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Sedimentary Geology 196 (2007) 201–215
Sedimentary Geology
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Formation, deformation and chertification of systematic clastic dykes in a differentially lithified carbonate multilayer. SW Iberia, Algarve Basin, Lower Jurassic

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Abstract

This work presents original field evidence for tectonically controlled calciclastic dyke injection and subsequent chertification in shelf carbonates during rifting in the south-westernmost part of the Eurasian continent in the Early Pliensbachian. It is shown by the detailed description of tectonic structures (faults and joints), stratigraphic discontinuities and by the distribution, orientation and morphology of the injected dykes of calciclastic sands into fine-grained carbonates, that these soft-sediment deformation structures were tectonically controlled. Extensional tectonics developed vertical tensile joints in semi-lithified, fine-grained limestones permitting upward and downward injection of loose calciclastic sands, forming clastic dykes. The frequency of injection structures along strata was constrained by the thickness of the layer into which the injection was occurring, which implies an elastic behaviour; however the curvi-planar shapes of the dykes and drop-like nodules attests the ductile behaviour of the fractured limestones, indicating transition from elastic to ductile response to deformation with time. The multi-layered system consists of just two lithotypes. The two display different mechanical behaviours, which evolved in time to more brittle conditions as lithification progressed, as shown by the cataclastic faulting and jointing of the previously formed soft-sediment deformation structures and strata under an analogous stress regime. The injection dykes were disrupted by a short-lived episode of compression, after which tectonic extension resumed, still in the Lower Pliensbachian. Sometime before the deposition of the Upper Pliensbachian a pervasive selective event of chertification occurred: only the calciclastic sandy layers, dykes and nodules were substituted by silica, thus enhancing the mechanical contrast between the primary sedimentary structures and the soft-sediment deformation structures. All the described events occurred during a time interval of approximately 2 Myr as constrained by ammonoid stratigraphy.

Keywords: Syn-sedimentary dykes; Differential lithification; Tensile joints; Chertification; Lower Pliensbachian

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0037-0738/$ - see front matter
doi:10.1016/j.sedgeo.2006.06.001

1. Introduction

1.1. Objectives

Soft-sediment deformation structures have been described from a variety of environments such as lacustrine, fluvial, aeolian, reef, continental shelf and continental slope (Haczewski, 1986; Roep and Everts, 1992; Guiraud and Plaziat, 1993; Pratt, 1994; Owen, 1995; Alfaro et al., 1997; Pope et al., 1997; Jones and Omoto, 2000; Moretti, 2000; Ken-Tor et al., 2001; Rossetti and Santos, 2003; Bachman and Aref, 2005). The lithologies involved in most of the published case studies are mudstones and sandstones, calciclastic limestones, dolomites and evaporitic sediments (Mohindra and Bagati, 1996; Alfaro et al., 1999; Molina et al., 1997; Matsuda, 2000; Jones and Omoto, 2000; Ken-Tor et al., 2001; Alfaro et al., 2002; Rossetti and Santos, 2003; McLaughlin and Brett, 2004; Bachman and Aref, 2005), and the invoked trigger mechanisms are generally, seismic activity, gravitational instabilities, overloading, unequal sediment loading, storm waves and cryoturbation (Cosgrove,

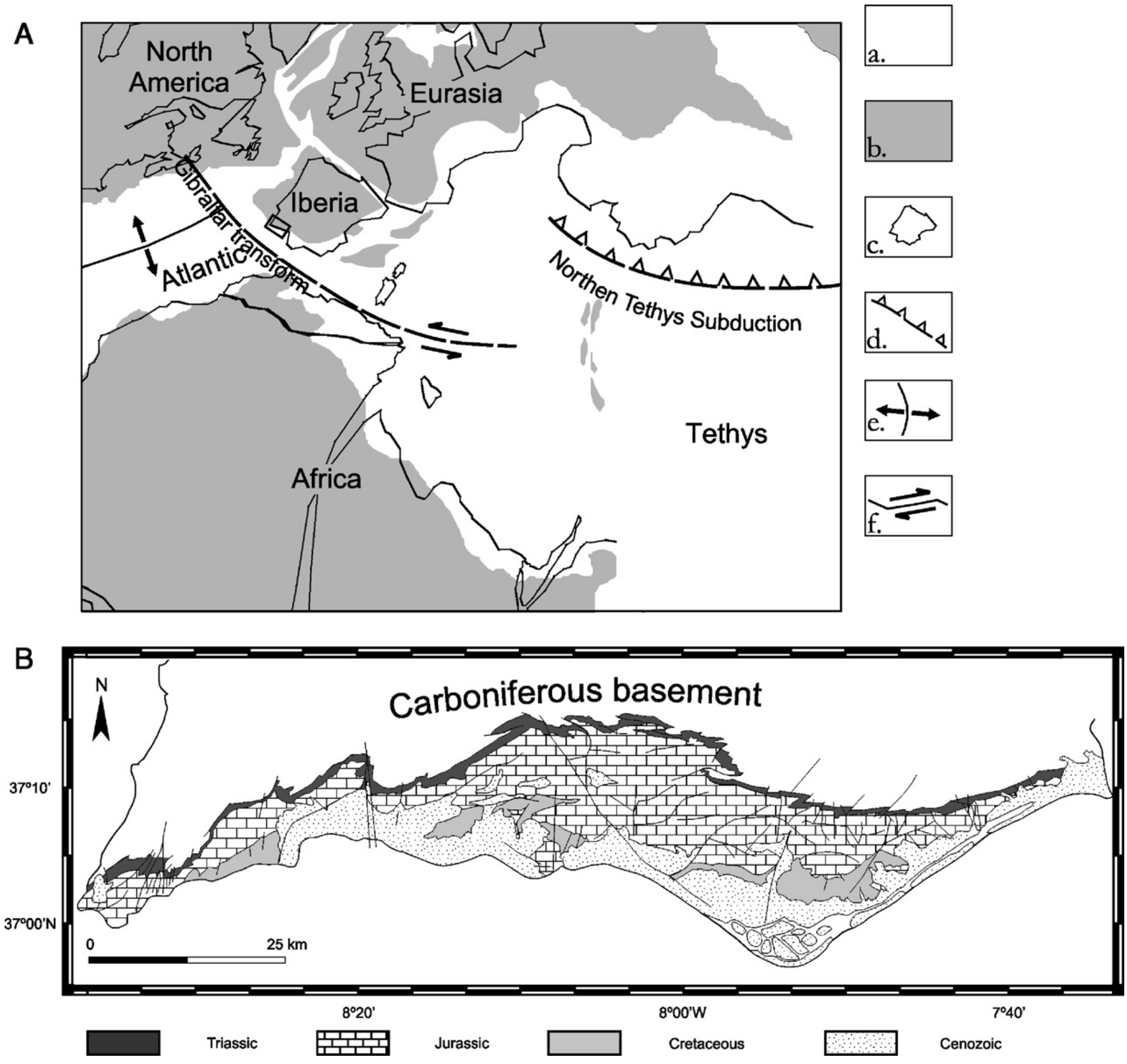

Fig. 1. (A) Schematic paleogeographic reconstitution of the West Mediterranean area in the Jurassic with the location of the Algarve Basin (small rectangle in the southwest Iberia): a – submersed areas; b – emerged areas; c – present day continent boundaries; d – subduction zone; e – rift; f– transform fault (adapted from Thierry, 2000). (B) Simplified geologic map of the Algarve Basin.

1995; Obermeier, 1996; Cosgrove, 1997; Molina et al., 1997; Alfaro et al., 2002).

