Mountain building in western Crete north and south of Sfinari


View of the village, Sfinari (distant bay) looking South

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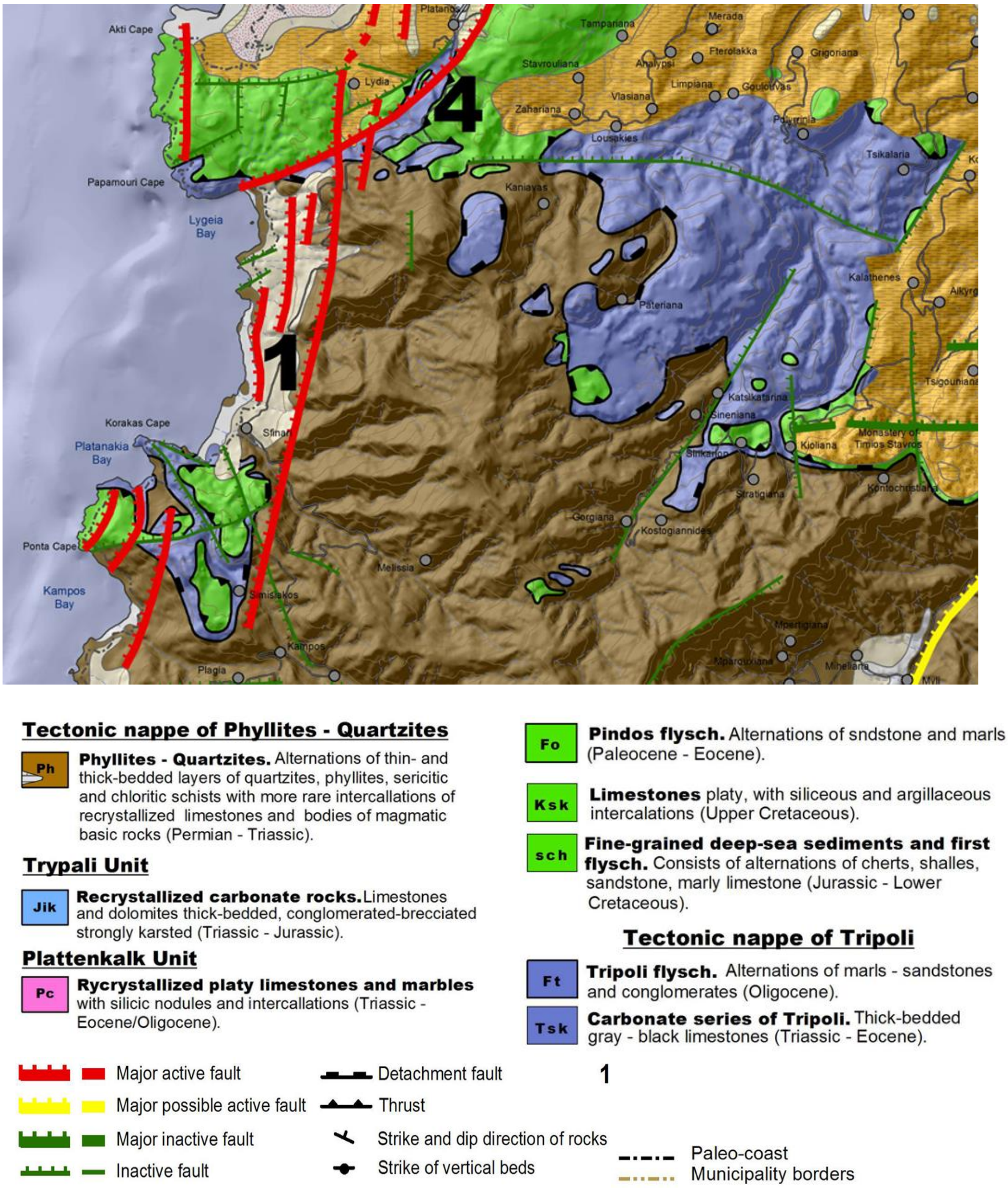
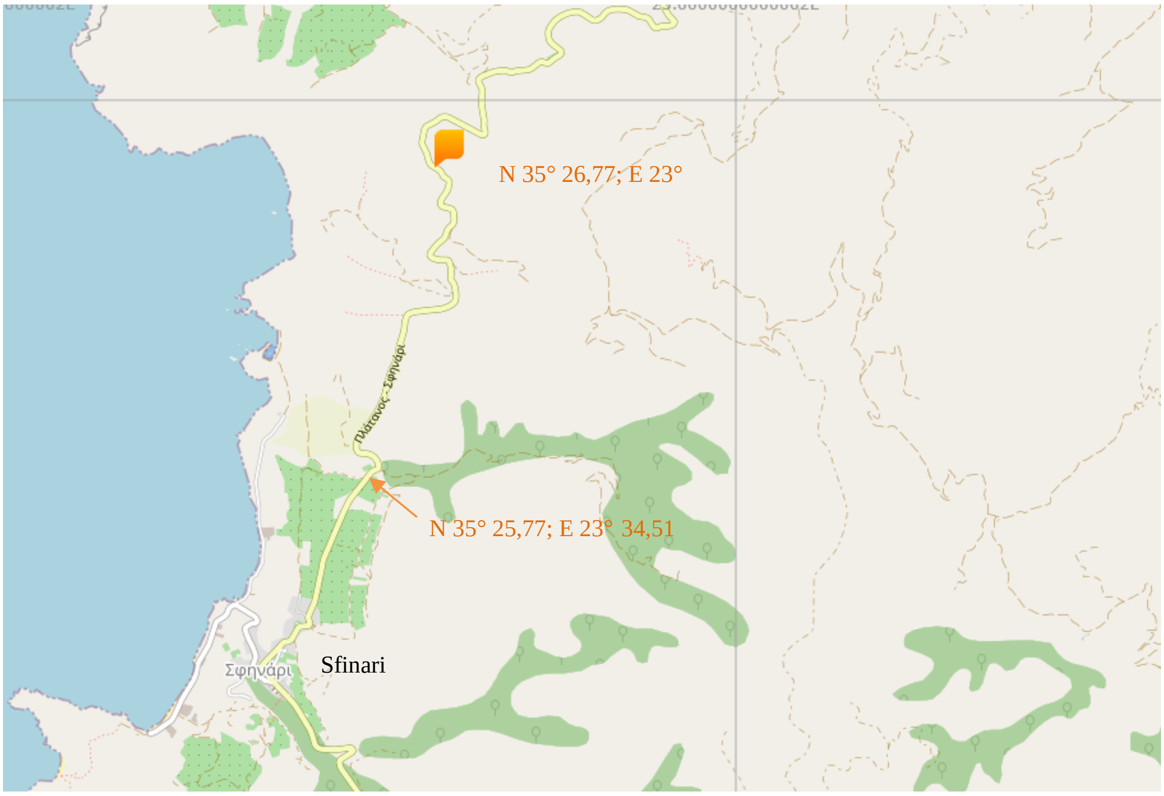
Berlin, March 2023

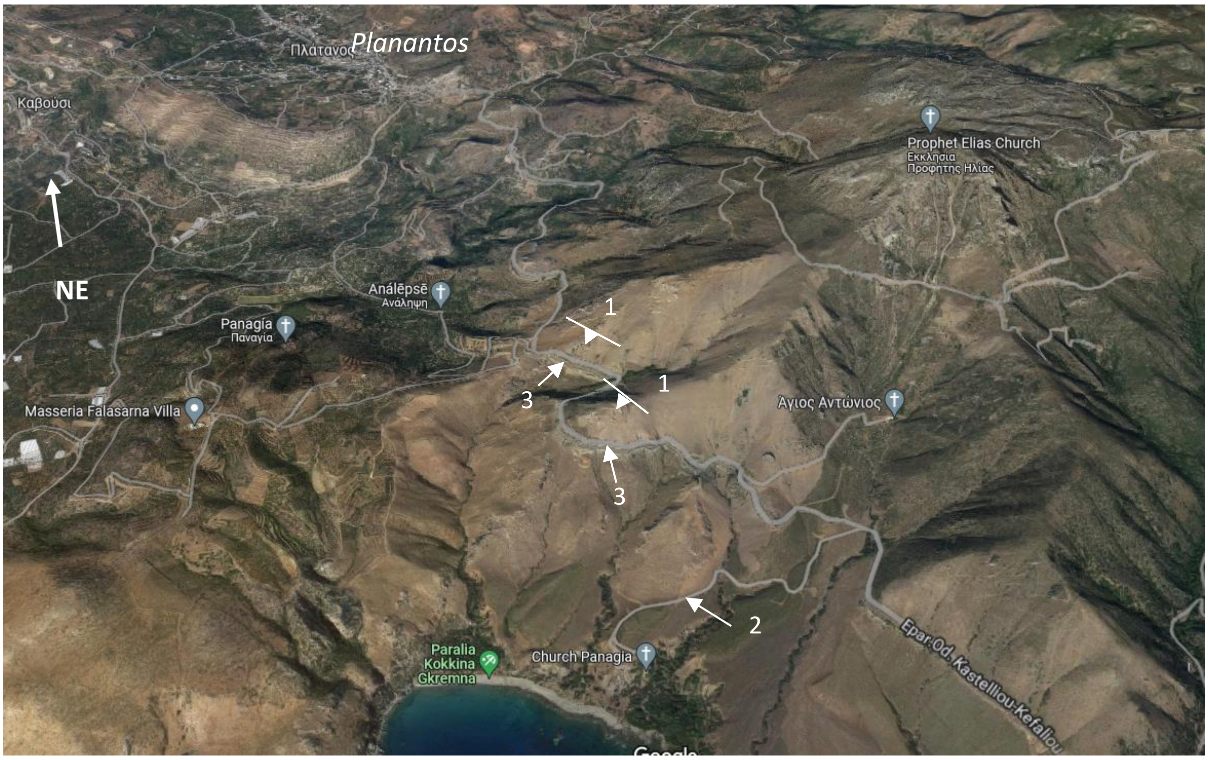
Geological map of part of the west coast of Crete, No.1



[D. Mountrakis et. al., 2012]

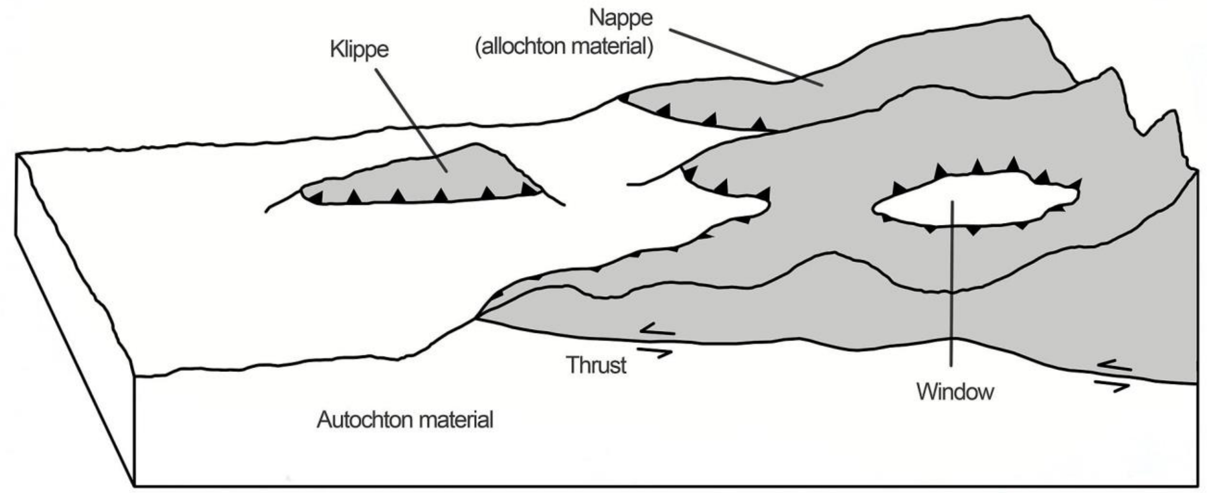


Location of Klippen and Phyllite-Quarzite outcrops



Area north of Sfinari and south of Planantos. 1: Miocene limestone nappe on Phyllite-Quarzite unit, 2: Tectonic breccia belonging to Miocene nappe, 3: Outcops of the Phyllite-Quarzite unit displaying predominantly brittle tectonics.

