The Plattenkalk Unit in the Talea Ori Mountains, Vossakos Traverse

View of the road leading south to the village of Kamos Doxarou. To the left road cuttings expose the Platy Marble Unit and the Vossakos folds. At this particular location the Giglios Beds may be observed.

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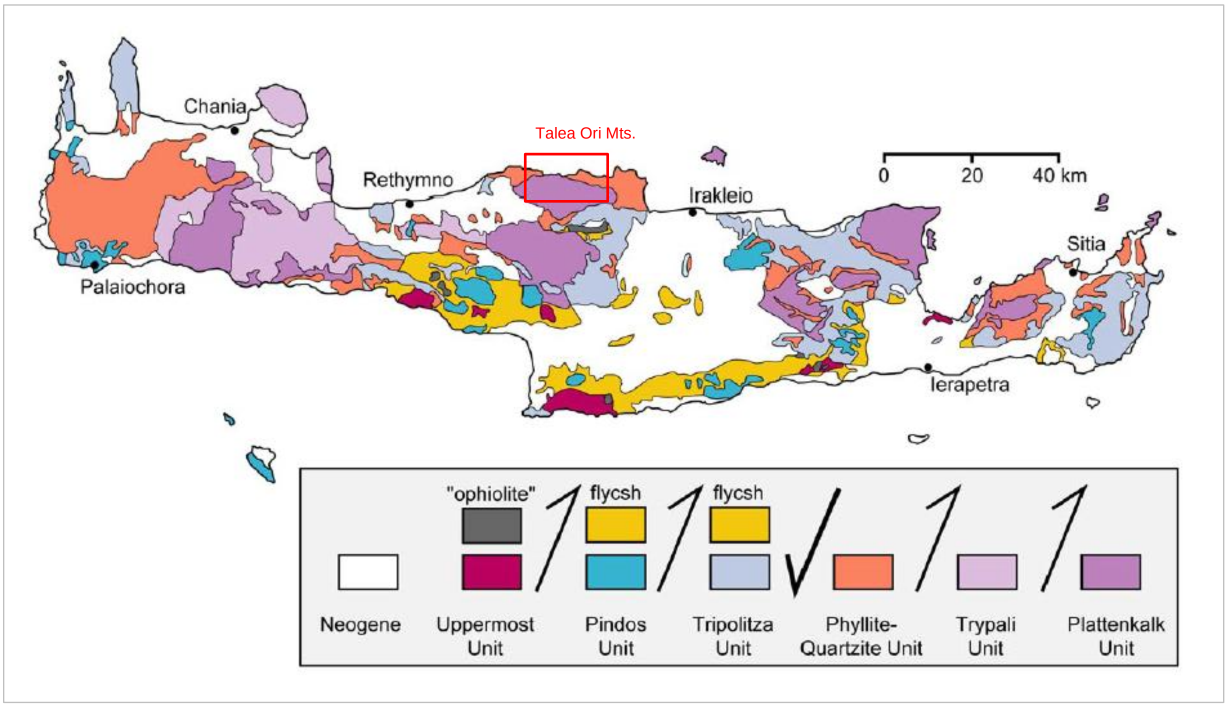
Berlin, December 2024

Contents

Appendix

1 Introduction - Plattenkalk Nappe (Platy Marble)

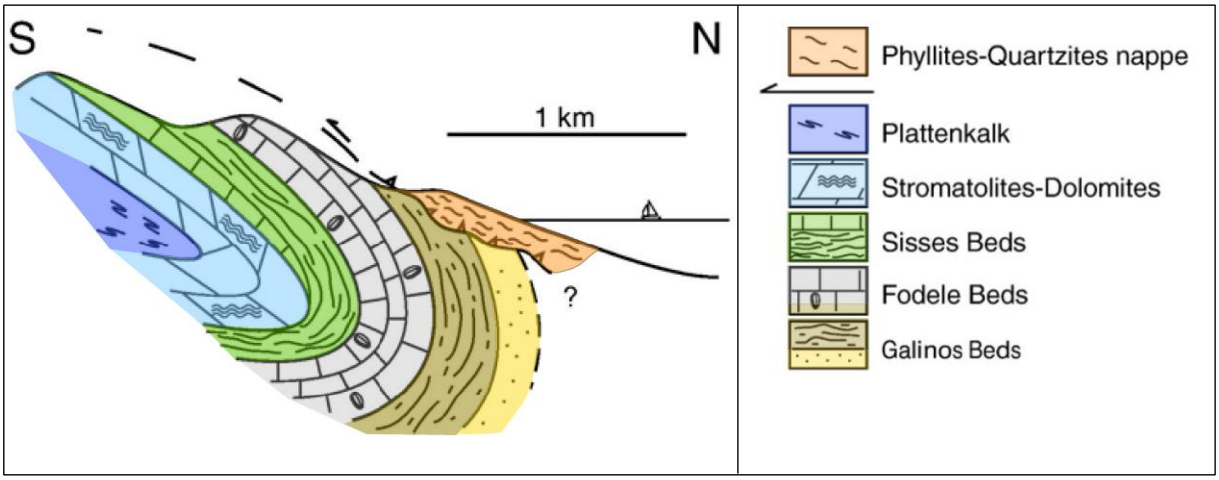


Simplified geologic map of Crete showing the remains of the different nappes. Arrows within the legend indicate major thrust planes. The bold arrow represents the Cretan Detachment, which divides the lower metamorphic Units from the less or non-metamorphic Pindos und Tripoliza Units. After Creutzburg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

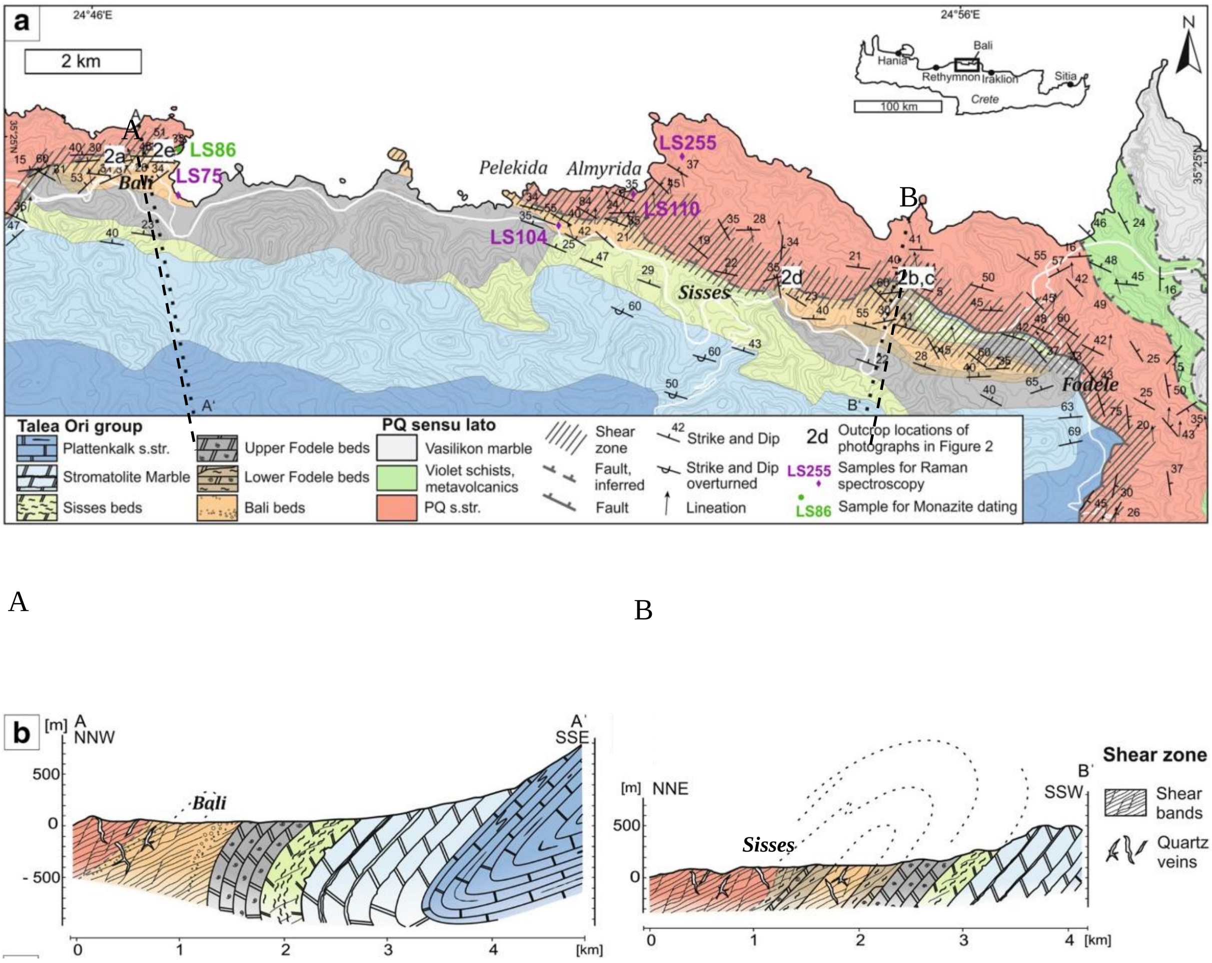
Subduction involves the horizontal convergence of two tectonic plates; however, researchers have long recognized that widespread extensional deformation is a characteristic of many subduction zones (e.g., Royden, 1993). The Hellenic subduction zone is one such region of overall plate convergence and widespread horizontal extension. As the lithosphere of the African plate dips to the north, deformation in the overriding European plate has thinned the continental crust, forming the Aegean Sea. Unlike in some subduction zones (e.g., Japan), extension in the overriding plate is not restricted to the back arc region; thinned continental crust underlies the Sea of Crete outboard of the volcanic arc. Although most of the attenuated continental crust is now submerged, the island of Crete provides a view of extensional deformation in the forearc region. Crete exposes a variety of variably metamorphosed sedimentary and volcanic units juxtaposed by thrust faulting during the Oligocene and later thinned by recent (and still active) normal faults. This tectonic thinning has exhumed high pressure-low temperature metamorphic rocks, which were only recently metamorphosed (~20 Ma) at about 35 km depth above the subducting slab. Thus, the island provides an excellent laboratory to study processes and consequences of syn-convergent extension. *[Rahl et. al., 2004]*