The aim of this paper is to show a case study of soft-sediment deformation structures whose development is controlled by the association of the prevailing stress regime with the differential lithification of the sediments. The structures formed in Lower Jurassic open marine, shallow-water platform carbonates, during rifting. The structures also experienced a short episode of tectonic compression and went through a process of partial to complete silicification. These events occurred in less then 2 Myr and are well constrained by ammonoide stratigraphy and well exposed unconformities. Another peculiar aspect of the described outcrops is the record of the progressive transition from soft-sediment to lithified sediment under two analogous stress regimes. The mechanisms of formation and the trigger mechanisms of intrusion of the dykes are also discussed.

2. Geological setting of the Algarve Basin

The Algarve Basin is located in the south-westernmost onshore part of the Eurasian continent and is made up of two superimposed Mesozoic and Cenozoic basins (Terrinha, 1998) that rest on top of a Carboniferous low grade metamorphic thrust belt of the Iberian Variscan orogen (Oliveira, 1990). The two basinal cycles are separated by the Turonian–Burdigalian hiatus which corresponds to the main phase of tectonic inversion and uplift of the Mesozoic rift basin (Terrinha, 1998; Lopes, 2002). The Mesozoic Algarve Basin and other contemporaneous basins of southern Iberia and their northern Africa neighbours resulted from the extensional tectonics associated with the break up of Pangea and development of the westernmost Neo-Tethys from Early Triassic to Late Cretaceous times (Terrinha, 1998).

The outcrops described in this paper (Fig. 1) are located on the geographical transition between the Western and Southern Iberian Margins (and its continuation along the Mediterranean Sea through the Strait of Gibraltar), i.e. on a structural high that separated the Atlantic and Tethyan rifted margins, where it is possible to inspect both sets of extensional faults: the Atlantic rift faults trending approximately N–S and the Tethyan faults trending approximately E–W (Terrinha, 1998). This structural high resulted in the deposition of condensed Jurassic and Lower Cretaceous stratigraphic sequences (~500 m) in which the rifting and inversion tectonic events can be more clearly seen than in the depocentres located further southeast where sequences can exceed 4 km (Lopes, 2002).

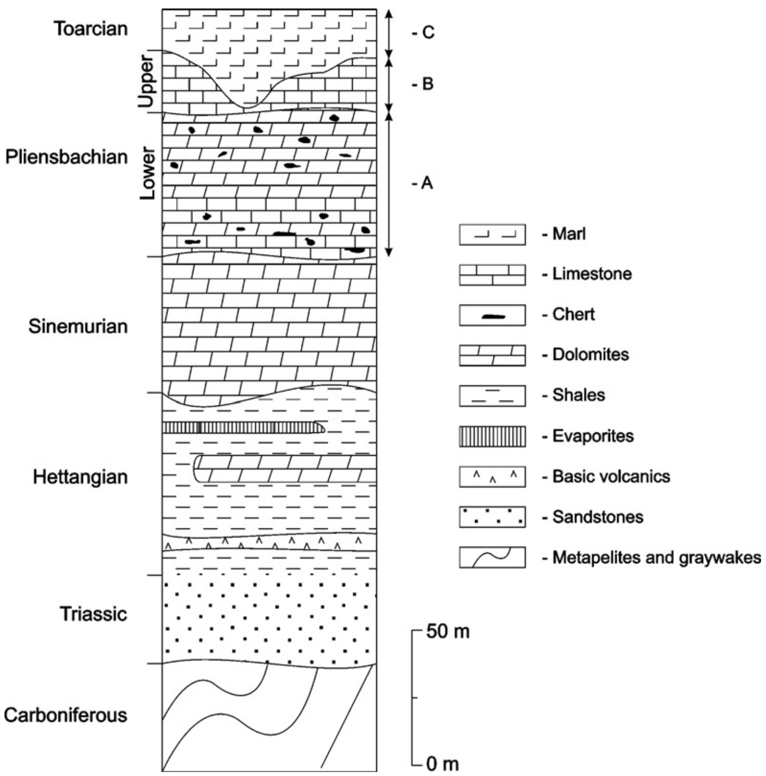

Fig. 2. Lithostratigraphic column of the Triassic–Lower Jurassic section of the western Algarve Basin. For the meaning of A, B, and C see text.

The Mesozoic sedimentary record of the Algarve Basin spans from Early Triassic (Palain, 1976) to Cenomanian times (Rey, 1983). The sedimentary environments evolved from continental in the Triassic (terrigenous siliciclastics) through confined littoral in Hettangian–Sinemurian times (red shales, dolomites and evaporites, Terrinha et al., 1990), to open marine from the Early Pliensbachian to the Toarcian (carbonates and marls with ammonoids) (Fig. 2). A sub-aerial tholeiitic volcanic event, related to the Central Atlantic Magmatic Province, spread out along the south Iberian margin and northwest Africa at the Hettangian–Sinemurian transition (Martins and Kerrich, 1998; Martins, 1991). Important hiatuses are observed from upper Toarcian to lower Bajocian, upper Callovian to middle Oxfordian and lower Berriasian. Based on the detailed chronology of the compressive structures observed throughout the Algarve Basin Terrinha et al. (2002) showed that the last two of these hiatuses and uplift events were associated with compressive tectonic episodes.

3. Stratigraphy

The soft-sediment deformation structures described in this paper are hosted within a 55 m thick package of limestones and dolomitic limestones with chert nodules of Lower Pliensbachian age. The lower stratigraphic boundary corresponds to a hard-ground that separates an underlying unit of extremely dolomitised sediments of Sinemurian age, from the overlying well bedded carbonates of the Jamesoni ammonoid biozone (Rocha, 1976). Most of the Lower Pliensbachian formation exhibits a paleontological association characteristic of the Ibex biozone and the uppermost Lower Pliensbachian Davoei biozone is absent. The upper boundary of this unit consists of an erosion surface and slight angular unconformity, overlain by the Upper Pliensbachian limestones and marls. The time interval of the discontinuity between Lower and Upper Pliensbachian units is fairly well constrained by the existence of ammonoids of the Stokesi biozone – of early Upper Pliensbachian age – above the discontinuity. Thus, the sediments hosting the cherts and soft-sediment deformation structures correspond to a time interval not larger than 2 Myr.

The Lower Pliensbachian primary sediments of the study area consist of fairly continuous 0.1 to 0.7 m thick layers of fine-grained limestones with variable small amounts of clay (less than 10%) containing abundant crinoid fragments and patches of bivalves alternating with discontinuous layers of calciclastic limestones, generally less than 0.1 m thick. Although generally discontinuous, some calciclastic layers can be followed for tens of meters.

