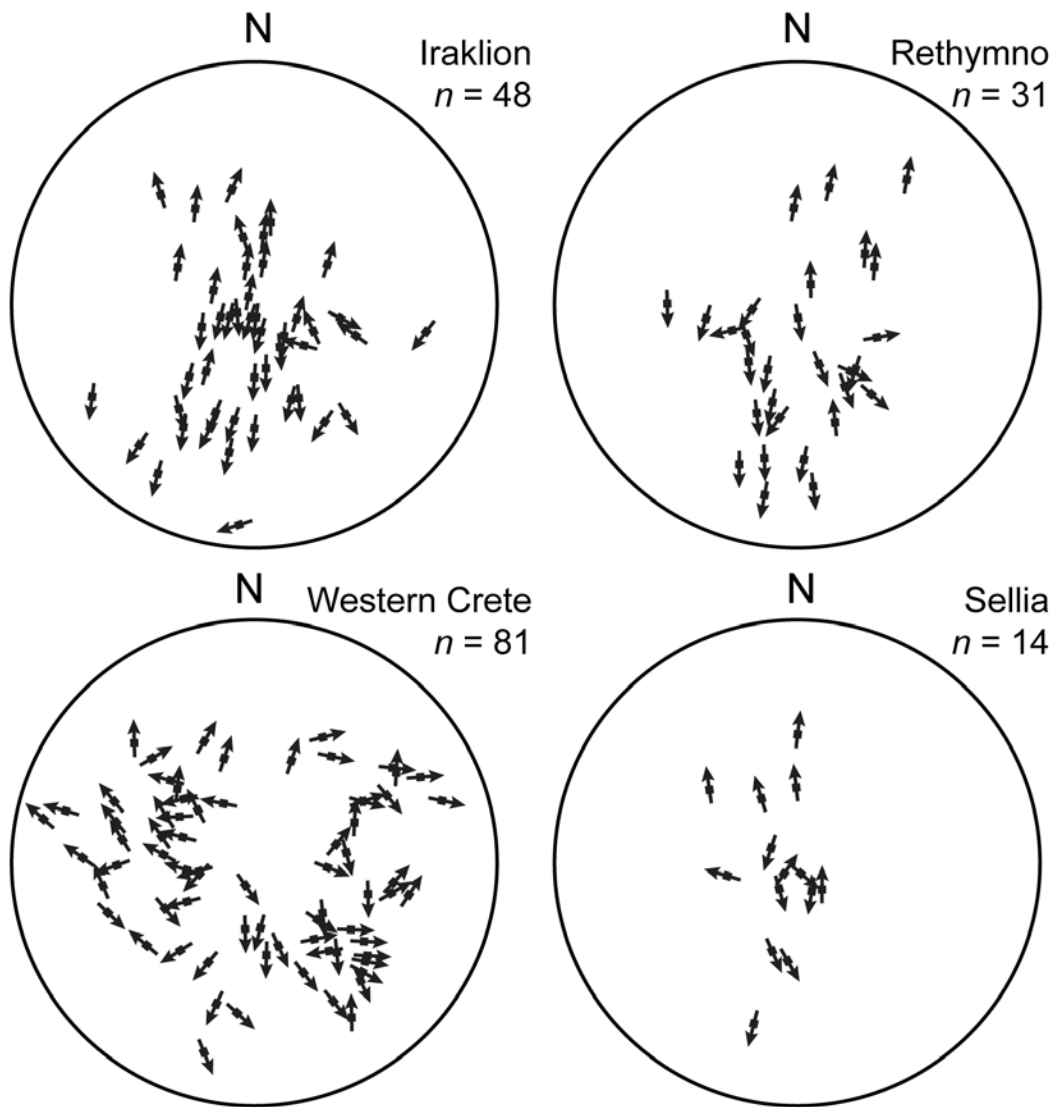


Figure 15 – Lower hemisphere stereonet showing slip-shear data for brittle fault zones within the Phyllite-Quartzite nappe in various areas in western and central Crete. Each square represents the pole to an individual fault plane. The arrow represents the direction of motion of the footwall block relative to a fixed hanging wall block. (Arrows pointing away from the center of stereonet indicate normal-sense slip.) Fault planes show a variety of orientations and nearly all record normal-sense slip.

Day 2, Stop 3: Folds in the Phyllite-quartzite nappe

GPS: N 35.38727

E 023.56462

Elevation: 313 m

Mesoscale folds are locally found in the Phyllite-Quartzite nappe (Greiling, 1982; Stöckhert et al., 1999; Thomson et al., 1999). Stöckhert et al. (1999) have recognized two sets of folds in western Crete, both of which can be observed at this location

(Figure 17). The older set consists of recumbent isoclinal folds, with fold axes trending NNE-SSW. These folds are associated with a nearly horizontal cleavage and a NNE stretching lineation. The second set of folds have axes that trend approximately E-W. The limbs of these folds are commonly thinned, and more competent quartz-rich horizons are boudinaged. Like the folds in the Plattenkalk, these are also attributed to deformation associated with Oligocene accretion of the Cretan nappes. In our experience, however, folds are not as common in the Phyllite-Quartzite nappe as they are in the Plattenkalk, and the fold geometry observed here may be a local feature. The curve in the road here provides a good opportunity to look at these folds in different orientations.

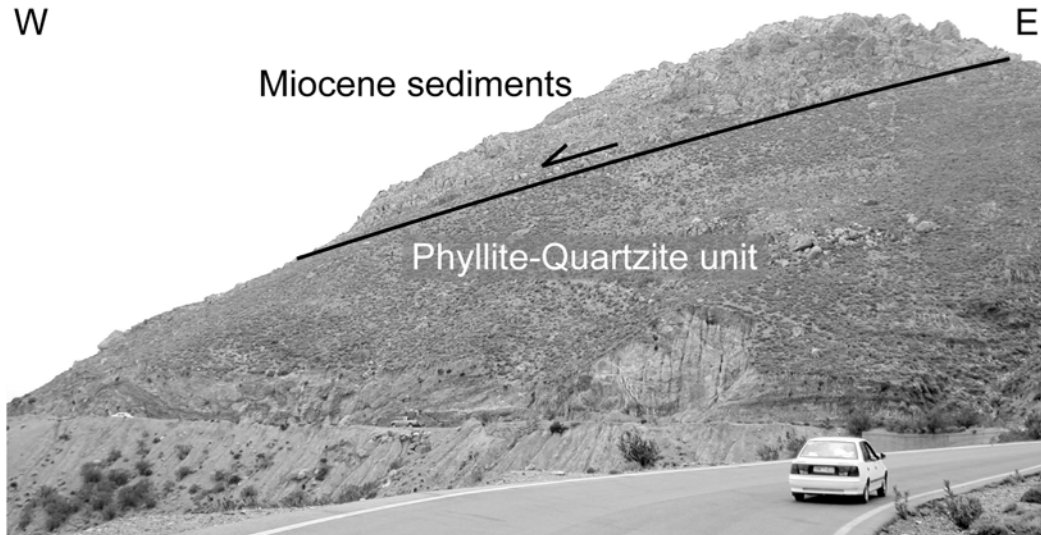


Figure 16 – Photo looking northward, showing the Cretan detachment fault with Miocene sediments in the hangingwall and Phyllite-Quartzite nappe in the footwall (Day 2, Stop 2).

