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Field Trip Guidebook

Volume 2

**Cretaceous and Cenozoic Events
in West Iberia Margins**

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ORGANIZED BY

Departamento de Ciências da Terra da Faculdade de Ciências e Tecnologia da Universidade de Coimbra
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Foreword

The western and south-western Iberian Peninsula margins (Portugal) are located in a key position for the understanding of the geological evolution of the Earth, in particular the steps that led directly to its present shape and functioning. In fact, two of the main domains of the planet history are articulated around Iberia - the Atlantic and the Tethys/Mediterranean – and were the locus of key events of the transition from the Mesozoic to the present status. From the early Cretaceous to our days, the final phases of the Pangea break-up and continental dispersion revolved to the Betic compression and the presumed incipient Atlantic subduction, and its sedimentological records are landmarks (literally) of this country. This volume includes field trips dedicated to outcrops revealing several clue events in such perspectives.

The Lower Cretaceous of the western margin reflects the northward progressive rifting, with the continental break-up of successive segments related to its main unconformities. The excellent exposures of the Lusitanian Basin depocenter (Lisbon region) allow detailed palaeogeographic and palaeoenvironmental reconstitutions, as well as the recognition of 2nd and 3rd order sequences, linked either to eustasy or regional tectonics. Further north, fluvial deposits are scoped in a proposal using Sedimentology to perform a sequential approach to continental deposits...perhaps contemporary of the explosive radiation of angiosperms.

The Cenomanian western carbonate platform provides extremely rich and diverse palaeontological assemblages, testifying the articulation between the Tethyan and Boreal realms. Refined palaeoenvironmental, biostratigraphical and sedimentological documents support the interpreted roles of eustasy, diapiric tectonics, dynamics of large rudists buildups, etc. Also noteworthy is the history of the research, with a centenarian regional stratigraphic network still in use, as in the classical sea-cliffs of the proposed stops.

Among the Cenozoic record, the western margin of Portugal is quite interesting, in particular the Neogene of the Lower Tagus basin, owing to its completeness. In the proposed field trip, Sedimentology s.s. is coupled with palaeontology and isotopic data to show how this is an interface area recording eustasy and wide scale changes in Atlantic climate and circulation, but also testifying the hinterland evolution under a compressive (Betic) regime.

Sedimentation and sediments deformation, essentially post-depositional, are the main issues of the field trip proposed to the Algarve region. Both reflect the proximity of the Azores-Gibraltar plate boundary and the West-Iberia continental margin, probably in transition to a convergent regime. Details on karstic evolution and palaeoseismites supports innovative ideas debated in face of expressive outcrops.

A final remark on the Meeting Theme: for the participants, there is no need to stress the importance of the basic Scientific knowledge, or the interest for Mankind of sedimentary resources...but for all the “others”? We believe it could be an opportunity to debate how our planet works - a way to prevent bad use, how linked are the natural systems, how life evolved – how fragile and circumstantial it is (we and our ancestors)...and even some pragmatic “details”, as how hydrocarbons – the base of nowadays energy - are tracked, and how sea level, climate, rivers, topography, etc. changes continuously and sometimes catastrophically.

The editors

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Pre-Meeting Field Trip A5

**Shallow marine to fluvial Lower Cretaceous of central Portugal:
sedimentology, cycles and controls**

J. REY
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Shallow marine to fluvial Lower Cretaceous of central Portugal: sedimentology, cycles and controls

J. REY¹ & J. L. DINIS²

1. INTRODUCTION

This field trip is an invitation to observe several relevant aspects of Early Cretaceous sedimentation across the Lusitanian Basin, during a time span contemporaneous of the first stages of oceanic crust formation in the several rift segments of the Iberia western margin. Stops will be centred in the study of the main depositional environments, as well as its bio- and lithofacies distributed in time and space, first as a carbonate ramp, from the Valanginian to the Lower Hauterivian, and after that as a rimmed carbonate platform (Lower Hauterivian to Albian) with an open distal platform, coral and stromatoporoid build-ups, sandy carbonate shoals and an inner platform with rudists. Other environments, interplaying with the previous in more proximal or low sea-level situations, include lagoons, littoral plains with dinosaur tracks, intertidal sandy bars, estuary, fluvial (and palaeosols). The most widespread environments are the fluvial ones, revealing the low rate of accommodation creation.

Special emphasis will be given in the influence of the cyclic variations of sea-level, observed at different scales, and based on environmental sections and their sedimentary record: i) organisation of the third order depositional sequences within the half transgressive and regressive cycles of second order; ii) composition of the

transgressive and high sea-level successions within the third order depositional sequences.

Concerning fluvial to paralic siliciclastics, the excellent exposures allow a detailed analysis of depositional architecture at several scales. The preliminary sedimentological interpretations points to mainly braided systems, dominated by braid bars bodies, but including some levels reflecting sinuous behaviour and preserved floodplain deposits and palaeosols, and operating in a low accommodation context.

The main unconformities and the levels immediately above are supposed to reflect important tectonic events linked to the episodes of the discrete northward opening of the Atlantic Ocean in the western Iberian margin.

2. THE LOWER CRETACEOUS OF THE LUSITANIAN BASIN

2.1. The basin morphology

The Cretaceous series are known between Aveiro, to the north, and the Sado River, to the south, but the chronological significance of the preserved deposits is diverse, allowing the division in two parts around the Caldas da Rainha or Nazaré parallels, being the Lower Cretaceous series well documented in the southern part. The basin was structured by Hercynian faults N/NE-S/SW orientated; wedged to the East by the Hercynian basement of the Hesperian Massif and rimmed at the western border by marginal basement horsts as the metamorphic and granitic Berlengas and Farilhões islands.

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The southern part of the basin comprises three morphological units (Rey, 1972) with a tilted block arrangement: i) a central trench with a complete (at the stage scale) and mixed (marine, non-marine; carbonated, siliciclastic) Lower Cretaceous succession. The depth and the rate of sedimentation increase to the southwest, and the depocenter is generally in the vicinity of Cascais; ii) eastern and western margins, with significant sedimentary hiatuses and essentially continental deposits.

2.2. The sedimentary infill and the palaeogeographic evolution

The Lower Cretaceous sedimentary series consists of (Fig.1, Pl. I):

- 9 formations between the Valanginian and the Albian, with dominant limestones in the Cascais and Sintra area (Rey, 1992). The total thickness is about 400 m;

- 10 formations north of the Espichel Cape and in the Ericeira area. Marine limestones alternate with lagoonal, brackish and estuarine dolomites and marls, and with fluvial conglomerates, sandstones and mudstones. The total thickness is about 250 m in the Ericeira area;

- 7 siliciclastic formations between Ericeira and Torres Vedras and east of Loures, mainly non-marine but including minor transitional to shallow marine deposits;

- exclusively siliciclastic deposits in the on-shore north of the Torres Vedras region, fluvial with rare signs of paralic environments, grading upward and westward to Cenomanian epicontinental carbonates; the age is poorly known and correlation difficult for isolated outcrops as the Cercal, Olhalvo, Alcanede and Galiota ones, and presumed as late Aptian to late Cenomanian for the Figueira da Foz Formation (north of the Nazaré parallel).