Most of the geologic units exposed on Crete were initially assembled by thrust imbrication during the Oligocene (Creutzberg and Seidel, 1975; Bonneau, 1984; Hall et al., 1984). That initial sequence was thinned dramatically by normal faulting, starting at ~15 Ma and continuing to the present (Thomson et al., 1998, 1999). Even so, the original nappe sequence formed during Oligocene thrusting is still readily apparent, as shown on the geological map above. The initial assembly of the nappes is widely thought to represent the collision of a microcontinent (known as Adria or Apulia plate) with the European margin (e.g., Robertson et al., 1991). However, note that this accretion of continental rocks may have been largely thin-skinned in that there is no break observed in the slab beneath Crete. Bonneau (1984) demonstrated that many units in Crete are present to the west and east into mainland Greece and Turkey, respectively. In contrast to other subduction zones, the structural style recorded in Crete is well organized, consisting of imbrication and folding of coherent units. *[Rahl et. al., 2004]*

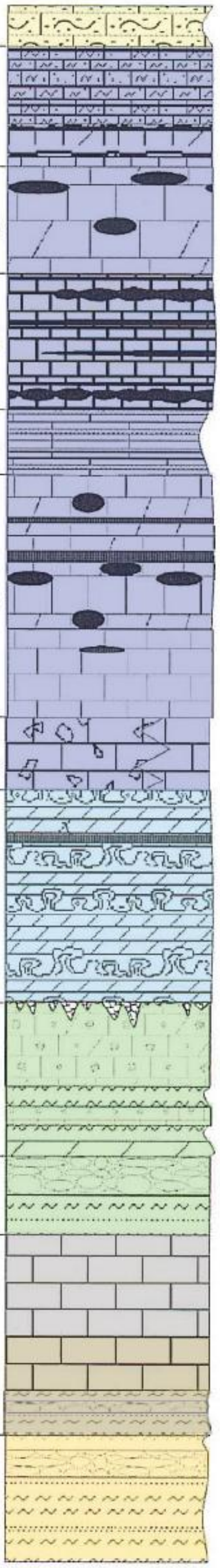
The structurally lowest unit exposed on Crete is the Plattenkalk nappe (called Ida nappe by some). The Plattenkalk Group is commonly overturned, and large recumbent folds are found in some areas. Nonetheless, the unit maintains a coherent well defined internal stratigraphy consisting of stromatolitic dolomite, carbonate breccia, and a distinctive sequence of platy, well-bedded carbonates with chert interbeds. The oldest Plattenkalk consists of shallow marine carbonates with Permian fusilinids (Bonneau, 1984). Younger parts of the section consist of carbonates and cherts that record a transition to deeper water conditions. The thickness of the Plattenkalk is debated, with estimates ranging from 1 and 2.5 km (compare Epting et al., 1972 and Hall and Audley-Charles, 1983). The Plattenkalk is devoid of siliciclastic input except for a thin, 10 – 30 m layer of flysch at the top of the section. In central Crete, foraminifera from this flysch have been dated by Bizon et al. (1976) as Oligocene (29.3 to 28.3 Ma). The Plattenkalk is thought to represent the sedimentary cover of the southern margin of the Adria continental block (sometimes referred to as the Apulia block), which presently underlies the modern Adriatic Sea. The Adria block is being overridden to west by the Apennine thrust belt and to the east by the Dinarides and Hellenides of the western Balkans and Greece. It also continues northward into the Alps, where it forms the highest structural unit in that Alpine collision zone. Bonneau (1984) correlated the Plattenkalk Group with the carbonates of the Ionian Zone of western mainland Greece. The Plattenkalk is thought to be a fully allochthonous nappe. Although its base is not observed on Crete, the presence of an active subduction zone beneath Crete supports the interpretation that it is an accreted tectonic slice. Furthermore, to the east in Rhodes, the Plattenkalk is found overthrust on a more inboard carbonate platform, associated with the “pre-Apulian domain” from mainland Greece (Bonneau, 1984). *[Rahl et. al., 2004]*



General structure of the Talea Ori window. The bedding is generally sub-vertical or overturned, dipping northwards. North-south section through a syncline between Bali Bay and Sisses. [Kock S. et al, 2007]



Tectonostratigraphy of the Talea Ori, central Crete, modified after Epting et al. (1972). The structural data and location of the shear zone are based on Seybold et al. (2019).

Formation	Description	Age
 5 **Kalavros Fm.**	Flysch: Calc-phyllite and phyllite	**Lower Oligocene**
4f–4a **Aloides Fm.** (Plattenkalk s. str.)	f: Thin bedded marble with red and green carbonate silt horizons and sponge remains e: Dark thick marble beds and white chert nodules d: Dark thin marble beds and interlayered white ropy chert layers (silica sponges) c: **"Gigilos-Beds"** Chert-claystone-carbonate sequence (phyllites and yellowish metasandstones) b: Dolomite marbles with chert nodules a: Carbonate breccias and white marbles, dolomitic breccias, stromatolite dolomite marbles	**Liassic – Eocene** Talea Ori: Cephalopods, Kuss, 1982 Krahl et al. 1988. Eastern Crete: Eocene foraminifers, Fytrolakis, 1972; rudists Wachendorf et al. 1980
3 **Mavri Fm.** (Stromatolith -Dolomite)	Dolomite or calcite marbles with stromatolites. Transgressional conglomerate at base	**Norian – Liassic** (Foraminifers, Epting et al. 1972)
2b, 2a **Sisses Fm.**	b: light dolomite marbles partly sericitic, oolite and oncolite marbles. Metabauxite in karst on top a: greenish to violet phyllites, carbonate meta-conglomerate, sericite quartzite	**Lower Triassic - Upper Triassic** a: conglomerate: Upper Triassic: miliolid Foraminifers Kock, 2007; Olenekian: conodonts König and Kuss 1980; Radiometric dating (Seybold et al., 2019)
1c **Fodele Fm.** (Talea Ori-Group)	c: Upper Fodele Fm. dark fossil-rich dolomite marble	**Middle to Upper Permian** Corals, fusulinids, bryozoa (Epting et al., 1972)
1b	b: Lower Fodele Fm. dark dolomite marble, black shales and metasandstones	**Upper Carboniferous - Lower Permian** Brachiopods, trilobites, etc. (König and Kuss 1980) Radiometric dating (Seybold et al., 2019)
1a	a: Bali and Galinos Beds black schists and phyllites, quartzites, meta-conglomerate, meta-chert, rare patch reef	

Stratigraphy of the Platy Marble Unit (Plattenkalk s. l.) in Central Crete after Manutsoglu et al. 2003, modified after Seybold et al, 2019

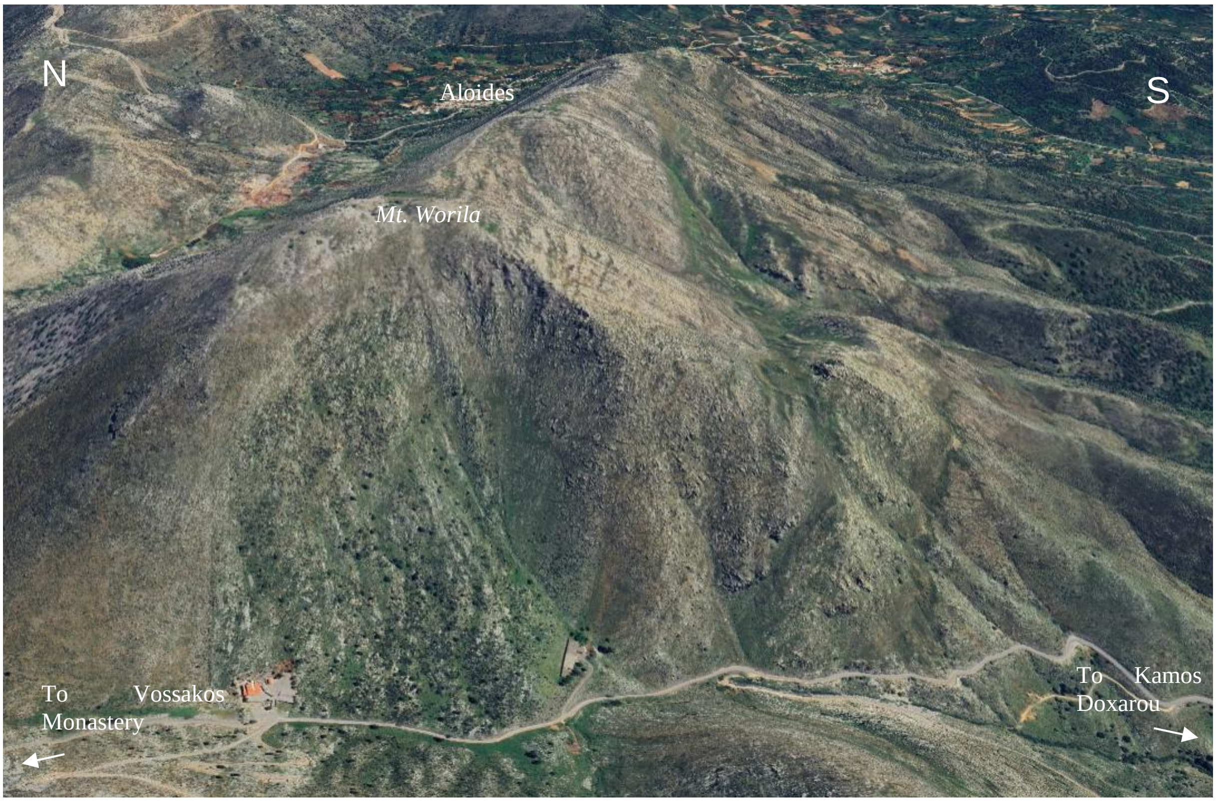
1.1 Talea Ori-Group

At the Talea Ori Mountains the Plattenkalk consists mainly of carbonate rocks from the shelf and deeper shelf areas. At the top of the stratigraphic sequence, the "Plattenkalk" is the thickest of the various formations. The Plattenkalk, which contains white chert nodules and layers is also known as the Aloides Formation and makes up a large part of the central mountain ranges on Crete e.g. the Ida Ori and Lefka Ori mountains. Only within the Talea Ori mountains do the lower formations of the section crop out – hence the name Talea Ori-Group. The sequence of predominantly carbonate rocks begins with the Fodele Formation, the basal part of which consists of the Galinos shales. The Fodele Formation is overlain by the Sissi Formation followed by Mavri Formation. The time line within the Talea Ori section begins approx. at the Upper Carboniferous and extends to the Liassic and Eocene. *[Rahl et. al., 2004]*

