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The Odivelas Palaeozoic volcano-sedimentary sequence: Implications for the geology of the Ossa-Morena Southwestern border

A sequência vulcano-sedimentar paleozóica de Odivelas: implicações para a geologia do limite Sudoeste da Zona de Ossa-Morena

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Keywords: Odivelas Limestone, Ossa-Morena Zone, Deformation phases, Variscan Orogeny.

Abstract: We present a preliminary characterization of the structure, stratigraphy and petrography of a sequence including the Odivelas Limestone near Covas Ruivas locality (Ossa-Morena Zone, Évora-Beja Domain). In this area, limestone and tuffites occur spatially associated with mafic meta-volcanic rocks presenting mineral parageneses typical of low grade metamorphism (*greenschists facies)* and with the occasional occurrence of silica iron exhalites (jaspers). Previous work dated the limestones at this locality as latest Emsian – early mid Eifelian. As indicated by folded and brittle structures, this Devonian sequence was affected by two Variscan deformation phases, which where individualized, related with regional orogenic D_2 and D_3 phases. The structural data indicate that the Odivelas Limestone stratigraphically overlays the metavolcanic rocks, although de contact between them was tectonically affected by a thrusting and/or shearing. Considering age assigned to the Odivelas Limestone and the fact that the effects of the regional D_1 are not visible in the studied area one can consider that, in this region, D_1 is pre-latest Emsian, i.e., Lower Devonian or older.

Palavras-chave: Calcários de Odivelas, Zona de Ossa-Morena, Fases de deformação, Orogenia Varisca.

Resumo: Neste artigo apresentamos uma caracterização preliminar da estrutura, estratigrafia, e petrografia da sequência que engloba os Calcários de Odivelas, perto da localidade de Covas Ruivas (Zona de Ossa-Morena, domínio Évora-Beja). Nesta área calcários e tufitos (Calcários de Odivelas) afloram associados a rochas metavulcânicas máficas, evidenciando um metamorfismo de baixo grau (fácies dos xistos verdes) e a ocorrências pontuais de jaspes. Trabalhos anteriores atribuíram estes calcários ao intervalo Emsiano terminal a Eifeliano médio baixo. A sequência Devónica estudada foi afectada por duas fases de deformação Variscas, correlacionáveis com as fases D_2 e D_3 reconhecidas a nível regional. Os dados estruturais indicam que os Calcários de Odivelas estão estratigraficamente suprajacentes às rochas metavulcânicas, embora o contacto esteja afectado tectonicamente por um cavalgamento e/ou desligamento. Considerando a idade atribuída aos Calcários de Odivelas e o facto de os efeitos da D_1 regional não se fazerem aqui notar, consideremos que nesta zona a D_1 é pré-Emsiano terminal, *i.e.*, Devónico inferior ou anterior.

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1. INTRODUCTION AND GEOLOGICAL SETTING

1.1. SW Iberian in European Variscan Belt

The Iberian Variscan belt (Fig. 1A) has been, for many years, the subject of numerous structural, metamorphic, magmatic and stratigraphic studies, aiming to understand its geodynamic evolution (Lötze, 1945; Silva *et al.*, 1970; Julivert, 1971; Ribeiro *et al.*, 1979; 1990; 2007; 2009; 2010; Andrade, 1983; Oliveira *et al.*, 1991; Fonseca, 1989; 1995; Fonseca & Ribeiro, 1993; Araújo, 1995; Fonseca *et al.*, 1999; Rosas, 2003; Rosas *et al.*, 2008).

The Ossa-Morena Zone (OMZ) is a major geotectonic domain located in the southern border of the Iberian

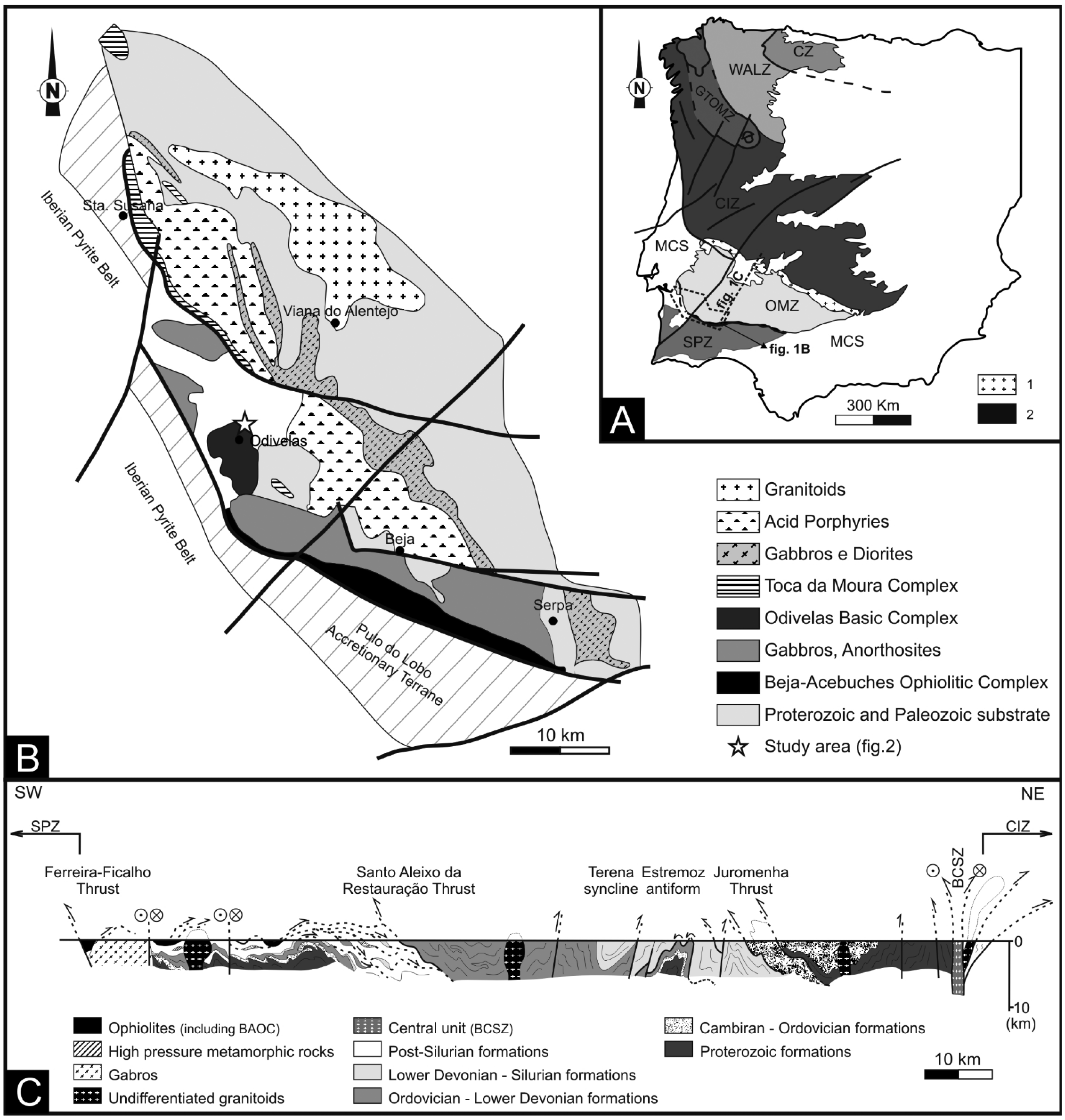


Fig. 1 – **A –** Iberian Peninsula main geotectonic divisions (adapted from Ribeiro *et al.*, 1979; 1990; San José *et al.*, 2004): MCS – Meso-Cenozoic rocks; OMZ – Ossa-Morena Zone, SPZ – South Portuguese Zone, CIZ – Central Iberian Zone, WALZ – West Asturian-Leonese Zone, CZ – Cantabrian Zone, GTOMZ – Galicia-Trás-os-Montes Zone, 1 – Pedroches Batholith axis, 2 – Beja-Acebuches Ophiolite; **B –** Simplified geological map of Ossa-Morena Zone / South Portuguese Zone northernmost contact, with location of study área (adapted from Oliveira *et al.*, 2006)**; C –** Geological profile on OMZ, SW Portugal: SPZ – South Portuguese Zone; CIZ – Centro Iberian Zone; BCSZ – Badajoz-Córdoba Shear Zone (Ribeiro *et al.*, 2007).