Continue driving southward. Approximately 7.6 km from Stop 3, the road will bend back towards the interior of the island, with a turnoff on the south side of the road. Park your vehicle in the turnoff, and walk along the access road that leads downhill. The evaporites for stop 4 are exposed along the side of this access road.

Day 2, Stop 4: Triassic evaporites in the Phyllite-Quartzite nappe (Optional stop)

GPS: N 35.35306

E 023.56938

Elevation: 510 m

Triassic evaporites are present locally in the Phyllite-Quartzite nappe. Here, the evaporites are recrystallized and deformed in a manner similar to the rest of the Phyllite-Quartzite rocks (Fassoulas, 2000). Evaporites of this age are common in Europe and are attributed to rift-related breakup of Pangea.

These Triassic evaporites are near the stratigraphic top of the Phyllite-Quartzite Group. As we will see at a later stop (Day 2, Stop 6), the phyllites and quartzites that characterize the lower part of the section, which is mainly Permian in age, grade upwards to more carbonaceous and evaporitic units.

Thomson et al. (1999) suggest that the Triassic evaporites served as a weak horizon that allowed the Phyllite-Quartzite nappe to detach from its overlying cover and to be deeply subducted. Several workers (Bonneau, 1984; Hall and Audley-Charles, 1983) have speculated that the Phyllite-Quartzite Group was originally overlain depositionally by the Tripolitza Group, which is one of the supra-detachment nappes. They note that there is no known stratigraphic overlap in these two units, with the youngest Phyllite-Quartzite being Triassic in age and the oldest Tripolitza being late Triassic.

Continue along the road, passing through the village of Kefali. Join the road that connects Chania and Elaphonisi, and drive towards Chania. Take the road towards the right that leads to the village of Strovles. Past this village, take the right fork in the road towards the village of Arhondiko and Voutas. After you reach Voutas, continue south towards Pal the road towards Palaiochora. Pull off in the hairpin turn approximately 3.1 km south of Voutas.

Day 2, Stop 5: The detachment fault near Voutas

GPS: N 35.27345

E 023.66350

Elevation: 277 m

Near the hairpin turn in the road south of Voutas is an excellent exposure of the Cretan detachment. The approximately 20 m wide fault zone

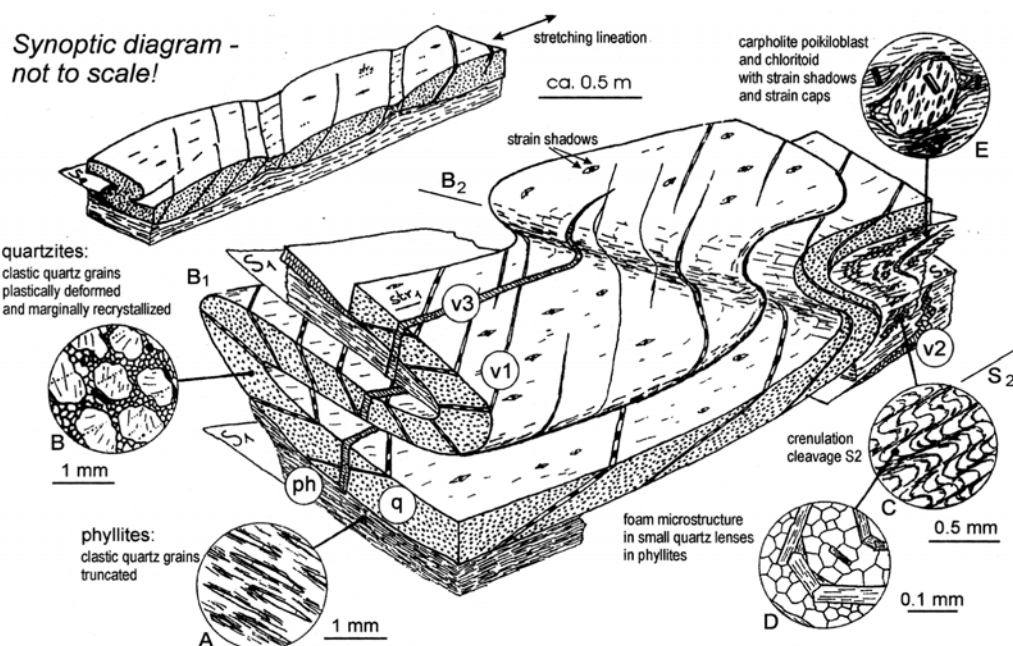


Figure 17 – Synoptic diagram illustrating deformation structures within the siliciclastic rocks of the Phyllite-Quartzite nappe of western Crete, from Stöckhert et al. (1999). B1, S1, str1 denote first-generation fold axis, schistosity, and stretching lineation. B2, S2 denote second generation fold axis and schistosity (crenulation cleavage in phyllites). Ph: phyllites; q: quartzites; v1: veins formed along original joints normal to bedding in sandstone; during deformation these veins were rotated and stretched, with boudinage of the original carbonate filling (black); quartz (white) was precipitated between the boudins; v2: veins formed during HP-LT metamorphism along S1 in phyllites, folded (B2) during progressive deformation and recrystallized; v3: later formed veins crosscutting all earlier structures, undeformed and with original microstructure preserved. Insets A to E reveal characteristic microstructures and their respective positions.

dips towards the south and contains several shear sense indicators consistent with top-S motion.

On the western side of the road, just prior to the sharp turn, a small access road runs downhill. Down this road are several exposures of the supra-detachment Tripolitza nappe. Note tectonic breccia which contains angular clasts of the Tripolitza.

An important question regards the motion along the major detachment fault. Jolivet et al. (1996) argue that extensional deformation in the Phyllite-Quartzite unit is consistently asymmetric, with a dominant top-N sense of shear, whereas Fassoulas et al. (1994) argue for a symmetrical deformation in central Crete. As discussed before,

our microstructural evidence suggests that deformation within the Phyllite-Quartzite unit is generally coaxial, at least on the micro-scale. Although Jolivet et al. (1996) admit that the deformation is “partly coaxial” on a small scale, they maintain that mesoscale observations are consistently top-N. However, the detachment here south of Voutas is clearly top-S. Recall that, when visible in the Psiloritis Mountains, the detachment is also top-S.

We agree with Jolivet et al. (1996) that there are many localities in which deformation appears to have a top-N shear sense. However, we do not believe that the deformation is systematically asymmetric. Therefore, we favor a model for the exhumation of the lower plate rocks along two separate detachment faults, one dipping northwards and the other dipping towards the south.

Return to Voutas, and take the road westward out of town towards Sklavapoula. The road to Sklavapoula is very windy and hilly. Upon reaching Sklavapoula, turn left towards Agioi Theodoroi.