The cherts are a product of the early diagenetic silicification of the coarser calciclastic and bioclastic units, i.e. the most permeable lithologies with respect to the fine-grained host limestones and dolostones, as shown by Ribeiro (2005) based on the observation of outcrops, petrography and scanning electronic microscopy (SEM). The coarse-grained primary lithologies interacted with silica-rich fluids that induced the dissolution of the primary calcite and precipitation of silica leading to the formation of silicified limestones when the process was not complete or the formation of cherts when the substitution was total. The photograph of Plate 1(I) was taken using a scanning electron microscope and shows crystals of primary calcite (Cc) with evidences of corrosion and pitting with secondary silica (S) precipitated in the voids. The silicification can be complete or partial from SEM scale to outcrop scale. In Plate 1(II) a mesoscopic example of different degrees of

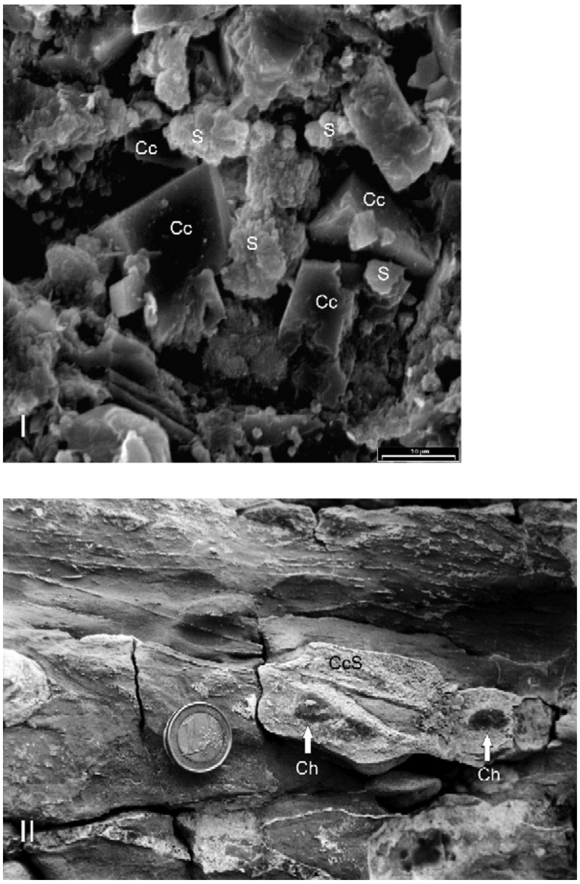

Plate 1. (I) SEM photograph of the primary calcite crystals (Cc) with evidences of pitting and corrosion, being replaced by secondary silica (S). (II) Calciclastic limestone replaced by silica. When the substitution is incomplete (CcS) some of the primary characteristics of the sediments are preserved. The complete substitution of the primary calciclastic limestone by silica originates the chert (Ch).

silicification is presented. In this picture a calciclastic limestone is partially replaced by silica (CcS) and preserves some of the primary characteristics of the sediment. However in two small areas (Ch) the replacement was complete and the observed lithology is now chert. The diagenetic evolution of this formation, which evolved from interbedded fine-grained and coarse-grained limestones to limestones and dolostones interbedded with chert nodules and layers, containing chert dykes was polyphase and complex (Ribeiro et al., 2004; Ribeiro, 2005).

There are two stratigraphic packages of chert interbedded in the Lower Pliensbachian carbonate sequence, one close to the base of the formation, 15 m thick, and another closer to the top, 20 m thick. Within these

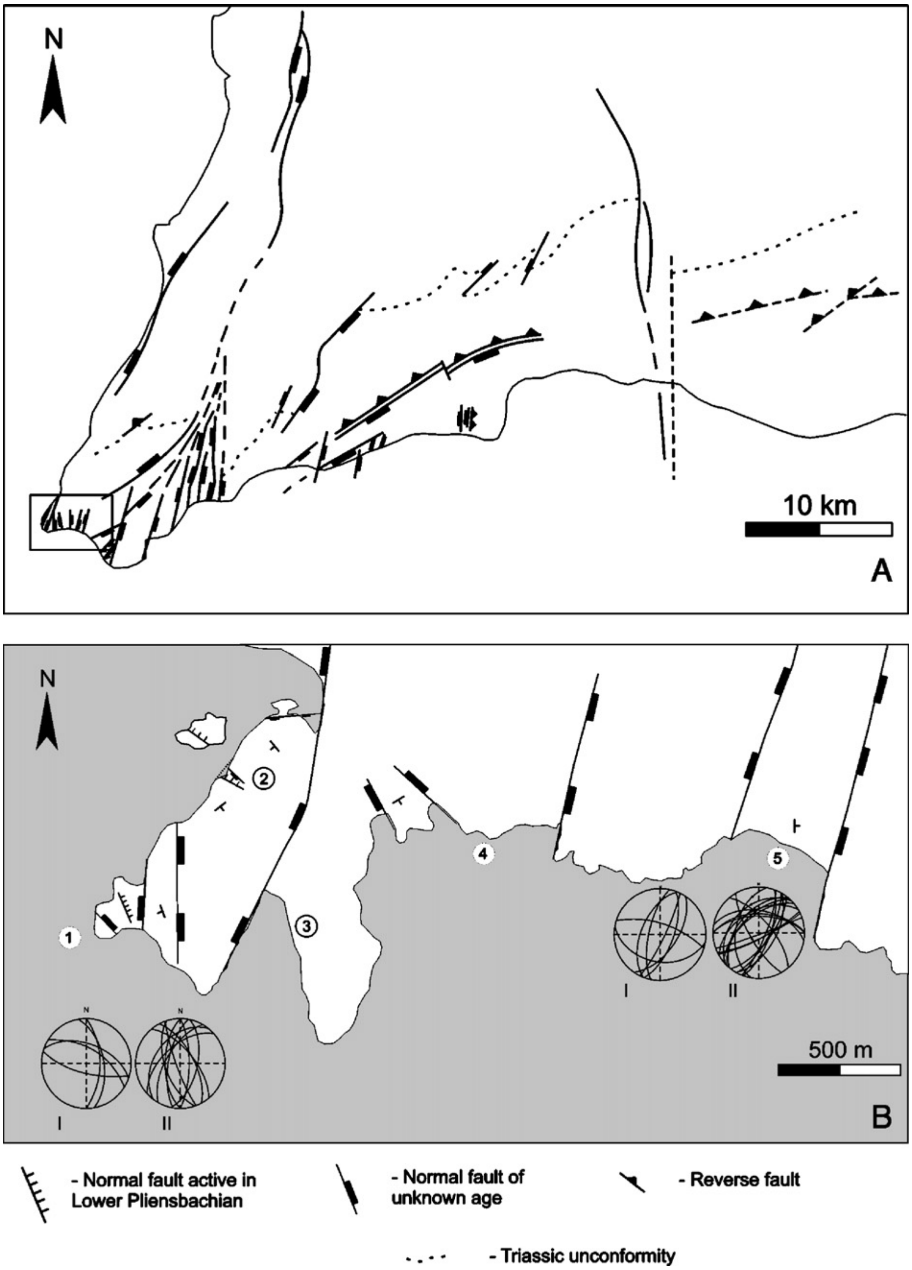

Fig. 3. (A) Structural map of the West Algarve Basin. Note the existence of approximately N–S and E–W trending normal faults related to the rifting of the Atlantic and Tethyan Continental Margins, respectively. (B) Structural map of the study area and location of outcrops described in the text: 1 – Cabo de São Vicente; 2 – Aspa; 3 – Foz dos Fornos; 4 – Forte do Belixe; 5 – Praia do Belixe. The stereoplots show (I) syn-sedimentary normal faults in soft sediments and (II) normal faults of unknown age.

packages, the chert amounts up to 15–20% of the whole rock volume. The limestones between these two chert packages consist of fine-grained limestone only (Fig. 2).

4. Structure and tectonics

This portion of the continental part of the Eurasia plate depicts a right angle between the Atlantic coast and the Gulf of Cadiz and records the intersection pattern of the rifting structures at the intersection of the Atlantic Ocean and the Mesozoic Neo-Tethys Ocean. Although the extensional faults found in the study area strike predominantly around N–S and E–W, i.e. parallel to both paleo-rift margins, sometimes the extensional strain was also accommodated on NW–SE or NE–SW trending extensional faults that resulted from synchronous stretching on both E–W and N–S direction (Terrinha, 1998). In the western part of the Algarve Basin, the study area, there is a clear predominance of the N–S extensional faults, which were active throughout the Middle–Late Jurassic and Middle Cretaceous (Terrinha, 1998), but possibly less active than the E–W fault set during Lower Pliensbachian times (Fig. 3A).

