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Genesis of deformation structures affecting Plio-Pleistocene sediments in the western Portuguese mainland (West Iberia). Implication on the regional neotectonics

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Genesis of deformation structures affecting Plio-Pleistocene sediments in the western Portuguese mainland (West Iberia). Implication on the regional neotectonics

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Abstract

A careful analysis of the morphology, geologic setting, stratigraphic distribution and sedimentological context of deformation features in sediments can give valuable information for the interpretation of their genesis, namely if they are of sedimentary or tectonic origin. This approach was used for the study area, located in the western region of the Portuguese mainland (West Iberian margin), where significant neotectonic and seismic activities occur, contrasting with the stabler interior of Iberia. Notwithstanding the regional neotectonics (intended as upper Pliocene to Recent), part of the deformation features observed in the studied deposits (Pliocene to Pleistocene in age) may be attributed to other phenomena but seismic shaking and/or active faulting. Slump beds and flame structures in coarse sand deposits are probably related to sedimentary overloading in high supply delta plain setting. Part of the deformation observed in the sediments that overlie Jurassic marls and limestones was probably induced by karst sink. This process is plausible even in Plio-Pleistocene sediments that overlie low carbonate content basement rocks, such as marl dominated successions. Thus, the abundance of deformation features affecting the studied Plio-Pleistocene sedimentary cover may lead to an overestimation of the regional active tectonics. The neotectonic activity in this area is reconsidered from the interpretation of the triggering mechanisms responsible for the observed deformation.

Keywords: Neotectonics; Soft sediment deformation; Karst sink; Seismotectonics; West Iberia margin

1. Introduction

The generation of deformation structures in sedimentary bodies requires the combination of several conditions. In the case of soft sediment deformation, it is necessary a triggering mechanism, a driving force responsible for reducing and/or surpassing the strength of the sediment allowing it to be disrupted, and a deformation process [1, 2]. The most common processes responsible for soft sediment deformation are liquefaction and fluidization [3]. Several driving forces can promote the deformation (unstable density gradients, lateral shear stress, gravitational forces, etc.), defining its final geometry [1]. The interpretation of the triggering mechanism is more problematic and has been the subject of several studies

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during the last years [2, 4, 5, 6, 7, 8]. It is even considered that the characterization of the triggering mechanism with a degree of certainty is a difficult and virtually impossible task, though it can be assessed by using criteria that exclude the less plausible mechanisms [2]. A detailed facies analysis and the reconstruction of the depositional environments for the deformed sediments can give valuable information on potential deposition related triggering mechanisms [4, 5].

Inference of the triggering mechanism is an essential task when dealing with the seismotectonic implications of deformation structures that can be attributed to neotectonic activity. Actually, interpretation of seismic induced structures, in particular the evaluation of their seismotectonic implications, is a matter that has received major attention from Earth Scientists [4, 5, 8, 9, 10]. Several authors even attempted to estimate the magnitude of the seismic shocks from the induced deformation features [10]. However, similar deformation structures can be related to other phenomena besides tectonic activity [2, 4, 5, 8, 11]. Sediment slide due to overloading and karst sink are two alternative processes.

The deformation observed in the studied sediments of Pliocene to Pleistocene age affects several sedimentary facies, from continental and coastal environments. It reveals both brittle and ductile rheology, and the two types are often found in the same outcrop. Deformation structures related to shortening and extensional strain regimes are also common, being helpful to infer sin-depositional and post-depositional deformation. In some places it was possible to recognize different deformation phases.

Although evidences of deformation affecting the upper Neogene cover sediments are widespread in the study area, the maps of instrumental seismicity and epicenter location of historical earthquakes indicate that the study region is presently subjected to a relatively low tectonic activity, at least when compared to the southern Iberia region [12, 13]. Two alternative hypotheses may be placed: either there are other mechanisms besides tectonic activity responsible for the observed deformation features; or the current and historical times constitute a period of seismic quiescence, reflecting characteristic earthquake behavior of individual seismogenic structures with long recurrence intervals. Hence, the determination of the triggering mechanisms responsible for the deformation can help to characterize present day seismic hazard. The genesis of the deformation is evaluated from the analysis of the geologic setting, the stratigrahic distribution and the depositional conditions of the deformed sediments, coupled with the geometry and kinematics of the outcropping structures.

2. Geologic setting

The studied Pliocene to Pleistocene sediments unconformably cover diverse (Triassic to Cretaceous) rocks of the Mesozoic Lusitanian Basin, and Precambrian and Paleozoic rocks of the Variscan Iberian Massif (Fig. 1), which constitute the geological basement of the central-

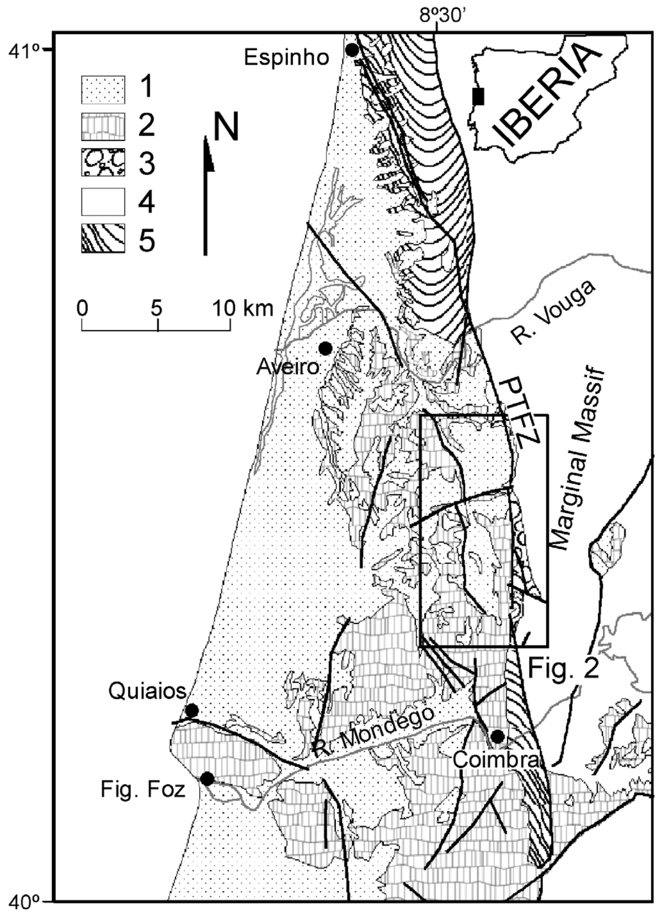


Fig. 1: Geologic setting of the study area (based on Oliveira *et al.*, 1992 [12]). **1**: Cenozoic; **2**: Mesozoic; **3**: Permo-Carboniferous; **4**: Precambrian and Paleozoic of the Central Iberian Zone of the Hesperian Massif; **5**: Precambrian and Paleozoic rocks of the Ossa Morena Zone of the Hesperian Massif; PTFZ: Porto-Tomar Fault Zone.

western sector of the Portuguese mainland territory. The boundary between the Variscan Iberian Massif, to the East, and the Mesozoic and Cenozoic sedimentary basins of the Atlantic margin of Central Mainland Portugal, to the West, corresponds to a kilometric wide, N-S trending structure named the Porto-Coimbra-Tomar fault zone (Fig. 1) [14]. This structure is a wide deformation zone with complex internal geometry that is related to Variscan and Alpine deformation. A vertical displacement of 100 to 200 meters since late Pliocene, along the contact between the Variscan Iberian Massif and the Mesozoic and Cenozoic sedimentary basins of the Atlantic margin is deduced from geomorphologic data [15, 16]. This displacement is responsible for the uplift of a roughly N-S trending ridge, named the Marginal Massif [17, 18]. The most frequent faults in this area are N0±20º trending sub-parallel faults. These structures are displaced by other, N20-70º oriented, faults. There are also N320±10º trending faults which, together with the former structures, delimit several subsiding areas expressed as low-lands in the topography (Fig. 2). Complex tilt movements, both parallel and transversal to the main structures,

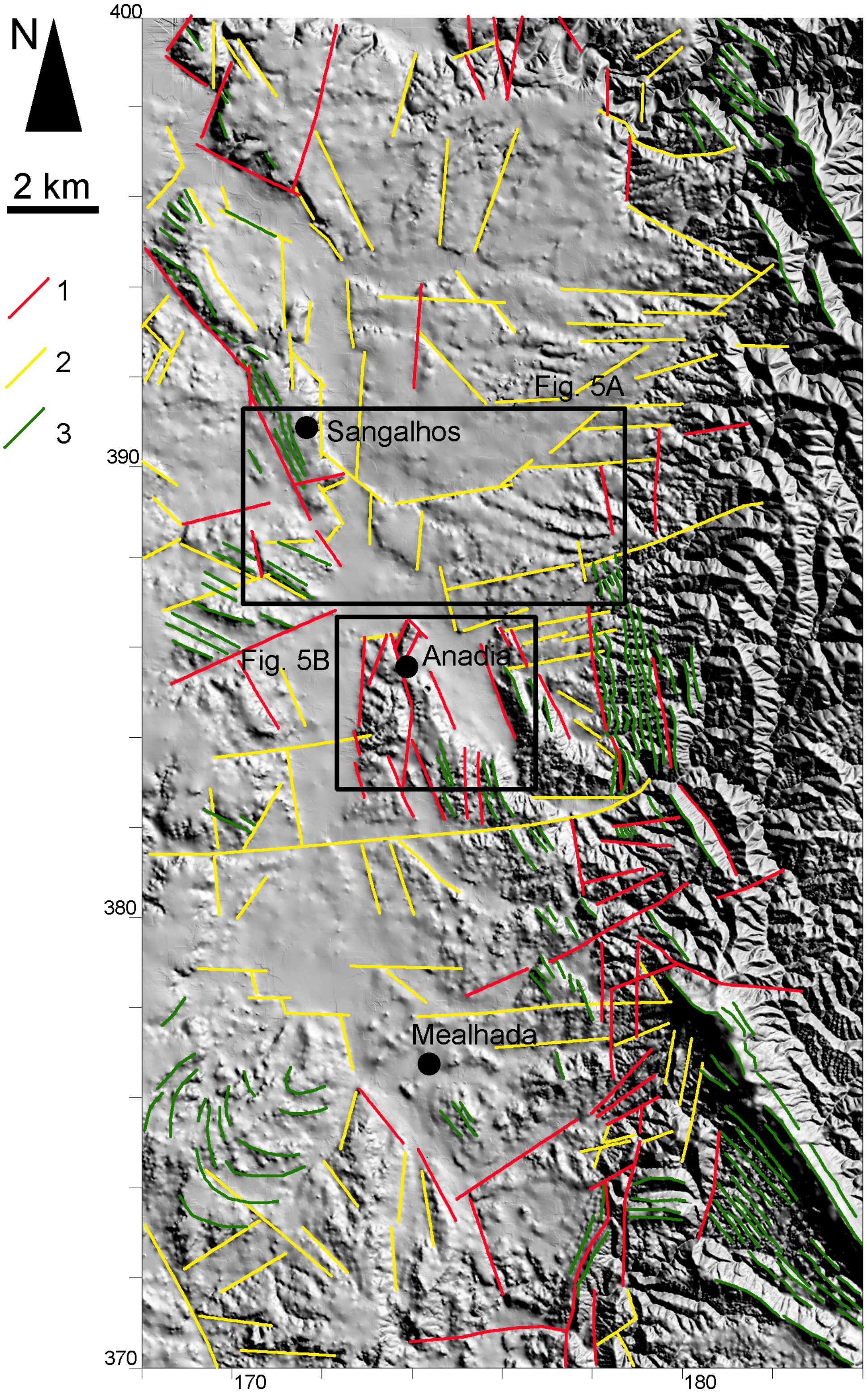


Fig. 2: Digital elevation model (DEM) of the study area and interpreted structural alignments. The topographic alignments may be associated with field recognized faults (**1**), probable faults or megajoints (**2**), bedding planes (**3**). Gauss coordinates.