Day 2, Stop 6: Deformed marbles within the Phyllite-Quartzite nappe

GPS: N 35.29030

E 023.61844

Elevation: 607 m

Between the villages of Sklavapoula and Agioi Theodoroi is a long roadcut that exposes Triassic carbonates of the Phyllite-Quartzite Group. These beds, known as the Kalamos formation, alternate between marbles and metapelites (Theye and Seidel, 1993). The petrology of these rocks has been described in detail by Theye and Seidel (1993). They report that limestones are lawsonite-bearing aragonite marbles, which in many cases have been partly or completely inverted to calcite marbles. The presence of unretrograded aragonite provides an important constraint on the P-T history of these rocks. Work by Carlson and Rosenfeld (1981) and Liu and Yund (1993) indicates that aragonite-bearing rocks must enter the calcite stability field at temperatures less than about 235°C for aragonite to survive transport to the surface without regression. This upper limit assumes very fast exhumation rates (~10 km/m.y.). Theye and Seidel (1993) have acknowledged that preservation of aragonite would imply much lower temperatures than the 350°C estimates from their thermobarometry study. This discrepancy is being studied at present by Matthew Manon and Eric Essene of the University of Michigan.

This outcrop is described by Stöckhert et al. (1999). The carbonate sequence is deformed into 100 m to km-scale asymmetric recumbent near-isoclinal folds. The long limbs of the folds have been shortened perpendicular to bedding, creating boudinage structures. Aragonite layers commonly contain veins filled with calcite and quartz oriented nearly perpendicular to bedding, some of which are slightly buckled. The veins are more competent than the aragonite layer, which lead to a characteristic “dog bone” structure during deformation. Stöckhert et al. (1999) describe fossils from these layers that are completely undistorted, demonstrating that even for the carbonate rocks in the Phyllite-Quartzite, the dominant deformation mechanism is pressure solution.

Continue following the road to Palaiochora for the evening.

DAY 3: BASIN DEVELOPMENT IN WESTERN CRETE

In the first two days of the field trip, we have emphasized the older history of Crete, focusing on the sub-detachment and supra-detachment nappes. Today we will examine the sedimentary record of exhumation.

Extensional deformation and exhumation started at about 15 to 10 Ma. Coincident with this activity was the development of many local sedimentary basins. The stratigraphic nomenclature for these sequences remains complex, particularly in western Crete. Meulenkamp et al. (1979) has defined six lithostratigraphic units that seem to work well in representing syn-extensional basins throughout Crete. As such, we will use the divisions and terminology of Meulenkamp et al. (1979) in our discussion here, while noting local names where appropriate. Figure 18 shows a composite stratigraphic section for the sediments of western Crete, adapted from Meulenkamp et al. (1979).

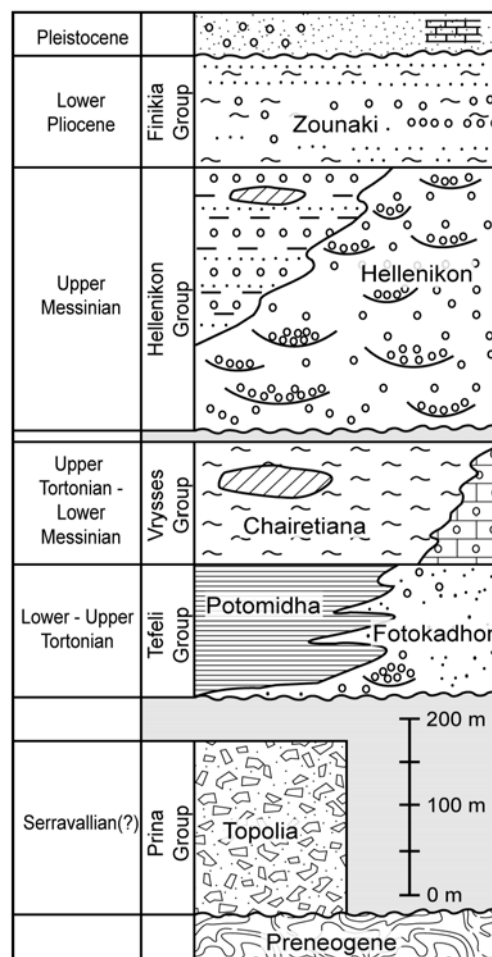


Figure 18 – Composite stratigraphic column for the sediments of western Crete, adapted from Meulenkamp et al. (1979).

Drive westward from Palaiochora along the coast road for about 2 km. Pull over in small lot on the left hand-side of the road.

Day 3, Stop 1: Wave-cut terraces along the southern coast at Koundoura beach

GPS: N 35.23722
E 023.66877
Elevation: 2 m

Recent uplift of the southwest coast of Crete can be observed west of Palaiochora at Koundoura beach. You can see there a flight of uplifted wave-cut terraces exposed north of the road. Wave-cut terraces are commonly formed as a result of sea level fluctuations (eustasy) and/or elastic deformation related to the earthquake cycle. The terrace is cut when the rate of uplift of the land matches the rate of rise in sea level. In this case, the surf zone can work for some time to cut the platform. Preservation of the terrace requires that the platform be rapidly uplifted above surf zone. This typically occurs due to coseismic uplift, but a rapid fall in eustatic sea level could produce the same result.

The terraces north of the road locally show a recessed notch that marks the transition from wave-cut platform to the seacliff that use to back the platform. Cemented sand and gravel are locally preserved on the wave-cut surface, and represent remnants of beach deposits. The older terraces would be located higher on the hillslope. These terraces have been eroded away, but at this location they can sometimes be detected by the presence of caves that hosted fresh water streams that drained to sea level.

The youngest uplifted terrace is located on the beach to the south of the road. Look for a low rocky pavement of beach rock exposed just above the modern shoreline. This pavement is composed of cemented conglomerate and sand, and contains cobbles derived from the adjacent Cretan nappes (e.g., Phyllite-Quartzite and Tripolitza). A broad wave-cut platform has been cut into the upper part of this beach rock unit.

Pirazzoli et al. (1982) have dated similar wave-cut terraces in western Crete the using ^{14}C method, and they have demonstrated that uplift is occurring at rates of up to 6 m/yr for over the past 2000 years. We will focus on the modern, continuing uplift in Crete tomorrow.

Return to Palaiochora and follow the main road north out of town. At the village of Plemeniana, turn left on the road towards Dris and Strovles. Continue northward through Strovles. Turn right

(towards Chania) when the road reaches an intersection. Drive north for roughly 4.5 km. The valley narrows to the north. Stop just after the memorial on the east side of the road at the bend. Walk south towards the tectonic contact between the Phyllite-Quartzite and the overlying breccia. Exercise extreme caution at this locality. The outcrops are exposed along a windy road, so be careful of traffic.

Day 3, Stop 2: The onset of Miocene sedimentation

GPS: N 35.41018
E 023.68315
Elevation: 301 m

Here, we can see the oldest Neogene sediments in western Crete exposed in the cliffs of the Topolia gorge. The Miocene(?) Topolia conglomerate is a poorly sorted, matrix supported conglomerate containing subangular to rounded clasts ranging in size from granules to cobbles. A variety of clasts-types are present, including grey to black limestones, dolomites, recrystallized limestones, radiolarites, calcarenites, and micritic limestone (Seidel, 2003). The conglomerate has been interpreted as an alluvial fan deposit, with debris flow and waterlaid deposits derived from a catchment to the south (Seidel, 2003). The unit is estimated to be about 500 m thick. The coarsest deposits are located in the southern part of the basin, near its base. Higher up-section, the sediments are predominantly debris-flows in the proximal and medial parts of the complex, accounting for 70 to 80% of the alluvial fan.

