The Uppermost Nappes - Ophiolites and the Preveli Nappe, Kerames to Korifi Mountain

View of the south coast as seen from a location SW of the village Kerames. In the foreground a dolerite dyke. The area is characterized by outcrops of serpentinite, and mafic to ultramafic rock. In the background are the Trio Petra and Agios Pavlos beaches.

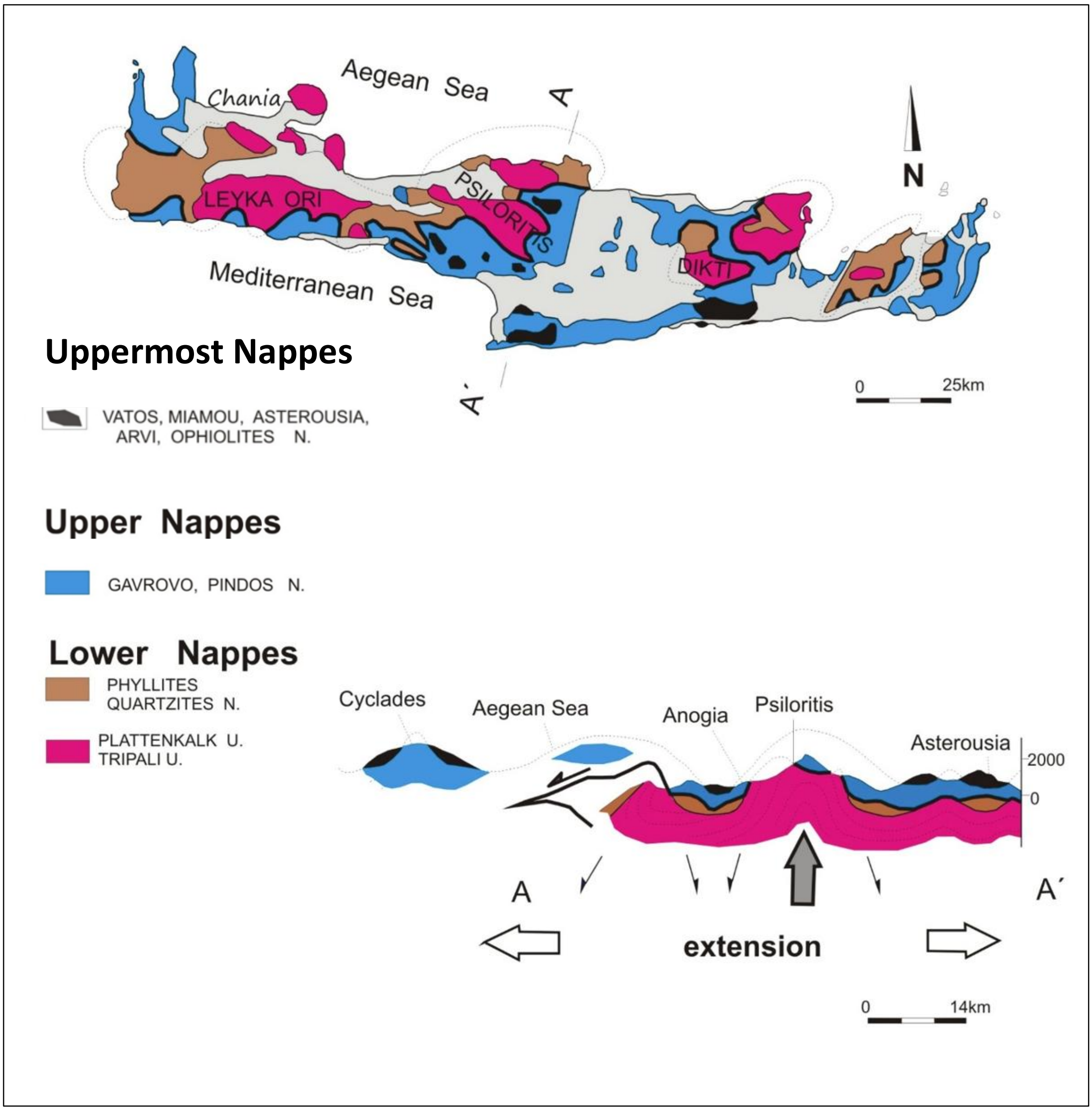
Compiled by George Lindemann, MSc.

Berlin, August 2024

Content

Appendix

1 Introduction



Journal of the virtual explorer _NW Crete_online

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 like 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]*.

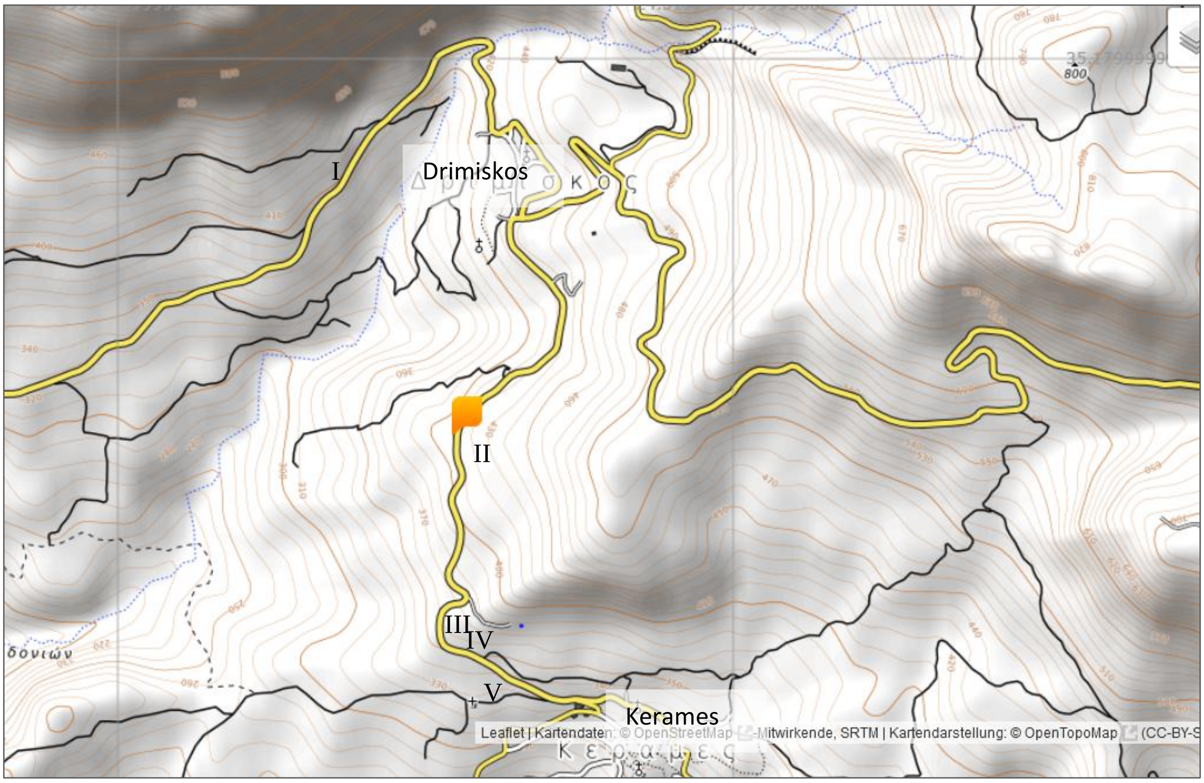
A characteristic feature of the Uppermost Nappes of southern and central Crete is the frequent occurrence of ophiolites, which represent pieces of oceanic crust predominantly from mid ocean ridges. Ophiolites consist of mafic to ultramafic plutonic and volcanic rocks and often exhibit deep marine sediments 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 on Crete, that overly the Tripolitza and the Pindos Nappes. *[Tortorici]*

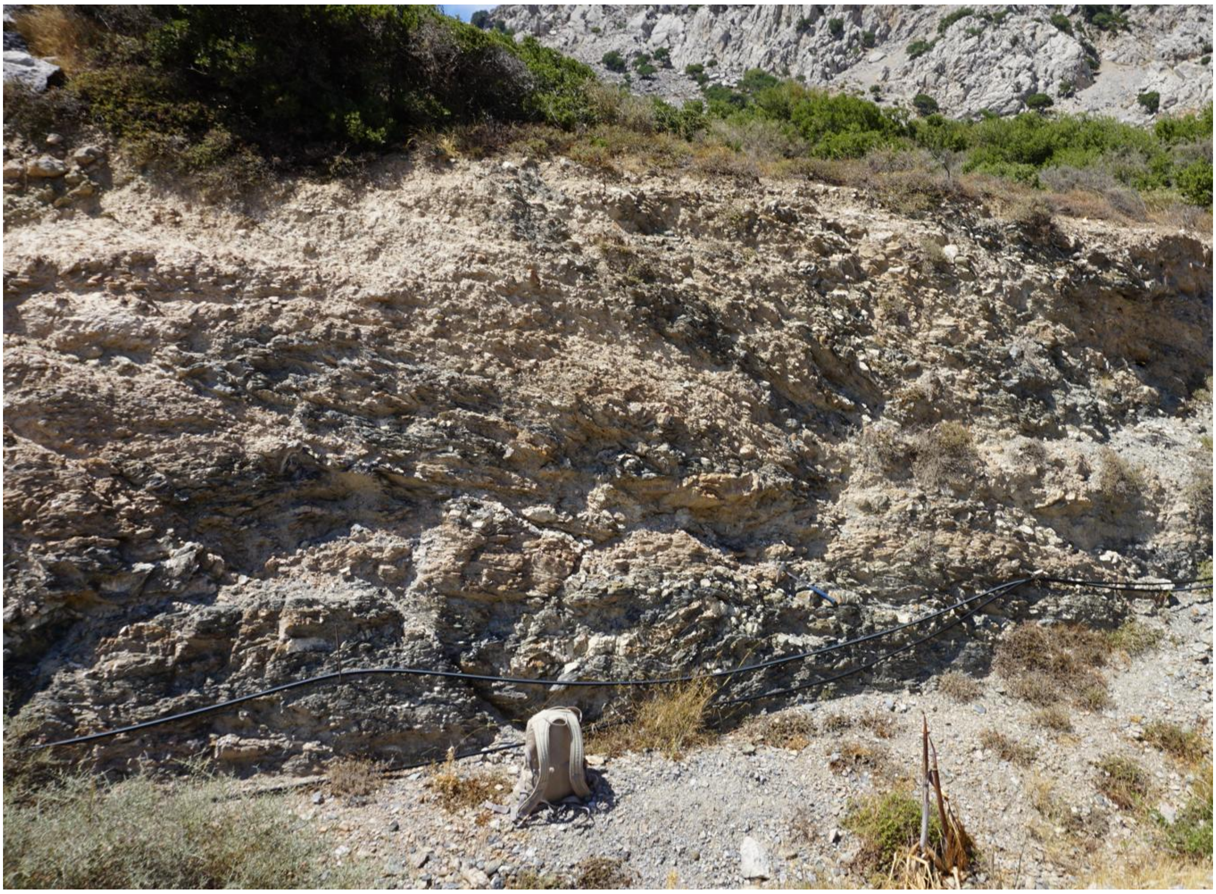
In the past numerous authors have attempted to group the rocks based on field evidence, petrology, and age into distinct nappes or have presented alternative olistostrome models (See Appendix).

Creutzburg and Seidel (1975) regarded the rocks above the unmetamorphosed Pindos flysch 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. Olistostrome is a chaotic unconsolidated sediment transported as a semifluid body by submarine gravity sliding or slumping. Similarly, Bonneau (1984) proposed that the rocks of the very low-grade metamorphic subunit formed an “ophiolitic olistostrome”, but also regarded there to be a mappable blue schist, high-grade metamorphic and ophiolite subunits forming individual nappes.

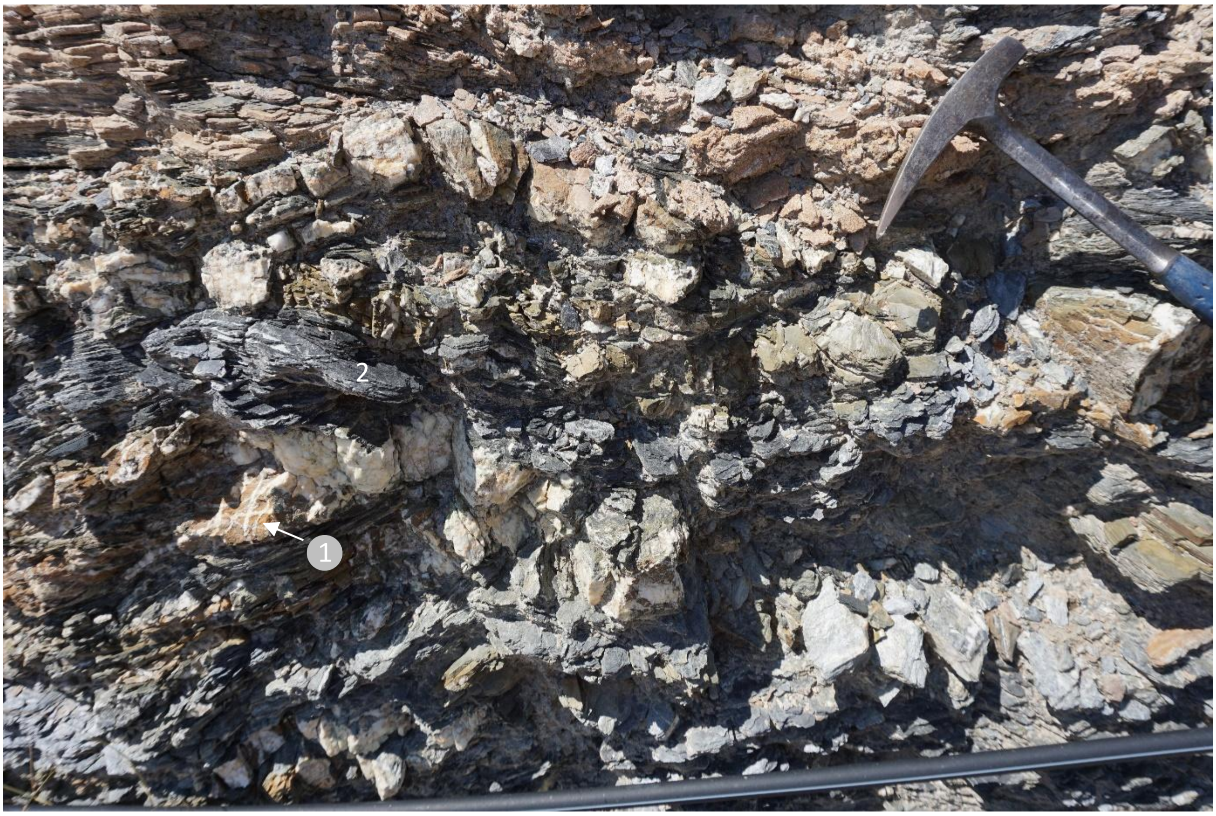
2 Undefined Upper Most Nappe



Location of outcrops, N35° 10‘12‘‘, E024°30‘39‘‘



Outcrop I: Calc-silicate schist



Outcrop I, Calc-silicate schist: 1: Light brown coloured calcite or other carbonate, 2: Phyllite
:

