My GeoGuide

Nappe sequence near Gonies and the Anogia Thrust in the Psilorites Mountains

The Windmill above Gonies village. On the left Vatos Phyllite and Ophiolites of the Upper Most Nappes

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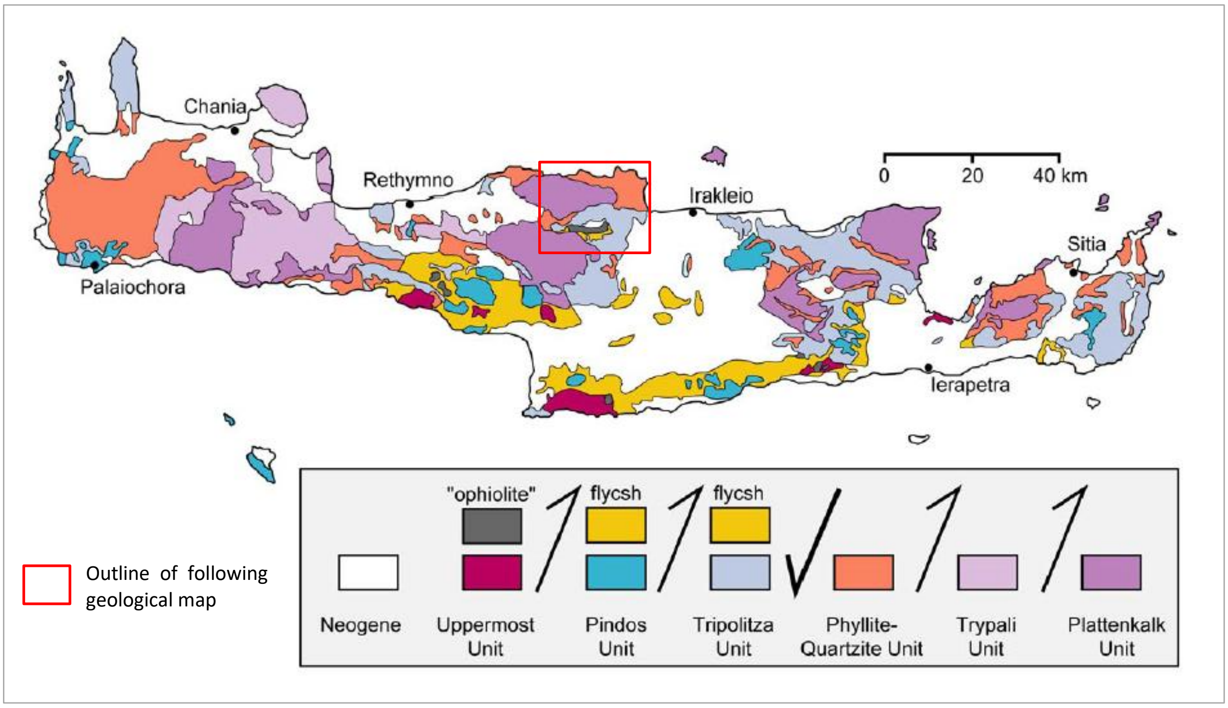
Berlin, January 2024

Content

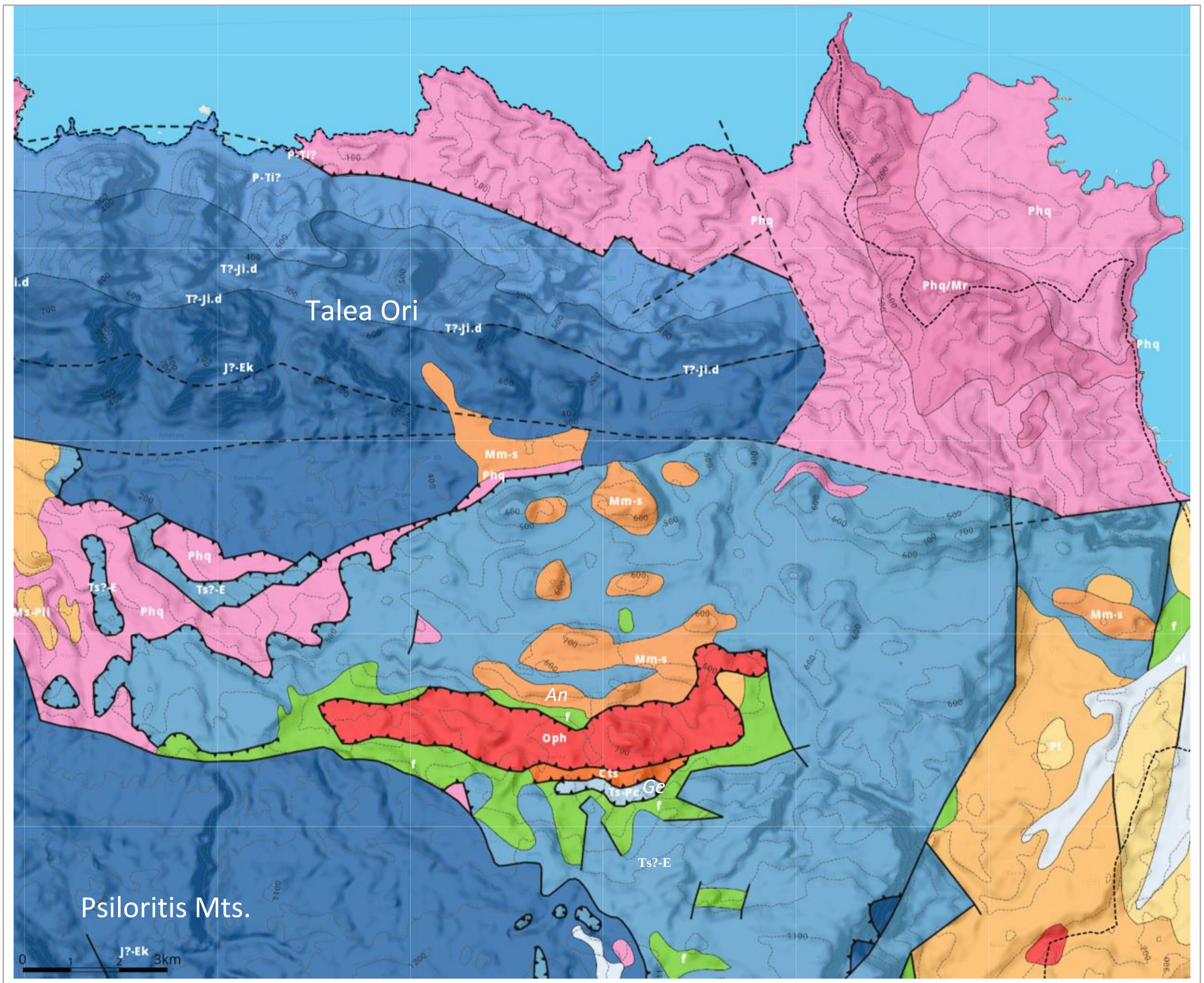
Appendix

1 Introduction



Simplified geologic map of Crete showing the remains of the different nappes. Arrows within the legend indicate major thrust planes. The bold arrow represents the Cretan Detachment, which divides the lower metamorphic Units from the less or non-metamorphic Pindos und Tripoliza Units. After Creutzburg et al. (1977) and Thomson et al. (1999). [J. M. Rahl et. al., 2004]

The island of Crete is located at the centre of the Hellenic arc and just over the active Hellenic subduction zone. It therefore plays a key role in the understanding the development of recent orogenic processes and the active geodynamic regime in the southern Aegean. At least seven main nappes exist on the island of Crete, all exposed at its central part, mainly around the Psiloritis mountains, which host the highest summit of the island at 2456 m. Most of these rocks have their equivalence or are the direct continuation of the mainland Greece geotectonic units. The backbone of the whole area of Crete forms the Plattenkalk or Kriti-Mani unit, which is considered as the metamorphic equivalent of the Ionian zone. It forms a continuous carbonate platform of more than 5 km thickness, developed from the Permian to Oligocene times. The whole stratigraphy is displayed in excellent outcrops in the area of Talea Ori mountains, the northern extension of Psiloritis; in several areas even in inverted form. Over the Plattenkalk and in direct tectonic contact are either the group of rocks that constitute the Phyllite-quartzites nappe, or if it is missing, the overlying Tripolitza carbonate nappe. Similar to the Plattenkalk the Phyllite-quartzites exhibits high pressure metamorphism of Late Oligocene/Early Miocene times.



Geological Map of the area between the Talea Ori and Psiloritis Mountains showing the ophiolites of the Uppermost Units An: Anogia Thrust, Ge: Gonies (Source: Psiloritis Geopark WebGIS).

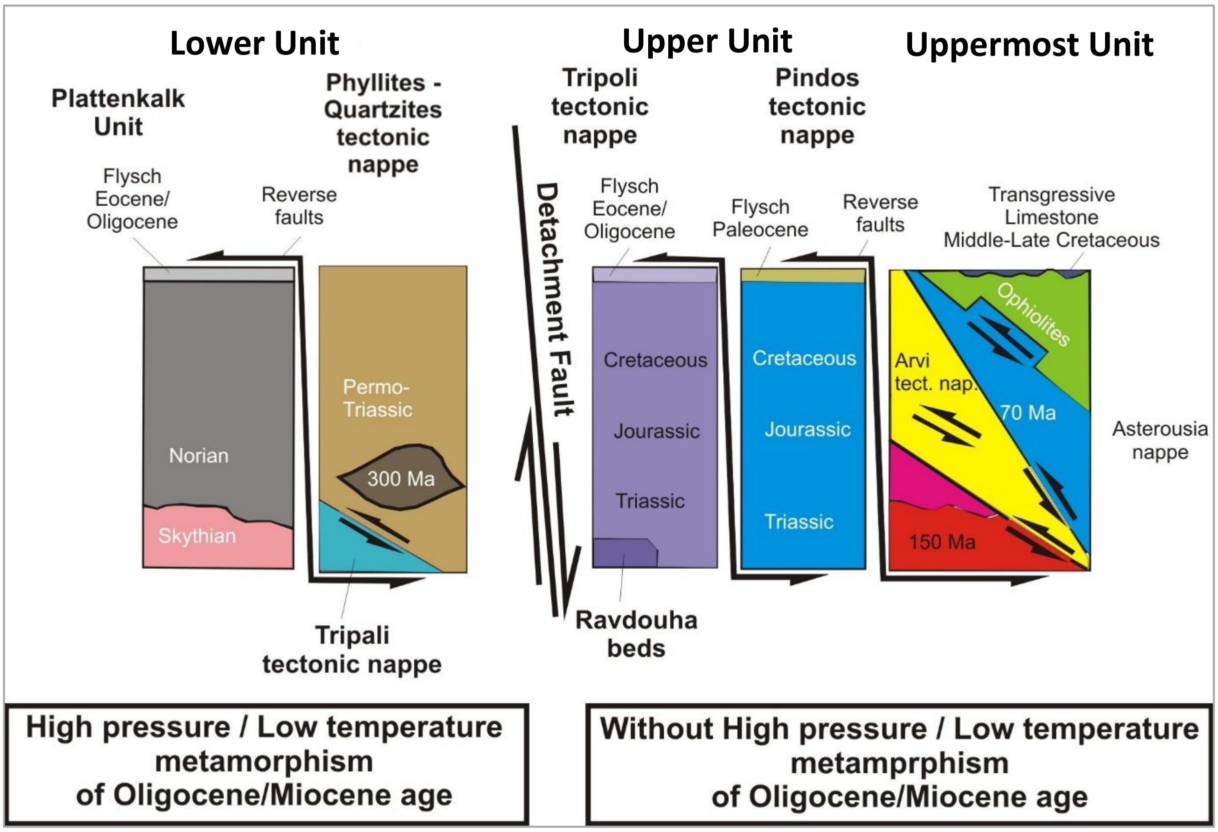
Oph	**Ophiolites** Upper Jurassic ophiolites comprising of serpentinites, pyroxenites, peridotites and gabbros.
Cts	**Vatos, Arvi nappes** Calcitic phyllites and reddish limestone of upper Jurassic age (Vatos) and pillow basalts and lavas of Upper Cretaceous age (Arvi).
ft	**Pindos nappe / Flysch** It includes Eocene tourbidites, fine bedded sandstone, clay, chert and limestone.
Ts-Pc	**Pindos nappe / Limestone** Deep marine sediments of greywacke, shale, radiolarites and platy limestone with cherty intercalations of Upper Triassic to Eocene age.
f	**Tripoliza nappe / Flysch** Upper Eocene sandstone, shale, conglomerates and brecciated limestone.
Ts?-E	**Tripoliza nappe / Limestone** A thick series of limestone and dolomite of Upper Triassic to Upper Cretaceous in age, deposited in a shallow marine environment with plenty of fossils.
Phq/Mr	**Phyllite-quartzite nappe / Marble** Thick bedded, white marble embedded in the phyllites. It appears in the area of Vasiliko and Fodele.

Phq	**Phyllite-quartzite nappe / Phyllites and quartzites** They comprise of Upper Paleozoic to Upper Triassic in age calcitic-, talc- and graphitic-phyllites, greenschists, quartzites, meta-andesites, meta-basalts, and meta-conglomerates.
P-Ti?	**Plattenkalk Unit / Fodele-Sisses beds** Crystalline limestone, white marble and dark shale of Permian to Lower Triassic in age with many typical fossils.
T?-Ji.d	**Plattenkalk Unit / Stromatolithic dolomite** Dark gray, laminated dolomites and stromatolites of Upper Triassic- Lower Jurassic age.
J?-Ek	**Plattenkalk Unit / Platy marble** Platy, grey marble with silica intercalations and lenses of Upper Jurassic to Oligocene in age. At the upper parts they pass gradually into a calcitic meta-flysch. At the area of Nida typical reddish silica layers appear also within platy marble.
Pl	**Pliocene sediments** Whitish to yellowish, fine bedded in places marls, bio-clastic limestone and sandy clay.
Ms-Pli	**Upper Miocene -Pliocene sediments** Bio-clastic, reefal limestone, alternated with whitish to yellowish marls and evaporites (anhydrite, gypsum).
Mm-s	**Middle Miocene sediments** Mainly terrestrial deposits of river and lagoonal sediments like conglomerates, sandstone, clay, lignites and limestone.

