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# GEOLOGY OF THE NORTHWESTERN KRANIA BASIN

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## Abstract

*We present a new map of 30 km$^2$ of the northwestern Krania Basin at 1:10,000 scale, including rocks of the Pindos Ophiolite Group and associated units, and the sedimentary fill of the Krania Basin. The Krania Basin is a flexural basin developed in the Middle – Late Eocene and filled first with alluvial fan conglomerates and later with turbidite sandstones and siltstones, following a deepening of the basin. Analysis of the clasts within the sediment, combined with paleoflow analyses, suggest sediment input from the eroding Pindos Ophiolite to the west. The Pindos Ophiolite Group is represented in the area by pillow lavas, sheeted dykes and serpentinized harzburgites of the Aspropotamos Complex. The ophiolite forms imbricated, thrust bounded blocks which show two phases of thrusting, corresponding to Late Jurassic and Eocene stages of ophiolite emplacement. We identify five stages of deformation within the basin itself, starting with Early - Middle Eocene syndepositional extensional faulting associated with the formation of the basin. This was followed by four stages of post-depositional deformation, starting with Late Eocene compression associated with basin closure, which caused thrust faulting and folding of the sediments. Oligocene dextral faulting with a thrust component affected the basin margins. Finally, two normal faulting events with different orientations have affected the basin since the Miocene.*

***Keywords:*** *Krania Basin, Pindos ophiolite, 1:10,000 map, sedimentology, structural geology*

**Περίληψη**

*Παρουσιάζεται ένας νέος χάρτης έκτασης 30 km$^2$ της βορειοδυτικής λεκάνης της Κρανιάς σε κλίμακα 1: 10.000, η οποία περιλαμβάνει πετρώματα του οφιολιθικού συμπλέγματος της Πίνδου και των συναφών ενοτήτων καθώς και των ιζηματογενών σχηματισμών της λεκάνης της Κρανιάς. Η λεκάνη της Κρανιάς είναι μια πτυχωμένη λεκάνη η οποία αναπτύχθηκε στο Μέσο-Άνω Ηώκαινο και η οποία αρχικά δέχθηκε αλλουβιακά ριπίδια και αργότερα τουρβιδιτικούς ψαμμίτες και ιλυόλιθους. Στη συνέχεια ακολουθεί εκβάθυνση της λεκάνης. Η ανάλυση των κλαστών των ιζημάτων, σε συνδυασμό με τις αναλύσεις παλαιοροής, υποδεικνύουν την συνεισφορά ιζημάτων από τη διάβρωση του οφιολιθικού συμπλέγματος της Πίνδου στα δυτικά. Το εν λόγω οφιολιθικό σύμπλεγμα στην περιοχή αποτελείται από μαξιλαροειδείς λάβες, φλέβες με στρώση και σερπεντινιωμένους χαρτζβουργίτες του συμπλέγματος του Ασπροποτάμου. Οι οφιόλιθοι εμφανίζονται σε φολιδωτά, οριζόντια οριοθετημένα τεμάχια. Τα τεμάχια αυτά χαρακτηρίζονται από δύο φάσεις επώθησης, που αντιστοιχούν στα στάδια τοποθέτησης των οφιολίθων κατά τη διάρκεια του Άνω Ιουρασικού και Ηωκαίνου αντίστοιχα. Επιπλέον, αναγνωρίζονται πέντε στάδια παραμόρφωσης εντός της λεκάνης, ξεκινώντας με τη συνιζηματογενή εφελκυστική ρηγμάτωση του Κάτω - Μέσου Ηωκαίνου, η οποία συνδέεται με τη δημιουργία της λεκάνης. Ακολουθούν τέσσερα στάδια μετα-αποθετικής παραμόρφωσης, με έναρξη τη συμπίεση του Άνω Ηωκαίνου που σχετίζεται με το κλείσιμο της λεκάνης, η οποία ακολουθείται από την επώθηση, ρηγμάτωση και πτύχωση των ιζημάτων. Η δεξιόστροφη Ολιγοκαινική ρηγμάτωση με επώθηση επηρέασε τα περιθώρια της λεκάνης. Τέλος, από το Μειόκαινο η λεκάνη επηρεάστηκε από δύο γεγονότα δημιουργίας κανονικών ρηγμάτων με διαφορετικό προσανατολισμό.*

***Λέξεις-κλειδιά:*** *λεκάνη της Κρανιάς, Οφιόλιθοι Πίνδου, χάρτης 1:10.000, ιζηματολογία, τεκτονική γεωλογία*

## 1. INTRODUCTION

This paper presents a new 1:10,000 scale map of the northwestern Krania Basin (KB). Previous mapping in the area has been done at a scale of 1:50,000 (Brunn, 1956; Koumantakis and Matarangas, 1979) and in some areas at 1:20,000 (Rassios and Grivas, 1999). The study area of 30 km$^2$ extends from the Tripimeni Petra in the southwest to the Venetikos River in the northwest, encompassing the villages of Mikrolivado and

Monachiti. Exposure is generally good in road cuts, rivers and dry streams, and on steep ground, but poor in forests and on flat ground. The Universal Transverse Mercator coordinate system (Zone 34S) is used throughout this paper.

The Mesohellenic Basin (MHB) is an intra-montane Oligocene – Miocene sedimentary basin that occupies central northern Greece and eastern Albania separating the Pindos Zone to the west and the Pelagonian Zone to the east. Sediment infill of the MHB is referred to as Mesohellenic Sediment (MHS) in this work. The Western Hellenic Ophiolite Belt outcrops around the margins of the MHB and the KB and is thought to be largely continuous in the subsurface (Rassios and Moores, 2006). It consists of Jurassic oceanic lithosphere which was initially emplaced to the northeast during the Jurassic (Robertson 2012) and underwent subsequent southwest verging backthrusting during the Eocene, thrusting the ophiolite over the Pindos Zone (Rassios et al., 1994). The KB, which contains the study area, is Eocene in age and located on the western margin of the MHB, separated from it by the dextral Eptahori Fault (Zelilidis et al., 2002).

Both the MHB and the KB have been described either as pull apart, transtensional basins (Zelilidis et al., 2002) or as piggyback, forearc basins (Ferrière et al., 2013). The formation of the KB in the Middle – Late Eocene is contemporaneous with the final emplacement of the ophiolite over the Pindos Zone. The Eocene strata were uplifted and deformed during and shortly after their deposition, generating reverse faults with a northeasterly vergence within the KB. The MHB developed in the Oligocene and was associated with NW – SE trending dextral faulting that also affects the KB. The final phase of deformation affecting the region is high angle normal faulting of variable strike, from the Late Miocene to the present (Kilias et al., 2015).

## 2. STRATIGRAPHY

The ophiolite stratigraphy adopted is that outlined in Jones and Robertson (1991), with tectonic disruption resulting in the outcrop patterns observed. The Krania Formation has previously been studied (e.g. Ferrière et al., 2004, Kilias et al., 2015) and mapped (Brunn, 1956) as one unit comprising basal conglomerates and overlying sands and shales. Here, we subdivide the formation into six distinct units – the Basal Conglomerate, Tripimeni Conglomerate, Limestone Conglomerate, the Lower and Upper Krania Members and a Limestone Breccia - and map their spatial extent. Though highly variable lateral thicknesses and faulting induce complex field relationships, we find the Basal Conglomerate to be the stratigraphically lowest unit in the Krania

Formation. Synchronous deposition of the Tripimeni Conglomerate in geographically distinct regions is probable, but the stratigraphy we suggest is consistent with every instance where the two conglomerates are in contact. Such contacts are rare and limited in extent: the conglomerates regularly pinch out and thicken. This is best expressed by the stratigraphic column in Fig. 1. The Upper and Lower Krania members are separated by an unconformity.