Massif (IM) which represents the largest and one of the most complete and continuous exposures of the Variscan Belt in Western Europe (Fig. 1A; LÖTZE, 1945; JULIVERT, 1971; RIBEIRO *et al*., 1990). The southern branch of the Iberian Variscan Belt comprises highly deformed exotic terranes of oceanic nature; these include the "Pulo do Lobo" Accretionary Terrane (PLAT), the Beja-Acebuches Ophiolite Complex (BAOC) (Fig. 1C; e.g., MUNHÁ *et al*., 1986; FONSECA & RIBEIRO, 1993; QUESADA *et al*., 1994; FONSECA *et al*., 1999) and the Internal Ossa-Morena Zone Ophiolitic Sequences (IOMZOS; FONSECA *et al*., 1999; ARAÚJO *et al.,* 2005; PEDRO *et al*., 2006; RIBEIRO *et al*., 2007, 2009; PIN *et al.,* 2008; PEDRO *et al*., *in press*).

The BAOC (ca. 30 km to the South of the studied area) has been regarded as an exotic oceanic terrain accreted to the Iberian Autochthon (OMZ), before the middle Devonian (Eifelian) (FONSECA & RIBEIRO, 1993; FONSECA *et al.,* 1999). It underlines the OMZ / South Portuguese Zone (SPZ) boundary (Fig. 1), being tectonically bordered in the South by the ductile/brittle Ferreira--Ficalho thrust.

It has been proposed (CRESPOBLANC & OROZCO, 1988; QUESADA *et al*., 1994; FONSECA, 1995, 1997; FONSECA *et al*., 1999; ALMEIDA *et al*., 2001; RIBEIRO *et al*., 2007; 2009; 2010) that a major ocean (Rheic) was closed by subduction/obduction leaving some remanent ophiolitic slices (e.g., the Lizard suture in SW England and the Beja Acebuches suture zone). Data acquired during the last decade clearly show that dismembered ophiolitic slices also crop out in the OMZ (the IOMZOS) which correspond to allochthonous klippen on top of lower Palaeozoic sequences (Fig. 1; e.g., PEDRO, 2004; ARAÚJO *et al.*, 2005; 2006; PEDRO *et al*., 2006; *in press*).

1.2. Synthesis of the OMZ Tectonostratigraphy

According to OLIVEIRA *et al*. (1991), the OMZ can be divided into five sectors, with distinct metamorphic and structural characteristics.

ARAÚJO (1995) adopted the proposal of OLIVEIRA *et al.* (1991) also taking into account the subdivision proposed by APALATEGUI *et al.* (1990) and defined the Évora-Beja domain, which includes the Montemor-Ficalho sector and BIC defined by OLIVEIRA *et al.* (1991).

The Évora-Beja domain is limited to the North by the Santo Aleixo Thrust (Fig. 1C; ARAÚJO, 1995) and to the South by a thrust that connects the OMZ with the Beja-Acebuches Ofiolitic Complex (BAOC) (Fig. 1A; FONSECA, 1989; FONSECA *et al.*, 1999). This domain is characterized by the abundance of acid, intermediated and basic intrusive rocks, whose genesis is related to the Rheic Ocean subduction below the OMZ (PEDRO, 2004; PEDRO *et al.*, 2006; RIBEIRO *et al.*, 2007). ARAÚJO (1995) considers that these intrusions span in time from the Middle Devonian to the Carboniferous and are only affected by the pulses of the last episode of Variscan deformation. The volcano-sedimentary formations have a large spectrum of ages, from Upper Proterozoic to Upper Palaeozoic. FONSECA & RIBEIRO (1992) and ARAÚJO (1995) identified three phases of Variscan tectonic deformation in this domain. This tectonic complexity and the rarity of geochronological data, make the stratigraphic correlation with the other sectors of the OMZ difficult.

Other important sub domain related with the studied area is the Santa Susana-Odivelas subsector and the Beja Igneous Complex (BIC) (OLIVEIRA *et al*., 1991). The BIC (Fig. 1B) is a plutonic association of variable compositions, from basic (Beja Gabbro-Dioritic complex, SILVA *et al*., 1970) to acid rocks (Baleizão-Alcáçovas porphyry, CARVALHO *et al*., 1971), also including volcano-sedimentary complexes (Odivelas basic Complex, Toca da Moura Complex, SANTOS *et al.*, 1987). The northern boundary with the Montemor-Ficalho Sector is not tectonic, while the southern contact coincides with the BAOC and the Ferreira – Ficalho Thrust (OLIVEIRA *et al*., 1991).

Recently, JESUS *et al.* (2007) and PIN *et al.* (2008) dated the plutonic bodies of the BIC. JESUS *et al.* (2007) dated the Layered Gabbroic sequence as lower-middle Mississippian (*ca*. 355 Ma to *ca*. 345 Ma) formed in the early stages of collision magmatism and the Baleizão Porphyry Complex as Pennsylvanian (*ca.* 300 Ma) and was considered to correspond to post-collision magmatism. PIN *et al.* (2008) obtained U-Pb zircon ages around 350 Ma interpreting them as reflecting the BIC intrusion in a late-collisonal transcurrent setting.

The volcano-sedimentary complexes have distinct tectono-stratigraphic characteristics, and occur separated in Santa Susana Odivelas sub-sector. Within this subsector, volcanic and plutonic complexes are present in the geo-transverse between Odivelas and Alvito. These include Odivelas Complex, Faro-Alvito Complex and Peroguarda Complex (Fig. 2). Considering its occurrence in the studied area, only the main Peroguarda Complex, will be described. ANDRADE *et al.* (1976) studied the Peroguarda Complex and sub-divided it in Casa Branca dolerites, Rebolado basalts and Mota Preta tuffs.