4.1. The Cabo de São Vicente outcrop

The study area at the Cabo de S. Vicente (labelled 1 in Fig. 3B) consists of small grabens and half-grabens trending around the typical N–S and E–W directions. Although the Lower Pliensbachian sediments outcrop in several locations the exposure conditions around the Cabo de S. Vicente make only three outcrops accessible: the Cabo (1 in Fig. 3B), the Aspa (2 in Fig. 3B) and the Foz dos Fornos (3 in Fig. 3B) outcrops.

At the three outcrops of the Cabo de S. Vicente the sedimentary record begins with a thick (more than 25 m) package of compact grey and brown dolostones of Sinemurian age. The dolomitisation is secondary and obliterates the primary sedimentary characteristics of the carbonate sediments (Rocha, 1976). The Sinemurian sediments are separated from the Lower Pliensbachian sediments by an unconformity visible in all the Cabo de S. Vicente area. The Lower Pliensbachian consists of a 30 to 35 m thick sequence of decimetric beds of dolomitic limestones with chert nodules, marly limestones and limestones. The faunal associations described by

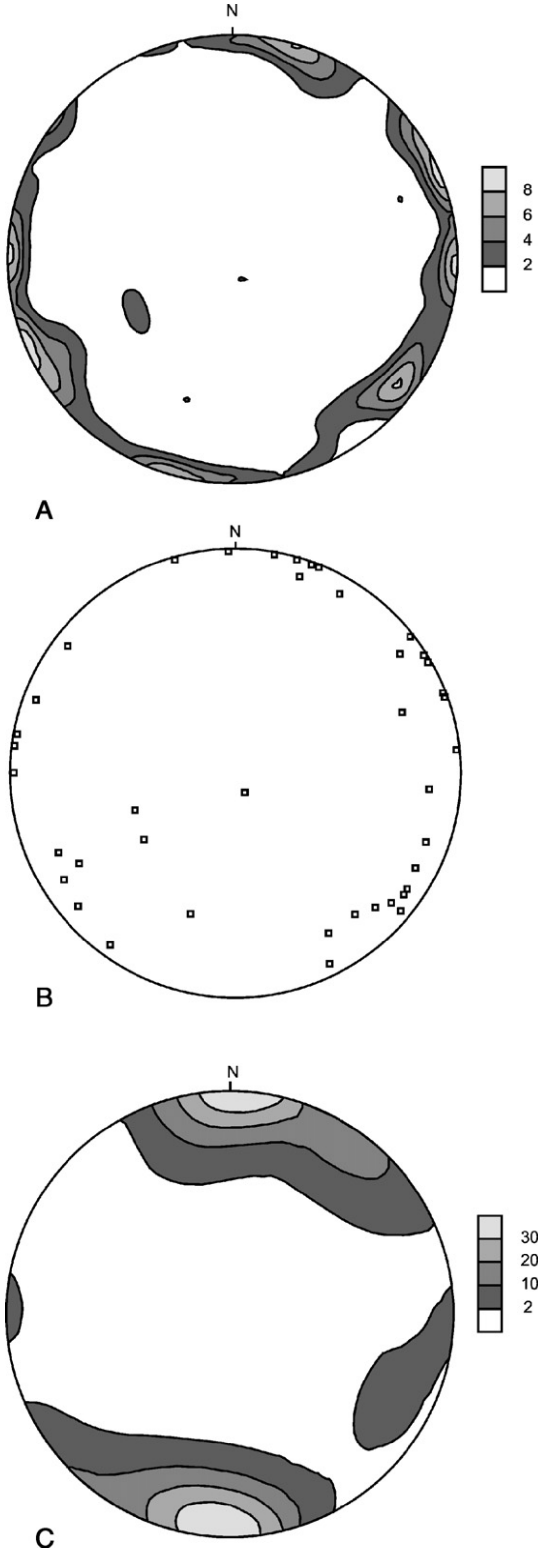

Fig. 4. Joints in the Lower Pliensbachian sediments. (A) Density contour diagram for the joints from the Cabo de S. Vicente outcrop (density values in percent; $n=97$). (B) π diagram of the poles of the joints filled with quartz from the Cabo de S. Vicente outcrop. (C) Density contour diagram for the joints filled with quartz from the Praia do Belixe outcrop (density values in percent; $n=81$).

Rocha (1976) are characteristic of the Early Lower Pliensbachian – Jamesoni biozone – and Middle Lower Pliensbachian – Ibex biozone. The dolostones display dolosparitic textures and the limestones can be sub-divided into micritic limestones and calciclastic (Ribeiro, 2005). The calciclastic limestones are lensoid in shape with a maximum thickness of 0.1 m, showing high energy structures, such as cross bedding.

The general structure of the Cabo outcrop (labelled 1 in Fig. 3B) consists of a N–S trending graben, within which the Lower Pliensbachian beds lie sub-horizontal. Only the basal 30 m of the formation are exposed resting on top of the Sinemurian dolomites and the deformation of the sediments is accommodated by the development of normal faults and joints. Two sets of striking N–S and WNW–ESE growth faults are present in this outcrop.

Various N–S trending faults, although not showing obvious syn-sedimentary criteria are also present, as well as NE–SW striking normal faults. All these faults show evidences of re-activation after lithification. Three joint sets approximately parallel to the main N–S, WNW–ESE and NE–SW fault trends are present (Fig. 4A); the first two sets consist of vertical tensile (Mode I displacement – Atkinson, 1987) joints, while the third set is made up of mixed Mode I (predominant) and Mode III displacement joints. Some of the joints (Fig. 4B) are infilled by 0.01 m thick quartz veins (Ribeiro, 2005). Many of the joints were injected by clastic–carbonate sediments, which indicate that the joints formed whilst the calciclastic layers were still un-lithified, as shown in detail later in this work.

The Aspa outcrop (labelled 2 in Fig. 3B) is a small WNW–ESE trending asymmetric graben in which the early Lower Pliensbachian sediments are brought into contact with the Sinemurian sediments (Fig. 5). These strata are strongly deformed by the later extensional and inversion tectonic events. One of the graben boundaries consists of a syn-sedimentary WNW–ESE striking normal fault sealed by the Lower Pliensbachian along which a 0.3 m thick quartz vein was emplaced; reactivation of this fault is shown by brittle deformation of the quartz vein.

The Foz dos Fornos outcrop (labelled 3 in Fig. 3B) consists of a half graben bound by the main N–S trending fault of this area, which cross cuts all the existent E–W trending faults and separates the Lower Jurassic from the Middle Jurassic and possibly continued its activity throughout the Lower Cretaceous; this is inferred by the steeply dipping Middle Jurassic bedding planes (>60°) and inspection of parallel faults on seismic reflection lines offshore further west (Terrinha et al., 2003). The Lower Pliensbachian shows a systematic WNW–ESE trending set of quartz filled tensile joints with occasional occurrence of thin (<5 mm thick) calciclastic injections, again indicating incomplete lithification of the whole sedimentary series at the time of joint formation (Terrinha, 1998).