Outcrop I: Phyllite

Outcrop I, Calc-silicate schist: 1: Light brown coloured carbonate, 2: Phyllite

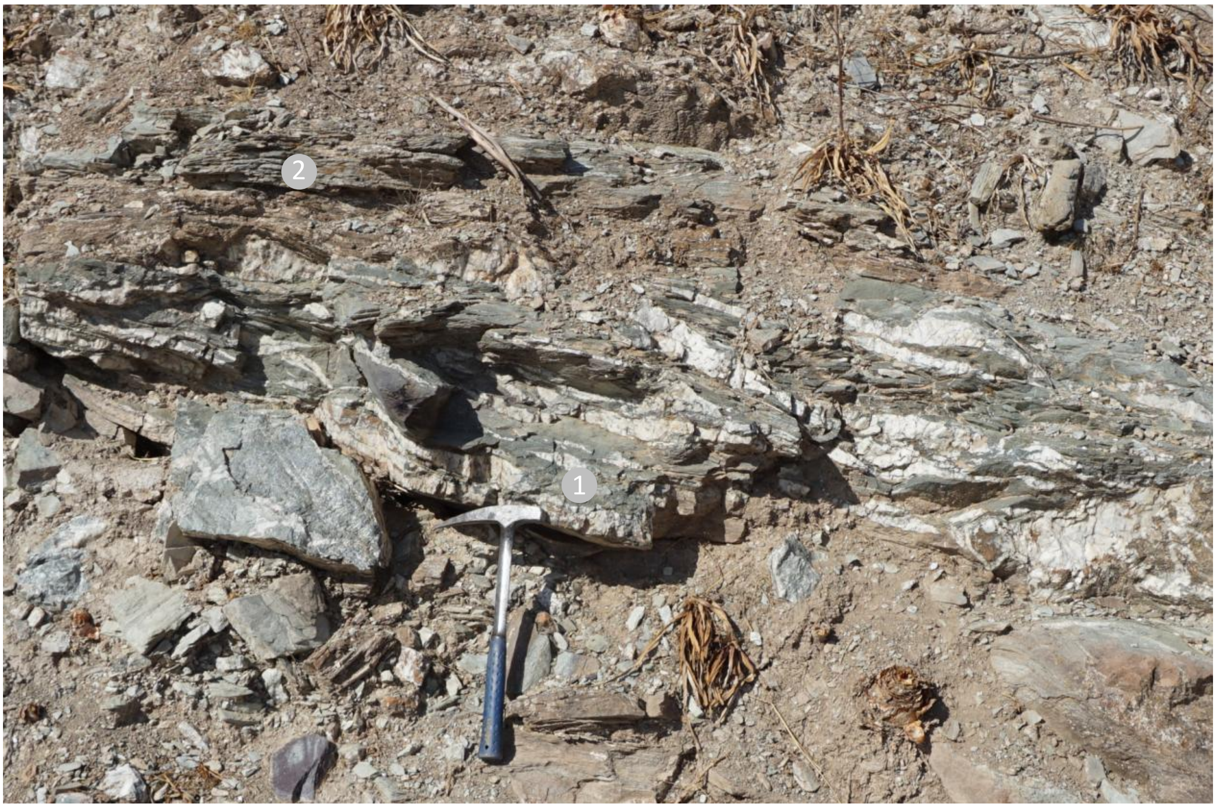
Outcrop I, Calc-silicate schist: 1: calcite is easily scratched with a steel nail.

Outcrop II: Two generations of folding, S_1: first foliation, D_2: approx. second direction of strain

Outcrop II: Scree from outcrop displaying schistosity and quartz veins as well as phyllite

Outcrop III, red and green silty phyllite

Outcrop III, sample of red silty phyllite, displaying muscovite on foliation planes.



Outcrop III, 1: mainly quartz veins in green schist, 2: Phyllite

Outcrop III: closeup of previous picture. Mostly quartz veins but also some calcite.

Outcrop III: top of foliation surface displaying quartz, muscovite and possibly chlorite, due to its green colour.

Outcrop IV: Calc-silicate schist

Outcrop IV: Calc-silicate schist, light brown layers consist either of calcite or possibly aragonite.

Outcrop IV: Calc-silicate schist. Sample displaying orange layers consisting of calcite (cal) or possibly aragonite (ar). Muscovite (mu) on foliation planes.

Location of outcrops [source of image : Google Maps]

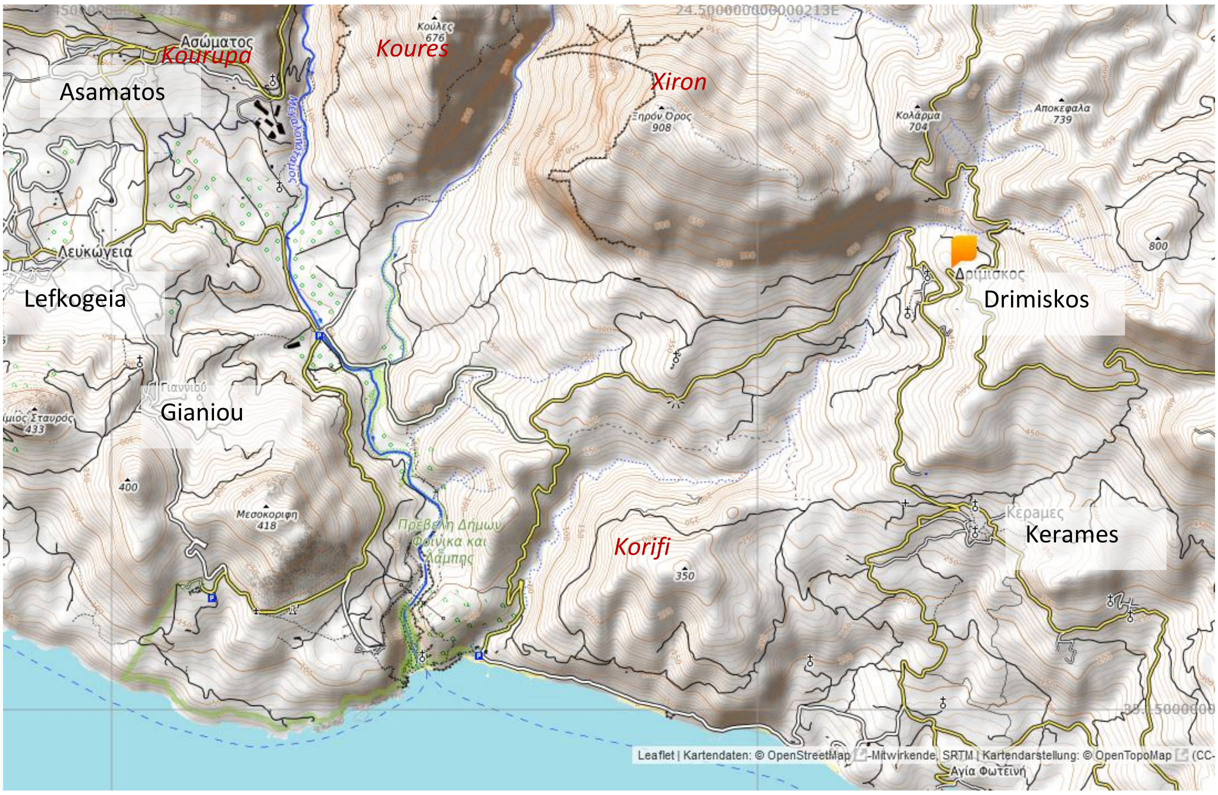
Outcrop V, undefined Upper Most Nappe: Calcsilicate schist and greenschist. 1: calcite, 2: greenschist

Outcrop V, undefined Upper Most Nappe: greenschist with quartz veins (see 2 Undefined Upper Most Nappe)

Outcrop V, undefined Upper Most Nappe, closeup of previous picture. 1: quartz veins, 2: greenschist (see 2 Undefined Upper Most Nappe)

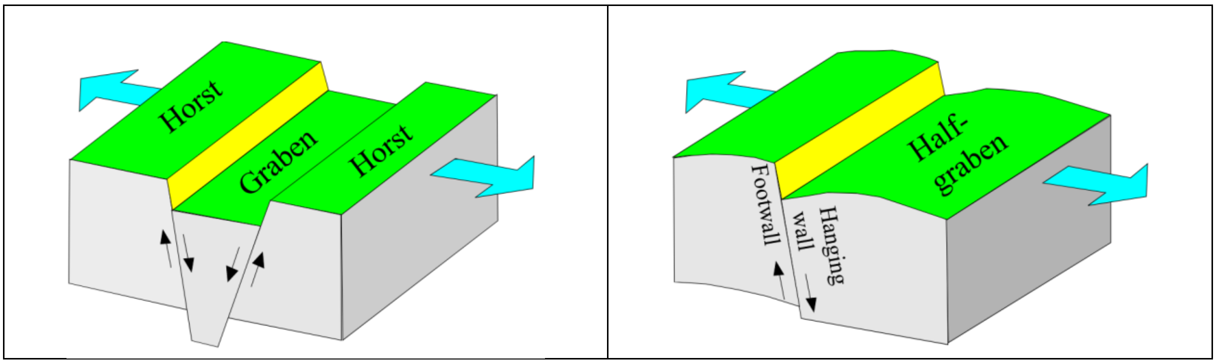
Outcrop V, undefined Upper Most Nappe, Calc silicate schist, 1: calcite, 2: greenschist (see 2 Undefined Upper Most Nappe)

3 Plakias Half-Graben



Topographic map of the area around Preveli

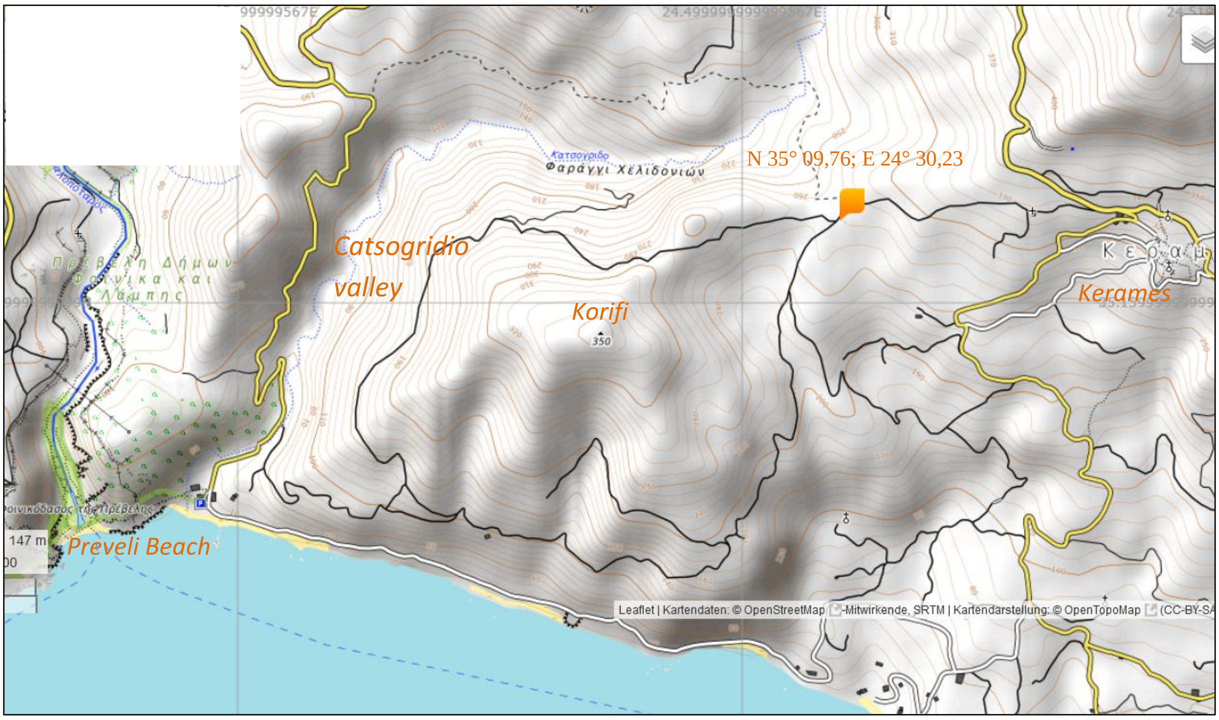
View of the Plakias half graben looking westwards from a point near Kerames (west of cemetery). 1: Small mountains at the south side of the graben (dog's teeth) are part of the half-graben and represent the Preveli Nappe. 2: The Koures Mt. consisting of Tripoliza limestone borders the northern side of the *graben and is part of the footwall. The foot wall has been slightly rotated creating a moderately dipping normal fault plane.*



Principle of Horst and Graben and Half-Graben structures caused by extensional forces

4 Ophiolites and Vatos Nappe West of Kerames

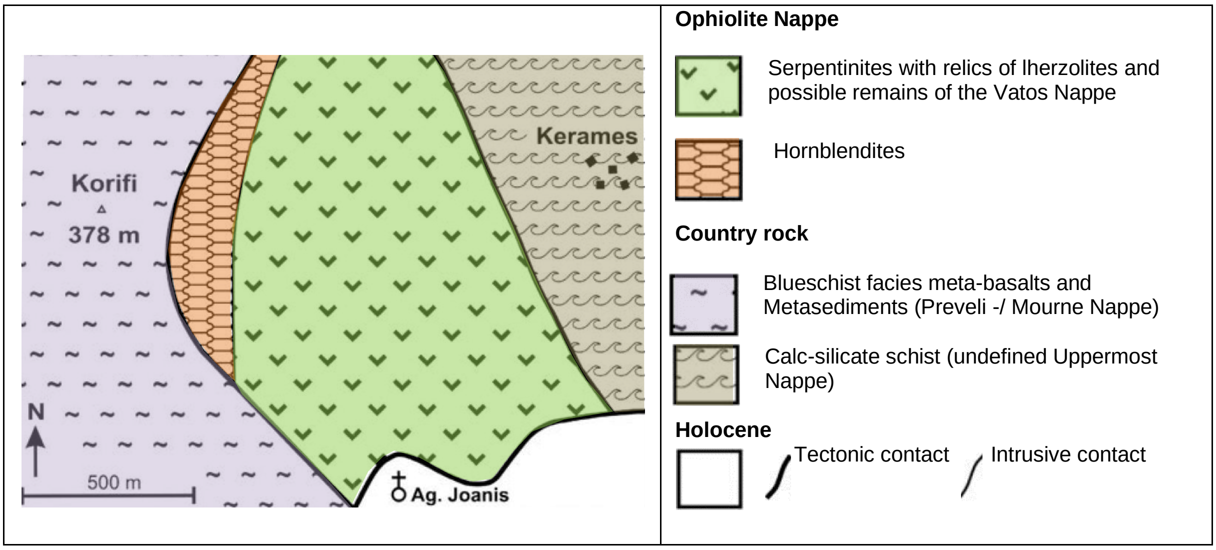
West of Kerames, lie the remains of the Uppermost Nappes. These consist of ophiolites in form of serpentinites and mafic/ultramafic intrusions, as well as the Vatos and the Preveli Nappes. A gravel road (N 35° 09.75; E 24° 30.42) branching off to the West from Kerames and leading past a cemetery provides a good view of the area and the existing serpentinized peridotite body. The serpentinite is reported to be of Jurassic age [Zulauf, 2023]. The serpentinites and mafic/ultramafic intrusions are thought to have been thrust on top of the Vatos Nappe as the Vatos Nappe is much young of Upper Cretaceous. The Korifi Mountain forms part of the Preveli Nappe.