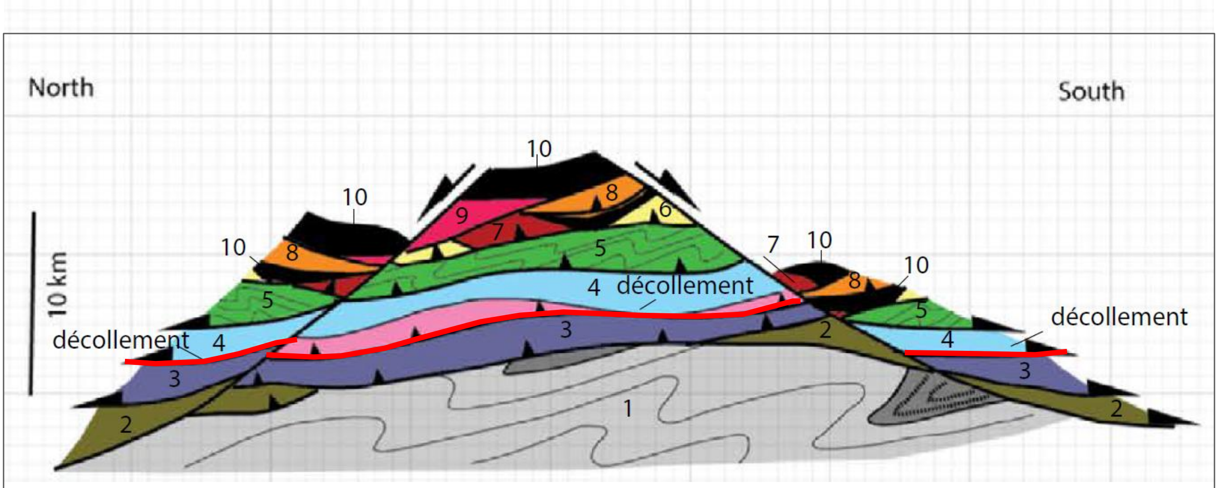
The Tripoliza carbonate nappe is characteristically non- or only slightly metamorphic. As in mainland Greece, the Pindos nappe in Crete has been emplaced above the Tripolitza nappe. It can be observed in several places such as in southern central Crete. All these external Hellenides rocks are overthrust by further nappes known as the Uppermost Unit that are thought to belong to the internal Hellenides. The Uppermost Unit is made up of a number of subunits consisting of blueschists (Preveli Group), medium temperature metamorphics and metabasalts (Spili group), pillow lavas (Arvi group) and only slightly metamorphic marine sediments (Vatos group). Part of the Uppermost Unit is thought by some to be an ophiolitic mélange, owing to its chaotic structure and association with oceanic crust. Remains of the Uppermost unit are scattered across central Crete mainly in the northern Psiloritis area, in southern central Crete, and in the Asteroussia Mts. As part of the Uppermost Unit the Asteroussia nappe consists of highly metamorphic crystalline rocks. The high temperature low pressure metamorphosed rocks are of Upper Cretaceous age and are considered to belong to the Internal Hellenides. Also, at the top of the Uppermost Unit lie upper Jurassic ophiolites, which are sparsely exposed in southern central Crete and at Gonies. *[Fassoulas C. et. al.]*

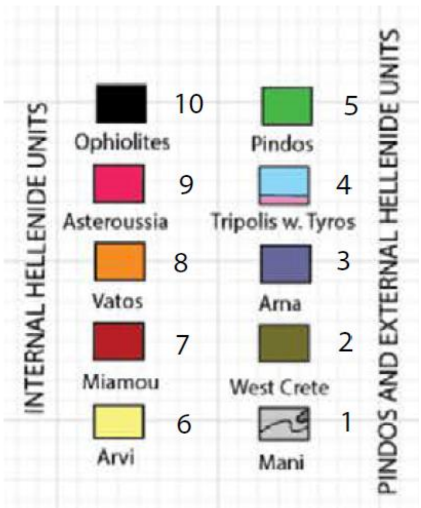
Table: Corelation of Units between Mainland Greece and Crete

Unit in Mainland Greece	Correlating Unit on Crete	Notes
Mani or Cretan-Mani Sequence	Plattenkalk (Lower Unit)	Parautochthonous carbonates; correlates with units in the Peloponnese.
Phyllite-Quartzite Unit (Peloponnese), Arna	Phyllite-Quarzite Nappe (Lower Unit)	Metamorphosed sediments; similar lithologies and metamorphic history.
Tripolis /Gavaros Zone (External Hellenides)	Tripolis Nappe (Upper Unit)	Carbonates and flysch; widespread in mainland Greece.
Pindos /Ionian Zone (External Hellenides)	Pindos Nappe (Upper Unit)	Deep marine sediments; extends from Crete to central Greece.
Internal Hellenides *(Manoutsoglou E., et. al. 2022)*	Uppermost Unit	Encompasses ophiolites and high-grade metamorphic rocks.



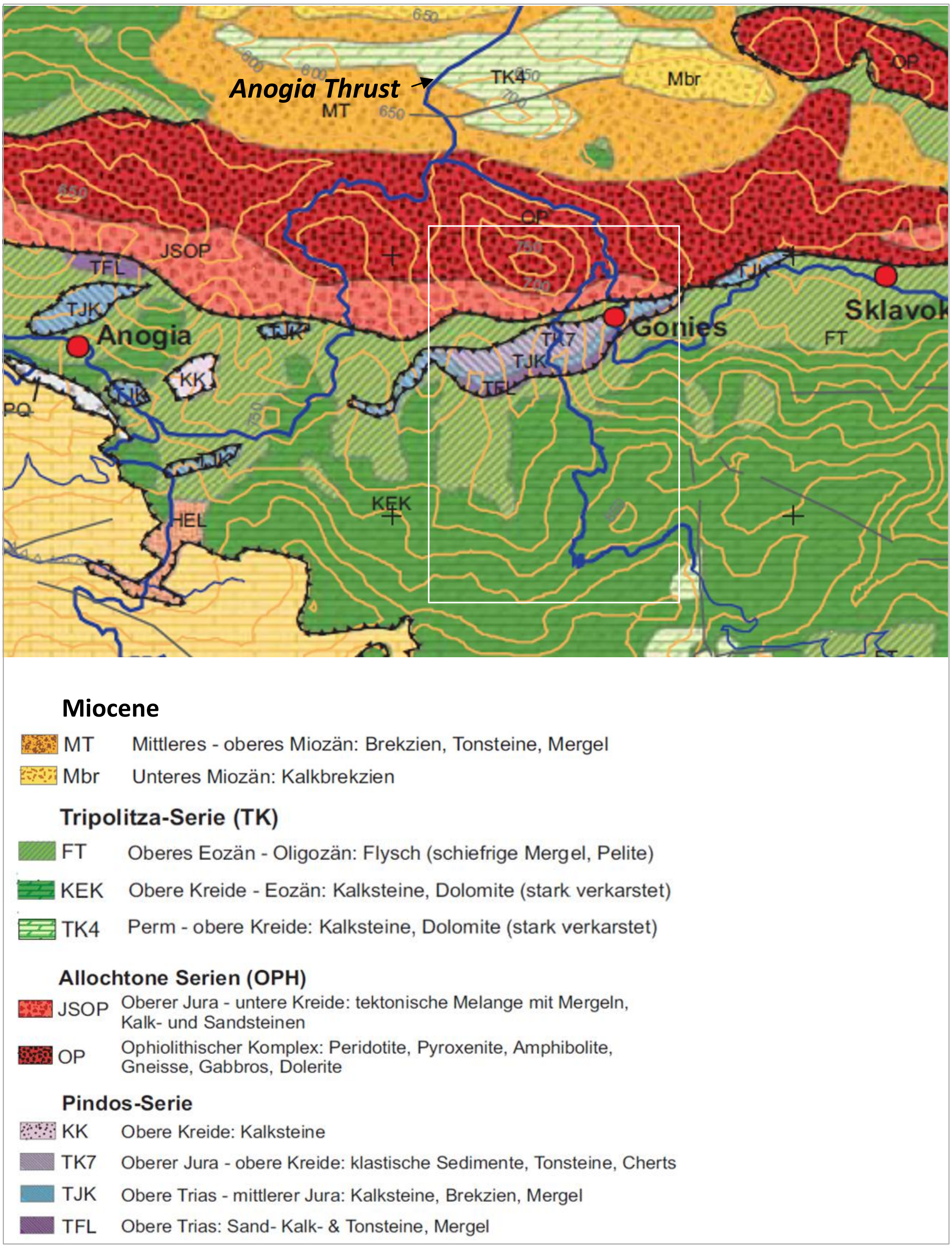
Source: [Journal of the Virtual Explorer | A dynamic review electronic Earth Science journal publishing material from all continents]()





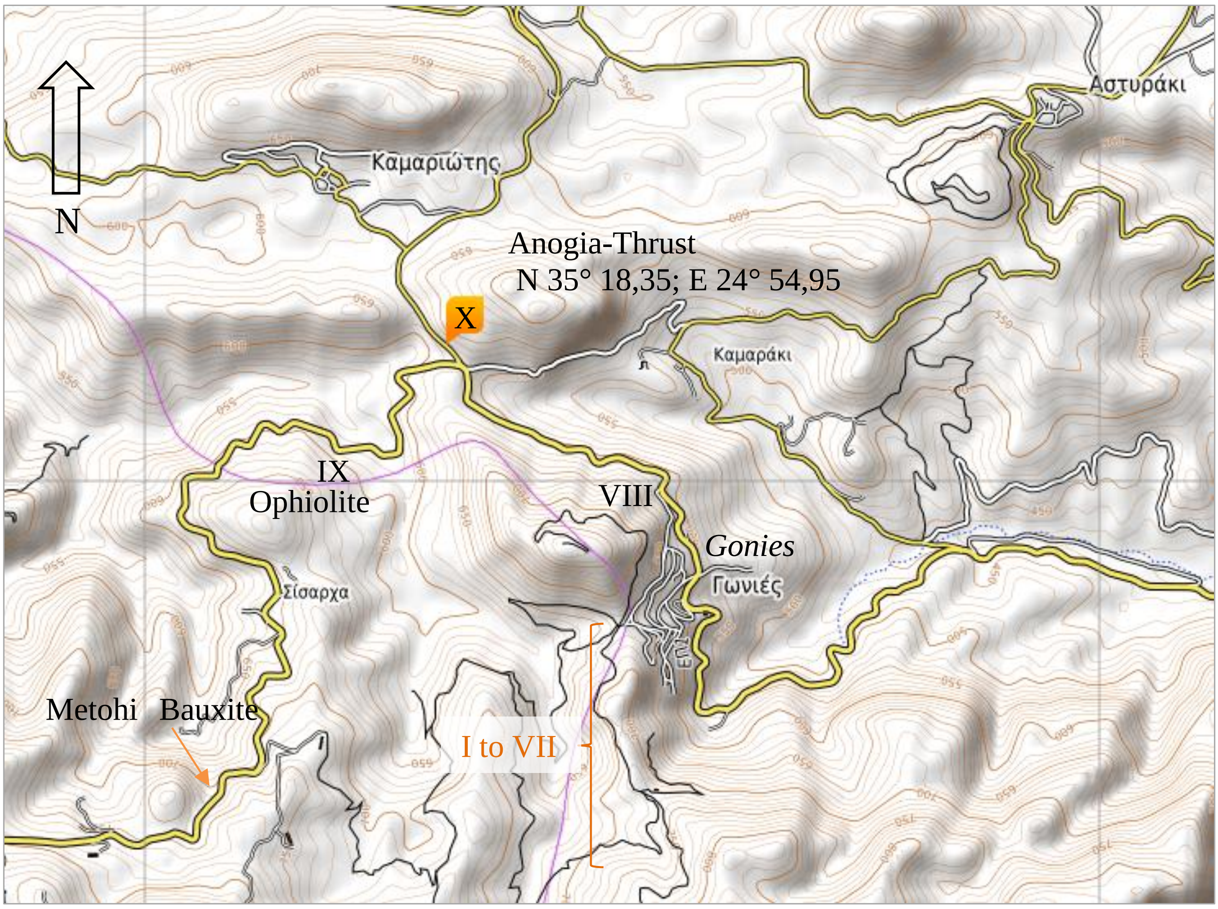
Generalized north-south cross section of Crete compiling data from different parts of major structural units formed during the Oligocene to Middle Miocene. Thrust sheets above the Pindos unit are inner Hellenide units. The Pindos unit was emplaced above the Gavrovo and Tripolis units and forms the upper crust above a décollement that separates upper crustal and lower crustal units. The lower crustal units are metamorphosed from lower to medium grade. They were emplaced in Oligocene to Middle Miocene time when they were subjected to extension and cut by normal faults that not only displaced the thrust sequence but also overprinted some of the thrust faults as normal faults. Redrawn from Papanikolaou and Vassilakis (2010). Burchfiel et al., 2017