Micocene Klippe



Schematic overview of a thrust system. The shaded material is called a nappe. The erosional hole is called a window or fenster. The klippe is the isolated block of the nappe overlying autochthonous material. [Wikipedia]

A klippe is a geological feature of thrust fault terrains. The klippe is the remnant portion of a nappe after erosion has removed connecting portions of the nappe. This process results in an outlier of exotic, often nearly horizontally translated strata overlying autochthonous or para-autochthonous strata. In the case of Crete the para-autochton are the Quarzite-Phyllite and Plattenkalk units. These themselves are thought of as being nappes exhumed from the lower continental crust (i.e. 35km) as a result of subduction and subsequent buoyancy uplift. Hence the term para-autochton.

Both the Quarzite-Phyllite and Plattenkalk units have undergone high pressure/low temperature (HP/LT) metamorphism owing to their former subduction. Both units can be discerned from overlying nappes by their metamorphic features and metamorphic mineral content. The fault or thrust plane between the overlying nappes (i.e. Tripoliza and Pindos nappes) and the lower units (i.e. Quartzite-Phyllite and Plattenkalk nappes) is referred to as the Cretan detachment fault.

At this location north of Sfinari, the Phyllite-Quartzite Unit is overlain by a low-angle fault that separates the Phyllite-Quartzite unit from an unmetamorphosed Miocene limestone unit. The limestone is reported to contains conglomerate carbonate clasts. The low-angle fault may be interpreted as a high-level part of the Cretan detachment.

Miocene klippe (1, yellow) on the Phyllite-Quarzite unit (2). There also appears to be a major active normal fault (3) running across the Phyllite-Quarzite unit (see geological map). The normal fault would explain why the Miocene klippe is still present and preserved from erosion.

Miocene klippe on the Phyllite-Quarzite unit. Phyllite-Quarzite outcrops are along the roadside

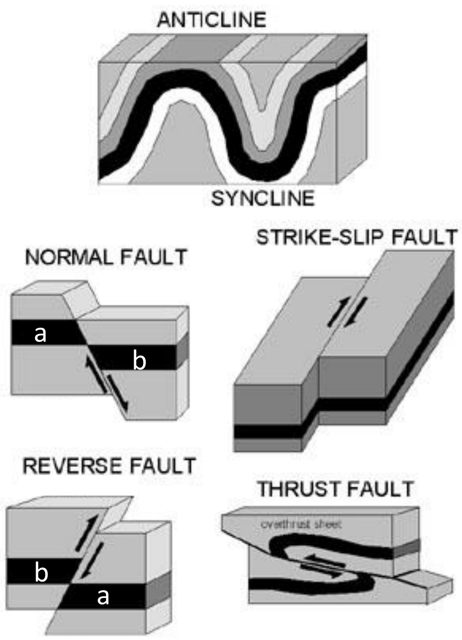
The conglomerate of the Miocene limestone klippe is matrix supported and consists mainly of rounded limestone cobbles, which look as if they may have been derived from the Tripolitza Group. A trail goes down from the main road, allowing a view of the fault zone beneath the klippe. There, in the lowermost part of the fault zone is a breccia that contains angular clasts of Phyllite-Quartzite material. Farther up in the fault zone, the Phyllite-Quartzite clasts disappear, and only limestone clasts are present.

Formation of the Phyllite-Quartzite Unit

The Phyllite-Quartzite Unit (PQ) on Crete comprises late Paleozoic to Triassic (Krahl et al., 1983) rock that was once marine siliciclastic, carbonaceous, and locally evaporitic sedimentary rock. The sedimentary rocks were deposited in shallow water on continental crust. The siliciclastic rocks occur mainly in the late Carboniferous to Permian stratigraphic units, whereas the Triassic strata consist predominantly of carbonate rocks. The microcontinent carrying this sedimentary pile underwent a prolongued diagenetic history and eventually encountered the Hellenic subduction zone in mid-Tertiary times approximately 36 to 32 Ma ago (Thomson et al., 1998, 1999).

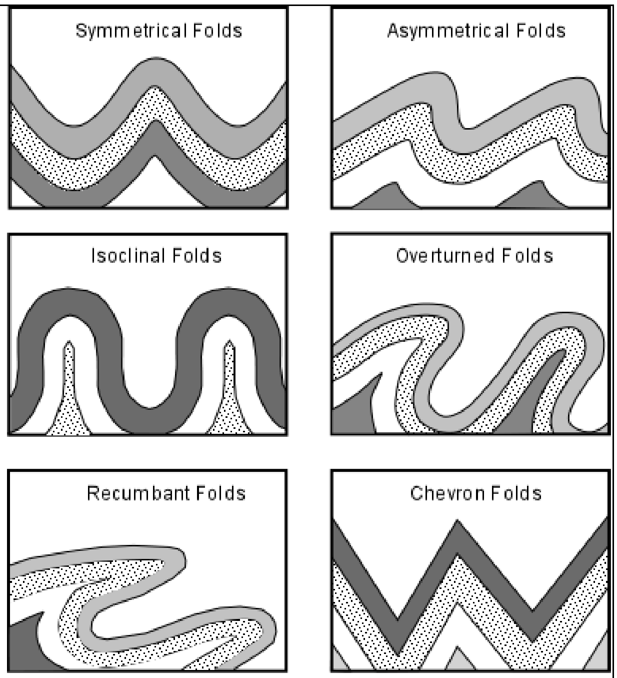
During the course of subduction, the stratigraphic units once overlying the mid-Triassic evaporites were detached and accreted at a shallow level; they are represented by the non-metamorphic rocks of the Tripolitza Unit (Bonneau. 1984; Thomson et al., 1999). In contrast, the PQ was carried down further by the subducted plate and was transformed by high pressure-low temperature (HP-LT) metamorphism. The siliciclastic rock consisting either of sandstone or shale was converted to quartzite and phyllite respectively, while the carbonate rock became marble. On its was way down the PQ became detached from its basement at a depth of about 35 km. Subsequently the PQ itself was underthrust by the more southerly part of the microcontinent, which was covered by the carbonate platform now forming the HP-LT metamorphic Plattenkalk Unit exposed beneath the PQ on Crete (B. Stockhert et al., 1998, Thomson et al., 1998, 1999).

Faults and Folds within the Phyllite-Quarzite Unit near Sfinari

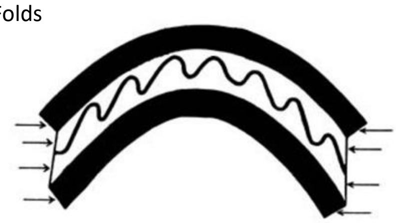


a: Foot wall, b: Hanging wall.

A Fault is a fracture where two blocks of rock are off-set. If the movement occurs rapidly it may be due to an earthquake. Fault blocks may also slip slowly, which is known creep.



Drag Folds



Drag folds occur within limbs of larger folds. Relatively thin incompetent (soft, weak, or compliant) layers that are usually embedded between thick competent beds form many smaller folds.

Numerous fresh cuts along the road provide an excellent view of both brittle and ductile structures in the Phyllite-Quartzite unit. Faults as opposed to folds are the result of brittle deformation near the crust's surface, where temperatures are low. Folds occur where pressures and temperatures are much higher; deeper with the earth's crust. Mesoscale isoclinal folds (i.e. folds that are tens of meters large) are locally present at several locations. However, these are not easily recognized, due to their size. Mostly only the limbs of the folds are visible. Some authors describe there being two generations of folds situated at right angles at this location (B. Stockhert et al., 1998).

Overview of one of several Phyllite-Quartzite out-crops. 1: possible thrust plane within the Phyllite-Quartzite.

A small isoclinal fold at this location. 1: Quartzite, 2: Phyllite. Isoclinal folds have acute angles between the limbs and are the result of extreme deformation (i.e. strain). The highly viscose quartzite layer was folded at approx. 400 °C at 30-35 km depth within the earth's crust during subduction.

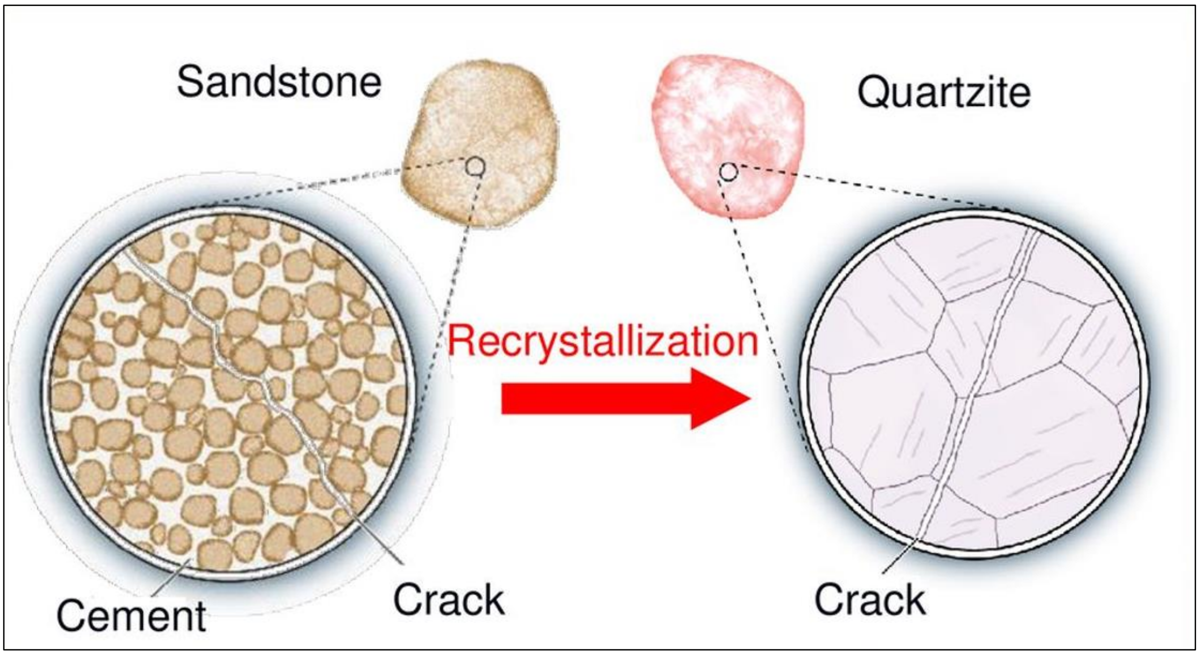
1: Quarzite, 2: Phyllite, 3: Former joints filled with quartz and later deformed when the quartzite beds were folded (hammer as scale).

Closeup of quartzite bed, showing fresh surface of hand sample. Quartzite is typically very hard. 2: Scratched with hardened steel nail you will see a minute trail of metal.

