The Uppermost Nappes: Ophiolites and the Vatos Nappe, Aktounda to Ardaktos

Road junction at the mountain ridge between Ardaktos and Kerames. The road cut reveals an outcrop of serpentinite, which was once ultramafic rock intruded at a mid ocean ridge.

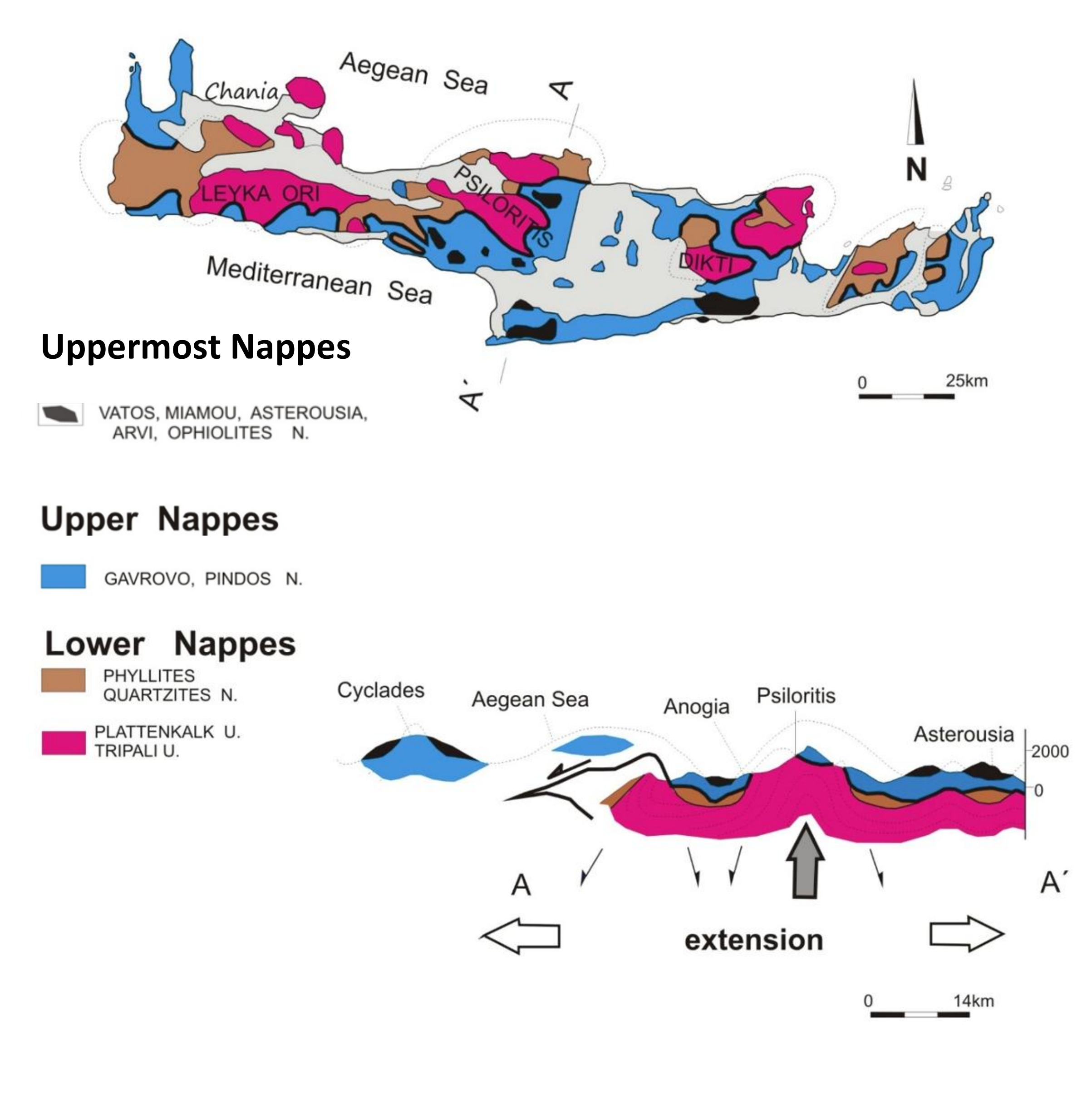
Compiled by George Lindemann, MSc.

Berlin, August 2024

Content

Appendix

1 Introduction



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A nappe or thrust sheet is a large sheetlike body of rock that has been moved tens or even hundreds of kilometres along a thrust plane from its original position. Nappes form in compressional tectonic settings such as continental collision zones or on the overriding plate in active subduction zones. The resulting structure may include large-scale recumbent folds, shearing along the fault plane, imbricate thrust stacks, fensters and klippes *[Wikipedia]*.

A characteristic feature of the Uppermost Nappes of southern and central Crete is the frequent occurrence of ophiolites, which represent pieces of oceanic crust predominantly from mid ocean ridges. Ophiolites consist of mafic and ultramafic plutonic and volcanic rocks and often exhibit deep marine sediments such as clay (i.e. phyllites), chert and deep marine limestones. The ultramafic rock is often hydrothermally altered to serpentinite.

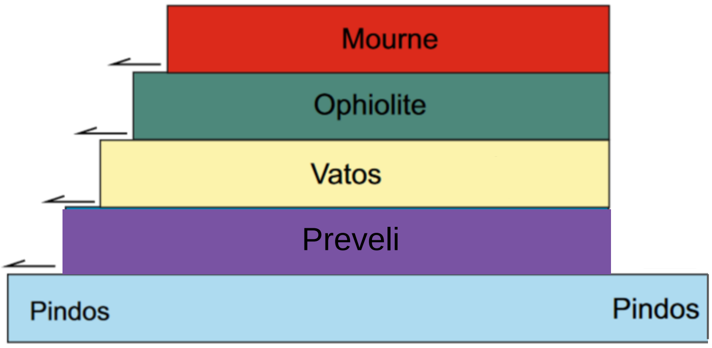
In central Crete, the ophiolite-bearing Uppermost Nappes feature tectonic mélanges that are indicated to be the result of chaotic thrusts within the accretionary wedge of a subduction zone (cf. Cowan, 1985; Hsu, 1968; Raymond, 1984). These assemblages of disrupted rocks, also known as the Cretan mélange (Langosch et al., 2000; Seidel et al., 1977, 1981) represent the highest tectonic nappes, that overly the Tripolitza and the Pindos Nappes. *[Tortorici]*

In the past numerous authors have attempted to group the rocks based on field evidence, petrology, and age into distinct nappes or have presented alternative olistostrome models (See Appendix).

Creutzburg and Seidel (1975) regarded the Uppermost Units as a "serpentinite-amphibolite association" forming a composite nappe. The nappe was thought to consist of ophiolites at the top, and of an assembly of three subunits, at its base made up of very low grade, blue schist, and high-grade metamorphic rock. Hall et al. (1984) interpreted the various rocks of the Uppermost Unit as exotic blocks of varying size within flysch. This was thought to be an olistostrome (i.e. chaotic unconsolidated sediments transported as a semifluid body by submarine gravity sliding or slumping). In contrast to Hall et al. (1984), Bonneau (1984) proposed that rocks of a very low-grade metamorphic subunit formed an "ophiolitic olistostrome" related to the rocks of the Pindos Unit, while mappable blue schist, high-grade metamorphic and ophiolite subunits were considered individual nappes. *[Thomson, 1999]*

Hornblende K-Ar ages from the ophiolite bodies within the Astrousia Mts. that also form the upper part of the Uppermost Unit reveal ages between 135 and 156 Ma (Seidel *et al.* 1977). This provides a Mid-Jurassic minimum age for the formation of these ophiolites. *[Thomson, 1999]*

Recent work by Zulauf et. al. 2023 divides the Uppermost Unit into several separate nappes, which rest on top of the Pindos or Tripoliza units. There are two different stacks one of which is exposed in the Preveli area (south central Crete) and the other in the Astrousia Mountains including the Arvi area. The nappe stack in the area of Vatos–Preveli–Spili–Gerakari consists of the Preveli, Vatos, Ophiolite and Mourne nappes (Zulauf et al. 2023a, see also Field Guide No XY concerning the Preveli Nappe).



Nappe stacks of the Uppermost Unit resting on top of the Pindos Unit of Crete. Zulauf et. al, 2023

Unit/nappe Type	Protolith age	Metamorphism Age	Metamorphism
Mourne	?	Blueschist/Amphibolite (3)	Upper Jurassic (3)
Ophiolite	Middle Jurassic (6)	Greenschist (10)	Upper Jurassic (9)
Vatos	Latest Jurassic - Upper Cretaceous (10, 11)	Greenschist (10, 18)	Upper Cretaceous (10)
Preveli	Late Permian - Upper Triassic (7, 8)	Epidote-Blueschist (12, 13)	Lower Cretaceous (14)

1) Reinecke et al. (1982), (2) Seidel et al. (1976), (3) Seidel et al. (1977), (4) Martha et al. (2017), (5) Martha et al. (2019), (6) Liati et al. (2004), (7) Bonneau and Lys (1978), (8) Zulauf et al. (2023a), (9) Koepke et al. (2002), (10) Malten (2019), (11) Koepke (1986), (12) Koepke et al. (1997), (13) Tortorici et al. (2012), (14) Zulauf et al. in prep., (15) Bonneau et al. (1974), (16) Robert and Bonneau (1982), (17) Palamakumbura et al. (2013), and (18) Karakitzios (1988)