are observed in association with the intricate fault pattern [19]. These tectonic movements affected the sedimentation and paleogeographic evolution of the area [19].

2.1. Tectonic framework

The study area is sited at the West Iberia continental margin, ca. 500 km north of the plate boundary between the Eurasian and the African (Nubian) plates, which extends roughly W-E from the Azores archipelago to the Straits of Gibraltar domain. The plates' interaction is characterized by a NW-SE to WNW-ESE convergence at a rate of 4-5 mm/yr that is responsible for an important tectonic activity, which is distributed across a 200-300 km wide deformation belt defining a plate boundary zone at the southern edge of Iberia [20, 21]. Intensive marine geosciences studies that have been performed since the 1990's, mostly based on seismic reflection data and detailed bathymetric surveys, have revealed important active faults that are the sources of the interplate seismicity, generally moderate but characterized by the occurrence of very large, although infrequent, events, as the $M_W > 8.5$, 1755, Lisbon earthquake [22, 23, 24, 25].

The epicentres distribution (Fig. 3) shows that Iberia has a nucleus with very low seismicity which is surrounded by margins with higher seismic activity, particularly the aforementioned southern margin and its eastern prolonging along the Betics, corresponding to the Iberia-Núbia plate boundary zone. Significant seismic activity also occurs to the east, along the Valencia trough and the Catalan ranges, in the north, along the Pirenean-Cantabrian margin, and in the west, along the West-Iberian margin.

This pattern of earthquake distribution points to the individualization of an Iberia microplate which is presently interacting with outlying lithosphere along its margins [16, 27, 28]. The Portuguese mainland and the study area in particular, are located at the seismicity belt of West-Iberia, in a

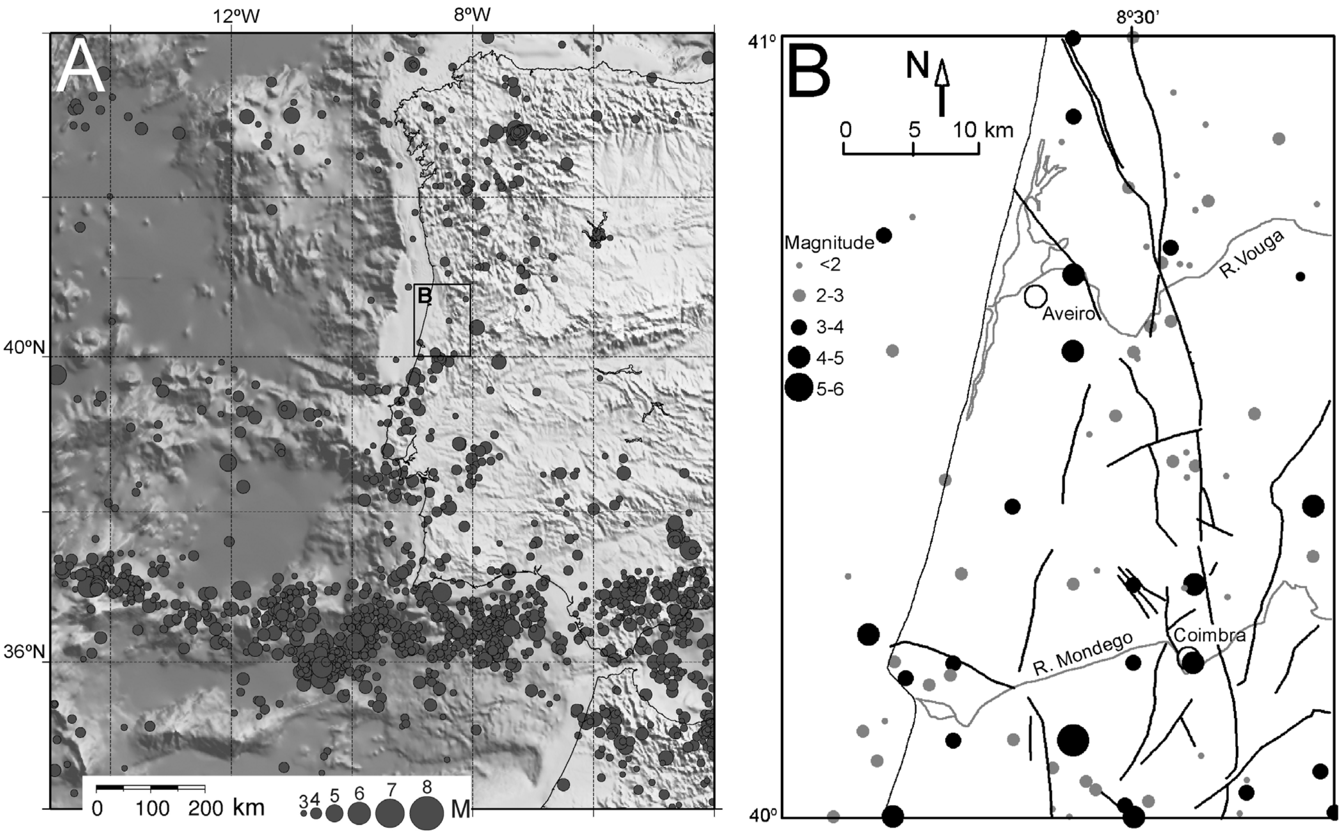


Fig. 3: Instrumental seismicity (based on Martins and Mendes-Victor, 2001 [26] and updates). **(A)** Seismicity recorded in West Iberia region during 1964-2007. **(B)** Seismicity in the study area and surroundings (area represented in figure 1) for 1912-2005. Notice that although the study area shows some seismic activity it is not as intense as in Lisbon and South Iberia-North Africa regions.

region where the Iberia-Núbia interaction along the Azores-Gibraltar, and the Atlantic-Iberia interplay along the Iberian western border result in a significant neotectonic activity, characterized by the presence of low slip rate active faults (<0.1-0.5 mm/yr; "slow faults") [16, 27, 28]. At least some of these faults have the potential of producing large, strongly damaging earthquakes, although with long return periods, which are absent in the human registry.

2.2. Plio-Pleistocene stratigraphy

The Plio-Pleistocene sedimentary succession in the study area constitutes a prograding sequence deposited in different environments, ranging from inner shelf to alluvial fans (Table 1, Fig. 4). It can be organized in four stratigraphic units that precede the present day fluvial incision [19]. Unit 1 (usually <5 m) is represented by well sorted, fine to very fine sand-size inner shelf deposits, which can pass upward to coarser coastal sediments. This unit is related to the Piacenzian transgressive phase that affected the entire West Iberia coastal margin. Unit 2A (2-20 m) consists mainly of sand and gravel fluvio-deltaic deposits that unconformably cover Unit 1. It is related to coastal progradation enhanced by high sediment supply. Unit 2B is preserved in the eastern, more proximal locations. It is made of fine grained mud deposits, sometimes with lignite beds, intercalated with immature sand and gravel deposits. This unit records lacustrine and alluvial deposition in a North-South elongated subsiding area. Unit 2B may be absent or reach 50 m in asymmetrical grabens or semi-grabens, indicating that its deposition was conditioned by fault related generation of accommodation space. Unit 3 (<5 m) is made of coarse grained alluvial deposits. This unit consists of an alluvial fan succession sourced from the Marginal Massif. Unit 3 extends across the limits of the North-South structural low sector and can be found in western, uplifted, locations. It marks the transition to the incision of the present day fluvial drainage system.

The available chronostratigraphic data for the studied deposits are imprecise. The lower transgressive deposits of Unit 1 have been considered equivalent to the upper Zanclean to lower Piacenzian fossil rich beds found in southern locations of the western Portuguese coastal area [29]. This hypothesis of equivalence is based on the presence of similar inner shelf facies [19]. The palynological content from a lignite bed of Unit 2B points to Piacenzian age [30]. Updating works on the microfossil content of Unit 2B are now being conducted.