Closeup of quartzite bed (1), displaying metamorphic muscovite (2). The glistening on the hand sample is caused by white mica grains. Loose piece of Phyllite (3). Radioisotope dating (K/Ar and Ar/Ar) from syn-metamorphic white micas suggest that metamorphism occurred between 21 and 24 Ma (Seidel et al., 1982; Jolivet et al., 1996).

Quartzite

Quartzite is a very hard rock composed predominantly of an interlocking mosaic of quartz crystals. The grainy, sandpaper-like surface is glassy in appearance. Minor amounts of former cementing materials, iron oxide, silica, carbonate and clay, often migrate during recrystallization, causing streaks and lenses to form within the quartzite. Quartzite is commonly regarded as metamorphic in origin. When sandstone is subjected to the great heat and pressure associated with regional metamorphism, the individual quartz grains recrystallize along with the former cementing material. The recrystallized quartz grains are roughly equal in size. The grains are so tightly interlocked that when the rock is broken, it fractures through the grains to form an irregular or conchoidal fracture. [Wikiwand]

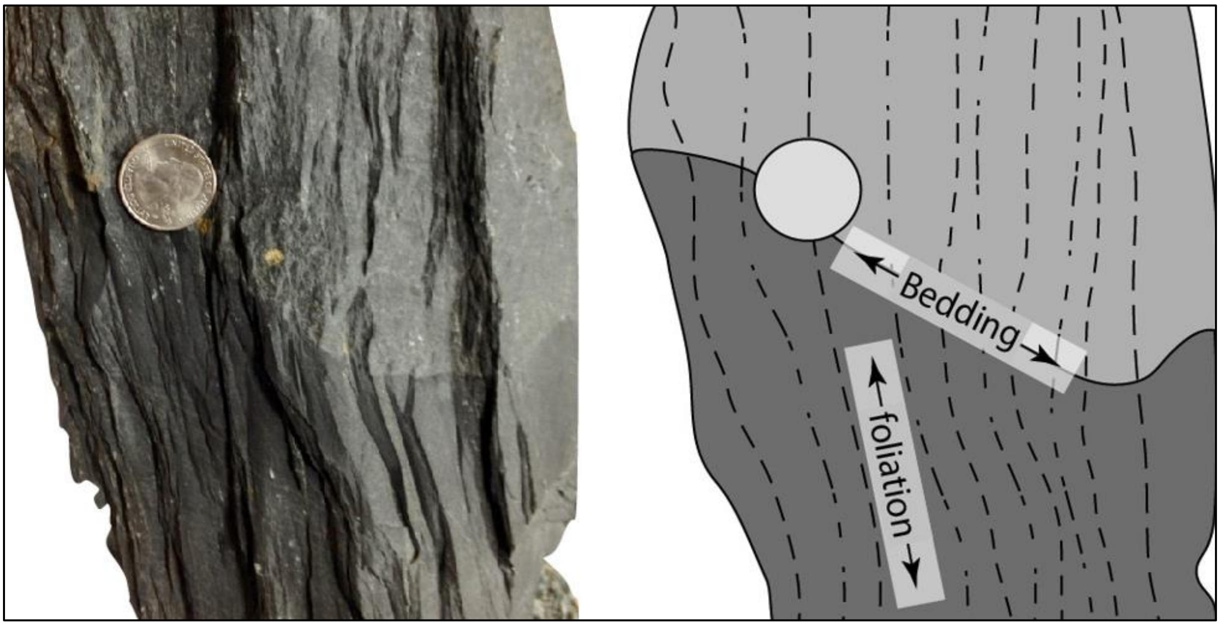


Recrystallization of sandstone to quartzite and reduction of surface area [Pierce Day, Metamorphic Structures]

Phyllite

Closeup of a phyllite bed. 1: Foliation (cleavage)

Phyllite is a foliated metamorphic rock in which platy minerals occurring for example in clay have grown larger and the surface of the foliation shows a sheen from light reflecting from the grains, perhaps even a wavy appearance, called crenulations.



Foliation vs. bedding. Foliation is caused by metamorphism. Bedding is a result of sedimentary processes. (Source: Peter Davis)

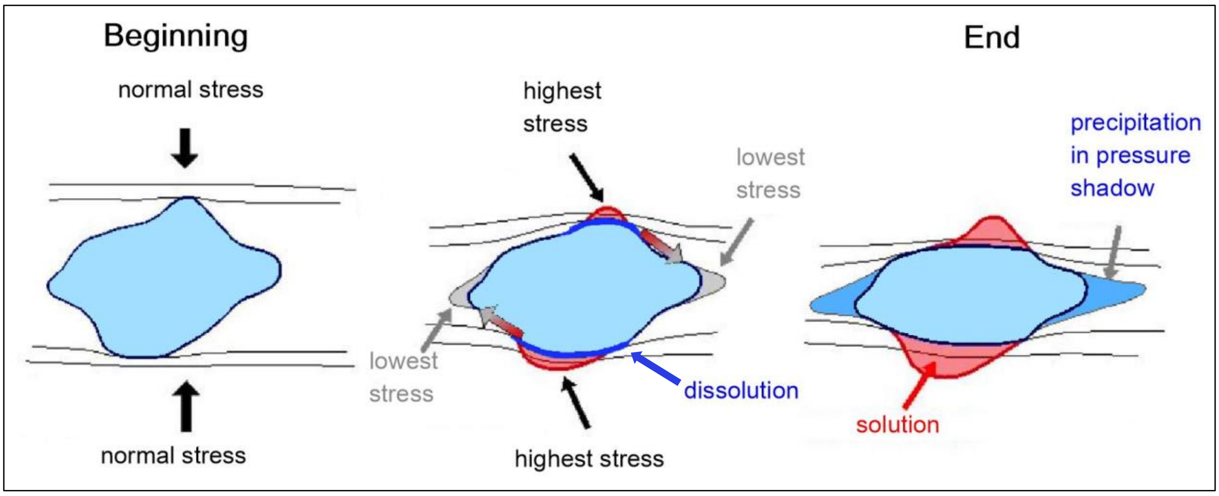
Boudinage structures (1) indicate extreme strain (i.e. deformation) perpendicular to the bedding plane and are a sign of pressure solution as a mode of material transport. In the lower half of the picture are smaller isoclinal folds indicated by folded quartzite beds (2). Phyllite (3).

Boudins appear predominantly within thin quartzite veins and layers within thicker phyllite layers and are attributed to pressure solution as a mode of material transport under elevated pressure and temperature. Boudinage represents shortening perpendicular to the bedding plane, whereas folding takes place due to shortening parallel to the bedding plane. Both mechanisms can work in combination when the bedding is rotated within the stress field – after folding the limbs lie perpendicular to the direction of stress enabling boudins to be formed.

Metamorphic Processes within the Phyllite of the Phyllite-Quartzite unit

In metamorphic rocks, quartz is often subject to pressure solution (also called dissolution and precipitation creep). This is a process where it is dissolved at crystal grain boundaries at elevated p-T conditions and recrystallizes in areas of less stress i.e. in the pressure shadow of larger crystal grains. Pressure solution is an important deformation mechanism that contributes to the formation of phyllites and schists in metamorphic rocks. Prerequisites for pressure solution are the presence of a fluid between the grains and normal stress. Combined with normal stress pressure dissolution changes the shape of the grains and can even lead to the complete dissolution of crystals. Different minerals respond differently to this mechanism depending on their pressure solubility.

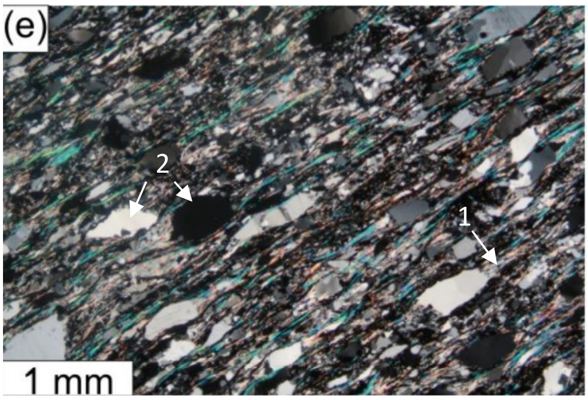
In the case of quartz dynamic stress causes larger grains to recrystallize into energetically more favorable small grains, which then have different crystallographic orientations. Less pressure-soluble minerals such as mica are selectively enriched as more pressure-soluble minerals such as quartz and calcite are dissolved and transported resulting in a metamorphic foliated structure also known as cleavage [Wikipedia].

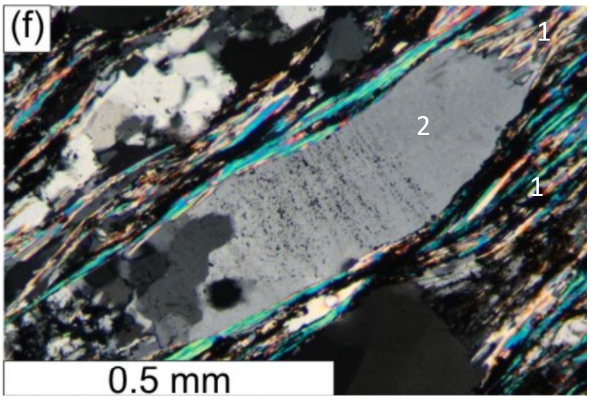


Process of dissolution precipitation creep [slightly modified from PETROgraph]

Flattening strains involving non-constant volume, or mass-loss processes (i.e. anisochoric processes) have been measured at several location in western Crete within the Phyllite-Quartzite unit.

Not all dissolved material has to be deposited in the pressure shadow of an adjacent large crystal. Some of it can also be transported away in the fluid along joints and fissures and precipitated elsewhere causing a loss of mass.



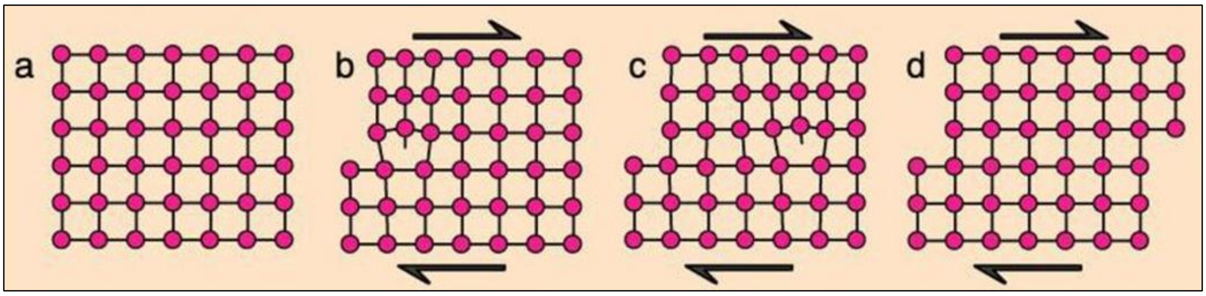


Microscopic structure of the phyllite rock: Clastic quartz grains (2) truncated by dissolution along interfaces with mica (1). Crossed polarizers. Phyllite-Quartzite Unit near Sfinari, Western Crete. [Sara Wassmann 2012]

Metamorphic Processes within the Quartzite of the Phyllite-Quartzite unit

In contrast to the phyllites, the microstructures of some the quartzites, that are particularly pure (< 3% mica), indicate a different mode of material transport known as **dislocation creep**. This is a process of internal crystal shape deformation associated with preferred orientation of the crystal lattice (i.e. mutual orientation of the quartz crystal's c-axis). The thick quartzite beds being more competent than the phyllite beds and therefore not so easily deformed must have been subjected to a higher stress field (B. Stockhert et al., 1998). Dislocation creep occurs under higher grade metamorphic conditions deep within the earths crust and is an important element of deformation at subduction zones.