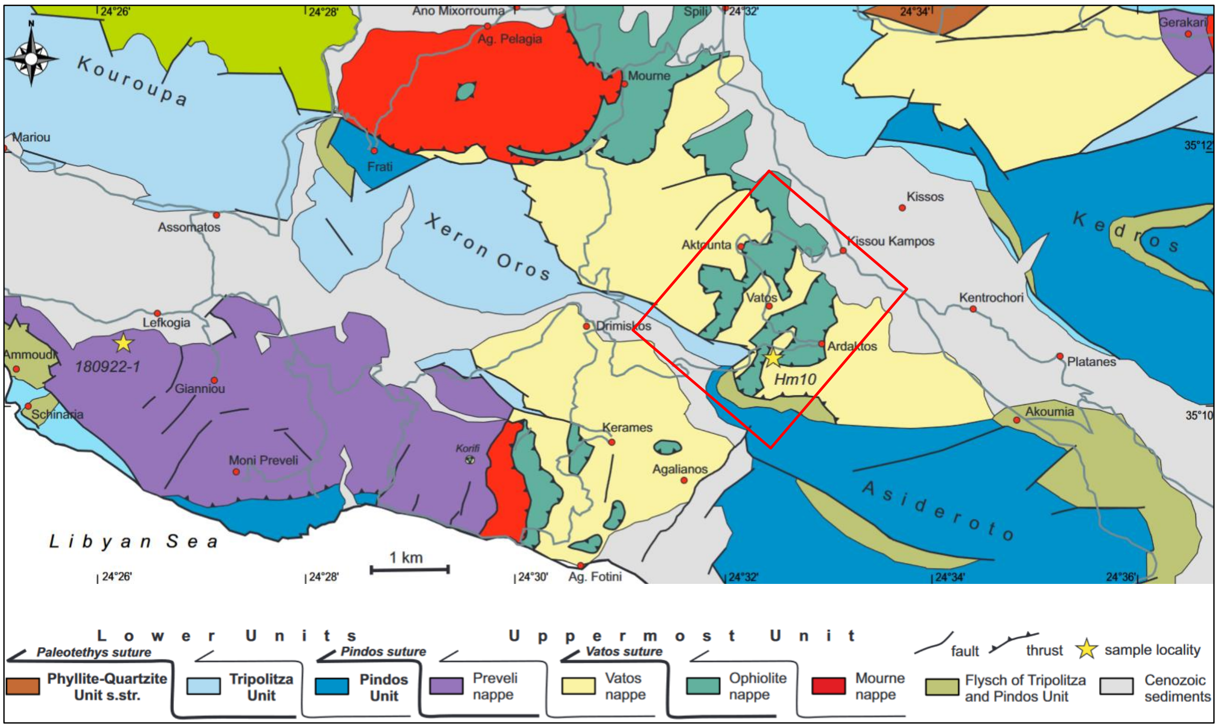
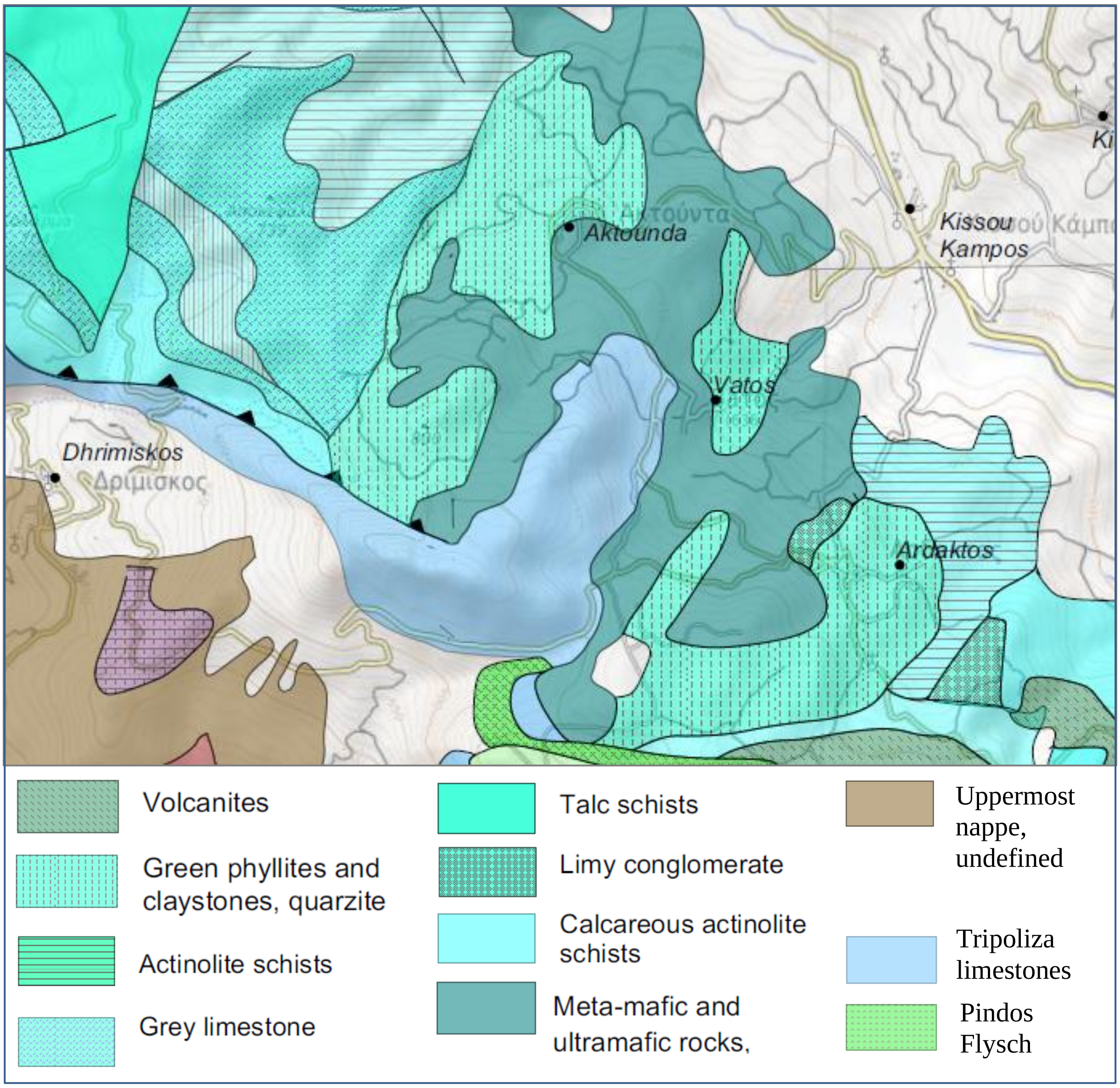


Fig. 2c Supplement. Geological map of the area Spili to Preveli (after Zulauf et al., 2023, and references therein). Red square indicates investigated area of this field guide.



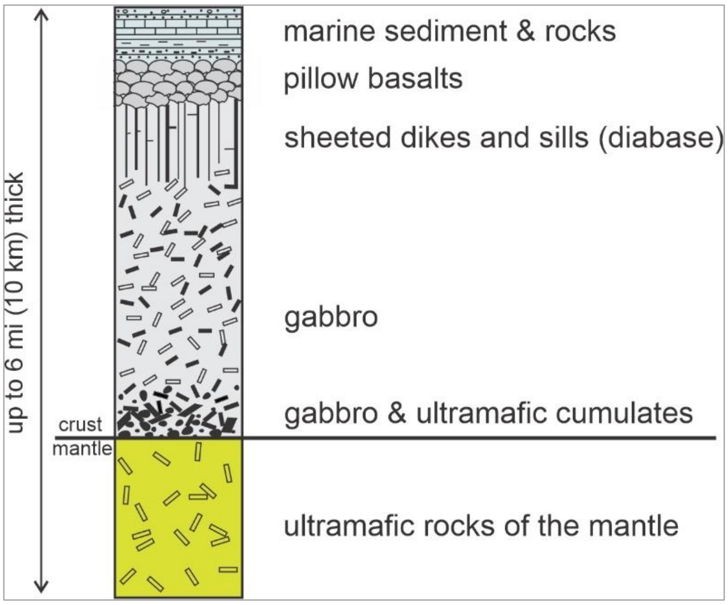
Geological map of the Vatos and Ophiolite nappes (after Krahle 1981, slightly modified)

1.1 Introduction to Ophiolites

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]

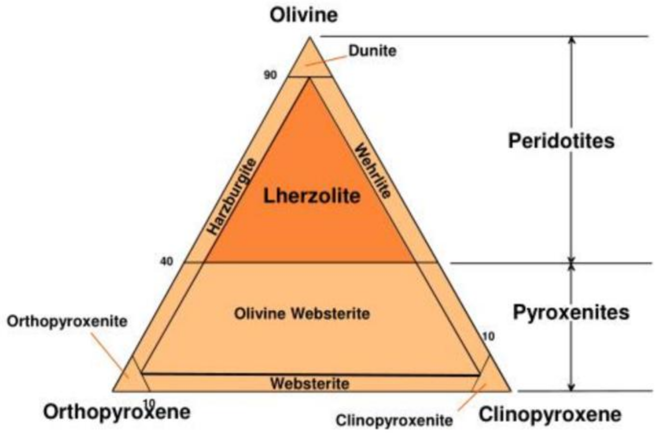
See Appendix

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.

Ophiolites have been identified in most of the world's orogenic belts. However, two components of ophiolite formation are under debate: the origin of the sequence and the mechanism for ophiolite emplacement. Emplacement is the process of the sequence's uplift over lower density continental crust. *[Wikipedia]*

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]*



Ternary diagram that classifies ultramafic rock according to the composition of olivine and pyroxenes. Peridotites, which are rocks containing more than 50 weight % olivine, include