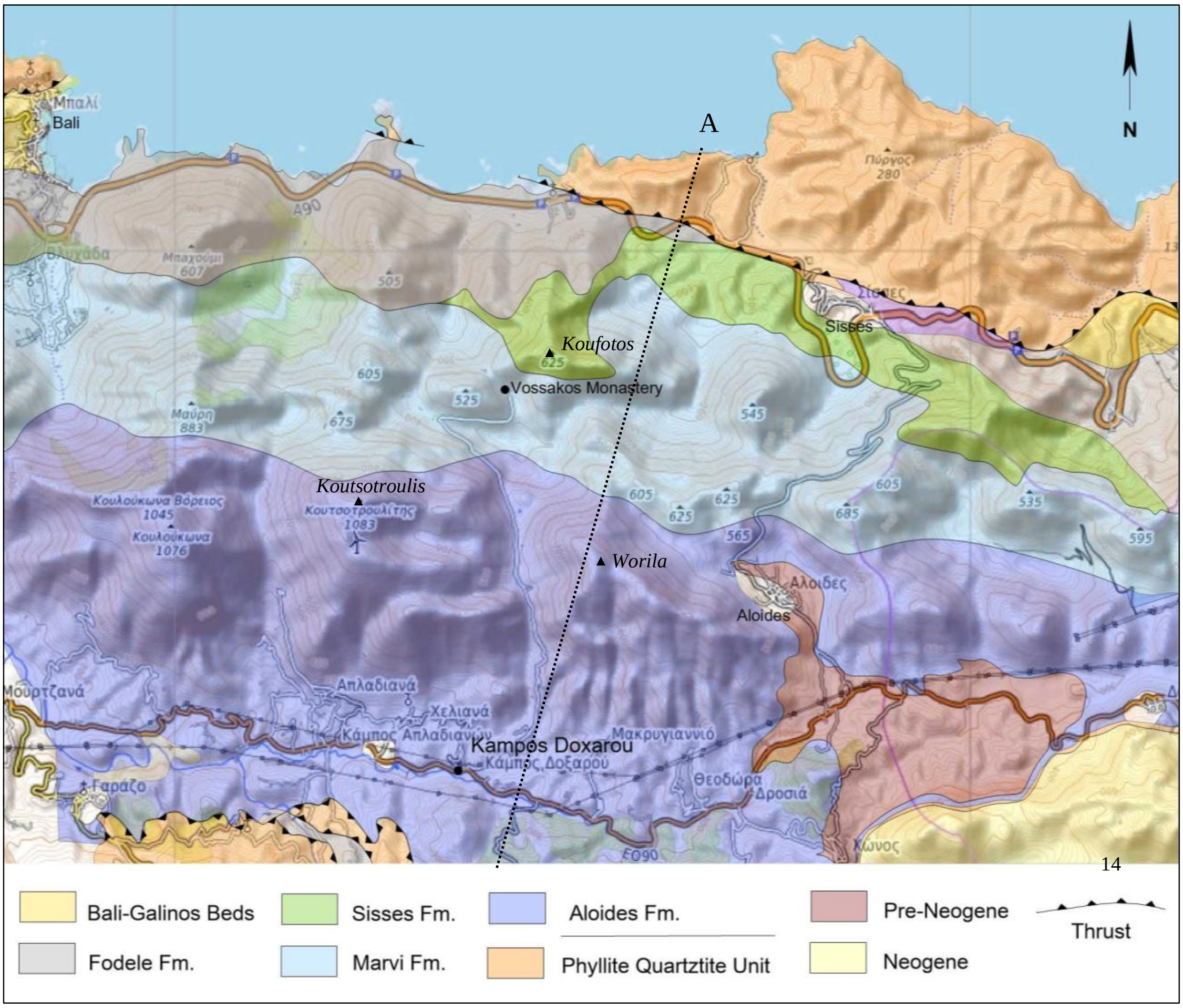
Near the Vossakos Monastery the Talea Ori Group is part of an overturned limb of a south-vergent fold and is therefore structurally inverted (Seybold et al. 2019). The following text describes the different Plattenkalk units encountered on a traverse from the village of Kamos Doxarou to the Vossakos Monastery and beyond beginning at the top most unit of the sequence.

1.2 Doxaro-Vossakos Traverse

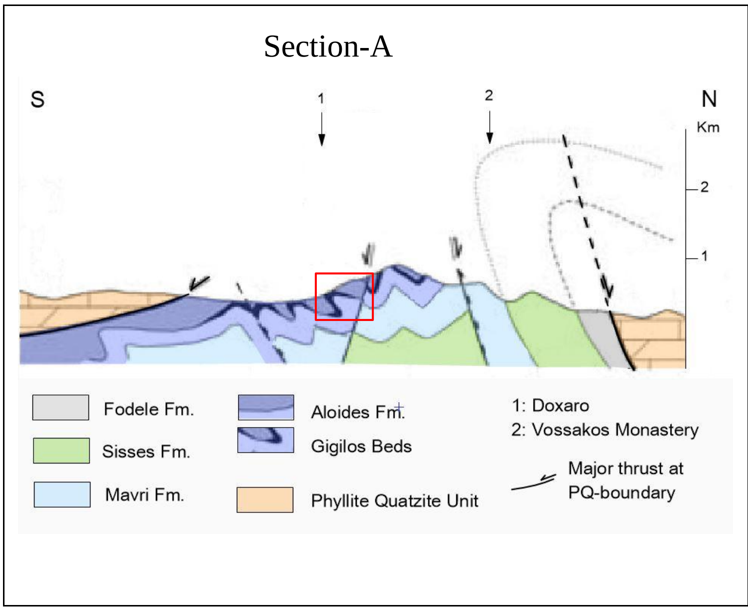
When starting from Kamos Doxarou, the Doxaro-Vossakos Traverse reveals the same formations as the Sisses-Aloides Travers, but in reverse order (see also My GeoGuide: No. 27 The Platy Mable in the Talea Ori Mountains, Sissi Traverse) - the younger rocks appear first then the older ones. Along the road leading from the village of Kamos Doxarou to the Vossakos Monastery, several subunits of the Aloides Fm. such as the yellowish “Gigilos” schist and white marbles can be observed. It is within the Aloides Fm. where some fine examples of semi-ductile folds are exposed. These have been termed the Vossakos Folds.

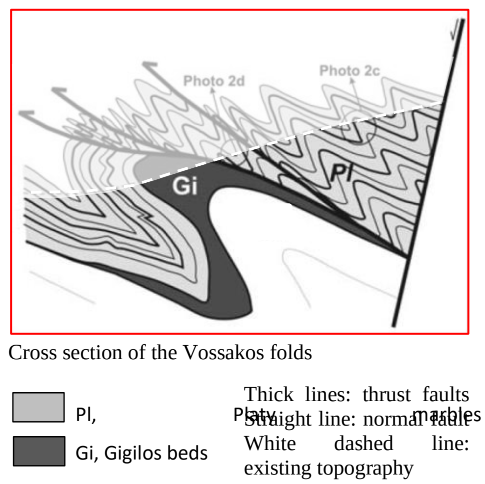


Location of villages and road cuts [Source of image Google Maps]



Geological map of the Talea modified after Kock, 2007





Cross section of the Vossakos folds

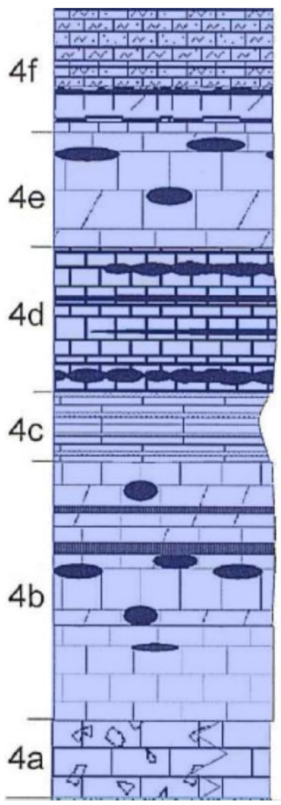
Deformation pattern in the Vossakos area, [Fassoulas C, 2004]. Note that in this drawing the major faults are deemed to be normal faults.

2 Aloides Formation (Plattenkalk s. str.)

2.1 Mt. Worila - South of Pass

Within the Aloides Formation, particularly in western Crete, there are thick siliciclastic beds, that have been transformed to phyllite during metamorphism. These rocks are known as the Gigilos Beds (see following Fig. - Section of the Aloides Formation).