4.2. The Forte do Belixe outcrop

This is the only outcrop (labelled 4 in Fig. 3B) where the whole Pliensbachian sequence is continuously

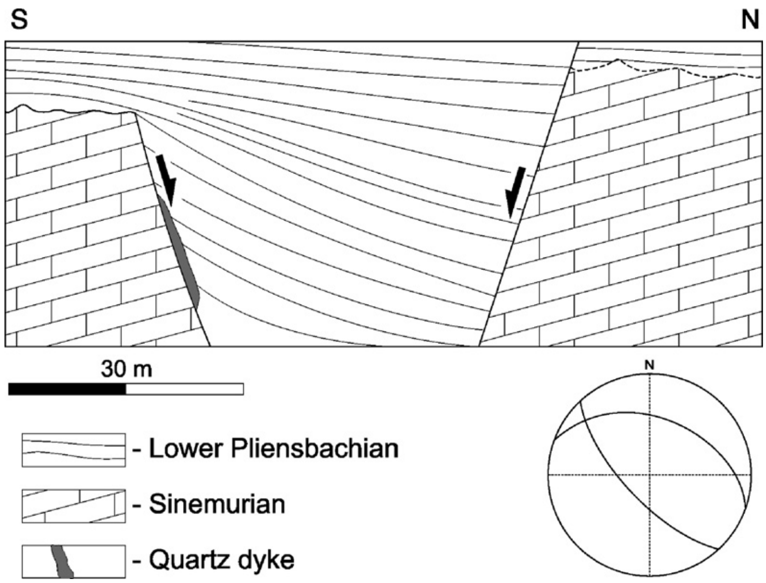

Fig. 5. Cross-section of the Aspa outcrop (for location see Fig. 3B).

exposed along a 15°ESE dipping monocline, from its lower contact with the Sinemurian through to its upper contact with the Lower Toarcian.

The compact Sinemurian dolomites are unconformably overlain by the Lower Pliensbachian limestones, marly limestones with chert nodules and dolomitic limestones with chert nodules. The fossils are abundant and the faunal association is characteristic of the Jamesoni biozone at the base of the Formation and of the Ibex biozone at the top (Rocha, 1976). The upper Lower Pliensbachian is absent and the sedimentation resumed in the Upper Pliensbachian (Stokesi biozone) with the sedimentation of limestones. The late Upper Pliensbachian times are marked by an erosional event after which the sedimentation resumed with the deposition of the Toarcian marls.

Structurally, this outcrop consists of a rotated faulted block, bound by N–S and NW–SE striking normal faults, across which the Lower Jurassic is brought into contact with the Middle and Upper Jurassic; no important internal extensional faults are observed and only minor inversion structures. However, WNW–ESE striking tensile joints, perpendicular to bedding, are common. These joints are generally filled by quartz or injected by clastic sediments.

4.3. The Praia Belixe outcrop

The Praia do Belixe outcrop (labelled 5 in Fig. 3B) hosts the majority of the soft-sediment deformation structures of the study area, which are well developed and display a prominent structural and tectonic control. This outcrop consists of a continuously exposed monocline sequence of Lower Pliensbachian through to part of the Lower Toarcian, dipping 5° to 15° to the east. The monocline lies within a tilted horst, bound by two NE–SW trending normal faults that have been steepened to vertical by post Middle Jurassic tectonic inversion events. The well exposed sedimentary section allows the inspection of unconformities and deformation structures that reveal a varied syn-sedimentary geological evolution of the Lower Pliensbachian sediments. The Upper Toarcian is absent and the contact with the Middle Jurassic is the result of the normal faulting mentioned above (Fig. 6).

Inspection of the outcrop shows the existence of a basal well bedded sedimentary unit, made up of limestones and dolostones of Lower Pliensbachian age containing chert (UNIT A), overlain by non-dolomitised limestones (UNIT B); these units are cut by an erosion surface overlain by UNIT C, which consists of marls of late Upper Pliensbachian and Lower Toarcian ages. All units were dated by Rocha (1976) based on their ammonoid fossil content.

Careful inspection of the sedimentary packages, their discontinuities, unconformities and faults allowed the mapping and dating of the tectonic structures, as follows:

i) UNIT A displays two extensional fault sets trending WNW–ESE and NE–SW. The former set predates the latter (Fig. 6);

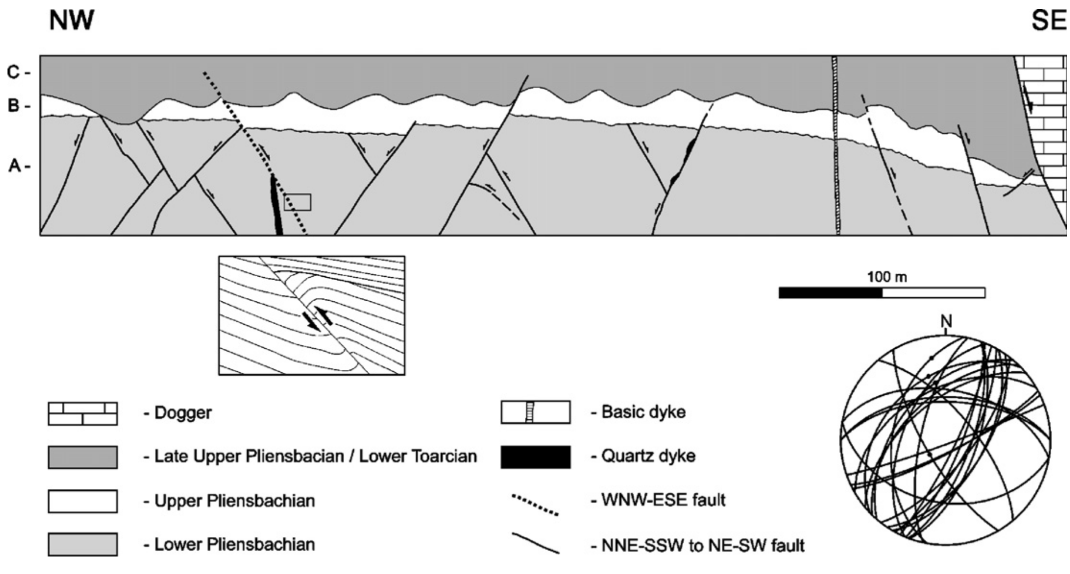

Fig. 6. Cross-section of the Praia do Belixe outcrop (for location see Fig. 3B). A, B and C refer to the Lower Pliensbachian, Upper Pliensbachian and Uppermost Pliensbachian–Toarcian units, respectively. The inset shows the detail of the inversion of a normal fault and synchronous formation of an unconformity. Extension resumed during Upper Pliensbachian as shown by the downthrown hanging-wall at the top of the fault section.

ii) both fault sets deformed un-lithified sediments of UNIT A; later they were re-activated during deposition of UNIT B – at a time when UNIT A sediments were already lithified – and ceased their activity during deposition of UNIT C (Fig. 6);

iii) the WNW–ESE trending extensional faults were inverted as dip–slip reverse faults as shown by the drag folds. This shortening event occurred during the deposition of UNIT A as indicated by the eroded crest of the drag-fold on the fault hanging-wall and the angular unconformity between the folded and the overlying strata within UNIT A (see inset in Fig. 6); the shortening event was followed by renewed extension as shown by the extensional offset at the top of the section at the base of unit B;

iv) the WNW–ESE extensional fault set is parallel to a well developed set of tensile joints, which are the oldest joints observable in UNIT A; these joints host chert nodules, chert dykes and calciclastic dykes whose description and formation is presented below;

v) the WNW–ESE faults do not contain quartz veins whereas the parallel extensional joints do (Fig. 4C); the quartz veins cross cut the chert layers and chert nodules;

vi) the NE–SW extensional faults are injected with quartz veins containing fluid inclusions, which yielded homogenization temperatures above 200 °C; these faults and quartz veins cross cut all the rocks of UNIT A;

vii) the NE–SW extensional faults were re-activated during the Lower Toarcian as was the WNW–ESE fault set although the evidence for this is only present at Cabo de São Vicente (Terrinha, 1998); the Middle Toarcian through Aalenian (?)–Bajocian hiatus across the whole Algarve Basin suggests that the end of the Lower Jurassic rifting event is associated to basin uplift.

viii) all fault sets were later cross cut by the N–S and NE–SW trending extension faults of Middle Jurassic through Lower Cretaceous age.