- harzburgite (mostly olivine and orthopyroxene),

- wehrlite (mostly olivine and clinopyroxene), and

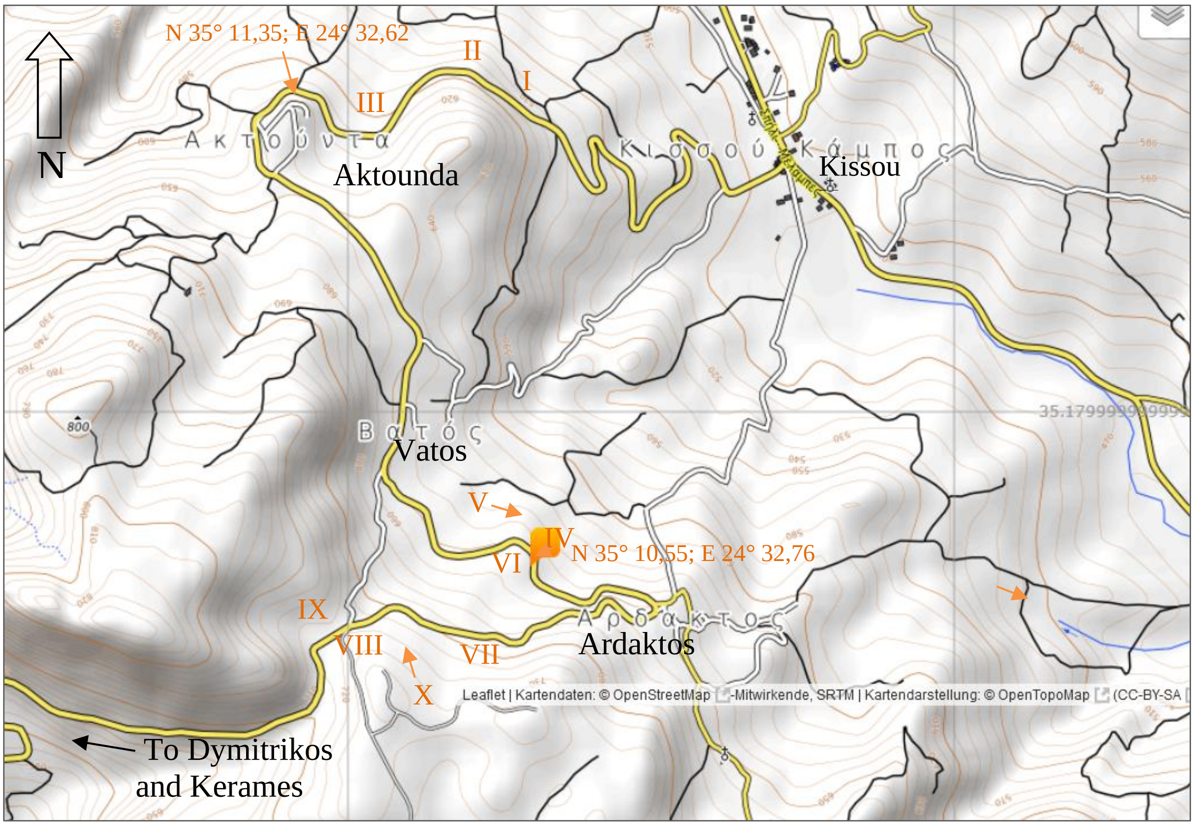
- lherzolites (subequal amounts of orthopyroxene and clinopyroxene).

As inferred from the chemistry of relict orthopyroxene and spinel, these ultramafic rocks may represent primitive, undepleted mantle formed under a slow-spreading midocean ridge. However, such lherzolite-type mantle is also indicated to occur in ophiolites of polygenetic origin, characterized by a long-lasting formation in an arc environment, which is indicated by isotope signatures (e.g., the Trinity ophiolite; Jacobsen et al., 1984; Gruau et al., 1995). *[Koepke J. et. al., 2004]*

2 The Ophiolite Nappe

The Ophiolite nappe at Aktounda to Ardaktos consists deformed meta-ultramafic rocks (serpentinite, meta-hornblendite) and meta-pillow basalt (Krahl et al. 1982). Relics of spinel lherzolite are thought to have originated from an undepleted mantle along a slow spreading oceanic ridge characterized by deformation at very high temperature (Koepke et al. 2002). Further south west of Kerames the serpentinite is overlain by metahornblendite, which was overprinted under greenschist-facies conditions (Koepke et al. 2002). See also My GeoGuide No. 20: The Uppermost Nappes: Ophiolites and the Preveli Nappe, Kerames to Korifi Mountain.

The first outcrops on the road to Aktounda from Kissou Kampos consist of olisthostrome or “wildflysch”. It contains mainly ultramafic clasts. *[Kull]*

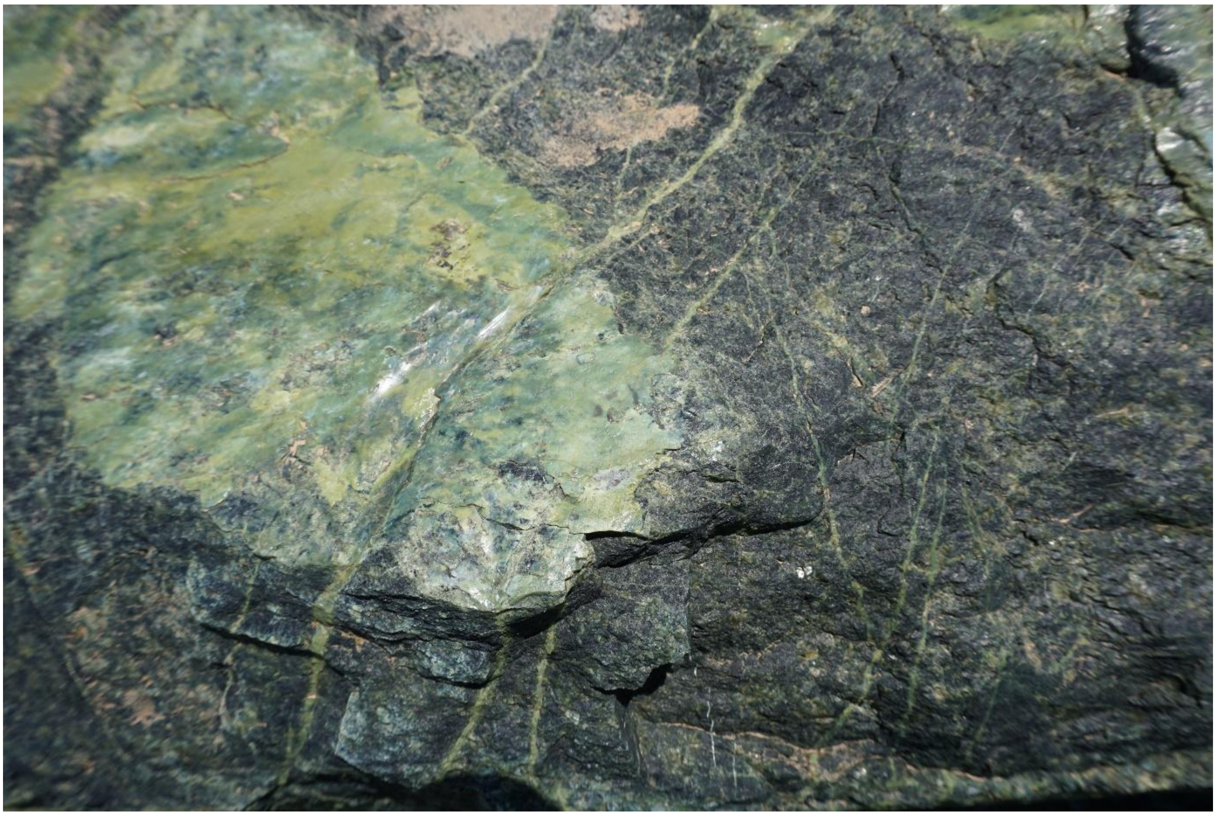


Location of Outcrops

Location of outcrops along the road to Aktounda

Outcrop I: Serpentinized ultramafic rock belonging to the Ophiolite Nappe (N 35° 11,35; E 24° 32,62). Serpentinization is the hydration and metamorphic transformation of ferromagnesian minerals, such as olivine and pyroxene, to produce serpentinite. Tortorici proposes that the ophiolites are tectonic mélanges formed at thrust planes within an accretionary wedge of a subduction zone (see Appendix).

Outcrop I: Serpentinite Boulder (20cm diameter). The intensive fragmentation has produced shear bodies with very smooth surfaces. Individual lherzolite relics with olivine indicate that the source rocks were probably from an undepleted mantle. The age is indicated to be Middle to Late Jurassic (163 ± 3 Ma.)



Outcrop I: Sample of serpentinized ultramafic rock displaying a typical light green "waxy" layer. The rock has multiple veins filled with serpentinite minerals such as antigorite, lizardite, chrysotile. Other minerals often formed are brucite, talc, Ni-Fe alloys, and magnetite. The mineral alteration occurs primarily at the sea floor at tectonic plate boundaries under hydrothermal conditions.

Outcrop II. Dolerite. The ultramafic rocks are intruded by gabbroic dikes ranging in composition from pyroxene gabbros to hornblende diorites [Koepke J. et. al].

Outcrop II. Closeup of previous picture. Dolerite (see Appendix)

Outcrop II. This breccia containing presumably dolerite clasts could be of volcanic or intrusive nature. Such igneous breccias are commonly found in shallow subvolcanic intrusions, where they are transitional with volcanic breccias. Intrusive rocks can become brecciated in appearance by multiple stages of intrusion, especially if fresh magma is intruded into partly consolidated or solidified magma. Note that there is no sign of serpentinization.

In general, igneous breccias can be divided into two classes:

- Broken, fragmental rocks associated with volcanic eruptions, both of the lava and pyroclastic type and
- broken, fragmental rocks produced by intrusive processes, usually associated with plutons or porphyry stocks.

Igneous breccias are also commonly found in shallow subvolcanic intrusions such as porphyry stocks, granites and kimberlite pipes, where they are transitional with volcanic breccias. Intrusive rocks can become brecciated in appearance by multiple stages of intrusion, especially if fresh magma is intruded into partly consolidated or solidified magma.