STAGE	T - R CYCLES	3rd ord. SEQU.	LITHOSTRATIGRAPHIC UNITS									
			Ericeira	Cascais, Sintra	Espichel Cape							
ALBIAN	up.	Al 11	GALÉ	Ponta da Galé	?							
		Al 10		GALÉ		Água Doce						
	Al 9											
	Al 8											
	Al 7											
	mid.	Al 6										
Al 5												
low.	Al 4	RODÍZIO	RODÍZIO	RODÍZIO								
APTIAN	up.					CRISMINA	Praia da Lagoa Ponta Alta Cobre	CRISMINA				
									Praia da Lagoa Ponta Alta Cobre			
	low.				Ap 4					CRISMINA	Praia da Lagoa Ponta Alta Cobre	CRISMINA
					Ap 3							
	Ap 2											
	Ap 1											
BARREMIAN	up.	Ba 6	REGATÃO	REGATÃO	REGATÃO							
		Ba 5										
	low.	Ba 4				GUINCHO	BOCA DO CHAPIM PAPO SECO AREIA DO MASTRO ROCHADOURO					
		Ba 3										
		Ba 2										
		Ba 1						RIBEIRA DE ILHAS	LADEIRAS Lagosteiros			
	Ba 0	RIBAMAR										
	Ha 7											
HAUTERIVIAN		Ha 6	PRAIA DOS COXOS	CABO RASO	MACEIRA							
		Ha 5										
		Ha 4										
		Ha 3										
VALANGINIAN		Ha 2	SANTA SUSANA	MACEIRA	MACEIRA							
		Ha 1	SÃO LOURENÇO	GUIA	GUIA							
		Va 7	VALE DE LOBOS	SERRADÃO	VALE DE LOBOS							
		Va 6										

Fig. 1 - The Lower Cretaceous lithostratigraphic units in the southern part of the Lusitanian Basin. Ages, depositional sequences and 2nd order transgressive-regressive cycles.

Fossils with a precise chronostratigraphical significance are scarce both in marine and non-marine deposits. However, using the concepts of sequence stratigraphy, it is possible to identify time lines and to draw up reliable correlations between outcrops of the various areas (Pl. I). That way, a short evolution of the sedimentary evolution of the basin can be suggested:

a) Latest Jurassic – Early Berriasian. The carbonate platform, located in the Cascais and Sintra area, corresponds essentially to brackish waters, as proved by the predominant sedimentation of limestones and marls with litiolids, charophytes and lagoonal ostracods (“Purbeckian facies”). This environment is surrounded to the north, east and south by a coastal plain – tidal or estuarine flats – with mudstone and sandstone deposition.

b) Late Berriasian – Early Valanginian. The coast line did not moved too much during this period. However, the marine carbonate platform becomes deeper, in accordance with the development of limestones and marls on an open and subtidal ramp, whereas a clear inflow of coarse, kaolinitic sandstones spreads in estuarine and fluvial environments.

c) Late Valanginian – Earliest Hauterivian. The relative sea level rises rapidly, and the areas of Cascais, Sintra and the Espichel Cape become an open shelf, reaching 40 to 50m deep at the Valanginian – Hauterivian boundary; jointly with a cephalopods sudden increase, the sedimentation was initially calcareous and turning marly. The sea progressed widely to the north and to the east, since tidal flats reach the vicinities of Torres Vedras and Alenquer.

d) Hauterivian – Early Barremian. With the maximum flooding during the early Hauterivian (Fig. 2), reefal buildups appear north of Espichel Cape and near Cascais, where the thick Cabo Raso reef was grown. Afterwards, the gradual infill of the basin resulted from the progradation of depositional systems. In the Cascais and Sintra area, the buildups are overlaid by limestones with rudists and dasycladaceae characterising an inner shelf. In the Arrábida hills and near Ericeira, subtidal inner marls and limestones with rudists and echinoids interbedded with lagoonal dolomites and sandstones, after minor sea level changes. Eastwards and northwards, braided alluvial plains draining westwards. The maximum regression was reached at the end of the early Barremian.

e) Late Barremian. The Lusitanian Basin is invaded by non-marine siliciclastic sediments, corresponding to

widespread fluvial environments, except in a narrow belt between Ericeira, Cascais and the Espichel Cape with predominant deposition of sandstones, mudstones and dolomite in coastal plains and estuarine domains.

f) Early Aptian. A progressive sea level rise allows the establishment of a new carbonate platform with buildups. In the western part of the Arrábida hills, in the Cascais and Sintra areas, as near Ericeira, the same succession is recognised, with a great uniformity of facies in space: limestones and marls of inner shelf are followed by coral and rudists biostromes associated to grainstones, then by marls with oysters indicating protected marshes. A coastal fringe of fine sandstones was identified to the north and to the east.

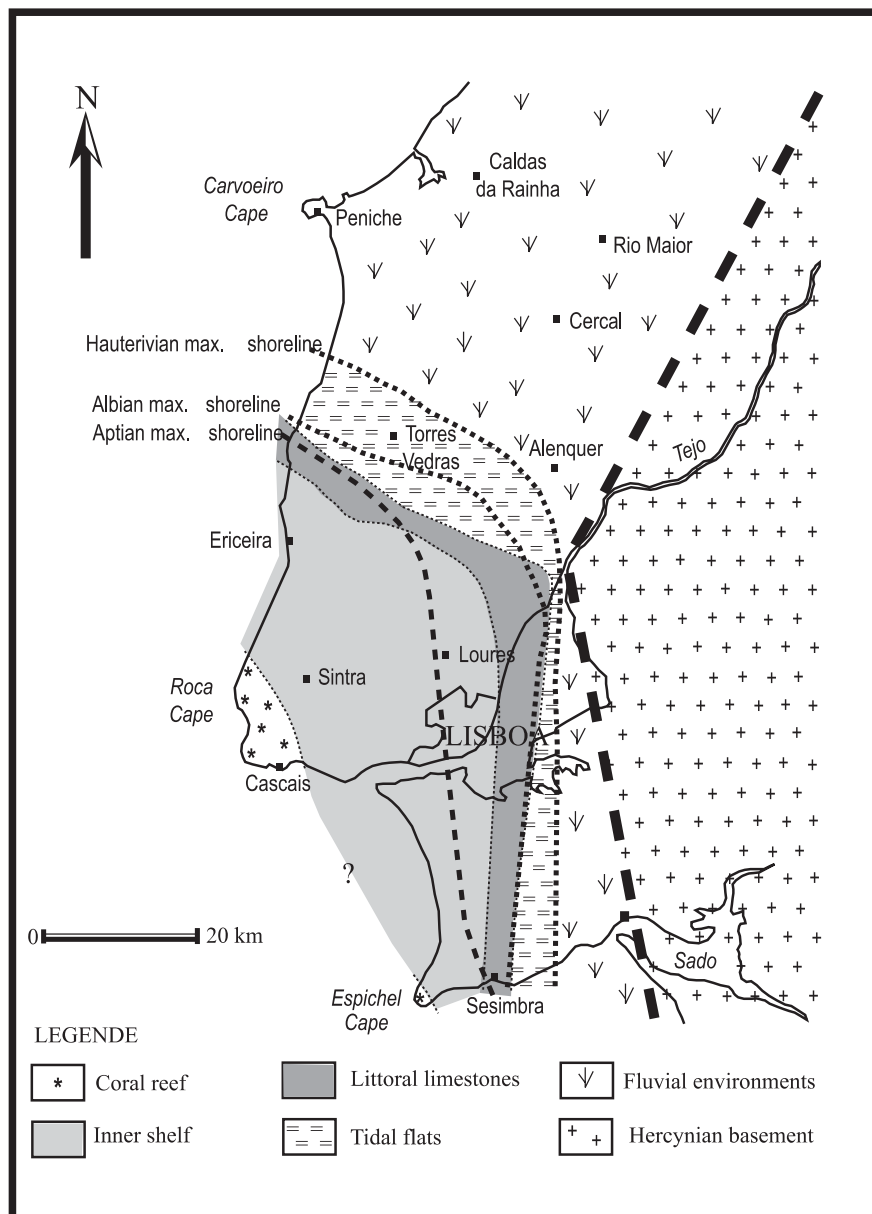
g) Latest Aptian – Early Albian *pro-parte*. An alluvial sedimentation in braided rivers suddenly occupies the entire basin. Conglomerates, sandstones and mudstones accumulate, cutting more or less deeply the underlying deposits. The palaeocurrent data, the morphology, size and nature of particles seem to indicate a double provenance: a prevalent westward contribution from the Hercynian basement and another from a granitic or gneissic relief located at the east (similar to the Berlengas block).

h) Albian. The beginning of a widespread long-term transgression is the main control of the Albian in the basin, while another carbonate platform with build-ups was created in the Cascais and Sintra area. Environmental belts can be deduced for this stage: entirely alluvial systems, siliciclastic coastal plain, marly inner shelf with oysters, orbitolinids and green algae, and a barrier of shoal calcarenites (grainstones) with rudist patch-reefs. During the Albian, these various facies shift north and eastwards. However, they remain restricted south of Ericeira (Fig. 2). A break marks the Albian – Cenomanian boundary, quite clear in the southern sector. Within the mainly fluvial deposits, changes in depositional systems slope interpreted from the grain-size and the depositional architecture distribution can reflect minor sea level changes and/or small vertical adjustments in the feeding area.