The complicated tectonic setting of Crete provides insight into the late orogenic development of the Hellenides, but also the early, Eocene - Oligocene mountain building process that emplaced the Asteroussia nappe belonging to the Internal Hellenides on top of the external Hellenide rocks - over the Pindos and Tripolitza Units. Well preserved thrusts, reverse faults and intense folding are well preserved in many of the tectonic units. The most outstanding geological process though, that can be studied in central Crete is related to the unroofing and the exhumation of the lowermost Plattenkalk and Phyllites-quartzite, high pressure metamorphic rocks in late Miocene times. The earlier Late Oligocene/Early Miocene underplating process is documented by a residual paragenesis with mainly carpholite, which can be found in the Phyllite-quartzites and Plattenkalk rocks, and is evident within the intense folding of the Plattenkalk series. According to zircon dating the unroofing of the lower, high pressure metamorphosed rocks was a very fast process. Most researchers agree that it took place during a north-south crustal extension which resulted in low-angle detachment faulting at the lower crust and high angle, listric normal faulting on the upper crust, which prevailed in all nappes at the latest exhumation stages. Due to this crustal extension all nappes have been re-emplaced, and all contacts of the upper nappes appear now as normal faults. A typical succession of rocks ranging from the Tripolitza to the uppermost Ophiolites nappes, that have been affected by the Late Miocene extension can be seen along the Gonies Section of the Psiloritis Geopark. *[modified after Fassoulas C. et. al.]*



Geological map of the area around Gonies

2 Anogia Thrust



Location of outcrops, X: Anogia Thrust, IX: serpentinite/ophiolites, VIII: serpentinite/ophiolites

Anogia Thrust, X: Detachment fault, Tripoliza auf Neogene [Source of image: Google Maps]

Outcrop X: Anogia Thrust exposed at a road cutting (Rucksack at edge of road for scale) 1: Tripoliza limestone, 2: Neogene marly sediments

Outcrop X, Miocene beds: soft sandy marly sediments

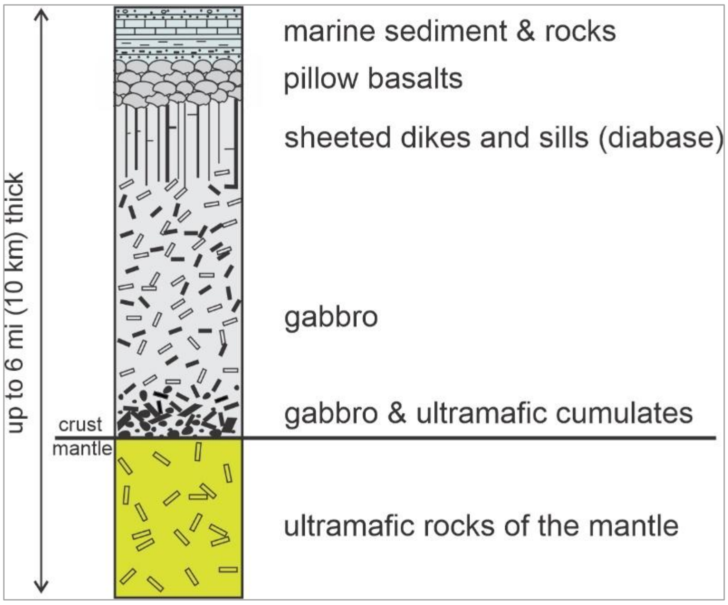
View of the Anogia thrust looking westwards from Outcrop X. 1: Tripoliza limestone, 2: Miocene sediments consisting of conglomertates, sandstones and marls towards the top. The tectonic contact of the Miocene sediments to the Tripoliza limestone is reported to have an approx. 5m thick cataclastic fault zone with limestone breccias. The thickness of the overridden Miocene sediments is approx. 100m. The Anogia Thrust extends for at least 11 km along strike.

3 Supra-Detachment Nappe Sequence Near Gonies

3.1 Ophiolites of the Uppermost Nappe

An ophiolite is a section of Earth's oceanic crust and the underlying upper mantle that has been uplifted and often emplaced onto continental crustal rocks. Their great significance relates to their occurrence within mountain belts such as the Alps and the Hellenides, where they document the existence of former ocean basins that have now been consumed by subduction. This insight was one of the founding pillars of plate tectonics, and ophiolites have always played a central role in plate tectonic theory and the interpretation of ancient mountain belts. The stratigraphic-like sequence observed in ophiolites corresponds to the lithosphere-forming processes at mid-oceanic ridges. Ideally, from top to bottom, the layers in the sequence are:

- Pelagic sediments: mostly siliceous oozes, calcareous oozes and red clays deposited since the crust formed.
- Extrusive sequence: basaltic pillow lavas show magma/seawater contact.
- Sheeted dike complex: vertical, parallel dikes that fed lavas above.
- High level intrusives: isotropic gabbro, indicative of a fractionated magma chamber.
- Layered gabbro, resulting from settling out of minerals from a magma chamber.
- Cumulate peridotite: olivine-rich layers of minerals that settled out from a magma chamber.
- Tectonized peridotite: harzburgite/lherzolite-rich mantle rock.



Cross section of a complete ophiolite sequence [Open Geology https://opengeology.org/petrology/13-metamorphism-of-mafic-rocks/#1371_Ophiolites_Serpentinites_and_Metaperidotites]

However, ocean crust can be quite variable in thickness and composition, and in places sheeted dikes sit directly on peridotite tectonite, with no intervening gabbros.

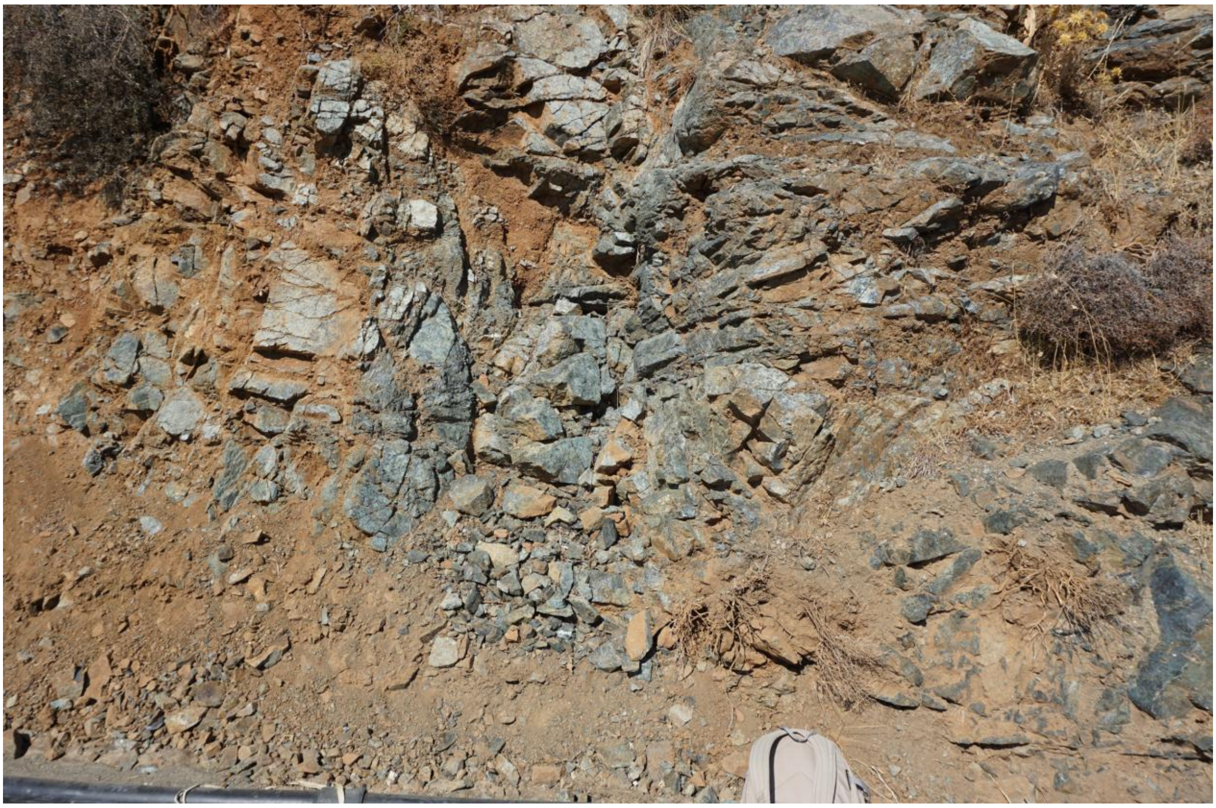
The dismembered ophiolites of Crete represent the southernmost outliers of the Jurassic ophiolite belt of the Dinarides/ Hellenides (Koepke et al., 2002). They form decameter- to kilometer- sized isolated bodies within the “Uppermost Unit”. The Cretan ophiolites mainly consist of serpentinites with relatively high contents of Al_2O_3 and CaO and include relics of spinel lherzolite. *[Koepke J. et. al., 2004]*

lherzolite is an ultrabasic igneous rock dominated essential by olivine and clinopyroxene and orthopyroxene in equal proportions. Accessory minerals include plagioclase, spinel, garnet, ilmenite, chromite and magnetite. Lherzolites are a peridotite and the main component of the upper mantle. Their aluminous phases change with pressure, with plagioclase present at low pressures, spinel at intermediate pressure and garnet at high pressure. *[Alex Stekeisen]*

See also the appendix of My GeoGuide “No. 21: The Uppermost Nappes: Ophiolites and the Vatos Nappe, Aktounda to Ardaktos” for further details on ophiolites

Outcrop IX, serpentinite (ophiolite). Alteration product of mafic and ultramafic rocks representing former oceanic crust.

Outcrop IX, closeup: serpentinite (ophiolite)

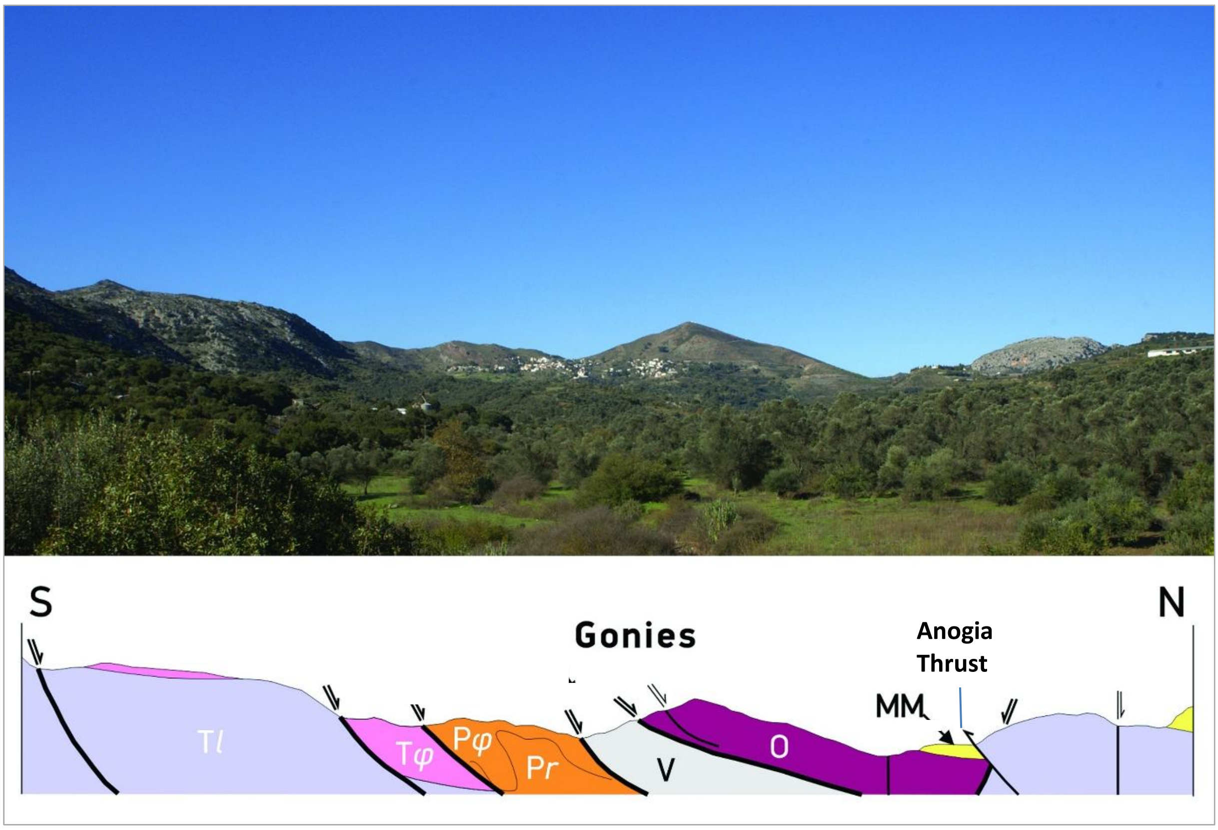


Outcrop VIII: Serpentinite (ophiolite)

Outcrop VIII: Serpentinite (ophiolite)

The hill above the village of Gonies is made of the ophiolitic rocks (O) that have been altered to serpentinite during hydrothermal processes within the oceanic crust (see cross-section below). Underlying the ophiolites are lavas of the Arvi nappe. Other rocks associated with the Uppermost Nappes are the Vatos nappe (V), which is exposed around the wind mill at the southern base of the hill. These consist of party folded grey phyllite. Following the dirt road southwards flysch sediments (Pφ) belonging to the Pindos unit consisting of red phyllite and sandstones with radiolarites may be observed. They are underlain by red platy limestone and the typical red chert of the Pindos Unit (Pr). Further down section the radiolarites are followed by further Pindos flysch. The red cyclic phyllite beds are reported to be of lowermost Triassic age. A normal fault forms a tectonic contact between the Pindos flysch and the Tripolitza nappe. Following the road south the Tripoliza nappe first displays yellowy-green flysch sediments partly containing coarse grained sandstones (Tφ). These extend further south into Tripoliza limestone beds (Tl).