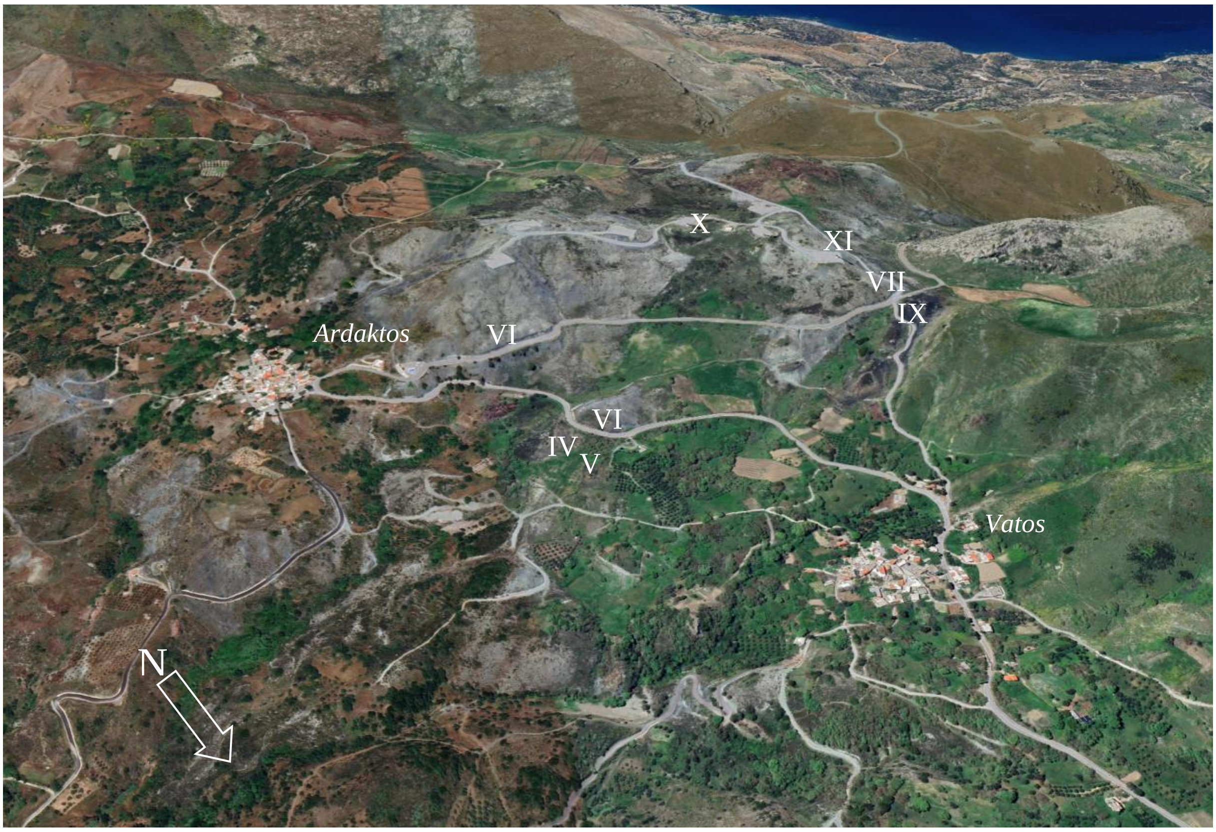
Clastic rocks in mafic and ultramafic intrusions can be formed by several processes such as:

- Consumption and melt-mingling with wall rocks, where the felsic wall rocks are softened and gradually invaded by the hotter ultramafic intrusion;
- accumulation of rocks which fall through the magma chamber from the roof, forming chaotic remnants;
- autobrecciation of partly consolidated cumulate by fresh magma injections or by violent disturbances within the magma chamber (e.g. postulated earthquakes);
- accumulation of xenoliths within a feeder conduit or vent conduit

[Learning Geology: Breccia (geologylearn.blogspot.com)]

Outcrop III, Phyllite belonging to the Vatos Nappe (see following section for details)

Outcrop III: Vatos Phyllite, displaying matrix supported clasts. Vatos Phyllite is described as weak metamorphic grey marls and sandy to clayey limestones of Upper Jurassic/Lower Cretaceous age [Kull]. It occurs within the Vatos Nappe and is thought to form the matrix of "Wildflysch" [Zulauf].



Location of outcrops

View of the Spili Graben. Flysch is reported to lie within the graben underlying Neogene and Pleistocene gravels. To the North, the graben is bordered by Tripoliza and Pindos limestone that form the ridges of the valley. 1: Smooth normal fault surfaces are recognizable at the base of the north-eastern side of the valley. The fault zone is still tectonically active today. Towards the S/SW (not visible in the picture) is a complicated patch work of rock representing the Mourne, Ophiolite and Vatos Nappes.

Outcrop IV: Serpentinized ultramafic/ mafic rock.

Outcrop IV: Close up of previous picture. 1: Rounded pieces of serpentinite are lodged between parallel layers of serpentinized brecciated mafic rock. Autobrecciation occurs when thick, nearly solid lava breaks up into blocks and these blocks are then reincorporated into the lava flow again and mixed in with the remaining liquid magma. The resulting breccia is uniform in rock type and chemical composition. Serpentinization and/or metamorphic overprint is likely to have taken place at a later stage. The round serpentinite clasts may have been picked up by the almost solidified lava flow, indicating different magmatic phases.

Outcrop IV: Close up of a brecciated mafic rock sample, probably an aphanitic basalt (i.e. contains only microscopic grains)

Outcrop V: Quartzite at a field below the main road

Outcrop V: Quartzite

Outcrop VI: Quartzite at road cutting.

Outcrop VI: Quartzite displaying numerous small cavities. The cavities could be the result of an alteration process whereby less stable or soluble minerals have been removed.

2.1 Serpentinites within the Ophiolite Nappe

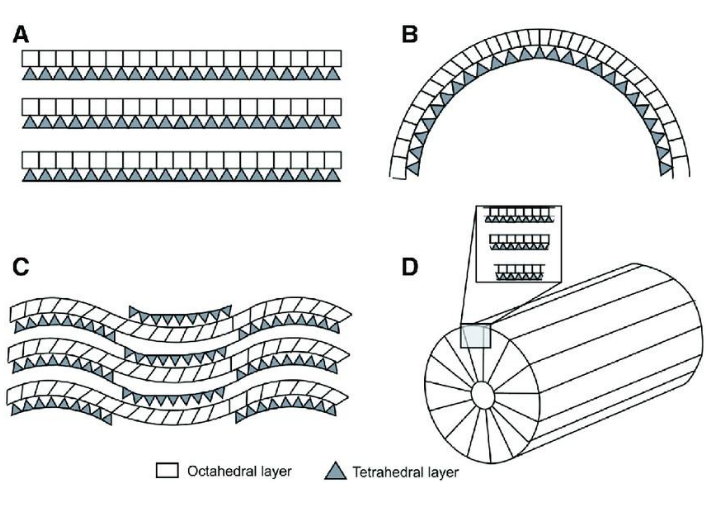
(N35°10‘28‘‘, E024°33‘45‘‘)

In many ultramafic plutonites, the olivine is serpentinised by inaction with fluids such as water or CO_2. It then looks black and is barely recognizable as olivine. Serpentinisation leads to the formation of a new fine-grained mineral mixture that cannot be determined macroscopically and forms a soft, dull black or dark green mass. The iron contained in the olivine is precipitated as finely dispersed magnetite. This magnetite is responsible for the black colour. *[Gesteinbestimmung]*

The serpentinite within the area displays lenticular structures with shiny shear surfaces. Relics of clinopyroxene and orthopyroxene (Opx) indicate that the serpentinite originated from mantle peridotite. Enstatite (Opx) + forsterite + water = serpentine. Water is needed to turn peridotite (olivine) into serpentinite. If CO_2 is also present within the system, it reacts with magnesium to form magnesite ($MgCO_3$), which often forms white veins within the serpentinite. Serpentinites form at transform faults and other displacements, wherever mantle material is able to react with water. Serpentinite is stable up to temperatures of 500-600°C, depending on pressure (approx.1 GPa). To the left and right of oceanic ridges for example, the temperature is low enough for serpentinite to form through crustal expansion. Alternatively, serpentinites may also develop in the forearc mantle of subduction zones when water is present.

2.1.1 Mineral content of Serpentinite

The serpentine minerals are greenish, brownish, or spotted minerals. They are used as a source of magnesium and asbestos, and as decorative stone. Serpentine minerals are polymorphous, meaning that they have the same chemical formulae, but the atoms are arranged into different structures, or crystal lattices. Chrysotile, which has a fibrous habit, is one polymorph of serpentine and is one of the more important asbestos minerals. Other polymorphs in the serpentine subgroup such as antigorite and lizardite may have a platy habit.



Schematic representation of the different serpentine polymorph structures. In principle serpentinite is a two 2-layer sheet silicate consisting of a tetraheadral layer (silica) and an octahedral layer (magnesium).

A) flat layer lizardite,
B) tubular chrysotile,
C) modulated antigorite (Mevel, 2003)
D) polygonal serpentine [adapted from (Cressey et al.,2010)]. [Source: Lacinska A. M. et. al., 2016]

Antigorite is the polymorph of serpentine that most commonly forms during metamorphism of wet ultramafic rocks and is stable at the highest temperatures—to over 600°C at depths of approx. 60 km. In contrast, lizardite and chrysotile typically form near the Earth's surface and break down at relatively low temperatures, probably well below 400°C. It has been suggested that chrysotile is never stable relative to either of the other two serpentine polymorphs.

Samples of the oceanic crust and uppermost mantle from ocean basins document that ultramafic rocks there commonly contain abundant serpentine. Antigorite contains water in its structure, about 13 percent by weight. Hence, antigorite may play an important role in the transport of water into the earth in subduction zones and in the subsequent release of water to create magmas in island arcs, and some of the water may be carried to yet greater depths.

2.1.2 Simplified Chemical Formula:

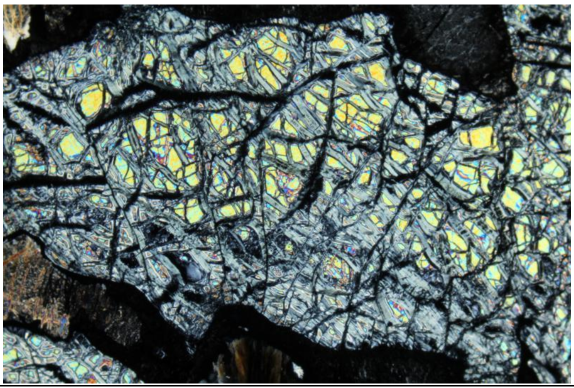
$Mg_3[Si_2O_5]\ (OH)^{-4}$,

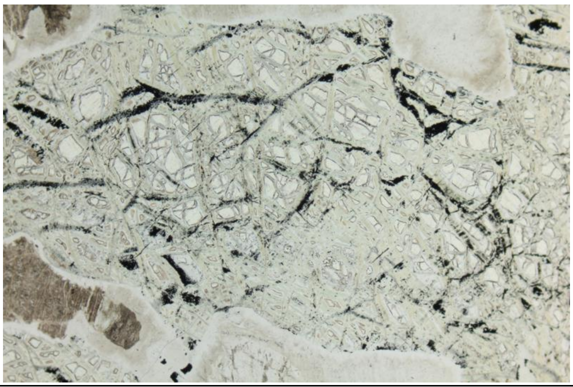
Serpentinite contains a greater amount of Mg compared to its protolith minerals (ratio Mg:Si serpentinite = 3:2, in comparison ratio Mg:Si olivine = 2:1, ratio Mg:Si Opx = 1:1). The OH-group within the formula indicates the presents of water within the crystal structure.