2.3. The geodynamic evolution

Six main stages are distinguished in the tectonic-sedimentary history of this basin during the Early Cretaceous, related with the geodynamic events in the western Iberian margin (Whitmarsh *et al.*, 1996; Srivastava *et al.*, 2000; Whitmarsh & Wallace, 2001):

Fig. 2 - Palaeogeographic sketch map of the southern Lusitanian Basin during the Early Hauterivian (sequence Ha 2) and boundaries of the Aptian and Albian transgressions.



a) The Neocimmerian crisis. This tectonic event occurs *circa* the Berriasian-Valanginian boundary. It is shown by: i) the deepening of the carbonate shelf in the vicinity of Cascais and Sintra; ii) the sudden arrival of coarse clastics (up to pebbles) to the borders of the marine platform; iii) the angular unconformity of Valanginian fluvial sandstones over Upper Jurassic beds (as next to Torres Vedras and Cercal). This event can be related with the final phase of rifting and eventual break-up of the Iberian segment of the margin.

b) The Late Valanginian – Early Barremian transgressive – regressive cycle. The Lusitanian Basin is stable. The deposits are arranged in a 2nd order transgressive-regressive cycle and twelve 3rd order depositional sequences. The maximum transgression at the 2nd order scale, during the Early Hauterivian, is expressed by the development of reefs near Cascais, by the sedimentation

of intertidal to supratidal marls in the Torres Vedras area, and by fine siliciclastic deposits in the non-marine environments.

c) The Middle Barremian crisis. This event is revealed in the calcareous series of Cascais by a karstic surface, and in the mixed series of Ericeira and the Espichel Cape by the arrival of an erosional mass of coarse sandstones. It seems to be related with a pulse of seafloor spreading rate on the Tagus segment and, perhaps, the onset of the oldest oceanic crust in the Iberian and/or Galicia segments.

d) The Late Barremian – Early Aptian transgressive – regressive cycle. This 2nd order cycle includes eight 3rd order depositional sequences. The maximum transgression at the 2nd order scale, during the middle Aptian,

is marked by a condensed interval. It is less important (in time and sedimentary volume) than in the previous cycle. It is contemporary of an anoxic event in the ocean (Jenkins, 1980).

e) The Austrian crisis. A major tectonic event occurs in the Lusitanian Basin in the late Aptian. A sudden influx of high-energy clastic and fluvial deposits covers the whole basin and previous series are more or less deeply eroded. Angular unconformities (of cartographic scale) are recognised, particularly in the eastern margin and the north of the basin (Dinis & Trincão, 1995; Cunha & Reis, 1995). This phase is probably related to an extensional pulse near the Galicia triple point, eventually the beginning of seafloor spreading in the Galicia northern margin. Nevertheless it is clear in several nearby basins (Hiscott *et al.*, 1990).

f) The Albian transgressive – regressive cycle. The gradual and generalised flooding of the southern part of the Lusitanian Basin during the Albian results of a global sea level rise well known on various stable platforms. It is followed by a short regression during the uppermost part of Albian (Vraconian). This cycle is realised by the stacking of eight 3rd order depositional sequences that were identified in coastal and inner-middle shelf environments. It includes two sub-cycles (Dinis *et al.*, 2002) that can be followed in the mainly fluvial siliciclastics of the northern part of the basin and reflects probably the Iberian plate transpression with Africa and Europe.

The comparison with other West European sedimentary basins shows the same number of 3rd sequences and a similar evolution on the North Atlantic margins (Rey



Fig. 3 - Road map and location of stops.



Fig. 4 - The cliffs of the Maceira cove. a: Western part (Guia); b: Eastern part (Mexilhoeira).



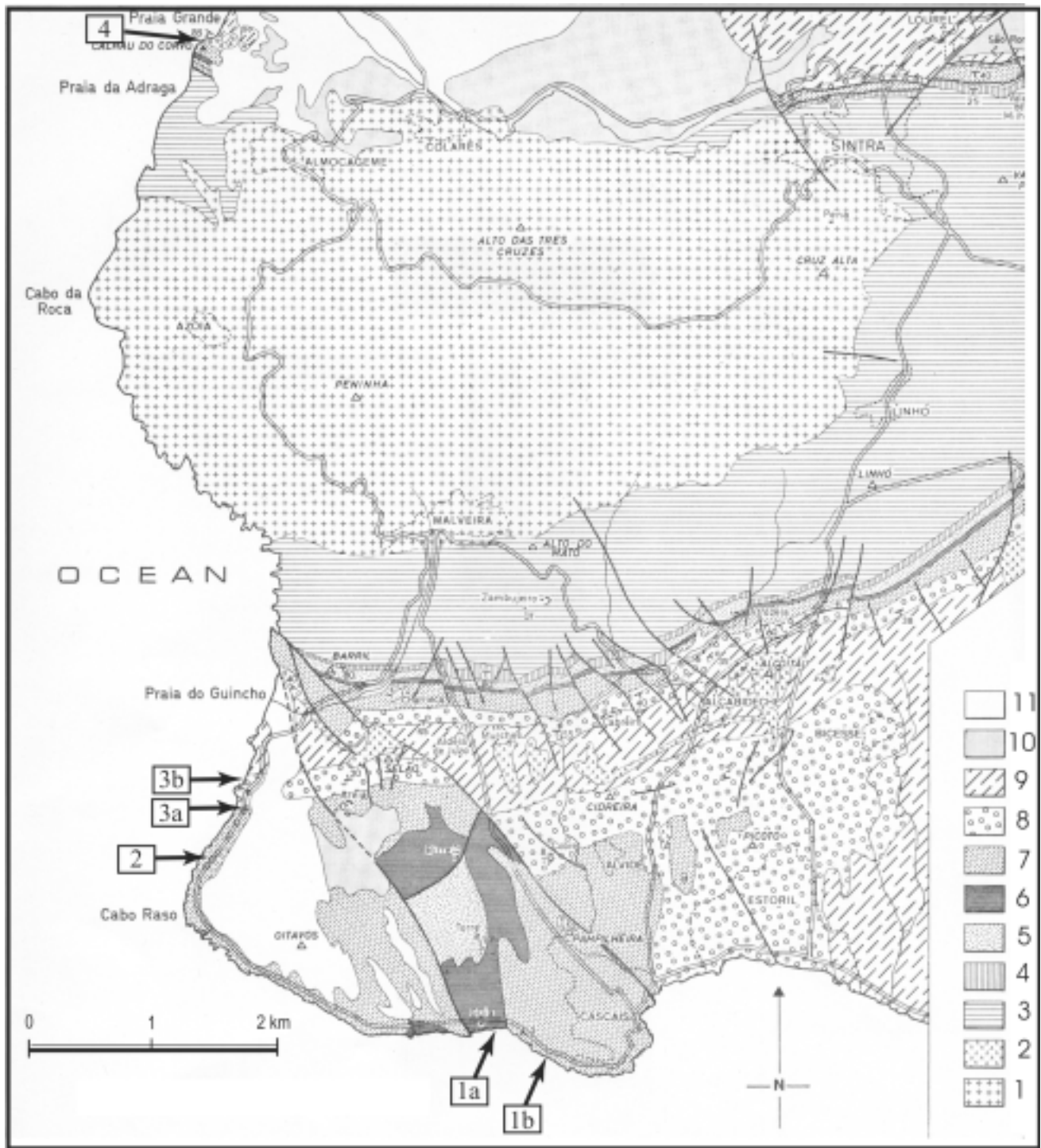


Fig. 5 - Geological map of the Cascais area and location of stops. 1: Granite, granodiorite; 2: volcanic rocks; 3: Jurassic; 4: Berriasian; 5: Valanginian; 6: Lower Hauterivian; 7: Upper Hauterivian-Lower Barremian; 8: Upper Barremian, Aptian, Lower Albian p.p.; 9: Albian-Cenomanian; 10: Tertiary; 11: Quaternary.