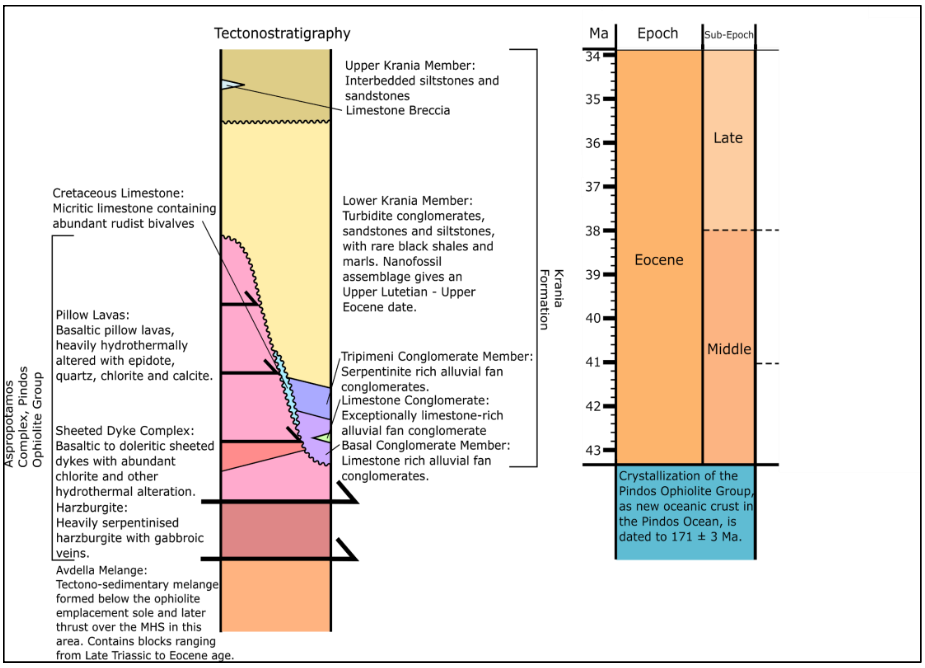


**Fig. 1:** Stratigraphic column of the mapped area.

## 3. LITHOLOGICAL DESCRIPTIONS

The descriptions in this section outline the superficial features of the units which allowed for their identification in the field. For detailed discussion of the Pindos Ophiolite, see Rassios (2011) and references therein, and the Avdella Melange is described in Ghikas et al. (2010). Detailed sedimentological descriptions of the Krania Formation are given in section 4, and the palaeontology of the Cretaceous Limestone is described in section 5.

### 3.1. Pindos Ophiolite

The ophiolite in this area comprised Harzburgite, Sheeted Dykes and Pillow Lavas.

**Harzburgite**

The estimated thickness of this unit is 200 m: its base is not visible in the area. In outcrop this unit weathers a distinctive green-brown colour and is composed of 30 % pyroxene, up to 10 % plagioclase needles and 60 % olivine. Weathering and serpentinisation are extensive. Intruding it are centimetre scale veins of paler gabbroic material.

**Sheeted Dykes**

This unit is faulted at its base, but grades into the Pillow Lavas above. Its maximum thickness in the area is about 60 m. It is a porphyritic basalt with grain-size variations that are systematic within metre-width dykes defined by fine-grained margins and coarser interiors.

**Pillow Lavas**

This lithology is estimated to be around 300 m thick in the area. Identified by lobate, rounded shapes with fissile chilled margins, it is sparsely phyric, containing occasional acicular plagioclase phenocrysts. Abundant amygdales are infilled by calcite and zeolite.

### 3.2. Avdella Melange

This lithology is estimated to be around 300 m thick in the area. Identified by lobate, rounded shapes with fissile chilled margins, it is sparsely phyric, containing occasional acicular plagioclase phenocrysts. Abundant amygdales are infilled by calcite and zeolite.

### 3.3. Cretaceous Limestone

This unit outcrops as an array of crags on a scale ranging from 100 to 200 m. Layers of poorly lithified breccia are also found, composed almost entirely of limestone. The limestone is generally micritic, with some regions of recrystallised sparite, and bedding is notably absent. Rudist bivalve fossils were evident in hand specimen, and in thin section foraminifera and bivalve bioclasts were abundant.

### 3.4. Krania Formation

**Basal Conglomerate**

Mostly clast supported (as shown in Fig. 2), the unit has a reddish-brown colour. It consists of poorly sorted, 1 - 40 cm diameter clasts, on average 60 % rounded limestone, 40 % sub-rounded igneous clasts (basalt, serpentinite and diorite) and trace amounts of chert, set within a coarse sand matrix of similar composition. Imbrication of clasts allowed bedding and paleoflow to be estimated.

**Fig. 2:** Basal Conglomerate showing imbrication of clasts. The discontinuity in clast size is interpreted as a bedding plane. Here, the conglomerate is clast supported.

**Tripimeni Conglomerate**

This conglomerate is heavily faulted, with a maximum thickness of around 600 m tentatively inferred. The unit is a clast supported, poorly sorted conglomerate and breccia: mostly dark blue in colour, it also contains white and green clasts. It is best distinguished where a coarse sandy matrix is exposed. Composed of 90 % ophiolitic and serpentinite clasts with 10 % sub-rounded limestone grains, these regions pass gradationally into crystalline regions formed of anhedral macroscopic serpentinite crystals. Other clasts include blocks of serpentinised harzburgite, 10s of metres across, while diorite, granite and basalt cobbles are exposed. Likely formed by hydrous alteration, regions of homogeneous serpentinite are commonly observed, distinguished from ophiolitic units by rare 1 - 10 cm diameter rounded limestone clasts.

**Fig. 3:** Tripimeni Conglomerate exposed in river bed. Large, angular, mafic igneous clasts are visible.

**Limestone Conglomerate**

The Limestone Conglomerate is exclusively found within the Basal Conglomerate with a maximum thickness of 45 m. It is composed of greater than 60 % angular limestone clasts up to 50 cm in size, plus a few percent rounded chert clasts and around 10 % rounded mafic clasts when present. The lithology is clast supported, with the lime-rich sandy matrix comprising only 10 % of the unit.

**Lower Krania Member**

The unit consists of conglomerates, sandstones and siltstones, deposited onto the Basal and Tripimeni conglomerates in places and directly onto the ophiolite elsewhere. The unit has a minimum thickness of 800 m and bed thicknesses are typically decimetre- or metre- scale. Sediment is a poorly sorted mixture of 50 % sub angular limestone clasts, 35 % rounded olivine grains and 15 % sub-rounded diorite or amphibolite clasts. Trace amounts of chert are also present. *Gordia* and *Paleophycus* trace fossils are present though extremely rare, while organic carbon is more abundantly preserved as coal. Fissile shale is present throughout in 5 cm thick beds. Asymmetric folding and slumping of sediments was evident throughout the area except in the far west.

**Limestone Breccia**

A 7 m thick breccia bed, which is very rich in 1 - 10 cm limestone clasts that were weathered during deposition, was mapped near the top of the Upper Krania Member.

**Upper Krania Member**

Greater than 600 m thick, this unit outcrops as thickly bedded sandstone or cyclically interbedded sandstone and siltstone: the gravel content in ordinary beds is much lower than in the Lower Krania Member.

## 4. SEDIMENTOLOGY

### 4.1. Basement and source rocks

The Krania Formation in this area is deposited directly onto ophiolitic basement. The unconformable and depositional nature of this contact is displayed in the Venetikos river, north of Mikrolivado, where both Basal Conglomerate and Lower Krania Member are seen in contact with ophiolite. The contact is highly non-planar due to erosion of the basement prior to deposition, though the absence of paleosol indicates that no significant pre-depositional weathering of the ophiolite is preserved. The bulk composition of the six members strongly suggests that local ophiolitic and limestone units constitute the source rocks for the formation, which is supported by measured paleoflow directions. With an absence of flows to the west, the dataset reveals source from the basin margins, as displayed in Fig. 4. Furthermore, a separate dataset shows northward flow in the southern half of the area and predominantly southward flow in the northern half of the area, with a small area in between with mixed flow directions.

### 4.2. Depositional environment

**Basal Conglomerate**

The sedimentology of the Basal Conglomerate is consistent with deposition in alluvial fans, supported by its highly variable lateral extent. Furthermore, its poor sorting, clast supported nature and distinctive horizontal stratification are most consistent with deposition by a sheet flood mechanism. Clasts are generally rounded and the matrix coarse, suggesting significant abrasion during transport to the fan. Finally, cement type within the conglomerate is dominantly fringing (Fig. 5), which implies deposition in the phreatic zone.

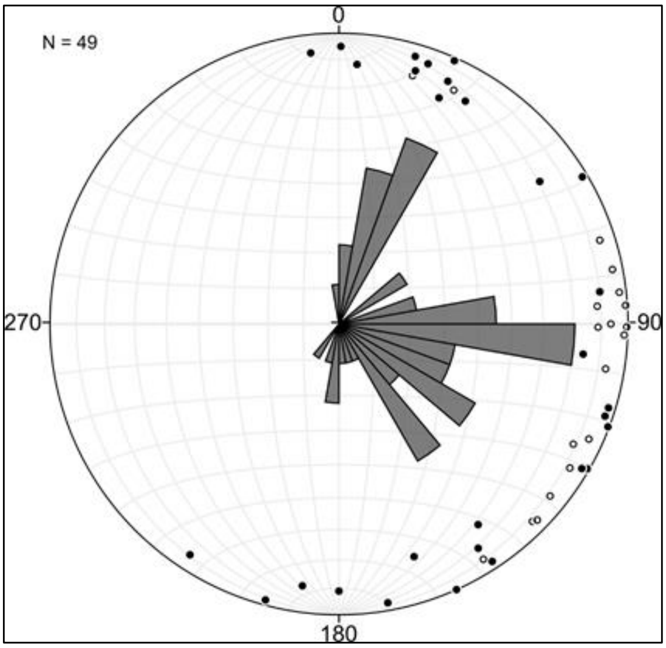


**Fig. 4:** Paleoflow measurements obtained from within the Krania Basin (n= 49). White circles indicate negative plunge.