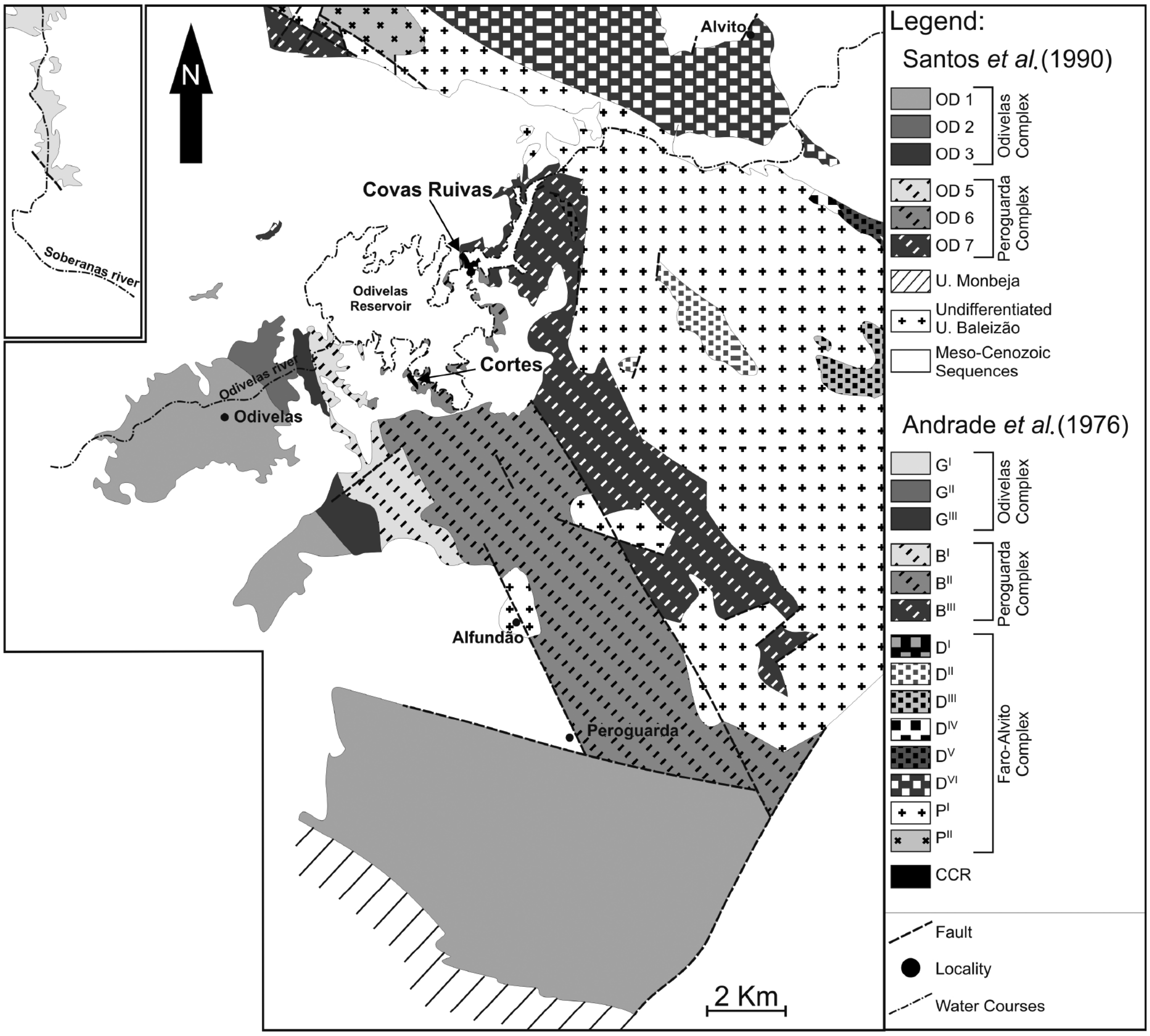


Fig. 2 – Geological map of the main igneous units surrounding the studied area: D^{I} – Monte Olival Dolerite, D^{II} – Monforte Diorite, D^{III} – Monte Novo Diorite, D^{IV} – Malcabrão Gabbros, D^{V} – Assentes Quartzo-diorite, D^{VI} – Alvito Quartzo-diorite, P^{I} – Monte Ruivo Porphyry, P^{II} – Castelo Ventoso Porphyry; B^{I} – Casa Branca Dolerite, B^{II} – Rebolado Basalts, B^{III} – Mota Preta Tufts; G^{I} – Herdade Grande Gabbros, G^{II} – Balona Gabbros, G^{III} – Gravitosa Gabbros; CCR – Odivelas Limestone. The white color represents Cenozoic cover (adapted from ANDRADE *et al.,* 1976; SANTOS *et al.,* 1990).

The Rebolado basalts are apparently placed at foot and hanging wall of the Odivelas Limestone. They are formed by massive spilitised lava flows (gradding to the Casa Branca dolerites) and by pyroclastic facies. SANTOS *et al*. (1990) referred to the Rebolado basalts as the unit OD-6 (Fig. 2). The geochemical studies performed by these authors allowed the confirmation of its orogenic character, as indicated by its calc-alkaline/tholeitic chemical signatures.

The Odivelas Limestone is a Devonian unit that occurs in scattered within and around the Beja Igneous Complex, Ossa-Morena Zone (BIC – OMZ) (ANDRADE, 1983). Its precise age has been a subject of debate for decades. The first stratigraphic study indicated an age of mid and/or late Devonian based on fauna rich in crinoids, corals, brachiopods and bryozoans, occurring at the Cortes locality (Fig. 2; CONDE & ANDRADE, 1974; ANDRADE *et al.*, 1976). However recent findings indicate that the age of these limestones is latest Eifelian-earliest Givetian at the Cortes locality (Fig. 2; MACHADO *et al.*, 2009) and latest Emsian – early mid Eifelian at the Covas Ruivas locality (Fig. 2; MACHADO *et al.*, 2010).

The occurrence of limestones, easily datable using precise stratigraphic criteria (e.g., conodonts, crinoids), associated with volcanic and sub-volcanic rocks offers an unique opportunity for the chronological characterization of the extrusive part of the BIC and for the dating of the various regional tectonic events.

Aiming a contribution for the understanding of the Late Palaeozoic evolution of the Southern border of the Ossa-Morena Zone, we present preliminary data on the structural geology, stratigraphy and petrography of the Covas Ruivas Devonian sequence, in the Northeastern bank of the Odivelas reservoir (Fig. 2), Beja district.

2. RESULTS

2.1. Stratigraphy and Petrography

A Palaeozoic volcano-sedimentary sequence that crops out in the Odivelas reservoir area shows distinct and unusual characteristics when compared with other OMZ sectors. In this chapter, we will describe the mineralogical, textural and stratigraphic characteristics of relevant outcropping rocks in the studied area.

2.1.1. Limestones and tuffites sequence (LTS)

The Covas Ruivas locality shows a long sequence of volcanic (see chapter 2.1.2.) and sedimentary rocks that crop out along the banks of the Odivelas reservoir (Fig. 3). In spite of minor faults and local lack of exposure, biostratigraphical data (MACHADO *et al.,* 2010) suggest that there are no major gaps or repetitions, indicating a fairly continuous sequence from the basal part (to the West) to the top (to the East). The total thickness of the LTS is estimated to be around 200m.

The volcanic lavic and pyroclastic rocks exposed to the West grade upwards into tuffs and tuffites. The basal part of the LTS is dominated by thin-bedded, quartz-rich tuffites with subordinate limestone beds. In the basal 9 m the limestone beds are made up by cm- to dm-thick crinoidal grainstones with few basinal elements (peloids, pelagic microfossils). From ca. 9 m up to 47 m the limestone beds (mainly calcimudstones) become rarer and laterally discontinuous. This is accompanied by an increase of the amount of organic matter and proportion of basinal elements. Accordingly the interbedded tuffites become darker and richer in siliceous detritus and pelagic fossils, locally forming radiolarite lenses. Above 47 m limestone beds become suddenly dominant over the tuffites (which become more calcite-rich) and are markedly coarser (wacke- to grainrudstones). The thickness of limestone beds increases (dm- to m-thick) and the proportion of reef-derived bioclasts becomes dominant, although basinal elements remain important components (10% to 40% of total allochems) (MACHADO *et al.,* 2010). This relatively long interval of basinal deposition followed by a sudden increase of calciclastic material in this time interval (earliest Eifelian) represent characteristic lithological features of the pre-basal Choteč event beds (see MACHADO *et al*., 2010 for details). In the interval from 47 m up to 57 m there are significant fluctuations on the thickness and lateral continuity of limestone beds and the interbedded tuffites. Above this interval and up to 80 m (top of the first part of the section) limestone beds have fairly constant thicknesses and lateral continuity and are generally dominant over the tuffites, which are calcite-rich although quartz remains an important component (MACHADO *et al*., 2010). The two other parts of the LTS are separated from the first part by an observational gap and a faulted zone respectively. The second and third parts of the sequence show a relatively monotonous sequence of dm-thick crinoidal grainstone beds interbedded with mm- to cm-thick laminated tuffite beds (MACHADO *et al.,* 2010). The top of the third part of the sequence is already late middle Eifelian in age (see MACHADO *et al.,* 2010 for details on conodont biostratigraphy and sequence interpretation).

2.1.2. Volcanic Sequences (VS)

In the described section, magmatic rocks are present in positions both geometrically below (Lower Volcanics – LVS) and above (Upper Volcanics – UVS) the Devonian sediments (Fig. 3).

The LVS can be subdivided in 3 major units taking into account the type of dominant volcanic material. They are: 1) pyroclastic heterometric lithified deposits where lithoclastic blocks can reach some 15 cm in diameter (see Plate 1F); 2) lava flows; 3) volcanic breccia (Fig. 3). This volcanic pile outcrops along some 300 m in the dam bank.