In the quartzites, the clastic quartz grains show undulose extinction or low angle subgrain boundaries parallel to the c-axis. These microstructural features are developed uniformly throughout the pure quartzite layers, suggesting that the internal microstructures of the clastic grains were not imported from their source rocks, but largely developed during deformation of the quartzite. The clastic grains are embedded in a matrix of recrystallized grains (Fig. 6A, B), with a variable grain size of about 50 ± 20 µm. The grain shapes are irregular and the high-angle grain boundaries are serrated [B. Stockhert et al., 1998].



Principle of dislocation creep as a form of deformation during metamorphose.

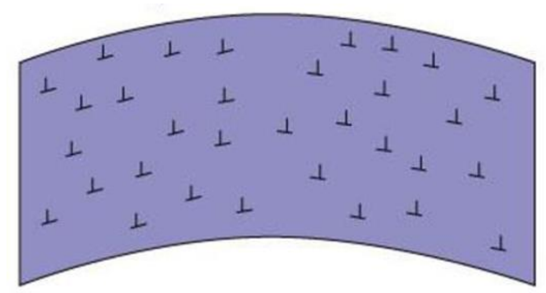
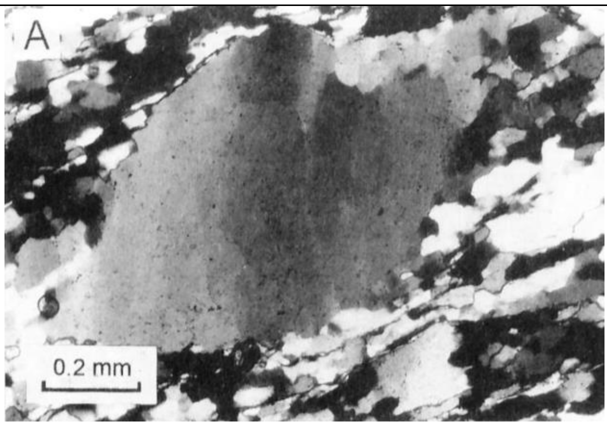
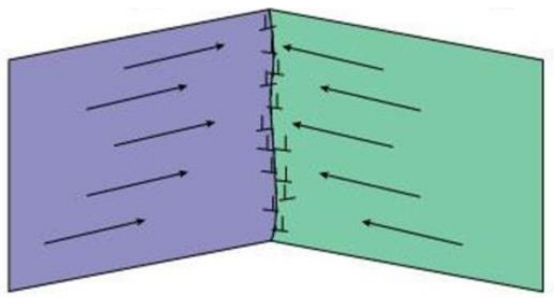


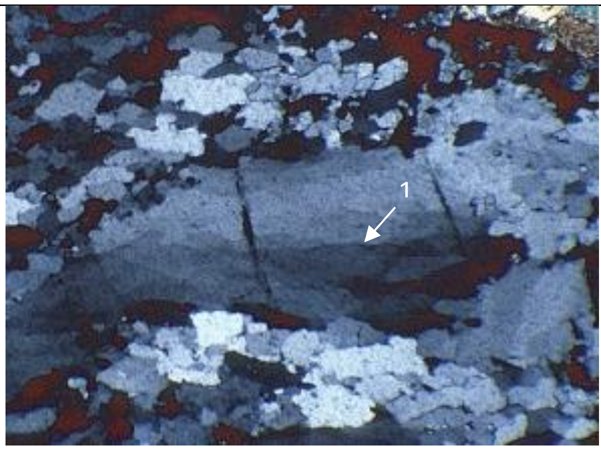
Illustration of a strained (deformed) crystal grain impaired by defects and dislocations resulting in undulose extinction [Pierce Day, Metamorphic textures].



Microstructures of clastic quartz grains in pure quartzites. The clastic grains display undulose extinction [B. Stockhert et al., 1998]



Recovery by dislocation creep produces two strain-free subgrains within the crystal. The subgrain boundaries form planar arrays separating relatively strain-free domains with a small angular mismatch.
[Pierce Day, Metamorphic textures]



Recovery and recrystallisation in quartz The large, strongly-strained quartz grain in the centre shows undulose extinction. Rather than varying smoothly across the grain, however, the extinction reveals a patchwork of subgrains (1). Above and below, quartz has recrystallised into new, smaller grains. Crossed polars. Field of view 1.75 mm. (Sample of Moine Meta-sandstone, Scotland).

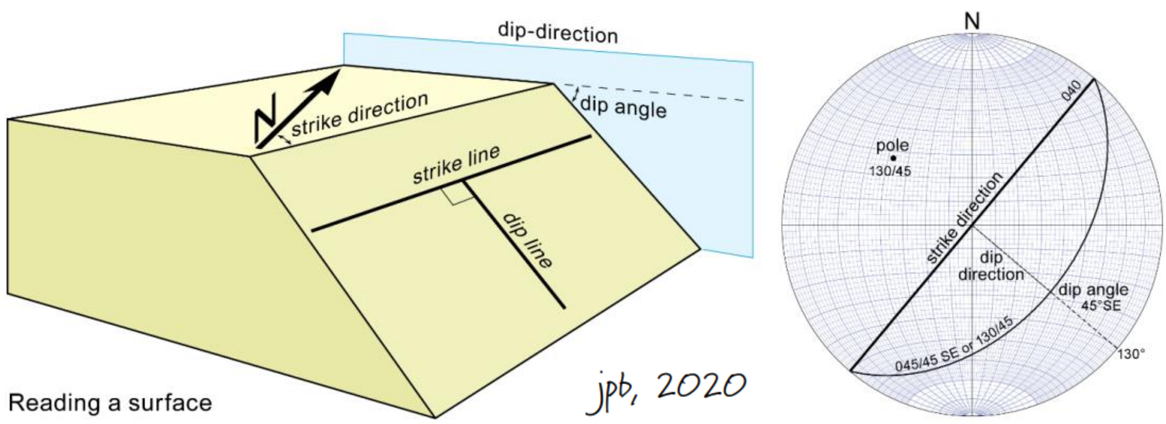
Brittle deformation and extension

Throughout Crete, the Phyllite-Quartzite nappe is commonly cut by numerous faults, which is interpreted as being the result of brittle extension of the earth’s crust during exhumation of the lower nappes (i.e. parautochton). The brittle extension followed subsequent to folding and metamorphism during a sperate tectonic phase, which is associated with the buoyant elevation of the Phyllite-Quartzite and Plattenkalk nappes. This relationship is well displayed along the West Coast road. These fault zones generally lie at a high-angle to the more gently dipping pressure solution cleavage in the Phyllite-Quartzite. Typically, they have gouge zones that are several cm to several m thick.

Overview of a Phyllite-Quartzite out-crop, displaying an array of faults. 1: Gauge zone, 2: Quartzite, 3: Phyllite

Outcrop next to the one shown above, displaying more faults. 1: Gauge zone, 2 Quartzite, 3: Phyllite

Closeup of previous photo. 1: Gauge zone, 3: Phyllite.



Cataclastites

There are a range of metamorphic rocks made along faults. Near the surface, rocks are involved in repeated brittle faulting that produces a material called *rock flour*. At lower depths, faulting creates cataclastites, which are chaotically-crushed mixes of rock material with little internal texture. At depths below cataclasites, where strain becomes ductile, mylonites are formed.

Gouge zone at a fault within the Phyllite-Quartzite unit, displaying cataclasts produced mainly by brittle strain.

Closeup of the above photo showing a chaotic mixture of loose small sharp-edged grains in a powdery matrix.

Phyllite Quartzite Rocks near Sfinari

At the village of Sfinari (N 35° 25.77; E 23° 34.51) there is series of rocks called the Sfinari Phyllite. They are the oldest in western Crete and based on conodont stratigraphy have been placed in Upper Carboniferous (Namurian/Westfalian boundary and middle Stephanian). They are dark to green chloritoid-bearing phyllites with thin marble beds [Dornsiepen & Manutsoglu 1994].

Elastic deformation vs. Plastic deformation

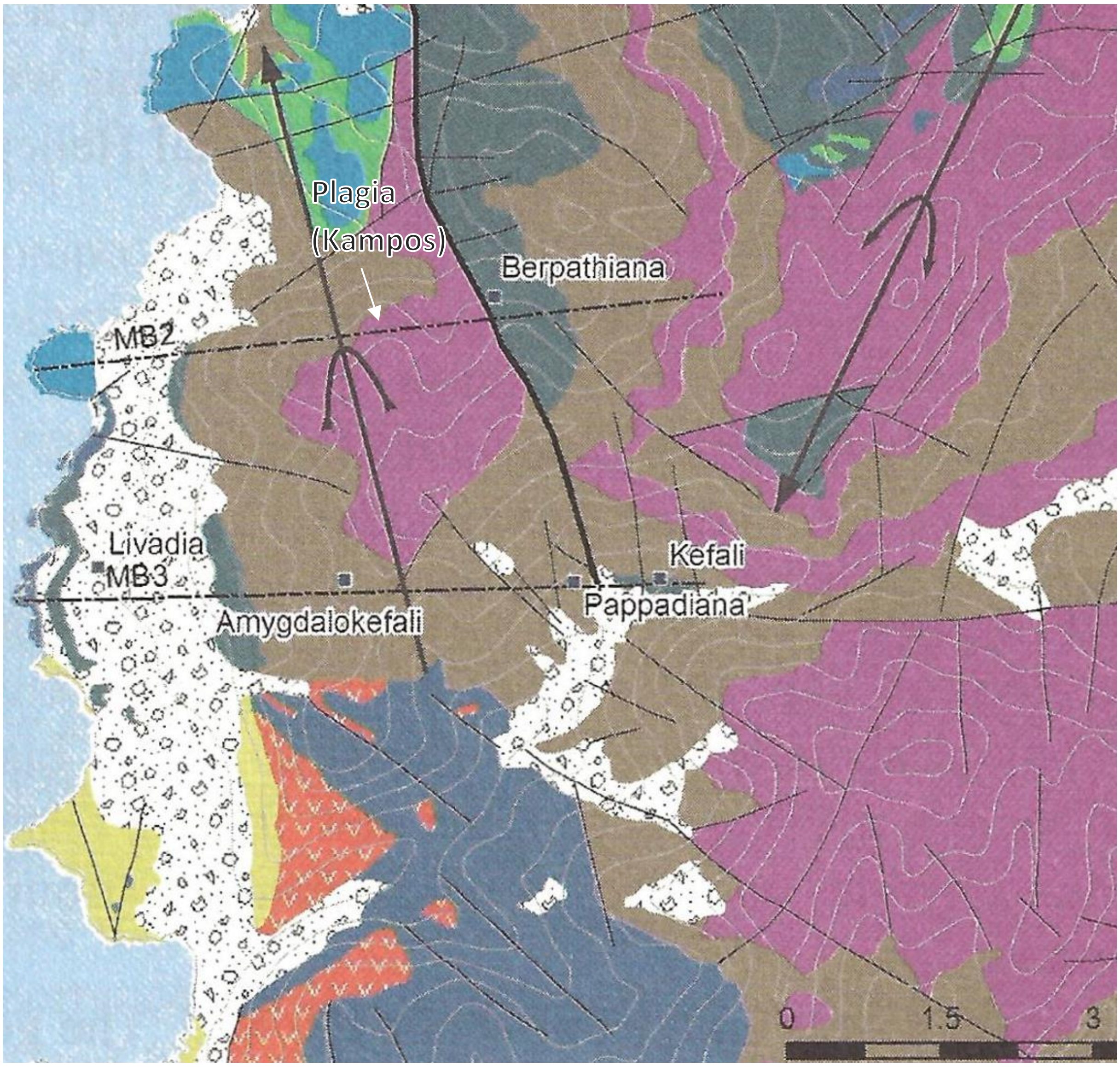
Low stress (i.e. forces) in a rock causes elastic deformation that is not permanent and therefore is not recorded in the rock record.