**Limestone Conglomerate**

The Limestone Conglomerate is deposited exclusively within the Basal Conglomerate and shares many of the same characteristics: it's depositional environment and mechanism are likely the same. However, it is exceptionally limestone-rich, interpreted as resulting from local variations in source rock composition.

**Tripimeni Conglomerate**

The outcrop pattern and highly variable lateral thickness of the Tripimeni Conglomerate is again consistent with deposition in alluvial fans. It is poorly sorted and clast supported, but in most cases lacks observable bedding or other sedimentary structures. In rare cases, however, there is clear evidence for a coarse sand matrix surrounding clasts of varying size (centimetres to metres) and angularity. Though later serpentinisation makes identification of the presence or absence of primary diagnostic features difficult, we favour deposition dominantly by a debris flow mechanism, based on the absence of horizontal stratification which is by comparison readily observed in the Basal and Limestone Conglomerates. The three compositionally distinct lithologies are likely a consequence of local variations in bedrock. The Basal and Limestone Conglomerates are likely derived from marine limestones and cherts which outcrop extensively in the east of the area, while the Tripimeni Conglomerate is sourced from ophiolitic units. The observation that the Basal Conglomerate underlies the Tripimeni Conglomerate may be a consequence of the need to remove overlying limestones and cherts before erosion of the ophiolite can begin. Spatially variable deposition of the conglomerates leads to deposition of the Krania Formation directly onto Ophiolitic basement in some locations.

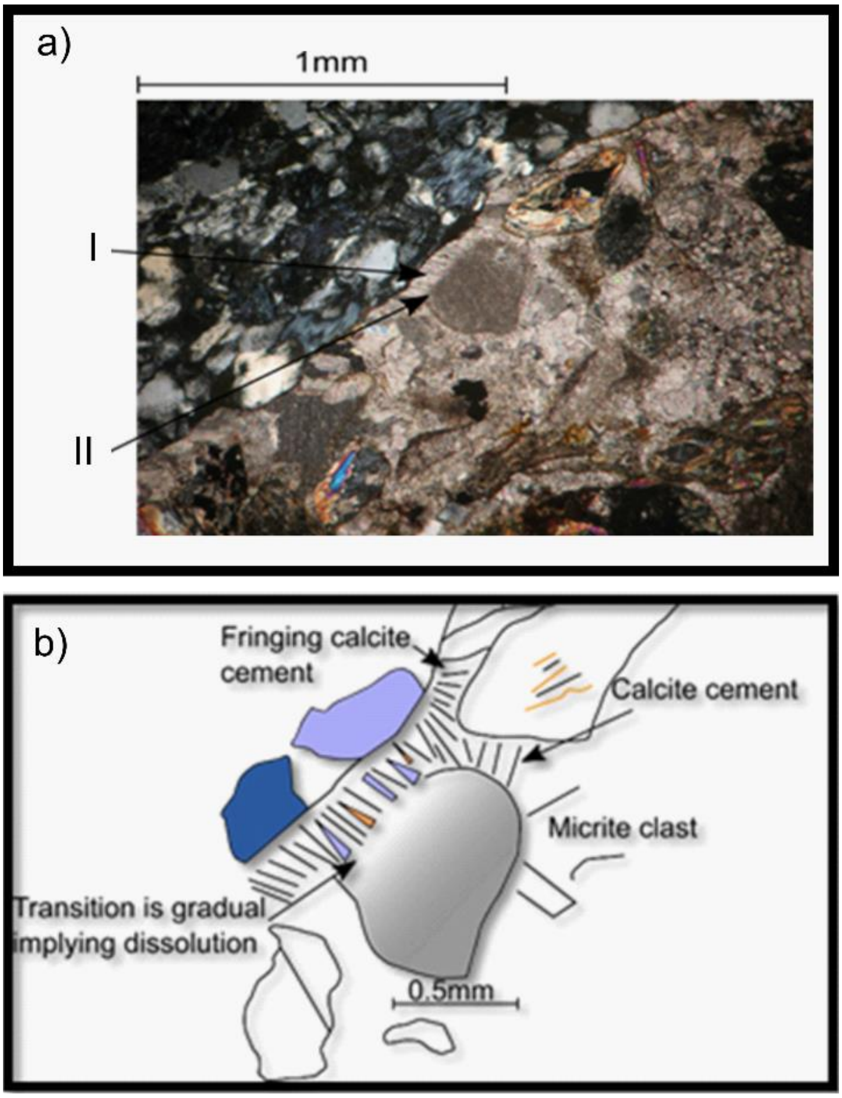


**Fig. 5:** a) Thin section photo showing the fringing cement texture observed in the Basal Conglomerate I) Calcite cement fringing a micrite clast. II) Transition is gradual implying locally-sourced cement by clast dissolution. b) interpretive sketch of a).

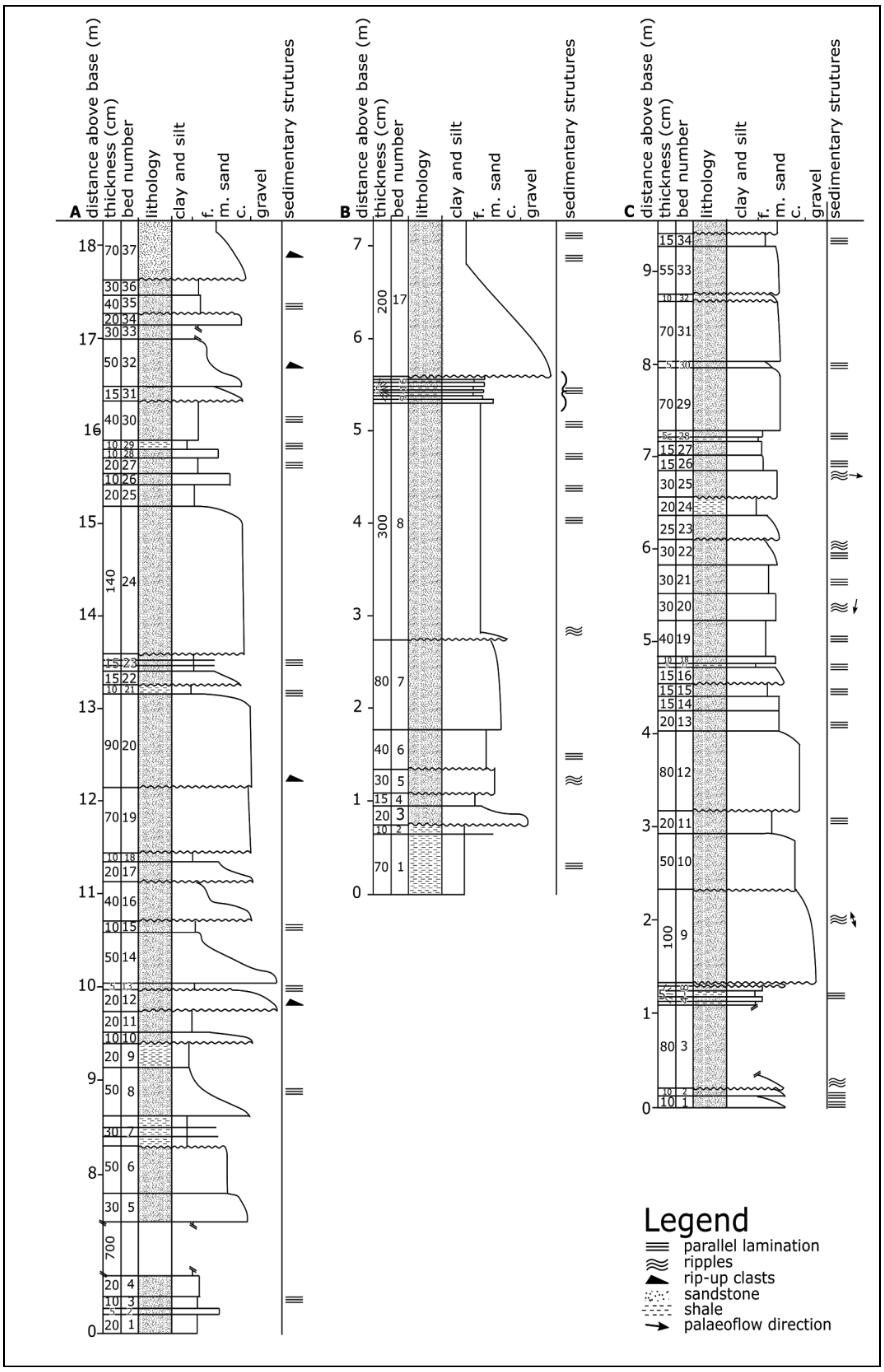


**Fig. 6:** Sedimentary logs of the Lower Krania Member. Left to right represents west to east, with localities at (0520514,4423080), (0521112,4420079) and (0522105,4422570). These are labelled A, B and C respectively.