Topographic map of the area west of Kerames

View of the ophiolites at the junction of two gravel roads (N 35° 09,76; E 24° 30,23). 1: Serpentinite, 2: Hornblendite, 3: Metasediments and meta-basalts (Preveli Nappe)

Location of outcrops west of Kerames. I: serpentinite, II: mafic pegmatite dike or zone, III: Vatos schist, IV: foliated hornbendite, V: hornblendite, VI: gabbro-diorite dike, VII: mafic pegmatite dike, VIII: dolerite dike [source of image : Google Maps]



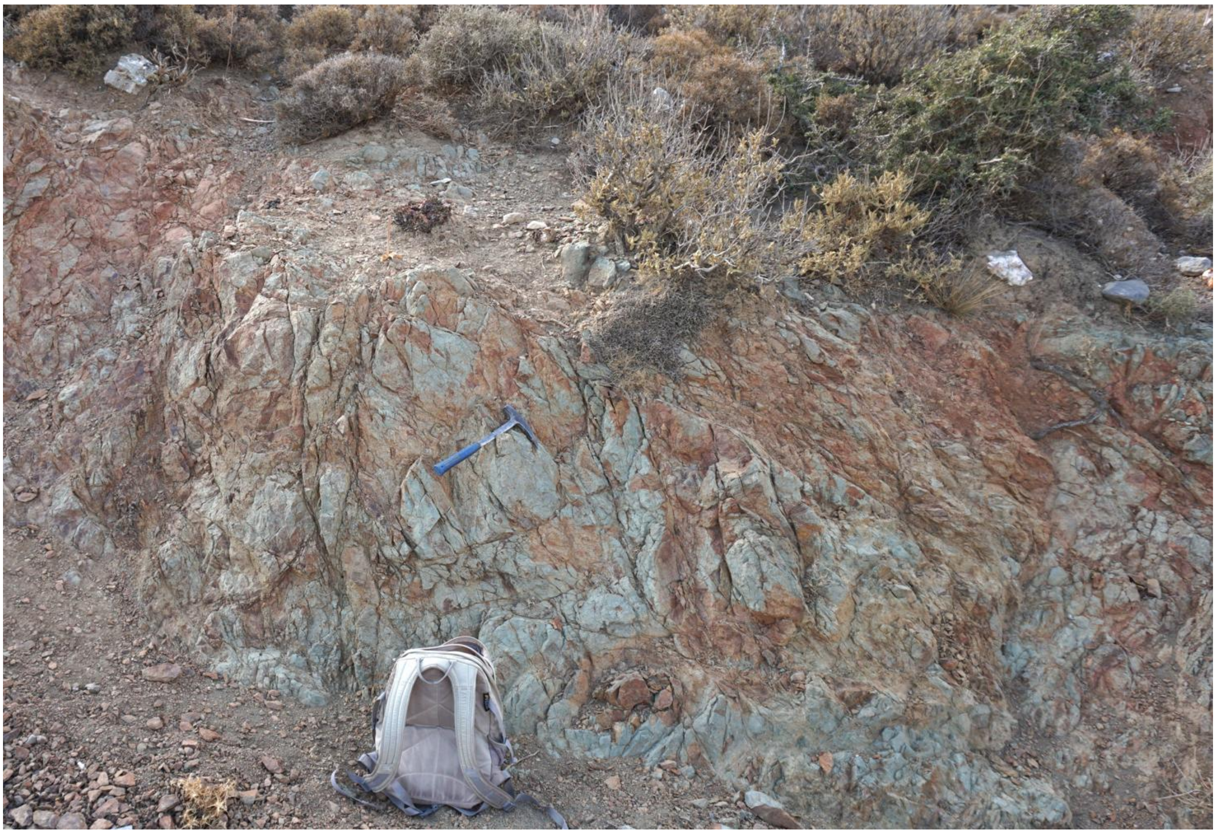
Rough sketch showing the existing rocks west of the Kerames village; modified after Bonneau and Lys (1978) and Skeries (1978) [Koepke J. et al., 2004], additional information concerning country rock Zulauf, (2023). Not to scale.

Outcrop I, serpentinite body. Serpentinite is poor in nutrients and does not weather to fertile soil. Most of the ultramafic bodies in the area have little vegetation although the climate provides adequate rain in Winter.

Outcrop I, serpentinite body. The serpentinite bodies at Kerames originated from lherzolite ultramafic rock. Characterized by high concentrations of Al_2O_3 and CaO these rocks are thought to have originated from primitive, non-depleted mantle material [Koepke J. et al., 2004]. The ultramafic rock may have formed at a slow-spreading midocean ridge (MOR).

Outcrop II, pegmatitic zone or dike. Highly altered pegmatitic zone or dike at the contact to the serpentinites. The mineral content consisting of chlorite and other minerals indicate alteration by metasomatic processes.

Outcrop II, pegmatitic zone or dike; closeup of previous picture. The mafic minerals have undergone alteration displaying probably chlorite and other secondary minerals. Rodingitization is reported to result in the formation of Calcium-aluminum-silicate minerals and can be considered here as a possible process of alteration See Appendix.



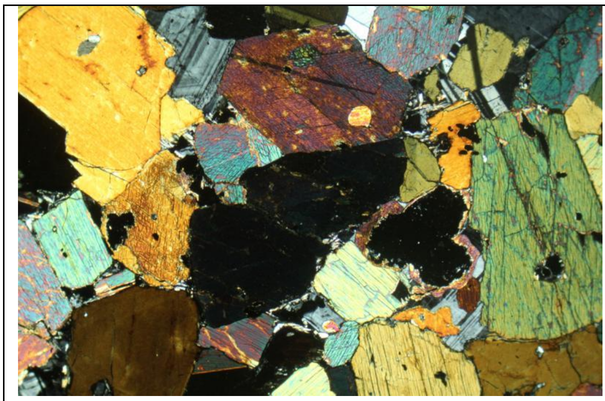
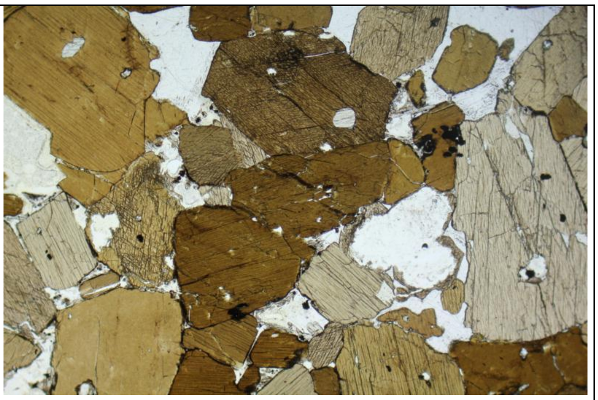
Outcrop III, possibly Vatos nappe: silty green clay stone

Outcrop III, possibly Vatos nappe, closeup of previous picture: silty green clay stone with conglomerate texture. The outcrop bears similarity with the Vatos schist found near the village of Vatos.

4.1 Hornblendite

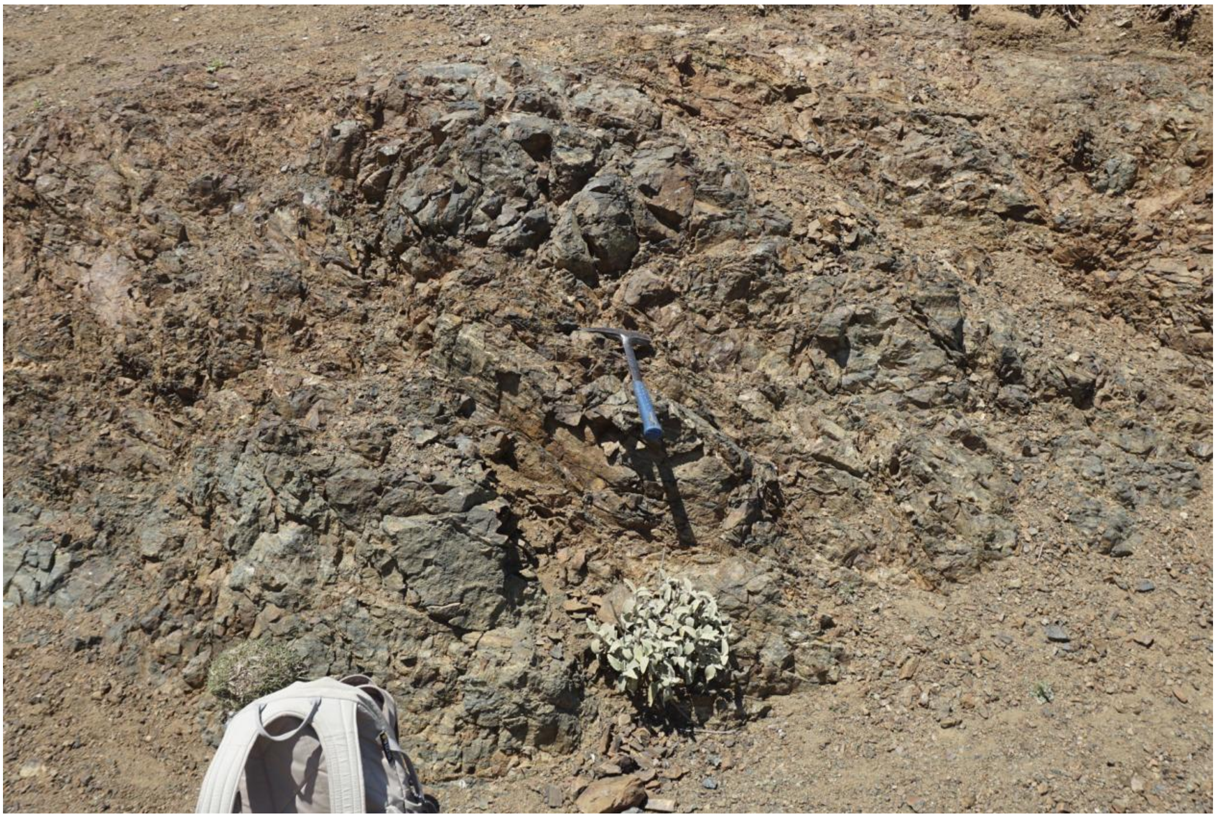
Further along the road towards the West is a hornblendite body (N 35° 09.76; E 24° 30.23). It belongs to the Ophiolite Nappe and dips towards E/SE at 40°-60° below the serpentinite.

A hornblendite is an ultramafic igneous rock dominated by hornblende. It is different from an amphibolite which is a metamorphic rock containing mainly amphibole and some plagioclase. Hornblendite is a rare rock most commonly associated with diorite, granite, gabbro and peridotite. They form as cumulates by settling of early crystallised hornblende in hydrous magmas, other related rocks such as hornblende gabbro and hornblende peridotite also form by accumulation.

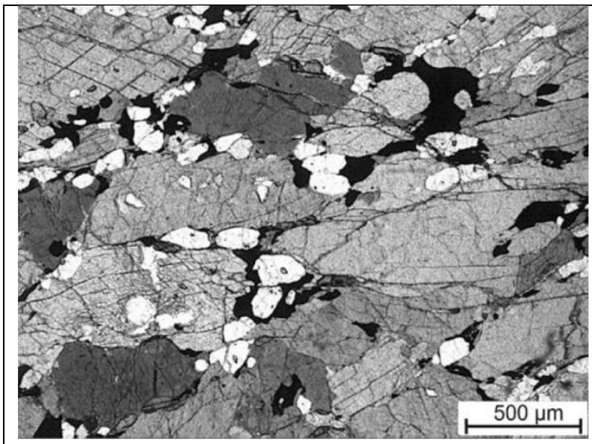
	
Hornblende crystals in a hornblendite rock. XPL image, 2x (Field of view = 7mm)/td>[Alex Strekeisen] https://www.alexstrekeisen.it/english/pluto/hornblendite.php	*A phaneritic texture occurs in plutonic rocks with tightly packed mineral grains of medium size (1 to 5 mm). The crystals are randomly distributed and not aligned in any consistent direction. A phaneritic texture is developed by the slow cooling and crystallisation of magma.*

The hornblendites of Central Crete, in their chemical composition, are reported to evolved from a typical iron- and titanium-rich gabbro as parent magma. Koepke, 2004 has shown that the hornblendites were generated by fractional crystallization and accumulation of amphibole in an iron-rich magma, which intruded the surrounding lherzolitic peridotites in Jurassic times *[Koepke, 1986]*.

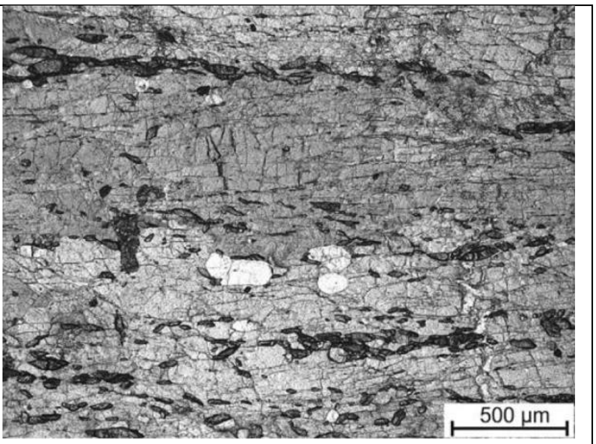
In the area the hornblendites are mostly foliated but some well-preserved less deformed phaneritic rock also occurs. The hornblendite forms a coherent sheet up to some tens of meters in thickness that lies beneath serpentinised lherzolites. In places it exhibits a marked layering due to alternating bands of black hornblendites and greenish coloured clinozoisite/epidote albite rocks. Like the serpentinites, the hornblendite is frequently intensely deformed, and forms meter-sized lenticular bodies in a cataclastic matrix of the same material *[Koepke, 1986]*.



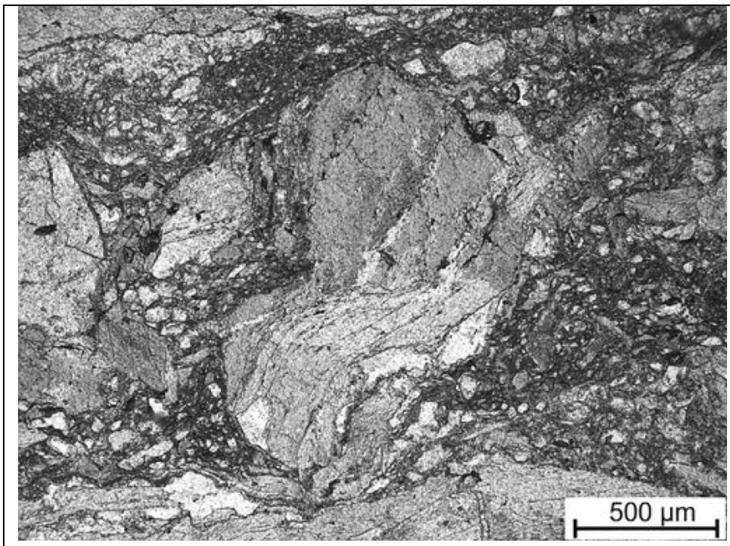
Outcrop IV, hornblendite. Highly deformed and metamorphically overprinted hornblendite schist displaying sigmoid shapes and shear stresses.