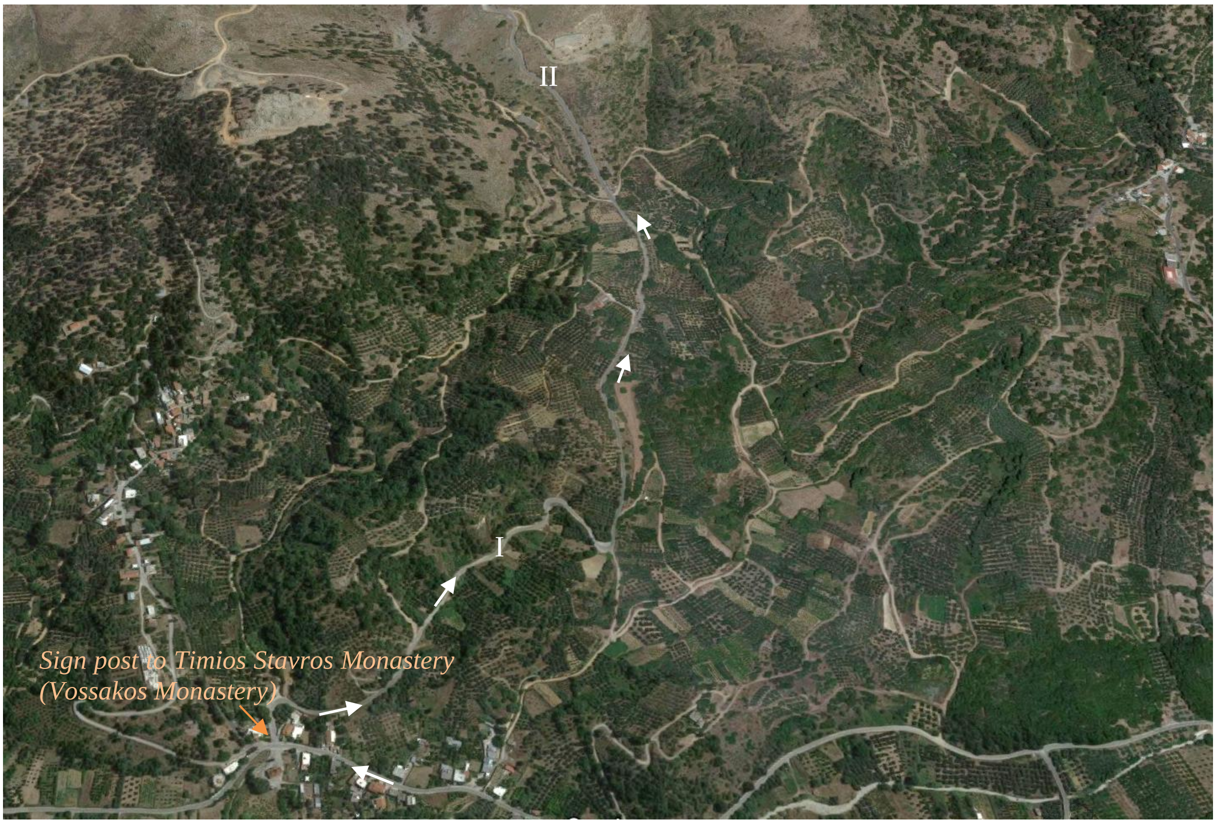
In general, the Aloides Formation forms a monotonous sequence numerous chert layers and chert nodules. At some locations the surfaces between the beds are regarded. The thick-bedded parts of the sequence are often coarsely recrystallized, while the thinner beds are finely crystallized. The coarse grained marbles are thought to represent pathways, along which fluids traveled and enabled recrystallisation during metamorphism. As metamorphism despite the HP/LT conditions is often not recognizable within the rock’s texture, the marble is regarded to have been very dry (Theye & Seidel 1991). The thickness of the Aloides Formation is quite considerable reaching more than 1500 m. The carbonate platform upon which the limestone was deposited is therefore thought to have subsided continuously over a long period of time. The Aloides Formation extends from the Lower Jurassic to the Lower Oligocene. Reddish limestone beds often occur in the upper most parts. In the upper part of the section sponge reefs with demosponges indicate that the water depth cannot have been more than 300-400 m (Manutsoglu et al. 1995a, Soujon et al. 1995) (see also No. 28 My GeoGuide: Platy Marble of the Psilorites Mountains). Many of the isolated chert nodules are probably poorly preserved sponges. Other fossils in the Aloides Formation are extremely rare. Turbiditic sequences are missing in the Aloides Formation indicating it to be a typical marine carbonate shelf formation.



Thin bedded marble with red and green carbonate silt horizons and sponge remains
Middle to thick bedded calcitic marbles, white chert-nodules and -layers
Dark thin calcitic marble beds and interlayered white ropy chert layers (possibly silica sponges)
Gigilos-Beds: Chert-claystone-carbonate sequence (phyllites and yellowish metasandstones)
Dolomite marbles with white chert nodules and layers Dark platy marble with white chert nodules Platy marble
Carbonate and dolomitic breccias passing into white marbles and stromatolite dolomite marbles

Section of the Aloides Formation at the Talea Ori Mountains, Central Crete [Manutsoglu et al. 2003]

In general, across the Island of Crete the Platy Marble Group (Plattenkalk s. l.) displays large fold structures with W-E trending fold axes. The upper part of the Platy Marble shows pronounced brittle tectonics, such as faulting and sharp hinged (zig zag) folds, while the lower parts of the unit tend to have undergone more ductile deformation. In the Vossakos area the Aloides Fm. displays ductile asymmetric, south-vergent folds. These folds have roughly east-west trending hinge lines, plunging gently towards the east. The rocks reportedly experienced higher temperatures than those exposed further south in the Psiloritis Mountains and therefore have rounded hinges as opposed to those folds found further south. Folding commenced during the Oligocene approx. 23 million years ago.



Route to outcrops from Kamos Doxarou. Outcrop I: Aloides Fm. [Source of image Google Maps]

Outcrop I: Aloides Fm. (possibly subunit 4e), gentally dipping sequence of dark marble and white silica layers "unfolded" at this location (rucksack at bottom centre for scale).

Overview of outcrops. II, III: Mesoscale recumbent folds, IV Vossakos Folds, V: Continuous zig zag folds, VI: Thick bedded marble and white chert layers, VII: Large calcite crystals in white marble, VIII: Gigilos Beds, IX: Hinge zone thickened by flexural slip, X: Marble quarry. Dashed line: normal fault or thrust plane [Source of image: Google Maps]

Outcrop II: recumbent folds. The right side of the outcrop reveals part of a hinge giving a three-dimensional view of the fold. Based on the numerous white chert layers the outcrop is thought to belong to the 4d subunit.

Outcrop III: mesoscale recumbent fold displaying asymmetric folding (subunit 4d). Box shows position of next picture

Outcrop III: thickening and buckle-folding at the inverted limb (Fassoulas C., 2004)

Outcrop IV Vossakos Folds. Chevron folds are a structural feature characterized by repeated well behaved folded beds with straight limbs and sharp hinges. Well developed, these folds display repeated sets of V-shaped beds. Inter-limb angles are generally 60 degrees or less. Chevron folding preferentially *occurs when the bedding regularly alternates between contrasting competences such as limestone and chert or sandstone and phyllite (subunit 4d).*

Outcrop IV: Closeup of previous picture. 1: Boudinage in Vossakos folds. Boudinage is the result of internal extensional stress within the layer.

Outcrop V: Tight, strongly asymmetric parallel zig-zag folds (chevron folds) consisting of a sequence of dark marble and white chert layers (subunit 4d).

Outcrop VI: Thick bedded grey marble and white chert beds of the Aloides Fm.

Outcrop VI: closeup of grey marble

Outcrop IX, 4b: Isoclinal folding and axial plane schistosity. The hinge zone is thickened by flexural slip along the limbs (Fassoulas C., 2004) (see Appendix).

2.1.1 White marble

Outcrop VII: White marble. The presents of white marble in this stratigraphic position between the beds of the Aloides Fm. indicates that the white marble was off-set along a major normal fault.

Outcrop VII: Closeup of coarse-grained white marble

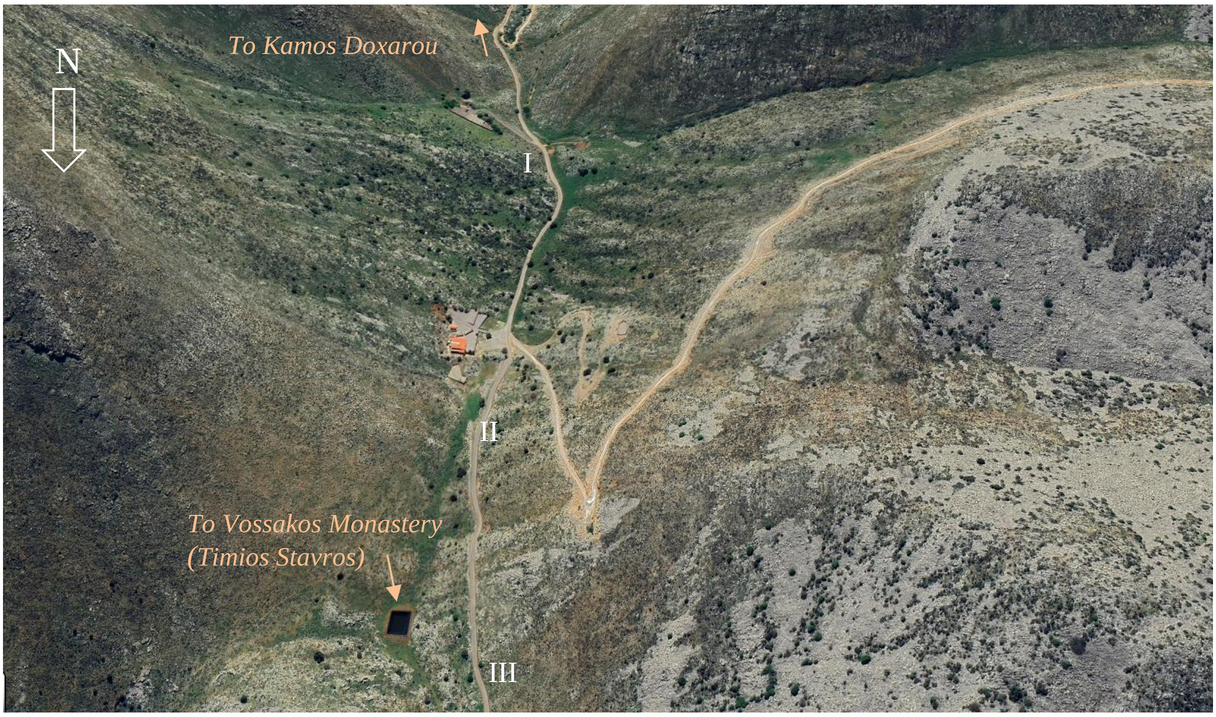
X: Marble quarry, in the background the Psiloritis Mts.

2.1.2 Gigilos Beds

Outcrop VIII: Gigilos Beds belong to the 4c subunit of the Aloides Fm. At this location the Gigilos Beds are highly weathered appearing as yellowy phyllite.

Outcrop VIII: Gigilos Beds. The phyllite has a high carbonate content indicating the protolith to have been a marl.