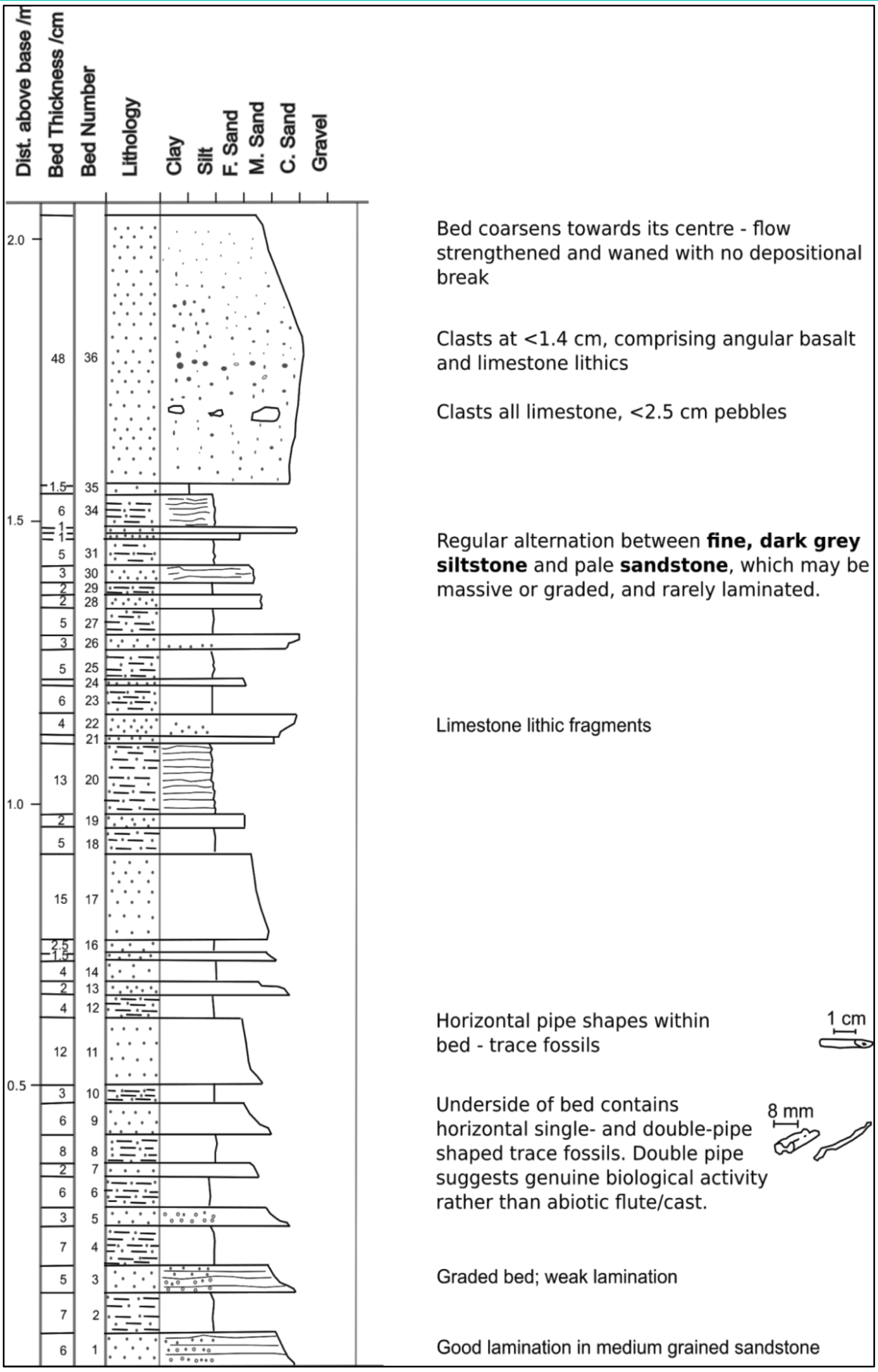


**Fig. 7:** Sedimentary log of the Upper Krania Member at (0523765, 4424061).

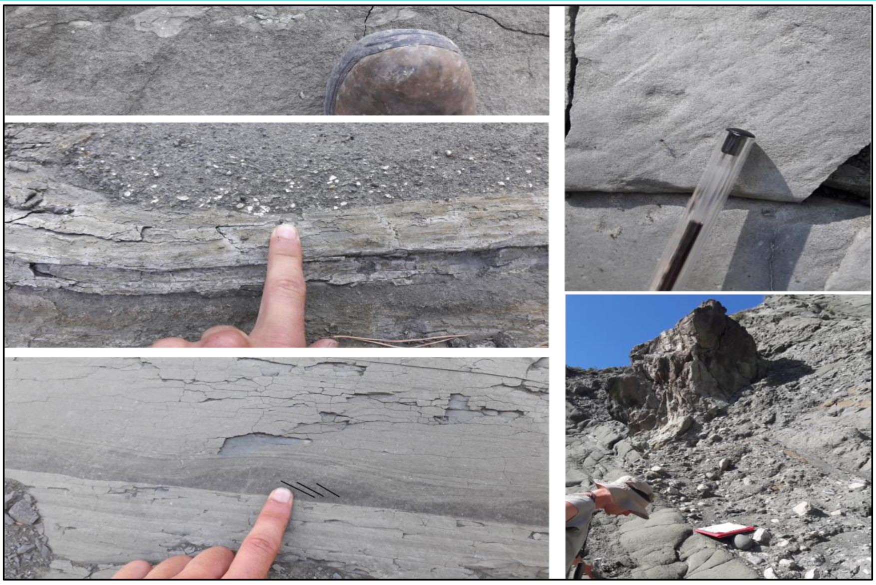

**Fig. 8:** Photos labelled down columns (a-c on left, d-e on right). In photos a-c, younging is up the page. Photo d is looking onto a bedding plane, and in e younging is to the top right. a) Randomly oriented rip-up clasts in medium sand. b) Overlying coarser bed scouring into fine sandstone, truncating laminations. c) Parallel laminated fine sand overlain by medium sand showing asymmetric ripples with highlighted cross lamination. d) Flute casts preserved in negative epirelief. e) Breccia bed in the Lower Krania Member containing a large olistolithic block 5mx3m in diameter.

### Krania Members

Zelilidis et al. (2002) recognise sands and shales at the base of the Krania Members, while our logging shows that the sediments of the Lower Krania Member are turbiditic, containing normally-graded beds displaying scoured bases and rip-up clasts (Figs. 6, 7, 8). In the Lower Krania Member, ripples are observed with rounded crests and convex-up foresets (Fig. 8), while hummocky cross-stratification (HCS) is also observed. This is consistent with deposition above storm wave-base, with ripples forming under combined unidirectional flow and wave-influenced conditions (Lamb et al., 2008). The absence of symmetric ripples implies post-depositional reworking of sediment by waves did not occur (Lamb et al., 2008), constraining water depth to greater than fair weather wave-base. Systematic changes in source proximity and water depth are difficult to assess, with local position relative to the basin margin the dominant factor. Away from the basin margin, gravel content certainly decreases, and bed thicknesses were observed to decrease somewhat. We emphasise that a greater sampling density would be required to separate the effects of putative changes in water level or simple proximity to the basin margin.

The east of the area displays outcrops of alternating sand and shale beds showing only sporadic grading (Fig. 7) and no ripples. Mapped as the Upper Krania Member, the unit lies above an unconformity surface identified in the field and can be distinguished by its low gravel content. Though deltaic facies have been inferred in the east of the Krania formation (Zelilidis et al., 2002), we cannot confidently assert this here. It is worth noting that such an interpretation is consistent with increasing source proximity with stratigraphic height. Both Krania members contain debris flow deposits (Fig. 8), with metre-sized blocks of chert, serpentinite, limestone and basalt within a coarse sand matrix. These represent intervals of high-energy deposition.

We find no evidence for salinity indicators in the west of the area. Putative chondrites burrows further east, however, imply the basin had - at least at points - a marine influence. Chondrites has been used to infer anoxic conditions in sediment pore waters (Bromley and Ekdale, 1984), which is consistent with abundant pyrite found in the Lower Krania Member (Fig. 9). Anoxic conditions in sediment pore waters may have arisen due to a high supply of organic matter from the water column. Pyrite is generally found with a circular cross section and an elongate expression parallel to bedding, possibly forming close to decaying organic matter and filling pre-existing burrows. Indeed, *Gordia* and *Paleophycus* traces are noted in the Lower Krania Member (Fig. 9), but elsewhere body and trace fossils are rare. This is expected in a turbidite succession, with the top layer of sediment eroded during the following events. In addition, rapid successive deposition will not give time for benthic communities to develop to their climax.

We suggest the following: the Krania Basin was initially filled by alluvial fan deposits which take distinct compositions controlled by source rocks. Subsequent turbidite deposition formed the Lower Krania Member, separated from the distinct Upper Krania Formation by an unconformity surface. The Krania Members may have been deposited in a marine setting, though this is far from confirmed and deposition is almost entirely consistent with both lacustrine and marine conditions.