et al., 2003). This evolution is partially different from the one recorded on Tethyan margins. Therefore, the depositional succession known in Portugal – that indicates strong sea level fluctuations created by tectono-eustasy – reflects the structural evolution of the West European craton and the northwards progressing rift-to-drift events of the Atlantic margins.

3. FIELD DESCRIPTIONS OF THE LISBON REGION

Considering the influence of the relative sea level changes at various scales on the sedimentary features and on the distribution of environments, the stratigraphic series and sedimentary bodies will be described with reference to 3rd order depositional sequences, systems tracts and parasequences.

3.1. Stop 1: The environments of a carbonated ramp during a transgressive half cycle: the sequences Va 6, Va 7 and Ha 1 from the Maceira cove

Between Boca do Inferno and the Guia lighthouse, the cliffs of the Maceira cove allow the analyse of the uppermost Valanginian and basal Hauterivian sequences (Figs. 3 to 5). The whole V6 sequence and the lower part of the V7 sequence outcrop well on the western cliff (Guia), while the upper part of the V7 sequence and the whole H1 sequence can be observed on the eastern cliff (Mexilhoeira).

3.1.1. Sequence Va 6 (Fig. 6)

S.B./T.S. A horizontal surface separates a bed of sandy limestone from a bed of sandy marls.

T.S.T.: It consists of three thinning-upward parasequences with a thin bed of black laminated sandy marls and a bed of bioturbated sandy limestone, topped with an oxidized surface. Decreasing thickness and increasing oxidisation and bioturbation to the following parasequence show the retrograding evolution on the upper shelf. The fauna is diverse: oysters, brachiopods, gastropods (*Nerinea*, *Natica*), bivalves (*Pterotrigonia*, *Fimbria*) and lituolids.

M.F.S.: The last and the thickest iron crust, with important horizontal bioturbation filled with limonite.

H.S.T.: It indicates a shallowing-up evolution in limestones and marls. The decreasing of depth is expressed by fossils (brachiopods and oysters in the lower part; algal films, *Ptygmatis* and oncolites in the upper part), by sedimentary structures (laminations and wave ripples in the lower part; pedogenic nodules and carbonate concretions in the upper part). The last level is bored by roots at the top, indicating a palaeosol.

3.1.2. Sequence Va 7 (Fig. 7)

S.B./T.S.: Exposure surface, followed by yellow sandy clays reworking the level below.

T.S.T.: The thickest systems tract. The basal part consists of laminated black marls alternating with undulated beds of marly limestones, or of silty-sandy, bioturbated and lignite-rich limestone, and of lenses of coarse sandstones. These deposits are arranged into a cyclic parasequence, in estuarine environments. They are topped by some thin beds of sandy limestones. The upper part is represented by undulated beds of oolitic and gravelly, sparitic cemented limestones from the middle and lower shelf. From the base to the top, the thickness of the last beds decreases and the faunal diversity increases (echinoids, brachiopods, bivalves, corals; ammonites and belemnites at the top). The oxidised surfaces are progressively more marked and closer. The density of bioturbation increases. These facts indicate a decreasing of the sedimentation rate and a deepening-up evolution.

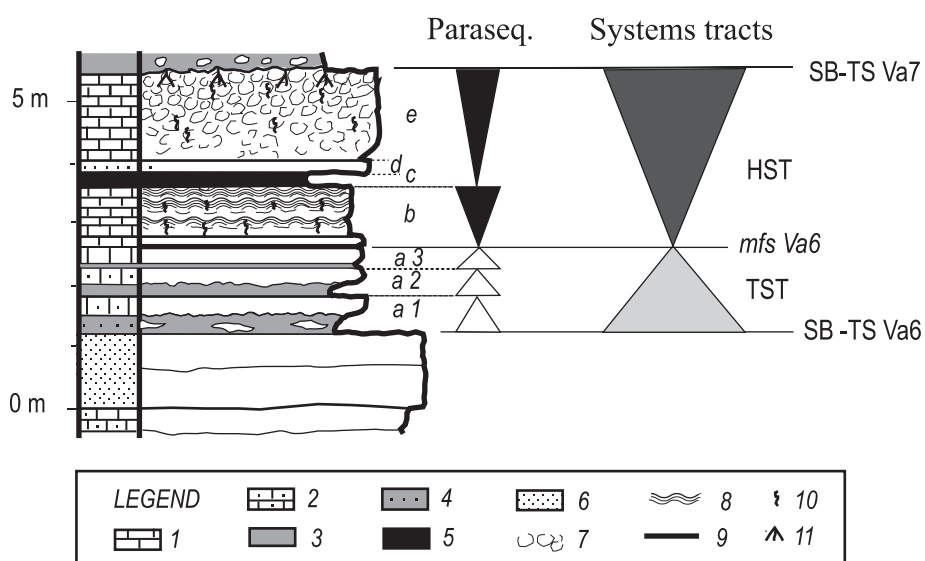


Fig. 6 - The sequence Va 6 at Guia. Lithology, stratonomy and sequential arrangement. 1: limestone; 2: sandy limestone; 3: marl; 4: sandy marl; 5: black mudstone; 6: sandstone; 7: pedogenetic nodules; 8: algal films; 9: oxidized surface; 10: bioturbations; 11: roots prints.