2.2 Mt. Worila - North of Pass



[Source of image Google Maps]

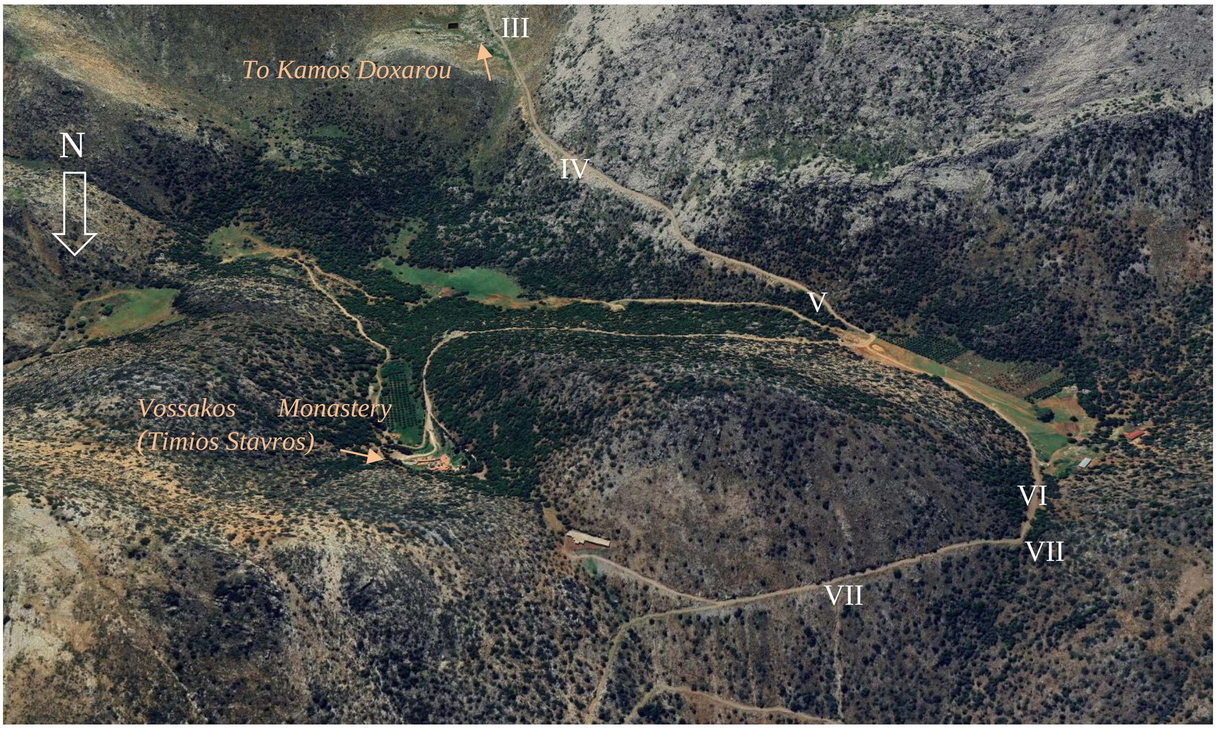
Outcrop I, 4b: Typical Aloides Fm displaying dark grey marble with white chert nodules

Outcrop II. Boulder at the side of the road possibly displaying recrystallized stromatolites. It is thought to belong to the 4a subunit of the Aloides Fm.

Closeup of a sample from Outcrop II. White marble

Outcrop III: light grey marble breccia thought to belong to the 4a subunit of the Aloides Fm.

Outcrop III, Close up of a sample from Outcrop III. White marble with small cavities. The cavities could be the result of dolomitization.



[Source of image: Google Maps]


Outcrop IV: white marble displaying hardly any structures besides the crystal grains


Outcrop IV: closeup of previous picture


Outcrop V: white marble


Outcrop V: closeup of sample of relatively coarse-grained white marble displaying numerous cavities

3 Mavri Formation

The Mavri and Sisses formations are exposed just north of the Vossakos Monastery and are accessible either on foot or with an appropriate vehicle by dirt road. At the base of the Mavri Fm. there is a transgressive conglomerate associated with a hiatus (see next Section). The matrix of the conglomerate contains an association of Norian–Rhaetian foraminifer species (Epting et al.,1972). The clasts consist of Sisses limestone (oolite), the Fodele limestone and clasts of stromatolitic dolomite. Overlying the conglomerate is a very thick, homogeneous, stromatolitic dolomite sequence, representing at least part of the Liassic (König and Kuss, 1980). Intraformational breccias and “birds eyes” structures (i.e. small cavities related to algae) are observed in many places. *[Kock, 2007]*

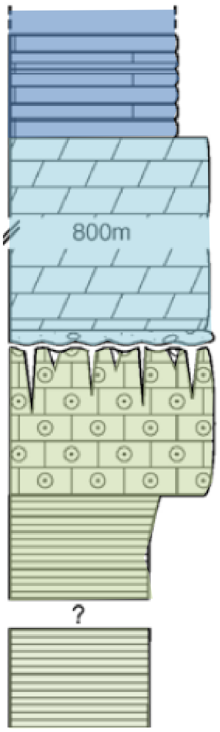
Outcrop VI, Mavri Fm: stromatolite dolomite

Outcrop VI, Closeup of previous picture displaying stromatolite dolomite with the characteristic wavy structure formed by stromatolites (arrow)

Outcrop VII, Mavri Fm: boulder of stromatolite dolomite

Outcrop VIII, View of the Koufotos Mtn. and the reported erosional boundary to between the Mavri Fm. and Sisses Fm (Kock, et. al., 2007). Mavri Fm. at the right side of the road.

4 Sisses Formation

4		**Aloides Formation** (Plattenkalk, Platy Marble) Dark thick marble beds and white chert nodules	Liassic – Eocene
3		**Mavri Formation** (Stromatolite Dolomite) Dolomite marbles with stromatolites. Transgressions conglomerate at base	Norian – Liassic
2		**Upper Sisses Beds** Light dolomite marbles partly sericitic, oolite and oncolite marbles. Meta-bauxite in karst on top	Late Triassic
		Lower Sisses Beds Greenish to violet phyllites, carbonate meta-conglomerate, sericite quartzite	Early to middle Triassic

Geological section showing the stratigraphic position of the Sisses Beds near the Koufotos Mtn. north of the Vossakos Monastery. At the Koufotos Mtn the Sisses Beds are relatively thick compared to other locations [Kock et. al.,2007]

Due to local erosion, significant thickness variations are recorded in this formation throughout the Talea Ori mountains. The sequence includes (from stratigraphic bottom to top): greenish–reddish marl and dolomite alternations, bivalve-bearing claystones, limestone conglomerate, claystones and massive, partly oolitic limestones. Conodonts found by König and Kuss (1980), reveal a Scythian age for the lower part of the formation. In the Koufotos area, palaeokarsts with lateritic fillings can be observed at the erosional boundary to the oolitic limestone. *[Kock et. al., 2007]*

4.1 Palaokarst with Lateritic Fillings

The contact with the overlying Mavri Formation is an angular unconformity arising from uplift and erosion. The former erosional surface displays karst cavities that are filled with lateritic/bauxitic material. Note that the bedding is inverse and that it steeply dips to the NNE. Some authors (Hall and Audley-Charles, 1983; Krahl et al., 1988) have proposed that the lower part of the Talea Ori unit (Galinos, Fodele and Sisses Beds) constitutes an individual block thrust onto the Stromatolites–Dolomites and Plattenkalk. Although it is clear that the contact between the Sisses Beds and the Stromatolites–Dolomites has subsequently undergone tectonic movement, its stratigraphic, erosive nature can still be observed in some locations such as at the Doxaro-Vossakos Traverse. *[Kock 2007]*.

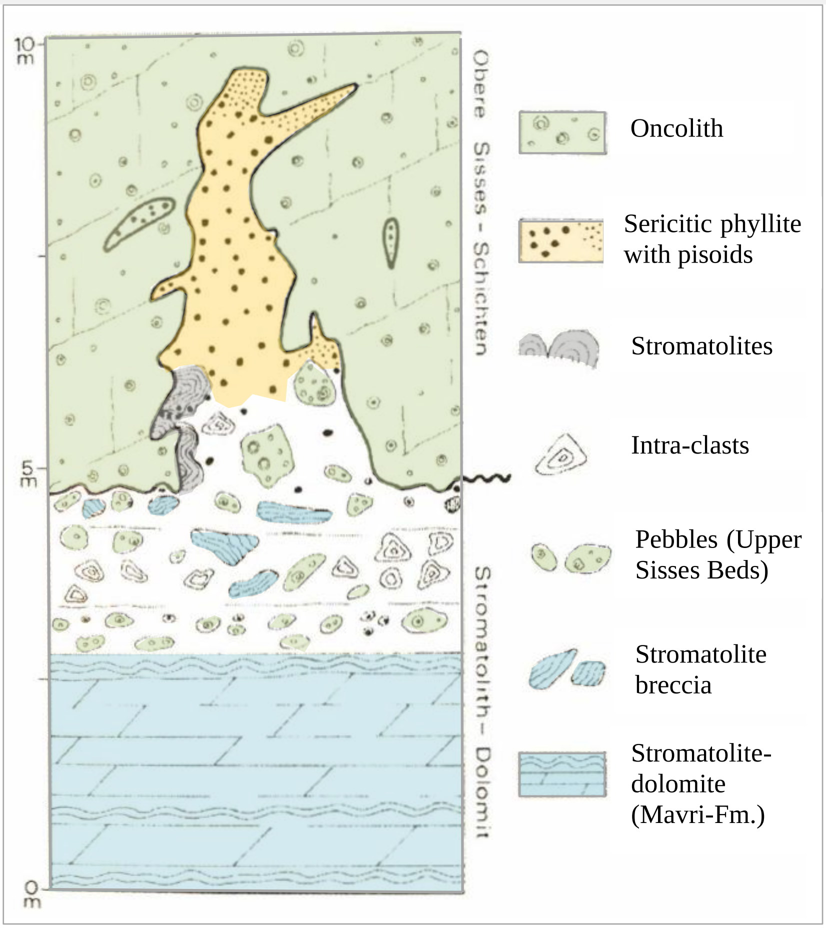
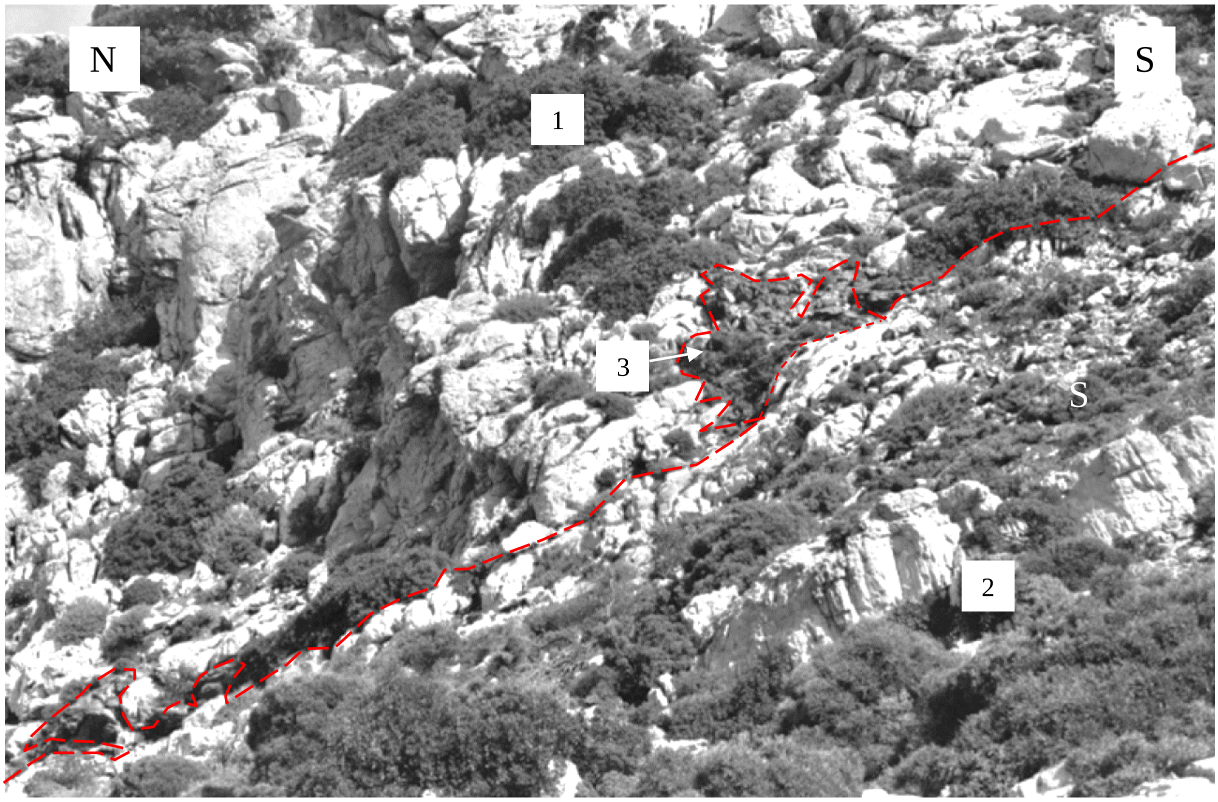