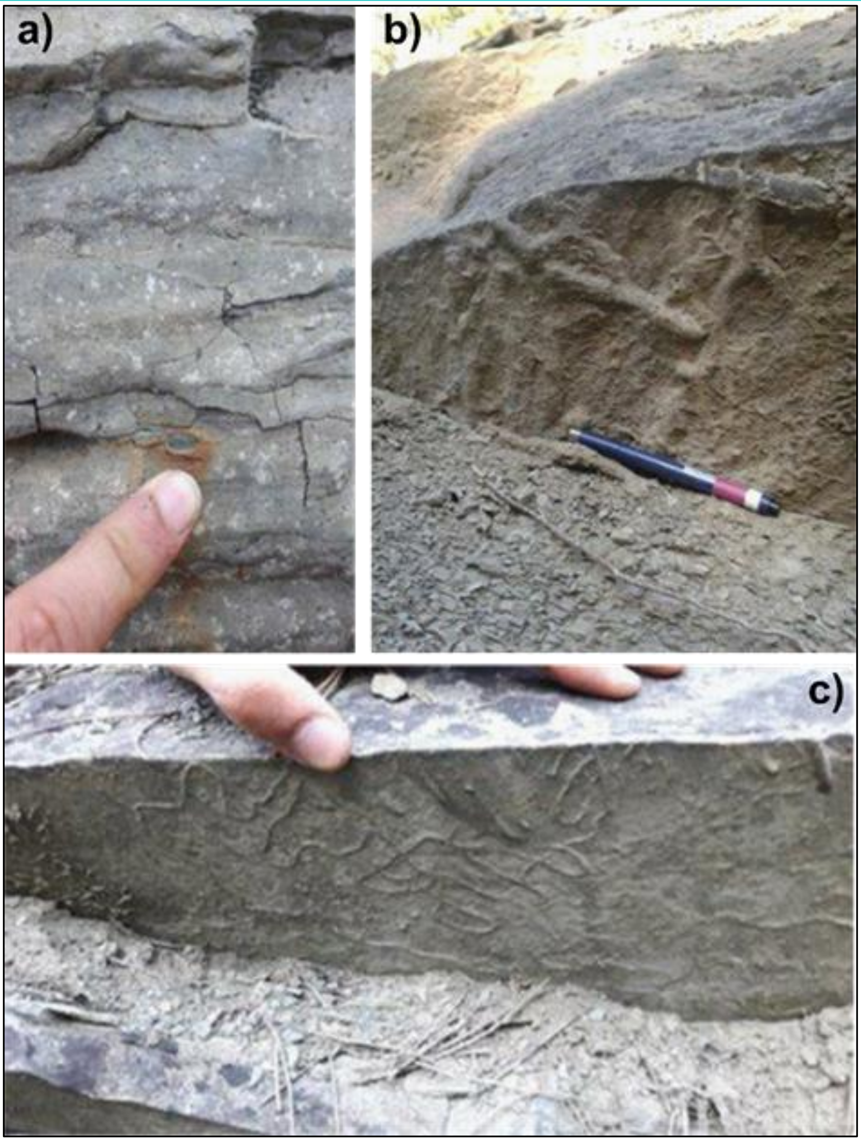


**Fig. 9:** a) Finger points to circular cross-section of pyrite nodule, facing direction is down dip of bedding. b) Paleophycus burrows in positive hyporelief. c) Convolute feeding trace - Gordia - preserved in positive hyporelief. All photos from the Lower Krania Member.

## 5. CRETACEOUS LIMESTONE: PALAEONTOLOGY

The Cretaceous Limestone contains abundant fossils. Colonies of reef-forming rudist bivalves are observed with individuals ranging in diameter from 1 to 5 cm. At least some of the rudists found within the limestone (Fig. 10) are of the genus *Milovanovicia*, from the family Radiolitidae, dating the limestone as Cretaceous, and more precisely, Turonian - Santonian (Dr Peter Skelton, pers. comm. 2017). Also present are gastropods, foraminifera, echinoderms, red algae, and oysters. Some of the foraminifera are benthic Miliolid foraminifera, restricted to inner ramps and platforms (Flügel, 2004). Furthermore, the limestone contains at least two genera of red algae: *Solenopora* and *Sporolithon*. Since these red algae photosynthesise, they must have lived within the photic zone, constraining water depth to less than 50 m (Dr Elizabeth Harper, pers. comm. 2017). We conclude that the Cretaceous Limestone was deposited as a platform carbonate.


**Fig. 10:** top) transverse section through a Milovanovica, Radiolitidae which has undergone partial diagenetic alteration to ferroan dolomite. Bottom) cross-section of a star-shaped Milovanovica, Radiolitidae (Sari and Ozer 2009).

## 6. STRUCTURE

### 6.1. Pindos Ophiolite

The ophiolite is dominated by four major faults, associated with two different faulting events.

The first event is recorded in northward-dipping faults which are well exposed in the Venetikos River north of Mikrolivado where clear fault planes are seen with a fault gouge zone up to 70 cm wide in the core. Extrapolation of these structures away from the river has been necessary due to poor exposure, constrained using regions of heavy brecciation. Shear indicators show shearing with a top to the south or southwest sense, suggesting a thrust motion on the major faults and implying they were originally westward-dipping before being rotated.

This thrust faulting is likely related to the Jurassic or Cretaceous emplacement of the ophiolite which is generally identified to be northeast verging but with a degree of variability (Rassios and Moores, 2006). The faults are inferred as meeting at depth to give blocks forming an imbricated thrust sheet which also contain smaller scale imbrication.

The close proximity of Pillow Lavas, Sheeted Dykes and Harzburgite around Mikrolivado shows that a large amount of telescoping of the original sequence has occurred, presumably during Jurassic-Cretaceous obduction. Prolific minor faulting with variable fault orientations is found in the Pillow Lavas (Fig. 11) reflecting the complicated kinematics of ophiolite emplacement. The mean orientation of these minor faults is close to the orientation of the major faults, however, suggesting that they were caused by the same tectonic event. Furthermore, a large southwest dipping thrust fault is exposed in the Venetikos river to the northeast of the region on which various exotic lithologies have been exhumed.

The second event in the ophiolite may be inferred from the fault separating the Harzburgite from the Pillow Lavas. This is now approximately oriented at 335/90 and rotates back to 334/50 when the regional tilting is considered. The fault must be contemporaneous with, or younger than, the Tripimeni and Basal Conglomerates since small blocks of these lithologies are found within the fault zone to the southeast of the Tripimeni Petra and in the same area there is evidence of top to the south shear with a right-lateral component suggestive of dextral strike-slip faulting with a thrust component. Further south, near the village of Krania, the contact between the Krania Formation and the ophiolite has been suggested to be a north-northwest to south-southeast trending dextral thrust fault of Mid – Late Eocene age (Kilias et al., 2015) putting a possible age on this faulting.

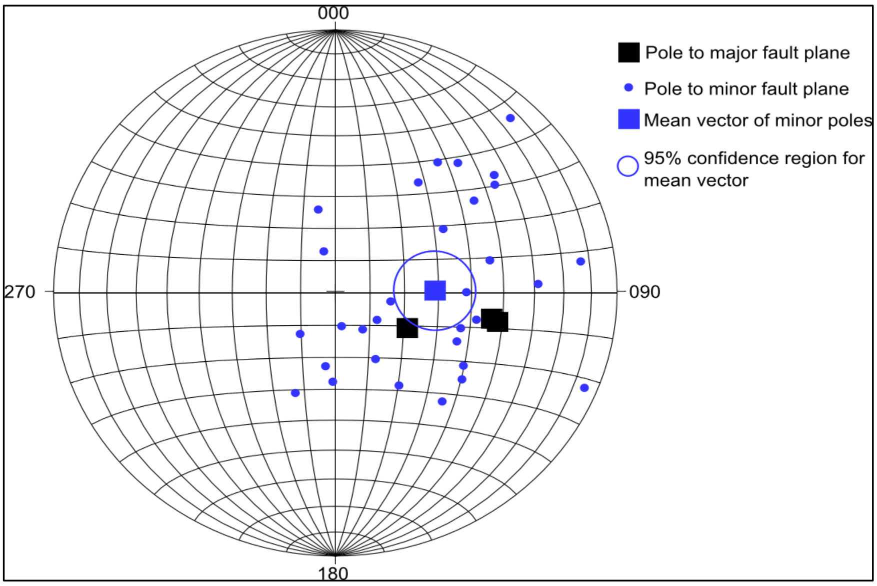


**Fig. 11:** Fault planes in the Pillow Lavas. The planes have been rotated anticlockwise by 40° around a 00/340 axis to correct for the tilting of the whole area which is inferred from the Krania Members. The mean orientation of the minor faults is similar to that of the major faults, which supports them being cogenetic.