Importantly, there are no clasts of the lower plate units within this basal unit. The limestones, dolomites, and recrystallized limestones appear to have been derived from the Tripolitza nappe (Figure 19). In fact, Seidel (2003) describes fossils within some of the clasts that correspond with those found in the Tripolitza unit. Other types of cobbles, including radiolarites, calcarenites, and micritic limestone are consistent with derivation from the Pindos unit. However, despite its current close proximity with the Phyllite-Quartzite unit, the Topolia conglomerate does not contain cobbles that may have been derived from this unit. This suggests that during deposition of the Topolia conglomerates, the lower plate rocks of the Phyllite-Quartzite unit were not exposed at the surface. However, the conglomerates were later faulted into directed contact with the Phyllite-Quartzite bedrock.

The age of the Topolia conglomerates is not clear, because a lack of fossils within the

Figure 19 – Photo of the Topolia breccia in western Crete. The Topolia belongs to the Prina Group, which are the oldest Miocene sediments in Crete. Clasts within the Topolia appear to be exclusively derived from the supra-detachment nappes.



complex itself prevents direct dating of the sediments. Sediments stratigraphically above the Topolia have been dated as about 9 Ma (Frydas and Keupp, 1996). Deposition of the alluvial fan may be related to extension associated with the Cretan detachment fault. Exhumation on the Cretan detachment fault is envisaged to have begun between 20 and 15 Ma, suggesting that the Topolia conglomerate was deposited sometime between 20 and 10 Ma (Seidel, 2003). The steep dip of the unit here implies significant rotation caused by uplift since the deposition of these sediments.

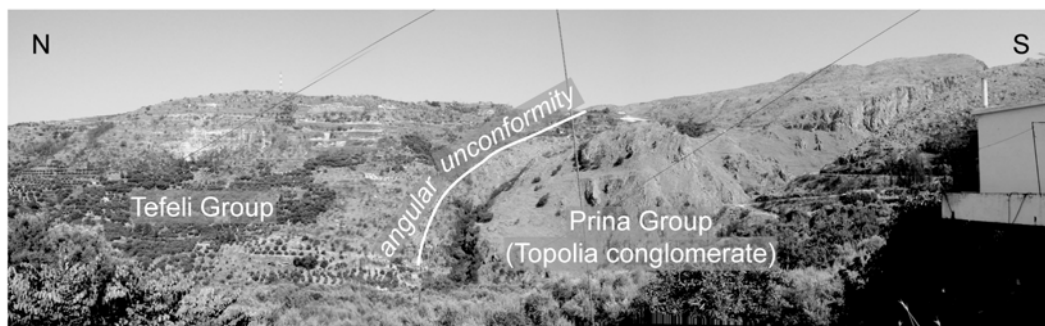
Following the road northward, the valley widens as we exit the gorge. Stop near the northern end of the next village, Voulgaro, for a view of the Topolia Conglomerate and the overlying Miocene sediments.

Day 3, Stop 3: View of early Miocene sedimentation

Looking towards the east, we have a nice view of the Miocene basin-fill sediments in Crete. Above the Topolia conglomerates are the horizontal sediments of the Tefeli Group (Frydas and Keupp, 1996). The oldest sediments of the Tefeli Group belong to the Fotokadhon Formation and record the initial phases of marine deposition in a shallow water environment. The Fotokadhon (also referred to by some authors as the Roka) is interfingered and overlain by the clays of the Potamida Formation, which contains calcareous nannoplankton indicative of a Tortonian age (NN10 – *Discoaster calcaris*). This demonstrates that in western Crete, the Fotokadhon formation has a depositional age of about 9 Ma.

A sedimentary hiatus followed deposition of the Topolia, as indicated by the presence of an angular unconformity between the Topolia and the overlying sediments of the Miocene Tefeli Group (Figure 20). The Fotokadhon grades both laterally and upsection into the Potamida formation. Above these formations are the well-bedded, sandy marls and interbedded clays of the Chairitiana Formation. The lower part of the Chairitiana of the Vrysses group corresponds to calcareous nannoplankton subzone NN 11a (latest Tortonian), while its upper

Figure 20 – Panorama looking eastward from the village of Voulgaro, showing the lower Miocene supra-detachment sediments in western Crete. An angular unconformity exists between the Topolia breccia and the overlying sediments of the Tefeli group.



part belongs stratigraphically to the subzone NN 11b (*Amaurolithus delicatus*), indicating an early Messinian age (Frydas and Keupp, 1996). Gypsum deposits are commonly found contained within the Chairetiana Formation. The consolidated marls of the Chairetiana are distinctive and thus serve as a useful marker horizon throughout northwestern Crete.

Continue driving north until you reach the National road at the eastern side of the town of Kissamos. Turn right and drive towards the east. Pass the road that heads off towards Nopigia (Day 2, Stop 1) and continue eastward for several km up onto the Rodopos Peninsula. About 10.5 km from Kissamos you will see a low concrete building on the right hand side of the road with "NISSAN" written on it. Turn into the large pullout near this building. There is a large outcrop of Phyllite-Quartzite rocks on the left-hand side of the road. Carefully cross the road and walk towards the east for about 100 m.

Day 3, Stop 4: Depositional contact on the Phyllite-Quartzite unit

GPS: N 35.5314
E 023.765

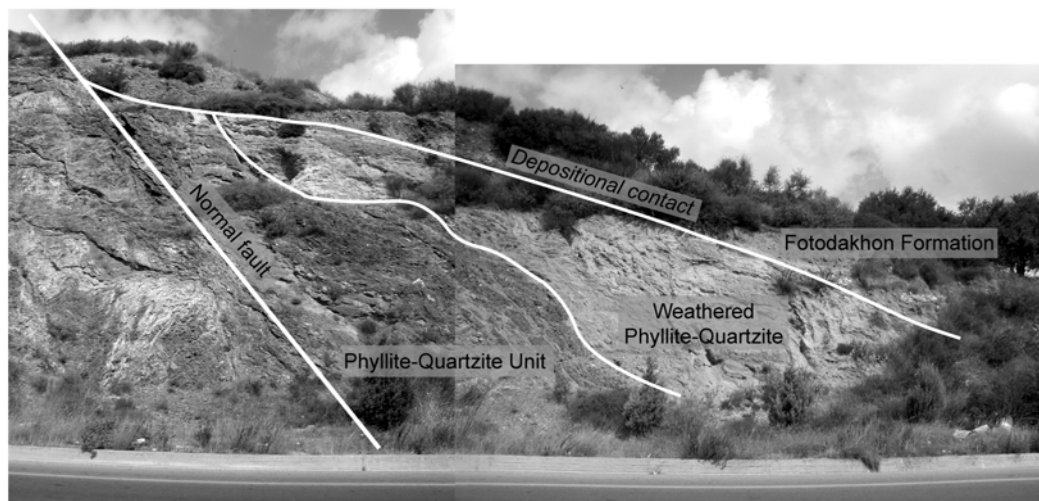
Here, sediments equivalent to the Fotokadhon clearly lie in depositional contact with the Phyllite-Quartzite unit. The contact between the bedrock of the Phyllite-Quartzite and the Miocene sediments is not a flat surface, but rather has an undulating, hummocky nature indicating that it was at the surface at the time. Furthermore, soil horizons also illustrate the depositional nature of the contact between the Phyllite-Quartzite and overlying

conglomerate. Walking eastward along the road, we can see that this conglomerate is clearly overlain by the Fotokadhon Formation sediments and the Chairetiana Formation. This clearly illustrates that the Phyllite-Quartzite was exposed at the surface prior to the NN10 Nannozone at about 9 Ma (Frydas and Keupp, 1996).