Late Pliocene to early Pleistocene sedimentary evolution had clear structural constraints. As a consequence of the tectonic role on the generation of accommodation space, the sedimentary architecture of Pliocene and Pleistocene sediments is strikingly different in uplifted and subsiding sectors [19]. Alluvial and lake deposits (Unit 2B) are preser-

Table 1: Facies characteristics (based on Dinis, 2004 [19])

Facies	Lithology	Geometry	Structures	Environment
FA	Very well sorted fine-grained sand; Intercalations of gravelly sand.	Tabular, laterally persistent; up to 2m thick.	Massive, horizontally laminated or ripple laminated.	Inner shelf.
FB	Moderately to well-sorted, fine sands to gravelly sands; dispersed well rounded pebbles.	North-south ribbons, lobes or tabular. Crops out westward and under thick mud sediments; up to 10m thick.	Oblique cross-stratification; Cases of polymodal palaeocurrent pattern; Cases of Gilbert type profile with basinward (W or NW) paleocurrents.	Dune and bar in coastal submarine and estuarine settings.
FC	Varied moderate to poorly-sorted sands and gravels; Mixture of well rounded and angular clasts.	Channel packages. Up to 2m thick and tens of meters wide.	Structureless, imbrication or cross-stratification; Normal grading.	Fluvial channel.
FD	Mud with intercalations of sandy mud and thin sands. Lignite beds.	Sheets, lenses or lobes (0.1-10m thick).	Structureless, ripple lamination or parallel lamination.	Lagoon, shallow lake and floodplain.
FE	Polymictic, poorly sorted, matrix or clast supported coarse gravel; Intercalations of sandy gravel.	Sheets or lobes (<3 m thick).	Structureless or crude horizontal bedding.	Alluvial fan.

ved exclusively in a N-S strip along the contact between the Variscan Iberian Massif and the Lusitanian Basin (Fig. 4). In this area there are several examples of late Pliocene to Pleistocene paleo-drainage paths along narrow grabens and of streams deflected northward against uplifted structural blocks [19]. Westwards, the sedimentary record is dominated by shelf and coastal plain sand and gravel deposits of units 1 and 2A. In places, Unit 2A lies directly over the older sedimentary basement. Here the orientation of the fluvial channels seems to be consequent with a general westward tilting of the coastal plain (Fig. 4). Due to relative uplift, the rate of generation of accommodation space must have been much lower in this region than in the eastern area and, therefore, amalgamated coarse grained deposits are dominant. These facts point to an important structural control on the organization of the sedimentary systems.

3. Deformation features

Although evidences of deformation are widespread in the study area, they seem to be more frequent in the Anadia and Sangalhos regions (Fig. 2). Deformation structures were observed in all stratigraphic units. The deformation features can be arranged on the basis of their outcrop geometry and kinematics, stratigraphic distribution and facies of the deformed sediments, and tectonic context and nature of the underlying basement rocks. In this paper we consider four groups of deformation structures.

3.1. Folding and fracturing

3.1.1. Tectonic coherent

Tectonically coherent fracturing and faulting features are found in the vicinity of regional structural alignments (Fig. 5). These features occur independently of the basement rock nature. They may affect only the older or all the Plio-Pleistocene sedimentary units. Most of the outcropping faults are sub-parallel or may be genetically related to mapped structural alignments. The offsets along fault planes are varied, but usually less than 1 m. Faulted contacts with thrusting of basement rocks over Pliocene sediments were recognized (Fig. 6A). In several places the Neogene sedimentary cover is densely fractured with some fractures showing upwards branching (Fig. 7). Fractures with no visible offset (joints) usually show a more scattered distribution. Particular sets of joints located in the proximity of larger faults were interpreted as tensile fractures resulting from slip in the major structures.

Evidences of strike-slip displacements, coupled with the occurrence of a high frequency of deformation features close to regional tectonic alignments, the low variability of their direction, and coincidence with the trend of those alignments, suggest that these structures were induced by tectonic processes. The high density of fractures in the Plio-Pleistocene cover, with upward branching, suggests that the surface expression of the active basement faults is diffuse, due to the distribution of the offsets along various slip surfaces in deformation bands.

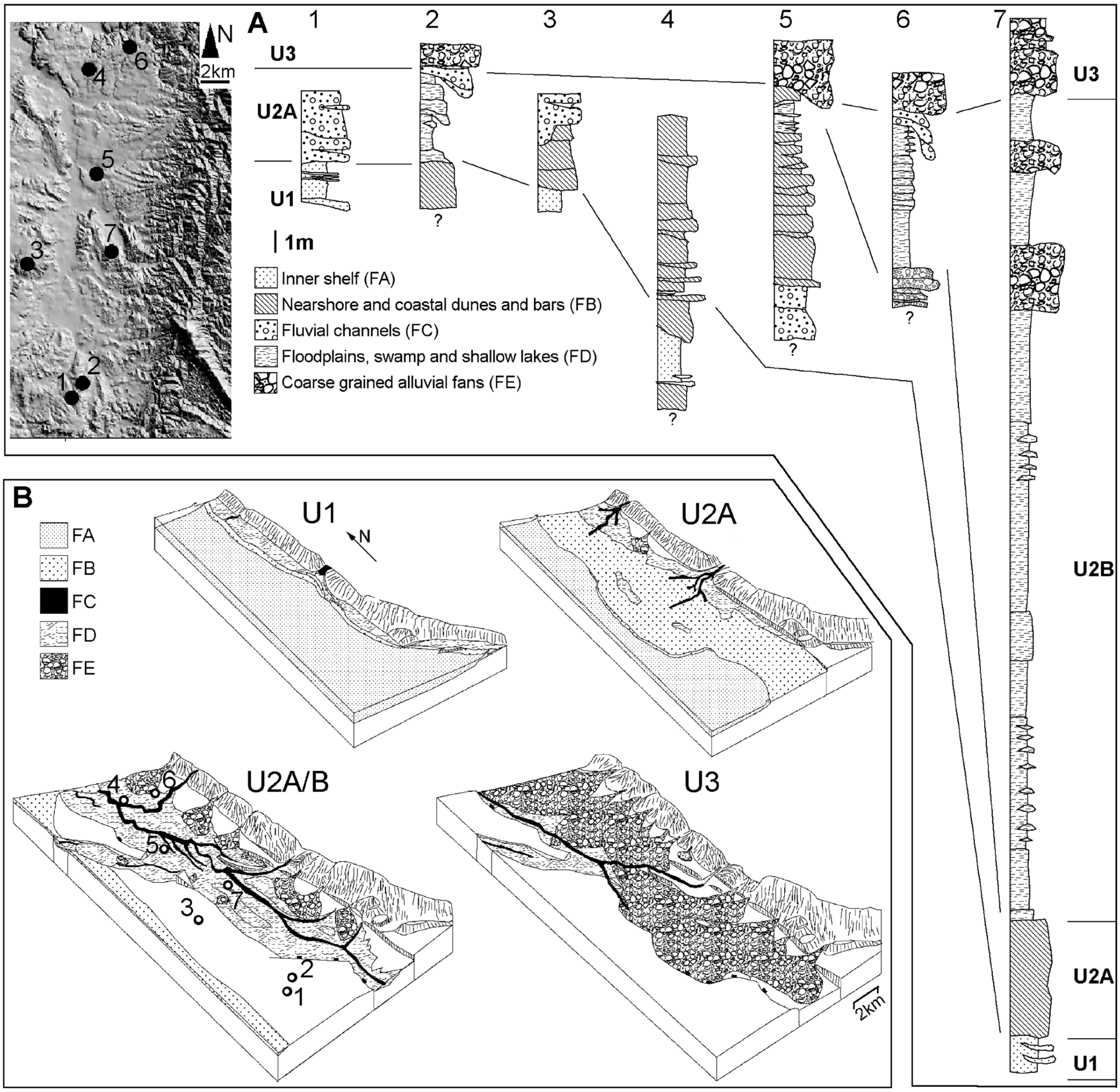


Fig. 4: Selected lithostratigraphic columns for the studied deposits (location in the DEM), with 5 differentiated facies **(A)**. Interpreted paleogeographic evolution comprising four stages **(B)**. During an initial phase, a transgression affected the majority of the Atlantic coastal margin (U1). The following progradation created a coastal plain mainly with fluvio-deltaic sediments (U2A). Tectonic movements developed two contrasting realms: an overall uplifted sector to the west and a subsiding sector to the east (U2A/B). The sedimentary record ends with alluvial fan deposits, mainly derived from the uplifted Marginal Massif (U3).

3.1.2. Non-tectonic

Non-tectonic fracturing and faulting are common in outcrops of Plio-Pleistocene sediments overlaying marl or limestone Jurassic basement rocks (Fig. 5). Locally the unconformity between the siliciclastic cover and the marl and limestone basement is outlined by a decalcification mud veneer (Fig. 8). In these outcrops the Plio-Pleistocene sediments are affected by many joints and small faults. A few fractures are rooted in the decalcification layer at the contact between the clastic cover and the carbonate basement. As a general rule the fracture density is higher in the clastic sedimentary cover than in the carbonate basement. Most fractures are vertical or almost vertical although they present highly scattered orientations and an intricate pattern is common (Fig. 9A). Normal offsets are the most frequent.

In several locations, Plio-Pleistocene units are seen in mechanical contact with Mesozoic marls and limestones

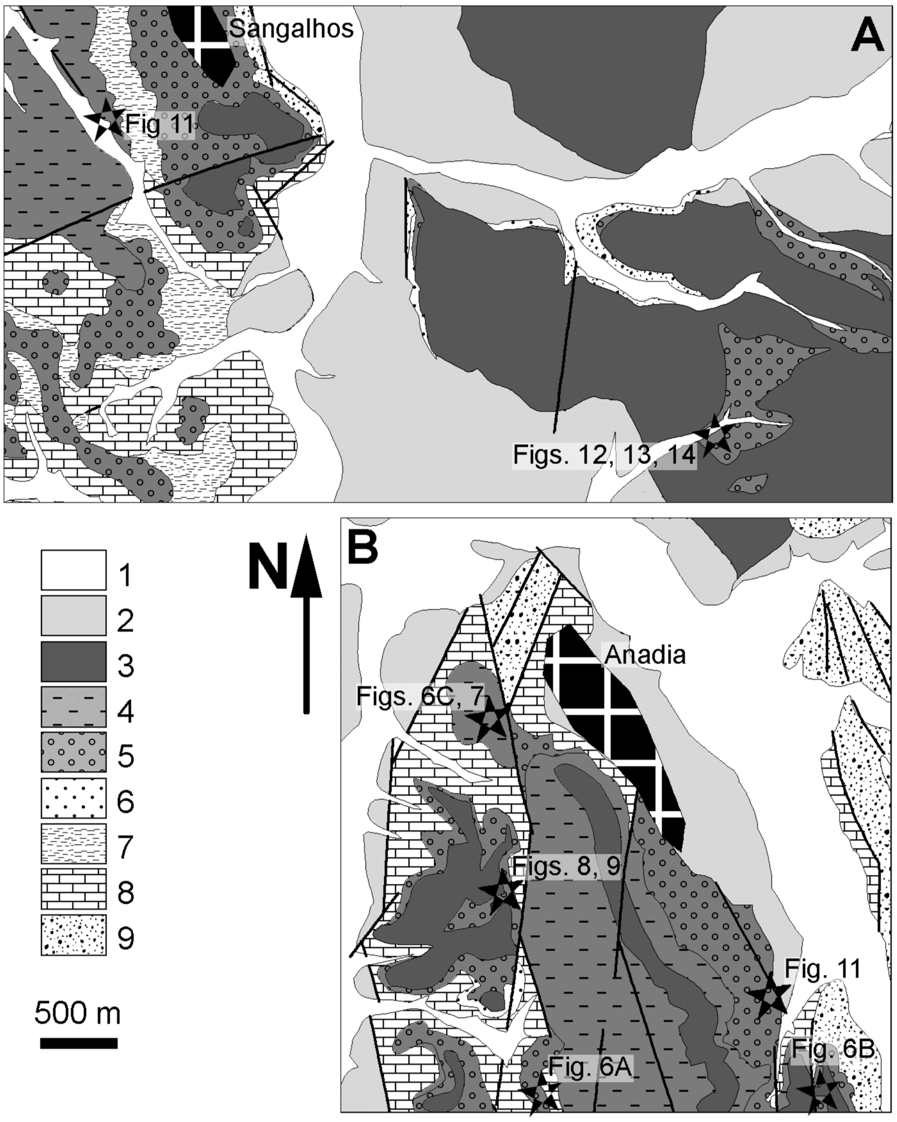


Figure 5: Geologic maps of sectors with frequent deformation features, and location of pictured sections. **1**: Alluvium; **2**: Terraces; **3**: Unit 3; **4**: Unit 2B; **5**: Unit 2A; **6**: Unit 1; **7**: Cretaceous siliciclastic rocks; **8**: Jurassic marls and limestones; **9**: Triassic siliciclastic rocks. Location of mapped areas in figure 2.