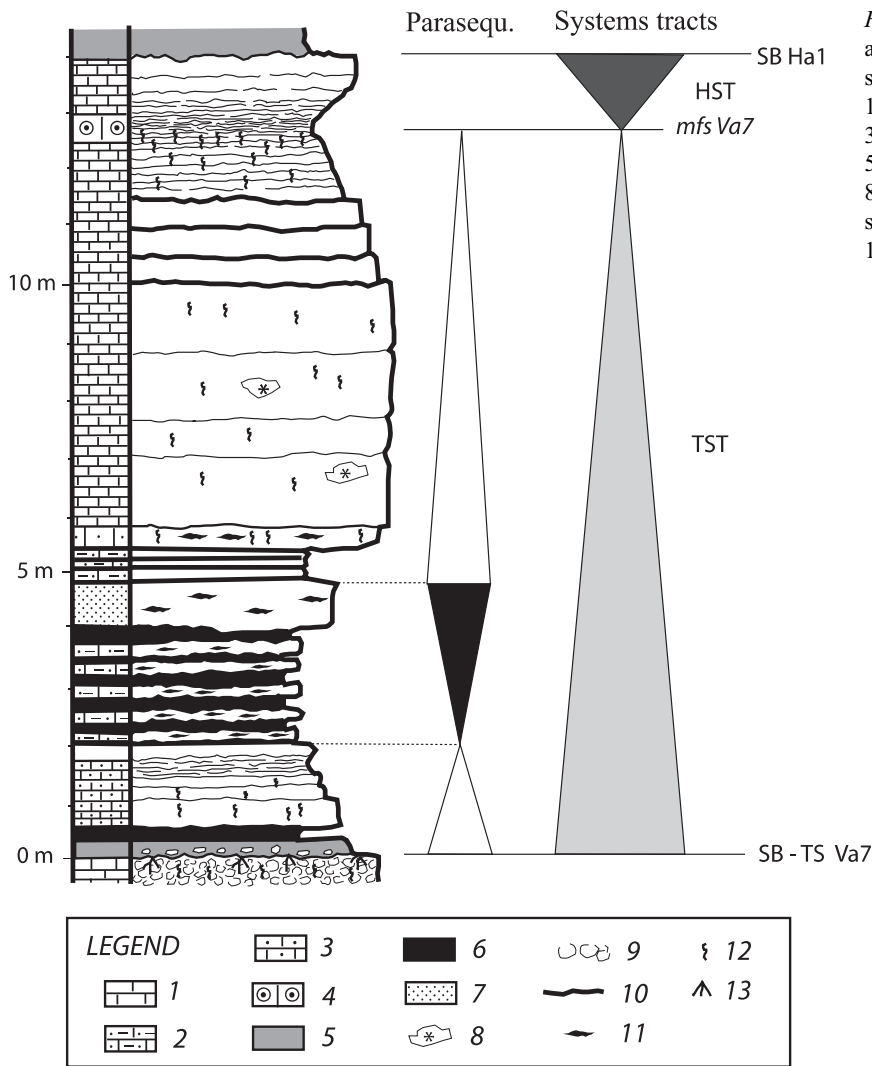


Fig. 7 - The sequence Va 7 at Guia (lower part) and Mexilhoeira (upper part). Lithology, stratonomy and sequential arrangement.

1: limestone; 2: sandy-muddy limestone; 3: sandy limestone; 4: ferruginous oolites; 5: marl; 6: black mudstone; 7: sandstone; 8: corals; 9: pedogenetic nodules; 10: oxidized surface; 11: lignite; 12: bioturbation; 13: root prints.

M.F.S.: The maximum flooding is represented by a condensed interval, with a very abundant and varied fauna (ammonites, Echinoids, brachiopods, corals (*Montlivaltiidae*), gastropods, bivalves...).

H.S.T.: It is materialised by a microclastic limestone with ammonites and nautiloids, stacked in some thickening-up beds with a sparse bioturbation. This systems tract disappears 500 m to the West, where it is replaced by an iron crust.

This sequence Va7 is a very typical backstepping sequence. The Valanginian – Hauterivian boundary is between the condensed interval and the H.S.T.

3.1.3. Sequence Ha 1 (Fig. 8)

S.B.: Irregular surface, with bores filled by the silty marls of the overlying level.

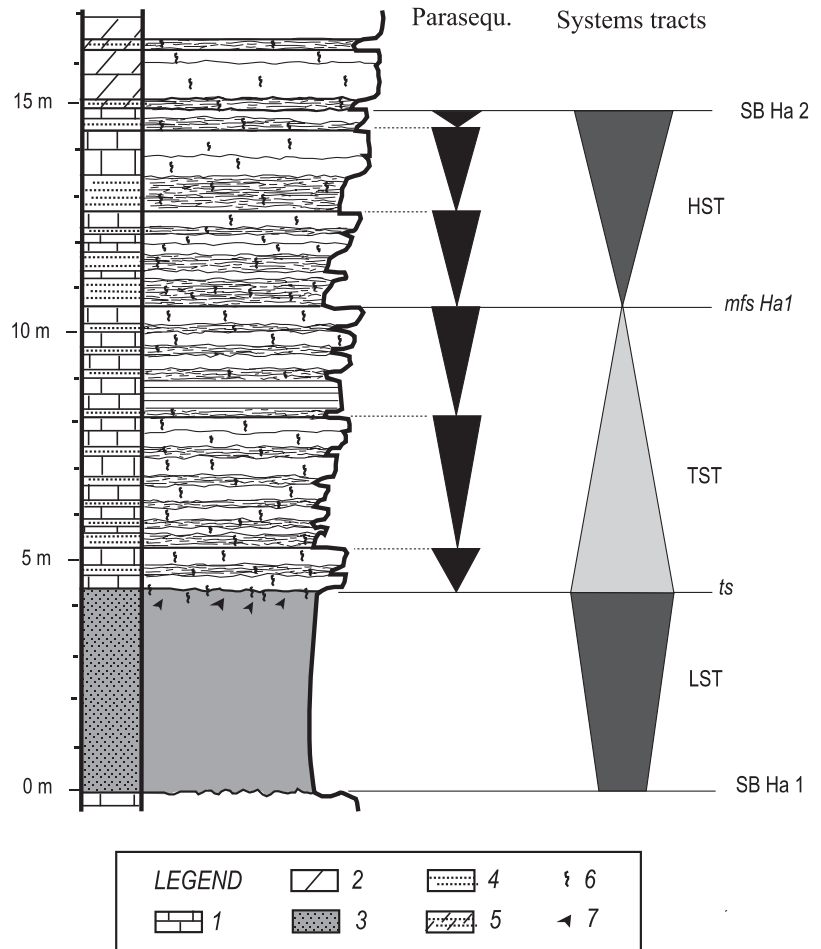
L.S.T.: massif grey silty marls with muscovite and (on the top) gypsum. Fossils are scarce and flattened.

1st F.S.: surface marked by the development of bioturbation on the terminal part of marls, and by the appearance of thin and discontinuous marly limestones.

T.S.T.: These two systems tracts consist of the stacking of six parasequences of muddy limestones and limestones. The limestone beds are thickening-up in each parasequence set. The stratification is irregular and the texture is nodular, in connection with horizontal bioturbation. The limestones contain pellets, intraclasts and bioclasts in a micritic matrix. The fauna is rich: ammonites, echinoids, brachiopods, gastropods (*Natica*) and bivalves. It is difficult to separate the T.S.T. from the H.S.T, due to the bad characterisation of the maximum flooding surface. This boundary may be located around the muddiest level. A toplap cuts this sequence.

This sequence Ha1 is an aggrading sequence developed in deep subtidal shelf (circalittoral environments).

Fig. 8 - The sequence Ha 1 at Mexilhoeira. Lithology, stratonomy and sequential arrangement. 1: limestone; 2: dolomite; 3: silty marl; 4: muddy limestone; 5: muddy dolomite; 6: bioturbations; 7: evaporite.



3.2. Stop 2: The reefal complex of the Raso Cape (Hauterivian)

The Raso Cape (Cabo) outcrops show a recrystallized and dolomitized complex, dated as Hauterivian. The original micritic cement is locally preserved. The palaeontological assemblage is constituted of stromatoporoids, chaetetids and scleractinia (*Astrocoenidae*, *Stylinidae*, *Thammasteriidae*, *Microsolenidae*, *Calamophyllidae*), associated with spines of *Cidariidae* and bioturbations.



Fig. 9 - Corals of the Cabo Raso Formation.

The colonies are encrusted, dome-shaped or ball-shaped (Fig. 9), with 10 to 20 cm average diameter. They are quite dense, with a bafflestone arrangement. Rather than a bioconstruction, this formation represents an active bioaccumulated unit. Its thickness should be about 50 m.

The sequence stratigraphy inner the Cabo Raso Formation is yet to be done. The only identified element is a level of limestone breccias, 2 m thick, known in all the Cascais and Sintra area, with broken, laminated and well-sorted fragments of rudists, nerineas and corals. The basal surface of this level could be considered as a surface boundary. Above this bed, limestones with large nerineas and rudists represent back reef environments.

3.3. Stop 3a: The lagoonal environments during a transgressive half cycle: the sequences Ba 3, Ba 4 and Ba 5 from the Crismina cliffs

The 2nd order transgressive – regressive cycle of Late Barremian – Early Aptian begins with three 3rd sequences (Fig. 10): the first two sequences are aggrading and settle in lagoonal area; the third sequence shows some shallow marine levels and therefore indicates a retrograding evolution.