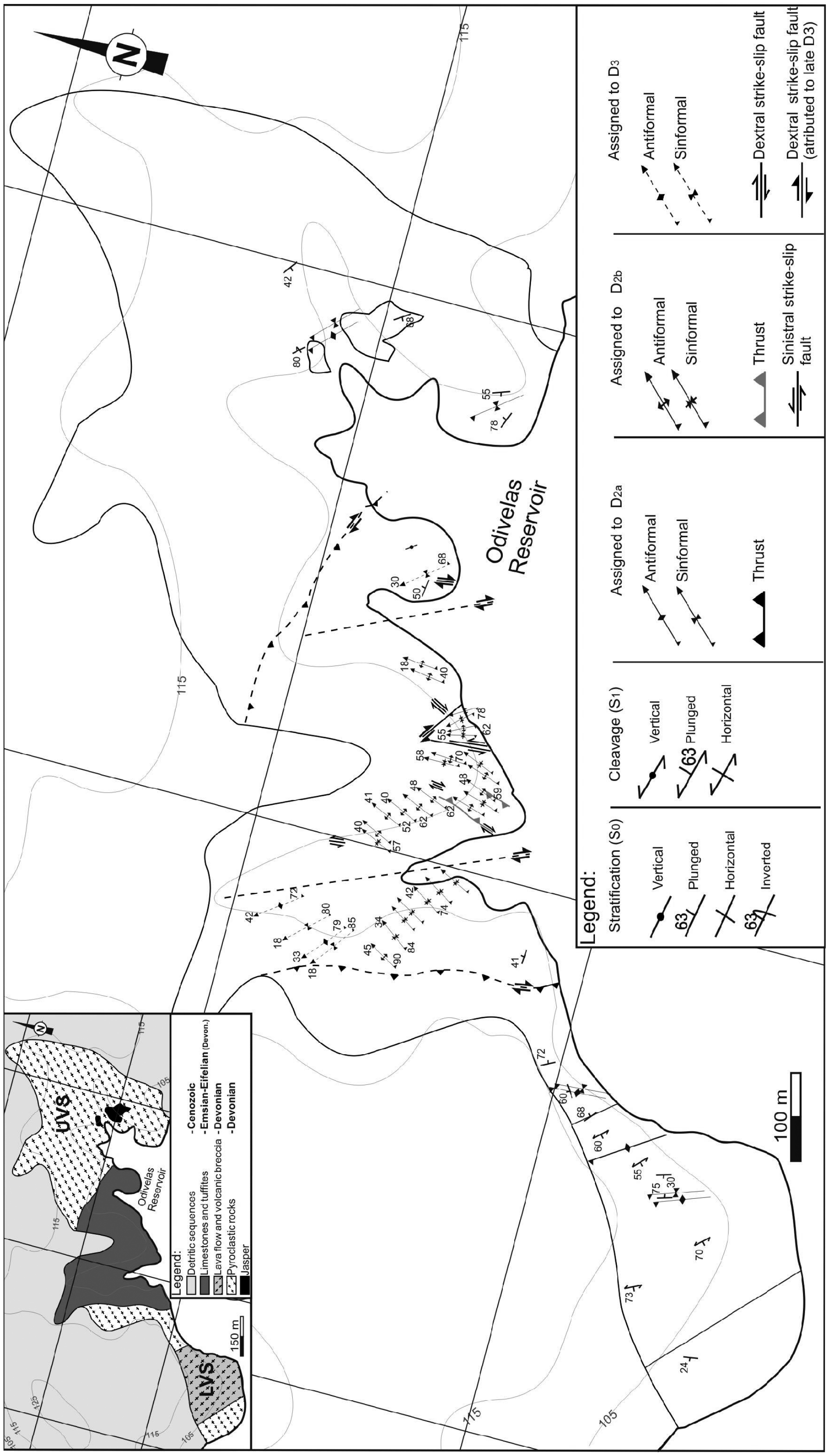


Fig. 3 – Simplified lithological and structural map of the studied area (see also MOREIRA *et al.*, 2010).

Micropetrographic observation of the lavic rocks reveals that, despite metamorphic blastesis, deformation was not significantly penetrative to obliterate the main textural aspects of the volcanic protoliths. The porphyritic character of most of the lavas is easily recognized, with phenocrysts set in fine grained (aphanitic) groundmasses, producing typical relict textures of the blasto-porphyritic type (see Plate 1H). Groundmasses frequently present fluidal arrangement of feldspar crystals reflecting flow of magma.

Groundmasses are dominated by plagioclase and alkali feldspar, while the phenocryst generation is mainly composed by plagioclase. Where the blastesis was less important, zoned clinopyroxene relict phenocrysts are sometimes preserved. Opaque minerals are partially transformed in titanite.

The preserved and/or inferred magmatic mineralogy suggest that the majority of the metavolcanic rocks are of basic to intermediate composition.

Small amygdales filled by pistacitic epidote, chlorite and carbonate are volumetrically important in some levels of the magmatic sequence. Judging by textural criteria, inside the amygdales epidote crystallization preceded the formation of chlorite.

The presence of actinolite, chlorite and epidote indicates that volcanic rocks were metamorphosed in conditions typical of the greenschist facies. It should be mentioned that the abundance of the Ca-amphibole is more important towards the top of the sequence.

The UVS are mainly pyroclastic (Fig. 3) and dominated by lappilli tuffs, which probably correspond to more distal volcanic facies than those preserved on the LVS. Interbedded in these tuffs some lava flows occur, petrographically identical to those described for the LVS, and some lenses of silica-iron exhalites (see Plate 1 D, E).

2.1.3. Silica iron exhalites

The Odivelas silica iron exhalites (jaspers) are made up of microcrystalline quartz and hematite with notable absence of detritic minerals. The proportions of quartz and hematite are variable (90-95 vol% and 5-10 vol%), respectively. The degree of recrystallization of jaspers varies to a great extent. Commonly, less recrystallized samples exhibit domains with well preserved primary textures. Of these, the most prominent are spherulitic structures (see Plate 1 E) and brecciation features.

When well-preserved, spherulites (0.03 – 0.5 mm) consist of a central core of anhedral grains of hematite sometimes intergrown with microcrystalline quartz. The core is involved by one or more concentric layers of microcrystalline quartz or chalcedony. Groups of spherulites are frequently coalesced, sometimes assuming a wavy form, reminiscent of plastic movements of soft, unconsolidated sediments.

Syn-sedimentary brecciation features are common throughout the whole jasper-bearing area. They are characterized by the presence of millimetric to centimetric fragments surrounded by a complex network of microcrystalline quartz. The morphology of fragments ranges from very sharp-edged to angular or subrounded shapes, frequently exhibiting mutually-fitting broken walls. The fragments are made up either by microcrystalline quartz (± hematite – type I) or by hematite (± microcrystalline quartz – type II). In type I fragments the presence of spherulites is common. Type II fragments often exhibit sets of polygonal cracks filled by microcrystalline quartz indicative of early-stage diagenesis dehydration-contraction phenomena. The relative abundance of the type II fragments suggests a post-diagenetic disruption of pristine banded structures.

Collectively, these features indicate that these Si-Fe rich sediments formed from the crystallization of silica iron oxyhydroxide gel, compatible with the low-temperature hydrothermal activity contemporaneous with their deposition.

2.2. Structure and Tectonics

2.2.1. Folded Structures

The studied area has a significant diversity of folded structures (Fig. 3; Plate 1 B, C). Three different families are recognized with respect to the folded structures' geometry (Fig. 4).

Turner & Weiss (1963) classification was used to describe the fold geometry, for the relation of hinge line and axial plane attitudes, and Fleuty (1964) classification, for the interlimb angle (Ramsay & Huber, 1983).

2.2.1.1. Family 1

From the geometrical point of view, this family is composed by two macro scale folds. The structure that

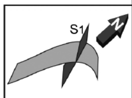
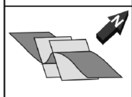
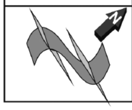
	Symmetry	Axial Surface	Hinge Line	Tightness	Cleavage	Vergency	Localization	Comments	Deformation phase
	Asymmetrical Long limb - 30ºSW Short limb - 70ºNE	Inclined N135º, 72ºSW	Slightly Plunging (15º) to SE	Open (Interlimb angle 80º)	Axial fracture cleavage N140º, 70ºSW	NE	Only in volcanic sequences	Two Macroscale folds. Part of the folds are covered by Cenozoic; Presents parasite folds;	D2a (family 1)
	Asymmetrical Long limb - 50ºNE Short limb - 71ºNW	Inclined (variable) N22º,70ºSE	Plunging 49º, N41º	Open (Interlimb angle 80 º)	-	NW	Only in limestones and tuffites	Hinge thickening on tuffites (more obvious) and limestones beds.	D2b (family 2)
	More or less symmetric - limb tilt 50º e 60º to NE and NW	Inclined (almost vertical) N310º, 80ºNE	Plunging 49º, N341º	Open (Interlimb angle 90 º)	-	SW	Only in limestones and tuffites	-	D3 (family 3)

Fig. 4 – Geometric and kinematic synthesis of folded families in the studied area.

crops out to the NW is not fully mapped, because the Cenozoic deposits cover part of the structure. These folds are located in edges of the studied area. They affect mainly the volcanic rocks.