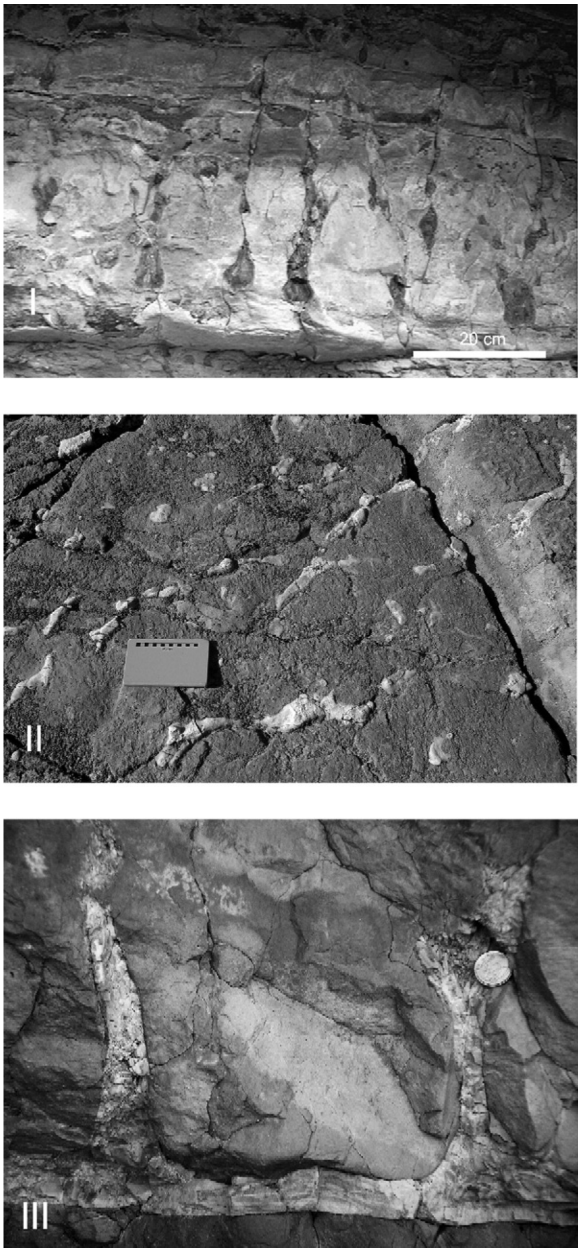

Plate 2. (I) Drop-like nodules of chert (dark gray) hosted in a layer of dolomitic limestone in a vertical perspective. (II) Chert nodules (light gray) in a limestone (dark gray) bedding surface. In this image the planar character of the nodules is visible. The scale bar in the notebook is 15 cm. (III) Upward injection of chertified calciclastic sediments (light gray) from a layer into the limestone.

5. The soft-sediment deformation structures

5.1. Characterisation

The Lower Pliensbachian of the study area exhibits various types of soft-sediment deformation structures. They occur in a systematic way and can be divided in two groups: i) load casts and pseudo-nodules resulting from a downward movement of the calciclastic material and ii) injection structures resulting from an upward movement of the same material. Because the calciclastic sands were replaced by silica (Ribeiro, 2005), most of the described structures are materialised in chert.

The chert beds frequently display irregularities both at their tops and bases. Irregularities formed at the bases of the beds evolved into load casts and elongated

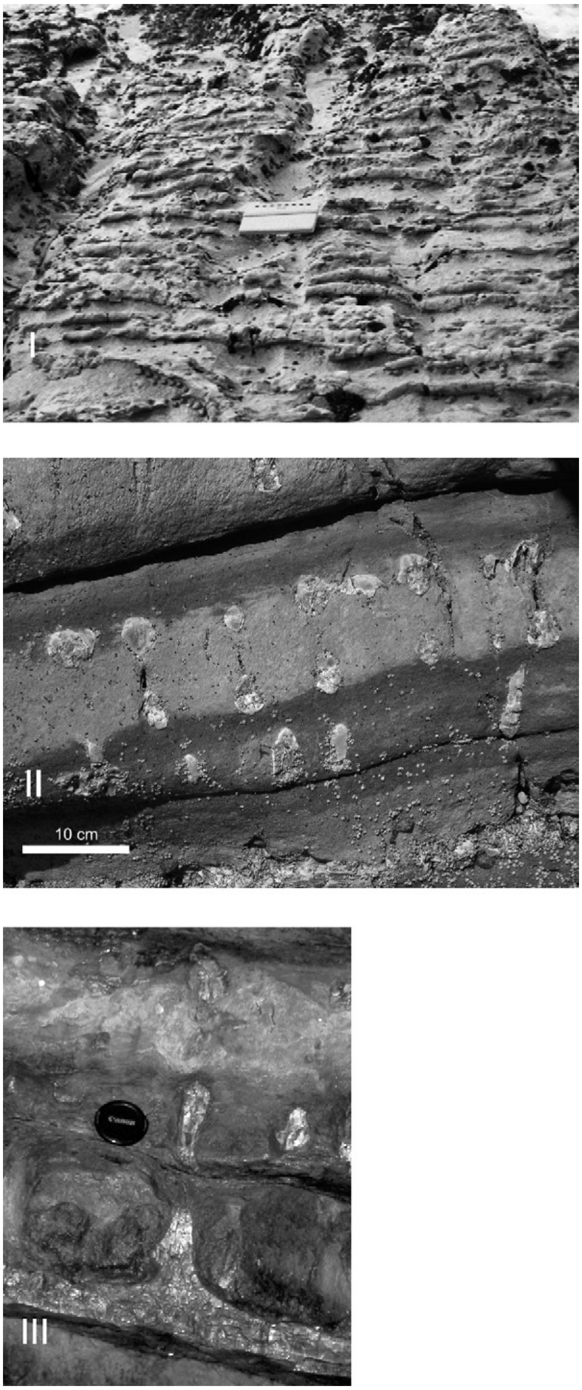

Plate 3. (I) Bed surface with a set of chert dykes (prominent parallel to notebook) perpendicular to bedding. (II) Set of chert nodules (light gray) with symmetric drop-like geometries within a single layer of dolomitic limestone. (III) Cusp and nodules aligned perpendicular to bedding.

pseudo-nodules that often pierced downwards through the whole of the thickness of the underlying layer (Plate 2(I)); however, when observed on a layer parallel section one becomes aware that these structures are planar and perpendicular to bedding (Plate 2(II)). Generally, there is a fracture, coated with quartz or chalcedony, connecting the chert pseudo-nodule and the chert mother layer. Another peculiar aspect is that these structures are periodically repeated along the host limestone beds, such as the joints formed in the brittle sediments.

The second type of soft-sediment deformation structure results from the upward movement of calciclastic material piercing the overlying layer. It is possible to find a continuous record of structures describing the various stages of formation of these structures. The top surface of the calciclastic layers (now preserved as chert) started to pierce through the fine-grained limestones; this stage is preserved by cusps (i.e. symmetric flame structures) at limestone–chert interfaces (Plate 2(III)). Once the initial resistance to piercing was overcome the injection evolved to a planar structure, i.e. a dyke of clastic material was injected upwards into the overlying limestone (Plate 3(I)).

Generally, after injection, the overlying layer recovered in a ductile manner and the connection between the injected material and the mother layer was severed in the same way as occurred with the pseudo-nodules (Plate 3 (III)). All that remains is a fracture coated by quartz. A peculiar sub-type of dykes that consists of drop-like, symmetric nodules is shown in Plate 3(II). These drop-like bodies are also planar and parallel to the remnant dykes.