Elastic deformation is the dominant form of deformation at shallow depths within the lithospheric crust because both the temperature and pressure are low. At shallow depth the crust is brittle and when the stress is large enough, failure by fracture or frictional sliding occurs. Viscous and plastic deformation typically arises deeper within the crust at higher pressures and temperatures. The folding in the Phyllite-Quartzite Unit is a good example of plastic deformation. In between the shallow elastic region and deeper viscous region there is a region that deforms through both mechanisms (both mechanisms accommodate roughly equal amounts of deformation). This region is referred to as the **brittle-ductile transition** and rocks here include cataclasites (features of viscous flow with pervasive micro-cracks), mylonites and pdeudotachylites (melting).

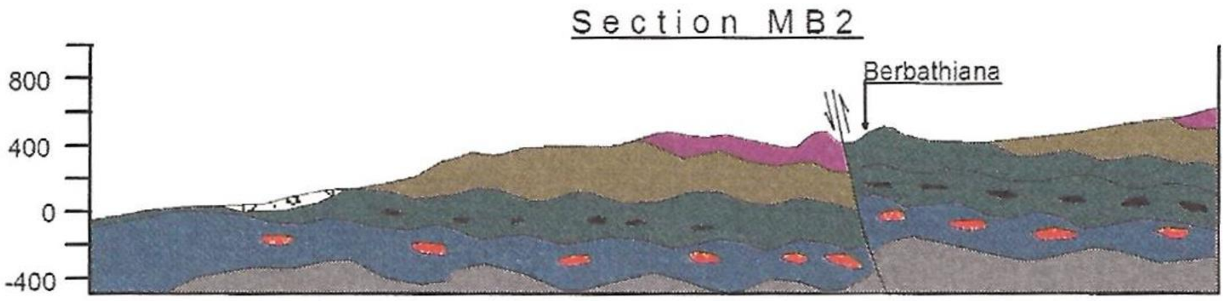
Even though elastic deformation is not preserved in the rock record, the elastic strength of the rock still supports the stress in the rock up until failure. For example, slow steady motion of tectonic plates causes elastic deformation in the region surrounding a fault. In between earthquakes this elastic deformation builds up until the stress inside the rock exceeds the friction on the fault. So, even though the elastic deformation is released in the surrounding rock, the history of this process is recorded by the earthquake [Billen M., 2021, LibreTexts, Geophysics].

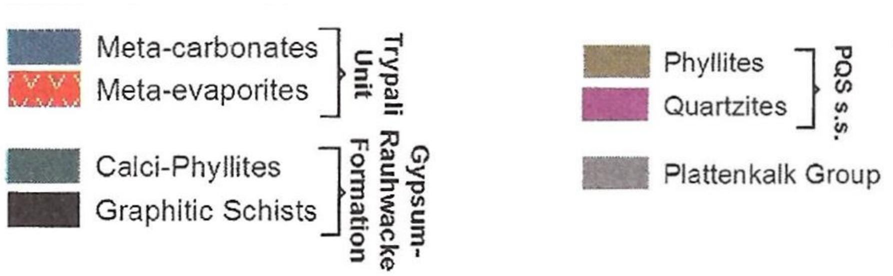
While we normally think of rocks as being hard, elastic materials that break, the rheology of rocks is in fact both elastic and viscous, but these two different responses occur at different time-scales. The short-term deformation of rocks is best described as elastic with brittle failure and applies to processes like earthquakes (timescale of 1-1000 years between events). This short term response dominates at cold temperatures and low pressures typically found in the crust. The long term deformation of the lithosphere and the mantle is best described as viscous flow. At the higher pressures and temperatures of the deep lithosphere and mantle, deformation of grains by viscous mechanisms occurs more readily than elastic or brittle deformation. However, it is a very slow process with mantle flow occurring at velocities of 1 to 20 cm/yr, at time-scales of a million years or more. It’s important to remember that at high pressure, the rocks are flowing but they are still solid. High temperatures (around 1200-1400 °C) and relatively lower pressure are needed to cause the rock to melt and form magma.

Geological map of part of the west coast of Crete

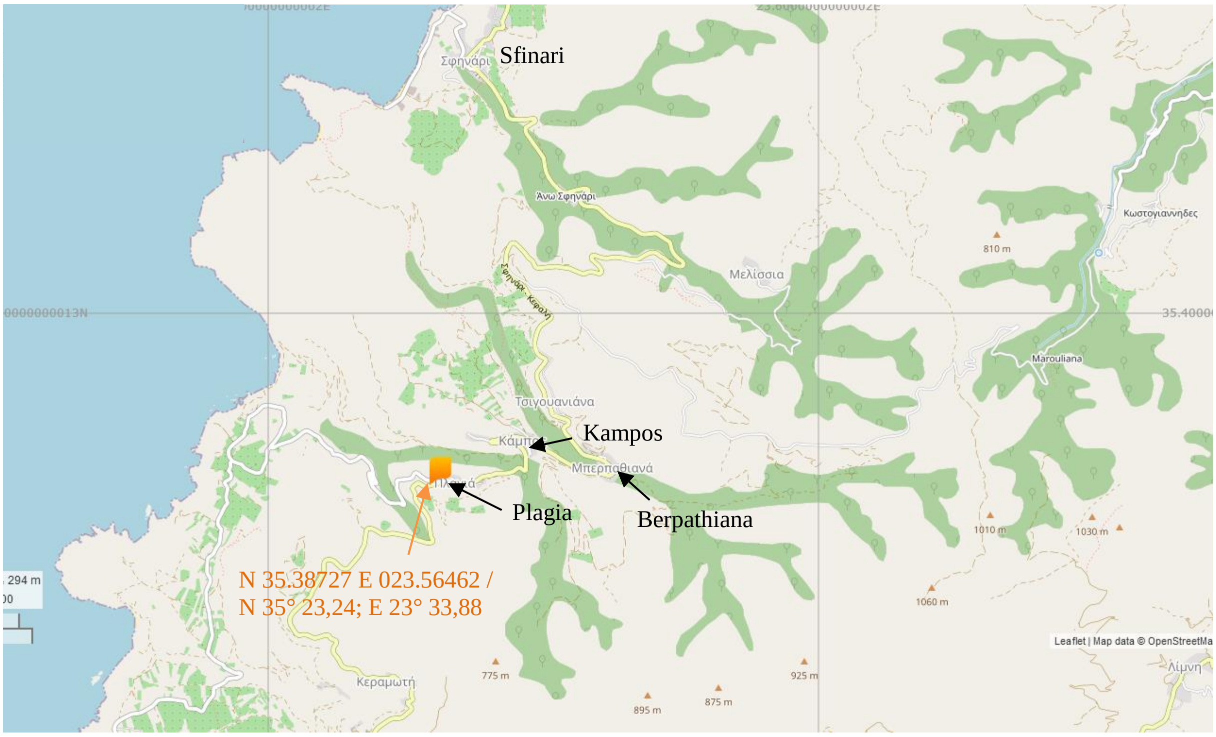


Geological map of part of the west coast of Crete, developed for hydrogeological research. [Manutsoglu E, et. al. , 2011]

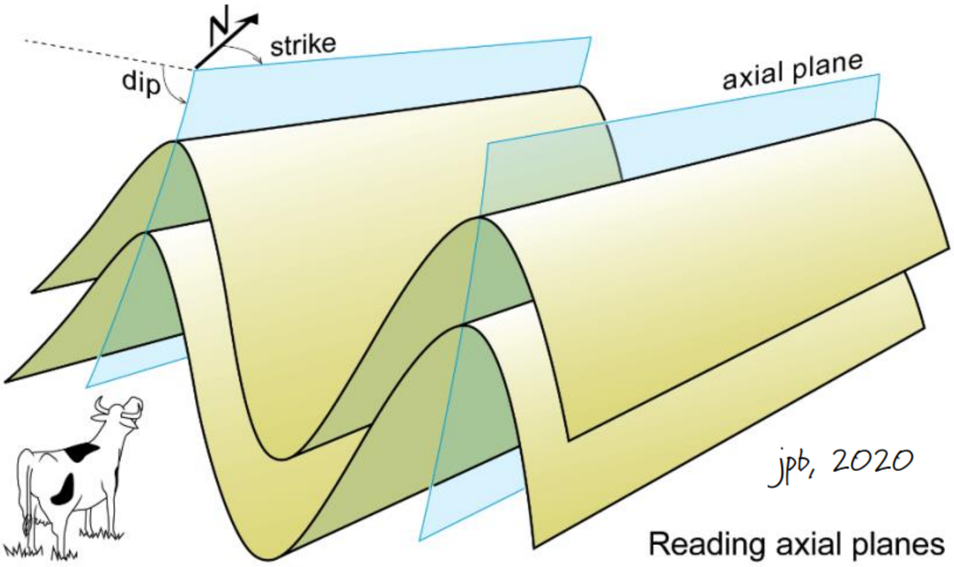




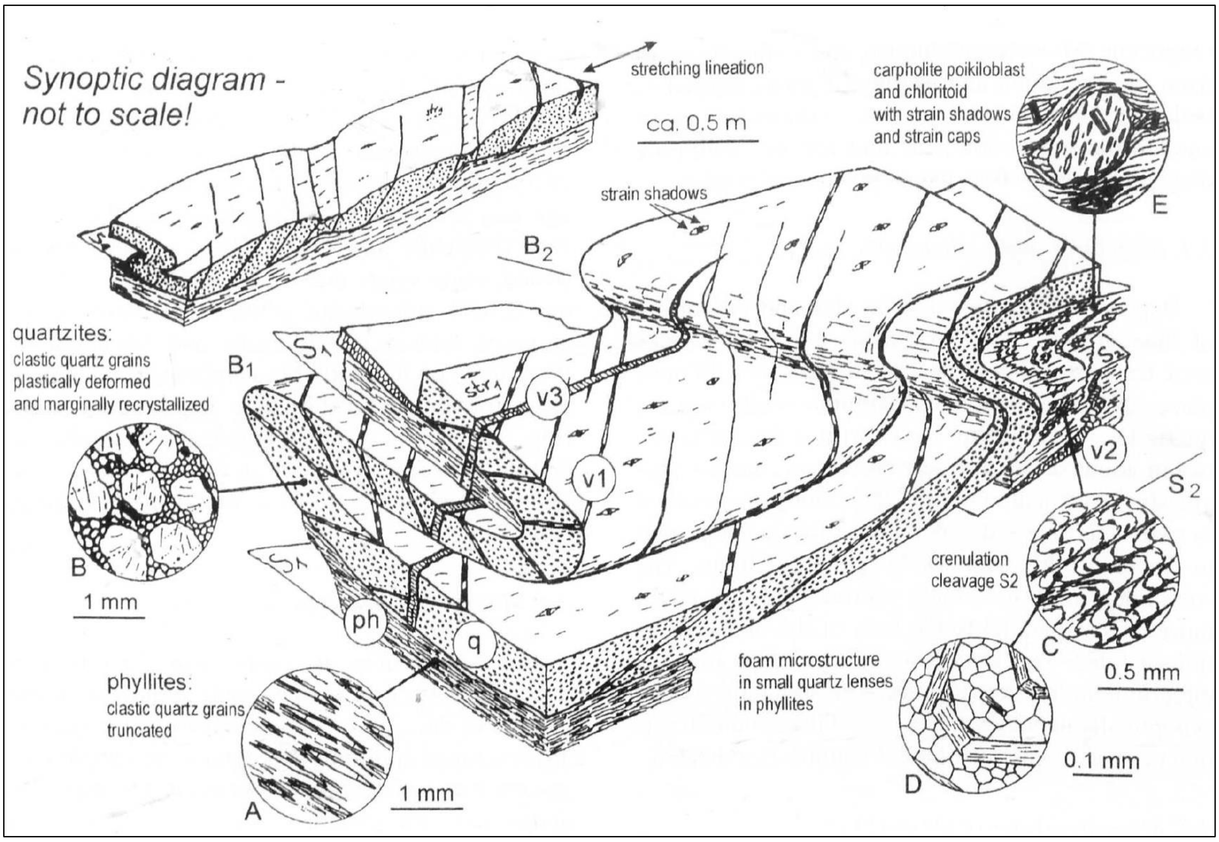
Two generations of folds in the Phyllite-Quartzite Unit near Plagia (Kampos)