The Sheeted Dykes, originally emplaced vertically, offer the chance to determine the original orientation of the ophiolite. Once corrected for later regional rotation, the primary orientation of the dykes is around 050/85 (shown in Fig. 12), meaning that there has been very little rotation around an axis parallel to the plane of the dykes. A smaller cluster of sub-horizontal intrusions may be interpreted as sills or dykes in a secondary orientation, though the latter interpretation would require a rotation of 90° about an axis perpendicular to the plane of original dyke emplacement. It should be noted that all of the dyke measurements are from only one major thrust block and so may not be truly representative of the ophiolite as a whole.

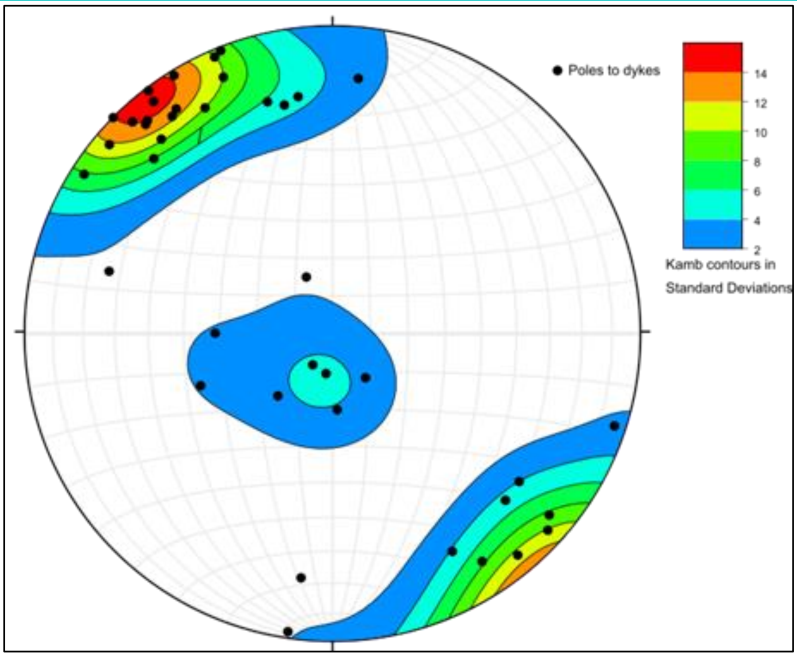


**Fig. 12:** Poles to sheeted dykes, rotated anticlockwise by 40° around a 00/340 axis to correct for the tilting of the whole area which is inferred from the Krania Members. There is a large cluster of sub-vertical dykes (around 050/85) and a smaller cluster of sub-horizontal intrusions.

### 6.2. Edge of Basin: West of Mikrolivado

The western margin of the basin, which is generally comprised of parallel beds rotated to 40°, is affected in the south of the area by two large north-northwest striking reverse faults which cut the sediments and the ophiolite. These are exposed to the northeast of the Tripimeni Petra, one in a steep gorge of Tripimeni Conglomerate and the other as a heavily brecciated band of the Lower Krania Member further downstream the Tripimeni river. Their displacements are estimated to be of the order of 75 m and 100 m respectively. The downstream-fault was inferred by the authors as the same fault which leads to overturned bedding approximately 1 km to the north.

Evidence of further reverse faulting can be found in the Lower Krania Member. A monocline of wavelength ca. 5m was found in the Tripimeni river downstream of the Krania road, for example. Decametric reverse faulting with a northeasterly vergence has been reported in this area of the KB before, and attributed to Late Eocene basin closure (Ferrière et al., 2013; Kilias et al., 2015): we attribute this faulting to this event also.

### 6.3. Edge of Basin: Ag. Demetreos and Ag. Nikolaos

Several phases of deformation have led to the formation of structures which can be seen in the Venetikos River valley north of Mikrolivado. The edge of the basin runs through the valley with depositional contacts of the Krania Formation onto the Pindos Ophiolite found to be parallel to sediment bedding striking at between 000° and 025°. The bedding steepens towards the basin edge to almost vertical at the steep cliffs of the Blades of Ag. Nikolaos. A pair of north-northwest striking low-angle Eocene thrust faults (Vamvaka et al., 2006) with a 30° dip are exposed on the eastern hillside, with the lower inferred as being truncated by the upper around Ag. Demetreos. These thrusts place the stratigraphically lower conglomerates above the Lower Krania Members and are seen to continue north beyond the Blades of Ag. Nikolaos where Melange is being thrust over the conglomerate units (depicted in Fig. 13).

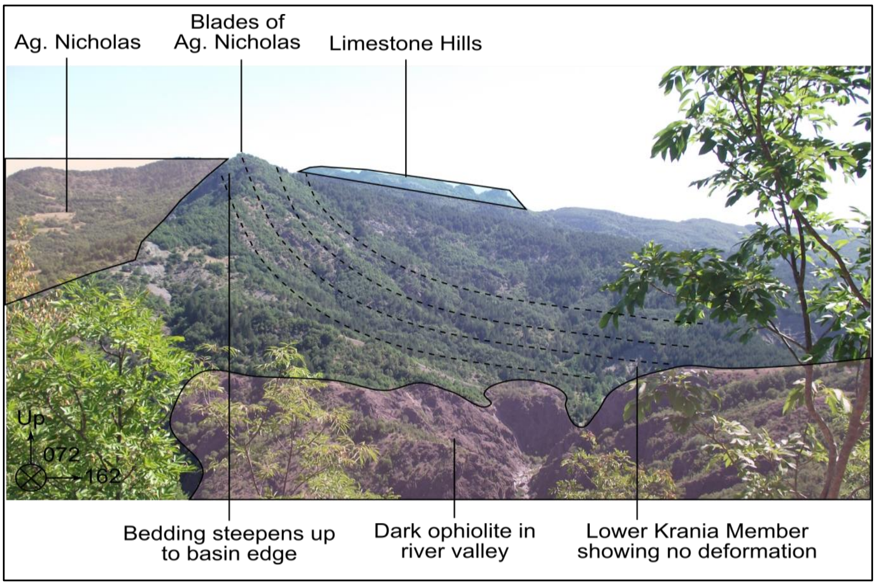


**Fig. 13:** A landscape photograph looking East at the Blades of Ag. Nikolaos and the Limestone Hills, with the Venetikos river in the valley, showing the shallowing of the bedding into the mapping area and the relationship of the bedding, limestone unit and ophiolite units at the basin edge.

These thrusts are offset by a series of three high-angle southwest-striking normal faults, dated as post-Late Miocene in line with parallel late-stage faulting reported across the MHB by Vamvaka et al. (2006). The northernmost of these can be observed on the hillside where Melange is separated from Tripimeni Conglomerate by a steep gully, confirming the high-angle orientation of the faults also inferred from their outcrop pattern.

To the north of these southwest-striking faults, Melange is seen to be faulted over the Krania Formation, as part of the Eocene thrusting regime. The fault is observed at the southwestern end of the Blades of Ag. Nikolaos as being in approximately the same orientation as the hillside, with two windows through the Melange seen. The nature of the northernmost high-angle fault appears to change: at the river it is seen as a high-angle fault whereas at the northeastern end of the blades a fault oriented at 044/36 can be seen. At the location where the two orientations meet, a large area of very heavily brecciated serpentinite is seen which is approximately 200 $m^2$ in size.

The Blades of Ag. Nikolaos are made up of steeply dipping Basal and Limestone Conglomerates. The contact between the Melange and the Tripimeni Conglomerate at the southwest end of the blades is seen to be faulted, and its orientation suggests that it is a continuation of the upper north-northwest striking fault. This contact, however, is striking south which may be due to the east-plunging folding.

### 6.4. Folding in the Ag. Demetrios area

East-plunging folding inferred across the region is supported by several observations. Changes in the bedding of the conglomerates, especially clear on the Blades of Ag. Nikolaos, (Fig. 14) suggest a fold axis orientation of approximately 32/074. Furthermore, structure contours of the southern Melange fault (truncated by another fault at (0523000, 4424800)) suggest a folded fault. The folding appears to be less pronounced moving into the conglomerate units, potentially because they are better lithified and so harder to fold. Similarly, no evidence for the north-northwest striking faulting is seen in the Lower Krania Member suggesting that a folding-based deformation mechanism becomes dominant, with minor faulting and parasitic folding seen in this poorly lithified sediment. Finally, the parasitic folding observed on the Monachiti road was found to systematically change in orientation (Fig. 15) around the folding proposed from the other evidence put forward. We attribute this folding to the same Late Eocene compressional event as the thrust faults which it folds.