3.3.1. Sequence Ba 3

This sequence consists of yellow, bioturbated, primary dolomites alternating with green or blue argillaceous marls. The deposition was in a restricted lagoon, mixing fresh and salt waters, without waves and tides. A thin bed of fine sandstone could point out the maximum flooding surface, and algal films the top of the H.S.T.

3.3.2. Sequence Ba 4

It is composed of the same facies as the previous sequence, with three parasequences of green clays and yellow dolomites in the T.S.T., and two parasequences in the H.S.T. Pedogenic alterations or rhizoconcretions appear on the top of H.S.T. dolomitic beds.

3.3.3. Sequence Ba 5

S. B./T.S.: Red clay underlying the uppermost bed of dolomites with root prints of the previous sequence.

T.S.T.: It consists of three deepening-up parasequences: the first one is made up of sandstones, bioturbated dolomites and clays; the second one consists of dolomitic

and bioturbated limestone with oysters covered by black marls; the third one is composed of yellow and nodular limestones with gastropods (*Trochactaeon*, *Pyrazus*, *Ptygmatis...*) echinoids and bivalves.

M.F.S.: Discontinuous interbed of green clay.

H.S.T.: It corresponds to a thick bed of white micritic and bioturbated limestone, with Bivalves, gastropods and green algae. The thickness of this system tract is very small (1 m).

The following sequence boundary, nowadays hid by a wall, puts the previous limestone in contact with siliciclastic deposit.

3.4. Stop 3b: The carbonate rimmed shelf environments during a transgressive half cycle: the sequences Ap 2 and Ap 3 at Crismina

3.4.1. Sequence Ap 2 (Fig. 11)

S. B./T.S.: A bored surface separates a bed of sandy limestone from a bed of green silty marls.

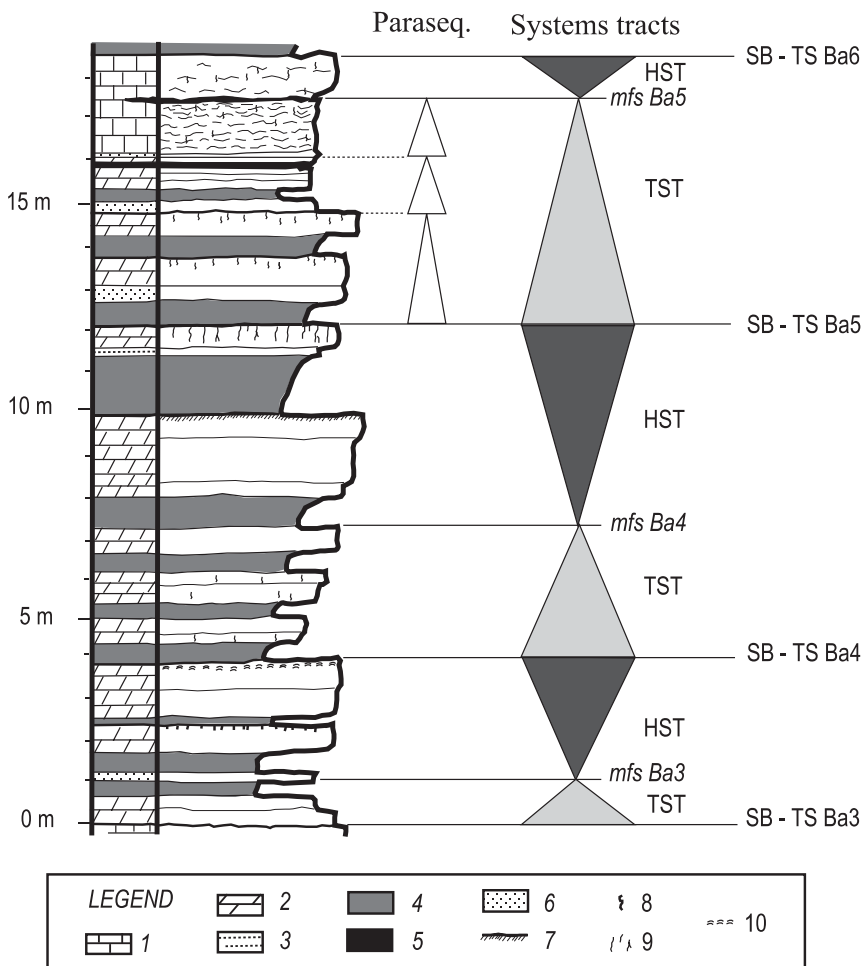
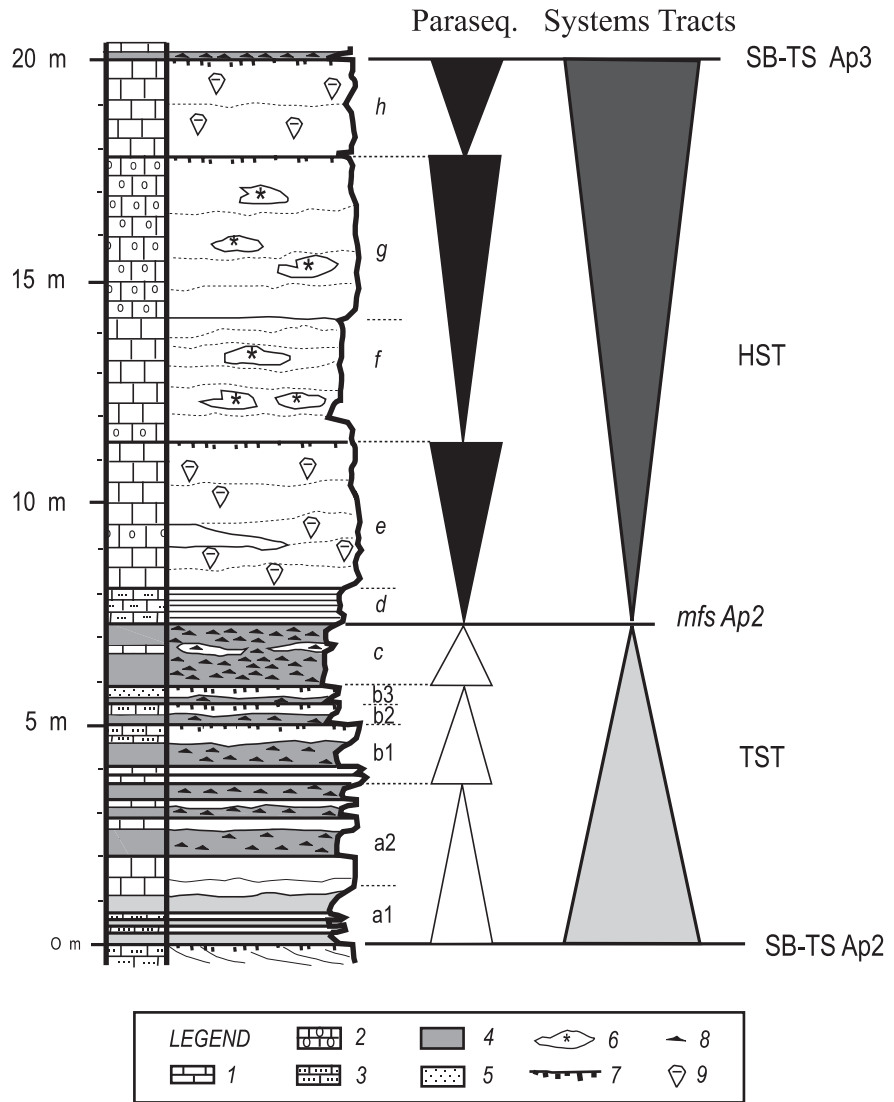


Fig. 10 - The Ba 3, Ba 4 and Ba 5 sequences at Crismina. Lithology, stratonomy and sequential arrangement. 1: limestone; 2: dolomite; 3: siltstone; 4: green and blue clay; 5: black mudstone; 6: sandstone; 7: pedogenetic nodules; 8: bioturbations; 9: rhizoconcretions; 10: algal films.