Two sets of mesoscale folds are locally found in the Phyllite-Quartzite nappe near Plagia (Stöckhert et al., 1999; Thomson et al., 1999). The older set consists of recumbent isoclinal folds, with fold axes trending NNE-SSW. These folds are associated with a nearly horizontal cleavage and a NNE stretching lineation. The second set of folds have axes that trend approximately E-W. The limbs of these folds are commonly thinned, and more competent quartz-rich horizons have boudinage structures. These are attributed to the deformation associated with Oligocene accretion of the Cretan nappes. At this location the road provides a good opportunity to look at these folds in different orientations due to the bend in the road.



Location of outcrops at the site near Plagia. I: Fold axes trending NNE-SSW (B_1) with nearly horizontal cleavage (S_1) and NNE stretching lineation (str_1), II and III: Second set of folds (B_2) with E-W trending axes.

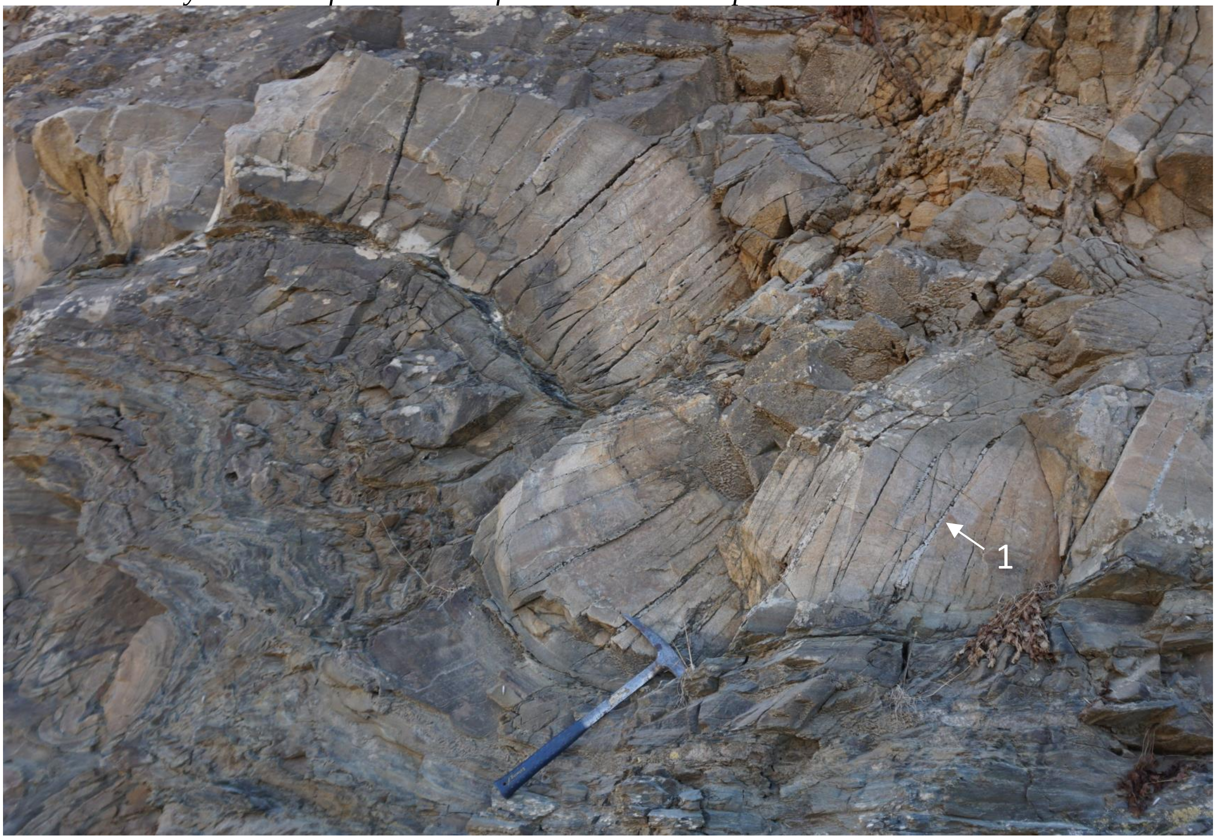


Synoptic diagram illustrating deformation structures within the siliciclastic rocks of the Phyllite-Quartzite nappe of western Crete, from B. Stockhert et al. B1, S1, str1 denote first-generation fold axis. schistosity and stretching lineation. B2, S2 denote second generation fold axis and schistosity (crenulation cleavage in phyllites). Ph: phyllites: q: quartzites: v1: veins formed along original joints normal to bedding in sandstone: during deformation these veins were rotated and stretched, with boudinage of the original carbonate filling (black): quartz (white) was precipitated between the boudins: v2: veins formed during HP-LT metamorphism along S1 in phyllites, folded (B2) during progressive deformation and recrystallized; v3: later formed veins crosscutting all earlier structures, undeformed and with original microstructure preserved, Note that the orientation of boudin necks and veins in quartzites is largely controlled by pre-existing discontinuities (joints) and not normal to the stretching lineation.

Insets A to E indicate characteristic microstructures and their respective position. (A) Clastic quartz grains in phyllites, flattened by pressure solution. (B) Clastic quartz grains in quartzite, slightly deformed by dislocation creep (subgrains, recrystallized grains, weak lattice-preferred orientation) and formation of microcracks (transgranular fluid inclusion trails normal to stretching lineation). (C) Crenulation cleavage related to second generation folds in phyllites, developed by pressure solution. (D) Close up of (C): micas form polygonal arcs around hinges of second generation folds and small quartz lenses between mica layer exhibit an interracial free energy controlled microstructure. (E) Carpholite poikiloblast that has overgrown clastic quartz grains flattened by pressure solution and earlier crystallized chloritoid. Deformation by pressure solution continued after growth of the minerals of the HP-LT paragenesis. as revealed by strain caps and strain shadows at the rigid carpholite and chloritoid blasts [B. Stockhert et al., 1998].

Outcrop I

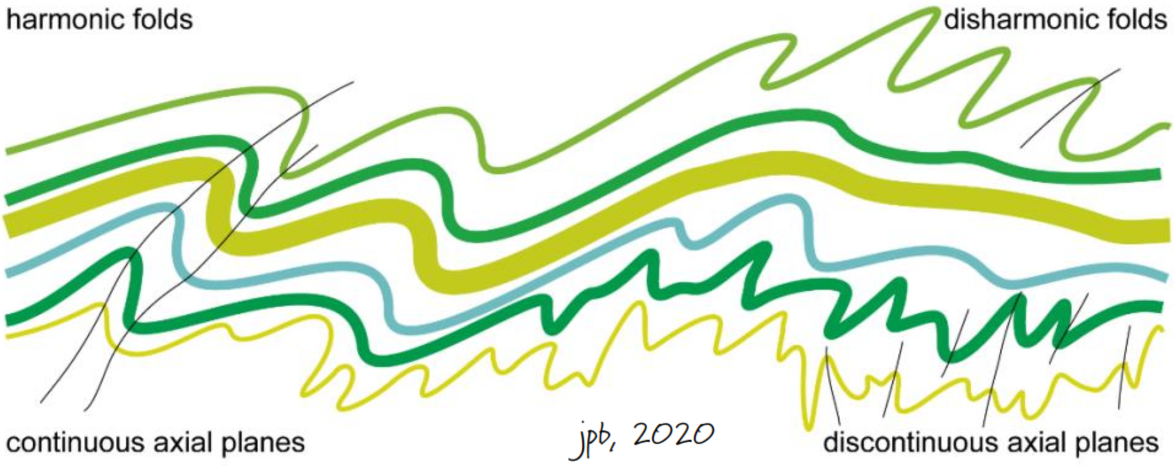
Overview of part of outcrop I. Recumbant isolclinal folds displaying predominantly Quartzite and shale beds. Only one limb of the isoclinal fold is visible in the picture.

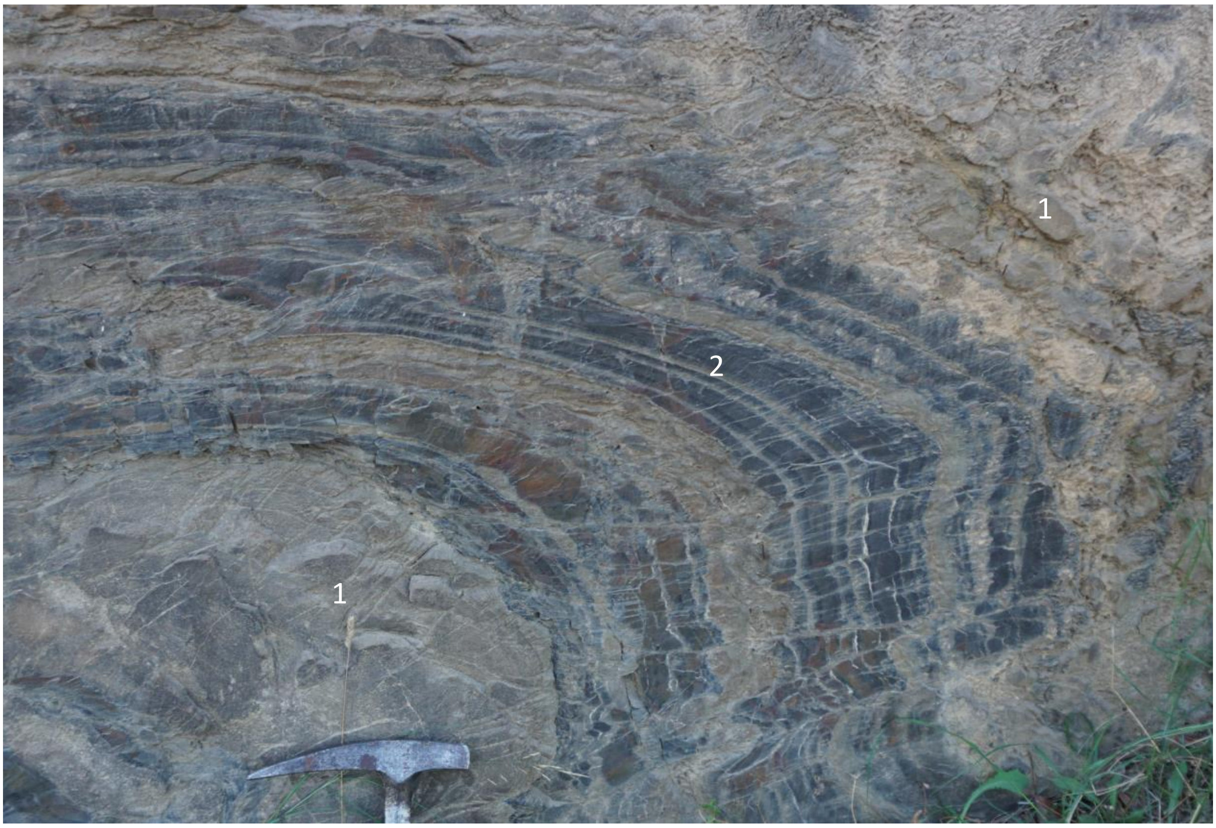


Outcrop I,1: Joints in quartzite that were originally at right angles to the bedding (v1). During folding the joints and their fillings of calcite and later quartz were tilted according to the curvature of the folds.