2.1.3 Mesh Texture of Serpentinzed Olivine

In olivine-rich rocks such as dunites and peridotites, serpentinization may or may not be complete. If the transformation is incomplete one can observe a serpentine mesh structure composed of a lattice work of longitudinally divided cross-fibre veins enclosing olivine cores. Each olivine grain may be divided into several cores which retain their original orientation without disturbance. Generally, the serpentine veins contain two parallel series of fibrous material separated by a black band of minute magnetite aggregates. If the transformation is complete, the inner parts of the mesh structure are filled by serpentine micro aggregates *[Alex Strekeisen]*.

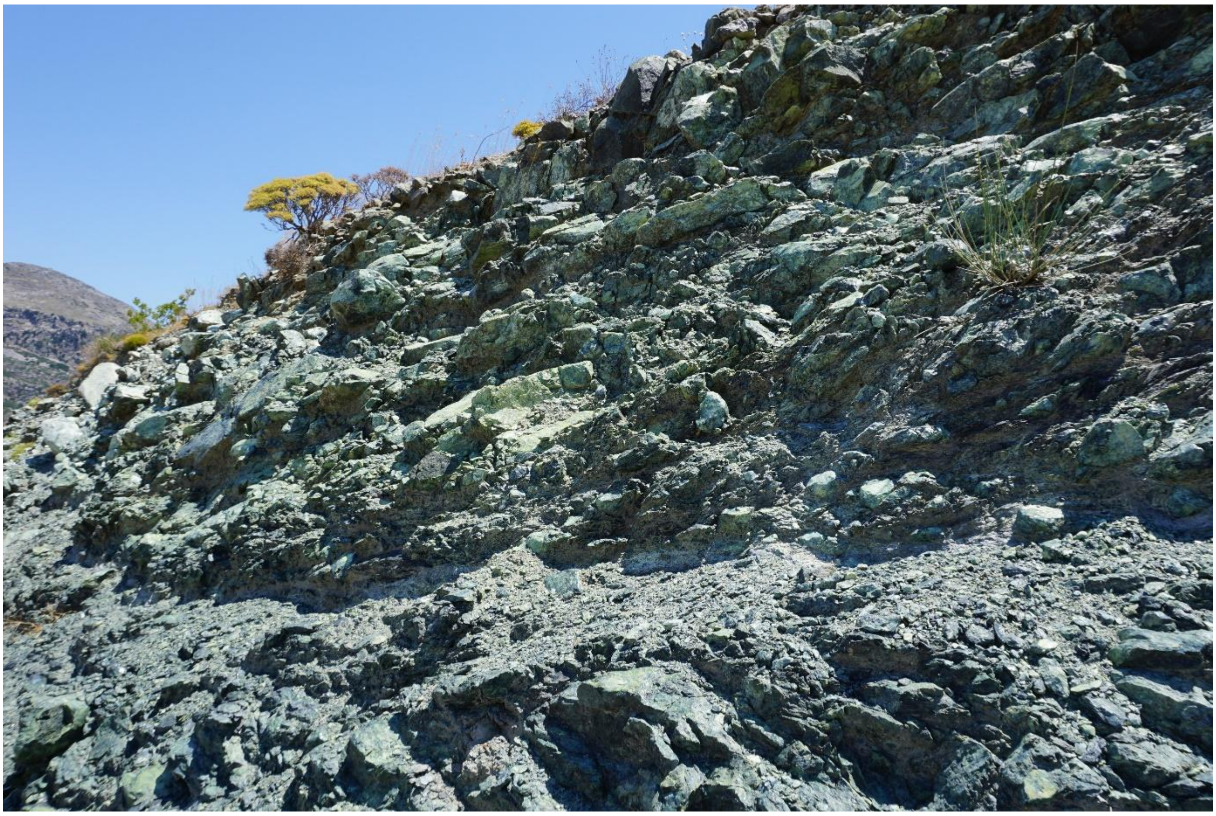
Serpentinised olivine appears black on fracture surfaces even if it is still partly fresh inside. The serpentinisation spreads along cracks within the olivine grains. When broken, the olivine disintegrates along these cracks filled with serpentinite, so that only the black serpentinite can be seen on the fracture surfaces. *[Gesteinsbestimmung]*





Mesh texture in olivine. Each olivine grain is surrounded by fibrous serpentine and black magnetite aggregates. PPL image, 2x (Field of view = 7mm) [Alex Strekeisen]	

Outcrop VII: Serpentinite



Outcrop VII: Serpentinite

Location of outcrops

Outcrop VIII. Highly serpentintized ultrabasic rock

Outcrop VIII. Serpentinite sample from the location above displaying light and dark green waxy layer that is found at frictional shear surfaces.

Outcrop IX: Mafic to ultramafic magmatic rock. Note that it is only moderately serpentinized/altered in comparison to the green serpentinzed ultramafic rock.


Outcrop IX: Mafic to ultramafic lava flow. Due to its dark colour in contrast to the green and cream coloured serpentinites there appears to be a least two different magmatic phases (see also outcrop V and XI).

2.2 Metamorphic Overprint of the Ophiolites and Associated Dykes

All the ophiolites on Crete show a more or less pronounced metamorphic overprint of the greenschist to epidote-amphibolite facies, evidenced by antigorite in the serpentinites. The metamorphic overprint is also evident in gabbro-dioritic dikes (see following section) that display a paragenesis of actinolite - epidote - albite. The metamorphic overprint of the Cretan ophiolites is assumed to have taken place during the Late Cretaceous or later. *[Koepke et al., 2002]*

Outcrop X: Highly sheared serpentinite containing lenses of light coloured breccias (arrow).

Outcrop X: light coloured breccia lens

Outcrop X: closeup of light coloured breccia lens. Neither the clasts nor the matrix contain calcite.

Outcrop X: closeup of sample from the light-coloured breccia lens. The dark specks are thought to be opaque phases such as magnetite or other accessories.

Outcrop XI: Mafic to ultramafic porphyritic rock.


Outcrop XI: Mafic to ultramafic porphyritic rock, which is thought to be subvolcanic located within a dyke.


Outcrop XI: close up

Outcrop XII: Light coloured breccia thought to be a metamorphic tuff breccia.

Outcrop XII: close up of previous picture, thought to be an altered tuff breccia.

Outcrop XII: close up of previous picture. The presents of a tuff breccia, which is caused by eruption, could indicate an emerged environment such a seamount volcano above sea level.

3 The Vatos Nappe

The Vatos Nappe consists of wildflysch-like Upper Jurassic to Upper Cretaceous metasediments and volcanics (Bonneau et al. 1977; Krahl et al. 1982; Malten 2019). Calcschists include Upper Cretaceous fossils and display a synkinematic crystallization of white mica, calcite, quartz and albite along the main foliation (Koepke 1986; Tortorici et al. 2012). Deformation microfabrics of quartz and calcite and the metamorphic index minerals, such as actinolite, chlorite, albite and white mica indicate lower greenschist-facies metamorphism (Karakitzios 1988; Malten 2019), which must be younger than Uppermost Cretaceous (the age of the fossils). Of particular interest are limy metaconglomerates that contain Late Jurassic/Early Cretaceous foraminifera, and which include rodingite and other ophiolitic pebbles (Krahl et al. 1982) (Malten 2019). U–Pb dating of calcite, which precipitated in pressure shadows behind rigid clasts of the limy metaconglomerate, yielded 76 ± 4 Ma (Malten 2019), which is probably the age of the greenschist-facies metamorphism. *[Zulauf, 2023]*

Outcrop XIII: Vatos wildflysch exposed WSW of Ardaktos. 1: Vatos phyllite, 2: Quartzite layer at the top

Outcrop XIII: 1: Vatos phyllite

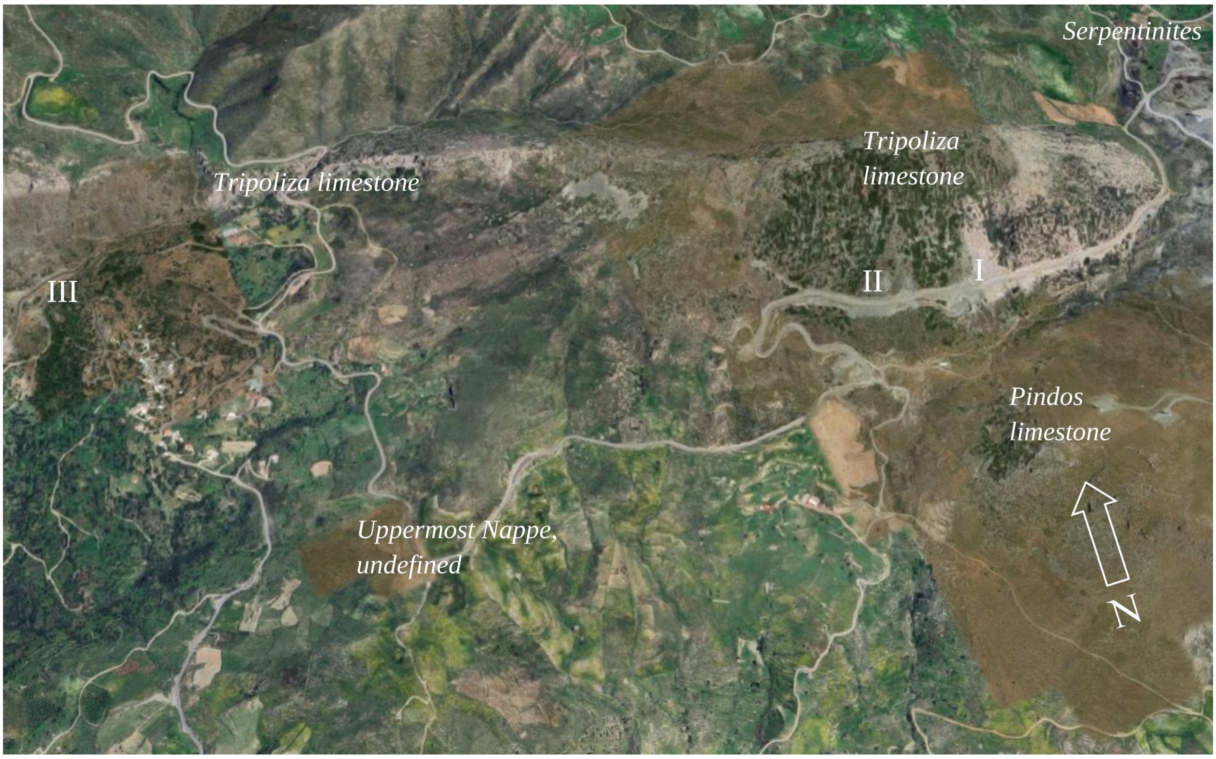
Outcrop XIII: 2: Quartzite layer, 3: pegmatite vein