The dimensions of these upward injected dykes are very variable, for instance those shown in Plate 3(I) are more than 0.5 m tall, 3 m long and less then 0.05 m thick. However, this proportion is highly variable, since height of the dykes depends on the thickness of the pierced layer

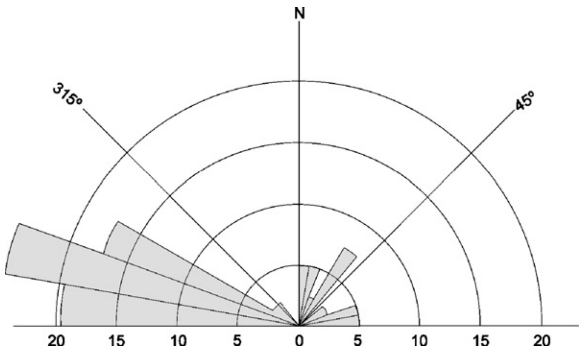

Fig. 7. Rose diagram of the chert dykes strike and chert nodules longest axis trend. The modal class is parallel to the strike of the syn-sedimentary normal faults. Note that there is also a concentration of values perpendicular to the main histogram class (see text for explanation). The horizontal axis values are in percentage.

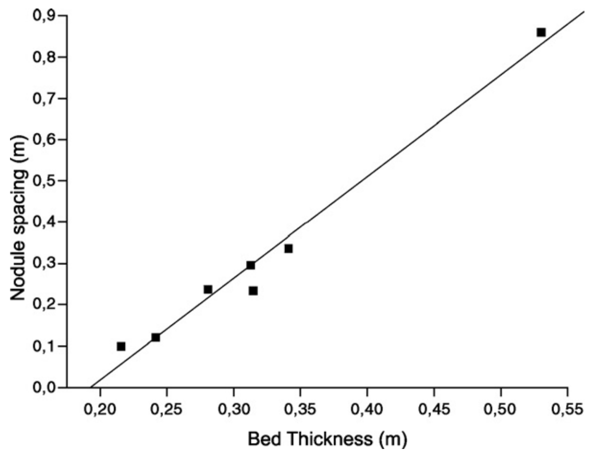

Fig. 8. Scatter plot of the fine-grained carbonate average bed thicknesses *versus* the average spacing of the hosted chert nodules. The observable linear correlation suggests the occurrence of calciclastic sediments injection along joints (see text for discussion).

and their strike length has subsequently been disrupted as discussed later. Sets of vertically aligned nodules, dykes and pseudo-nodules cutting across the various beds are frequent (Plate 3(III)); the relevance of this vertical continuity of injection structures is discussed below. They are generally connected by fractures with thin quartz coatings of the walls.

5.2. Position, distribution and orientation of the soft-sediment deformation structures

The Lower Pliensbachian contains two stratigraphic packages of calciclastic limestones (see Fig. 2, Section 2) that have been replaced by chert. The silicification increased upwards in the stratigraphic sequence and eastwards. At the Praia do Belixe outcrop (5 in Fig. 3B), the easternmost outcrop, there are only rare remnants of the calciclastic material, whilst in the westernmost exposure at Cabo de São Vicente (1 in Fig. 3B) this material is common. Here only the lower stratigraphic section is exposed. The soft-sediment deformation structures formed mainly at the upper calciclastic sub-package of the Praia do Belixe outcrop and were extensively replaced by chert. The upper chert stratigraphic package is also well exposed at the Forte do Belixe outcrop (4 in Fig. 3B); however, there are less soft-sediment deformation structures compared to the Praia de Belixe outcrop, probably because there is less strain accumulated in the Forte do Belixe monocline.

The injection structures (i.e. dykes and aligned nodules) exhibit a well organized spatial distribution both in orientation and frequency. The strike of the dykes and the trend of the nodules' long-axis measured on bedding surfaces (Fig. 7) cluster around WNW–ESE, parallel to the strike of some of the syn-sedimentary normal faults and of the oldest tensile joints present at the outcrops. Despite the concentration of dykes around the WNW–ESE strike, there are a significant number of nodules that strike at right angles to this trend. Like the nodules making up the dykes these nodules are also oblate but much shorter. The distribution of the nodules, when observed in vertical sections is not random and a plot of the average nodule spacing against the host limestone average bed thickness shows a strong correlation between the two variables (Fig. 8).

The lower calciclastic package, which was also replaced by chert does not display the deformational features present in the rest of the formation. Both the calciclastic and chert nodules, have highly irregular shapes, with no particular spatial organization and the dykes are rare (Ribeiro, 2005). The very well spatially organized nodules and dykes are only found on the upper Lower Pliensbachian sediments.

6. Discussion

6.1. Progressive lithification

The Lower Pliensbachian of the study area shows four different lithotypes that correspond to different early diagenetic processes: i) fine-grained limestones, formed by compaction and cementation of crypto-crystalline calcium carbonate and traces of clay (type I limestones); ii) calcium carbonate cemented coarse-grained clastic and bioclastic limestones (type II limestones); iii) cherts, formed by a process of dissolution of the type II limestones and the precipitation of silica; and iv) dolomites, formed by substitution of the primary carbonates during the interaction with a mixture of land derived fluids with marine fluids (Ribeiro, 2005). The dolomitisation was preceded by the silicification.

The lithification of limestones was the first diagenetic process to initiate; however, this lithification process, mainly accommodated by compaction and cementation was not vertically uniform nor did it occur simultaneously in the two interstratified fine-grained and coarse-grained limestones types. The injection structures which are only observed in the upper part of the Lower Pliensbachian are a strong indication that the lower part of the sequence was already at least partially lithified when the injection structures formed. For the upwards and downwards injection of calciclastics, a different mechanical behaviour of the two limestone sediments is required. The type I limestones had to be more cohesive in order to develop extensional fractures and the type II limestones had to be completely uncemented and subjected to a high fluid pressure.

Since type I limestones contain only traces of clay it is not likely that the cohesive behaviour was the result of the clay content. It is argued therefore that the resistance to planar fracturing was caused by partial cementation by $CaCO_3$. After the formation of the soft-sediment deformation structures and faulting, total cementation of the two limestone types was accomplished before the deposition of the Upper Pliensbachian limestones, probably during the formation of the Lower–Upper Pliensbachian unconformity.

6.2. Age of lithification

Although there are four lithotypes defined in this paper (limestones type I and II, and their alteration products chert and dolomites) only the lithification for the limestones is discussed because the chert and dolomites are secondary in origin formed by two epigenetic events. The chertification and dolomitisation both occurred before deposition of the Upper Pliensbachian since this formation is not affected by either process.

Although the chertification occurred also before the Upper Pliensbachian discontinuity, it is not known whether the lower chert and upper chert units formed simultaneously after compaction of the whole limestone series (and before dolomitisation) or whether the chertification of the lower unit preceded the upper unit. It is important to note that the cross cutting relationship between chert nodules is not a reliable chronological criterion since coarse clastic materials (permeable media) can be separated from fine-grained material (impermeable media) by a thin impermeable film, both on horizontal bedding surfaces and within the vertical injected dykes.