1: Stylolites parallel to bedding within a quartzite bed. 2: Joint filled either with calcite or quartz (v1). Stylolites indicate that material has been removed by pressure dissolution, in a deformation process that decreases the total volume of rock. Minerals which are insoluble in water, such as clays, pyrite and oxides remain within the stylolites and make them visible. Stylolites occur mainly in homogeneous rocks such as carbonates, cherts, and sandstones and may be the result either of diagenesis or tectonic stress. The amplitude of the stylolites is formed parallel to the main direction of stress (σ1)

Disharmonic folds within shale between quartzite beds.





Disharmonic folds within shale between quartzite beds. Some of the thinner lighter layers may be of limestone. 1: quartzite, 2: shale


Overview of part of outcrop I, displaying isoclinal folding of quartzite and shale beds.

Closeup of Quartzite sample.

Thin shale beds between quartzite beds. 1: Quartzite, 2: There appears to be two cleavage planes within the shale.

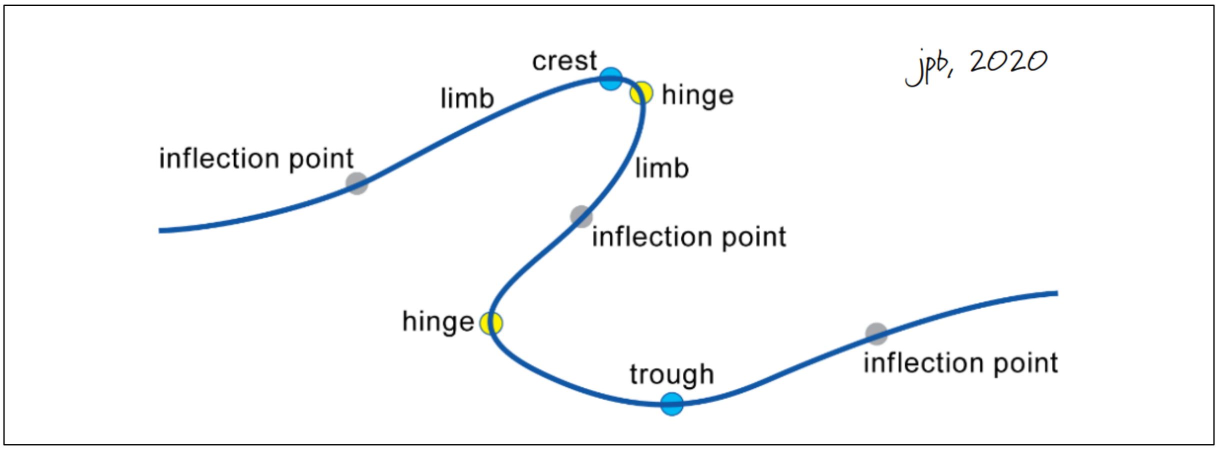
Outcrop II

Overview of outcrop II. Thick quartzite beds at the top and mainly thin shale and quartzite beds lower down.

The combination of two fold generations has created a complicated not easily determined pattern of hinges and limbs.

Outcrop III

Overview of outcrop III showing phyllite and quartzite beds. In the centre of the picture some of the layers are in a vertical position indicating the hinges of the first-generation folds. Second generation folds are hard to define at this spot.



Basic terminology of folds

Vertical phyllite and quartzite beds shown in the picture above. The cleavage of the phyllite at this spot indicates that this may be the hinge of the fold.

Thin phyllite and quartzite beds with possible third generation vein. 1: Phyllite with cleavage, 2: Quartzite, 3: possible v3 vein cross cutting quartzite and phyllite layers folding

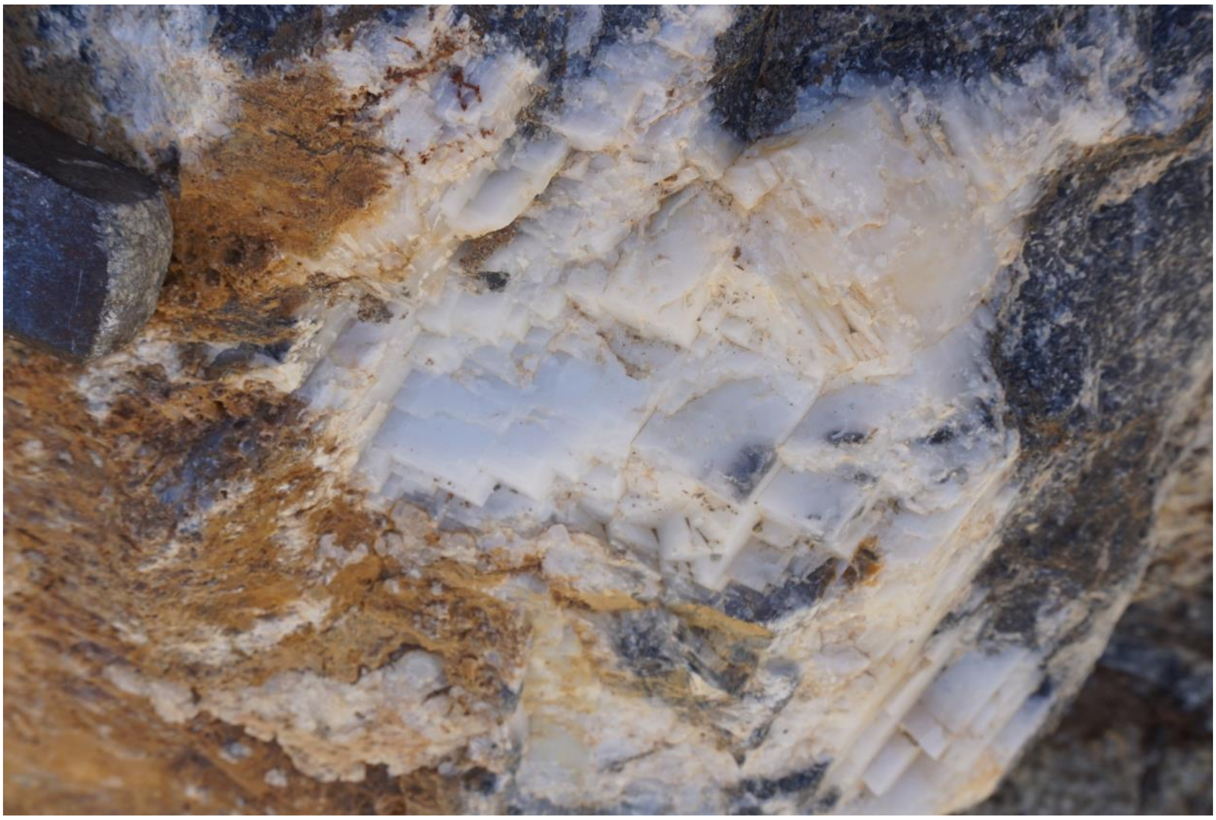

Phyllite hand sample showing two generations of cleavage (S_1 and S_2).

Overview displaying another part of outcrop III and second-generation folds. Quartzite and phyllite beds similar to outcrop I, but some of the quartzite beds are thinned out and have boudinage structures (1).

Outcrop IV

Large unfolded block of dark marble. It has straight veins that were probably once joints and numerous irregular shaped veins.

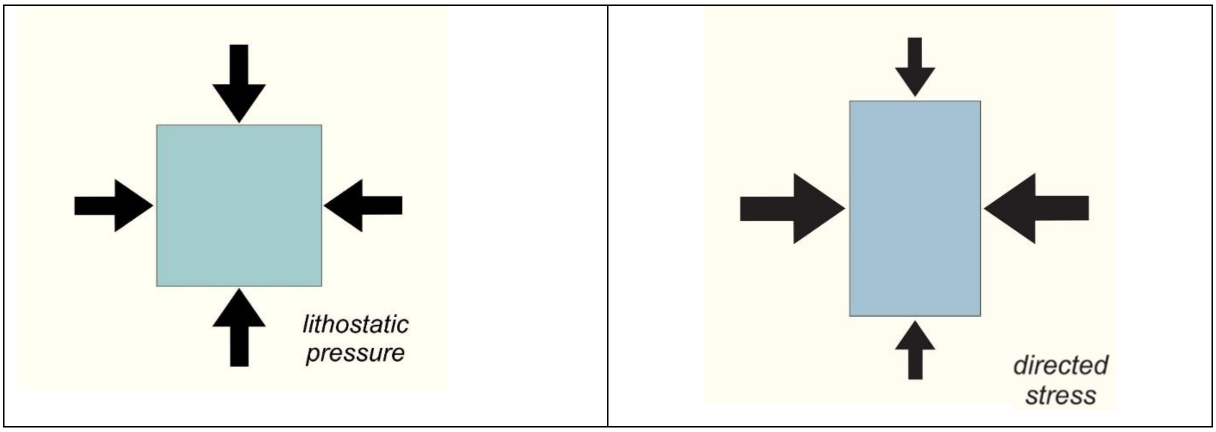
1: Irregularly shaped veins that could have once been fossil shells destroyed by recrystallisation. 2: Piece of siliclastic rock ?



Calcite vein filling displaying characteristic rhombic cleavage.

Confining pressure vs. directed pressure

At higher temperatures, and under higher confining pressures, rocks are more likely to undergo ductile deformation, when subjected to directed stress (e.g. due to tectonic movement during subduction). Confining pressure is the stress that a material experiences uniformly from all sides as a result of the weight of material above and around it. Within Earth, the confining pressure is due to the weight of overlying rocks. Confining pressure due to the overweight of rocks is called lithostatic pressure. Directed stress, sometimes called differential pressure, is also a force applied to an object, but the force is not the same in all directions. Unlike lithostatic pressure, high levels of directed stresses are not sustained for long because rocks deform to reduce the stress. Directed stress, thus, is commonly associated with rock folding or faulting.