Abb. 27: Unconformity between the Sisses- and Mavri-Formation in an inverse position, with karst cavity containing metabauxite. (Epting et al. 1972).

The figure above shows a drawing of a karst cavity and its filling within the Sisses Fm. It contains products of laterite weathering such as yellow to red-brown meta-bauxites, and pebbles of limestone and sericite schists that were deposited during a transgression. The karst filling has been dated and is considered to be of Norian age. The mineral carpholite found within the meta-bauxite indicates HP/LT metamorphism of the Platy Marble Unit (Talea Ori Group, Plattenkalk). The karst filling is important as there are no other existing indicator minerals within the marble sequences (Seidel et al. 1982). *[Kull]*



Outcrop on the western slope of the Koufotos Mtn. north of the Vossakos Monastery showing the stratigraphic and erosive contact between the Sisses Beds (1) and the Stromatolites–Dolomites (2). Palaeokarst with laterite/bauxite filling are visible at the interface (3). The stratigraphy is overturned; [Kock S. et al, 2007]. The palaeokarst is not visible along the Sisses-Aloides Traverse, but it is exposed at the Koufotos mountain.

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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

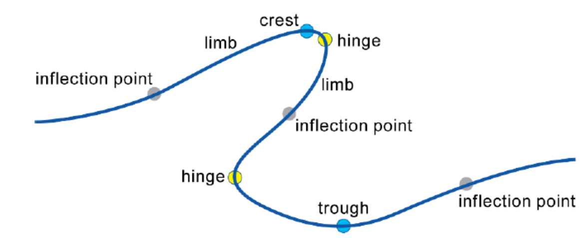
Folds

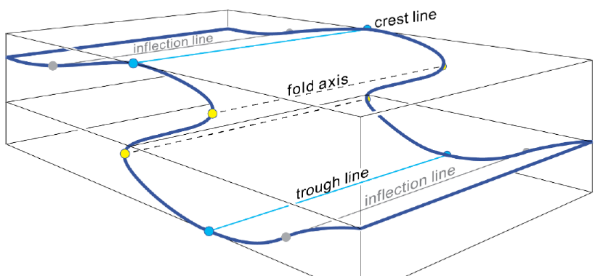
Basic geometrical definitions

A proliferation of terms has developed to describe the considerable variation of fold morphology. For convenience, it is easier to first define folds on a single surface.

Morphology of a folded surface: Hinge, limb, inflections

The radius of curvature of a folded surface varies progressively from point to point. The point with the smallest radius of curvature is the hinge. It is flanked by two areas with a larger radius of curvature: the limbs. The inflection points are points of zero curvature, where the sense of curvature changes from a convex to a concave line. They usually are aligned on either limb of a fold. If the limb has a straight segment, its midpoint is conveniently taken as the inflection point.



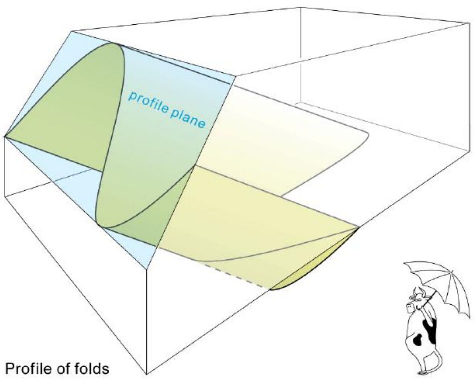


Cylindrical fold

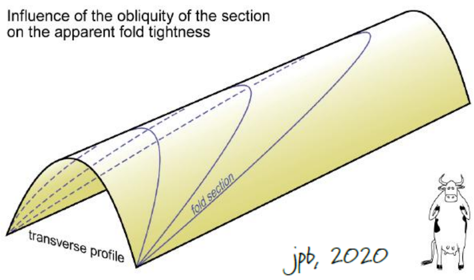
Folds are three-dimensional structures. Connecting the hinge points on a specific folded surface defines the **hinge line** or **fold axis**. The limb may thus be redefined as the fold segment between a hinge line and the adjacent **inflection line**, which is the locus of inflection points. The fold axis is the most important structural element of a fold because it shows the direction of maximum continuity of this fold.

Profile

The **transverse profile** (or simply profile) of a fold is the section drawn perpendicular to the fold axis and axial surface; this contrasts with a geological section which is normally drawn in a vertical plane.

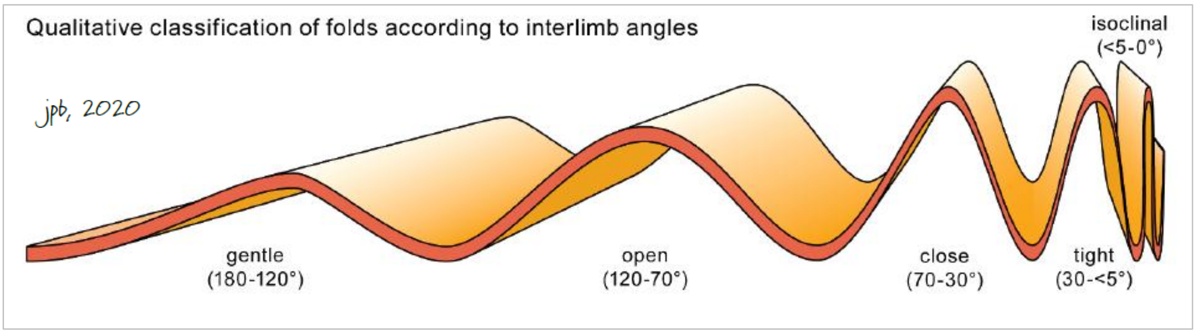


Profile of folds



Burg, Jean-Pierre, 2018: Script to Structural Geology, ETH Zürich

The profile is an ideal reference plane used to describe and measure all geometrical characteristics of the fold: height or amplitude, wavelength, tightness, roundness. Indeed, these aspects vary with the angular relationship between any section plane and the folded surface.



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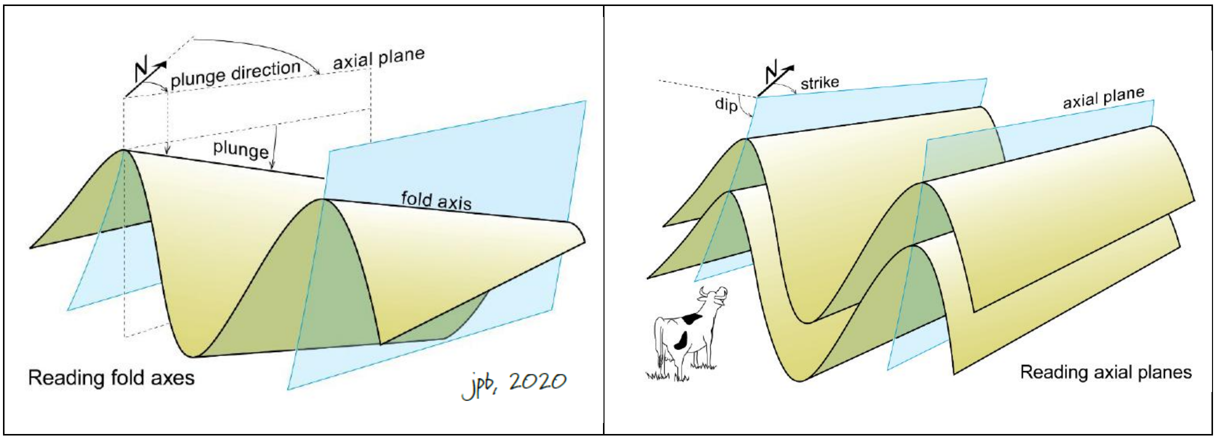
Fold axis orientation

The orientation of a fold axis is expressed by its **plunge** and its **plunge azimuth**:
- The **plunge** is the inclination measured from the horizontal in the imaginary vertical plane containing the hinge line.
- The direction of plunge (the trend) is the strike (azimuth, the bearing relative to North) of the imaginary vertical plane that contains the hinge line and the direction in which the downward inclination occurs.