Outcrop XIII: 2: Quartzite layer, 3: pegmatite vein displaying quartz (qu) and feldspar (fsp)

The quartzite layer exposed in Outcrop XIII forms the uppermost part of a flysch sequence and displays primary bedding in form of light and dark layers. The bedding plane dips moderately towards NNE. The quartzite is cut by a subhorizontal brittle shear plane, which shows a N–S trending striation. The quartzite was formed from badly sorted sandstone. Apart from quartz, there are other components such as white mica, plagioclase, chlorite and opaque phases.

	*Quartzite from within the Vatos flysch. The large quartz grains show evidence of deformation portrayed by serrated grain boundaries and stylolites, which indicate both dislocation and dissolution–precipitation creep as the main deformation mechanisms. [Zulauf G. et. al., July 2023*

4 Tripoliza Limestone



Location of Outcrops

Outcrop I: Steeply dipping Tripoliza limestone, which is interpreted as part of a tilted horst structure.

Outcrop II: Tripoliza limestone. Dark bands of organic carbon in well banked limestone beds.

Outcrop II: Closeup of previous picture displaying finely laminated algae concretions.

The route to Dyrimiskos leads through a narrow, steeply inclined horst of Tripoliza limestone. Behind the highest point in the mountain ridge, there are fresh outcrops of steeply dipping Tripoliza limestone. The well banked dark and very fine-grained limestone partly displays shell fragments, but no cherts, which is one feature that can be used to differentiate between Tripolitza and Pindos limestone. Tripolitza limestone is also usually significantly darker than Pindos, as it contains a higher proportion of organic carbon. The presents of fossils such as nummulites (foraminifera) and rudists (bivalves) indicate deposition in a shallow water, shelf environment.

View of the Asideroto Mountain from Outcrop II looking East. A: Ridge belonging to Asideroto Mtn, Pindos limestone, B: Asideroto Mtn, Pindos Limestone, C: Kenda Mt, Tripoliza Limestone

Outcrop III: Cataclastic Tripoliza limestone located just below the village of Dimitrikos. The highly crushed and ground texture indicates the presents of a major brittle shear zone, possibly representing the base of the Tripoliza Nappe.

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6 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	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

Mélange, Flysch (Wildflysch)

https://www.wikiwand.com/de/Tektonische_M%C3%A9lange

A mélange is a large-scale breccia characterized by a lack of continuous bedding and the inclusion of fragments of rock of all sizes, contained in a fine-grained deformed matrix. The mélange typically consists of a jumble of large blocks of varied lithologies. Both tectonic and sedimentary processes can form mélange. Mélange occurrences are associated with thrust faulted terranes in orogenic belts. A mélange is formed in the accretionary wedge above a subduction zone. The ultramafic ophiolite sequences which have been obducted onto continental crust are typically underlain by a mélange. Smaller-scale localized mélanges may also occur in shear or fault zones, where coherent rock has been disrupted and mixed by shearing forces.

Large-scale mélanges formed in active continental margin settings generally consist of altered oceanic crustal material and blocks of continental slope sediments in a sheared mudstone matrix. The mixing mechanisms in such settings may include tectonic shearing forces, ductile flow of a water-charged or deformable matrix (such as serpentinite), sedimentary action (such as slumping, gravity-flow, and olistostromal action), or some combination of these. Some larger blocks of rock may be as much as 1 kilometre (0.62 mi) across. *[Wikipedia]*

Wildflysch is defined as a turbiditic, mass-flow sediments with numerous ill-sorted exotic clasts. In modern terminology wildflysch is better referred to as a form of diamictite.

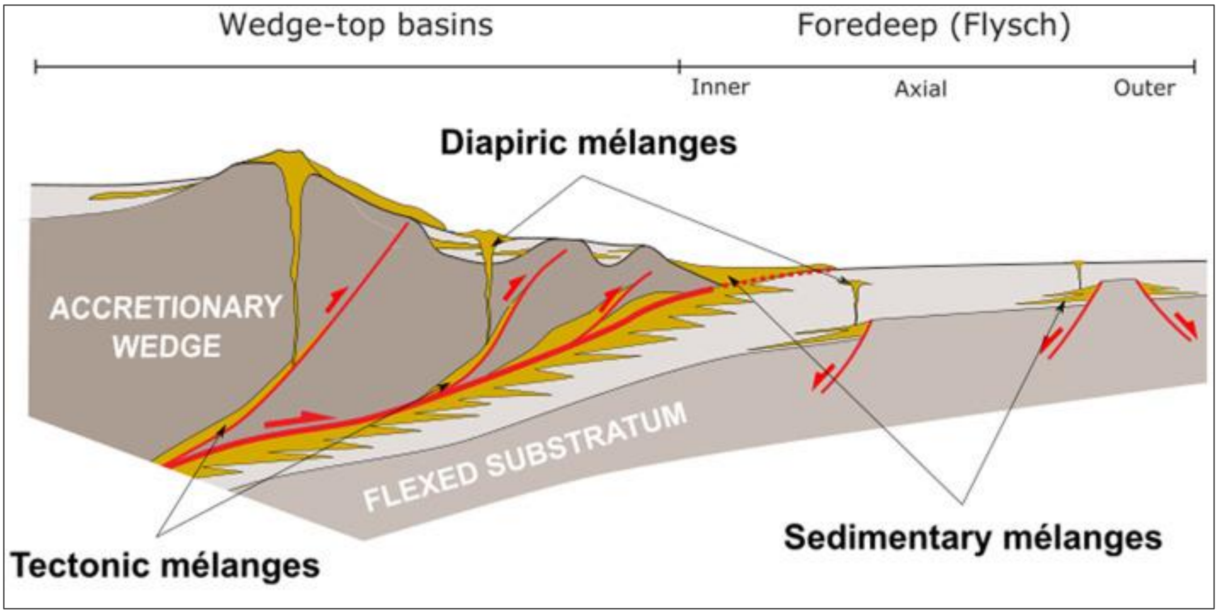


Fig. 2. Schematic cartoon illustrating a profile across an ideal Alpine-type foreland basin (flysch to molasse) with subdivision of the depositional settings and the typical localisation of sedimentary, tectonic and diapiric mélange units. Note the reworking and recycling of these processes and products through time (modified from Festa et al., 2019). [Ogata K. et. al., 2013]