Walk eastward on the road. Bedding in the Miocene sediments is initially steep but becomes more gentle away from the contact. Continuing eastward for several hundred meters, there is a break in the section across the bridge. After this, the section is still conglomerates, but they fine upward into buff-colored thinly bedded marls of the Potamida Formation. Note on the southern side of the road, we can see that these sediments are capped by the marls of the Tortonian-Messinian Chairetiana Formation.

The presence of cobbles of the Phyllite-Quartzite unit in the sediments can be used to constrain the thermal evolution of the high-pressure, low-temperature rocks of Crete. Here, we summarize the results that we have discussed thus far throughout the trip (Figure 7). This discussion is a modification of the results of Thomson et al. (1999). The Oligocene flysch at the top of the Plattenkalk unit shows that it was still at the surface at ~29 Ma. However, the sediments of the Phyllite-Quartzite are thought to have been deposited to the north of the

Figure 21 – Photo of depositional contact between the Miocene sediments of the Fotokadhon Formation and Phyllite-Quartzite unit. The Fotodakhon is about 9 Ma based on calcareous nannoplankton, providing a minimum exposure age for the Phyllite-Quartzite nappe in western Crete



Plattenkalk platform, meaning that the rocks of the Phyllite-Quartzite were deeply subducted prior to 29 Ma. Estimates of the rate of plate convergence during the Oligocene and the inferred paleogeography suggest subduction of the Phyllite-Quartzite sometime between 36 to 32 Ma. White mica Ar-Ar ages, interpreted here as sub-closure crystallization ages, suggest peak metamorphic conditions were reached between 24 and 20 Ma (Jolivet et al., 1996). Previously published zircon He ages reveal cooling through 240°C at about 18 Ma. New zircon He ages show that the lower plate rocks cooled below 210°C around 10 Ma. Finally, conglomeratic deposits indicate exposure of the HP rocks at the surface at about 8.5 Ma. This suggests that the HP-LT rocks in the footwall of the Cretan were experienced a two-phase exhumation history, with modest exhumation between 20 and 10 Ma, followed by a rapid pulse of exhumation that ultimately led to the exposure of Phyllite-Quartzite bedrock at the surface by 9 Ma.

Drive westward back towards Kissamos. Turn left onto the Old National Road towards Chania (about 1 km before reaching Kissamos). Drive for 0.4 km and then turn right on the road towards Sirikari. Drive another 0.9 km, following the main road. Pull off and park at the hairpin turn in the road.

Day 3, Stop 5: The Chairetiana Formation, Messinian evaporite, more recent faulting

GPS: N 35.48585
E 023.67421

Here, laminated marls of the Chairetiana Formation outcrop along the side of the road. At the point in the hairpin turn, a small trail leads towards the west. A few meters from the road, the pavement exposes gypsum associated with the Messinian event. These rocks are exposed at 100 m elevation, thus requiring at least that much uplift since about 5 Ma.

This locality also provides an excellent view of the recent and still active normal faults. In western Crete, these are often oriented N-S. Looking to the north, we can see Kissamos Bay, a north-south oriented waterway bounded by long north-south peninsulas. These are bounded by N-S trending normal faults that have dropped the rocks beneath Kissamos Bay down relative to the rocks of the long peninsulas.

Continue north along the main road for another 0.6 km. Turn right on the road up to Marediana (the sign is only in Greek). Drive for 0.5 km and part at the hairpin turn off on the right-hand side of the road. Walk a short distance along the access road (50 m) to the outcrop.

Day 3, Stop 6: Messinian evaporite within Chairetiana formation (Optional stop)

GPS: N 35.48310
E 023.67142
Elevation: 105 m

Large (~20 cm) clasts of gypsum crystals can be found in the low outcrops alongside road. The gypsum appears to have been fragmented by depositional reworking. They are found as interbeds in finely laminated marls, indicating a marine depositional environment.

Return to the National Road and drive eastward to Chania for the evening.

DAY 4: SAMARIA GORGE

We will conclude our field excursion with a hike down the beautiful Samaria Gorge. This 16 km gorge is considered the longest in Europe. We will take a bus from Chania to the Omalos Plateau and to the entrance to the Samaria Gorge Park. The hike through the gorge typically takes 4 to 6 hours and ends at the village of Agia Roumeli. From there, we take a ferry eastward to Chora Sfakion where buses will take people back to Chania.

The road to the park passes first through the Phyllite-Quartzite nappe and then into the Trypali nappe after the village of Lakoi. The Samaria Gorge itself is cut into the Plattenkalk nappe. The whitish peaks at the head of the gorge are the Gigilos beds, the massive white marbles in the Plattenkalk that are stratigraphically above the Triassic stromatolitic dolomite but below the typical "platy limestone" horizons (Figure 11). The structure of the Lefka Ori region is depicted in Figure 22, which shows that as we hike down towards the coast we will travel progressively upsection in the Plattenkalk stratigraphy.

The hike begins at an elevation of 1250 m. The surrounding peaks are typically at about 2000 m and reach as high as 2450 m. The first five km of the hike are through the Gigilos beds, the massive white marbles in the Plattenkalk unit that we say at Day 1,

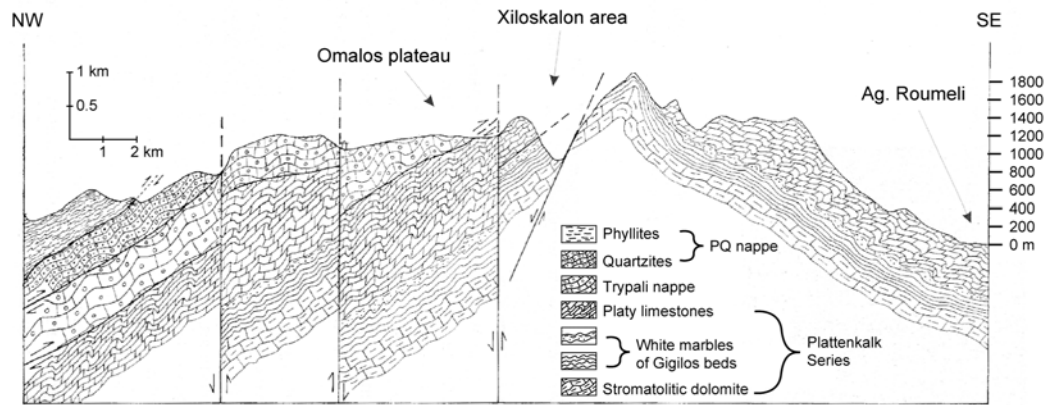


Figure 22 – Cross section of the Lefka Ori Mountains in the Omalos area of western Crete, modified from Fytrolakis (1980) and Fassoulas (2000). The Samaria Gorge begins in the Xiloskalon area and ends at Agios Roumeli