Fig. 11 - The Ap 2 sequence at Crismina. Lithology, stratonomy and sequential arrangement.

- 1: mudstone, wackestone;
- 2: grainstone; 3: sandy limestone;
- 4: marl; 5: sandstone; 6: corals;
- 7: hard ground; 8: *Palorbitolins*;
- 9: rudists.



T.S.T.: It consists of three thinning-up and deepening-up parasequences sets: the first one associates green marls, sandy limestones, bioturbated micrites and calcareous marls with *Palorbitolina*; the second one consists of sandstones, marls with *Palorbitolina* and sandy limestones topped by hard-grounds; the last is composed of brown calcareous marls, very rich in *Palorbitolina* (“orbitolinite”), that constitute a condensed interval.

M.F.S.: Top of the condensed interval.

H.S.T.: Above thin beds of sandy limestone, this systems tract essentially shows grainstones and micrites with rudists (*Requienidae*, *Caprotinidae*), nerineas, or with stromatoporoids and scleractinia (*Montlivaltiidae*, *Astrocoenidae*, *Faviidae*). Some surfaces with dissolved shells indicate subaerial exposures. Therefore, the sedimentary floor is very close to the sea level, and the relative sea level rise only creates an accommodation that is quickly infilled by the bio-sedimentary production.

3.4.2. Sequence Ap 3 (Fig. 12)

This sequence is a backstepping sequence that points out the maximum transgression of the Late Barremian – Early Aptian 2nd order cycle, during the Deshayesi Zone.

S.B./T.S.: Hardground with burrows, iron crust and silicified fossils.

T.S.T.: It consists of six retrograding and thin parasequences including a lower bed of orbitolinids - rich marls, an upper bed of limestone and topped by an oxidised surface. It is a condensed interval, with abundant and diverse fossils: echinoids, brachiopods, bryozoa, bivalves, gastropods, rudists, *Palorbitolina*, *Praeorbitolina*, etc.

M.F.S.: The last and most marked oxidised surface.

H.S.T.: It comprises two parts: i) a lower part of mudstones and wackestones with *Choffatella* and

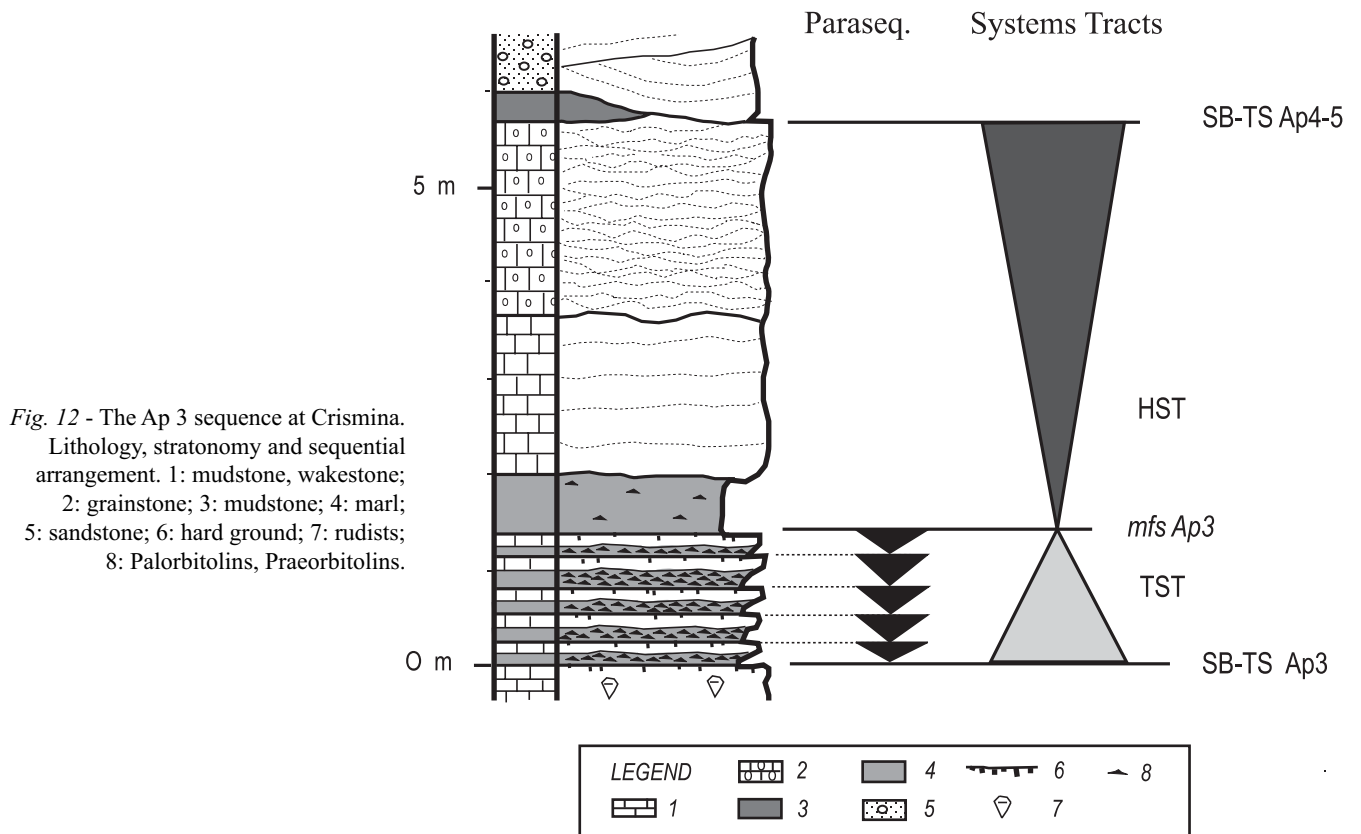


Fig. 12 - The Ap 3 sequence at Crismina. Lithology, stratonomy and sequential arrangement. 1: mudstone, wackestone; 2: grainstone; 3: mudstone; 4: marl; 5: sandstone; 6: hard ground; 7: rudists; 8: Palorbitolins, Praeorbitolins.

palorbitolins, topped by a large-waved surface; ii) an upper part of grainstones with cross-stratification. The bioclasts of rudists, corals and echinoderms are the most numerous. The level should represent the fore-reef of a biostrome identified some kilometres to the East, North to Cascais.

The top of this systems tract is an oxidised surface that represents the Ap4 sequence boundary. At Crismina, this last sequence only consists of a thin and discontinuous bed of green marl, truncated by the erosive and fluvial sandstones of the Rodízio Formation (latest Aptian – early Albian).

3.5. Stop 4: The carbonate protected shelf environments during a regressive half cycle: the sequence Ap 4 at Praia Grande do Rodízio

This last marine sequence of the Late Barremian – Early Aptian cycle is generally eroded by the underlying sandstones. However, it well appears on the cliffs South to Praia Grande do Rodízio, with the following characters (Fig. 13):

S.B./T.S.: micro-karstified and brecciated surface topping the grainstone of the sequence Ap 3.

T.S.T.: It includes an alternation of black marls with oysters and argillaceous bioturbated limestones with *Choffatella* and *Palorbitolina*. These facies are associated into shallowing-up parasequences bounded by hardgrounds. The first hardground shows remarkable dinosaur footprints (*Iguanodon* and *Megalosaurus*, Madeira & Dias, 1983).

M.F.S.: This surface is expressed by the top of the last and most important marl level.

H.S.T.: It comprises three beds of yellow argillaceous limestones with oysters, separated by yellow and purple marls. These highstand deposits are cut by the basal erosive surface of the Rodízio Formation.