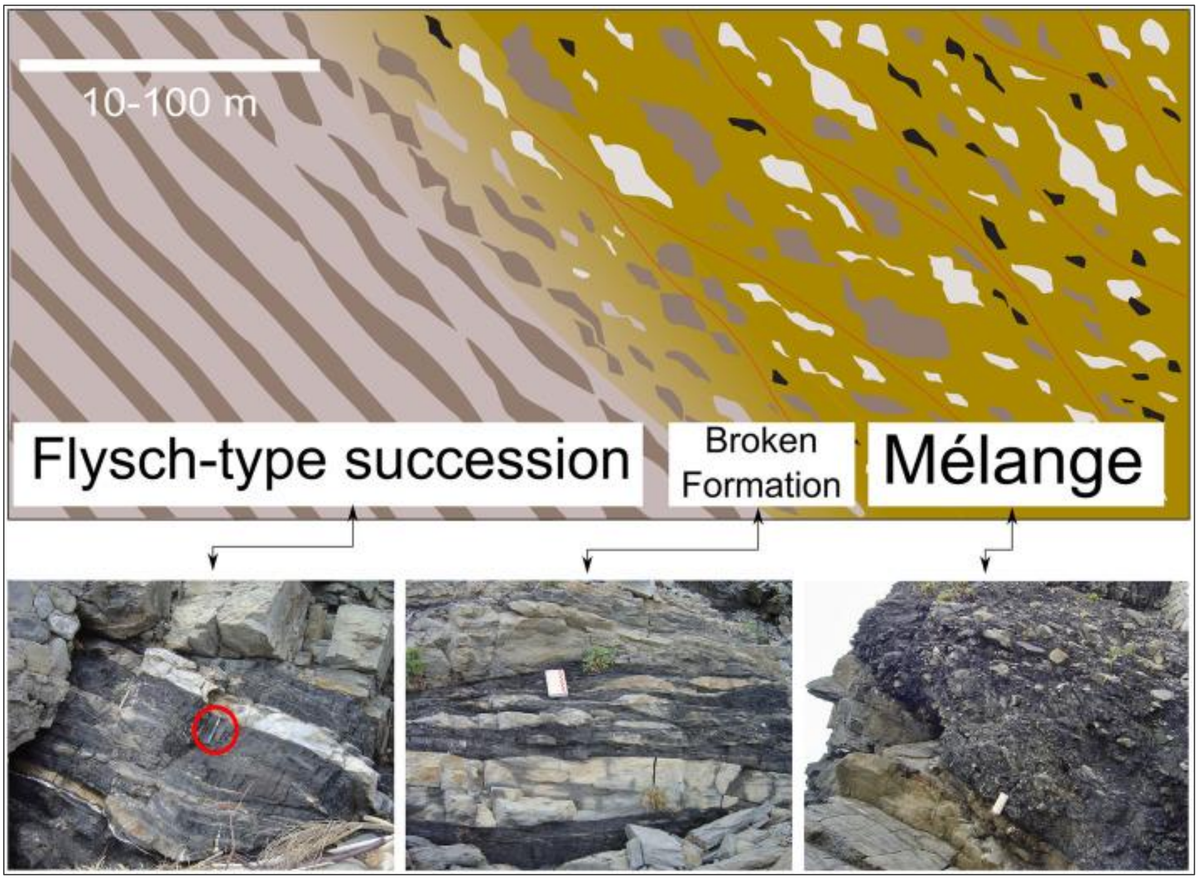


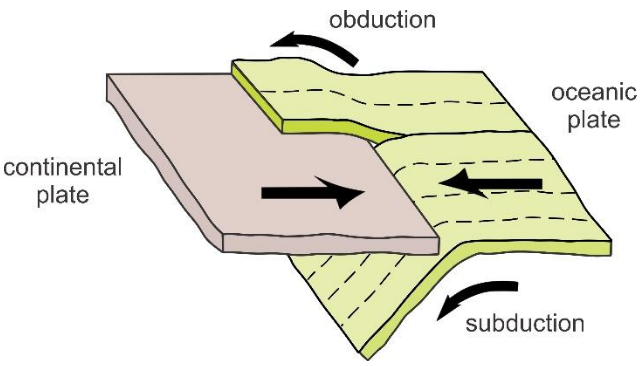
Fig. 1. Conceptual diagram showing a gradual transition from an undeformed, flysch-type lithology to a mélange (mixed rock) passing through a broken formation (un-mixed rock; modified from Festa et al., 2019). Outcrop examples are from the Eocene Canetolo Formation exhumed at the Cinque Terre, La Spezia (Northern Apennines, Italy). [Ogata K. et. al., 2013]

Ophiolites and Serpentinites

https://opengeology.org/petrology/13-metamorphism-of-mafic-rocks/#1371

Earth's mantle is made of ultramafic rocks. Peridotites, including mostly lherzolite, harzburgite, and dunite, dominate. At mid-ocean ridges, these ultramafic rocks rise to fill space created when plates diverge. This leads to decompression melting, creating mafic magmas that crystallize in the upper part of the lithosphere as gabbros, or that erupt as seafloor basalts.

Overall, the oceanic lithosphere consists of a thin part of the uppermost mantle (ultramafic), overlain by mafic and ultramafic crust, in turn overlain by thin layers of seafloor sediments. The lithosphere cools and becomes denser during seafloor spreading, and so sinks. Thus, the youngest seafloor and highest elevations are at the ridges, and the oldest seafloor and lowest elevations at ocean margins (Figure 3.21). The oceanic lithosphere may not be the same composition and structure everywhere, but all evidence suggests that there are some standard components. The evidence comes from many sources including drill core, seismic studies, laboratory experiments, grab sampling from the ocean floor, and dredging. The best information, however, may come from studies of ophiolites. The term ophiolite derives from the Greek words ophio (snake) and lite (stone), referring to the commonly green color of the rocks that make up ophiolites.



Obduction and subduction occurring at a subduction zone

Where oceanic and continental plates converge, oceanic lithosphere generally subducts beneath the continent. Sometimes, however, pieces of oceanic lithosphere are scraped off and added to a continent. A process called obduction. Because of obduction, ultramafic rocks, formerly pieces of the mantle, can be found in ophiolite complexes. These complexes are wedges of ocean crust and mantle, now exposed in outcrops at the surface, that were thrust onto continental margins above subduction zones. Ophiolites are very important to geologists because they provide direct evidence for the nature of Earth's oceanic crust and upper mantle. Although there are many ophiolites around the world, most are small or very fragmented. Most ophiolites are 10s or 100s of millions of years old because the processes of sea-floor spreading and obduction are slow. The Macquarie Island Ophiolite, the youngest known, is still more than 10 million years old.

Every ophiolite provides a partial cross section of the oceanic lithosphere. When the information from many ophiolites is combined with other evidence, a standardized model of the oceanic lithosphere, emerges. A "complete" ophiolite would include all the layers of rock the standardized model. These layers, which correspond to sediments and igneous rocks created by seafloor spreading, make up a cross section of the oceanic lithosphere. At its top, muds and other debris typically overlie hard rock. These sediments, which increase in thickness from mid-ocean ridges to ocean margins, may eventually lithify to form shale or chert. A layer of basalt, often containing pillow lavas, worm-like bodies formed during submarine eruptions, underlies the sediments. These basalts, like many ocean floor basalts, are commonly highly altered by interaction with seawater.

Magmas that create ocean-floor basalts rise from magma chambers below, following fractures and creating vertical, parallel, mafic dikes. The many dikes produce a sheeted dike complex beneath the basalts. At still greater depth, a thick layer of gabbro is the remains of once liquid basaltic magma chambers. The gabbro layer, accounting for most of the oceanic crust by volume, typically contains mafic to ultramafic cumulates in its lowest levels. Figure 13.36 is a photo of gabbro layers in the Troodos ophiolite – the browner layers contain significant amounts of orthopyroxene compared with the lighter colored layers. Peridotites (mainly harzburgites and lherzolites) of the oceanic mantle can be found beneath the gabbro layer.

13.36 Layered gabbros from the Troodos Ophiolite in Cyprus

The peridotites in the oceanic lithosphere are very high-temperature rocks and consequently the minerals they contain are unstable under normal crustal conditions. High-temperature minerals are especially unstable in the presence of water. So when uplifted to become part of the oceanic lithosphere, original olivine and pyroxenes typically are metamorphosed by seawater to produce a variety of different hydrous minerals. Further hydration also occurs during tectonism associated with subduction zones.

During continental collision and obduction, ophiolites may become incorporated into mountain belts to become bodies of rock known as alpine peridotites. These peridotite bodies range from small slivers to large plutons. Often the peridotites are dismembered pieces – all that remain of once more complete oceanic lithosphere. Alpine peridotites typically contain serpentinized rocks.

Ophiolite emplacement

Wikipedia

WIKIPEDIA
Die freie Enzyklopädie

There is yet no consensus on the mechanics of emplacement, the process by which oceanic crust is uplifted onto continental margins despite the relatively low density of the latter. All emplacement procedures share the same steps nonetheless: subduction initiation, thrusting of the ophiolite over a continental margin or an overriding plate at a subduction zone, and contact with air.

Scientists have drilled only about 1.5 km into the 6- to 7-kilometer-thick oceanic crust, so scientific understanding of oceanic crust comes largely from comparing ophiolite structure to seismic soundings of in situ oceanic crust. Oceanic crust generally has a layered velocity structure that implies a layered rock series similar to that listed above. But in detail there are problems, with many ophiolites exhibiting thinner accumulations of igneous rock than are inferred for oceanic crust. Another problem relating to oceanic crust and ophiolites is that the thick gabbro layer of ophiolites calls for large magma chambers beneath mid-ocean ridges. However, seismic sounding of mid-ocean ridges has revealed only a few magma chambers beneath ridges, and these are quite thin. A few deep drill holes into oceanic crust have intercepted gabbro, but it is not layered like ophiolite gabbro.

Beyond issues of layer thicknesses mentioned above, a problem arises concerning compositional differences of silica ($SiO2$) and titania ($TiO2$). Ophiolite basalt contents place them in the domain of subduction zones (~55% silica, <1% $TiO2$), whereas mid-ocean ridge basalts typically have ~50% silica and 1.5–2.5% $TiO2$. These chemical differences extend to a range of trace elements as well (that is, chemical elements occurring in amounts of 1000 ppm or less).

Additionally, the crystallization order of feldspar and pyroxene (clino- and orthopyroxene) in the gabbros is reversed, and ophiolites also appear to have a multi-phase magmatic complexity on par with subduction zones. Indeed, there is increasing evidence that most ophiolites are generated when subduction begins and thus represent fragments of fore-arc lithosphere. This led to introduction of the term "supra-subduction zone" (SSZ) ophiolite in the 1980s to acknowledge that some ophiolites are more closely related to island arcs than ocean ridges. Consequently, some of the classic ophiolite occurrences thought of as being related to seafloor spreading (Troodos in Cyprus, Semail in Oman) were found to be "SSZ" ophiolites, formed by rapid extension of fore-arc crust during subduction initiation.

A fore-arc setting for most ophiolites also solves the otherwise-perplexing problem of how oceanic lithosphere can be emplaced on top of continental crust. It appears that continental accretion sediments, if carried by the down going plate into a subduction zone, will jam it up and cause subduction to cease, resulting in the rebound of the accretionary prism with fore-arc lithosphere (ophiolite) on top of it. Ophiolites with compositions comparable with hotspot-type eruptive settings or normal mid-oceanic ridge basalt are rare, and those examples are generally strongly dismembered in subduction zone accretionary complexes.

Metamorphic Reactions

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Different Kinds of Reactions

Under any pressure and temperature, the most stable mineral assemblage is the one with the lowest Gibbs free energy. So, when a rock is heated or squeezed, chemical reactions occur that may consume old minerals and create new ones. These reactions may be of several types. The table seen here gives examples of different types of metamorphic reactions. By convention, the low-temperature mineral or assemblage is to the left of the equal sign; the high-temperature products are to the right.

Solid-solid reactions involve no H_2O, CO_2, or other vapor phase. The first example of a solid-solid reaction contains only two minerals, both Al_2SiO_5 polymorphs. This reaction may occur when a metamorphosed shale is heated to high temperature. But most metamorphic reactions involve more than two minerals, and many involve H_2O or CO_2. The second solid-solid reaction is more typical and involves four minerals.

Dehydration reactions and decarbonation reactions, such as the examples in this table, liberate H_2O and CO_2, respectively. Hydration reactions and carbonation reactions consume H_2O and CO_2, respectively.