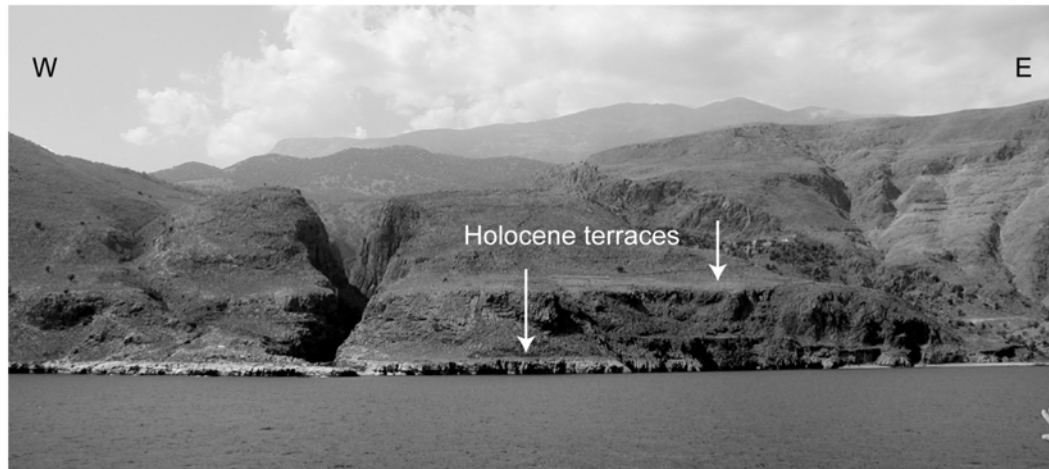
Stop 8. The impressive cliffs along the sides of the gorge offer impressive exposures of folds developed in the platy member of the Plattenkalk. The platy rocks drop down to the valley floor about 2 km above the abandoned village of Samaria. For the remainder of the gorge, the platy member is well exposed, revealing numerous south vergent asymmetric folds, similar to the Plattenkalk folds we saw earlier in the trip (Day 1, Stop 7).

The boat trip from Agia Roumeli at the mouth of the gorge to Chora Sfakion provides several views of other south-flowing gorges that

have been cut into the southwest flank of Crete. This south slope also has several uplifted wave-cut terraces (Figure 23), similar to the uplifted terraces that we saw east of Palaiochora (Day 3, Stop 1). Many of these terraces have been dated by Pirazzoli et al. (1982), who describe terraces of 8 distinct ages. Throughout western Crete, these are typically uplifted from 3 to 8 m over the past 2000 to 4000 years. If sustained, these values would be equivalent to uplift rate of 0.8 to 4 km/m.y..

The southwest coast of Crete appears to coincide with a south-dipping normal fault, with the hangingwall offshore to the south. Thus, the southwest coast of Crete represents the footwall of this fault. The uplifted coastal terraces and the deeply incised gorges along the southwest coast indicate that slip on this normal fault seems to be occurring primarily by rise of the footwall, rather than by fall of the hangingwall, as commonly assumed. This distinction is possible here because the geomorphic features allow us to measure displacements relative

Figure 23 – Photograph of uplifted Holocene terraces along the western coast of Crete



to sea level. This evidence of footwall rise during normal faulting suggests that horizontal extension may be related to basal accretion or underplating at depth beneath the island. The HP-LT metamorphism of the sub-detachment nappes indicates that they were accreted by underplating at depths of ~35 km. Continued underplating seems required to account for the 35 km of rock that presently underlie the Plattenkalk and Phyllite-Quartzite units. The record of rapid uplift along the south coast is also consistent with continued accretion through underplating.

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Reference cited

Angelier, J., Lyberis, N., Le Pichon, X., Barrier, E., and Huchon, P. (1982). The tectonic development of the Hellenic Arc and the Sea of Crete: a synthesis. *Tectonophysics* 86, 159-196.

Armijo, R., Lyon-Caen, H., and Papanastassiou, D. (1992). East-west extension and Holocene normal fault scarps in the Hellenic arc. *Geology* 20, 491-494.

Bizon, G., Bonneau, M., Leboulenger, P., Matesco, S., and Thiébault, F. (1976). Sur la signification et l'extension des <<massifs cristallins externes>> en Péloponnèse méridional et dans l'Arc égéen. *Bulletin de la Société géologique de France* 18, 337-345.

Bonneau, M. (1984). Correlation of the Hellenic Nappes in the south-east Aegean and their tectonic reconstruction. In: Dixon, J.E., and Robertson, A.H.F., eds. *The Geological Evolution of the Eastern Mediterranean*. Geological Society, London, Special Publications 17, 517-527.

Brix, M.R., Stöckhert, B., Seidel, E., Theye, T., Thomson, S.N., and Küster, M. (2002). Thermobarometric data from a fossil zircon partial annealing zone in high pressure-low temperature rocks of eastern and central Crete, Greece. *Tectonophysics* 349, 309-326.

Buick, I. (1991). Mylonite fabric development on Naxos, Greece. *Journal of Structural Geology* 13, 643-655.

Carlson, W.D., and Rosenfeld, J.L. (1981). Optical determination of topotactic aragonite-calcite growth kinetics: metamorphic implications. *Journal of Geology* 89, 615-638.

Creutzburg, N., Drooger, C.W., Meulenkamp, J.W., Papastamatiou, J., Seidel, and Tataris, A. (1977). General Geological Map of Crete (scale 1:200,000). *Institute of Geological and Mining Exploration (IGME), Athens*.

Creutzburg, N., and Seidel, E. (1975). Zum Stand der Geologie des Präneogens auf Kreta. *Neues. Jb. Geol. Paläont. Abh.* 149, 363-383.

Dewey, J.F. (1988). Extensional collapse of orogens. *Tectonics* 7, 1123-1139.

Dewey, J.F., Helman, J.L., Turco, E., Hutton, D.W.H., and Knott, S.D. (1989). Kinematics of the western Mediterranean. In: Coward, M.P. and Dietrich, D., eds., *Alpine Tectonics*. Geological Society, London, Special Publications 45, 265-283.

Duermeijer, C.E., Nyst, M., Meijer, P.Th, Langereis, C.G., and Spakman, W. (2000). Neogene evolution of the Aegean arc: paleomagnetic and geodetic evidence for a rapid and young rotation phase. *Earth and Planetary Science Letters* 176, 509-525.