Black & white picture: (a) thin section of the hornblendites displaying relic magmatic cumulate texture: hypidiomorphic hornblende crystals (i.e. magnesio-hastingsite) with late-stage crystallised interstitial apatite and ilmenite.



Black & white picture: (b) strongly sheared hornblendite-schist exhibiting extreme flattening of crystal grains. Primary ilmenites were changed into tiny lenticular titanites which were aligned in disrupted strings at the contact between the platy amphibole grains (pargasite), accompanied by apatite.



Black & white picture: (c) sigmoid ductile deformed hornblende porphyroclast (pargasite) in a brittle shear zone; the cataclastic matrix consists of disrupted pargasites (same composition as the porphyroclast), titanite, and prehnite. A slow-spreading ocean ridge environment is suggested by the crystal-plastic deformation observed in some hornblendites. Such features are regarded as results of ductile deformation of the young, still hot oceanic lithosphere in high temperature shear zones, probably related to near-ridge transform faults. [Koepke, 1986].

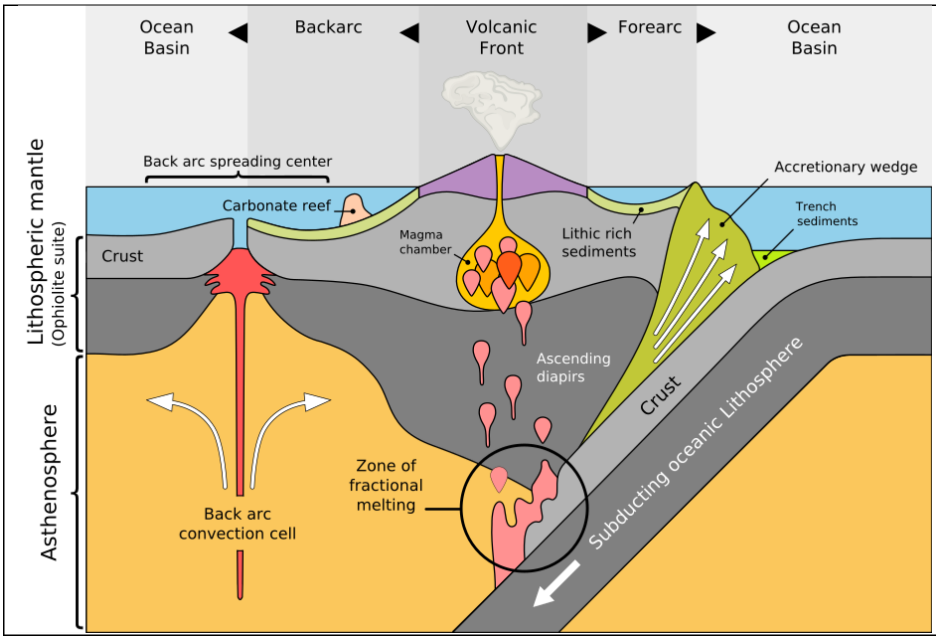
The above thin sections show that amphiboles initially crystallised in an evolved, Fe-rich magma, were accumulated probably on the floor of a magma chamber. Late-stage ilmenite and apatite filled the interstices. Then, the residual melt was squeezed out, probably due to compaction or high temperature shearing processes within the newly formed lithosphere. A post-magmatic, high-temperature shearing is often visible, indicated by an extreme flattening of the primary amphibole (predominantly pargasite) which did not change its composition (thin section, b). Some amphiboles show signs of plastic deformation like sigmoid deformation (thin section, c) *[Koepke, 1986]*.

Outcrop IV, hornblendite. Sample of hornblendite schist

Outcrop V, hornblendite. A less tectonically disturbed hornblendite

Outcrop V, hornblendite. Close up of previous picture showing phaneritic texture of the rock, which consists mainly of hornblende.

Major and trace element geochemistry reveal a magmatic evolution to extreme iron-rich liquids with high amounts of incompatible elements, consistent with a model of fractional crystallization of magmas with amphibole as dominant phase. The magmas were generated under strongly reducing conditions with water contents > 4 wt% permitting the onset of amphibole crystallization probably at temperatures around 1000°C. The assumed high-water contents in the initial melts indicate a subduction-related environment for the formation of the hornblendites. Note that the generation of magmatic hornblendites requires a melt with high water content, which is not sufficiently abundant at a "normal" mid-ocean ridge, where the MORB liquids show low water contents, varying between 0.1 and 0.5 wt% (e.g., Michael and Chase, 1987; Sobolev and Chaussidon). Therefore, it is likely that the location for the intrusion of the hornblendite crystal mush occurred in a high-temperature shear zone environment in a subduction related back-arc MORB setting *[Koepke, 1986]*.



Cross-section through the shallow part of a subduction zone showing the relative positions of an active magmatic arc and back-arc basin. Based on the hornblendite's high water content, it is likely to have been formed in a back-arc environment.

[Wikipedia, https://www.wikiwand.com/en/articles/Back-arc_basin]

U–Pb SHRIMP dating of zircons of the meta-hornblendite yielded 163 ± 3 Ma, which is interpreted as crystallization age (Liati et al. 2004). K–Ar ages of hornblende yielded ca. 156 Ma meaning that the hornblendite was formed close to the Middle/Late Jurassic boundary and emplaced and cooled down to T < 500 °C still during the Late Jurassic *[Koepke, 1986]*.

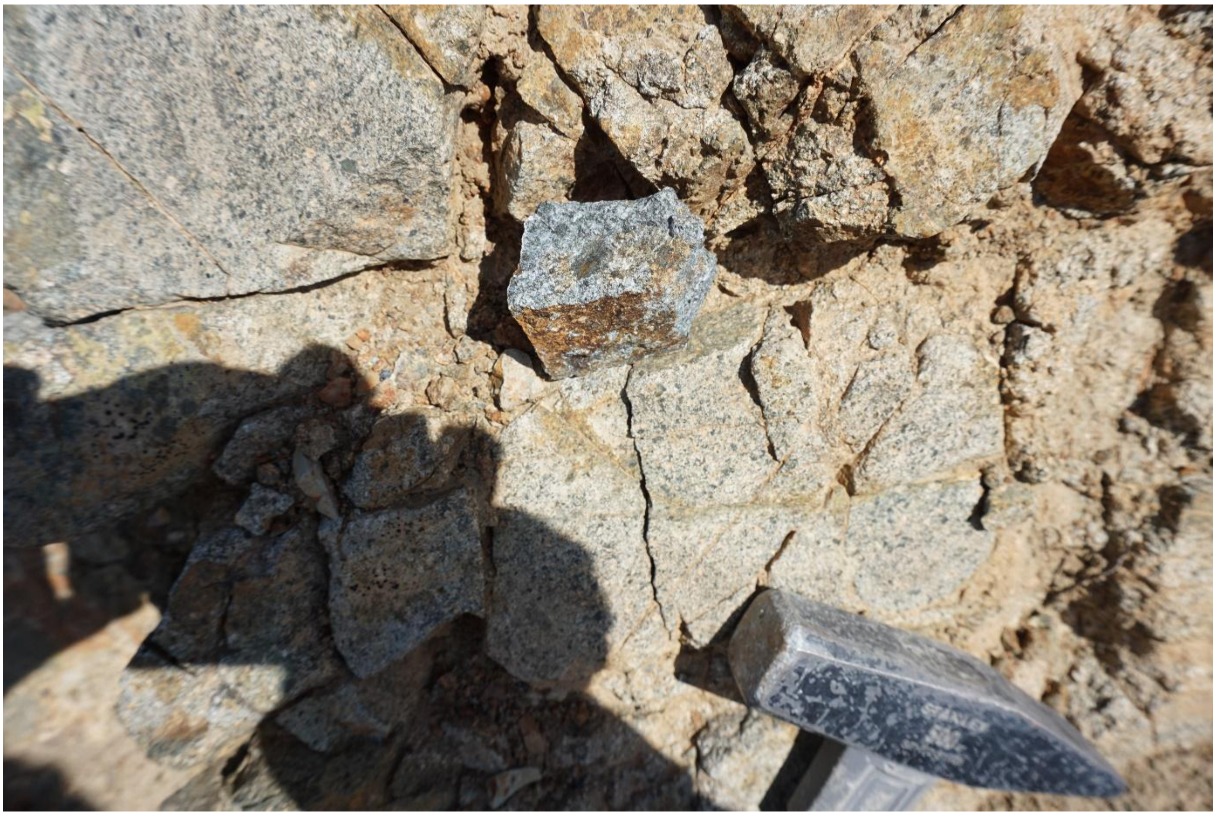
4.1.1 Rodingitization of hornblendites

In literature, rodingites are described as fine grained, dense, white, green or pink coloured rocks rich in calcium-aluminum-silicate minerals (Grossular, Phrenite, Wollastonite, Vesuvianite, Diopside, Tremolite, Chlorite, Spinel, Calcite, Quartz). They are locally developed along the boundary between serpentinite and country rock or as dikes, and sills, completely enclosed by serpentinite. Rodingites are also frequently found at the contacts between serpentinites and mafic or felsic intrusive bodies. Rodingite rocks are metasomatic products of alteration. Mass transfer, during metasomatic processes, is believed to be due to circulation, of an aqueous fluid of high pH and high Ca^{2+} and OH- contents during serpentinization *[Alex Strekeisen]*.

Near Kerames the contact between the hornblendite and serpentinite sometimes displays a decimeter-thick zone of rodingite. The intrusion of hornblendite magma is thought to have been prior to the process of serpentinization/ rodingitization, which later occurred in the high heat flow environment of a ridge system.

4.2 Gabbro, Dolerite and Pegmatite Dykes

The peridotites of Kerames, like those of other Cretan ophiolites, are intruded by gabbro-dioritic dikes that are mainly meter-sized, and range in composition from pyroxene gabbros to hornblende diorites. The dominance of hornblende as crystallizing phase in these rocks as well as the geochemical signature indicate a subduction-related origin for the dikes (Koepke et al., 2002). K-Ar amphibole dates obtained by Koepke et al. (2002) for the gabbro-dioritic dikes within the Kerames peridotites are around 20 Ma younger than those of the hornblendites. The gabbro-dioritic dikes were intruded into the peridotites after the serpentinization process. This is indicated by the formation of hydrous high-temperature parageneses at the contact between gabbro-dioritic dikes and serpentinites (presence of anthophyllite) and by the lack of rodingitization in the gabbro-diorite dikes. *Koepke et al*



Outcrop VI: Gabbro-diorite

Outcrop VI: Closeup of previous picture. The gabbro-diorite displays a porphyritic texture, i.e. there is a distinct difference in the size of the crystals.1: The large phenocryst appears to be a zoned pyroxene e.g. Augit.

Outcrop VII, pegmatite dike :1: chilled boarder, 2: large crystals in centre of dike

Outcrop VII, pegmatite dike: weathered sample from centre of the pegmatite dike, displaying large crystals. The absents of quartz and the presents of hornblende indicates mafic to ultramafic composition. The brown orange colouring of the hornblende is due to weathering and probably arises from the oxidization of Fe.

Outcrop VII, pegmatite dike: chilled boundary of the pegmatite dike

Outcrop VIII: Dolerite dike

Outcrop VIII, dolerite dike. Dolerite is a mafic, holocrystalline, subvolcanic rock chemically equivalent to volcanic basalt or gabbro. In terms of mineral composition, gabbro and dolerite are identical, however as subvolcanic rock dolerites are much more finely grained.

At the contact between hornblendites and peridotites intense rodingitization developed, whereas none of the numerous gabbro diorite dikes show signs of rodingitization. This indicates that the phase of ocean floor alteration resulting in serpentinites and rodingites was prior to the dike intrusion event.

Location of outcrops south-west of Kerames.

Outcrop IX: Pegmatite dike displaying reduction of crystal size towards the walls of the country rock.

Outcrop IX, closeup of a sample from the pegmatite dike

Outcrop X: Weathered dolerite dike

Outcrop X: Closeup of previous picture

4.3 Ultramafic intrusion

Outcrop XI: Ultramafic intrusion. The dome shape is now less obvious because of faulting.

Outcrop XI: ultramafic intrusion

Outcrop XI, ultramafic intrusion: weathered surface of the ultramafic intrusion

Outcrop XI, ultramafic intrusion: closeup of the ultramafic intrusion

4.4 Serpentinite

Outcrop XII, serpentinite: highly sheared serpentinite body. Rounded loose boulder of serpentinite in the centre of the picture.

Outcrop XII, serpentinite: typical snake like pattern occurring in serpentinites in an advanced stage of alteration.

4.5 Vatos Nappe?

The country rocks surrounding the serpentinite bodies are similar in appearance to the rock formation near the Vatos, Aktounda and Ardaktos villages and are therefore thought to be part of the Vatos Nappe, which exists there. In general, 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).

Of particular interest are limy metaconglomerates, that display rodingite and other ophiolitic pebbles (Krahl et al. 1982), and which contain Late Jurassic/Early Cretaceous foraminifera (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 owing to the age of the existing fossils. U–Pb dating of calcite that 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 Nappe: country rock: possible olistostrome consisting of shale matrix and limestone clasts. Large boulders are made of grey marble. 1: shale matrix, 2: marble. Owing to the size of the marble clasts this formation could also be called "Wild Flysch" similar to the country rocks found near Vatos.