(Fig. 10) and the clastic cover is folded in a few hundred meters wide, low amplitude anticlines and synclines.

Plio-Pleistocene deposits are also found infilling dissolution holes in Jurassic basement rocks (Figs. 9B and 9C). When viewed in outcrop, these cavities define meter-scale wide and deep voids along the contact between the Cenozoic clastic cover and the Jurassic, or constitute isolated holes in the Jurassic rocks. The cavities may be almost spherical or elongated in section view, intersecting the Jurassic basement stratification and usually developing parallel to joints that affect these rocks (Fig. 10C). The clastic infill constitutes a chaotic material with massive fabric and it is frequently not easy to recognize the typical appearance of the parent Plio-Pleistocene units.

A few aspects, such as the variable orientation of the deformation structures in the sedimentary cover, the occurrence of more deformation features in the Plio-Pleistocene cover than in the Jurassic basement rocks, the occurrence of faults rooted in decalcification clayey horizons and the presence of cavities filled with material identical to (or reworked from) the sedimentary cover, all favor the premise of a deformation triggered by cryptokarst induced subsidence. The often disorganized structure of the cavities infill suggests intermixing of the Plio-Pleistocene sand cover as it sagged into the cavities, while the presence of a decalcification clay layer between the permeable siliciclastic cover and the carbonate rocks points to continuing cryptokarst processes [11, 31, 32]. The high permeability and porosity of the siliciclastic sediments and their intermittent water saturation must have played an important role, increasing the weathering potential of the percolation waters. Albeit the karstification of the Jurassic rocks, there is no topographic expression of the dissolution and infilling of the cavities by the clastic overburden, such as the presence of sinkholes. This fact points to a relatively old cryptokarst evolution and to a presently small near surface karstification of the Jurassic rocks.

3.2. Clastic dykes

One of the most common sediment deformation features in the study area are clastic dykes. These are more frequent in the older Plio-Pleistocene units, but may affect the totality or part of the younger Unit 3 sediments (Fig. 6C). In other cases, the dykes are sealed by the younger units. They are common in silty-clayey, sand and gravel lithologies. The more conspicuous dykes can be 0.1 to 0.2 m wide and reach more than 2 m in height, in cross-section view. They are composed of diverse materials, though fine grained, silty-clayey materials prevail. In some outcrops, it is possible to observe the connection with the fine-grained sedimentary bed that fed the dyke. One of the observed fracture related dykes shows complex morphology, narrowing close to coarser sediments, becoming wider in fine grained deposits, where it reaches 50 cm (Figs. 6B and 7). The clastic dykes commonly appear intruding along faults or joints, suggesting that the fractures have guided the intrusion, which was promoted by liquefaction and upwards escape of fluidized sediment.

Although not exclusive, the most outstanding clastic dykes affect clay rich successions of Unit 2B, with intervened mud and gravel beds, deposited in alluvial settings.

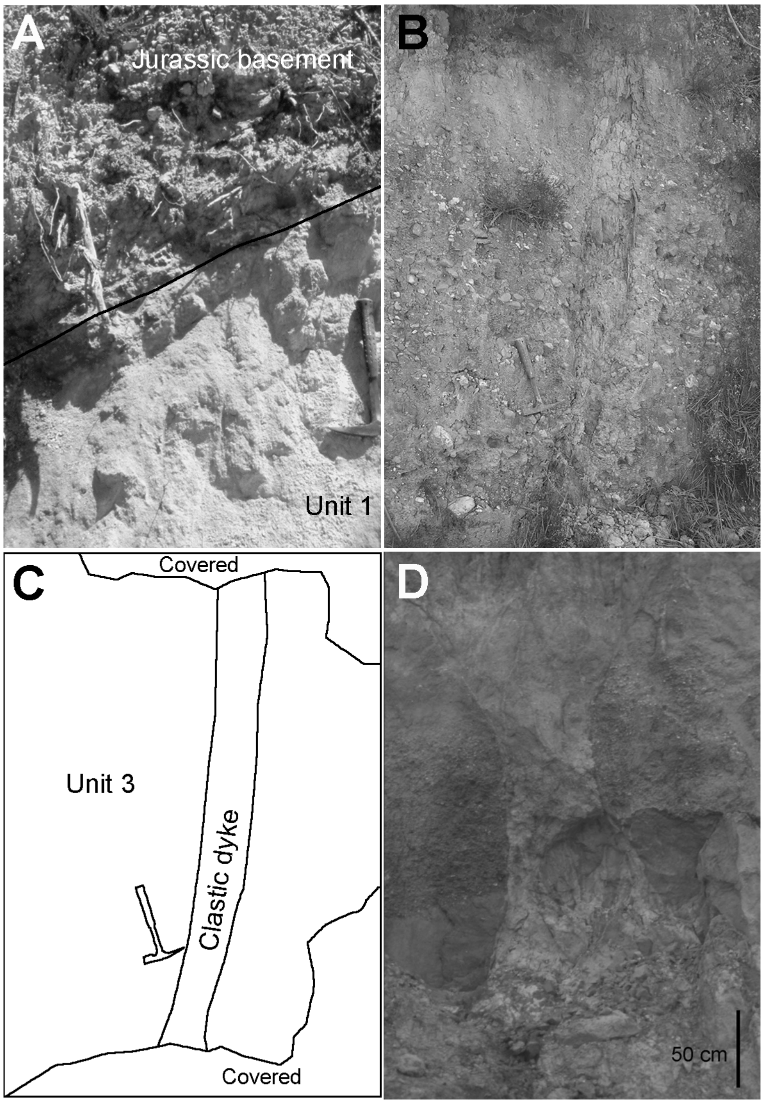


Fig. 6: Examples of tectonic related deformation. Jurassic rocks overthrusting inner shelf (Unit 1) Pliocene deposits **(A)**; Dyke intruding Unit 3 coarse grained sediments **(B)** and interpretation **(C)**; the orientation of this dyke is parallel to an approximately N-S mapped fault; Fine grained clastic intrusion that widens in fine grained host sediments and narrows in gravel deposits in unit 2B **(D)**.

Several aspects support a tectonic origin for these structures, namely: 1) the dykes are found in sediments coeval of the tectonic differentiation of a subsiding sector between the uplifted coastal margin, to the west, and the Marginal Massif, to the east; 2) they tend to have parallel wall geometry and orientation sub-parallel to mapped faults; 3) they are found in the proximity of some regionally important active fault zones. These aspects suggest that at least some of the observed larger clastic dykes are related to liquefaction and fluidization processes induced by seismic events.

Small scale dykes are also found in sandy, fluvio-deltaic, sediments of Unit 2A. In these cases their origin may be attributed to depositional events in which sediment overloading was able to increase the pore pressure enough to generate fluid extrusion and sediment fluidization.

3.3. Diapir-like deformation

In one outcrop located near Anadia, close to the interception of N355º and N330º mapped faults (Fig. 5B), a NNW-SSE elongated, approximately 10 m wide wedge of pre-Neogene mudstones was identified in mechanical contact with surrounding Plio-Pleistocene sands (Fig. 11). The northern contact shows a relatively planar fault geometry trending N340º, 70ºNE, which is sub-parallel to one of the mapped faults direction mentioned above, and presents low dip slickenside striations (pitch 10º to 20º S). The Plio-Pleistocene sediments are folded in anticline, dipping away from the mudstone body, and are affected by two fracture systems. Some fractures measured in this sedimentary cover are sub-parallel to the main mechanical contact with the argillaceous core (N340º, usually dipping ENE), although most of the fractures in the surrounding Plio-Pleistocene sands are high angle normal faults, trending N015º-035º, 55º-70º NW, with offsets of a few centimeters to decimeters. These fractures often present a clay fill that can reach 10 to 25 cm in width. Some faults with listric geometry were also found (Fig. 11).

The occurrence of exotic clayey material in the nucleus of roughly antiformal structures such as the one described above suggests that the cover sediments were affected by ductile flow of underlying pre-Neogene clayey sediments protruding along fracture weakened locations. The clayey material probably "intruded" the sediments above and generated bending folds (forced folds) in the cover deposits. The normal faults, sometimes with listric geometry, must have been triggered by the ductile flowage of the "intruding" material, for space accommodation.

3.4. Slumping

Slump beds are occasionally found in well rounded, coarse sand grain-size delta lobe facies of Unit 2A. They were evident in an outcrop located east of Anadia, exposed along an approximately 1 m thick deformation band extending for at least 100 m (Figs. 12 to 14). Slumping structures affecting a discrete decimeter thick mud horizon with intense deformation were easily recognized within this deformation band (Fig. 12), which coincides with a slight shift in the direction of the coarse sand deposits accretion foresets, from a mean N270º direction to a mean N305º (Fig. 13). Several low angle detachment surfaces (dipping 5º to 10º) were recognized within the deformation band. These detachment surfaces are associated with a few thrusts and thrust related folds verging approximately N225º (Figs. 12 and 14A).