Duermeijer, C.E., Krijgsman, W., Langereis, C.G., and ten Veen, J.H. (1998). Post early Messinian counter-clockwise rotations on Crete: implications for the Late Miocene to Recent kinematics of the southern Hellenic Arc. *Tectonophysics* 298, 177-189.

Epting, J., Kudrass, H-R., Leppig, U., and Schafer, A. (1972). Geologie der Talea Ori/Kreta. *Neues Jb. Geol. Paläont. Abh.* 141, 259-285.

Fassoulas, C., Kiliass, A., and Mountrakis, D. (1994). Postnappe stacking extension and exhumation of high-pressure low- temperature rocks in the island of Crete, Greece. *Tectonics* 13, 127-138.

Fassoulas, C. (2000). Field Guide to the Geology of Crete. *Natural History Museum of Crete, Publ.*, 104 p., Heraklio.

Fassoulas, C. (2001). The tectonic development of a Neogene basin at the leading edge of the active European margin: the Heraklion basin, Crete, Greece. *Journal of Geodynamics* 31, 49-70.

- Fortuin, A.R., and Peters, J.M. (1984). The Prina Complex in eastern Crete and its relationship to possible Miocene strike-slip tectonics. *Journal of Structural Geology* 6, 459-476.
- Frydas, D., and Keupp, H. (1996). Biostratigraphical results in Late Neogene deposits of NW Crete, Greece, based on calcareous nannofossils. *Berliner geowiss. Abh.* E 18, 169-189.
- Fytrolakis, N. (1980). The geological structure of Crete: Problems, observations, and conclusions. *Habil. Thesis, Nat. Techn. Univ. Athens*, 143 p.
- Greiling, R. (1982). The metamorphic and structural evolution of the Phyllite-quartzite nappe of western Crete. *Journal of Structural Geology* 4, 291-297.
- Hall, R., Audley-Charles, M.G., and Carter, D.J. (1984). The significance of Crete for the evolution of the Eastern Mediterranean. In: Dixon, J.E., and Robertson, A.H.F. (eds) *The Geological Evolution of the Eastern Mediterranean*. Geological Society, London, Special Publications 17, 499-516.
- Hall, R., and Audley-Charles, M.G. (1983). The structure and regional significance of the Talea Ori, Crete. *Journal of Structural Geology* 5, 167-179.
- Hatzfeld, D., Martinod, J., and Bastet, G. (1997). An analog experiment for the Aegean to describe the contribution of gravitational potential energy. *Journal of Geophysical Research* 102, 649-659.
- Huguen, C., Mascle, J., Chaumillon, E., Woodside, J.M., Benkheilil, J., Kopf, A., and Volskonkaia, K.A. (2001). Deformation styles of the eastern Mediterranean Ridge and surroundings from combined swath mapping and seismic reflection profiling. *Tectonophysics* 343, 21-47.
- Jackson, J. (1994). Active tectonics of the Aegean region. *Annual Reviews of Earth and Planetary Science* 22, 239-271.
- Jiménez-Munt, I., Sabadini, R., Gardi, A., and Bianco, G. (2003). Active deformation in the Mediterranean from Gibraltar to Anatolia inferred from numerical modeling and geodetic and seismological data. *Journal of Geophysical Research* 108, doi:10.1029/2001JB001544.
- Jolivet, L., Goffe, B., Monie, P., Truffert-Luxey, C., Patriat, M. & Bonneau, M. (1996). Miocene detachment in Crete and exhumation P-T-t paths of high-pressure metamorphic rocks. *Tectonics* 15, 1129-1153.
- Karakitsios, V. (1987). Sur la signification de la série de Trypali dans la région de Sellia en Crète occidentale (Grèce). *CR Acad. Sci. Paris*, 304 (II/3), 123-128.
- Kastens, K.A. (1991). Rate of outward growth of the Mediterranean Ridge accretionary complex. *Tectonophysics* 199, 25-50.
- Kilias, A., Falalakis, G., and Mountrakis, D. (1999). Cretaceous - Tertiary structures and kinematics of the Serbomacedonian metamorphic rocks and their relation to the exhumation of the Hellenic hinterland (Macedonia, Greece). *International Journal of Earth Sciences*, 88, 513-531.
- Kissel, C., and Laj, C. (1988). The Tertiary geodynamical evolution of the Aegean arc: a paleomagnetic reconstruction. *Tectonophysics* 146, 183-201.
- Knapmeyer, M., and Harjies, H.P. (2000). Imaging crustal discontinuities and the downgoing slab beneath western Crete. *Geophysical Journal International* 143, 1-21.
- Kopp, K., and Ott, E. (1977). Spezialkartierungen im Umkreis neuer Fossilfunde in Tripali und Tripolitzakalken Westkretas. *Neues Jb. Geol. Paläont. Monat.*, 4, 217-238.
- Kopf, A., Mascle, J., and Klaeschen, D. (2003). The Mediterranean Ridge: A mass balance across the fastest growing accretionary complex on Earth. *Journal of Geophysical Research*, 108, doi:10.1029/2001JB000473.
- Krahl, J., Kauffmann, G., Kozur, H., Richter, D., Förster, O., and Heinritzi, F. (1983). Neue Daten zur Biostratigraphie und zur tektonischen Lagerung der Phyllit-Gruppe und der Trypali-Gruppe auf der Insel Kreta (Griechenland). *Geologische Rundschau* 72, 1147-1166.
- Krahl, J., Richter, D., Forster, O., Kozur, H., and Hall, R. (1988). Zur Stellung der Talea Ori im Bau des kretischen Deckenstapels (Griechenland).