They are cylindrical asymmetric with the long limb (SW limb) tilting ca. 30º to SW and the short limb (NE limb) has slopes 70º to NE. The axial plane has a NW-SE direction, plunged about 72º to SW. This fold family has a hinge line slightly inclined (about 15º) to SE. These are open folds, with an interlimb angle of around 80º. It is possible to classify these folds as inclined slightly plunging, using the relation between the hinge line and the axial plane. These are folds with normal polarity forming essentially antiforms-anticlines. The polarity is determined based essentially on the presence of graded bedding on pyroclastic beds.

These "macro-folds" generated parasitic folds. The parasitic folds observed within the long limb are also cylindrical asymmetric, with long limb tilting 40º to SW and the short limb tilting 75° in the opposite direction (Fig. 5). The parasitic folds are closed, with an interlimb angle of around 65º. They have an axial plane with a direction of around N130º, plunging about 70° to SW. The parasitic folds located on the short limbs are closed (interlimb angle 65°), with hinge lines plunging 50° to N; the axial plane direction is approximately N-S, with slopes around 80º to W. They are asymmetric, with long limb plunging 70º to the NE and the short limb 60º to the opposite direction.

An axial plane cleavage (S1) with a pervasive attitude N140º, 70°SW is observed in these folds (Fig. 6).

Cinematically the macrofolds have a very clear vergency to the NE quadrant, with evidence both in folding and the associated axial plane cleavage (S_1). The vergency of the parasitic folds could be differentiated: in the long limb the vergency is to NE as the macrofold, in the short limb the vergency is to E.

If we assume that the tensional forces required to form these folds are perpendicular to axial planes, then the direction of maximum compression (σ1) would be, if the deformation was coaxial, approximately SW-NE.

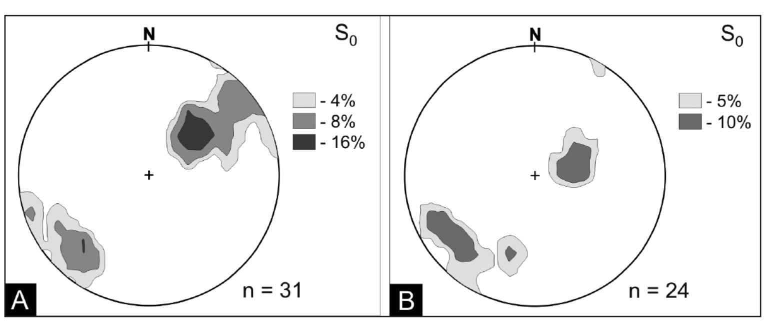


Fig. 5 – Family 1 folds: Schmidt diagram (lower hemisphere) with diagram point density: A – S_0 on Upper Volcanic Sequence (31 measurements); B – S_0 on Lower Volcanic Sequence (24 measurements).

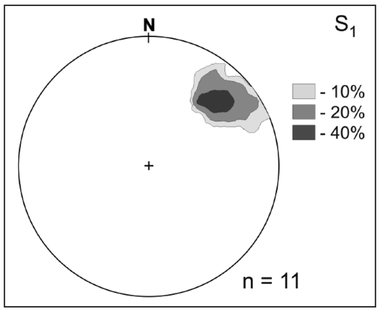


Fig. 6 – Family 1 folds: Schmidt diagram (lower hemisphere), with diagram point density that shows the poles of axial plane cleavage (S1) on volcanic rocks, with greater frequency around N140º, 70ºSW.

2.2.1.2. Family 2

This family includes most of folded structures present in the studied area. These structures generate sequences of folds with alternating antiformal and sinformal sequences, which are also anticlines and synclines respectively, considering they all have normal polarity.

Geometrically, the folds of family 2 have the axis plunging to NE. Folds are open (interlimb angle of around 80º), with hinge lines dipping 40° to NE (Fig. 7A); they are cylindrical asymmetric, with the long limb plunging 50° to NE and short limb 71º to N-NW. They have an average axial plane attitude N22º, 70º SE (Fig. 7B). These folds are classified based on the relationship between the hinge line and axial plane attitudes to inclined plunging. In some antiformal structures, it is possible to observe deformation accommodation by reverse faulting.

Locally a fold system subtype is identified in this family. This sub-type is characterised by closed folds (opening angle is 70°), cylindrical asymmetric, with long limb tilting 40° to NW and the short limb that plunges approximately 70° to NE. The hinge line tilts about 60º to NW-NNW and its axial plane shows a direction of N345º tilting about 65° to W. Thus, these folds are classified as inclined plunging.

From the kinematic point of view, these folds present vergency to NW. This vergency is mainly observed in the field, but only quantified from the projection data in a stereogram. The axial planes plunge to SE with geometric vergency to NNW-NW.

It should also be noted that these folds' family have a thickening of layers in its axis, more evident in tuffite beds, also possible to observe in limestone beds as well.

The subtype has an axial plane inclined to SW. These structures have vergency to NE-ENE. The explanation for the low spatial dispersion of this subtype will be addressed in subsequent chapter. The folds of this subtype have a very clear thickening of beds in their axis.

This family has an axial plane with NNE direction, plunging to SE, with vergency to WNW. If deformation was coaxial, these folds would be explained by tensional forces with maximum compression direction ESE-WNW.

These folds subtype that has vergency to NE-ENE, are restricted to a deformation corridor located between two faults, and this is possibly the reason for the recorded divergence. Geometrically, the two subtypes are quite identical.

2.2.1.3. Family 3

The detailed mapping shows that the geometry of this family, located in the NW area of the LTS, consists on cylindrical folds, more or less symmetrical, with the tilt

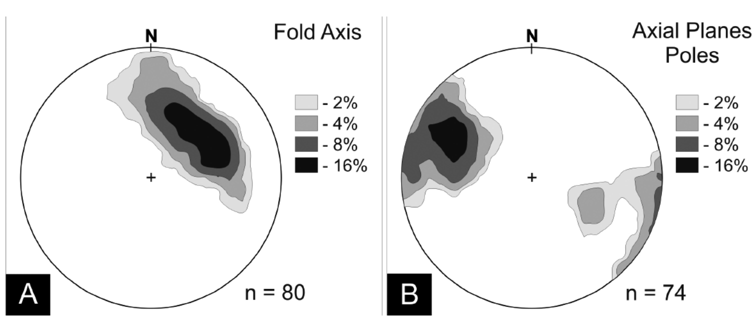


Fig. 7 – Family 2 folds: Schmidt diagram (lower hemisphere) with diagram point density: A – hinge lines, with greater frequency around 49º, N41º (80 measurements); B – axial plane poles, with greater frequency around 20º, N292º (74 measurements).

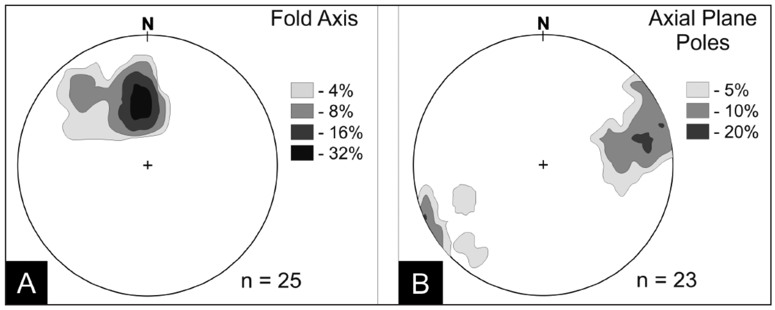


Fig. 8 – Family 3 folds: Schmidt diagram (lower hemisphere) with diagram point density: A – hinge lines, with greater frequency around 49º, N341º (25 measurements); B– axial plane poles, with greater frequency around 15º, N72º (23 measurements).

values of the limbs ranging from 40º to 60º in opposite directions (NE and NW). It is characterised by open folds (interlimb angle around 90º), with hinge lines dipping 40° to NW (Fig. 8A) and axial plane with a prevailing attitude N310º, 80º NE (Fig. 8B).

A large sinform fold, located in the eastern border of the LTS with volcanic rocks (UVS), is also part of this family. It presents a hinge line plunging 30° to NW, with an axial plane with a direction 310º, tilting 66° to NE. This sinform has parasitic folds. All the mapped folds have normal polarity.