Under the above referred slump bed there are poorly defined, small scale water escape structures, approximately 15 cm high (Fig. 14B). These structures are identified in disrupted sand bodies that intrude the overlaying cross-stratified sediments. A

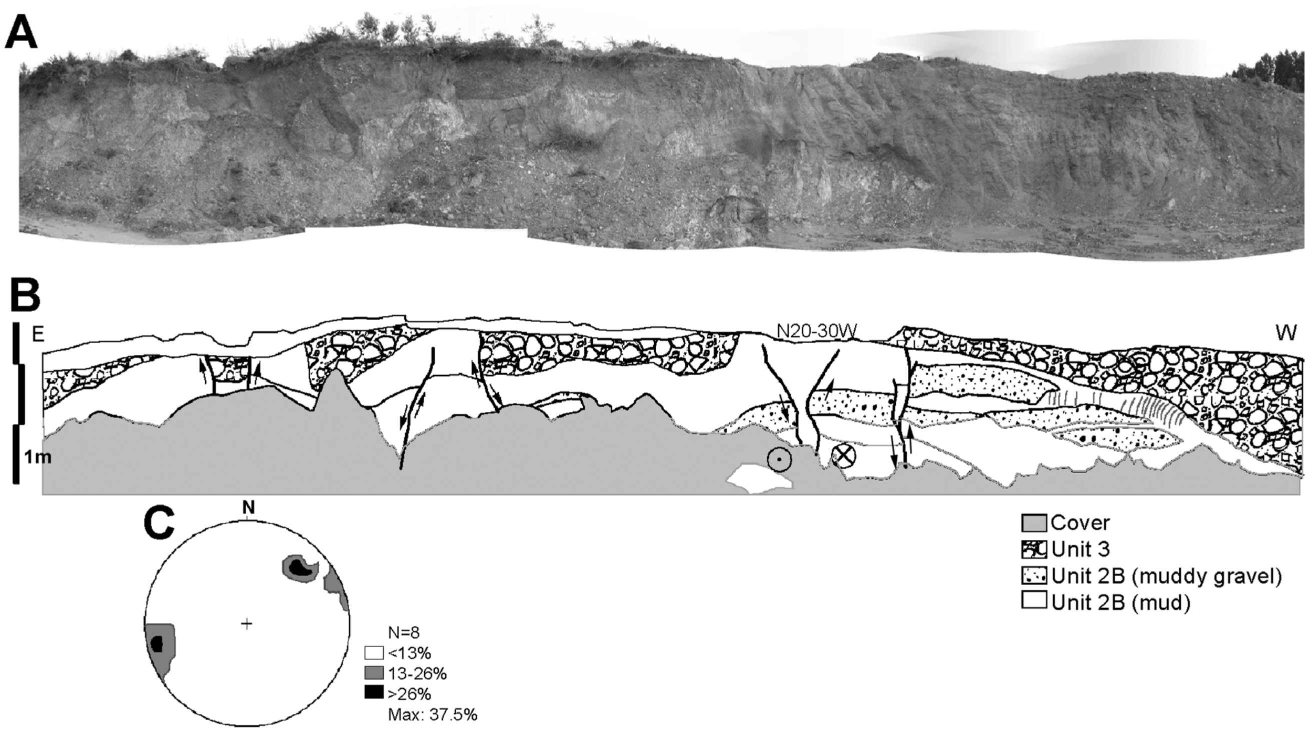


Fig. 7: Photomosaic of Unit 2B and Unit 3 deposits affected by faulting **(A)** and interpretation **(B).** Notice the upward bifurcation of some faults. A detail of N20-30W fault is found in figure 6B. Fractures are represented in a lower hemisphere equal area stereographic plot **(C)**. Note de predominance of fractures trending NNW-SSE to NW-SE. This outcrop is located in the vicinity of a southward tilted horst limited by faults with similar orientation. Location in figure 5.

few compressive and extensional small faults were also observed in the sand unit affected by the slump bed. The difficulty in identifying good reference horizons made uneasy the recognition of faults, which may be more frequent than detected. Nevertheless, some were easily recognized because they cut decimeter long outsized clay clasts. The observed fault offsets are only of a few centimeters and frequently tend to decrease upwards.

The slump structures must be impelled by lateral driving forces, due to the presence of a slope, acting in sediments with plastic behavior [1, 6, 8]. The well developed folds, thrusts and detachment surfaces are indicative of compression in the toe of the slump bed. Because water escape structures were found below the slump structures it may be argued that these two structures were genetically related. Water escape structures can result from the upward movement of water and sediment particles after liquefaction and fluidization in sandy sediments limited by overlying low-permeability levels [33, 34]. In this case they may result from overburden induced by the slumping phenomena..The faults observed in the slumped section cannot be related to karst sink phenomena, since the Plio-Pleistocene sediments stand over non-soluble clastic Triassic sediments. The reverse faults may be directly related to the compression at the toe of the slumps, while the normal faults may result from the space accommodation due to the concomitant lateral spreading.

The slope failure phenomena responsible for the observed deformation could be triggered by several mechanisms like seismic shaking, storms and flood surges, among others [1, 2, 6, 8]. Because these phenomena are common in high sediment influx settings, like deltaic environments, independently of the tectonic activity [35], the genesis of the slump structures in the studied sediments could be attributed to aseismic processes. However, the location of the slump bed in a section close to a regional structural alignment (Fig. 5) and the fact that these delta lobe facies in other sections do not show similar deformation features suggest that this deformation may have been triggered by seismic shaking penecontemporaneous to sedimentation.

4. Genesis of the deformation features

4.1. Tectonic activity

When the geometry and kinematics of the studied mesoscale structures affecting the cover sediments are consistent with the regional tectonic pattern and stress field, they can be related to tectonic activity with some confidence [10, 11]. In what concerns the study region, several geophysical and geological indicators point to an overall NW-SE directed maximum horizontal compressive stress acting across the Portuguese mainland territory (West Iberia Atlantic margin) since late Pliocene [12, 27]. Accordingly, structures showing compatibility with this stress field, particularly those that have some geometrical relationship with, and/or are located close to regional active tectonic structures can be confidently considered of tectonic

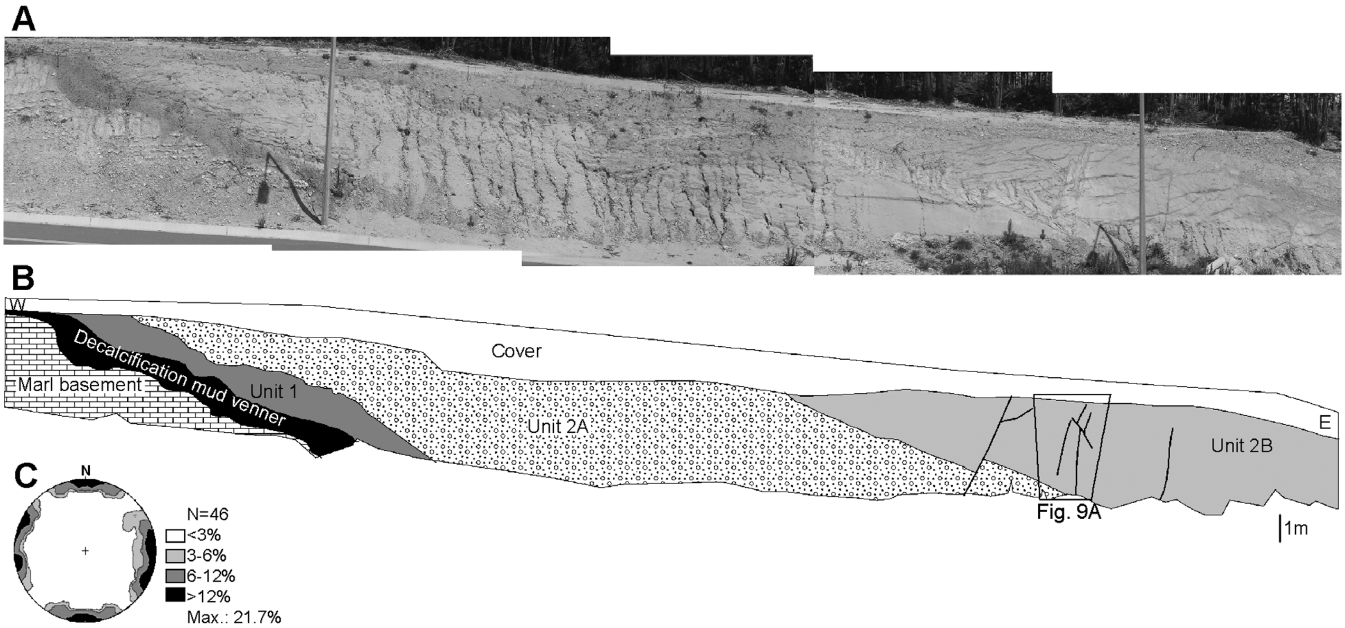


Fig. 8: Photomosaic of tilted Neogene deposits over a Jurassic sedimentary succession dominated by marl rocks (A) and interpretation (B). A mud horizon outlines the contact between the Jurassic rocks and overlying Plio-Pleistocene sediments. Fractures measured at this outcrop are represented in a lower hemisphere equal area stereographic plot (C). Note the high scattering of the fractures orientation. Location in figure 5.