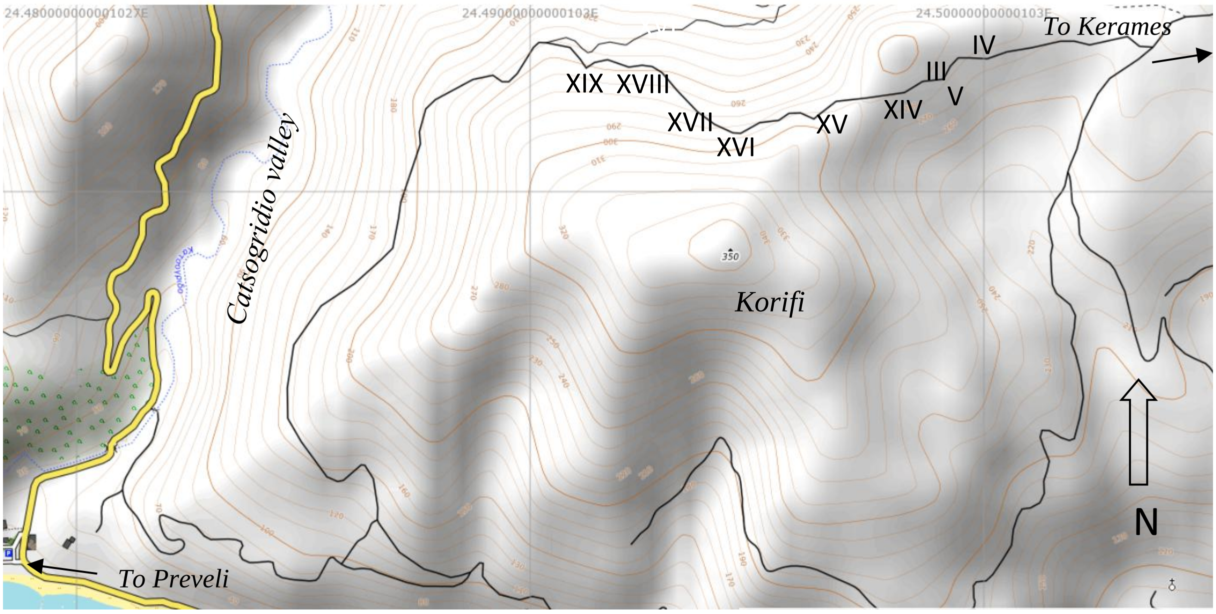
Outcrop XIII, Vatos Nappe: possible olistostrome displaying matrix supported texture, 1: shale matrix, 2 limestone/marble clasts

Outcrop XIII, Vatos Nappe: stratified marble boulder

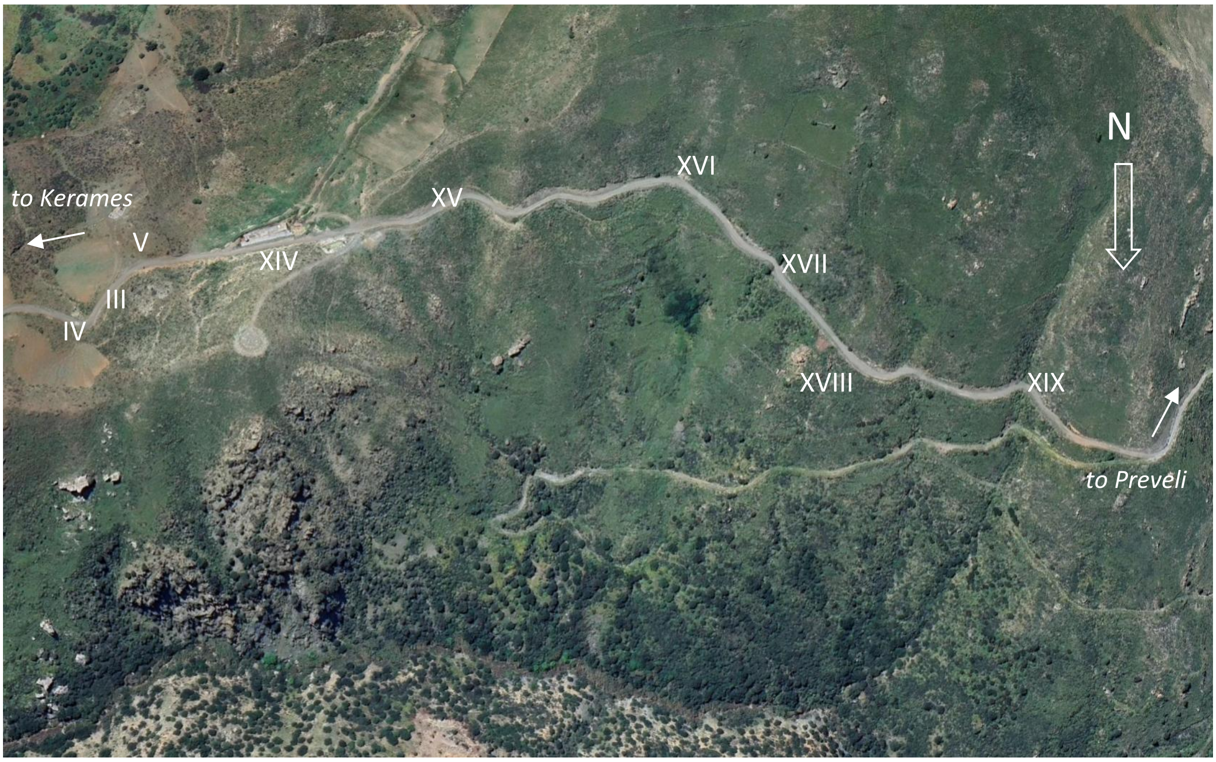
5 Preveli Nappe at Korifi Mountain

In the vicinity of the Korifi Mt. the hornblendites are tectonically separated from the country rock that consists of amphibolites, schists and marbles. Note that authors in the past have classified the Uppermost Nappes quite differently, so that some regard the Korifi mountain to belong to the Preveli Nappe, while others have proposed that the Preveli Nappe is associated with the “Vatos Nappe”. Further north in the Spili-Graben and near the villages Vatos, Aktounda and Ardaktos geologists (Zulauf et. al, Koepke) have mapped the remains of a Mourne Nappe, a Vatos Nappe and an Ophiolite Nappe. The Uppermost Nappes are thought once to have had a much wider extent than and today’s remaining relict nappes and it remains to be seen how they can be correlated with each other under general consensus.

West of the Ophiolites and hornblendites, the summit of the Korifi mountain is formed by the Preveli Nappe, which consists of mica schist, amphibolite and phyllite. In general, the Preveli Nappe comprises high-pressure metamorphic metabasalts and metasediments at the base, and flyschoid sediments at the top. Fossils found further south (Bonneau & Lys 1978) indicate Permian age. According to Koepke et al. (1997) these rocks were metamorphosed under the epidote-blueschist facies conditions.



Location of outcrops



Location of outcrops at Korifi Mtn. III: Vatos schist, IV: foliated hornbendite, V: hornblendite, XIV: calcschist, XV: greenschist, XVI: limestone and phyllite schist, XVII: greenschist, XVIII: limestone/marble conglomerate, XIX: greenschist [source of image: Google Maps]

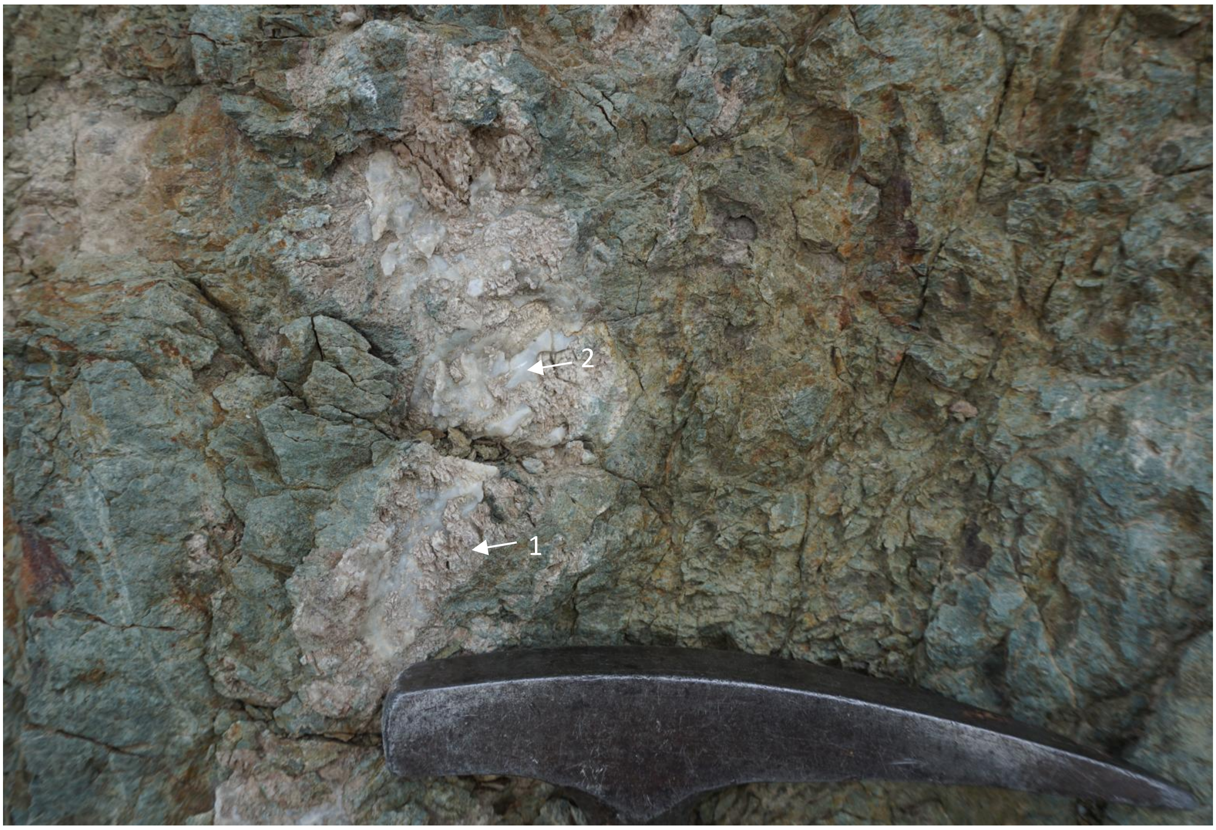
Outcrop XIV: calcschist

Outcrop XIV: calcschist

Outcrop, XV: green schist

Outcrop, XV: green schist (epidote-blueschist facies?)

Outcrop, XV, greenschist. 1: quartz veins 2: greenschist



Outcrop XV, greenschist. 1: calcite vein in greenschist with two different generations of calcite. 1: first generation idiomorph crystals, 2: second generation filling.

Outcrop XVI, phyllite schist. 1: phyllite schist, 2: calcschist

Outcrop XVI, , phyllite schist. Closeup of previous picture showing phyllite schist.

Outcrop XVII, greenschist

Outcrop XVII, closeup of previous picture

Outcrop XVII, greenschist

Outcrop XVII, greenschist, closeup of previous picture

Outcrop XVIII, limestone/marble conglomerate

Outcrop XVIII, limestone/marble conglomerate

Outcrop XIX: greenschist

Outcrop XIX, greenschist. Sample displaying crenulation cleavage, which is a sign of different foliation planes.

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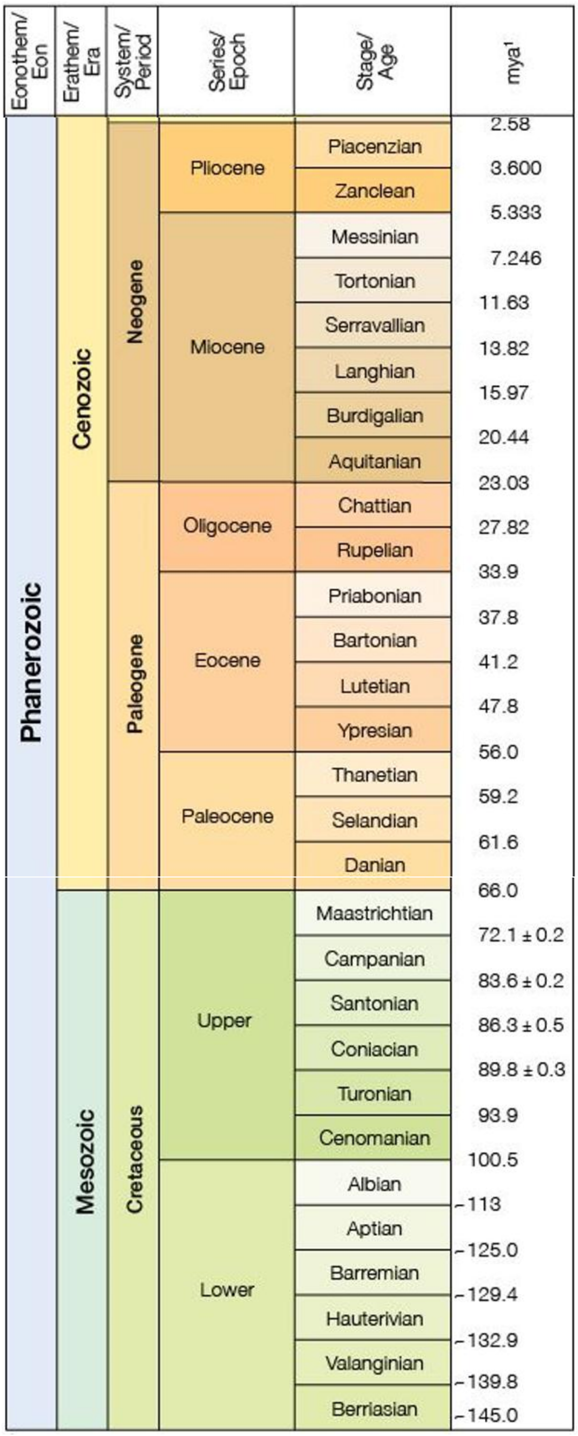
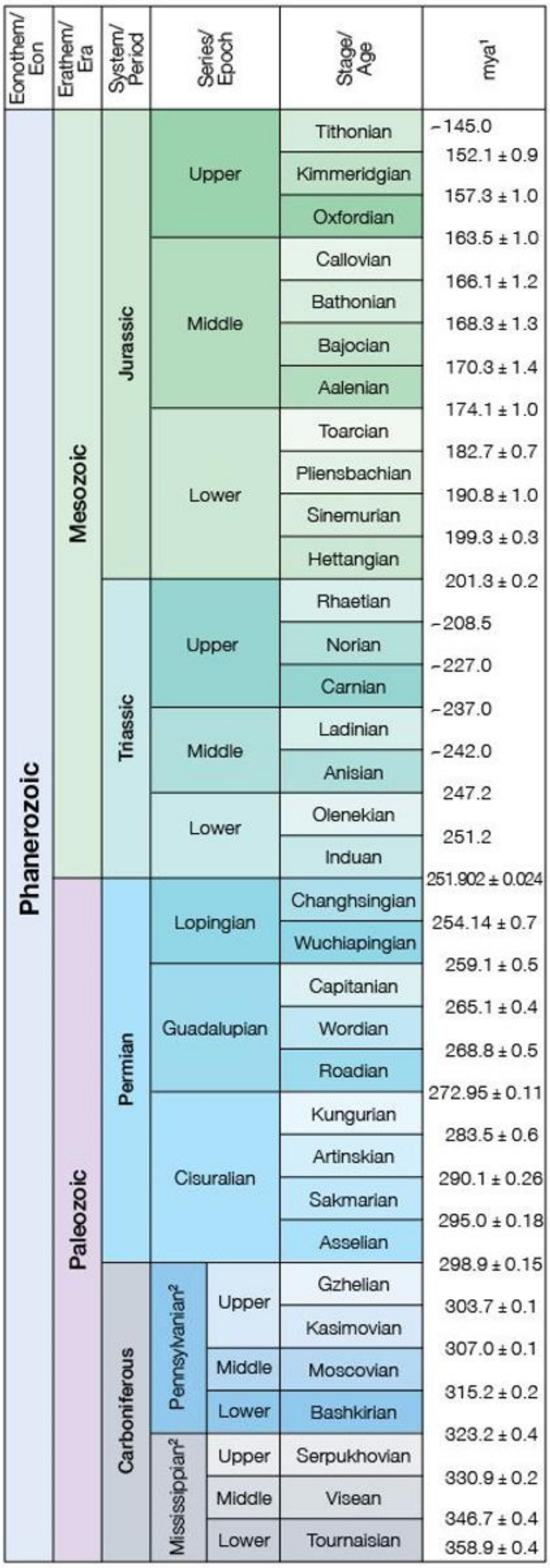
6 Appendix