This family has, in general, axial planes slightly plunged to the NE quadrant, with a fold vergency to SW. The large sinform on the East of LTS has the same vergency. Consequently, this family has a geometrical vergency to SW, with axial planes plunging to the NE quadrant and a direction approximately NW-SE. These structures can be explained, in view of a dominant compressive field, where the maximum compression tensor has a direction approximately NE-SW, perpendicular to the axial planes.

It should be noted that fold families 2 and 3 are present only in LTS. The family 1 could only be detected in the massive volcanic sequence. This is interpreted as resulting from distinct rheological characteristics of the different rock types (LTS *vs.* VS), which may have induced the genesis of a tectonic discontinuity (thrust/shearing?) during Variscan deformation events.

2.2.2. Brittle Structures

Some of the brittle structures are interpreted as deduced, which explains some inconsistencies in the structural interpretation.

Similarly to the more ductile structures, the brittle structures are divided into fault families, with distinct geometric and kinematics characteristics (Fig. 9). The fragile structures are divided into four families:

I. Parallel faults and/or overlap limit LTS-UVS and LTS-LVS;
II. Faults with direction N10° to N45°;
III. Faults with direction N-S;
IV. Faults with direction N315° to N350°;

I – *Parallel faults and/or overlap limits LTS-VSB and LTS-VST*

This family includes two faults separating the LTS from the VS. In the borderland between LTS and LVS, the contact is sharp, and the attitude of the LVS stratification (S_0) is distinct from the LTS. This contact can be interpreted as a brittle/ductile dextral strike-slip fault. This fault has a direction N340º, plunging to W.

A difference in the folding geometry between the LTS and the UVS can be observed. This differentiation, caused by different rheological behaviours, is similar to the one observed at the limit between the LTS and LVS. The limit is also interpreted as a fault with a general direction of N290º, with a fault plane inclined to SW (Fig. 3).

Cinematically, both faults could be assumed to be dextral brittle/ductile strike-slip or brittle/ductile thrust with vergency to NE. The kinematics could correspond to a previous deformation phase (which are grouped in fold family 1 – see chapter 2.2.1.1) or a late D_3, but always related with a significantly diverse rheological behaviour.

	Strike	Dip	Kinematics	Fault family	Deformation phase
	N290º and N340º	40º SW	Trust with vergency to NE	I	D2a
	N10º to N20º	Vertical	Dextral strike-slip fault	II	D3
	N45º to N20º	Variable but more than 38º	Trust with vergency to NW	II	D2b
	N6º and N20º	Variable but more than 38º	Trust with vergency to SE	II	D2b
	N-S	Vertical	Sinistral strike-slip fault	III	D2b
	N320º to N350º	Vertical	Dextral strike-slip fault	I and IV	(Late D3 - Upper Carboniferous)

Fig. 9 – Geometric and kinematic synthesis of four brittle structures families in the study area.

The presence of these thrusts with vergency to NE, along the limit between the LTS and LVS, accompanied by folds with axial plane sub-parallel to the fault, points to the association of these two types of structures with the same deformation episode (thrust criteria). In this case, the stress field would be dominantly compressive in order to generate the folding. If we consider that the deformation is coaxial, then the direction of maximum compression ($\sigma1$) would be perpendicular to the axial plane.

Alternatively, if this is interpreted as a ductile/fragile dextral strike-slip structure, a compressive $\sigma1$ approximately with N-S direction should be considered.

II – *Faults direction N10° to N45°*

The detailed mapping points out that this family is composed by faults with directions between the N6º to N45º. The slope is very variable, but always above 38º to NNW or to SSE. This family is only identified in the LTS.

The mentioned faults have different drives. Four combinations of different horizontal and vertical component can be distinguished:

– Reverse faults with vergency to NW, without horizontal component (only visible in one case) – direction N45º;

– Reverse faults with a NW vergency, with dextral horizontal component (only visible in one case) – direction N20;

– Reverse faults with vergency to SE, without horizontal component (in four places) – direction between N6º to N20º;

– Dextral strike-slip fault, without vertical component (visible in three places) – direction between N10º to N20º.

There are evidences of two distinct fields of tension that would explain the two main types of faults:

– The reverse faults are formed by ductile deformation accommodation. These faults are associated with folded structures grouped in family 2 (see chapter 2.2.1.2.; Fig. 10). Considering that the deformation is coaxial, a stress field near vertical $\sigma3$, $\sigma2$ SW-NE and $\sigma1$ NW-SE is required to form this fault association (Fig. 10B).

– To explain the dextral strike-slip faults observed in some places, a stress field with maximum compression near NE-SW is required.

III – *Faults with N-S direction;*

Two main faults of this family correspond to the limit of the deformed zone, referenced on fold family 2, and have, usually, a vertical slope. These faults appear to be of similar origin. In the studied region, it only affects the LTS. The faults of this family show the same relative kinematics. One of the faults drags the stratification, showing clear sinistral horizontal movement (Fig. 11).

With respect to the dynamic analysis of structures grouped in this family, they all show a similar origin,

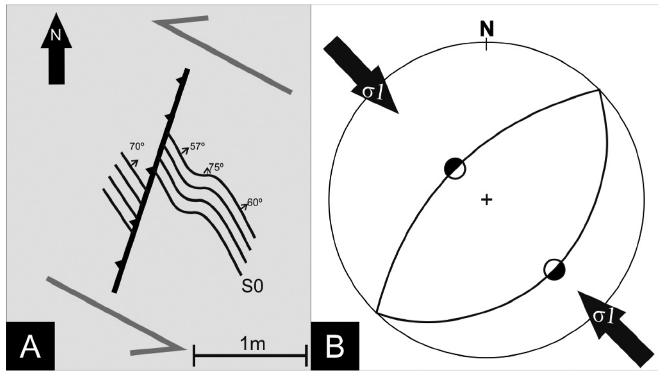


Fig. 10 – **A –** Schematic and interpretative map that shows the relationship between the family *2* folded structures and the N20º thrust with vergency to SE, framed within a sinistral transpressive regime; **B –** Thrust families projections that are include in this fault family (horizontal movements not recorded).

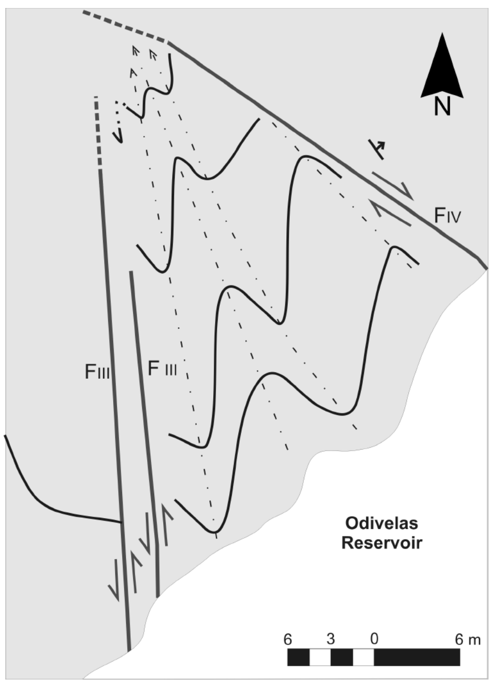


Fig. 11 – Schematic and interpretative map of the deformation in a deformation corridor, which promotes a rotation on folds grouped on family 2 (in this corridor, hinge lines plunges to N350º, since in family 2 hinge lines plunges to NE quadrant; F_{III} fault family III and F_{IV} fault family IV).

since they all have similarities in their geometrical and kinematic characteristics.

If we consider a coaxial deformation, although the conjugate faults have not been found in mapped area, we can infer on the possible direction of maximum compression, responsible for the formation of these faults; σ1 with a SE-NW direction.

IV – *Faults direction N315° to N350°*

In this family, that includes five faults, all the faults that have directions between N315° and N350° are subvertical. Regarding the kinematics of these faults, they have a dominant horizontal component (vertical component not deduced); however it is necessary to subdivide this family into two groups:

– Fault of direction N315°, vertical, with sinistral strike-slip component (very scarce);

– Fault direction N320° and N350°, with vertical dextral strike-slip component (more abundant).

The sinistrogiral strike-slip structures can be explained by a stress field where the maximum compression direction was close to WNW-ESE, while the dextral strike slip faults, imply a σ1 with approximately N-S direction.