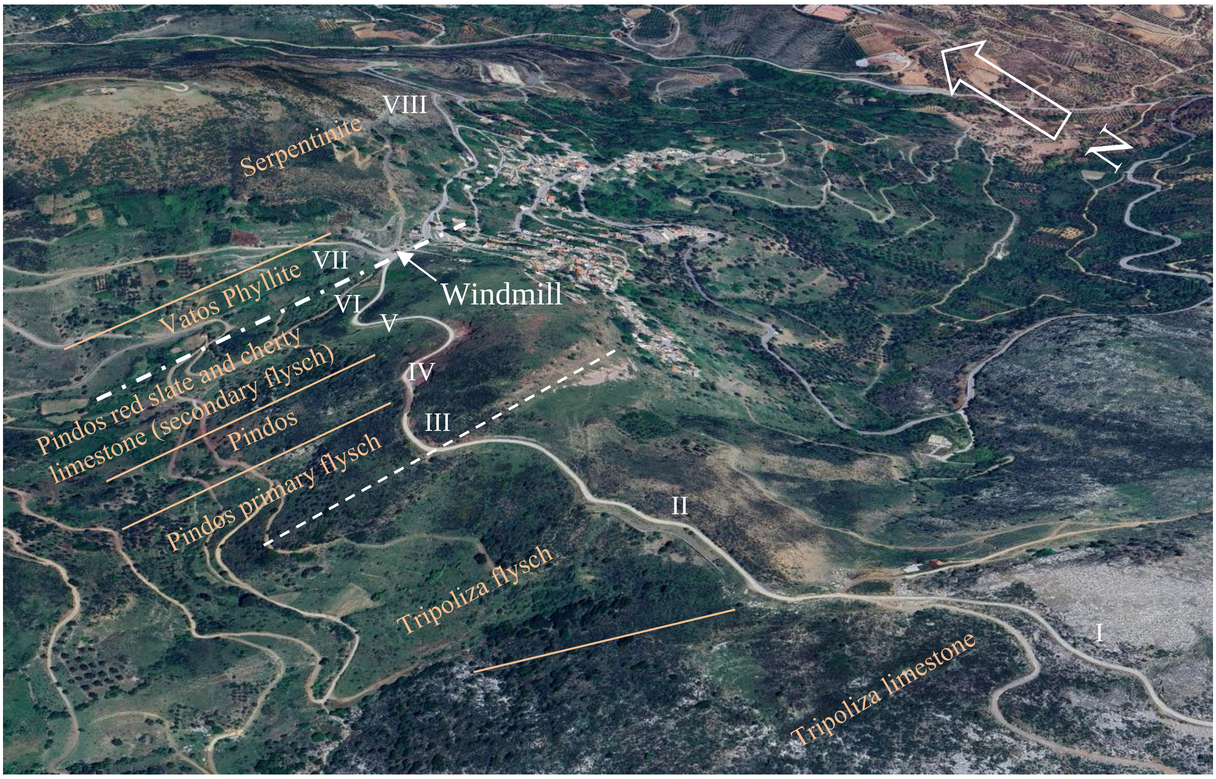
Younger Miocene sediments are marked in the following cross section in yellow (MM). *[Psiloritis Geopark WebGIS]*



North-South cross-section of the Gonies region. (Source: Psiloritis Geopark WebGIS) [PSILORITIS GEOPARK WebGIS App]

The dirt road leading to the windmill provides an opportunity to view a nearly continuous section that exposes all of the supra-detachment nappes. The traverse, walking downhill towards the north and the windmill, provides a view of the various units moving up-section. 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 nappe.

The section ends at Gonies and the windmill, so start approx. 800m further south down the road.



Location of outcrops. Dashed line indicates fault off-setting Pindos and Tripoliza flysch deposits. Dash-dot line represents thrust plane between Upper and Uppermost Nappes.

3.2 Tripoliza Nappe

3.2.1 Tripoliza Limestone

Outcrop I: Tripoliza Unit, limestone displaying karst weathering (see arrow).

Outcrop I, closeup: Tripoliza Unit, limestone

Outcrop I: Tripoliza Unit, limestone displaying numerous bivalve shells and other fossils

Outcrop I: Tripoliza Unit, limestone. Weathering has revealed various carbonate clasts orientated along the bedding plane.

3.2.2 Tripoliza Flysch

The Tripoliza limestone grades upward into Tripolitza flysch consisting of Eocene sandstone and minor conglomerates. It was deposited by submarine landslides (turbidites) in a subsiding basin. The siliciclastic sediments indicate 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 typically on the order of 100 to 200 m thick. The entire thickness of the Tripolitza is about 700 m thick. Typical features of flysch are graded bedding and rhythmically interbedded sandstones and shales.

Flysch results initially from tectonic activity such as earth quakes that dislodge thick instable sediments from the continental shelf. Tremors trigger the transport of instable sediments as turbidity currents that flow rapidly down continental slopes into deep marine basins. These currents carry a mix of sand, silt, clay, and rock fragments, which settle out in graded layers. Between turbidity currents the background sedimentation of the deep marine environment continues often forming shale layers and the cyclic nature of a flysch sequence.

Outcrop II: Transition from Tripoliza limestone to Tripoliza flysch. 1: Tripoliza flysch, 2: Tripoliza limestone slope debris.

Outcrop II, Tripoliza flysch: cyclic sequence of sandstone and shale

Outcrop II, Tripoliza flysch: shale interbeds representing background sedimentation or a distal part of turbidite fan.

Outcrop II, Tripoliza flysch: Sample displaying two different sandstones at a possible erosional contact. 1: sandstone with calcite matrix, 2: coarse sandstone, 3: erosional contact, 4: possible plant remains

Outcrop II: Tripoliza flysch, bed consisting of coarse sandstone (possibly grauwacke).

A typical sandstone formed by turbidity currents is graywacke. It often displays graded bedding—coarser materials at the bottom and finer grains at the top. Other features are angular fragments of quartz, feldspar, and lithic debris embedded in a fine matrix. The presents of feldspar indicates that the sediment was rapidly eroded and deposited, with minimal chemical weathering**.** This is because feldspar is less chemically stable than quartz and weathers quickly. Similarly, lithic grains (i.e. pieces of pre-existing rocks) in graywacke are a sign of the immaturity and proximity to the source. See My GeoGuide No. 13 "Tripoliza Flysch and Pindos Nappe at Chametoulo" for further details on flysch.

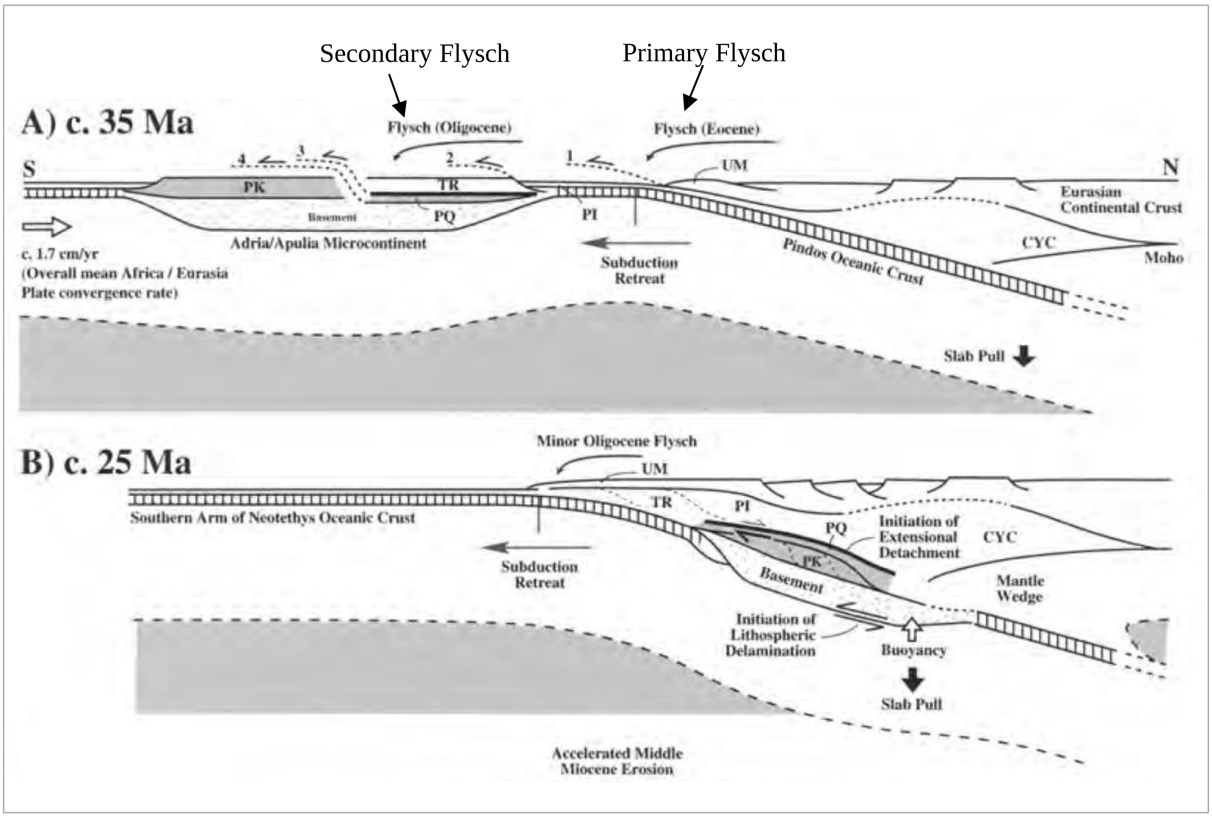
3.3 The Pindos Nappe

Continuing downhill (and upsection), one encounters cyclic Primary Pindos Flysch sediments consisting of red chert, limestone and red shale beds. These are followed by Jurassic red ribbon cherts of the Pindos nappe. The Pindos unit contains radiolarites and chert with shale interbeds, which is exposed for about 150m along the road. At the end of the sequence the ribbon cherts grades into Secondary Pindos flysch, which consists here of about 75 m of siliciclastic sandstone, shale, and minor ribbon chert.

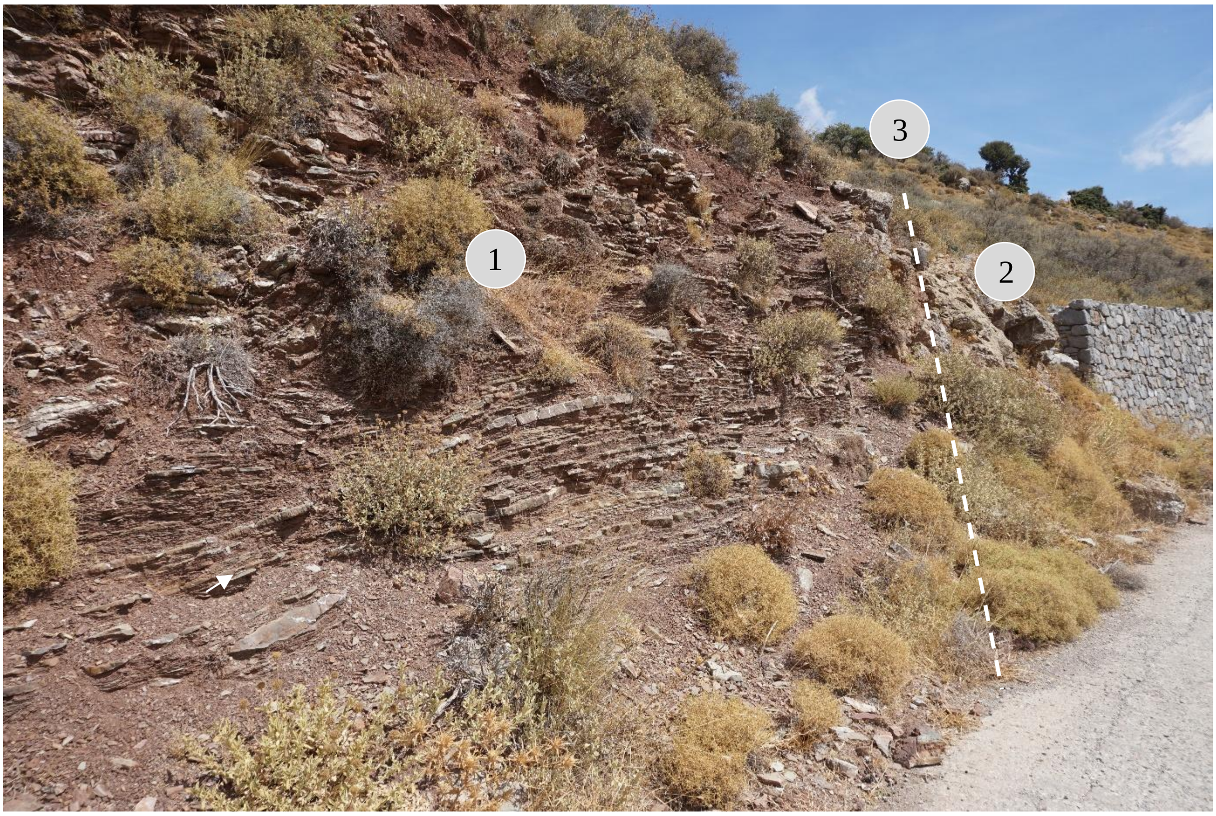
3.3.1 Primary Pindos Flysch vs. Secondary Pindos Flysch

Primary Pindos flysch is typically found beneath the Pindos nappe and represents the initial sedimentation in the Pindos Ocean basin. Secondary flysch is younger and was deposited during thrusting events. It is often found between nappes of the Tripolitza and Plattenkalk Units

Feature	**Primary Pindos Flysch**	**Secondary Pindos Flysch**
Stratigraphic Position	Lower flysch unit	Upper flysch unit
Age	Late Cretaceous to Paleocene	Eocene to Oligocene
Depositional Environment	Deep marine basin	Foreland basin during nappe thrusting
Tectonic Role	Records early convergence and basin formation	Records nappe emplacement and deformation
Examples on Crete	Flysch beneath the Pindos nappe	Flysch between Tripolitza and Plattenkalk nappes



Thomson, S.N., Stockhert, B. & Brix, M.R., 1999: PK: Plattenkalk Unit, PQ: Pyllite -Quarzite Unit, TR: Tripoliza Unit, PI: Pindos Unit, UM: Uppermost Unit.