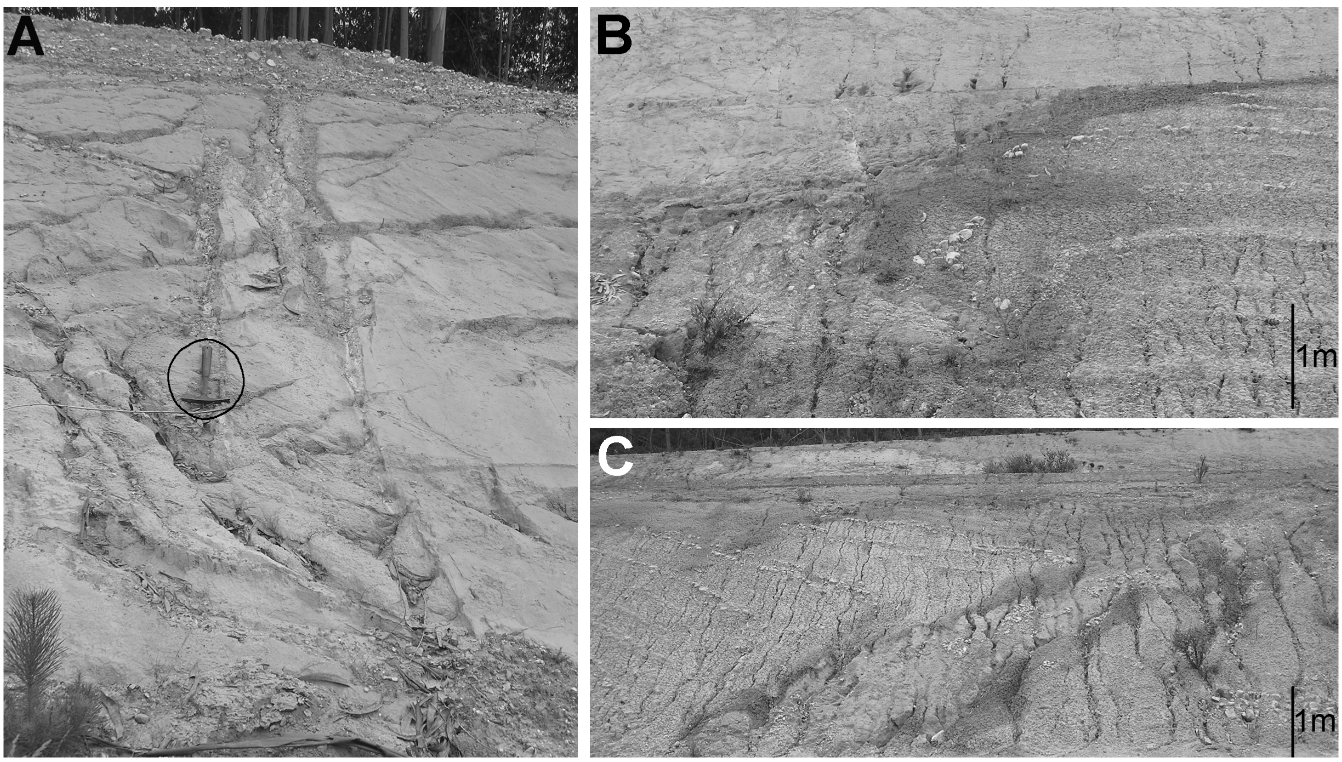


Fig. 9: Detail of some deformation features observed in the outcrop represented in figure 8. An intricate fault pattern can be found **(A)**. Decalcification mud veneer between the Pliocene and Jurassic units **(B, C)**. When cavities have a clastic infill the mud veneer is still found along the contact between the infill and the Jurassic sediments **(C)**. The cavities are elongated diagonally to the bedding plane orientation.

cause. Thus, deformation features in the study area interpreted as tectonic in origin are more frequent in the proximity of some regional structural alignments and active faults, trending parallel or showing a close relationship to these structures, and their kinematics is in agreement with the regional stress field.

There are also some soft deformation structures (sand dykes and slumping) that may be related to liquefaction and fluidization processes triggered by seismic activity. Not withstanding the possibility of a relationship with distant earthquakes, as seismically induced deformation

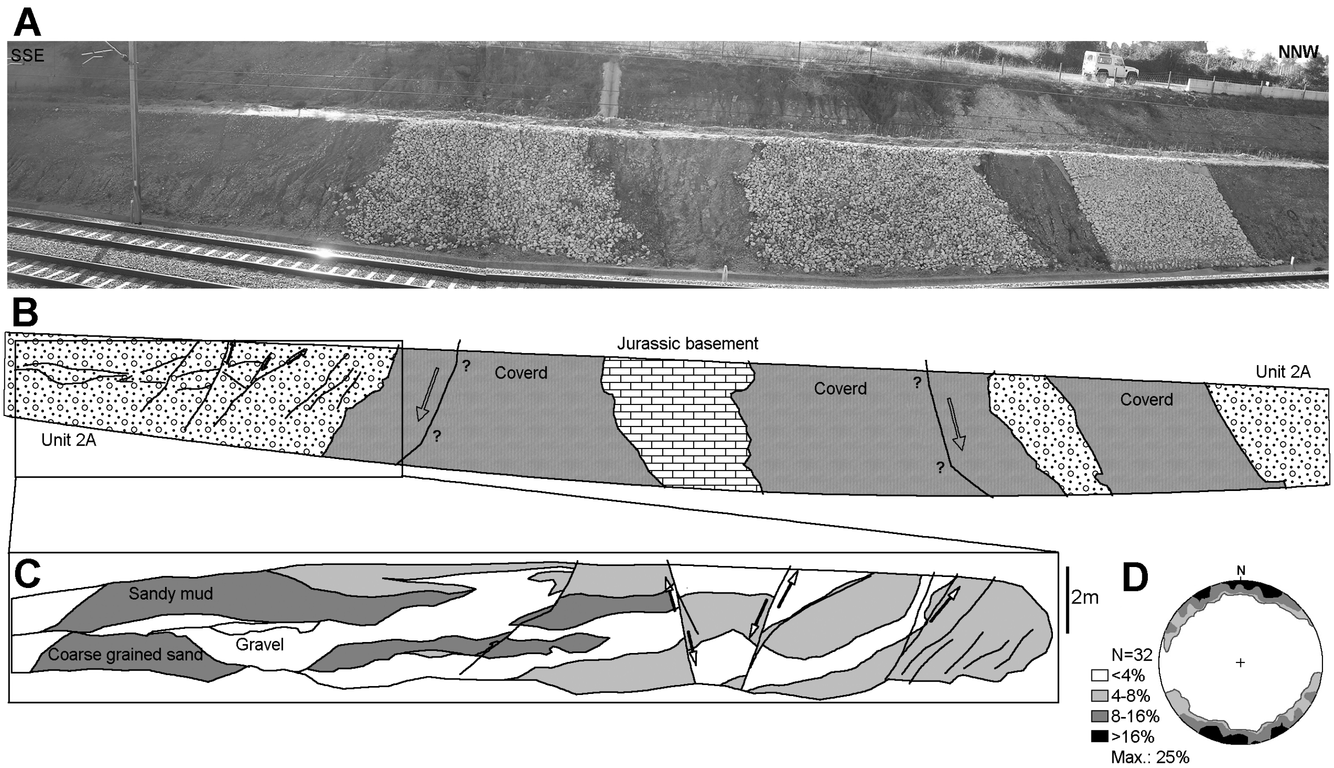


Fig. 10: Photomosaic of deformed Plio-Pleistocene sediments in mechanical contact with a Jurassic marl and limestone basement (**A**) and interpretation (**B**). Detail of the southern portion of the outcrop (**C**). Close to the Jurassic basement the faults density is higher. Fractures measured in this outcrop are represented in a lower hemisphere equal area stereographic plot (**D**). Notice the high scattering of the fractures orientation. Location in figure 5.

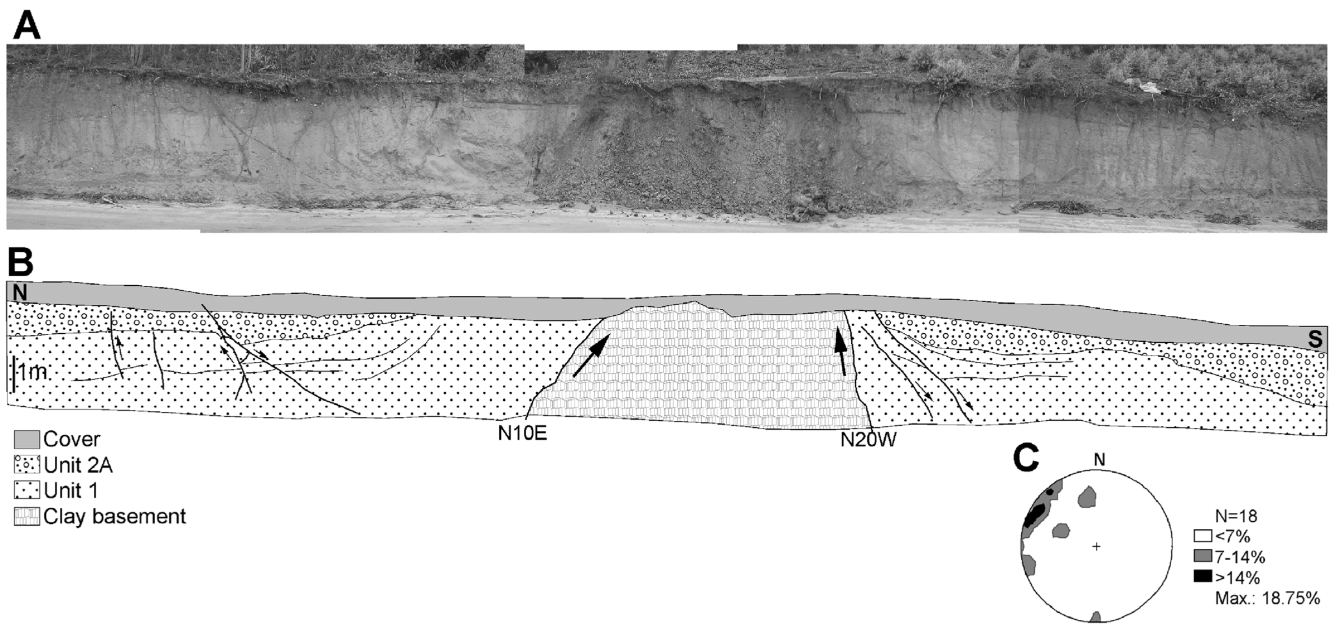


Fig. 11: Photomosaic of Unit 1 and Unit 2A deposits intruded by a pre-Neogene argillaceous body (**A**) and interpretation (**B**). The Pliocene deposits are folded and faulted. The faulting is more intense close to the intruded body. In the southern part of the outcrop it is possible to recognise an antithetic listric fault. Fractures are represented in a lower hemisphere equal area stereographic plot (**C**). Most faults trend NNE-SSW, which is sub-parallel to the direction of the southern mechanical contact of the pre-Neogene clay rich body. Location in figure 5.

may also be related to "far-field" (more than 100 km distant) high magnitude events [4, 5, 36], the proximity to an active structure supports the seismogenic hypothesis, as the deformation features may thus result from lower magnitude, although more frequent earthquakes.

Another evidence of neotectonic activity in the study area is the above referred presence of tectonically triggered plastic flowage processes of pre-Neogene argillaceous rocks, which developed anticline forced folds in the overlying Plio-Pleistocene sediments.