Metamorphic reactions involve changes in mineralogy or in mineral composition. A mineral assemblage is at chemical equilibrium if no such changes are occurring. If the assemblage has the lowest Gibbs free energy possible for the given conditions, it is at *stable equilibrium*. In principle, all rocks tend toward stable equilibrium. Whether they reach it depends on many things, including temperature, grain size, and reaction kinetics. If reactions cease before a rock has reached stable equilibrium, the rock is at *metastable equilibrium*. Many metamorphic rocks contain metastable minerals.

Examples of Metamorphic Reactions
Solid–solid reactions:
andalusite = sillimanite $Al_2SiO_5 = Al_2SiO_5$
grossular + quartz = anorthite + 2 wollastonite $Ca_3Al_2Si_3O_{12} + SiO_2 = CaAl_2Si_2O_8 + 2\ CaSiO_3$
Dehydration reactions:
muscovite + quartz = K–feldspar + sillimanite + vapor $KAl_2(AlSi_3)O_{10}(OH)_2 + SiO_2 = KAlSi_3O_8 + Al_2SiO_5 + H_2O$
kaolinite + 2 quartz = pyrophyllite + vapor $Al_2Si_2O_5(OH)_4 + 2\ SiO_2 = Al_2Si_4O_{10}(OH)_2 + H_2O$
Hydration reaction:
enstatite + 2 H_2O = 2 brucite + 2 quartz $Mg_2Si_2O_6 + 2\ H_2O = 2\ Mg(OH)_2 + 2\ SiO_2$
Carbonation reaction:
forsterite + 2 CO_2 = 2 magnesite + quartz $Mg_2SiO_4 + 2\ CO_2 = 2\ MgCO_3 + SiO_2$

We call a stable mineral assemblage representative of a given set of pressure-temperature conditions a *paragenesis*. When conditions change, metamorphic reactions may create a new paragenesis as some minerals disappear and others grow. Such reactions may be prograde or retrograde. Most of the reactions in the table above are prograde, but the two examples of carbonation and hydration reactions are retrograde reactions (involving orignial high-temperatue minerals reacting to form low-temperature minerals) that often affect mafic rocks.

Prograde metamorphism involves the breakdown of minerals stable at lower temperature to form minerals stable at higher temperature. Some prograde reactions are solid-solid reactions, but most involve the release of H_2O or CO_2 that flow along cracks or grain boundaries. As temperature increases, minerals containing H_2O or CO_2 become increasingly unstable, causing dehydration or decarbonation, and the release of H_2O or CO_2 as intergranular fluid. If we ignore H_2O and CO_2, we find that most prograde metamorphism is nearly *isochemical*, meaning that the rock is the same composition before and after metamorphism. Sometimes, however, flowing fluids and metasomatism can be the dominant forces controlling metamorphism.

Retrograde metamorphism is, in many ways, just the opposite of prograde metamorphism. Typically, H_2O- and CO_2-free minerals react with fluids to produce hydrous or carbonate minerals. Mg-silicates such as forsterite (Mg_2SiO_4), and enstatite ($Mg_2Si_2O_6$), for example, may react to form talc or serpentine (both hydrated Mg-silicates), brucite (Mg hydroxide), or magnesite (Mg carbonate), at low temperature. In contrast with prograde reactions, retrograde reactions are often quite sluggish. They may not go to completion and frequently do not reach stable equilibrium. Sometimes retrogression only affects parts of a rock or parts of some grains in a rock.

Metamorphic reactions involve changes in mineralogy or in mineral composition. A mineral assemblage is at chemical equilibrium if no such changes are occurring. If the assemblage has the lowest Gibbs free energy possible for the given conditions, it is at *stable equilibrium*. In principle, all rocks tend toward stable equilibrium. Whether they reach it depends on many things, including temperature, grain size, and reaction kinetics. If reactions cease before a rock has reached stable equilibrium, the rock is at *metastable equilibrium*. Many metamorphic rocks contain metastable minerals.

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Greenstones (Metabasalt)

Greenstones, which are a specific kind of hornfels, form by metamorphism of basalts. Figure 8.34 shows a 9-cm wide sample of greenstone from Ely, Minnesota. The greenish color is due to chlorite or epidote that grew during metamorphism.

Figure 8.35 Greenstone outcrop of in Italy. The rock originated as an ocean-floor basalt, and contains rounded structures called pillows, indicative of submarine eruption.

Metamorphosed Ultramafic Rocks

Ultramafic rocks come from Earth's mantle. Most of the examples of metamorphosed ultramafic rocks that can be observed are in o*phiolites*, slivers of Earth's oceanic crust and mantle uplifted and accreted onto continents.

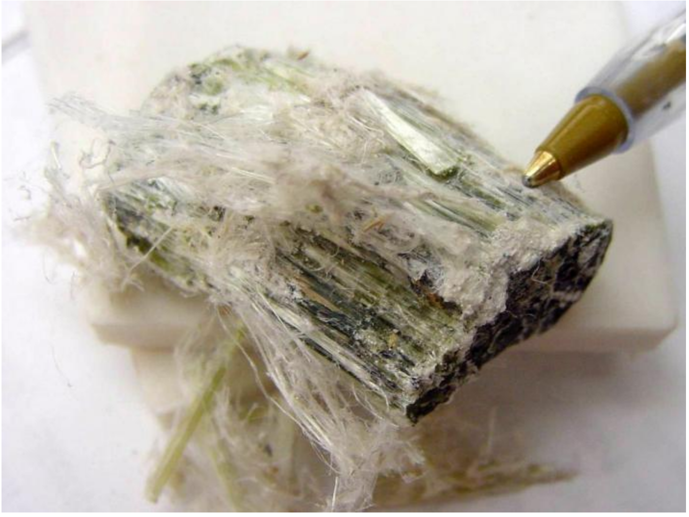
Mantle rocks are high-temperature, high-pressure rocks that typically contain olivine, clinopyroxene, and orthopyroxene when unweathered. When weathered or metamorphosed at low temperature, the original minerals often react to create low-temperature minerals. Hydration and carbonation reactions occur and produce hydrous and carbonate minerals. Magnesium oxides and hydroxides may also form. Thus, unless metamorphic temperatures are very high, metamorphism of ultramafic rocks produces low-temperature minerals from high-temperature minerals, essentially retrograde metamorphism.

Common Minerals in Metamorphosed Ultramafic Rocks	
	minerals
low grade	talc $Mg_3Si_4O_{10}(OH)_2$
	brucite $Mg(OH)_2$
	magnesite $MgCO_3$
	serpentine $Mg_6Si_4O_{10}(OH)_8$
	olivine $(Mg,Fe)_2SiO_4$
	anthophyllite $(Mg,Fe)_7Si_8O_{22}(OH)_2$
	anthophyllite $(Mg,Fe)_7Si_8O_{22}(OH)_2$
	garnet (pyrope-almandine) $(Mg,Fe)_3Al_2Si_3O_{12}$
	clinopyroxene (diopside) $CaMgSi_2O_6$
high grade	orthopyroxene (enstatite) $Mg_2Si_2O_6$

The most common minerals in metamorphosed ultramafic rocks. Low-grade minerals are at the top of the table and high-grade minerals at the bottom. Ultramafic rocks are very Mg-rich, and often contain much olivine. Because of their chemistry, Mg-silicates such as talc, serpentine, anthophyllite, diopside, and enstatite are also common in these rocks.

8.72 Outcrop of a serpentinite containing primarily antigorite. Antigorite, the most common serpentine mineral that forms during metamorphism of ultramafic rocks, is stable over a wide range of metamorphic conditions. The weathered outcrop in Figure 8.72 is a typical occurrence. The green color and the texture are diagnostic.

8.73 Lizardite with stichtite from Tasmania. The specimen is 8.6 cm across Lizardite (Figure 8.73), named for occurrences in the Lizard Complex, Cornwall, and chrysotile (Figure 8.74) are less common and form exclusively at low pressures or at Earth's surface. The green lizardite in the middle photo above contains pinkish inclusions of stichtite, a rare magnesium chrome carbonate.



8.74 Chrysotile, the most common asbestos mineral. Chrysotile is one of the few recognized asbestos minerals; fine fibers are easily seen in Figure 8.74. The other asbestos minerals are amphiboles.

8.75 7 cm wide sample of talc

8.76 Brucite

8.77 Magnesite forming from ultramafic rock near Turin, Italy

While serpentine commonly dominates very low-grade ultramafic rocks, several other minerals may also be present. The photos above show talc (hydrated Mg-silicate), brucite (Mg-hydroxide), and magnesite (Mg-carbonate). These minerals, common in metamorphosed ultramafic rocks, can also form in metacarbonates.

At higher metamorphic grades, ultramafic rocks may contain olivine, anthophyllite, enstatite, periclase, or spinel. And, at the highest grades, garnet and pyroxene become stable. Minerals in high-grade ultramafic rocks are the same as the minerals in rocks of the mantle (where pressure and temperature are great). In effect, mantle rocks originated as high-grade metamorphic rocks.

Metamorphosed Sandstones (Metapsammites)

Compared with metamorphosed pelites, *metamorphosed sandstones*, also called *metasandstones* or metapsammites, are often nondescript. Normal sandstones are mostly quartz, perhaps with some feldspar. When metamorphosed, they still contain quartz and feldspar because these minerals are stable at all metamorphic grades. At low grades, metasandstones typically appear massive and homogeneous, containing light-colored quartz and feldspar grains. The rock seen here (Figure 8.49) is an example. Sometimes small micas and other dark minerals may be scattered evenly throughout.

At higher grades metasandstone may recrystallize with quartz grains growing together and becoming coarser. This produces a quartzite, a hard, nonfoliated metamorphic rock. In quartzites, the once separate quartz crystals become massive quartz with no visible grain boundaries. As this happens, original sedimentary textures are obliterated. Pure quartzites are generally white or light colored (like the one in Figure 8.49) but iron staining often adds a red or pinkish coloration. Figure 8.37, earlier in this chapter, shows another example of an unremarkable quartzite.

8.49 Metamorphosed sandstone from South Australia. Sample is 10 cm across

So, many metasandstones have unexciting mineralogy, but if the original sandstone contained some clay, any of the minerals that can be in metapelites may be present. Quartz usually dominates, and the amounts of other minerals depend on how much clay was in the protolith. Foliation, typical of metapelitic rocks, is usually lacking in these rocks because micas are generally absent.