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

Origin of Magma

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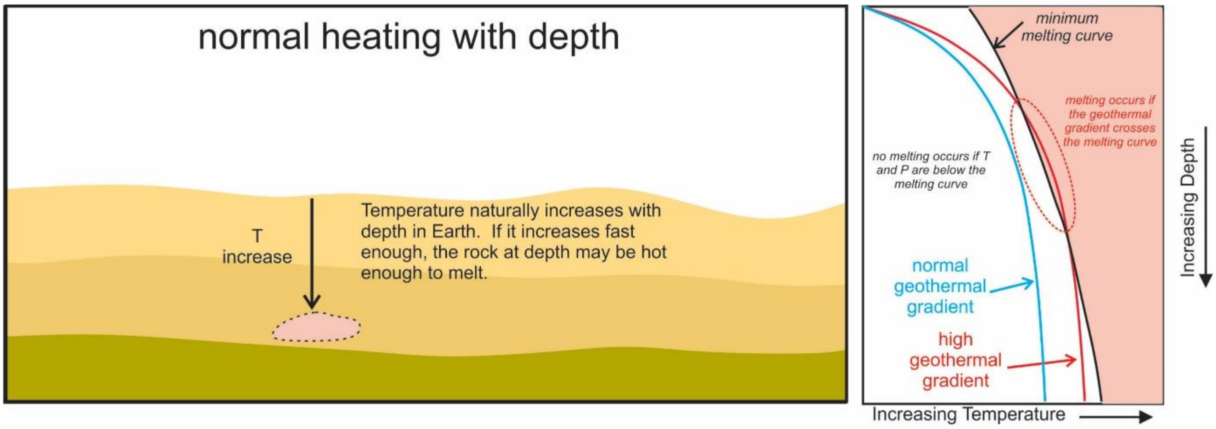
https://opengeology.org/petrology/03-magma/

Just as ice melts when temperature goes above 0 ºC, rocks will melt if heated to temperatures above their melting temperatures. To accomplish this rock melting requires extra heat, and that poses a problem. Where is the extra heat to come from? Although radioactive decay of potassium, uranium, thorium, or other radioactive elements may create small amounts of heat, most of Earth's heat is left over from the original time of formation. This residual heat flows from Earth's interior to dissipate at the surface, and Earth has been cooling for more than 4.5 billion years. In some places, flowing magma delivers extra heat, but the origin of the heat necessary to initially create the magma is problematic.

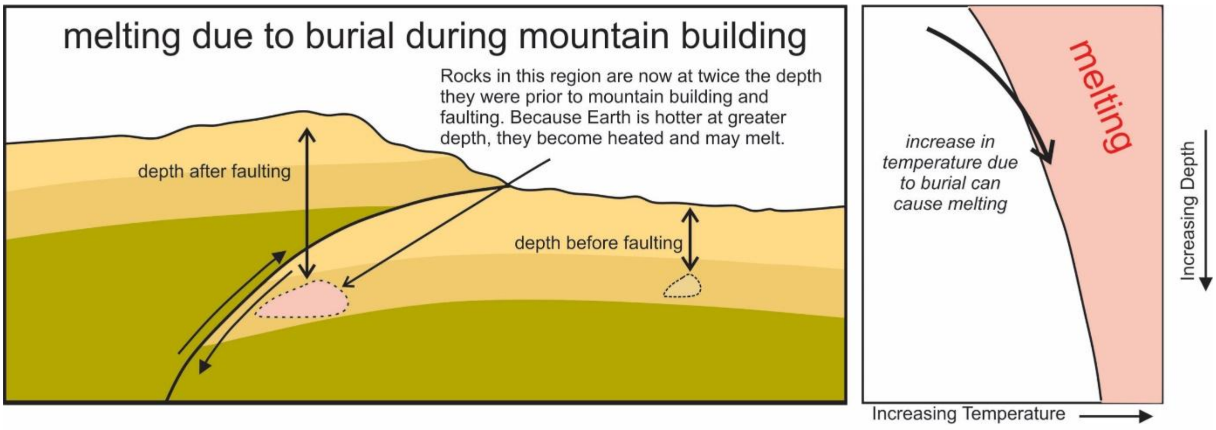


3.18 Heating with depth in Earth

Earth's average *geothermal gradient* (the rate at which temperature increases with depth in Earth) is about 25 ºC/km near the surface. It is not the same everywhere – some places, such as mid-ocean ridges or hot spots like Yellowstone, have very high gradients, sometimes exceeding 50 ºC/km. Other places, such as centers of old continents, have low gradients. The blue and red lines in the temperature-depth diagram on the right of Figure 3.18 are *geotherms*. The geotherms show schematically how temperature increases with depth for a place with an average gradient and for a place with a high gradient.

The solid black curve in the temperature-depth diagram of Figure 3.18 is a typical *melting curve*; it shows the minimum temperatures at which melting can occur. If temperature-depth conditions plot in the red part of the diagram, rocks will melt. The minimum melting temperature increases with depth but so do temperatures along the geotherms. In principle, if the geothermal gradient is high enough, the temperature may exceed the melting curve (shown where the red "high" geothermal gradient line crosses the black melting curve in Figure 3.18). Yet, even at mid-ocean ridges or hot spots, the gradient is generally insufficient for this to happen and cause melting. Consequently, most of Earth's mantle is unmelted.

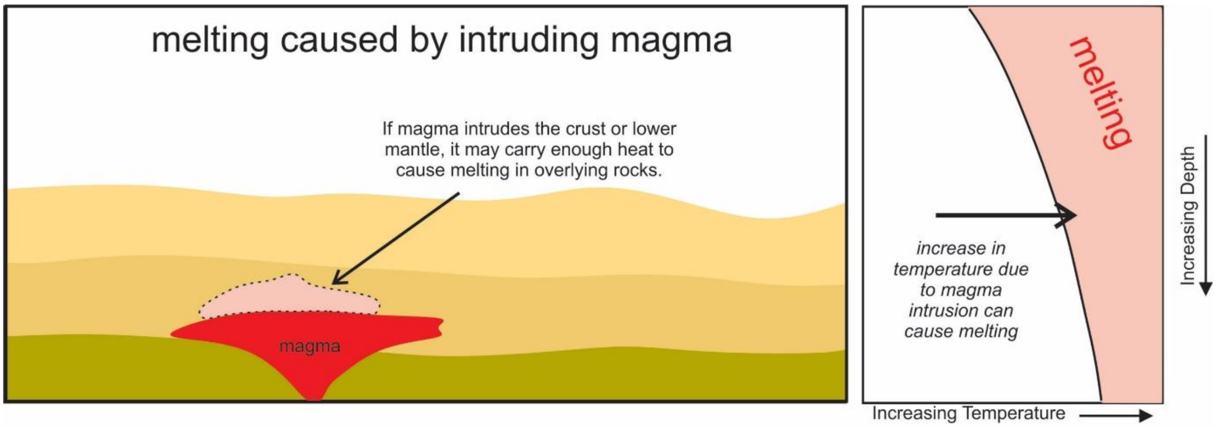
Melting Caused by Mountain Building



3.19 Melting during mountain building

Because a normal geothermal gradient cannot lead to melting, other mechanisms must be responsible in most cases. For instance, *tectonism* associated with mountain building can occasionally cause melting when rocks are buried by folding or faulting – because heating naturally accompanies burial (Figure 3.19). The pressure-temperature diagram on the right side of this figure shows that if burial-induced heating is great enough, melting may occur when temperatures cross into the melting field (shown in pink). Some granites in continental regions undoubtedly form by melting of sediments and sedimentary rocks, once at or near the surface, that melted after being carried to depth.

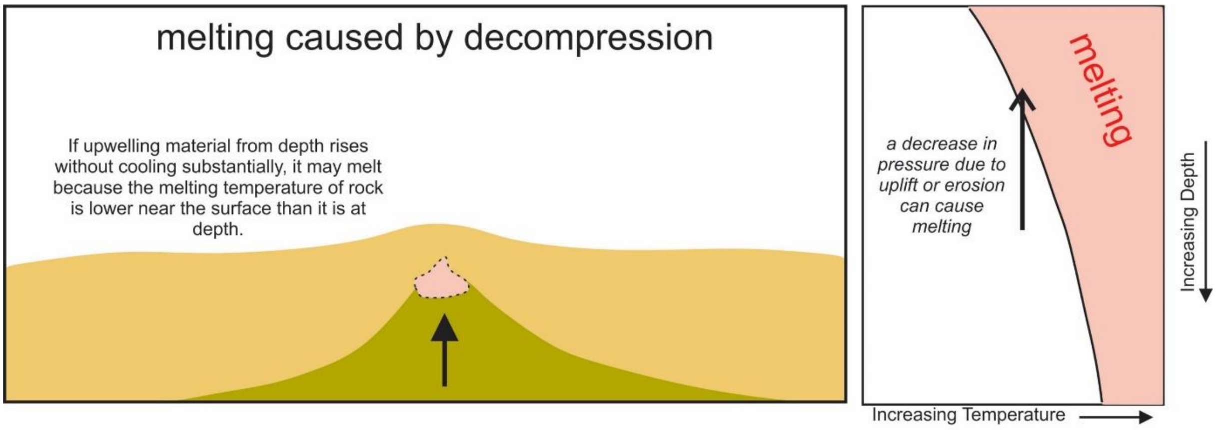
Melting Caused by Intruding Magma



3.20 Melting caused by magmatic intrusion

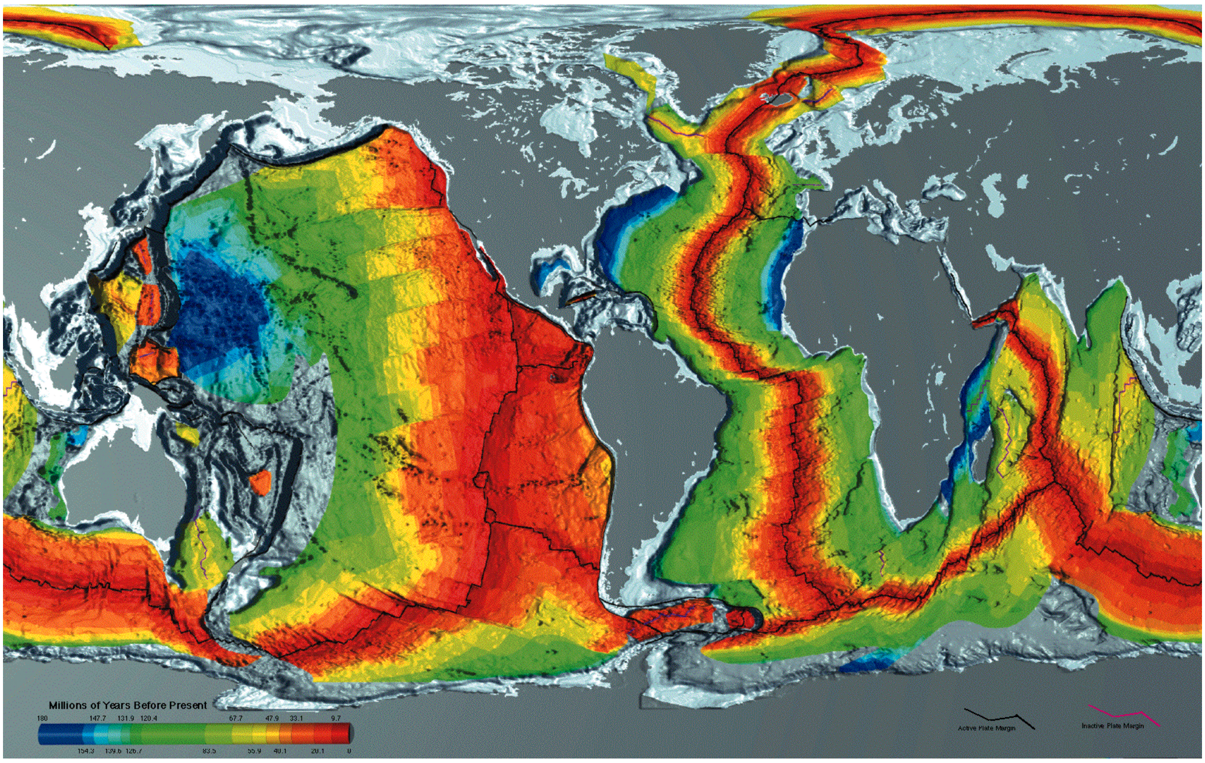
Intruding magma is a very efficient mechanism for delivering heat that can cause melting (Figure 3.20). As shown by the black arrow in the pressure-temperature diagram, heat from magmas can cause temperature to increase without any increase in pressure. As discussed earlier in this chapter, beneath Yellowstone National Park, rising magmas from the mantle are hot enough to cause overlying crustal rocks to melt. In subduction zones, magmas rising above a subducting plate may cause melting in the overlying continental lithosphere, creating silicic magmas that erupt in subduction zone volcanoes or crystallize underground to become plutons.

Decompression Melting



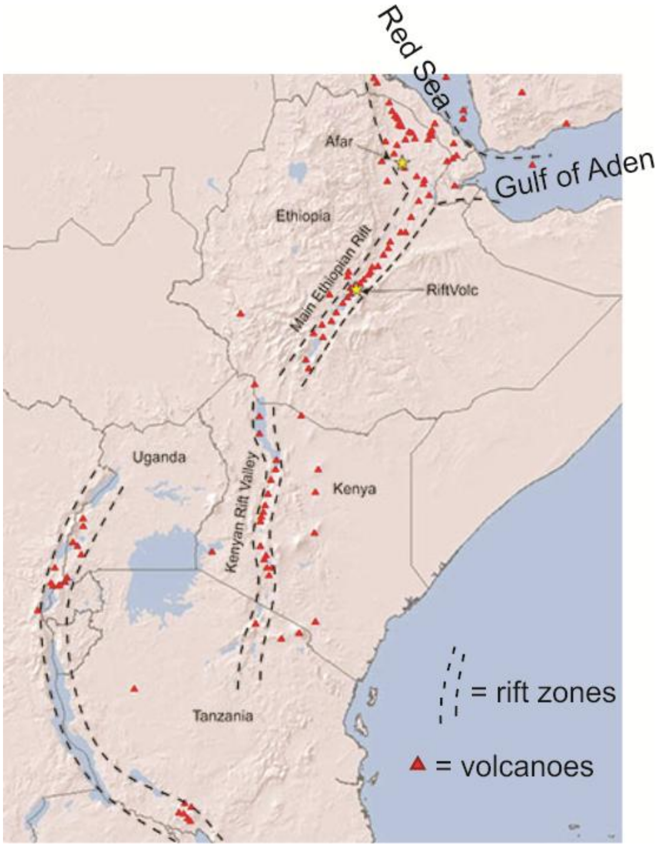
3.21 Melting caused by decompression

The processes described above may all cause melting, but they do not account for the widespread melting that occurs at mid-ocean ridges. There, an additional, and very significant, mechanism promotes melting – a decrease in pressure (Figure 3.21). Rising mantle moves up to fill the void created by sea floor spreading. As shown by the black arrow in the pressure-temperature diagram, the resulting pressure decrease leads to melting because rock melts at lower temperature when at low pressure, compared with high pressure. This process, called *decompression melting*, generates more magma than any other Earth process.