Outcrop III: Primary Pindos Flysch. Bonneau interprets this as "primary flysch" that was deposited *during the Cenomanian. 1: Pindos Flysch, 2: Tripoliza flysch, 3: major fault displacing the younger Primary Pindos flysch on top of the Tripoliza nappe.*

Outcrop III: Primary Pindos Flysch, displaying rhythmic deposition of chert, shale, sandstone and limestone

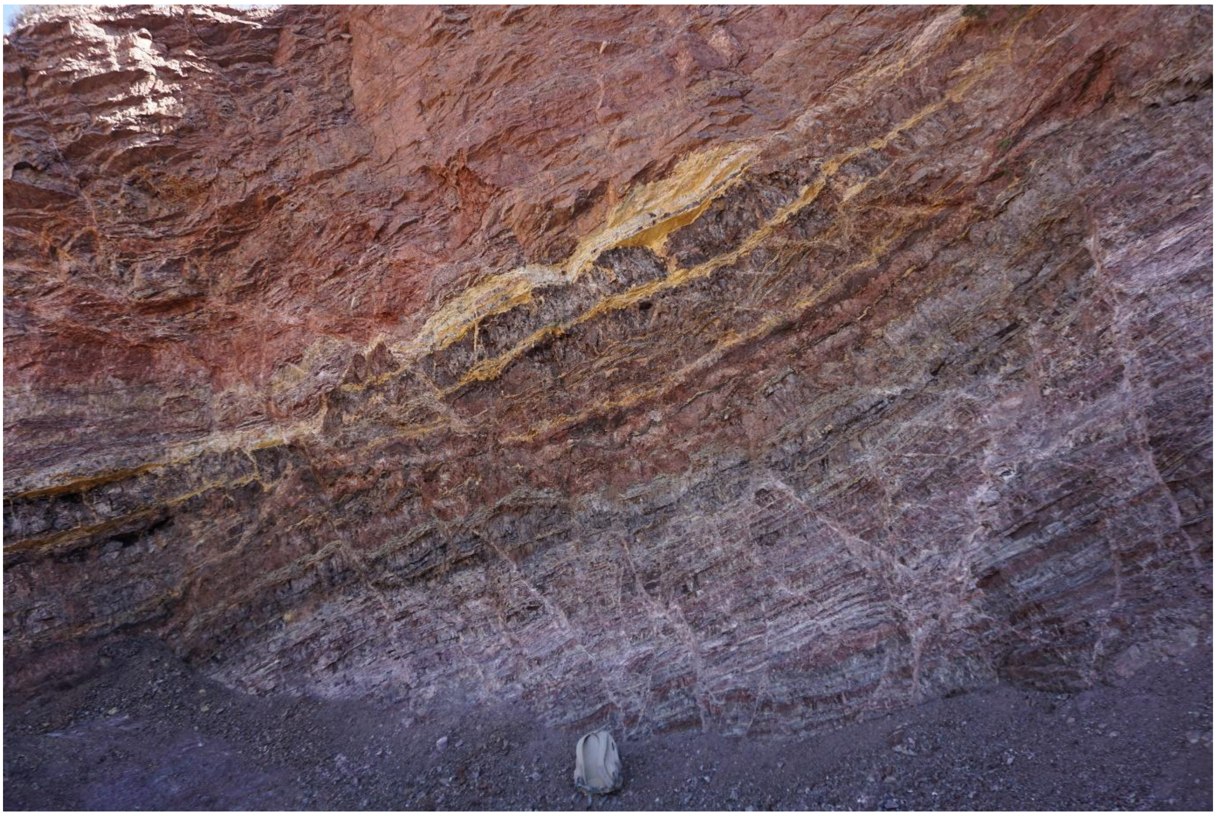
Outcrop III, Pindos Flysch, closeup of previous picture: chert bed

Outcrop III: Primary Pindos Flysch, close up of sandstone bed. Sandstone with calcite matrix

Outcrop III: Primary Pindos flysch, close up of limestone bed

3.3.2 Pindos Chert and Radiolarites

Outcrop IV: Pindos Unit, chert beds

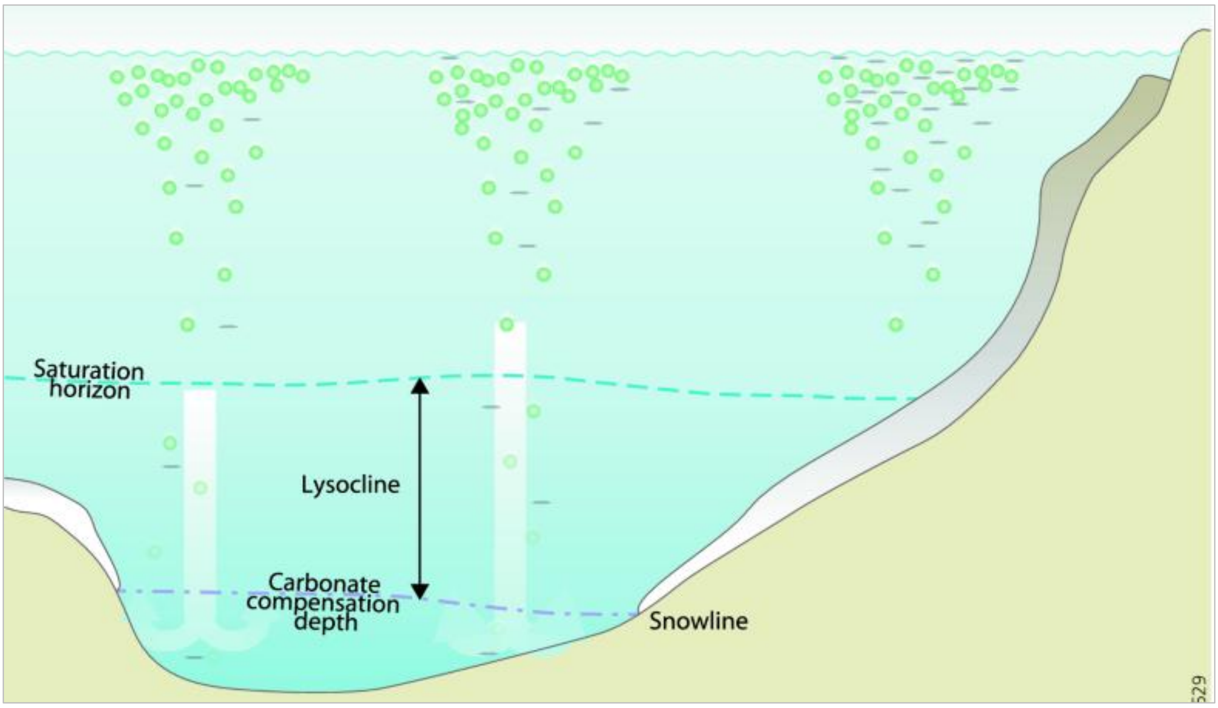


Outcrop IV: Pindos Unit, chert beds

Outcrop IV: Pindos Unit, chert beds

Note that other locations on Crete for example at the Kredos Mountain and the coast line between Souda and Rodakino in Central Crete display extensive sequences of Pindos limestone. The limestone is usually creamy white and sometimes slightly pink.

Carbonate compensation depth

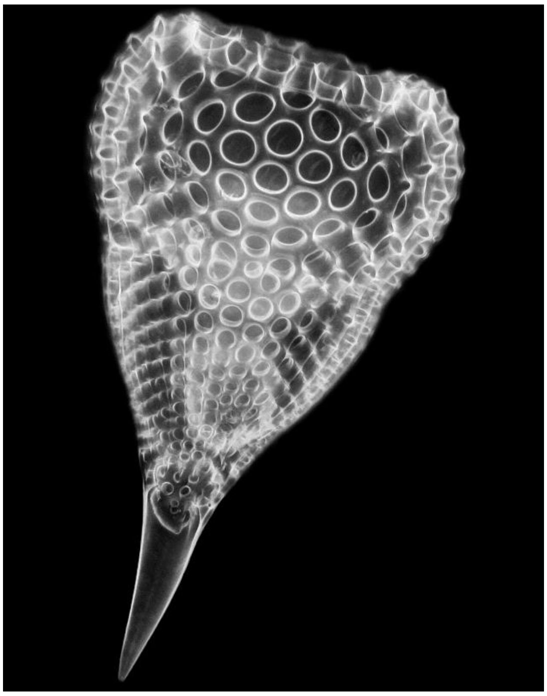


Carbonate compensation depth and the lysocline range [Source: Wikipedia, Carbonate compensation depth - Wikipedia]

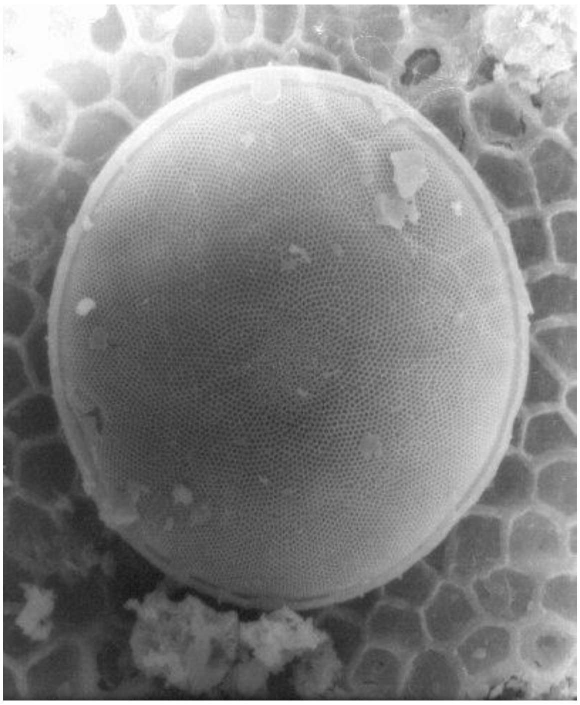
Above the carbonate compensation depth (CCD) calcium carbonate from shells and skeletons of marine organisms (like foraminifera and coccolithophores) settles and accumulates on the seafloor. At the CCD the rate of carbonate supply equals the rate of dissolution. Below the CCD cold temperatures, high pressure, and elevated CO_2 levels increase carbonate solubility, causing complete dissolution of $CaCO_3$ before it reaches the seafloor. The CCD varies by region—deeper in the Atlantic (~5000 m), shallower in the Pacific (~4000 m).

Siliceous ooze

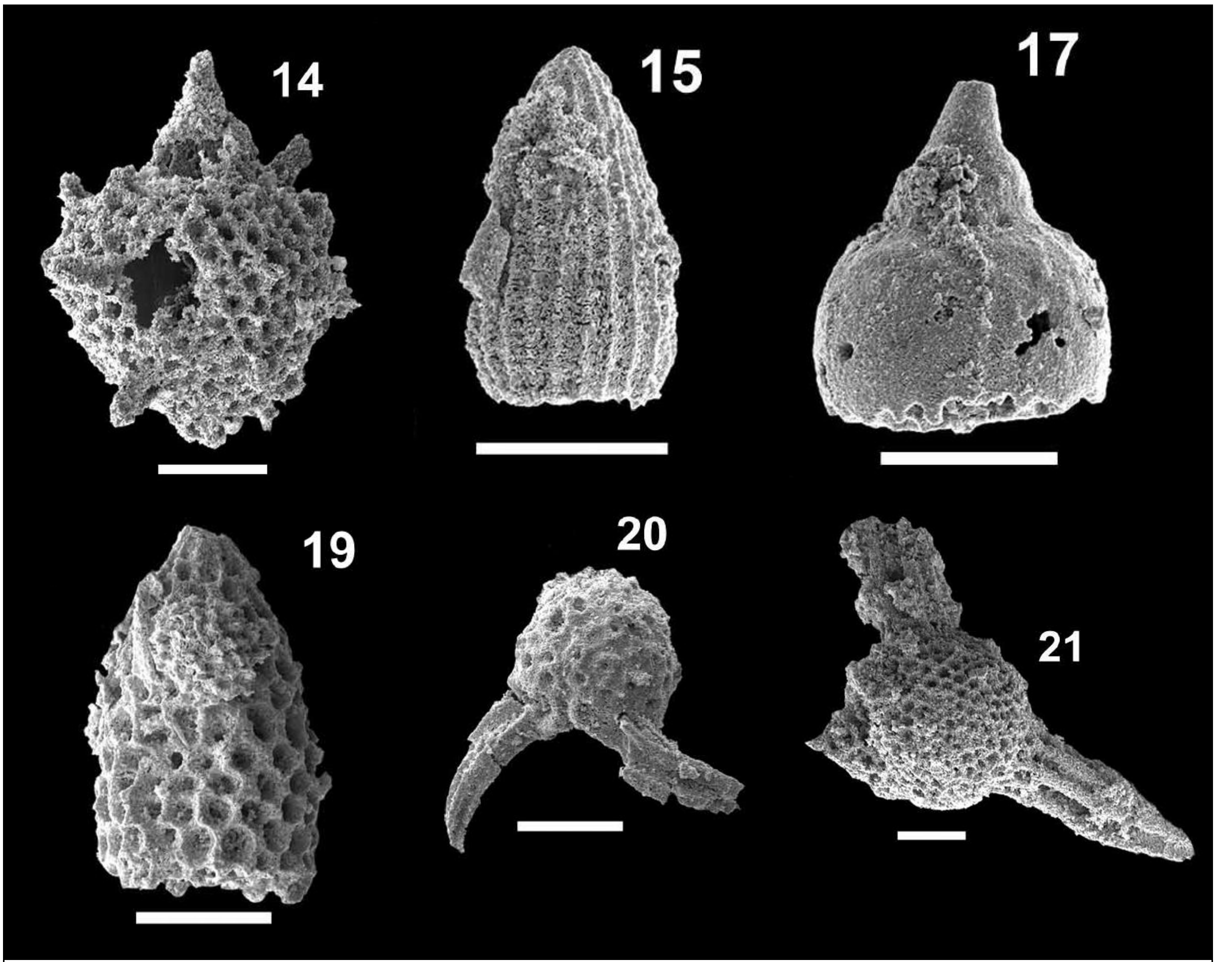
Siliceous ooze is a type of biogenic pelagic sediment located on the deep ocean floor, which can form besides other locations below the CCD. Siliceous oozes are the least common of the deep sea sediments, and make up approximately 15% of the ocean floor. Oozes are defined as sediments which contain at least 30% skeletal remains of pelagic microorganisms. Siliceous oozes are largely composed of the silica based skeletons of microscopic marine organisms such as diatoms and radiolarians. Other components of siliceous oozes near continental margins may include terrestrially derived silica particles and sponge spicules. Siliceous oozes are composed of skeletons made from opal silica $SiO_2 \cdot nH_2O$, as opposed to calcareous oozes, which are made from skeletons of calcium carbonate ($CaCO_3 \cdot nH_2O$) organisms (i.e. coccolithophores). Silicon (Si) is a bioessential element and is efficiently recycled in the marine environment through the silica cycle. Distance from land masses, water depth and ocean fertility are all factors that affect the opal silica content in seawater and the presence of siliceous oozes. *[Source: Wikipedia, Siliceous ooze - Wikipedia]*