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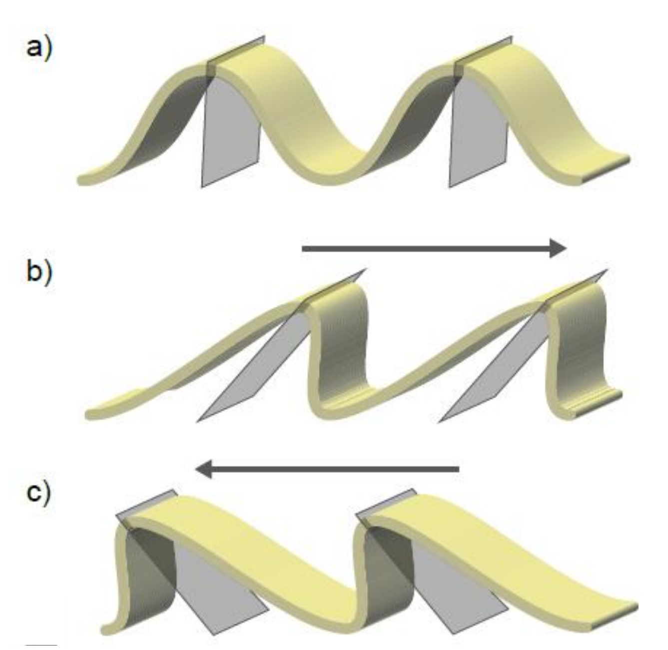
A general rule is that both the trend and plunge of minor order folds can be used for extrapolation in fieldwork. They also indicate the trend and plunge of first-order folds of the same generation.

Axial plane orientation

The axial planes cut the hinge zone of a folded surface along the fold axis. The orientation of the axial plane is expressed as a dip and strike or, in a more compact form, as a dip and direction of dip. As for any surface, the strike is the trend of the horizontal line contained in the surface; the angle dip is the angle between the surface and the horizontal plane. The direction of the dip of a surface is the trend of the line perpendicular to the strike of the surface looking down the dip.

Fold symmetry and vergence



a) Symmetric fold with neutral vergence.
b) Asymmetric fold verging to the right.
c) Asymmetric fold verging to the left

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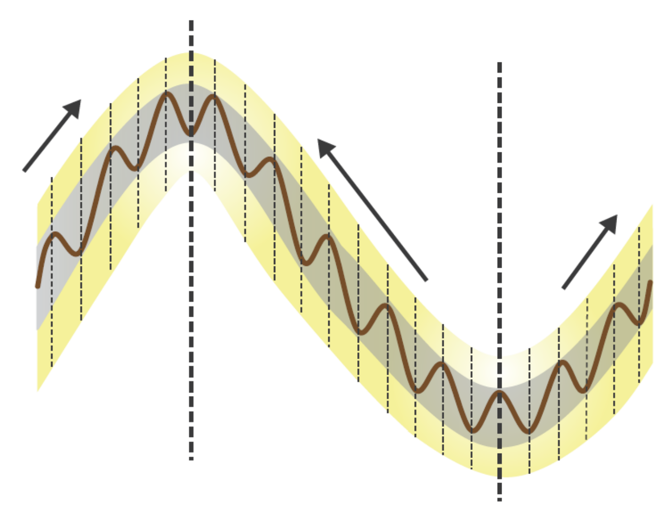
The symmetry of a fold relates to its limb lengths. Symmetric folds have equal limb lengths and the two sides of the fold are mirror images. Asymmetric folds have a shorter and a longer limb. A series of folds with the same asymmetry are said to have vergence. The direction of vergence is determined by the sense of displacement of the upper limb relative to the lower limb. When viewed down plunge, a fold verges to the right where there is apparent clockwise rotation of the short limb and to the left where there is apparent anti-clockwise rotation of the short limb.

Minor folds



Geometric relationships between first and second order
folds. Black arrows give direction of vergence of the minor folds.

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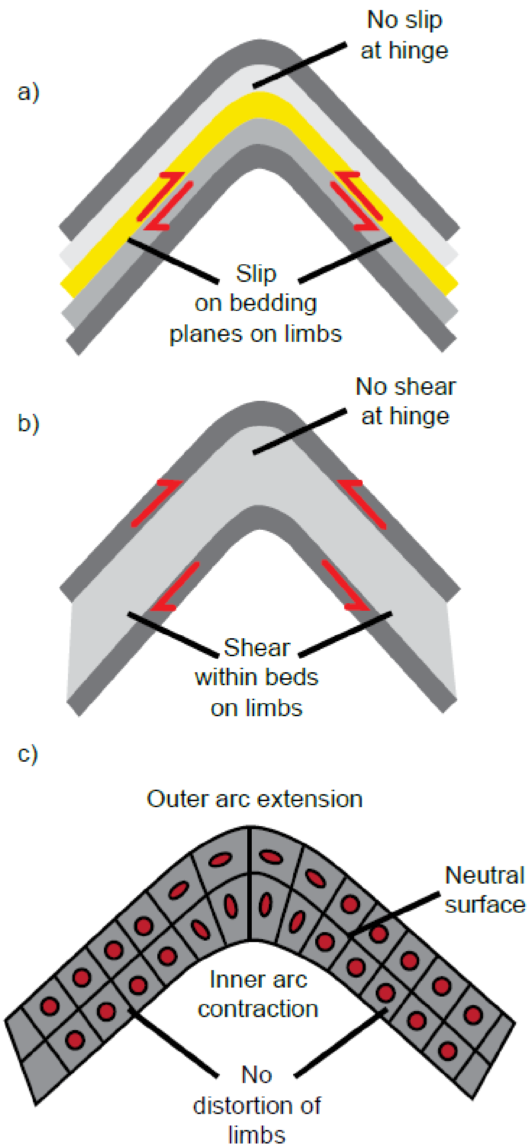
Large scale (or first order) folds often have minor (second order) folds associated with them, which formed during the same phase of deformation. These folds form in the same stress field and so have the same geometric features. The axial planes of the minor and large-scale folds are parallel. Within the hinge zones of the larger folds the minor folds are symmetric. On the limbs of the large-scale folds the minor folds are asymmetric. The minor folds verge towards the hinge zone of the larger scale antiform.

These geometric relationships are particularly useful in the field where the vergence of the minor folds can be used to predict the position of larger scale folds.

How rocks fold



a) Flexural slip. b) Flexural flow.
c) Tangential longitudinal strain

Rock layers accommodate layer parallel shortening by folding. This shortening may be taken up either along the limbs with slip concentrated between or within the layers (flexural slip or flow) or concentrated around the hinge zone (tangential longitudinal strain). In both cases bed thickness will be maintained.

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Flexural slip and flow

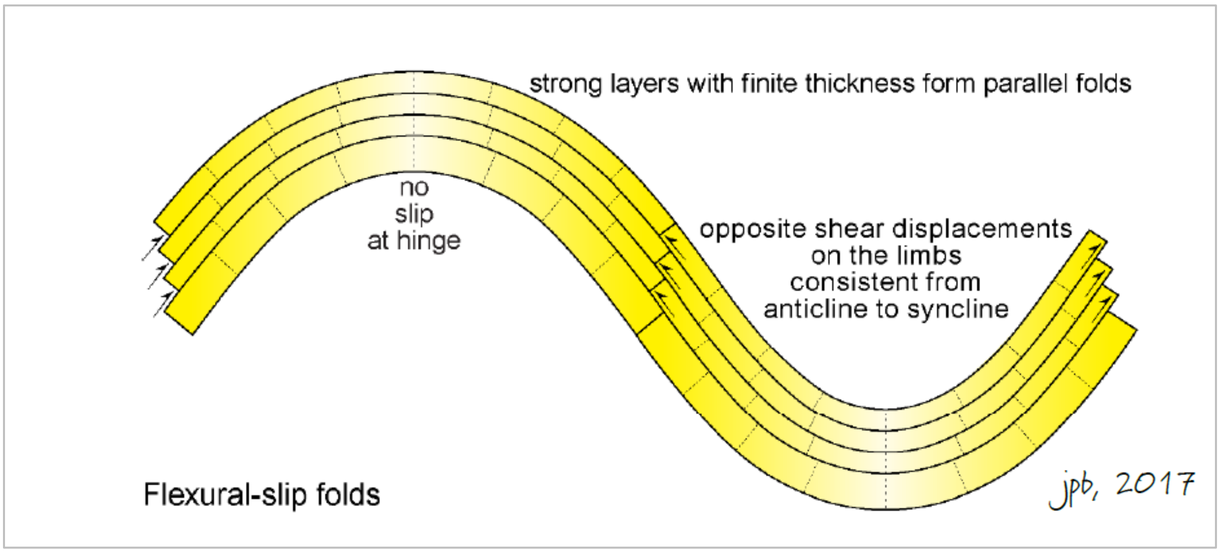
In flexural slip and flexural flow strain is concentrated in limbs and dies out towards hinge. The limbs show opposite senses of shear. Flexural slip occurs in well layered rocks usually in the brittle regime. The slip is concentrated between beds or along incompetent layers (e.g. shale). Flexural flow occurs in the ductile regime and strain is evenly distributed across the limb.

Tangential Longitudinal Strain (TLS)

Tangential longitudinal strain (also known as orthogonal flexure) occurs in more homogeneous, thicker, less well layers rocks (figure c). Strain is concentrated in hinge zone with no strain along limbs. It results in outer arc extension and inner arc contraction across the hinge zone. Areas of extension and contraction are separated by the neutral surface along which there is no strain. Veins often develop around the outer arc of the hinge zone where it is stretched and pressure solution cleavage (figure 55) or small scale thrust faults in the inner hinge zone where it is compressed.

Influence of discontinuities: Flexural-slip and flexural flow

A multilayer can be a pile of competent layers separated by surfaces of discontinuity or alternating layers of highly contrasting competence. The mechanical consequence is that the competent layers on either side of the surface of discontinuity or of a weak layer may easily slide relative to each other. This shear “decoupling” of layers allows a fold to accommodate a greater flexure than if the stack deforms as a single layer.