3.22 Age of the ocean floor: red (young) to blue (old)

So, decompression melting is the key mechanism producing magmas at mid-ocean ridges, which, although we don’t generally see them, are the most active volcanic settings on Earth. In these settings, partial melting of rising solid rocks produces basaltic lavas that erupt on the ocean floors, and upon cooling, are added to the spreading oceanic lithosphere. Thus the youngest oceanic crust – shown in red in Figure 3.22 – is next to mid-ocean ridges and the oldest oceanic crust (light and dark blue) is found along ocean basin margins.

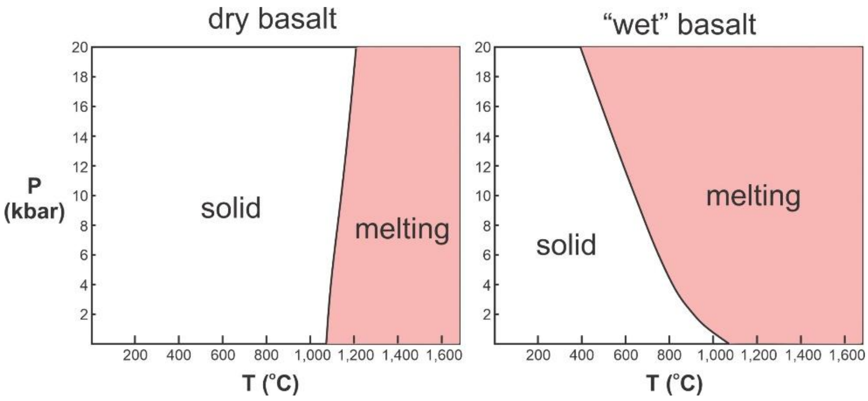


3.23 Volcanoes of the East African Rift

Decompression melting also leads to igneous activity where continental rifting occurs, for example along the East African Rift. The East African Rift is an elongate zone that includes the Kenyan Rift and the Main Ethiopian Rift (Figure 3.23). Along this narrow zone, the African continent has begun to split apart; the rift may eventually be the site of a new ocean basin similar to the Red Sea or the Gulf of Aden, also shown in Figure 3.23. Besides causing melting at rift zones, decompression contributes to the melting associated with more localized hot spots, like those under Yellowstone, where warm rocks move upwards due to buoyancy.

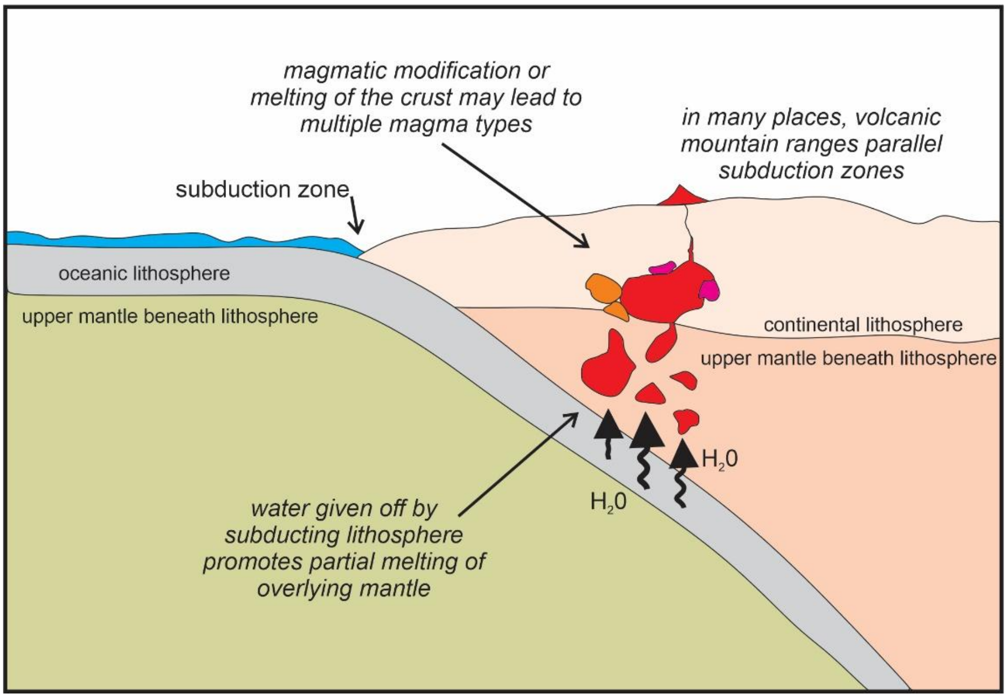
Flux Melting

Principles of thermodynamics tell us that two things combined will often melt at a lower temperature than they would individually. An analogy is a mix of ice and salt, which we all know melts at a lower temperature than ice alone. Similarly, rocks will melt at lower temperatures when they contain water, CO_2, or another volatile compared to when dry.



3.24 The effects of water on melting temperature of basalt

The presence of a small amount of water can change magma melting temperatures by hundreds of degrees. Figure 3.24 compares the conditions that cause melting (shown in red) of a basalt that contains no water and one saturated with a small amount of water. “Dry” basalt begins to melt at temperatures in excess of 1,000 °C. Depending on pressure, melting of “wet” basalt may begin at temperatures significantly lower. Because water and other volatiles lower melting temperatures in the same way that a *flux* is used to lower the melting temperature of metals, this additional mechanism for melting is called *flux melting*. If a rock is already hot, addition of only a small amount of water can promote melting. Water is the most important geological flux, but CO_2 and other gases also promote melting in some settings.

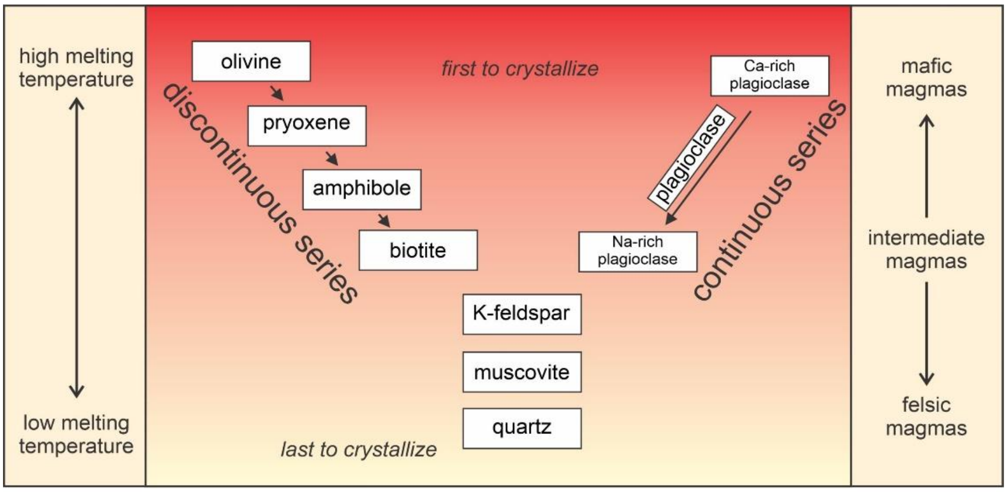


3.25 Flux melting in a subduction zone

Flux melting is especially important in subduction zones (Figure 3.25). Subducting oceanic lithosphere contains hydrous minerals that react during metamorphism to form anhydrous minerals. Consequently, water is released and migrates upwards into hot overlying mantle. This water lowers the melting temperature, causing partial melting of the ultramafic mantle to produce mafic (basaltic) magma. This magma then migrates upwards and may reach Earth’s surface. It may also promote additional melting in the uppermost mantle or crust, or may become modified to produce magmas of different compositions.

Bowen’s Reaction Series

When rocks melt or magmas crystallize, things are generally more complicated than when a single mineral melts or crystallizes. Most rocks contain more than one mineral, and different minerals melt or crystallize at different temperatures. N. L. Bowen, an early 20th century petrologist, conducted many laboratory experiments and was the first to compare melting and crystallization temperatures of common igneous minerals in rocks.



3.*33 Bowen's Reaction Series*

Bowen's Reaction Series (Figure 3.33) depicts Bowen's fundamental findings. We call it a *reaction series* because during melting (or crystallization) solid minerals continuously react with surrounding liquid.

Mafic magmas crystallize at high temperatures and felsic magmas at lower temperatures. Bowen's series shows the relative liquidus temperatures for common minerals for magmas of different compositions. As shown by the series, olivine crystallizes at the highest temperature (from mafic magmas) and quartz at the lowest temperature (from felsic magmas).

One sometimes confusing thing about this series is that it depicts liquidus temperatures for minerals when they crystallize from magmas of complex chemistry. Crystallization temperatures are different for the individual minerals if they are by themselves. For example, ignoring minor complications involving polymorphs, quartz may not begin to crystallize from a granitic melt until other minerals have formed and temperature drops to less than 1,275 ^{o}C. In contrast, quartz will crystallize from a melt of 100% SiO_2 composition at a much higher temperature (in excess of 1,700 ^{o}C).

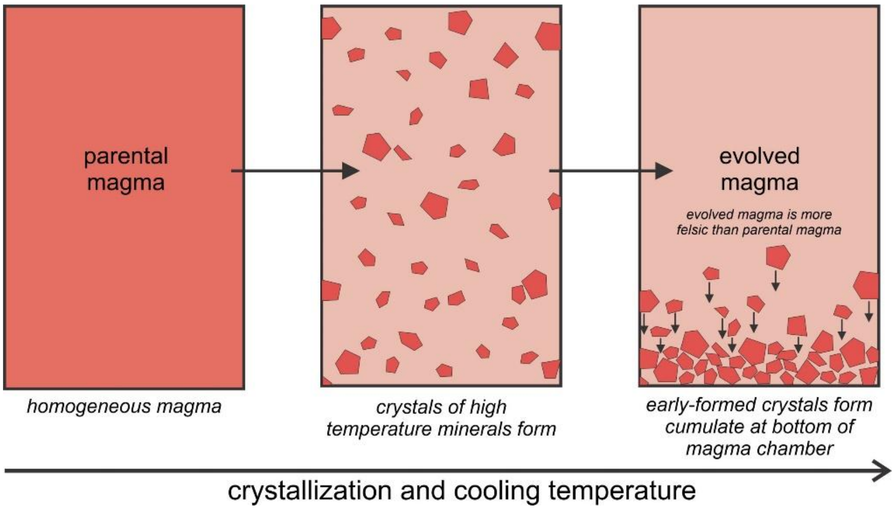
Plagioclase melts incongruently over a range of temperature depending on its composition; Bowen called the plagioclase side of the diagram the *continuous series*. Most other minerals, also depending on their compositions, melt sequentially over more restricted temperature ranges (*discontinuous series*). Bowen found that (mafic) minerals common in ultramafic and mafic rocks have the highest liquidus and solidus temperatures, and (silicic) minerals that are common in silicic rocks have the lowest. Consider a cooling magma: as temperature decreases, minerals higher up in the series crystallize first followed by minerals lower down. We call minerals that melt and crystallize at high temperatures *high-temperature minerals*; those that melt and crystallize at low temperatures are *low-temperature minerals*. The specific minerals that crystallize, however, vary with magma composition.

Although melting and crystallization in the order depicted by Bowen's Reaction Series seems straightforward, there are many complications. No magmas follow the entire series – most crystallize only one or a few of the minerals in the series – and some magmas crystallize minerals that are not part of the series. Furthermore, some minerals melt (and crystallize) congruently and some do not. Additionally, many melting and crystallization reactions involve more than one mineral reacting together. For example, at Earth's surface, anorthite melts at about 1,560 ^{o}C and diopside melts at about 1,390 ^{o}C. A rock that contains both anorthite and diopside, however, will begin to melt at around 1,270 ^{o}C, a much lower temperature than the melting point of either individual mineral.

For most magmas, crystallization begins at some maximum temperature and continues over a range of temperatures until everything is solid. As the process continues, different minerals form at different temperatures and the composition of the magma changes. The opposite occurs during heating of a rock. Melting generally begins by melting of low-temperature minerals followed sequentially by melting of higher-temperature minerals until all is liquid. During this process, the melt continually changes composition. Bowen's reaction series is a model that serves to remind us that the minerals that form depend on magma composition, that different minerals melt and crystallize at different temperatures, that mafic minerals tend to crystallize before silicic ones, and that silicic minerals melt at lower temperatures than mafic minerals do. The series does not, however, apply in detail to any known magma or rock composition. And, as pointed out above, the series only applies to rocks and magmas – it does not tell us about the melting and crystallization temperatures of individual minerals when they are by themselves.

Fractional Crystallization

Fractional crystallization, the opposite of partial melting, occurs when a magma partially crystallizes and the remaining magma becomes segregated from the crystals. In these circumstances, the new *evolved magma* will have a different composition from its *parental magma*. The evolved magma, which is more silicic than its parent was, may move upwards, leaving the high-temperature (mafic) minerals behind. Fractional crystallization, like partial melting, has been a key process contributing to differentiation of Earth.



3.*37 Diagram showing fractional crystallization*

Fractional crystallization may occur when newly formed crystals sink to the bottom of a magma chamber and no longer stay in equilibrium with the melt. Figure 3.37, a schematic diagram, shows the principles involved. While cooling, a parental magma crystallizes some high temperature minerals. These minerals eventually sink to the bottom of the magma chamber, leaving an evolved magma above. Because high-temperature minerals are mafic, the evolved melt is more silicic (less mafic) than the original parent magma. During this process, a *cumulate* rock forms at the bottom of the magma chamber, and the evolved magma may move upwards and become completely separated from the cumulate.