6.3. The role of tectonics and the trigger mechanism

The formation of the herewith described calciclastic intrusions resulted from an extensional stress field, the Early Jurassic rifting that deformed the still un-lithified Lower Pliensbachian sediments of the study area. This is inferred from the constant orientation of the soft-sediment deformation structures and the vertical prolongation of downward and upward propagating planar clastic injection, all parallel to the strike of the syn-sedimentary normal faults at the different outcrops. It is also notable that the ductile deformation of the dykes by the episode of transient tectonic inversion that stretched the dykes by forming disrupted bodies of softer material within a harder one. The occurrence of the minor population of nodules (around NE–SW) striking perpendicular to the majority ones (around WNW–ESE) raises two questions: i) did the two joint sets develop during two coaxial extensional events with permuted horizontal minimum and intermediate principal stresses? or ii) did the two sets develop during a single event of extension? The lack of cross cutting relationships between the two sets suggest that the minor population did not form in the same way as the major population, that is to say, perpendicular to a direction of stretching.

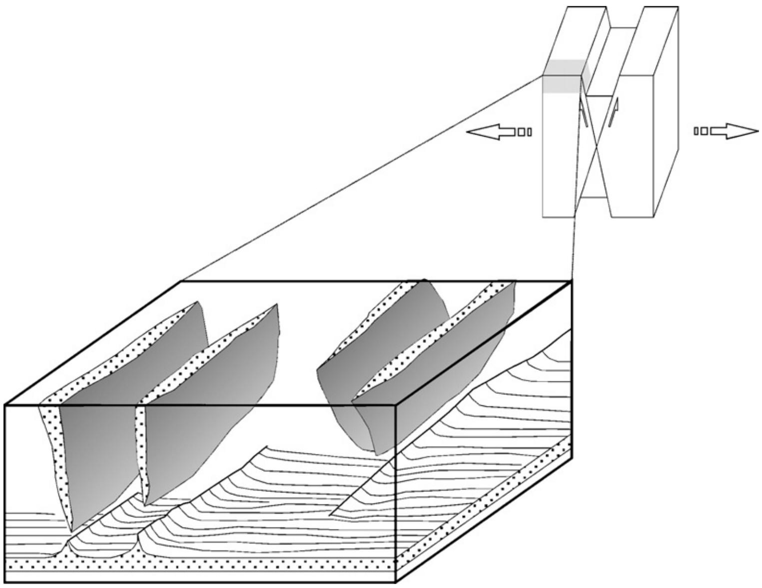

Fig. 9. Sketch of the injection of calciclastic sediments into joints and relation to the syn-sedimentary normal faults.

Alternatively, the small population of NE–SW joints, at approximately 90° to the main WSW–ESE joints can be interpreted as cross-joints. Thoroughly studied by several authors (e.g. Hodgson, 1961; Hanckock, 1985; Bai and Gross, 1999; Fabbri et al., 2001; Bai et al., 2002) the cross-joints are smaller, abut against the systematic joints and may develop at different angles to them, although for each geological context the angle between cross-joints and systematic joints has a small dispersion. Two mechanisms have been proposed for the formation of one main joint set bounding a secondary one oriented at a high angle. The couple of systematic joints and cross-joints may result from two diachronous deformation events with different stress regimes (e.g. Fabbri et al., 2001), or may develop during a single deformation event as was demonstrated theoretically by Bai et al. (2002). The lack of nodules (injected along the cross-joints) cross cutting the dykes (injected along the systematic joints), in the study area, is the best evidence for a synchronous development of the systematic joints and the cross-joints in a single deformation event.

It is important to consider what triggered the upward and downward migration of loose and fluidized calciclastic sands along planar structures, i.e. tensile joints, developed in the enclosing, semi-lithified, fine-grained limestones. In doing so it should be recalled that all these soft-sediment deformation structures and tectonic structures are planar and that they show a consistent strike. It is suggested that stretching perpendicular to N100°–110° produced tensile joints in the semi-lithified type I limestone layers; when these joints propagated upwards or downwards across the boundaries between type I and type II limestones, the coarse-grained un-lithified layers, which contained overpressured fluids were injected downwards or upwards, respectively, into the type I limestones forming load-casts or dykes.

The dependence of joint spacing on layer thickness (Fig. 8) has been demonstrated by numerous field examples (e.g. Price, 1966; Gross et al., 1995; Underwood et al., 2003) who have also considered the mechanical basis for this relationship. The data in Fig. 8 show that at the time of dyke injection the semi-lithified fine-grained limestone acted as an elastic solid.

Although earthquake vibration is the classic trigger mechanism for the injection of clastic dykes this mechanism is not envisaged by the authors as a principal mechanism for the formation of the injected clastic dykes in the study area. This is because the outcrop at Forte de Belixe does not contain systematic injection structures, or extensional faulting. It lies within 1 km of the Praia de Belixe outcrop, which is extensively cross cut by extensional faults. It is proposed that the injection of clastic dykes into the type I fine-grained semi-lithified limestones occurred during tectonic stretching and propagation of tensile joints that preceded the formation of the normal faults (Fig. 9).

The WNW–ESE orientation of the main injection structures and normal faults is sub-parallel to the main Lower Jurassic extensional faults of the Algarve Basin, which trend approximately E–W to ENE–WSW in the central and eastern part of the basin, respectively. The transient tectonic inversion event is only well documented in the study area. However, transient tectonic inversion events at basin scale were documented for the Callovian–Oxfordian and Jurassic–Cretaceous transitions (Terrinha et al., 2002), which suggests that this type of event has some cryptic relationship with the rifting process of the Algarve Basin.

The possibility of gravity tectonics causing both stretching and transient compression cannot be proved nor excluded because of lack of data. However, it should be emphasized that no mass wasting deposits of this age were reported anywhere in the Algarve Basin, which implies that in the case of a gravity driven process, the study area would then lie close to the source (or headscarps) areas. In any case, the coincidence of orientation of the main faults (WNW–ESE) and the regional structure of the Algarve Basin would imply that the gravity process was itself controlled by the geometry of the rift basin.

7. Conclusions

The conclusions of this study can be summarised as follows:

i) The load casts and derived pseudo-nodules, dykes and nodules derived as a result of disruption of dykes, all formed by the intrusion of calciclastic, loose sands into the fine-grained limestones;
ii) The systematic and pervasive occurrence of dykes and related structures by subsequent deformation of still soft-sediment, indicated injection of confined sands containing overpressured fluids;
iii) The commonly observed vertical continuity between upward injected dykes and downward intrusion of elongated load-casts and pseudo-nodules, both types aligned with WNW–ESE trending extensional joints, indicates that the calciclastic sands were injected upwards and downwards into the extensional joints;
iv) The systematic orientation of WNW–ESE extensional faults, extensional joints, dykes and load-

casts (and structures derived from these) together with the demonstrated Lower Pliensbachian age of all mentioned structures is a strong indication of tectonic control on the formation of the soft-sediment structures;

v) All the soft-sediment deformation structures and associated faults occurred during the Lower Pliensbachian;

vi) The chertification of the calciclastic bodies, including the stratigraphic horizons and described soft-sediment deformation structures occurred before the deposition of the Upper Pliensbachian;

vii) The reactivation of the WNW–ESE extensional faults and joints, breaking through the already lithified and chertified soft-sediment deformation structures demonstrates the continuation of the tectonic regime after lithification of the whole Lower Pliensbachian formation, before the deposition of the Upper Pliensbachian.

viii) The soft-sediment deformation structures described in this work formed at some time during the upper part of the Lower Pliensbachian under the active control of extensional tectonics, suffered a transient episode of tectonic inversion, an event of erosion and lithification, an event of chertification and later dolomitisation, all before the deposition of the Upper Pliensbachian, i.e. probably during a time interval of 2 Myr:

ix) The silicification is responsible for the preservation of the majority of the soft-sediment deformation structures observable in this formation.

Acknowledgments

The authors are grateful to John Cosgrove, Eugenio Carminati and Fabrizio Storti for their comments and criticism, which helped to improve the quality of the manuscript. John Cosgrove is specially thanked for correcting and improving the English.

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