- Zeitschrift der deutschen geologischen Gesellschaft* 139, 191-207.
- Kuss, S.E. (1980). Führer zur Kreta – Exkursion. *Geologisch – Paleontologischen Institut Universität Freiburg*.
- Küster, M., and Stöckhert, B. (1997). Density changes of fluid inclusions in high-pressure low-temperature metamorphic rocks from Crete: A thermobarometric approach based on the creep strength of the host minerals. *Lithos* 41, 151-167.
- Le Pichon, X., and Angelier, J. (1979). The Hellenic arc and trench system: a key to the neotectonic evolution of the Eastern Mediterranean area. *Tectonophysics* 60, 1-42.
- Le Pichon, X., and Angelier, J. (1981). The Aegean Sea. *Philosophical Transactions of the Royal Society, London*, 300, 357-372.
- Le Pichon, X., Lallemand, S.J., Chamot-Rooke, N., Lemeur, D., and Pascal, G. (2002). The Mediterranean Ridge backstop and the Hellenic nappes. *Marine Geology* 186, 111-125.
- Lister, G.S., Banga, G., and Feenstra, A. (1984). Metamorphic core complexes of Cordilleran type in the Cyclades, Aegean Sea, Greece. *Geology* 12, 221-225.
- Liu, M., and Yund, R.A. (1993). Transformation kinetics of polycrystalline aragonite to calcite: new experimental data, modelling, and implications. *Contributions to Mineralogy and Petrology* 114, 465-478.
- Makris, J. (1978). The crust and upper mantle of the Aegean region from deep seismic sounding. *Tectonophysics* 46, 269-284.
- Makris, J., and Stobbe, C. (1984). Physical properties and state of the crust and upper mantle of the eastern Mediterranean Sea deduced from geophysical data. *Marine Geology* 55, 217-254.
- McClusky, S. et al. (2000). Global Position System constraints on plate kinematics and dynamics in the eastern Mediterranean and Caucasus. *Journal of Geophysical Research* 105, 5695-5719.
- McKenzie, D.P. (1972). Active tectonics of the Mediterranean region. *Geophysical Journal of the Royal Astronomical Society* 30, 109-185.
- McKenzie, D.P. (1978). Some remarks on the development of sedimentary basins. *Earth and Planetary Science Letters* 40, 25-32.
- Meijer, P.Th. and Wortel, M.J.R. (1997). Present-day dynamics of the Aegean region: A model analysis of the horizontal pattern of stress and deformation. *Tectonics* 16, 879-895.
- Meulenkamp, J.E., Dermitzakis, M., Georgiadou-Dikeoulia, E., Jonkers, H.A., and Böger, H. (1979). *Field Guide to the Neogene of Crete*. Publications of the Department of Geology and Paleontology, University of Athens, A, 32.
- Meulenkamp, J.E., Wortel, M.J.R., van Wamel, W.A., Spakman, W., and Hoogerduyn Strating, E. (1988). On the Hellenic subduction zone and the geodynamic evolution of Crete since the late Middle Miocene. *Tectonophysics* 146, 203-215.
- Meulenkamp, J.E., van der Zwaan, G.J., and van Wamel, W.A. (1994). On Late Miocene to Recent vertical motions in the Cretan segment of the Hellenic arc. *Tectonophysics* 234, 53-72.
- Pirazzoli, P.A., Thommeret, J., Thommeret, Y., Laborel, J., and Montaggioni, L.F. (1982). *Tectonophysics* 86, 27-43.
- Platt, J.P. (1986). Dynamics of orogenic wedges and the uplift of high-pressure metamorphic rocks. *Geological Society of America Bulletin* 97, 1037-1053.
- Pomoni-Papaioannou, F., and Karakitsios, V. (2002). Facies analysis of the Trypali carbonate unit (Upper Triassic) in central-western Crete (Greece): an evaporite formation transformed into solution-collapse breccias. *Sedimentology* 49, 1113-1132.
- Robertson, A.H.F., Clift, P., Degnan, P.J., and Jones, G. (1991). Palaeogeographic and paleotectonic evolution of the eastern Mediterranean Neotethys. *Palaeogeography, Palaeoclimatology, and Palaeoecology* 87, 289-343.
- Royden, L.H. (1993). The tectonic expression of slab pull at continental convergent boundaries. *Tectonics* 12, 303-325.

- Sannemann, W., and Seidel, E. (1976). Die Trias-Schichten von Rawdoucha/NW-Kreta. Ihre Stellung im kreitischen Deckenbau. *Neues Jb. Geol. Paläont. Mh.* 4, 221-228.
- Seidel, M. (2003). Tectono-sedimentary evolution of middle Miocene supra-detachment basins (western Crete, Greece). *Ph.D. Dissertation, University of Köln*. 116 pages.
- Seidel, E., Okrusch, m., Kreuzer, H., Raschka, H., and Harre, W. (1976) Eo-alpine metamorphism in the uppermost unit of the Cretan nappe system – petrology and geochronology – Part 1. The Lendas area (Asteroussia Mountains). *Contributions to Mineralogy and Petrology* 57, 259-275.
- Seidel, E., Kreuzer, H., and Harre, W. (1982). A Late Oligocene/Early Miocene high pressure belt in the External Hellenides. *Geologisches Jahrbuch* E23, 165-206.
- Schwarz, S., and Stöckhert, B. (1996). Pressure solution in siliciclastic HP-LT metamorphic rocks - constraints on the state of stress in deep levels of accretionary complexes. *Tectonophysics*, 255, 203-209.
- Spakman, W., Wortel, M.J.R., and Vlaar, N.J. (1988). The Hellenic subduction zone: a tomographic image and its geodynamic implications. *Geophysical Research Letters* 15, 60-63.
- Stöckhert, B., Wachmann, M., Küster, M., and Bimmermann, S. (1999). Low effective viscosity during high pressure metamorphism due to dissolution precipitation creep: the record of HP-LT metamorphic carbonates and siliciclastic rocks from Crete. *Tectonophysics* 303, 299-319.
- Ten Veen, J.H., and Meijer, P.T. (1998). Late Miocene to Recent tectonic evolution of Crete (Greece): geological observations and model analysis. *Tectonophysics* 298, 191-208.
- Ten Veen, J.H., and Postma, G. (1999a). Neogene tectonics and basin fill patterns in the Hellenic outer-arc. *Basin Research* 11, 223-241.
- Ten Veen, J.H., and Postma, G. (1999b). Roll-back controlled vertical movements of outer-arc basins of the Hellenic subduction zone (Crete, Greece). *Basin Research* 11, 243-266.
- Theye, T., and Seidel, E. (1991). Petrology of low-grade high-pressure metapelites from the External Hellenides (Crete, Peloponnese): A case study with attention to sodic minerals. *European Journal of Mineralogy* 3, 343-366.
- Theye, T., Seidel, E., and Vidal, O. (1992). Carpholite, sudoite, and chloritoid in low-grade high-pressure metapelites from Crete and the Peloponnese. *European Journal of Mineralogy* 4, 487-507.
- Theye, T., and Seidel, E. (1993). Uplift-related retrogression history of argonite marbles in Western Crete. *Contributions to Mineralogy and Petrology* 114, 349-356.
- Theye, T., and Seidel, E. (2001). Bauxite in hochdruckmetamorphen Einheiten der externen Helleniden. *Braunschweiger geowiss. Arb.* 24, 225-233.
- Thomson, S.N., Stöckhert, B., and Brix, M.R. (1998a). Thermochronology of the high-pressure metamorphic rocks of Crete, Greece: implications for the speed of tectonic processes. *Geology* 26, 259-262.
- Thomson, S.N., Stöckhert, B., Rauche, H., and Brix, M.R. (1998b). Apatite fission-track thermochronology of the Uppermost Tectonic Unit of Crete, Greece: Implications for the post-Eocene tectonic evolution of the Hellenic subduction system. In: Van den haute, P., and De Corte, F. (eds). *Advances in Fission-Track Thermochronology*, 187-205. (Kluwer Academic Publishers.)
- Thomson, S.N., Stöckhert, B. and Brix, M.R. (1999). Micocene high-pressure metamorphic rocks of Crete, Greece: rapid exhumation by buoyant escape. In: Ring, U., Brandon, M.T., Lister, G.S. & Willett, S.D. (eds) *Exhumation Processes: Normal Faulting, Ductile Flow and Erosion*. Geological Society, London, Special Publications, 154, p. 87-107.
- Willett, S.D. (1999). Rheological dependence of extension in wedge models. *Tectonophysics* 305, 419-435.
- Wortel, M.J.R., and Spakman, W. (2000). Subduction and slab detachment in the Mediterranean-Carpathian region. *Science* 290, 1910-1917.