Fig. 11 represents an area where three faults are observed, two of which are N-S sinistrogiral strike-slip structures (fault family III) and the other is a dextral strike-slip with direction N320° (belonging to this family). The activity of these faults cause a rotation on folded structures, which led to a reorientation of the original structure. These folds were similar to the folds of family 2, rotated by the action of these three faults.

3. DISCUSSION AND CONCLUSIONS

3.1. Correlation with OMZ Structures

In the OZM, the first Variscan deformation event (D_1) has been interpreted as corresponding to the installation of BAOC to the North with the consequent generation of a flake tectonic geometry (ARAÚJO *et al.,* 1993; FONSECA & RIBEIRO, 1992). It is considered to have lasted from the Early to Mid Devonian (Fig. 12 and 13), but may have started in the Silurian in the SW parts of the Évora-Beja Domain (*cf.* FONSECA & RIBEIRO, 1993; ARAÚJO, 1995; RIBEIRO *et al*., 2007, 2009, 2010). Considering the age assigned to the Odivelas Limestone (latest Emsian-early mid Eifelian; see MACHADO *et al.,* 2010) and the fact that the effect of the regional D_1 is not visible in the studied area one can infer, in this region, D_1 is pre-latest Emsian, i.e., Lower Devonian or older (Fig. 12).

In the studied area, the presence of folds with axial planes aligned NW-SE and vergency to NE (family 1) is interpreted as resulting from the first pulses of the 2nd regional deformation phase (D_{2a}).

As the subduction continued the second deformation phase (D_2) was generated (ARAÚJO, 1995) (Fig. 12). Regionally the D_2 folds, in the Évora-Beja Domain, have N-S to SW-NE direction and are asymmetric with W to NW vergency, suggesting a constrictive system (DIAS & RIBEIRO, 1995, RIBEIRO *et al*., 2010). The D_2 hinge lines have a great dispersion, due the effect of the third deformation episode (D_3) (ARAÚJO, 1995).

In the Covas Ruivas area we grouped folds with similar geometry in family 2 described in chapter 2. However, as they show a low dispersion of the hinge lines, standing to NE quadrant, they are coherent with D_{2b} structures described by ARAÚJO (1995) and RIBEIRO *et al.,* (2010).

The third deformation phase in the OMZ (D_3) has presumably a Famennian-Visean (Late Devonian--Mississippian) age (FONSECA, 1995). During this phase the intracontinental deformation mechanisms were predominant, with a maximum compression direction close to NE-SW (ARAÚJO, 1995, DIAS & RIBEIRO 1995).

This last episode of deformation is represented in the studied area by faults family III which have a N10º direction, with dextral strike-slip movement. No previous works in OMZ refer faults with these geometric and kinematic characteristics. However, BASILE AND DIAS (2008) have reported the presence of dextral strike-slip

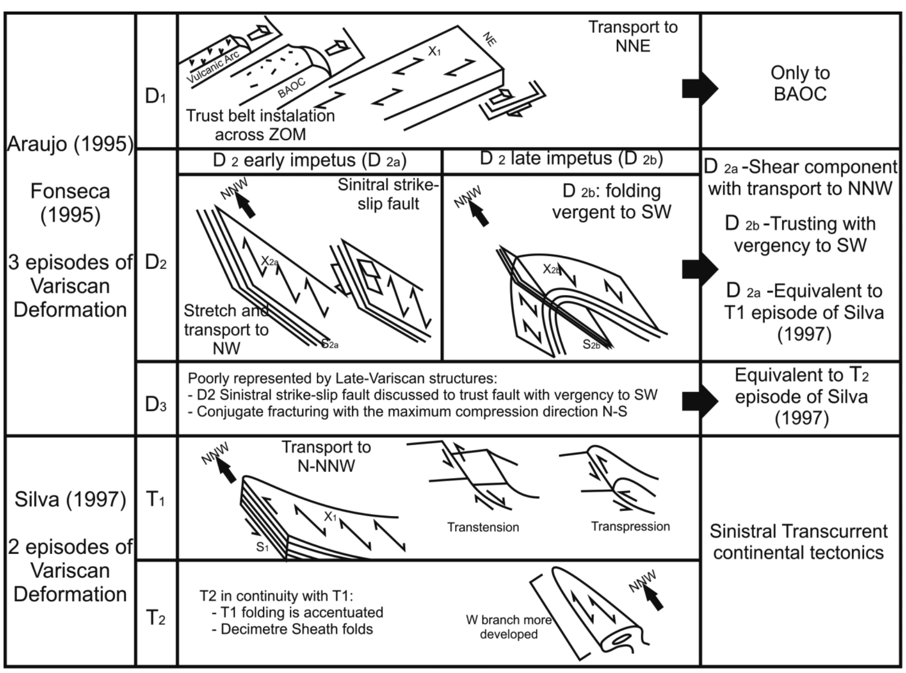


Fig. 12 – Summary geometric and kinematic characteristics of the main Variscan deformation episodes in Évora-Beja Domain according to different authors (adapted from ROSAS, 2003).

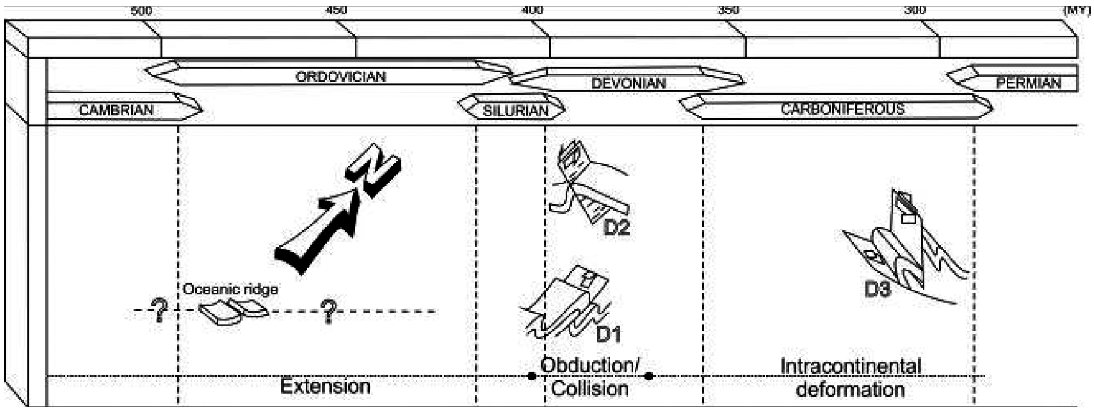


Fig. 13 – Dating of the main events of deformation in the Iberian Peninsula in SW Iberia (Dias & Ribeiro, 1995).

faults with NNE-SSW direction in the SPZ. According to these authors, these faults are associated with later stages of D_3.

Finally it is important to refer that the dextral strike--slip faults with direction between N350º and N315º described above (see chapter 2.2.1), can be related with Porto-Tomar-Ferreira do Alentejo shear zone and the Santa Susana Shear Zone with which they share kinematic and geometric similarities. Almeida *et al.* (2006) and Oliveira *et al*. (2007) refer that the later was active at least since the Pennsylvanian.

The distinct folds families present in different units (family 1 in VS and family 2 and 3 in LTS) is interpreted as resulting from distinct rheological characteristics of the different rock types, which may have induced, during Variscan deformation events, the genesis of the observed tectonic discontinuity (thrust/shearing?) between them.

Moreover, the observed structures are consistent with models developed for the southern border of the OMZ (e.g., Andrade 1983; Santos *et al*., 1987; 1990; Jesus *et al.*, 2007; Araújo *et al.*, 2005; Fonseca, 1995; Fonseca & Ribeiro, 1993; Fonseca *et al.* 1999; Ribeiro *et al.,* 2010).