A recent radiolarian, 160x magnified [Source: Wikipedia Siliceous ooze - Wikipedia]



Recent centric diatom, magnified x150 [Wikipedia Siliceous ooze - Wikipedia]



Radiolarians found in samples taken from Pindos Flysch exposed in the area of Etolia (Central Greece). The upper levels of the flysch sediments display debris flows and slide blocks (olistostromes and olistoliths) that exhibit ophiolitic material and radiolarian cherts. Samples have been etched with hydrochloric and hydrofluoric acid at different concentrations. Scale bar=50μ.

Middle-Late Jurassic: 14) Arcanicapsa sp. cf. A. leiostraca (Foreman), VR 1; 15) Archaeodictyomitra sp. cf. A. patricki Kocher, VR1; 17) Eucyrtidiellum unumaense s.l. Yao, VR 12; 19) Pseudodictyomitrella tuscanica (Chiari, Cortese and Marcucci), VR1; 20) Saitoum pagei Pessagno, VR1; 21) Tripocyclia smithi Pessagno and Yang, VR1.

[Botziolis, C. et. al., 2023]

Radiolaria

Fossil evidence suggests that radiolarians first emerged during the late Cambrian as free-floating shallow water organisms. They did not become prominent in the fossil record until the Ordovician. Radiolarites evolved in upwelling regions in areas of high primary productivity and are the oldest known organisms capable of shell secretion. The remains of radiolarians are preserved in chert; a by-product of siliceous ooze transformation. Major speciation events of radiolarians occurred during the Mesozoic. Many of those species are now extinct in the modern ocean. Scientists hypothesize that competition with diatoms for dissolved silica during the Cenozoic is the likely cause for the mass extinction of most radiolarian species.

Diatoms

The oldest well-preserved diatom fossils have been dated to the beginning of the Jurassic period. However, the molecular record suggests diatoms evolved at least 250 million years ago during the Triassic. As new species of diatoms evolved and spread, oceanic silica levels began to decrease. Today, there are an estimated 100,000 species of diatoms, most of which are microscopic (2-200 µm). Some early diatoms were larger, and could be between 0.2 and 22mm in diameter.

The earliest diatoms were radial centrics, and lived in shallow water close to shore. These early diatoms were adapted to live on the benthos, as their outer shells were heavy and prevented them from free-floating. Free-floating diatoms, known as bipolar and multipolar centrics, began evolving approximately 100 million years ago during the Cretaceous. Fossil diatoms are preserved in diatomite (also known as diatomaceous earth), which is one of the by-products of the transformation from ooze to rock formation. As diatomaceous particles began to sink to the ocean floor, carbon and silica were sequestered along continental margins. The carbon sequestered along continental margins has become the major petroleum reserves of today. Diatom evolution marks a time in Earth's geologic history of significant removal of carbon dioxide from the atmosphere while simultaneously increasing atmospheric oxygen levels. *[Source: Wikipedia Siliceous ooze - Wikipedia]*

3.3.3 Secondary Flysch

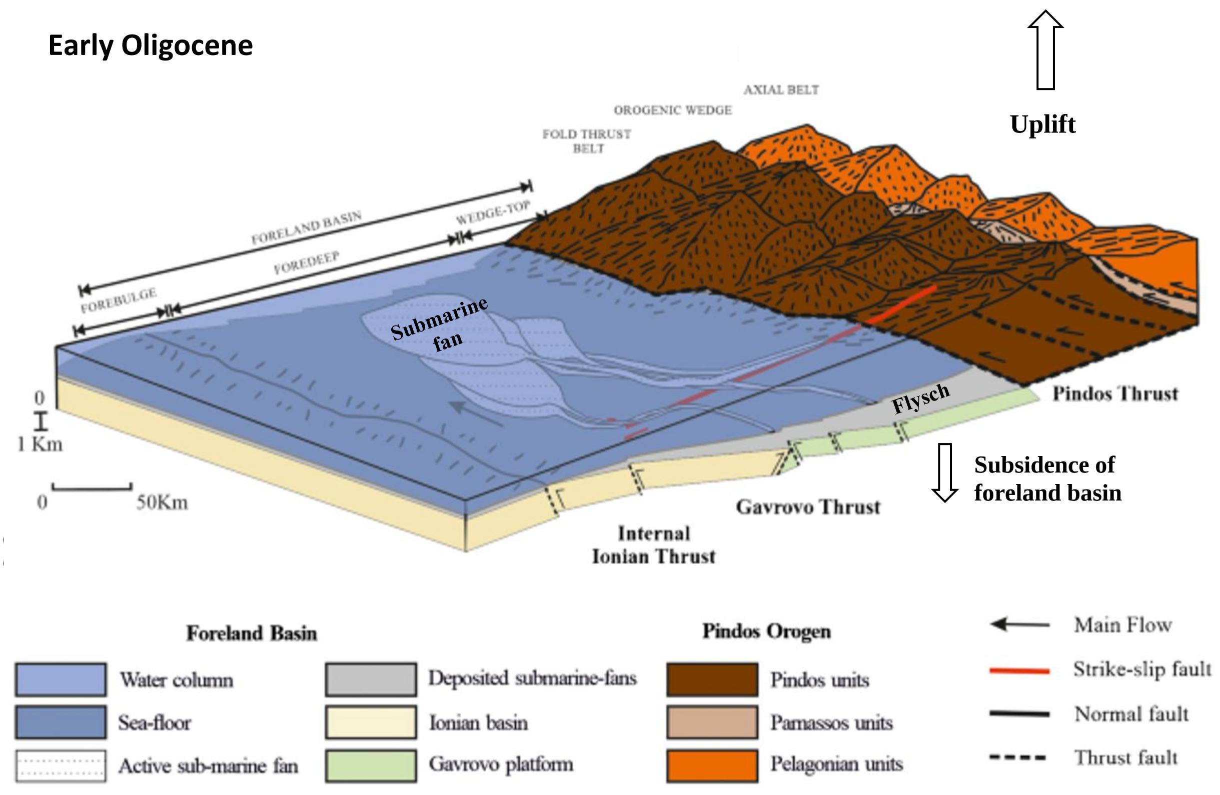
Outcrop V, Secondary Pindos Flysch: grey cherty limestone and red shale beds indicated to be secondary Pindos flysch that was deposited during the Eocene.

Outcrop V, Secondary Pindos Flysch: grey cherty limestone

Outcrop V, Secondary Pindos Flysch: red shale representing the background sedimentation in a deep marine environment.

Outcrop VI, Secondary Pindos Flysch: sandstone bed (grauwacke) thought to be a turbidite deposit.

Outcrop VI, closeup of previous picture. Secondary Pindos Flysch, sandstone sample possibly grauwacke



Model showing how Secondary Pindos Flysch was deposited in the Hellinides (mainland Greece) during the Early Oligocene. The concept is based on a continually subsiding foreland basin in which turbidites were deposited as sub-marine fans. Note that the subduction of the oceanic crust is not shown in this model. In principle this mechanism can be applied to Crete, however, the internal Hellenides (Parnassos and Pelagonian units) are not found on Crete except for the remains of the Uppermost Unit (e.g. Asteroussia nappe). Modified after Botziolis, C. et. al., 2023.

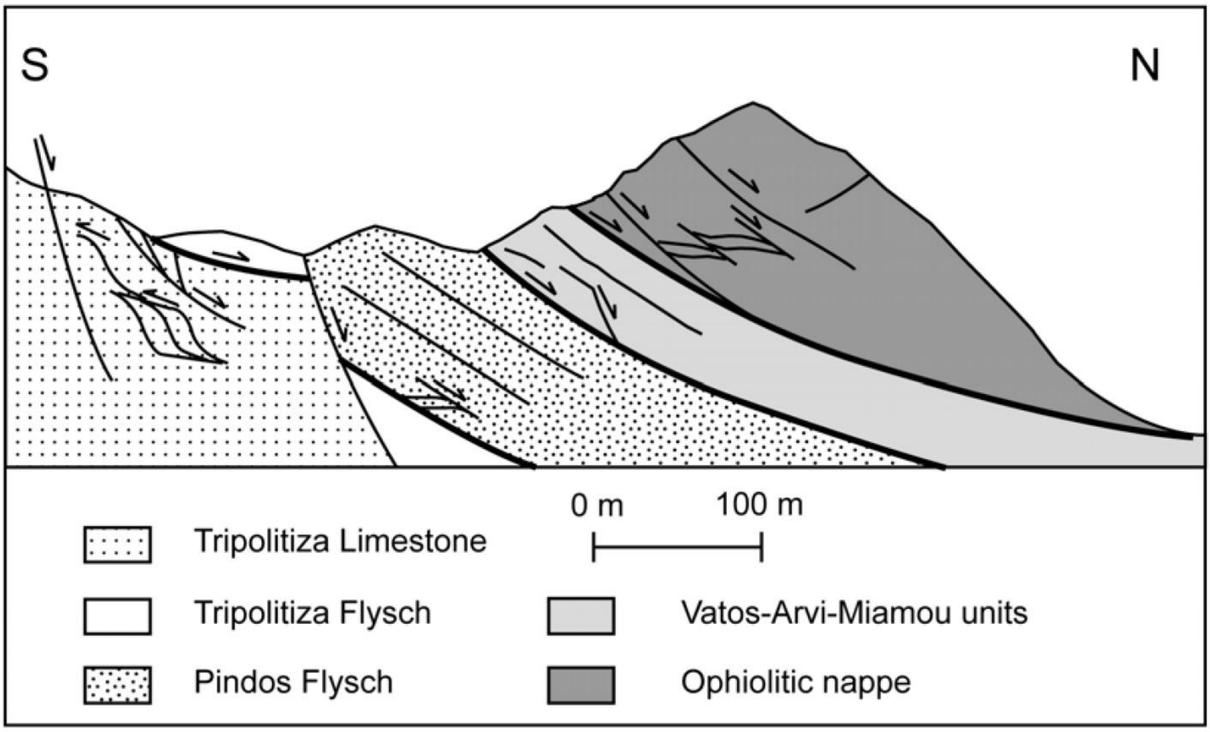
3.4 The Uppermost Nappes

At the windmill a lithologically heterogenous assemblage of rocks characteristic of the Uppermost nappe is exposed. The first exposures are calcareous-siliciclastic sediments of the Vatos subunit. Continuing north along the left fork of the road there are outcrops of serpentizied ultramafic rocks and subordinate mafic rocks. There are 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]*.