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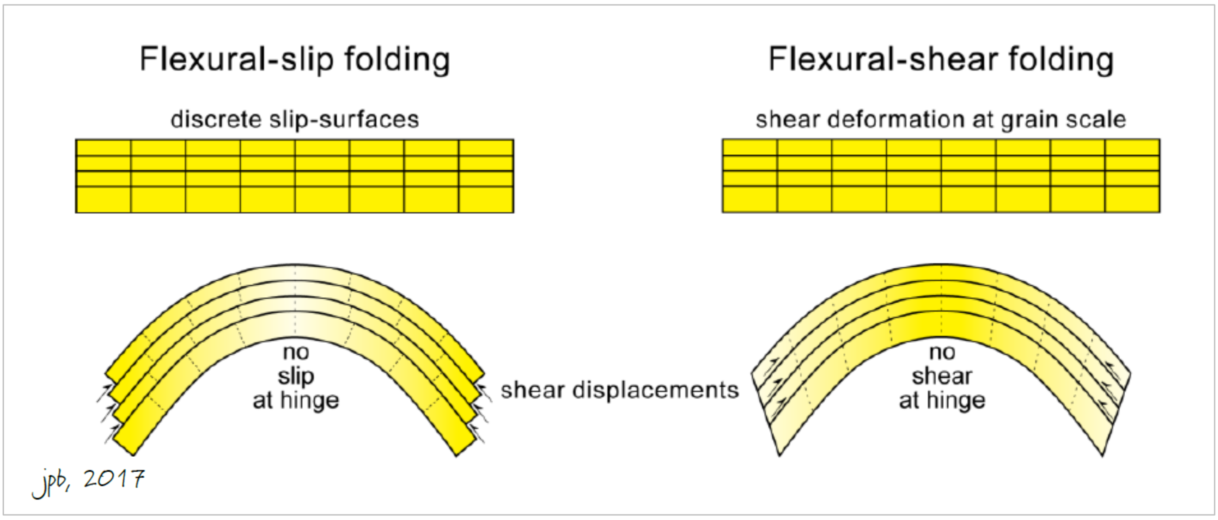
Flexural-slip describes discrete faulting, usually coincident with bedding planes and accompanying folding. A classical simulation is to bend a book or pile of paper sheets; increasing bending about the fold axis is accommodated by increasing slip between the pages of the book or sheets of the pile. The thickness of individual sheets does not change, meaning that each sheet makes a parallel fold (i.e. layer surfaces remain parallel). Slip is an important part of folding because layer-parallel stresses increase with increasing rotation of the limbs.

When the shear stress exceeds shear resistance of weak layers or layer boundaries, the strong layers in the limbs slip over each other towards and usually perpendicular to the hinges, which are fixed from layer to layer. Therefore, slickensides and fibrous mineral growth or other movement indicators showing reverse dip-slip on bedding planes within fold limbs are common criteria for flexural slip.

Flexural flow describes bedding-parallel shear homogeneously distributed within the ductile layer being folded between stiffer layers. Like for flexural slip, bedding-parallel shear in limbs is opposite across the axial plane. The strain pattern due to hingeward shear tends to develop thickened hinges between thinned limbs, i.e. flexural-flow folds are mostly **similar**. Flexural-flow is sometimes applied to the weak layers that take up bedding-parallel motion within larger parallel folds, generally under low metamorphic grade. In this case, the stiff, active layers tend to keep their thickness throughout the deformation to produce and control the overall shape of concentric and/or parallel folds while the incompetent layers undergo flexural flow. In order to maintain similarity from bed to bed, ductile material moves out of the limbs into the hinges. Natural examples of such similar folds show intense foliation in the fold limbs, which dies away from limbs towards hinge zones. The intensity of shear strain depends on fold shape and position within the fold, with shear strain equal to limb dip in radians.



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Oolite

Oolite is a sedimentary rock formed from ooids, spherical grains composed of concentric layers. Strictly, oolites consist of ooids of diameter 0.25–2 millimeters; rocks composed of ooids larger than 2 mm are called pisolites. The term oolith can refer to oolite or individual ooids. *[Wikipedia]*

Modern ooids from a beach on Joulter Cays, The Bahamas

Thin-section of calcitic ooids from an oolite within the Carmel Formation (Middle Jurassic) of southern Utah

Ooids are most commonly composed of calcium carbonate (calcite or aragonite), but can be composed of phosphate, clays, chert, dolomite or iron minerals, including hematite. Dolomitic and chert ooids are most likely the result of the replacement of the original texture in limestone.

They are usually formed in warm, supersaturated, shallow, highly agitated marine water intertidal environments, though some are formed in inland lakes. The mechanism of formation starts with a small fragment of sediment acting as a 'seed', such as a piece of a shell. Strong intertidal currents wash the 'seeds' around on the seabed, where they accumulate layers of chemically precipitated calcite from the supersaturated water. The oolites are commonly found in large current bedding structures that resemble sand dunes. The size of the oolites reflect the time that they were exposed to the water before they were covered with later sediment. *[Wikipedia]*

Stromatolites

WIKIPEDIA
Die freie Enzyklopädie

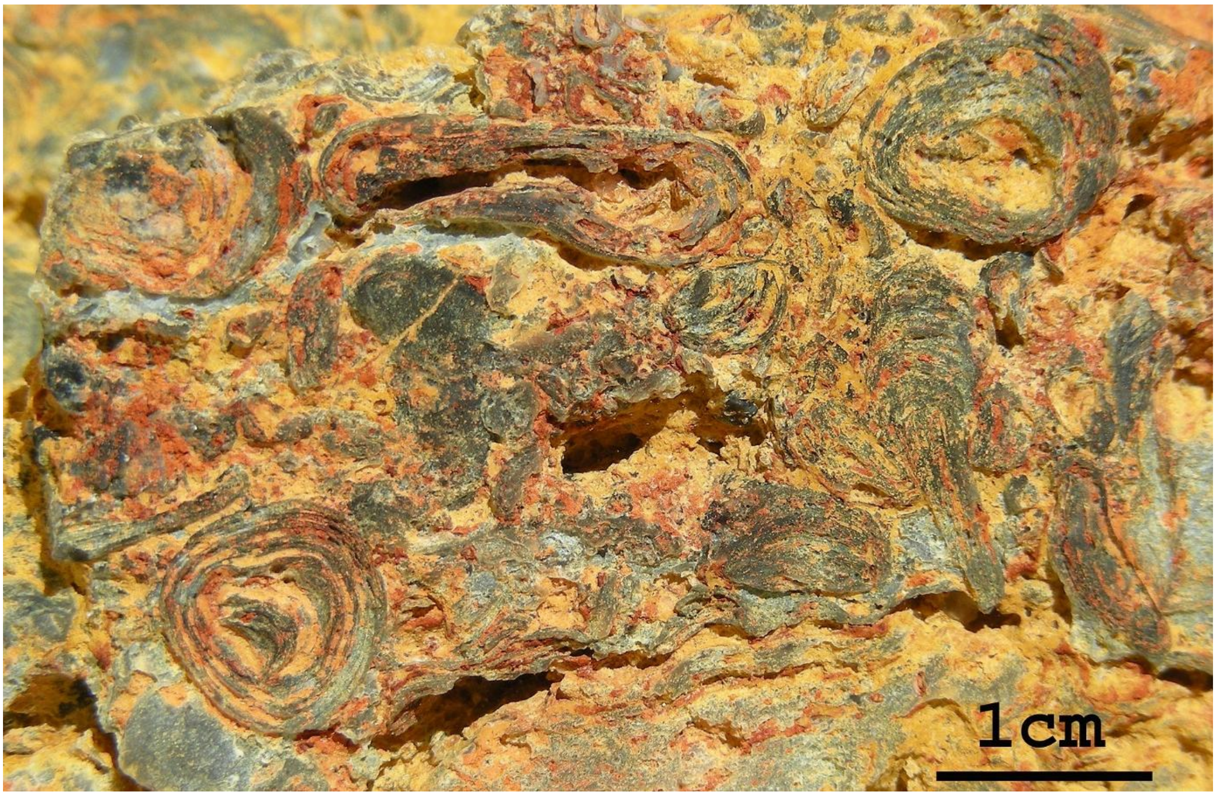
https://www.wikiwand.com/en/Stromatolite

Stromatolites are layered sedimentary formations that are created mainly by photosynthetic microorganisms such as cyanobacteria, sulfate-reducing bacteria, and Pseudomonadota (formerly proteobacteria). These microorganisms produce adhesive compounds that cement sand and other rocky materials to form mineral "microbial mats". In turn, these mats build up layer by layer, growing gradually over time. A stromatolite may grow to a meter or more. Fossilized stromatolites provide important records of some of the most ancient life. As of the Holocene, living forms are rare.

Modern stromatolites in Shark Bay, Western Australia

Stromatolites are a major constituent of the fossil record of the first forms of life on earth. They peaked about 1.25 billion years ago and subsequently declined in abundance and diversity, so that by the start of the Cambrian they had fallen to 20% of their peak. The most widely supported explanation is that stromatolite builders fell victim to grazing creatures (the Cambrian substrate revolution); this theory implies that sufficiently complex organisms were common over 1 billion years ago. Another hypothesis is that protozoans such as foraminifera were responsible for the decline, favoring formation of thrombolites over stromatolites through microscopic bioturbation.

Proterozoic stromatolite microfossils (preserved by permineralization in silica) include cyanobacteria and possibly some forms of the eukaryote chlorophytes (that is, green algae). One genus of stromatolite very common in the geologic record is Collenia.



Paleoproterozoic oncoids from the Franceville Basin, Gabon, Central Africa. Oncoids are unfixed stromatolites ranging in size from a few millimeters to a few centimeters

The connection between grazer and stromatolite abundance is well documented in the younger Ordovician evolutionary radiation; stromatolite abundance also increased after the end-Ordovician and end-Permian extinctions decimated marine animals, falling back to earlier levels as marine animals recovered. Fluctuations in metazoan population and diversity may not have been the only factor in the reduction in stromatolite abundance. Factors such as the chemistry of the environment may have been responsible for changes.

While prokaryotic cyanobacteria reproduce asexually through cell division, they were instrumental in priming the environment for the evolutionary development of more complex eukaryotic organisms. They are thought to be largely responsible for increasing the amount of oxygen in the primeval Earth's atmosphere through their continuing photosynthesis (see Great Oxygenation Event). They use water, carbon dioxide, and sunlight to create their food. A layer of polysaccharides often forms over mats of cyanobacterial cells. In modern microbial mats, debris from the surrounding habitat can become trapped within the polysaccharide layer, which can be cemented together by the calcium carbonate to grow thin laminations of limestone. These laminations can accrete over time, resulting in the banded pattern common to stromatolites. The domal morphology of biological stromatolites is the result of the vertical growth necessary for the continued infiltration of sunlight to the organisms for photosynthesis. Layered spherical growth structures termed oncolites are similar to stromatolites and are also known from the fossil record. Thrombolites are poorly laminated or non-laminated clotted structures formed by cyanobacteria, common in the fossil record and in modern sediments. There is evidence that thrombolites form in preference to stromatolites when foraminifera are part of the biological community.

The Zebra River Canyon area of the Kubis platform in the deeply dissected Zaris Mountains of south western Namibia provides an extremely well exposed example of the thrombolite-stromatolite-metazoan reefs that developed during the Proterozoic period, the stromatolites here being better developed in updip locations under conditions of higher current velocities and greater sediment influx. *[Wikipedia]*