Other Processes Explaining Variations in Magma Composition

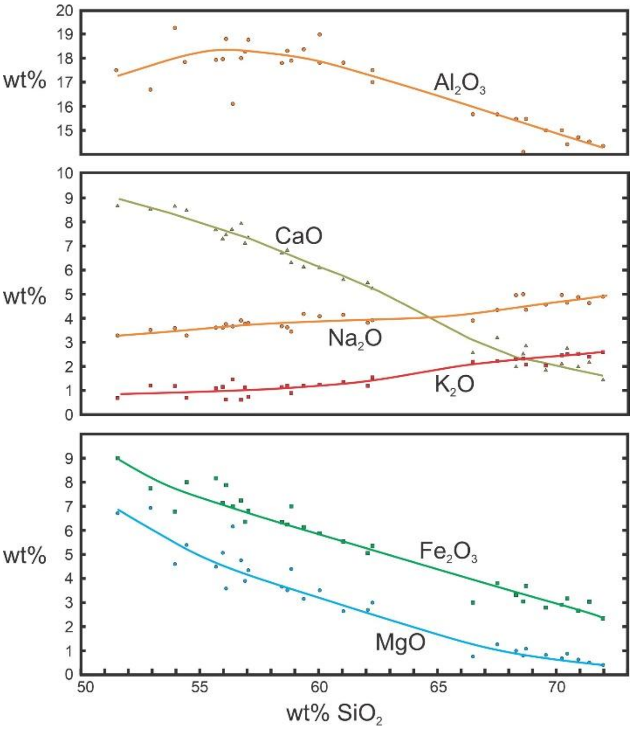
Fractional crystallization is undoubtedly the most important process that changes magma composition after a magma forms. Other mechanisms, however, also lead to changes. For example, in some settings, hot magmas may melt surrounding rocks and assimilate them into the magma. Generally, we think of this *assimilation* occurring when mafic magmas encounter more silicic rocks, because mafic magmas may be hotter than the silicic rock's melting temperature is. So, assimilation can make magma more silicic and is most likely to occur in the (silicic) crust. Some volcanic rocks contain crustal xenoliths, inclusions of rock fragments incorporated as solid pieces into the melt; often the xenoliths show evidence of partial melting. It is no stretch to assume that sometimes xenoliths melt and mix in completely. Some geochemical data, too, supports the idea that crustal material has been incorporated into a mantle-derived melt.

Parental Magmas and Differentiation

The majority of magmas, evolve from some parental magma. They are evolved melts, not melts having the composition created during initial melting. Subsequently, as crystallization progresses, magma compositions follow what is called a *liquid line of descent*, producing a series of magmas of different compositions as fractional crystallization removes specific minerals from the melt.

If solid mantle melted directly, either partially or completely, to create magma, the magma would be called a *primary magma*. Primary magmas have undergone no differentiation and have the same composition they started with. Specifically, if they come from the ultramafic mantle, and were not subsequently modified, they must have a very high Mg:Fe ratio and be enriched in Cr and Ni just like mantle rocks, and petrologists use these and other characteristics to test if magmas could be primary magmas. Most magmas fail the tests, and primary magmas are exceptionally rare, or may not exist at all. Some magmas and rocks, however, come close to being primary, and petrologists describe them as *primitive*, meaning they have undergone only minor differentiation.

Parental magmas may be primary or primitive. The only requirement is that they lead to magmas of other compositions. If a collection of melts with different compositions evolve from the same parent, they form a *magma series*. Although the melts have different compositions, they will share some chemical characteristics, especially trace element compositions and isotopic ratios. A challenge for petrologists is to study the compositions of an inferred magma series to learn the composition and source of the original parent.



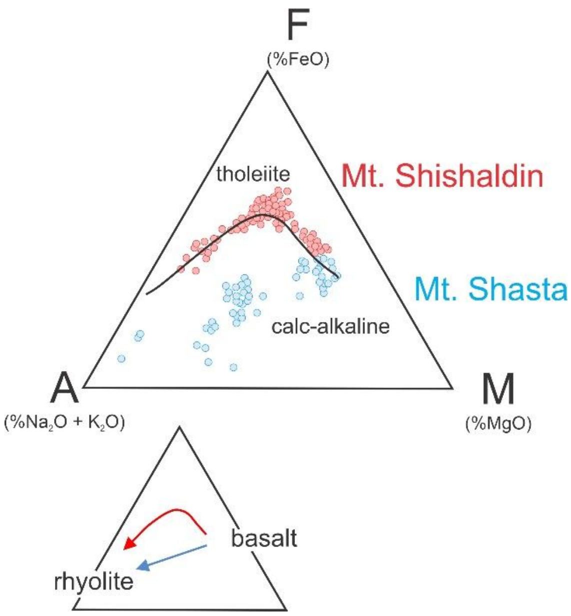
3.40 Harker diagram for volcanic rocks of the Crater Lake region, Oregon

Typically, petrologists begin their quest by obtaining analyses of the rocks and plotting the results on different kinds of *composition diagrams*. For example, *Harker diagrams*, first used in 1902, have SiO_2

content as the horizontal axis and other oxides plotted vertically (Figure 3.40). SiO_2 is chosen as the abscissa because it generally shows the most variation of all oxides, and because it relates closely with magma temperature and the amount of fractional crystallization.

When looking at Harker diagrams, the principles are that (1) if derived from a common parent, rock compositions should trend smoothly across a diagram; and (2) the most mafic composition is closest to the parent magma composition. So, if a Harker diagram reveals smooth trends, it is possible that all the magmas derived from the same parent and that the low SiO_2 end of the graphs are closest to the magma's parent composition. Harker diagrams are only one kind of composition diagram; many others with different oxides on the axes are commonly used.

Figure 3.40 shows a well-studied Harker diagram for volcanic rocks from near Crater Lake, Oregon, based on the data of Howell Williams (1942). Each point represents a different volcanic rock from the same region; the horizontal axis shows the SiO_2 content of the rock and the vertical axis the amount of other oxides present. The solid lines show the smoothed trends. The smooth trends are evidence that the different rocks may have derived from the same original parental magma. The Crater Lake magmas range from *basalt* (on the left side of the diagram) to *rhyolite* (on the right side). Based on the trends shown, Williams concluded that the magmas all came from a common parent magma and that they evolved by fractional crystallization. The basalt composition is closest to that parent.



3.41 AFM diagrams for rocks from Mt. Shishaldin, Alaska, and from Mt. Shasta, California

A second commonly used way to look at magma composition is to plot compositions on an AFM diagram (Figure 3.41). AFM diagrams ignore SiO_2 and instead look at alkali (Na_2O + K_2O), iron as FeO (assuming it is not Fe_2O_3), and MgO content. The triangle corners are: A = alkali oxide weight %, F = FeO weight %, and M = MgO weight %. Many studies have found that magma series follow one of two trends, the *tholeiite* trend or the *calc-alkaline* trend, and we easily see these on an AFM diagram. Figure 3.41 is an AFM diagram comparing rocks from Shishaldin Volcano (Aleutian Islands) and Shasta Volcano (California). Each point represents an analysis of an individual rock. Shishaldin is an island arc volcano associated with an oceanic plate subducting under another oceanic plate. Shasta is a continental margin volcano where an oceanic plate is subducting under a continental plate. The Shishaldin data follow a tholeiite trend, depicted by the solid line and red arrow that initially moves toward the F-corner before curving downward toward the A-corner. The Shasta data follow a calc-alkaline trend (depicted by the blue line that heads directly toward the A-corner).

Whether tholeiitic or calc-alkaline, originally mafic magmas can produce rocks ranging from basalt to rhyolite, as the bottom triangle in Figure 3.41 shows. At both Shishaldin and Shasta Volcanos, more primitive parental magmas were mafic and the later evolved magmas were silicic. They differ, however, because tholeiitic magmas become iron-rich as they evolve, moving initially toward the F apex of the triangle. Calc-alkaline trends go directly from basalt to rhyolite.

The trends on an AFM diagram reveal clues about the environments in which the magmas differentiated. The difference between calc-alkaline and tholeiite trends is due to the oxidation state of iron. If iron is mostly oxidized, magnetite (Fe_3O_4), a mineral that contains oxidized iron (Fe^{3+}), crystallizes early from a melt. If the iron is mostly reduced (existing as Fe^{2+}), magnetite does not crystallize. In calc-alkaline magmas, the iron is oxidized, leading to crystallization of magnetite. Consequently, when mafic minerals crystallize, iron is removed from the magma as fast as magnesium and the melt's Fe:Mg ratio remains about constant during differentiation. In tholeiitic magmas, olivine and pyroxene crystallize first and magnetite may not crystallize at all. Olivine and pyroxene have high Mg/Fe ratios compared with melt, and the magma becomes enriched in iron during the initial stages of crystallization. Calc-alkaline magmas are dominant in *andesitic-type* subduction zones, such as California's Cascade Mountains. Mt. Shasta is an example. Tholeiitic trends occur mostly in island arcs, such as the Aleutian Islands, and Shishaldin Volcano is an example.

Pegmatites

https://geologybase.com/pegmatites/

Pegmatites are intrusive or plutonic igneous rocks with wholly crystalline, unusually large, or coarse crystals that may sometimes interlock. These rocks form from a low-viscosity fluid phase rich in volatile compounds formed during the crystallization of the last magma portion. Usually, pegmatitic rocks will have crystals that are at least 1 cm in diameter, with the larger ones more than 3 cm and some as large as 10 m. Pegmatite crystals may vary in size, but some pegmatites may be equigranular – have nearly equal crystal sizes. Also, they may be spatially zoned and directionally oriented.

Most pegmatites have quartz, feldspar, and mica, with a silicic composition similar to granites. However, this texture can be shown by mafic rocks like basalt, gabbro, syenite, etc. Pegmatite or pegmatitic texture only describes the texture, not the mineral composition. Therefore, we may use the rock name as a prefix to indicate the mineral composition. For instance, granite pegmatite, diorite pegmatite, or gabbro pegmatite suggest that the said rock has granite, diorite, and gabbro composition, respectively. However, there is no unanimity on naming.

Pegmatite bodies are smaller than usual magma intrusions. Therefore, they are commonly found as small pockets, veins, dikes (including dike swarms), or lenses, especially at the margins of large intrusions and batholiths. Geographically, pegmatites occur anywhere in the world, and they are commonly found on cratons and greenschist-facies (Barrovian facies) metamorphic belts.

Since most pegmatites have a granitic composition, pegmatites are pinkish, gray, or whitish. However, you can find those that are colorless or darker (dark gray or black) if they have mafic minerals. Also, depending on the unique mineralogy, you will find some with red, brown, green, or other colors.

Mineral composition

Most simple pegmatite will have mainly granite to granodiorite composition. Such rocks will mostly have large crystals of feldspar, quartz, and mica (like lepidolite, biotite or muscovite). Also, they may have any other minerals associated with granitic rocks like aplite. However, some will have nepheline syenite composition. It is like granitic with quartz replaced with nepheline, a feldspathoid. And in rare cases, you may have pegmatite with intermediate to mafic rock composition.

Most complex pegmatites have a composition like granite but are enriched with some incompatible elements and minerals not typically found in most igneous rocks. Some may also be of intermediate to mafic composition. Incompatible elements are those with a too-large or small atomic number or very high or low ionic charge or size. So they cannot participate in the rock formation. Instead, they will remain dissolved in water and eventually form their minerals. Examples of incompatible elements in pegmatites are uranium (U), niobium (Nb), boron (B), yttrium (Y), phosphorus (P), fluorine (F), lithium (Li), zirconium (Zr) and beryllium (Be). Others are Strontium (Sr), Barium (Ba), scandium (Sc), Tantalum (Ta), thorium (Th), cesium (Cs), Tin (Sn), Titanium (Ti), Bismuth (Bi), and some Rare Earth Minerals (REEs).

These incompatible elements may form minerals such as pollucite (Cs), spodumene (Li), Beryl (Be), tantalite-columbite (Ta, and Nb), tourmaline (Li, B), cassiterite (Sn). Also, complex pegmatites may also have beryl, lepidolite, topaz, apatite, garnet, emerald, spodumene, tourmaline, mica, monazite, and fluorite. Other minerals are amblygonite, triphylite, molybdenite, scapolite, columbite etc. These minerals are not typical in ordinary igneous rocks. Lastly, the large crystal size of some pegmatitic rocks makes it hard to determine mineral composition accurately.

Formation of pegmatites

At the initial stages following a significant magma intrusion, minerals with higher melting points will start crystallizing and slowly be depleted from the magma. As the process continues, the remaining magma is increasingly enriched with minerals with lower melting point as well as with, water, and volatiles. These volatile compounds include carbon dioxide, boron, fluorine, chlorine, and phosphorus. Also, any incompatible minerals or elements will remain in this melt, specifically in water. At the final stage of magma crystallization, the exceptionally high amount of dissolved water causes a *phase separation*. So, that ultimately a last magma melt phase and a fluid phase or superheated water phase rich in silica, alkalis, volatiles, and incompatible trace elements develops. The fluid phase is what forms pegmatites. The presence of volatiles, i.e., water, carbon dioxide, chlorine, and fluorine, will tremendously lower the viscosity of the fluid phase (hydrous fluid). So, ions or molecules can move, migrate, or diffuse quickly to the crystal growth site. Also, the crystal growth rate (ions or molecules joining a growing crystal) exceeds nucleation (formation of new nuclei or sites for crystallization). Otherwise, the pegmatites would have many much smaller crystals.

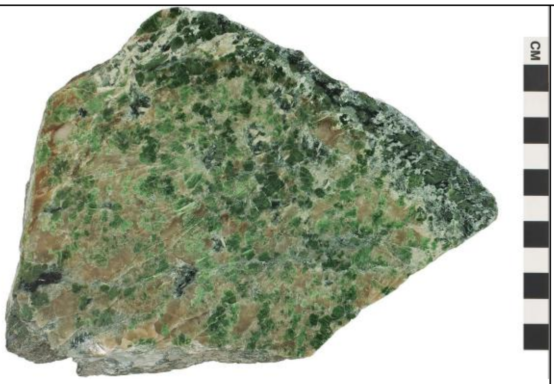
Lastly, pegmatite formation does not have to be associated with intrusive rocks or magma. Such cases may be from melting or anatexis of metamorphic rock under high pressure and temperature. The melting will make fluids, volatiles, and trace elements sweat out, forming a fluid phase pocket seen in felsic gneiss.

Emeralds (green), schorl tourmaline (black), and garnet (pinkish-reddish) in quartz (medium-brown glassy) and feldspar (grayish) Crabtree Pegmatite from Devonian of North Carolina, USA. Photo credit: James St. John, Wikimedia, CC BY 2.0.

Rodingite

https://www.wikiwand.com/en/articles/Rodingite

Rodingite is a metasomatic rock composed of grossular-andradite garnet, calcic pyroxene, vesuvianite, epidote and scapolite. Rodingites are common where mafic rocks are in proximity to serpentinized ultramafic rocks. The mafic rocks are altered by high pH, Ca2+ and OH− fluids, which are a byproduct of the serpentinization process, and become rodingites. The mineral content of rodingites is highly variable, their high calcium, low silicon and environment of formation being their defining characteristic. Rodingites are common in ophiolites, serpentinite mélanges, ocean floor peridotites and eclogite massifs. Rodingite was first named from outcrops of the Dun Mountain Ophiolite Belt in the Roding River, Nelson, New Zealand.



Rodingite from Maryland

A rodingite dyke (white) in serpentinite (green) in the Dun Mountain Ophiolite Belt, New Zealand

Mineralogical composition

Rodingites consist of various Ca-Al silicates, including grossular, diopside, vesuvian, epidote and wollastonite. Chlorite, actinolite and tremolite are often found in quartz-poor source rocks, while alkali feldspar and xonotlite are formed in quartz-rich source rocks. In the fresh state, the rock is often light to medium gray, with high contents of diopside, actinolite or chlorite also green, with a predominance of alkali feldspar also pink.