Normal and shear stress within folds

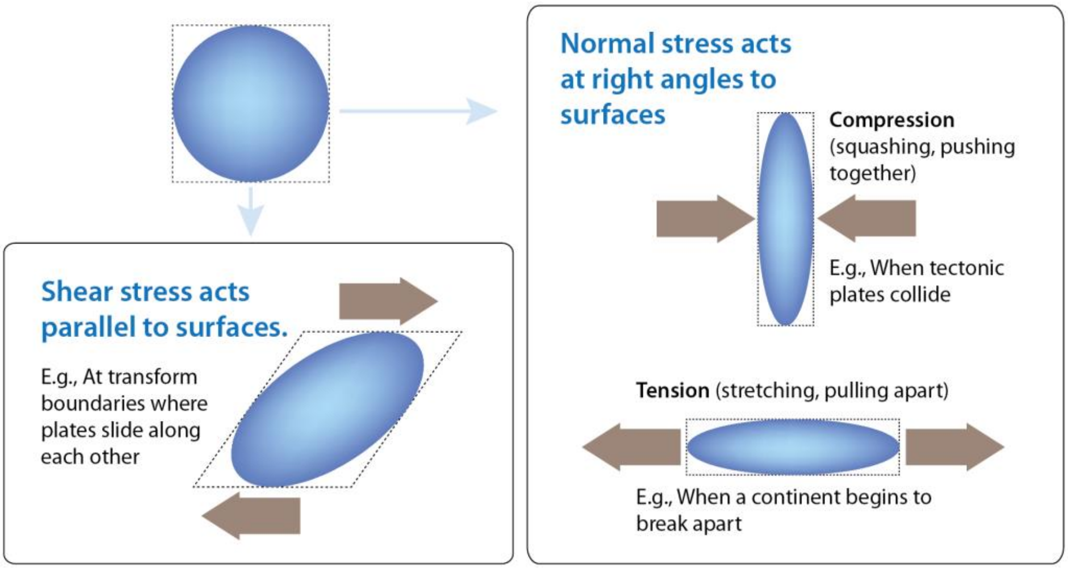
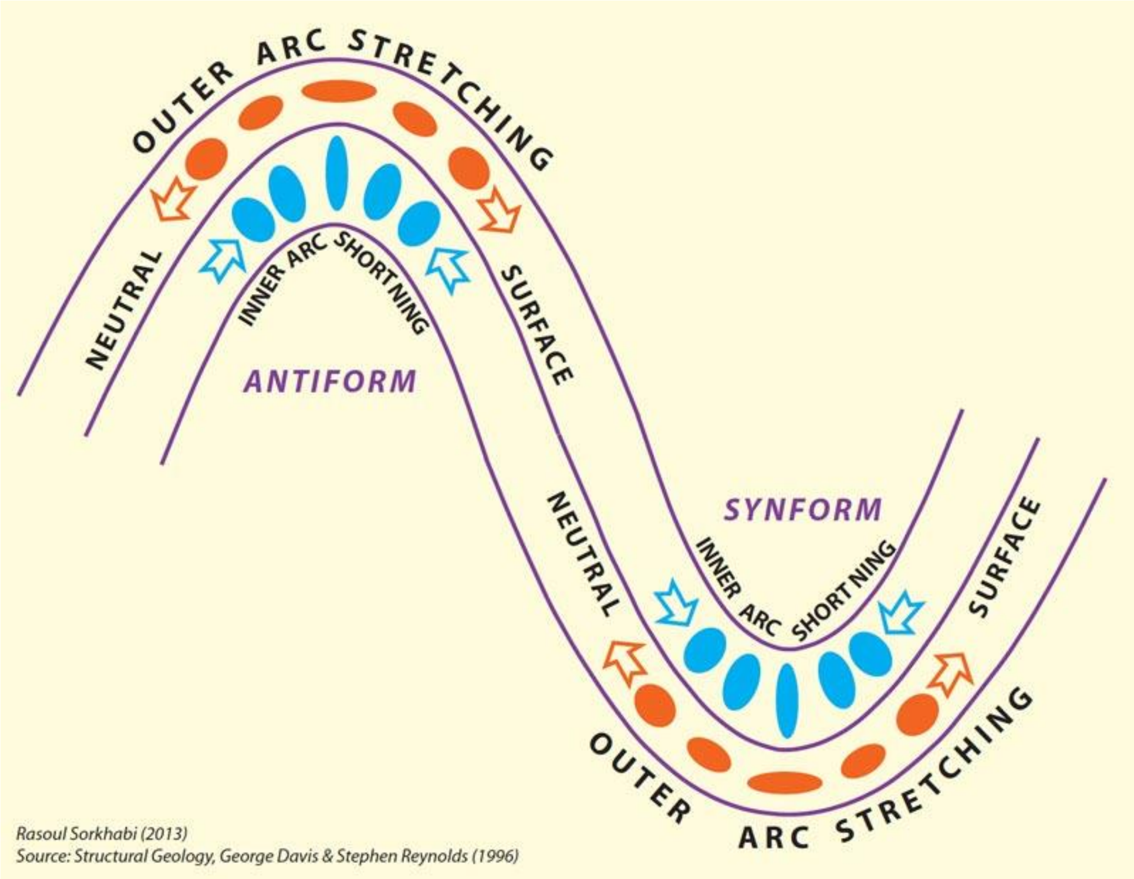
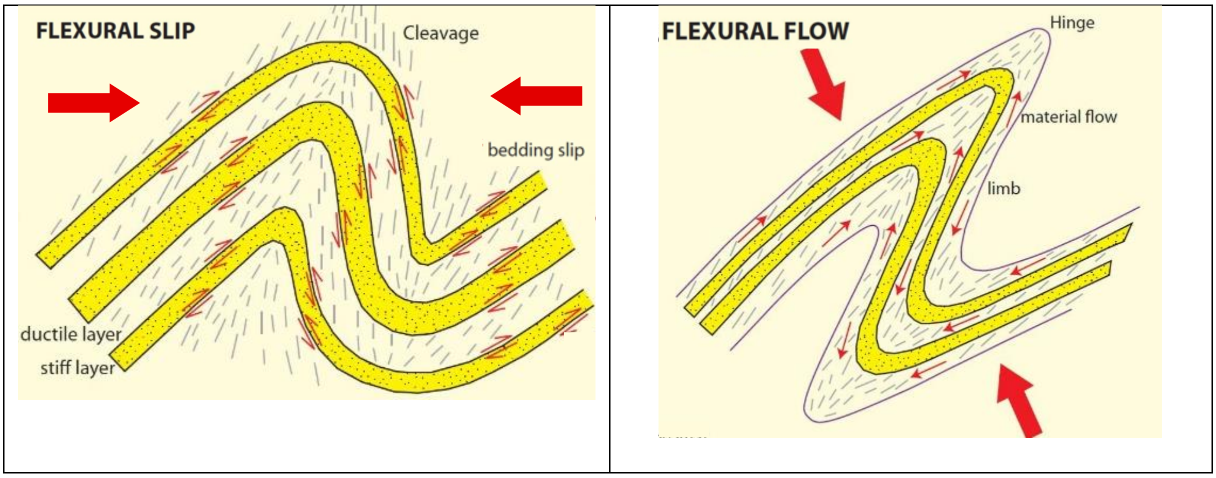


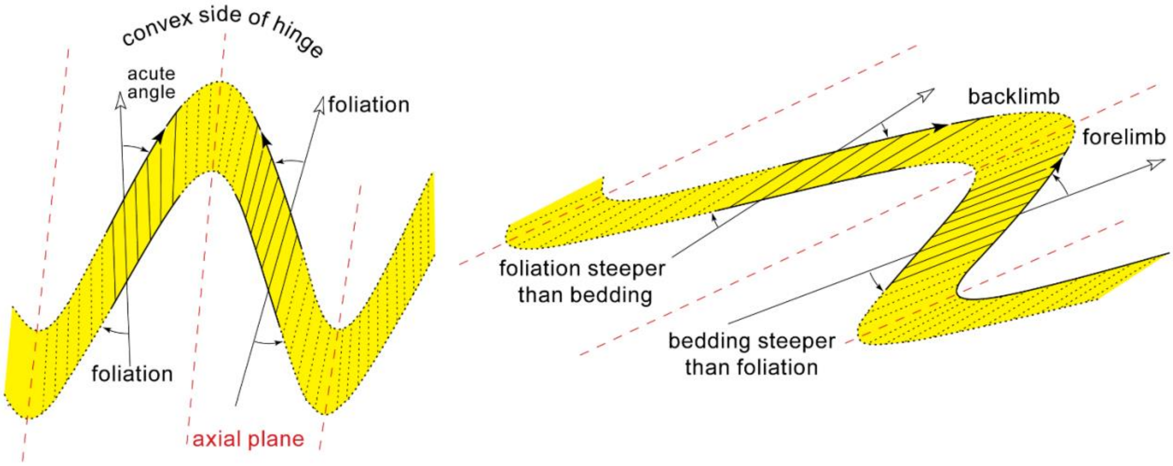
Figure 13.2 *Rocks can be affected by normal stress (compression and tension) or shear stress. [Source: Karla Panchuk (2016) CC BY 4.0, Physical Geology , First University of Saskatchewan edition]*





Two modes of flexural folding: layer-parallel flexural slip of sedimentary rocks in a parallel fold at relatively lower temperatures and pressures; and flexural flow of weak layers in a simple fold. (Modified from Robert Hatcher, Structural Geology, 1995).

In flexural-flow folds, rock material in incompetent layers flows from fold limbs toward fold hinges, and therefore appreciable thickness changes occur in the rock layer. Obviously, flexural-flow requires more ductility contrast between layers than flexural slip. Flexural flow produces similar folds in the weak layers.



Angle relationship between bedding and foliation in fold profiles, Foliation jpb, 2020

References

Billen M., 2021, LibreTexts, Geopghysics (https://geo.libretexts.org/Courses/University_of_California_Davis/GEL_056%3A_Introduction_to_Geophysics/Geophysics_is_everywhere_in_geology.../01%3A_Rheology_of_Rocks/1.03%3A_Elastic_Deformation)

Manutsoglu E, Steiakakis E., 2011: Contribution to the hydrogeological research of Western Crete, Technical University of Crete, Department of Mineral Resources Engineering, University Campus, Akrotiri, 73 J 00 Chania, Hellas.

Davis P., An Introduction to Geology, Free Textbook for College-Level Introductory Geology Courses (https://opengeology.org/textbook/6-metamorphic-rocks/) Salt Lake Community College

jpb, 2020, Website Folds

Kull U., 2012: Kreta, Sammlung geologischer Führer

Krahl, J., et al. ,1983: Neue Daten zur Biostratigraphie und zur tektonischen Lagerung der Phyllit-Gruppe und der Trypali-Gruppe auf der Insel Kreta (Griechenland). - Geol. Rundsch. 72 1147-1166.

Nichols G., 2009: Sedimentology and Stratigraphy

Rahl J. M. et. al.: Exhumation of high-pressure metamorphic rocks within an active convergent margin, Crete, Greece: A field guide, Jeffrey M. Rahl, Charalampos Fassoulas2, and Mark T. Brandon1, 1 – Department of Geology and Geophysics, Yale University, New Haven, Connecticut 06511, U.S.A. 2 – Natural History Museum of Crete, University of Crete, Heraklion 71409, Greece

Seidel M., 2003: Tectono-sedimentary evolution of middle Miocene supra-detachment basins (western Crete, Greece), Ph.D. Dissertation, University of Köln.

Stöckhert B. et. al, 1998: Low effective viscosity during high pressure metamorphism due to dissolution precipitation creep: the record of HP-LT metamorphic carbonates and siliciclastic rocks from Crete, Faculty of Earth Sciences. Ruhr-Universität, Bochum

Thomson et. al. S. T., 1999, Miocene high-pressure metamorphic rocks of Crete, Greece: rapid exhumation by buoyant escape, Institut für Geologie, Ruhr-Universität Bochum,

Wassmann S., 2012 Geländekurs Kreta, Ruhr-Universität, Bochum Institute of Geology, Mineralogy and Geophysics

Wassmann S., 2012: Low stress deformation along the plate interface in subduction zones - the microstructural rock record, Ruhr-Universität, Bochum Institute of Geology, Mineralogy and Geophysics