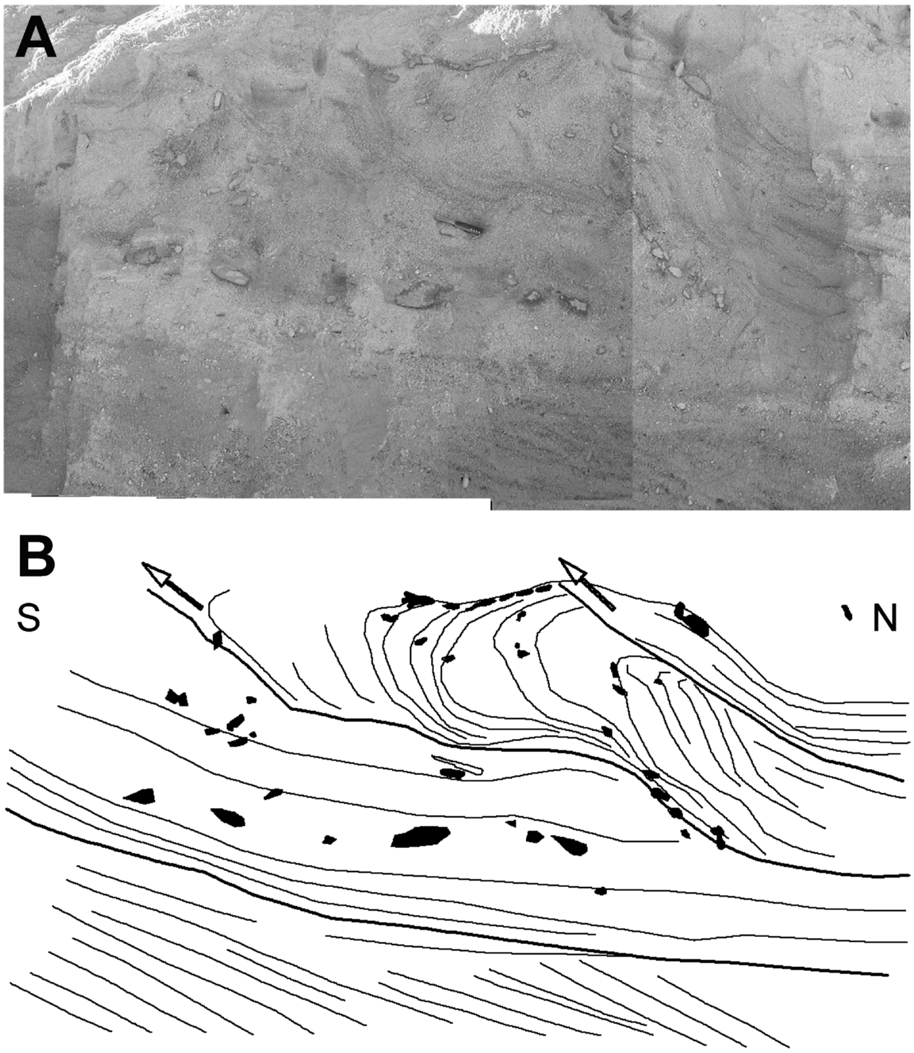


Fig. 12: View of a slump bed affecting foreset accretion deposits **(A)** and interpretation **(B)**. Several detachment surfaces can be observed. A mud bed is fragmented due to the deformation. Knife for scaling. Location in figure 5.

The regional neotectonic activity inferred from the soft sediment deformation structures is also supported by the late Pliocene to present day drainage pattern. At least since unit U2B the trunk drainage tends to be adapted to N-S elongated structural lows; and the present day streams that drain from the east are frequently deflected northward against elevated structural blocks while they enter the complex N-S directed structural corridor (Figs. 2 and 4).

4.2. Slope failure

Slumping structures are usually formed soon after deposition of the sedimentary body and before its lithification [6, 35]. In places where the outcrop structures show a consistent geometrical and kinematical relationship with the accretion of foreset beds, the deformation can be related to sediment sliding in a high sediment supply depositional setting, triggered by seismic shaking, overloading or storm events, among other mechanisms [2, 6, 8].

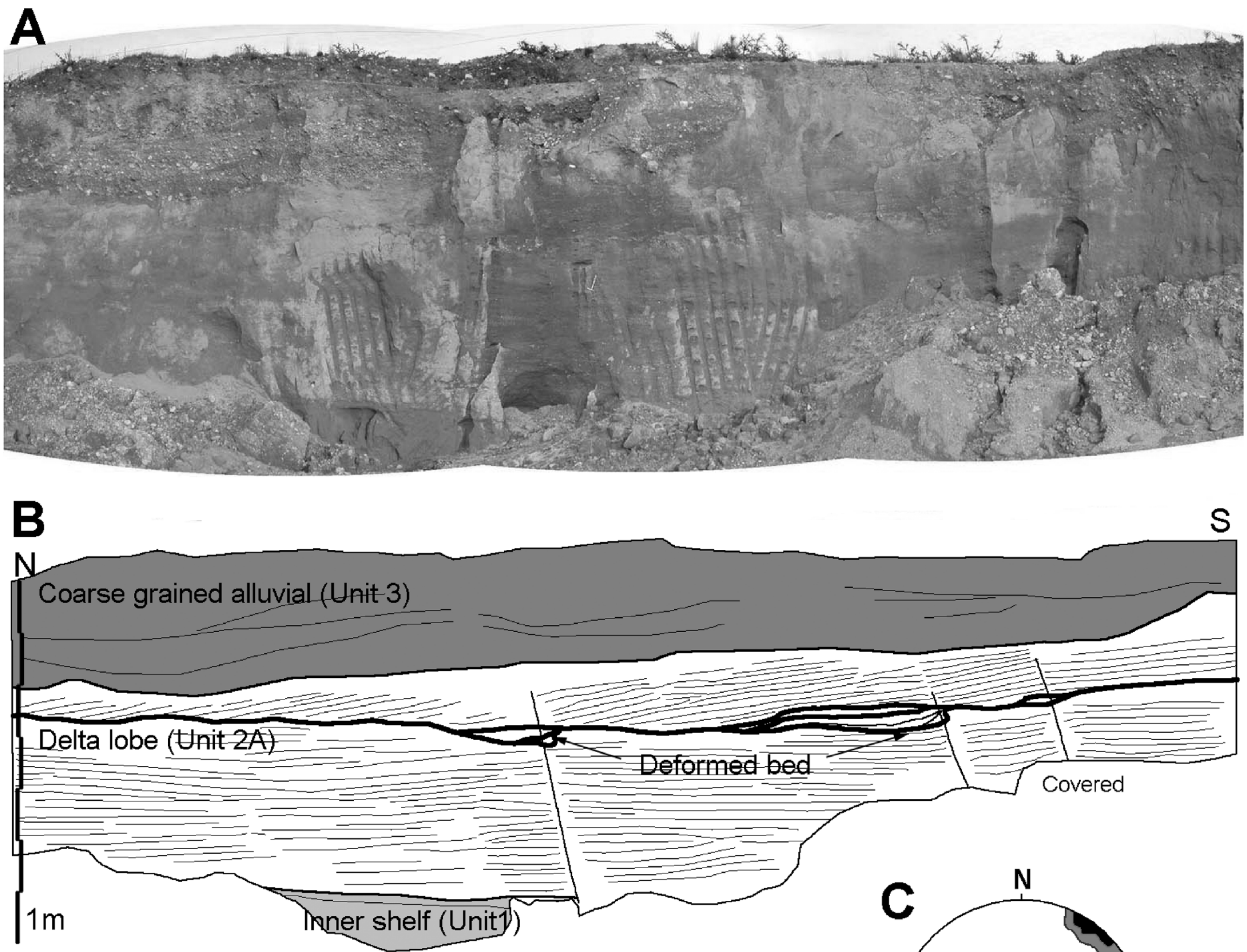
In the studied deposits this group of structures is exclusive of foreset accretion facies deposited in environments with relatively high sediment supply. This is the case of the slump beds in delta lobe facies confined to a specific stratigraphic level. In summary, when fault offsets diminish rapidly upwards, the structures pattern can be related to the direction of foreset accretion and the deformation is concentrated along specific stratigraphic horizons, the deformation may be interpreted as sin-depositional and associated to sedimentary processes.

4.3. Gravitational sink due to cryptokarst development

Studies of the deformation affecting clastic deposits overlaying soluble rocks in southern Portugal have suggested a strong interference of cryptokarst evolution on the development of deformation structures in the cover sediments [11]. Effectively, though the south Portuguese region shows relatively intense, instrumental and historical, seismic activity (Fig. 3), that interpretation was considered the most likely for the majority of the regional deformation structures affecting the Plio-Pleistocene deposits [11].

As a general rule, in the present study area the dissolution related deformation is characterized by a high variability in the orientation of the fractures, and the faults and folds geometry and kinematics are not compatible with the tectonic structure of the underlying Mesozoic rocks and the regional tectonic alignments, neither with the regional stress field. Other distinguishing features are the occurrence of more intense deformation in the Plio-Pleistocene cover than in the carbonate basement rocks, the presence of a clayey veneer between the carbonate rocks and the Plio-Pleistocene cover and the occurrence of faults rooted in this clayey veneer. As a result of dissolution, outcrop scale anticlines occur in the Plio-Pleistocene sediments, with Jurassic limestones and marls in the nucleus that correspond to static karst pinnacles (or other elevated forms) present between sagged sectors.

Cryptokarst induced deformation is found even in sediments that cover low carbonate content basement units, such as the early Jurassic marl dominated successions. Hence, cryptokarst mechanisms may be also postulated for sediments overlaying these low carbonate content rocks. Karst



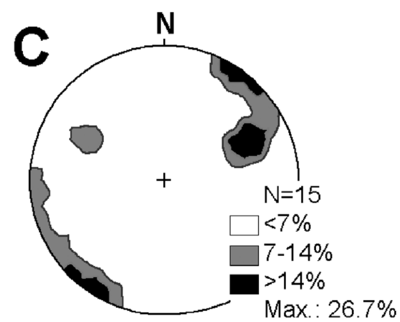


Fig. 13: Photomosaic of the outcrop with a slump bed **(A)** and interpretation **(B)**. The accretion direction shifts at the deformed bed. Fractures measured at this outcrop are represented in a lower hemisphere equal area stereographic plot **(C)**. There is a concentration of faults trending almost perpendicular to the direction of accretion. Location in figure 5.

sink triggering of deformation can only be a priori excluded in Plio-Pleistocene sediments that overlay Triassic or older basement rocks, which are unsuitable for dissolution.