3.4.1 Vatos Phyllite

Outcrop VII, Uppermost Nappe: Vatos Phyllite

Outcrop VII: Uppermost Nappe: Vatos Phyllite



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

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5 Appendix

Geological Time Scale

Eonothem/ Eon	Erathem/ Era	System/ Period	Series/ Epoch	Stage/ Age	mya[1]
					2.58
Phanerozoic	Cenozoic	Neogene	Pliocene	Piacenzian	3.600
				Zanclean	5.333
			Miocene	Messinian	7.246
				Tortonian	11.63
				Serravallian	13.82
				Langhian	15.97
				Burdigalian	20.44
				Aquitanian	23.03
		Paleogene	Oligocene	Chattian	27.82
				Rupelian	33.9
			Eocene	Priabonian	37.8
				Bartonian	41.2
				Lutetian	47.8
				Ypresian	56.0
			Paleocene	Thanetian	59.2
				Selandian	61.6
				Danian	66.0
	Mesozoic	Cretaceous	Upper	Maastrichtian	72.1 ± 0.2
				Campanian	83.6 ± 0.2
				Santonian	86.3 ± 0.5
				Coniacian	89.8 ± 0.3
				Turonian	93.9
				Cenomanian	100.5
			Lower	Albian	~113
				Aptian	~125.0
				Barremian	~129.4
				Hauterivian	~132.9
				Valanginian	~139.8
				Berriasian	~145.0

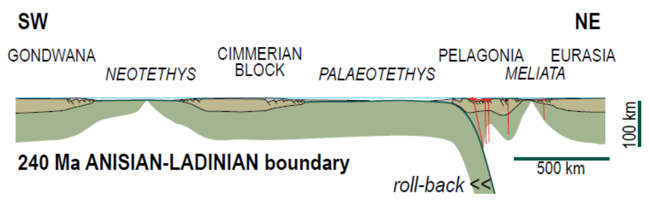
Eonothem/ Eon	Erathem/ Era	System/ Period	Subsystem	Series/ Epoch	Stage/ Age	mya[1]
						~145.0
Phanerozoic	Mesozoic	Jurassic		Upper	Tithonian	152.1 ± 0.9
					Kimmeridgian	157.3 ± 1.0
					Oxfordian	163.5 ± 1.0
				Middle	Callovian	166.1 ± 1.2
					Bathonian	168.3 ± 1.3
					Bajocian	170.3 ± 1.4
					Aalenian	174.1 ± 1.0
				Lower	Toarcian	182.7 ± 0.7
					Pliensbachian	190.8 ± 1.0
					Sinemurian	199.3 ± 0.3
					Hettangian	201.3 ± 0.2
		Triassic		Upper	Rhaetian	~208.5
					Norian	~227.0
					Carnian	~237.0
				Middle	Ladinian	~242.0
					Anisian	247.2
				Lower	Olenekian	251.2
					Induan	251.902 ± 0.024
	Paleozoic	Permian		Lopingian	Changhsingian	254.14 ± 0.7
					Wuchiapingian	259.1 ± 0.5
				Guadalupian	Capitanian	265.1 ± 0.4
					Wordian	268.8 ± 0.5
					Roadian	272.95 ± 0.11
				Cisuralian	Kungurian	283.5 ± 0.6
					Artinskian	290.1 ± 0.26
					Sakmarian	295.0 ± 0.18
					Asselian	298.9 ± 0.15
		Carboniferous	Pennsylvanian[2]	Upper	Gzhelian	303.7 ± 0.1
					Kasimovian	307.0 ± 0.1
				Middle	Moscovian	315.2 ± 0.2
				Lower	Bashkirian	323.2 ± 0.4
			Mississippian[2]	Upper	Serpukhovian	330.9 ± 0.2
				Middle	Visean	346.7 ± 0.4
				Lower	Tournaisian	358.9 ± 0.4

Geodynamic Model by Stampfli

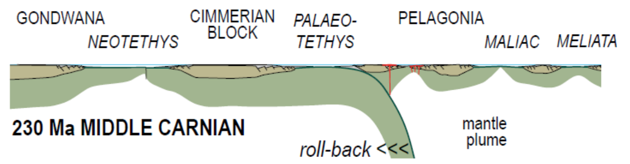
For numerous years, the group headed by Prof. Stampfli at the university of Lausanne has been working on plates tectonic reconstruction models. The models are constantly evolving due to the addition of new data.

Breakup of Pangea and the Cimmerian Collision

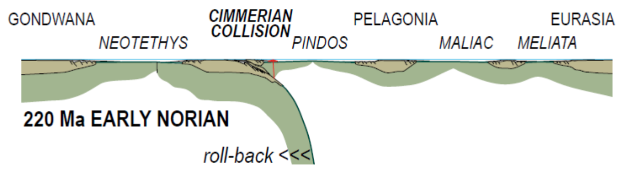
Since the mid Permian (Roadian), the Paleotethys was being subducted northwards under the Eurasia plate. By the Lower Triassic the Cimmerian terrane had already detached itself from the Gondwana northern margin and in doing so initiated the opening of the Neotethys. At the same time Southern Eurasia was strongly subjected to extensional stress due to the roll-back of the Paleothetys oceanic plate during subduction. This caused the opening of a number of back-arc basins starting with the Küre basin in the East. During the Middle Triassic, the Meliata and the Maliac basins opened causing the detachment of the Pelagonian terrane. Owing to the northward subduction of Paleothetys oceanic plate, and the resulting Cimmerian collision at the beginning of Late Triassic, the western part of Neotethys stopped spreading. At the same time, the Pindos back-arc basin opened within the southern Pelagonian margin. Between middle Carnian and early Norian, the Paleotethys was completely subducted and the Cimmerian collision had taken effect on the whole of the western Tethys. The Neotethys started to subduct under the Huğlu back-arc, but when its ocean ridge reached the subduction trench, the subduction was halted and transferred to the North by transform faults, leaving the Huğlu basin open. Further north, the intra-oceanic subduction of the Maliac ocean under the Izmir-Ankara ocean began, which triggered the formation of Vardar supra-subduction-zone (SSZ) since the Rhaetian *[Stampfli, 2010]*.



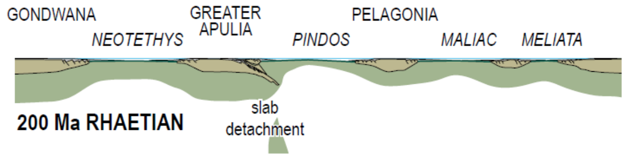
During the Anisian-Ladinian the Paleotethys was subducting northwards against Greater Pelagonia causing lower plate extension, which was offset by spreading of the Neotethys ridge. Additionally, the Paleotethyan slab was rolling back, inducing extension in the upper plate too. The Meliata back-arc basin was already spreading separating the Pelagonian microcontinent from the Eurasian margin, whereas the Maliac basin had just started its synrift stage.



Because of the Cimmerian collision further Eastward of this section (see section line on the following figures) the Neotethys ridge in this region stopped spreading. However, extension is still active in the back-arc realm, where the spreading ridge switched from the Maliac basin to the beginnings of the newly created Pindos basin (see next Figure).

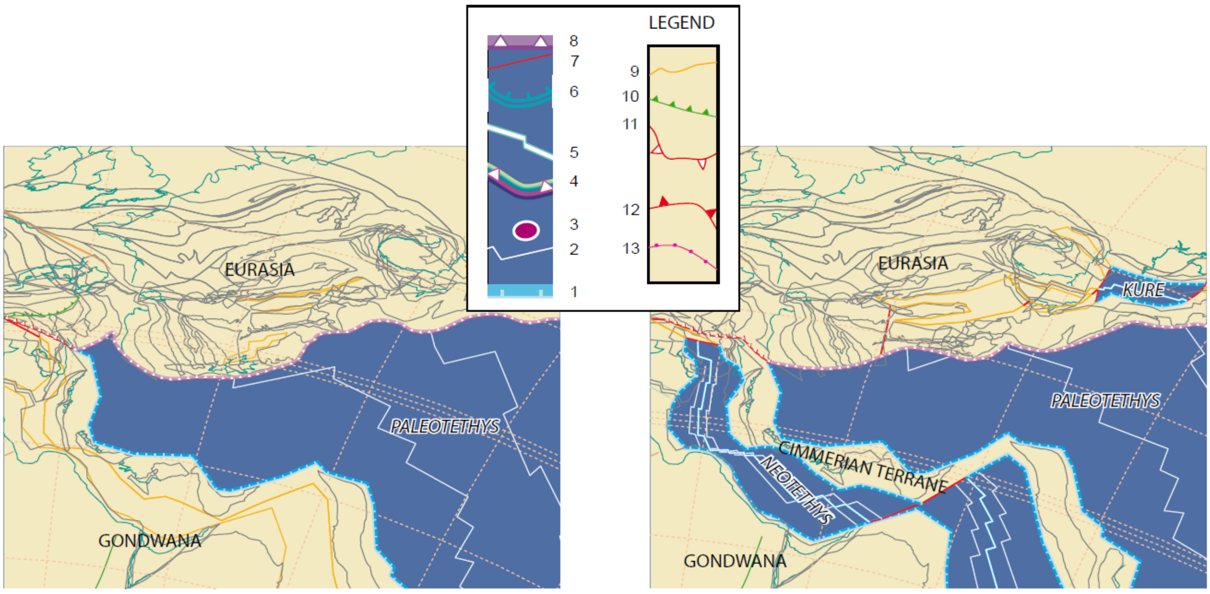


By the early Norian the Paleotethys ocean and its arc was completely consumed and collided with the Cimmerian block causing a phase of mountain building called the Cimmerian collision. The Pindos ridge stopped spreading

	and sedimentation within the basin becomes pelagic. At about 210Ma, the Paleotethyan subducted slab detaches and the Pindos basin started to subside as a whole.
	*During the Rhaetian the thermal subsidence stage of the Pindos basin continued accompanied by the deposition of the Drimos Formation. [Stampfli, 2010]*

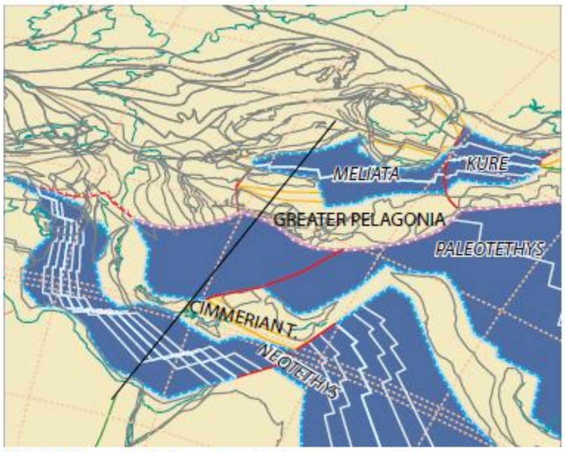
The Eo-Cimmerian is a term used in geology to describe a period of time. It refers to the Late Triassic period, which occurred between approximately 237 and 201 million years ago. During this period, the Cimmerian orogeny took place, which was a series of tectonic events that led to the collision of the Gondwana-derived Cimmerian terrane with Eurasia. The Eo-Cimmerian deformation and metamorphism was very weak and was largely overprinted during Alpine subduction and collision. *[AI: Bing]*

Today, the Cimmerian fold belt is a string of folded mountains that are integrated into the Alpine Mountain belt, which stretches from the Balkans to south-east Asia. However, the Cimmerian orogeny took place more than 100 million years before Alpine mountain building. The Cimmerian terrane split off from the northern edge of south-east Pangaea (Gondwana) in the Permian period. The fragment rotated like the hand of a clock, but anti-clockwise, around a point at its north-western end, which was located in the area of today's Carpathian Mountains. The western Cimmerian mountains contain predominantly overprinted Variscan and Cadomian crust. *[Wikipedia]*

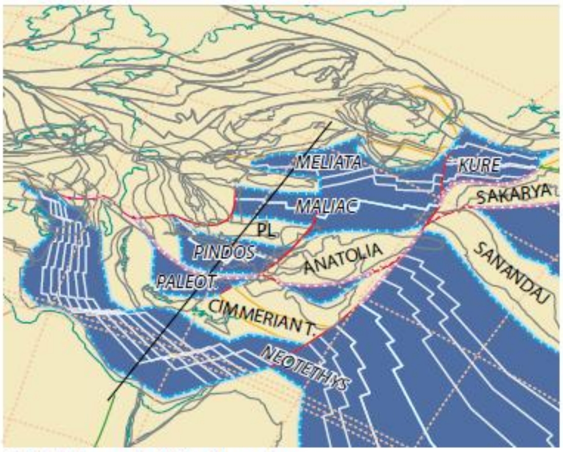


270 Ma - Roadian 250 Ma - Olenekian

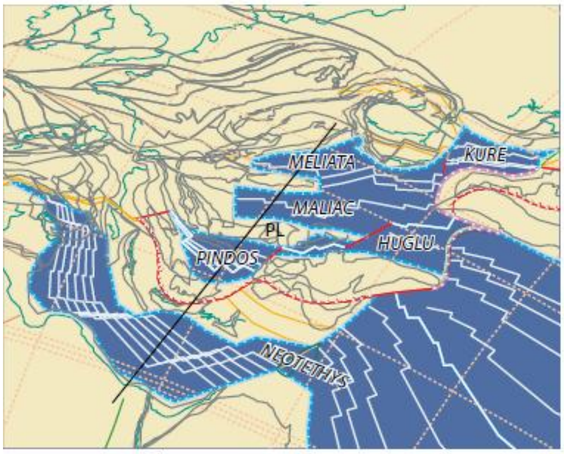
Plates tectonic model from Middle Permian to Late Triassic. modified after Stampfli & Hochard (2009) and Moix (2010). 1. passive margin; 2. magnetic anomalies or synthetic anomalies; 3. seamount; 4. intraoceanic subduction/arc complex; 5. spreading ridges; 6. obductions; 7. Transform and strike-slip faulting; 8. subduction zone; 9. rifts; 10. inversion zones; 11. collision zones; 12. activethrusts; 13. sutures. (PL = Pelagonia). The black lines represent the section