3.2 Palaeogeography

The Limestones and Tuffites sequence (LTS) indicates calciturbidite deposition associated with a reef system, probably at the base of a slope setting. This is indicated by the reef-originated bioclasts forming the limestones and the occurrence of interbedded tuffites which probably represent basinal deposition between turbidite events (Machado *et al*., 2010). The close asso-

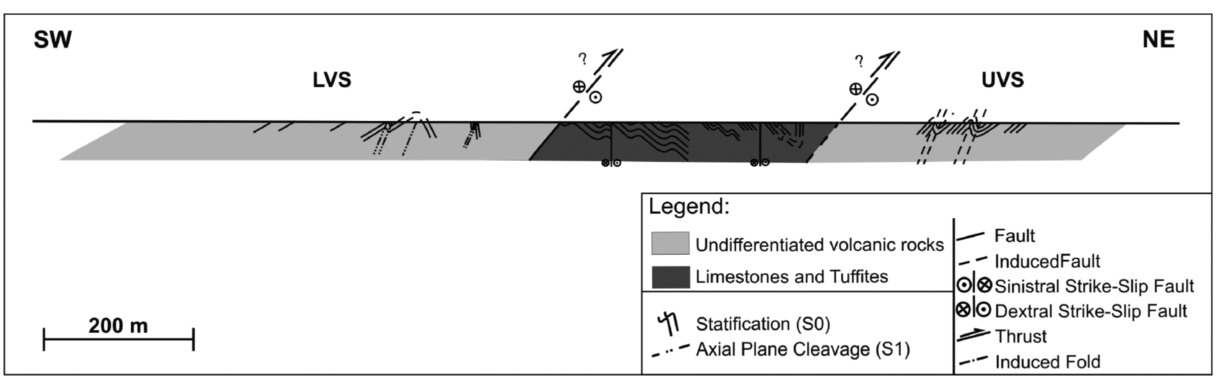


Fig. 14 – Simplified geological profile for the studied area.

ciation of a reef system with volcanic rocks suggests the existence of a relatively isolated area of carbonate sedimentation. There is no evidence of the existence of a carbonate platform. Similar tectonosedimentary settings were described in the Bohemian massif (e.g., GALLE *et al*., 1995, HLADIL *et al*., 1994, 1999; CHLUPÁČ & HLADIL, 1992) and Rheno-Hercynian zone (e.g., FLICK *et al*., 2008; KÖNIGSHOF *et al*., 2010).

The discrete occurrences (few square meters) of silica iron exhalites in association with the volcanic sequences indicate the existence of low temperature, diffuse hydrothermal activity. Moreover, well preserved primary textures observed in many samples of these exhalites suggest that these sediments formed from the crystallization of silica iron oxyhydroxide gel.

The occurrence of spatially related Middle Devonian volcanic basic to intermediate rocks, iron-silica ores and reef-related limestones at the Covas Ruivas site can be compared with other occurrences elsewhere in the European Variscides. The Lahn and Dill synclines in the Rheno-Hercynian zone (Germany) show extensive Devonian volcanic rocks mostly of submarine facies (BREITKREUZ & FLICK, 1997; FLICK *et al.,* 2008; NESBOR *et al.,* 1993). During the Middle Devonian several reefs developed in areas of volcanic islands and seamounts (FLICK *et al*., 2008; KÖNIGSHOF *et al*., 2010), as recorded by reef and peri-reefal facies. Several iron ore occurrences are known from the same synclines (Lahn-Dill ores) which had economic importance in the recent past (FLICK *et al*., 1990; 2008). The resemblance of the petrology of these ores with the ones from the BIC is striking. The same type of ores is also present in the Moravian-Silesian part of the Bohemian massif within thick Middle Devonian volcanic successions (VÁCLAV KACHLÍK, pers. com.). It thus seems that a similar tectono-magmatic-sedimentary setting prevailed during the Lower-Middle Devonian in wide spread areas of what are now the European Variscides.

The precise dating of the limestones and the definition of their stratigraphical position on top of the LVS will allow, once the geochemical affinities of the magmatic rocks are known (*work in progress*), to better constrain the evolution of the Southern domains of the Ossa-Morena Zone.

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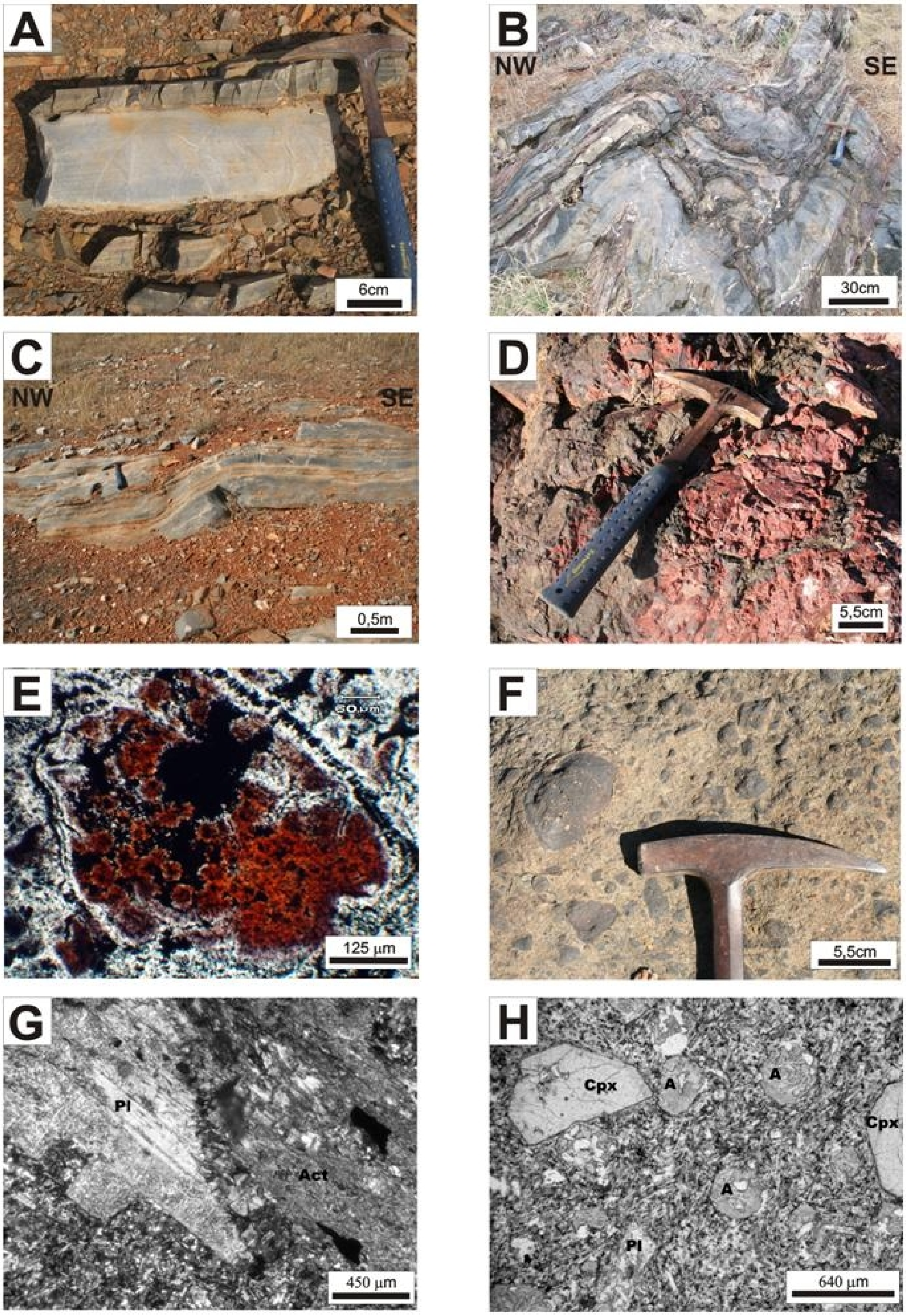

Plate 1 – Field and microscopic images of the Odivelas Volcano-sedimentary sequence (Covas Ruivas locality)

A. Graded bedding in limestone. There is a tuffitic bed on the top.
B. Fold showing deformation accommodation for fault.
C. Family 2 fold, affected by two small dextral strike.slip faults.
D. Jasper deposits in volcanic rocks.
E. Spherullites structures, in jaspers, envolved by hematite (microscope image).
F. Detail of a coarse pyroclastic deposit.
G. Relict phonocrystal of albitized plagioclase (Pl) and prophyroblast of actinolite (Act) set on a fine grained matrix mainly formed by feldspar.
H. Blasto-porphyritic rock showing relicts of clinopyroxene phenocrystals (Cpx) and of albitized plagioclase (Pl) set on a feldspatic matrix. Amygdalaes (A) of epidote and chlorite are also visible.