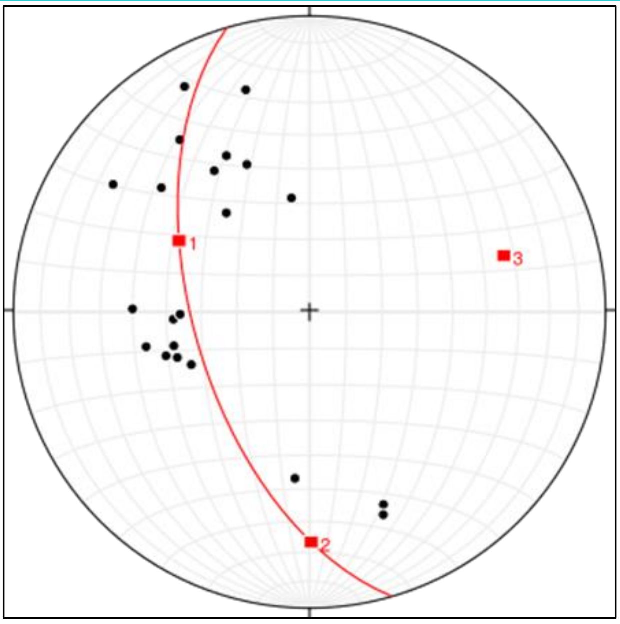


**Fig. 14:** Stereographic projection showing poles to bedding planes around Ag. Demetreos (black). Bingham analysis (red) suggests an approximate fold axis orientation of 32/074.

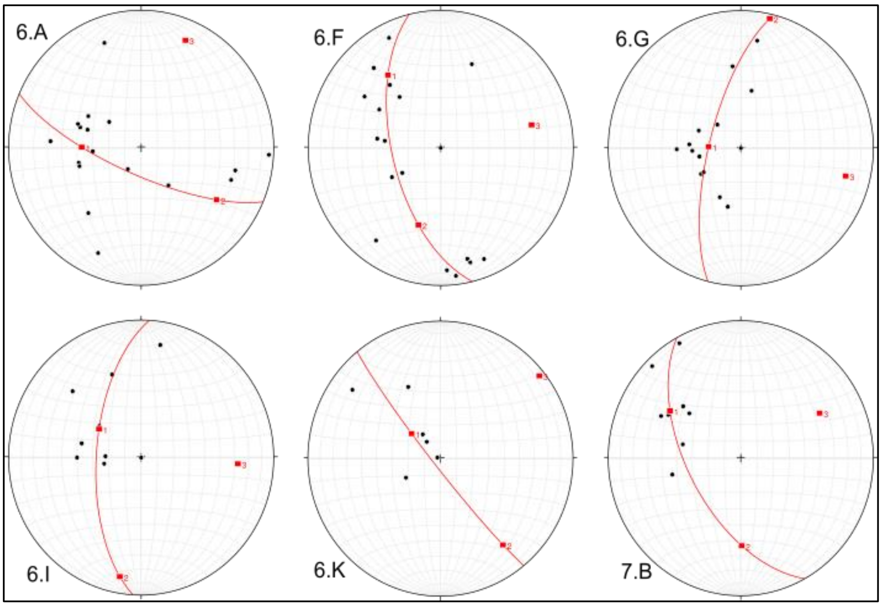


**Fig. 15:** Stereographic projections showing the poles to bedding in the Lower Krania Member (black) at six folded outcrops on the road to Monachiti. Bingham analysis (red) at each locality shows variations in the orientations of the fold axes. The localities are numbered South-West to North-East, with 6.A closest to Mikrolivado and 7.B closest to Monachiti. The Krania Members young northeast.

The folding is only observed to the north of the 4422 grid line. Some deformation in this region is observed to be syndepositional, with some clearly syndepositional faults observed in the roadcuts on the Mikrolivado-Monachiti road.

**6.5. Ag. Demetrios**

The region around Ag. Demetrios (0521600, 4423700) shows a non-planar outcrop pattern. Three potential hypotheses tested for this include northeast-southwest trending folding, faulting of the contact and deposition of the Lower Krania Member onto a non-planar surface of Basal Conglomerate. This last hypothesis is most likely: folding is unlikely since there is no folding observed in this orientation despite almost complete cross-sectional exposure of pristine cliffs of the Lower Krania Member, while no signs of faulting were observed in the area.

**6.6. Limestone Hills**

Many outcrops of Cretaceous limestone can be found between Mikrolivado and Monachiti (e.g. 0522000, 4423900), some of which cause topographic highs due to their relatively high resistance to erosion. There are several suggestions for how the limestone outcrops are related to the surrounding Lower Krania Member including: faulting after Lower Krania Member deposition; deposition of the Lower Krania Member amongst the limestone unit which was deposited previously onto the ophiolite; and each limestone hill comprising of a single olistolith which has fallen into the Lower Krania Member during its deposition (Jones and Robertson, 1991). There is evidence for faulting in other Cretaceous limestones of the MHB, where Mesohellenic Sediment bedding steepens to near vertical near the Limestone (e.g. the Portitsa Gorge; Rassios, 2011).

However, this is not seen in our area suggesting the limestone may not be related to the Lower Krania Member by faulting. The contact seen to the east of Ag. Demetrios contains clasts of several lithologies suggesting that deposition of these lithologies occurred amongst the limestone. Multiple outcrops also reveal that the limestone bedding lies parallel to bedding in the Lower Krania Member. Furthermore, structure contours of the limestone/Lower Krania Member contacts are parallel to this. The consistent alignment of the Lower Krania Member and limestone bedding suggests that the limestones making up the hilltops are unlikely to be olistoliths, since limestone bedding would likely be more random in this case. We therefore propose that the Lower

Krania Member is deposited onto the eroded surface of a limestone unit which is itself deposited onto the ophiolite basement.

**6.7. Eptachori Fault**

The northwest-southeast striking serpentinite sliver in the northeast is parallel to the major dextral strike slip faults in the region, and is likely part of the Eptachori fault (Kilias et al., 2017). The serpentinite contacts both dip at around 30º, shallowed from the expected steep attitude of regional strike slip faulting. Slickencryst lineations on fault surfaces give major strike-slip and minor dip-slip components, while smaller dextral transfer faults within the area imply a dextral sense, in agreement with previous work on the Eptachori fault.

There is a bimodal distribution of slickencryst lineation directions, indicating two principal stress directions. The northwest-southeast lineations are consistent with the strike of the serpentinite outcrop and with the long axes of the prolate serpentinite shear pods, all suggesting the same strike-parallel slip.

**6.8. Basin margin faulting in the north of the area**

Reverse faulting of Pillow Lavas over Basal Conglomerate is seen striking southwest (north of Monachiti) and southeast (1 km further east, between Monachiti and Trikomo). These faults postdate the Lower Krania Member so are attributed to the Late Eocene compressional phase recognised across the basin.

The faults within the Krania Formation group with a steep north-northwest to south-southeast strike, both consistent with Vamvaka et al. (2006) and parallel to the basin opening direction. This, and the scarcity of northwest dipping faults, suggests that northwest-southeast extensional tectonics occurred by soft sediment deformation and transfer faulting.

Also notable is the scarcity of northwest dipping faults, which points to a lack of northwest-southeast oriented dip-slip events and thus suggests a strike slip northern margin to the basin. These observations fit the northwest-southeast dextral normal (Wilson, 1993) fault scarp at Orliakas.

## 7. GEOLOGICAL HISTORY

Late Triassic rifting of the Apulian and Pelagionian microcontinents in the Palaeotethys Ocean produced oceanic crust, forming the Pindos Ocean (Ghikas et al., 2010). Subsequently, west-dipping Jurassic subduction within the Pindos Ocean brought together these microcontinents (Robertson, 2012). Slab rollback caused back-arc spreading and assimilation of supra-subduction zone trace element signatures comparable to most present-day island-arc tholeiites (Smith and Rassios, 2003; Capedri et al., 1980). During subduction, offscraping and tectonism formed the Avdella Melange in the accretionary prism, a subduction melange in the sense of Festa et al. (2010).

Eastward emplacement of the oceanic crust and accretionary prism onto Pelagonia (Jones and Robertson, 1991) occurred during the Jurassic, evidenced by the now northward-dipping faults observed in this study within the Pindos Ophiolite Group (Fig. 16.1). In the Cretaceous the northern Pindos Ocean was fully sutured, stalling subduction and leaving a small basin in the south for deposition of the Orliakas Group limestone (Fig. 16.2). In the early Cenozoic the Krania Basin opened (Fig. 16.3, 16.4) while the Pindos Ocean continued to subduct. Accommodation space was created during this phase of compression by either crustal flexure to form a piggyback basin, or strike-slip faulting to give a pull-apart basin (Zelilidis et al., 2002). Syndepositional normal faulting is seen within the sediments but it is not exclusively diagnostic of either mechanism for basin formation. The eastward palaeoflow directions of this study indicate a palaeotopographic high immediately to the west of the mapped area and the actively tectonic, scarped landscape present during deposition of the basin fill would have provided ample sources for the large limestone blocks observed around Monachiti.

Our study identified a series of four post-depositional tectonic events, consistent with four of the five phases proposed by Vamvaka et al. (2006). Fault relationships observed in the field confirmed relative timings of each phase. Phase one was caused by broadly east-west oriented compression, consistent with continued basin closure in the Late Eocene. Phase two showed broadly north-south oriented compression, with north-northeast trending dextral strike-slip compression observed, noted by others as an Oligocene transpressional event (Vamvaka et al. 2006). In the Mikrolivado area we find this phase of N-S compression to result in NW/SE-striking thrust faulting and SE-dipping folding (Fig. 17.5). The faulting then became extensional (Fig. 17.6) with phase three due to east-west oriented extension and phase four showing north-south oriented extension, likely associated with Late Miocene to present day tectonics of northern Greece.