5. Discussion and conclusions

Previous regional studies indicate the occurrence of neotectonic activity along the study region, namely in some N-S trending active faults, of which the major one is the Porto-Tomar fault [16, 27, 37]. In several places it is possible to recognize a high frequency of deformation features. However, part of them seems to be related to non-tectonic processes. Some of the meso-scale deformation structures may be related to sin-depositional processes, as the slump beds and flame structures found in fine to coarse sand delta deposits. Also part of the deformation observed in the cover sediments, in areas where they overlie Jurassic carbonate rocks, could be instigated by dissolution sink along the unconformity between the Mesozoic basement and the covering clastic succession. This can be expected even in areas where low carbonate content marls are the most frequent basement lithology.

Several facts demonstrate an important structural conditioning on the organization of the regional sedimentary systems since late Pliocene to upper Quaternary and indicate significant neotectonic activity. The tectonic activity was more intense during a period of major individualization of uplifted and subsiding areas along the coastal margin, probably in the late Piazencian to early Pleistocene, when units 2B and 3 were deposited, although it apparently continued up to the Present as suggested by the tectonic control on the drainage network.

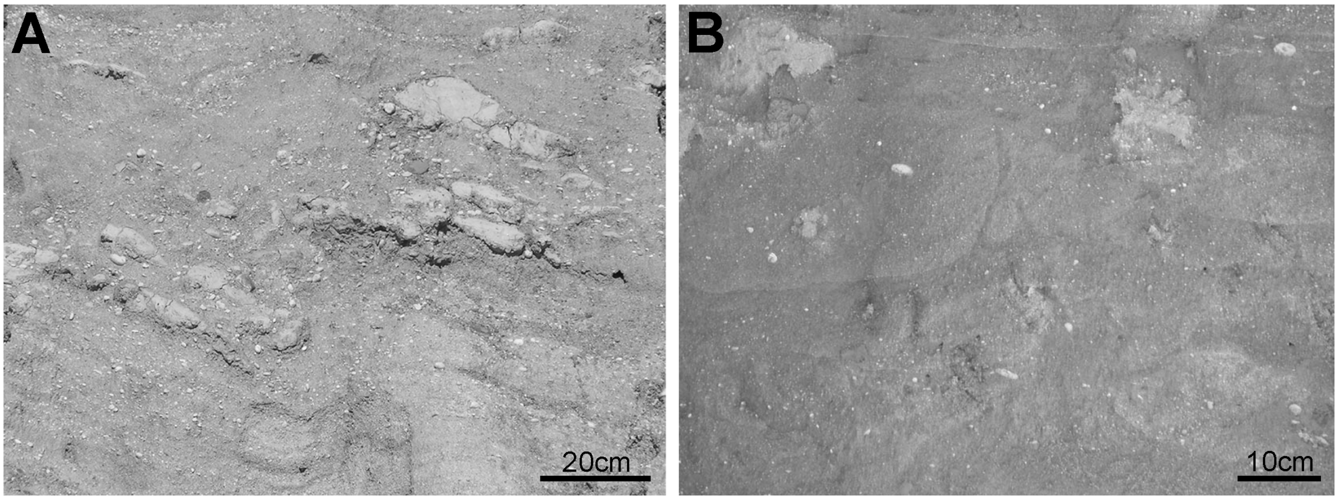


Fig. 14: Detail of some structures associated with the slump bed shown in figure 13. Small recumbent folds **(A)** and water escape structure **(B)**.

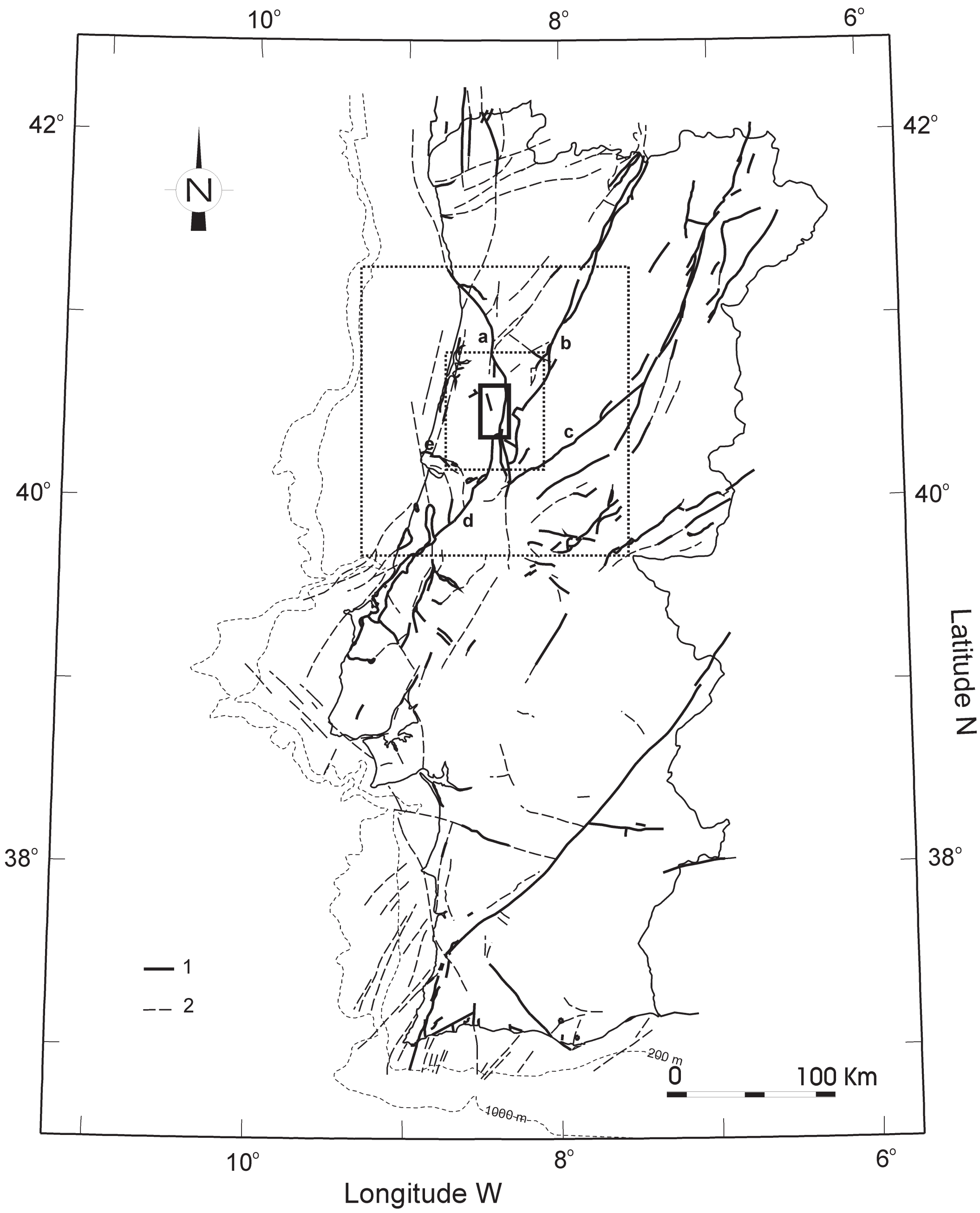


Fig. 15: Simplified version of the Neotectonic Map of Mainland Portugal, adapted from Cabral and Ribeiro [37] and Cabral [16].1: Fault evidencing tectonic activity in the last ca. 2 Ma; 2: Idem, probable. Grey frame: Study area; Dotted frames: 20 and 70 km distance reference areas.
a: Porto-Tomar fault;
b: Penacova-Verin fault;
c: Seia-Lousã fault;
d: Nazaré-Pombal fault;
e: Serra da Boa Viagem fault.

The evidence of surface or near surface faulting affecting Pleistocene sediments and the occurrence of seismites in the study area, indicate moderate to high magnitude paleoseismic activity, which should be considered for the regional seismic hazard assessment, particularly considering the proximity to the cities of Aveiro and Coimbra and their densely populated surroundings. Effectively, coseismic surface rupture typically occurs during shallow earthquakes with magnitude ≥ 5.5 – 6, and a similar magnitude lower boundary of about 5.5 is usually necessary for liquefaction effects to become relatively common at locations near the seismic source, given adequate conditions of high liquefaction susceptibility [36]. It must be noted, however, that high energy, distant earthquakes can also be responsible for the liquefaction related deformation observed in the area.

As referred above, mainland Portugal experiences a moderate seismicity characterized by small events (M<5.0) although occasional moderate to large earthquakes occur, the biggest being related to the Azores-Gibraltar plate boundary zone at the SW, such as the M 8.7, 1755 "Lisbon earthquake" [22, 23, 24, 25]. However, in what concerns the study area, if we exclude this strong far-field seismicity (more than 400 km distant) and consider the regional intraplate environment, the seismic activity gets significantly lower, and strongly damaging earthquakes are scarce or virtually absent in the human registry. The map of maximum intensities for the historical and instrumental seismicity in mainland Portugal [38] shows a value of VII MMI in the study area, which results from the effects of the 1755 earthquake. No "local" earthquakes have produced this level of shaking in the study area. This reflects the intraplate tectonic environment, with low slip rate active faults (<0.1-0.5 mm/yr) implying long recurrence times (ca. 5,000 - 200,000 years) for maximum (M 6-7) earthquakes [16, 27].

According to Obermeier [4, 36], given optimal liquefaction conditions, the maximum distance from earthquake epicenter (for shallow focus events, <50 km depth) to farthest surface evidence of liquefaction-induced ground failure is ca. 20 km to 70 km for M_W 6 to 7, respectively. Considering the empirical regression of fault surface rupture length and moment magnitude by Wells and Coppersmith [39], these values of magnitude imply minimum fault, or fault segment lengths of ca. 8 km and 41 km respectively. Taking those "buffer widths" around the study area boundary we conclude that several known active faults having the potential to produced earthquakes of such magnitudes lay within the necessary distance, the major ones being the above referred Porto-Tomar fault, which crosses the study area, and the Penacova-Verin, Seia-Lousã, Nazaré-Pombal and Serra da Boa Viagem faults [16, 37] (Fig. 15).

The discrepancy between the absence of historical earthquakes and low level of the instrumental seismicity, and the geological evidences of seismic triggered processes suggests a characteristic earthquake behavior of the major regional seismogenic faults, with the earthquakes seismic cycles being much longer than the available data series of seismic activity [40, 41, 42]. In other words, it may signify that the area was affected by relatively strong seismic events, though with return periods much larger than the instrumental or historical seismic record.

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