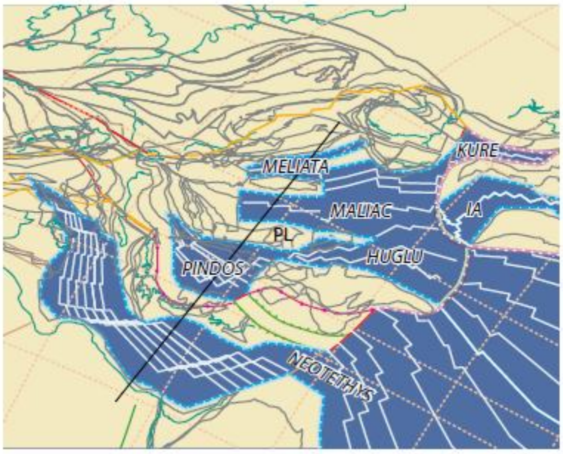
240 Ma - Anisian-Ladinian boundary



230 Ma - middle Carnian



220 Ma - early Norian

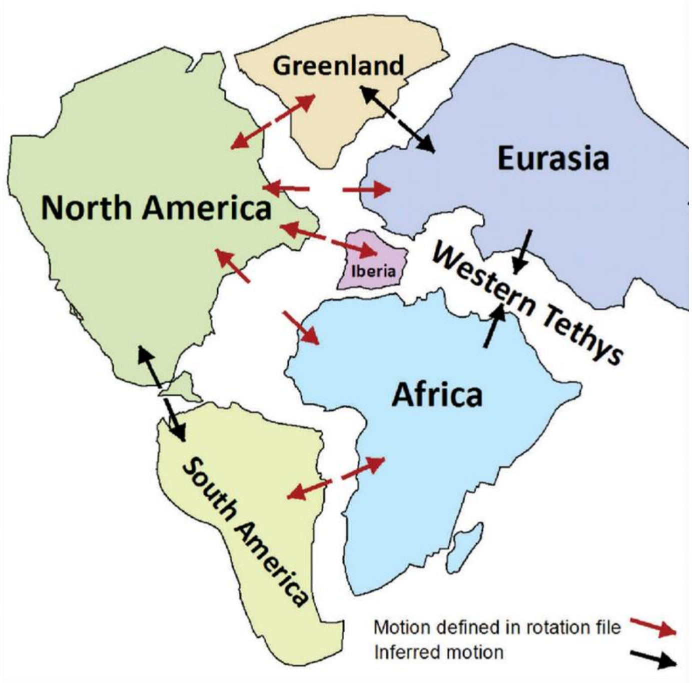


200 Ma - Rhaetian

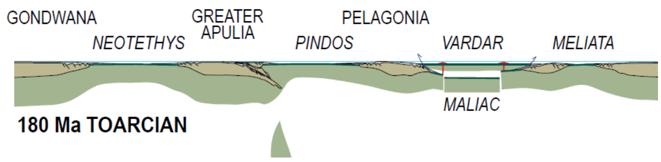
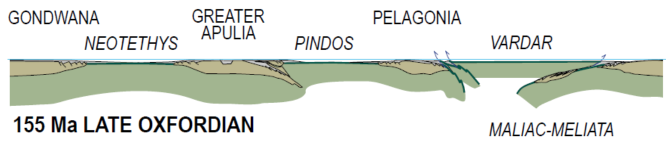
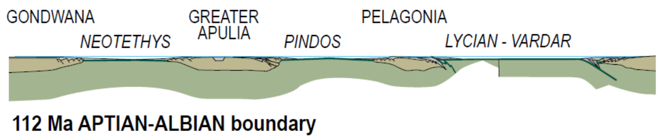
[Source: Stampfli, 2010]

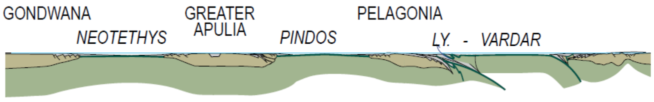
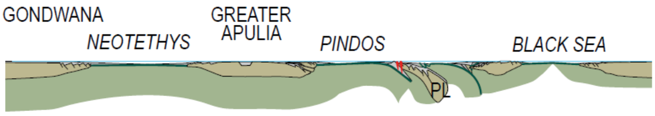
Closure of the Vadar and Pindos Basins and the Alpine Orogeny

During the Middle Jurassic the Vardar basin completely replaced the previous back-arc basins and at the same time its ridge was obducted (i.e. was pushed upwards) onto the Pelagonian microcontinent from East to West. In the late Oxfordian a new ocean ridge was formed between the Huğlu and Vardar basins. The newly formed Lycian basin was part of Vardar basin. Towards the end of Early Cretaceous, the Vardar oceanic crust was subducted northward under the Rhodope inducing the opening of the Black Sea. Simultaneously, the Huğlu subducted under the continuation of the Neothethyan trench causing the obduction of the Lycian slab in Cenomanian-Turonian time. In the East the Lycian nappes stopped at the Anatolian margin. In contrast, in the West, the Pelagonian microcontinent was significantly extended which enabled the propagation of the arc with the obduction front into the Pindos basin (subduction propagation). At the end of the Cretaceous the Vardar supra-subduction-zone (SSZ) was closed and the Pindos slab (consisting of oceanic crust) was subducted toward the East. During the Tertiary the geometry of the Hellenides underwent no major changes. After the subduction and sealing of the Pindos (ocean), the geodynamics was controlled by the Neotethyan (Eastern Mediterranean) subduction, which initiated significant extension within the whole Hellenic orogenic wedge.

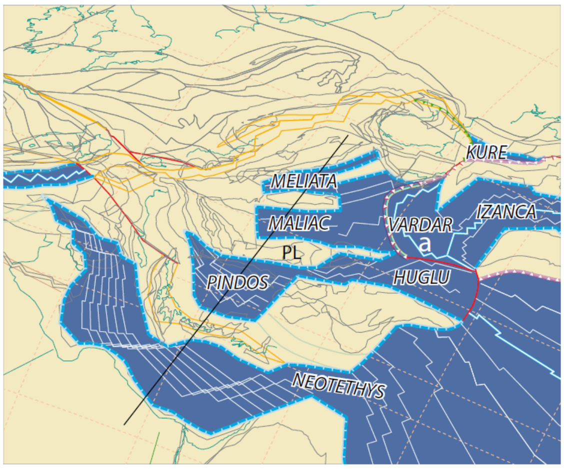


Schematic illustration of the Western Tethys region in the framework of surrounding plates from the Jurassic onwards [Hosseinpour M, et al. 2016.]

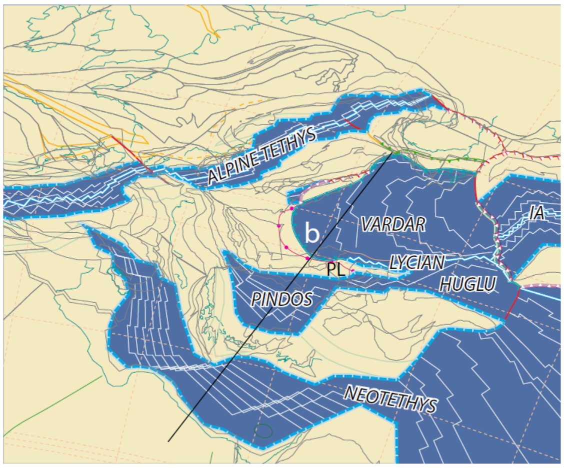
	*Sedimentation into the Pindos is going on with a continuous evolution to more pelagic facies. The radiolarites mark the passage from the Drimos to the Lesteena Fm. North of Pelagonia, the Maliac subducts toward the East providing room for the supra-subduction Vardar ocean.*
	*The deep-sea Lesteena Fm is still depositing in the Pindos domain. Maliac and Meliata are completely subducted, when the Vardar ridge diachronically obducts along the Pelagonian northern margin from the East to the West, forming the largest ophiolitic belt of the Hellenides. Elements of the lower plate like oceanic seamounts or slices of the margin succession are scraped and carried with the Vardar nappe. This event causes an important flexuration marked by the deposition of Oxfordian-Kimmeridgian radiolarites.*
	*The Lambia Fm is forming in the Pindos basin. Sediments are mostly pelagic limestones, Degnan & Robertson (1998) invoke the presence of strong current because of the reworking of intrabasinal deposits in calc-rudites and calc-arenites. North of Pelagonia, the passive margin is reset and the young Lycian ridge is spreading. The remnant Vardar oceanic floor subducts toward the North, under*

	the Rhodope, after a subduction reversal event (Bonev & Stampfli, 2003)
 **90 Ma TURONIAN**	*The Lycian ridge is already obducting on the Eastern Pelagonian margin, picking up the Vardar ophiolites again. This event provokes the bulging of the microcontinent, which implies a partial emersion that stops the pelagic deposition. On the other hand, shallow water conditions enhance the development of platforms (i.e. the Arvi creamy limestones). The ophiolitic front is accompanied by an arc, which constitutes a precursor of the Asteroussia s.s., propagating to the Pindos northern margin. The flexural bulge was subject to erosion, which is regarded as the source for the Pindos first flysch. Further-flexuration followed the obduction inducing the deposition of pelagic limestones Turonian to Santonian in age.*
 **70 Ma MAASTRICHTIAN**	*The remaining Vardar Ocean is completely subducted causing the opening of the Black Sea within the Eastern Rhodope and Pontides. Pindos subduction starts a roll-back stage inducing extension on the active margin. Stampfli suggests that the Arvi lavas that were associated with the deposition of Pindos basin red shales and pelagic limestones were emplaced in this context.* *PL = Pelagonia*
66 - 33 Ma PALEOCENE-OLIGOCENE	*During the Paleocene-Oligocene period, the Pindos slab subducts until the collision of the arc with the Greater Apulia terrane. This stage is marked by the continuous deposition of the Pindos second flysch.*

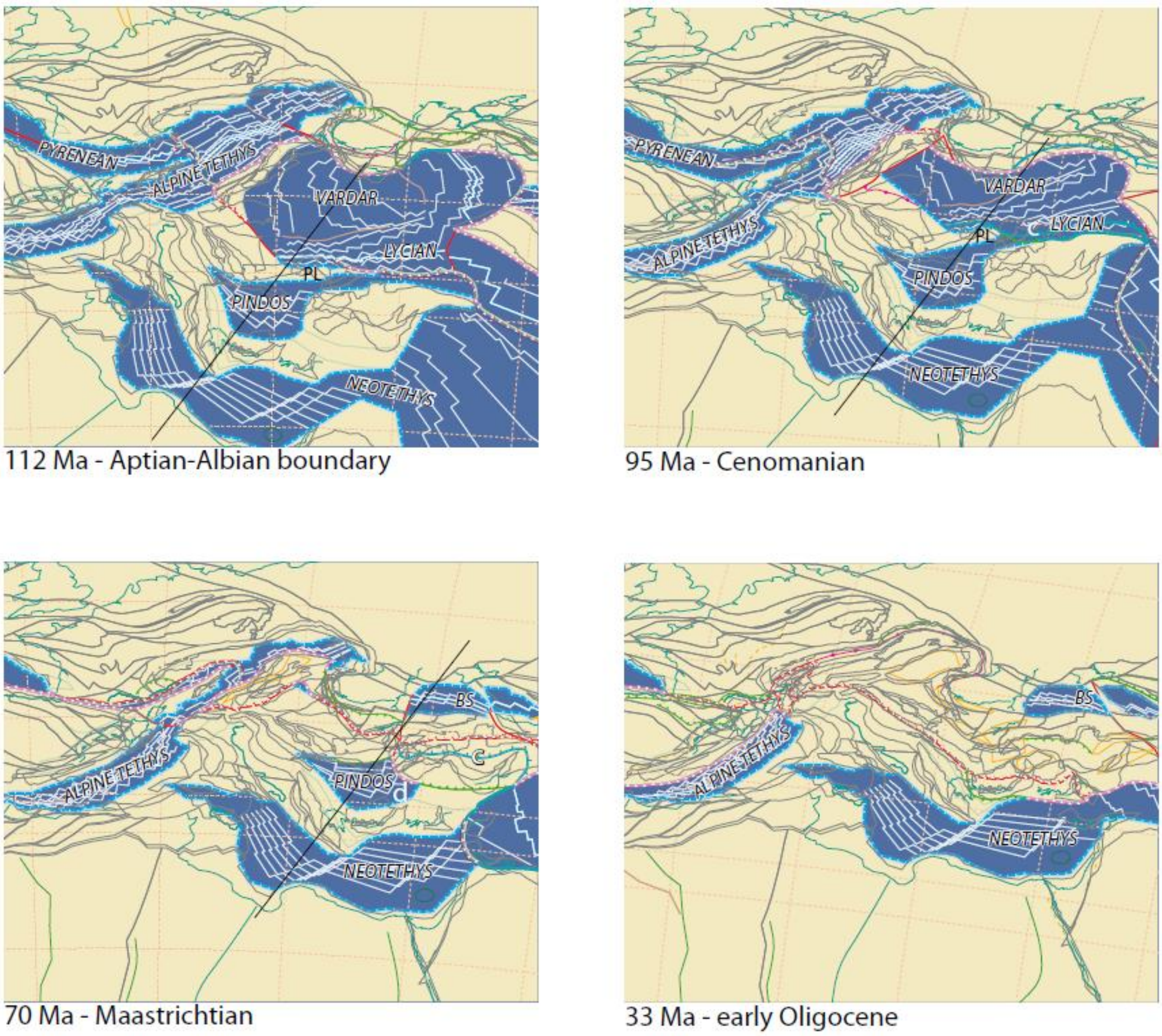
[Source: Stampfli, 2010]



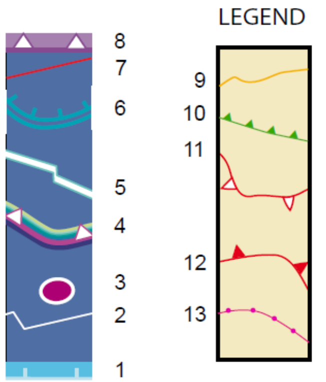
180 Ma - Toarcian



155 Ma - late Oxfordian



Plates tectonic model from Early Jurassic to Oligocene modified after Stampfli & Hochard (2009) and Moix (2010). ***a)*** *Vardar SSZ spreading;* ***b)*** *Vardar obduction;* ***c)*** *Lycian obduction;* ***d)*** *arc propagation into the Pindos (IA = Izmir-Ankara, BS = Black Sea).*



1 passive margin; 2. magnetic anomalies or synthectic anomalies; 3. seamount; 4. intraoceanic subduction/arc complex; 5. spreading ridges; 6. obductions; 7. Transform and strike-slip faulting; 8. subduction zone; 9. rifts; 10. inversion zones; 11. collision zones; 12. activethrusts; 13. sutures. (PL = Pelagonia).

[Source: Stampfli, 2010]