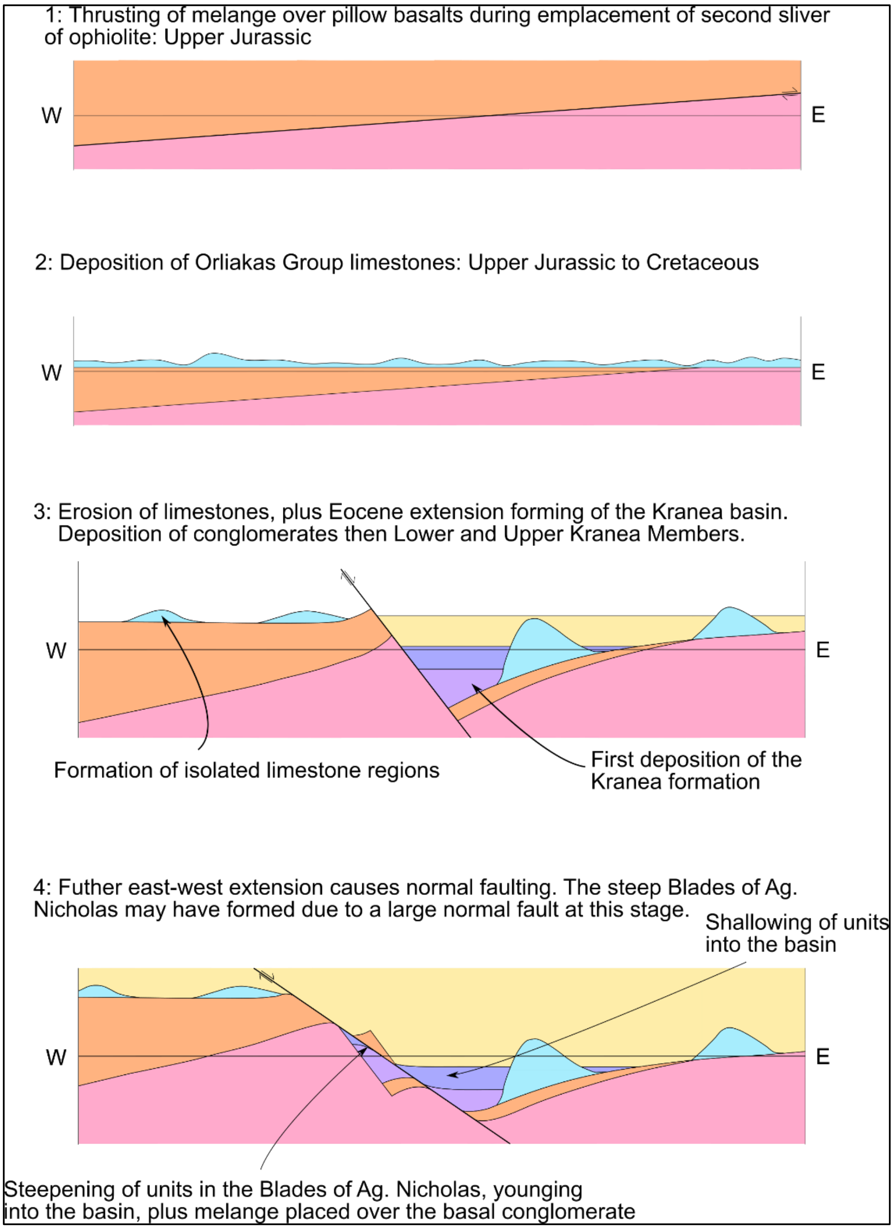


**Fig. 16:** The first four phases of the geological history of the Krania basin as inferred from observations in the study area. Colours represent units as described in the key on the map.

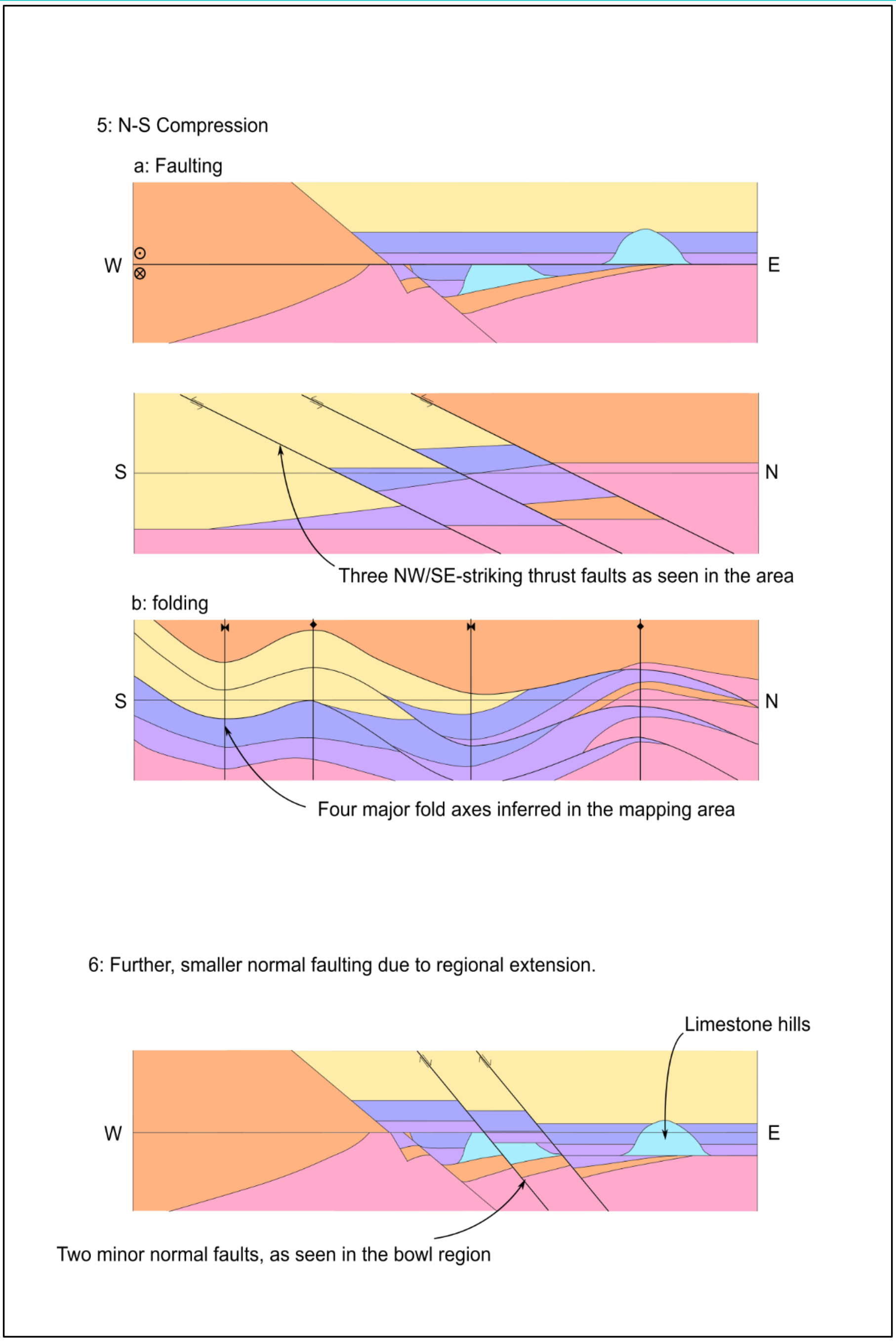


**Fig. 17:** Two further phases of the geological history of the Krania basin as inferred from observations in the study area. Colours represent units as described in the key on the map.

Alluvial deposits overlie the bedrock and appear to post-date all faulting. The erosion in this area appears to be dominated by steep-sided river valleys, with an absence of glacial erosion. The Pindos Mountains were at the southerly limit of the most recent extensive glaciation (Hughes, 2004). Although this glaciation reached as low as 1000m above sea level at Mount Olympus, 100 km east of Mikrolivado (Smith et al., 1997), closer to the area of fieldwork (at Vasilitsa, 16 km northwest of Mikrolivado) it only occurred above 1865 m (Hughes 2004).

## 8. CONCLUSIONS

We present a new geological map of part of the Krania Basin and its western margin, revealing initial infilling by alluvial fan deposits and subsequent turbidite deposition. The Krania Members show features consistent with either lacustrine or marine deposition, and we cannot distinguish these hypotheses. We identify four stages of post-depositional deformation affecting the Krania Basin, building on the work of Vamvaka et al. (2006): these features do little to distinguish the basin as either of pull-apart or piggyback origin. Our work corroborates regional tectonic hypotheses and provides evidence-based assessments of basin genesis and form.

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