3.6. Stop 5: The lagoonal environments during a regressive half cycle: the sequence Ha 6 near Ribamar

The cliffs next to the village of Ribamar shows the facies and stratal patterns of some 3rd depositional sequences during the Late Hauterivian – Early Barremian regressive half cycle (Fig. 14). We will only examine the sequence Ha 6 (Fig. 15), deposited in a lagoonal and estuarine context.

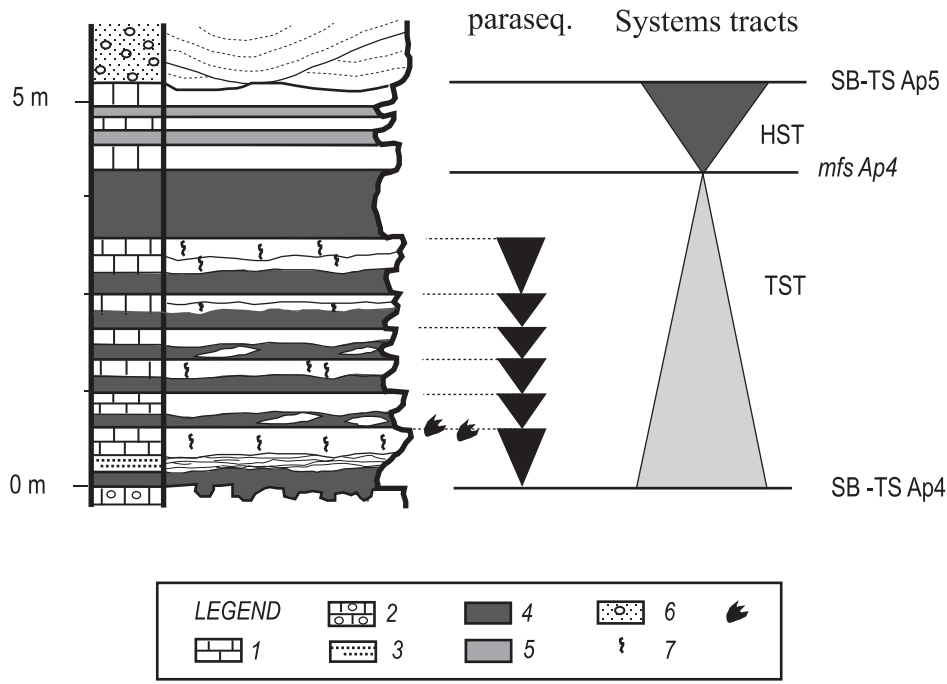


Fig. 13 - The sequence Ap 4 at Praia Grande do Rodizio. Lithology, stratigraphy and sequential arrangement. 1: mudstone, wackestone; 2: grainstone; 3: muddy limestones; 4: black marl; 5: yellow and purple marl; 6: sandstone; 7: bioturbations; 8: dinosaur footprints.

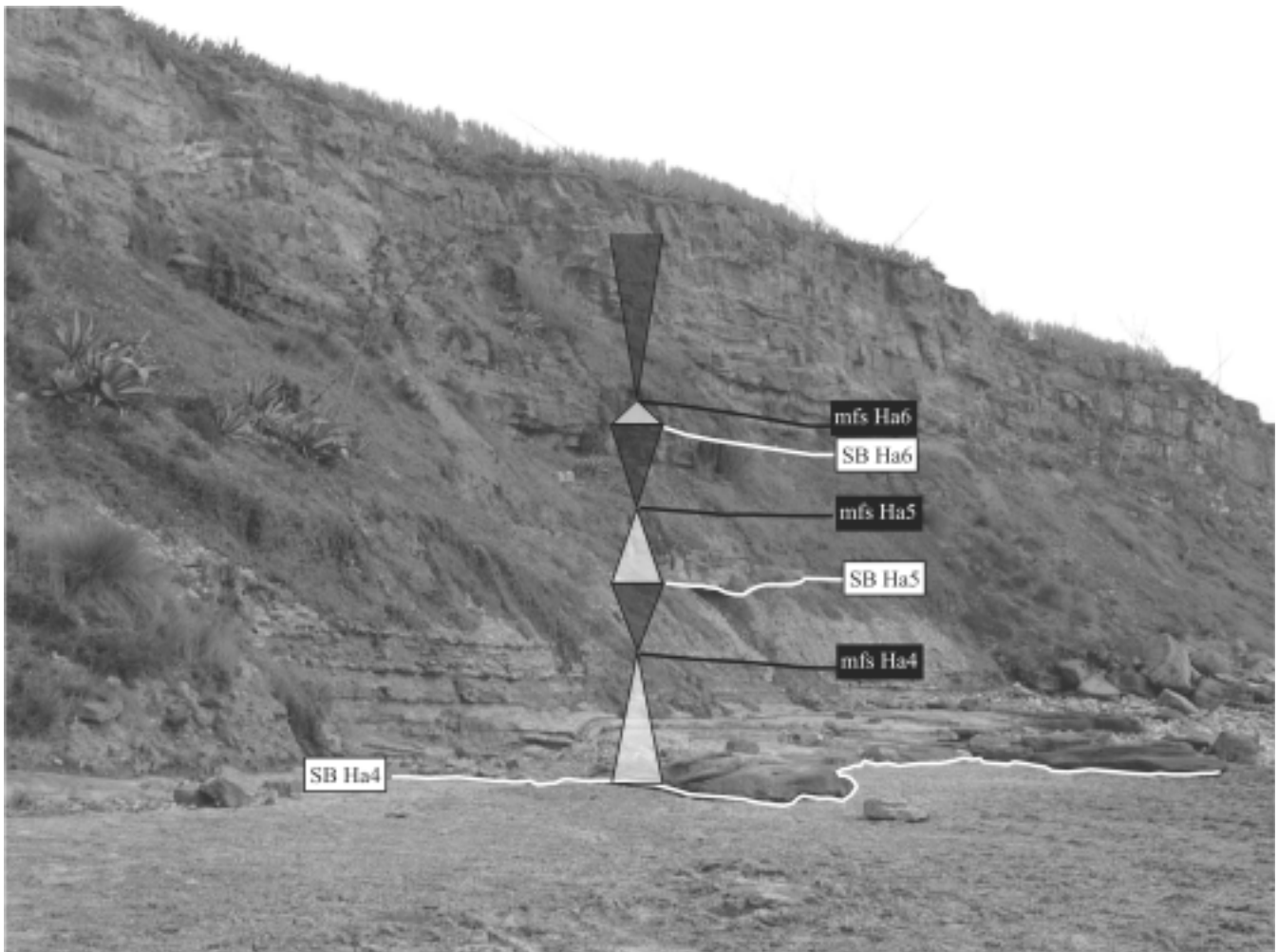


Fig. 14 - The cliffs near Ribamar and the sequences Ha 4, Ha 5 and Ha 6.

S.B./T.S.: Surface that separates sandstones with cross-stratification from horizontal beds of sandy limestones.

T.S.T.: Thinning-up sandy limestones interbedded with sandy bioturbated marls. The bioturbation increases from the base to the top of this systems tract. The faunal assemblage is represented by oysters, trigoniids, serpulids and *Choffatella* (subtidal inner shelf).

M.F.S.: The thinnest and most bioturbated bed of sandy limestone.

H.S.T.: It includes three parts, in a shallowing-up evolution: i) a lower part with horizontal beds of sandy

vacuolar dolomites and fine sandstones (restricted lagoon) – from the base to the top, the thickness of beds increases and the bioturbation decreases; ii) a middle part of fine sandstones and black clays with lignite – the stratification is sub-horizontal and irregular (tidal flats); iii) an upper part of channel coarse sandstones, and black, lignite-rich marls, deposited in estuarine environments.

Above these siliciclastic sediments, the first horizontal beds of sandy limestones represent the T.S.T. of the following sequence (Ba 0).

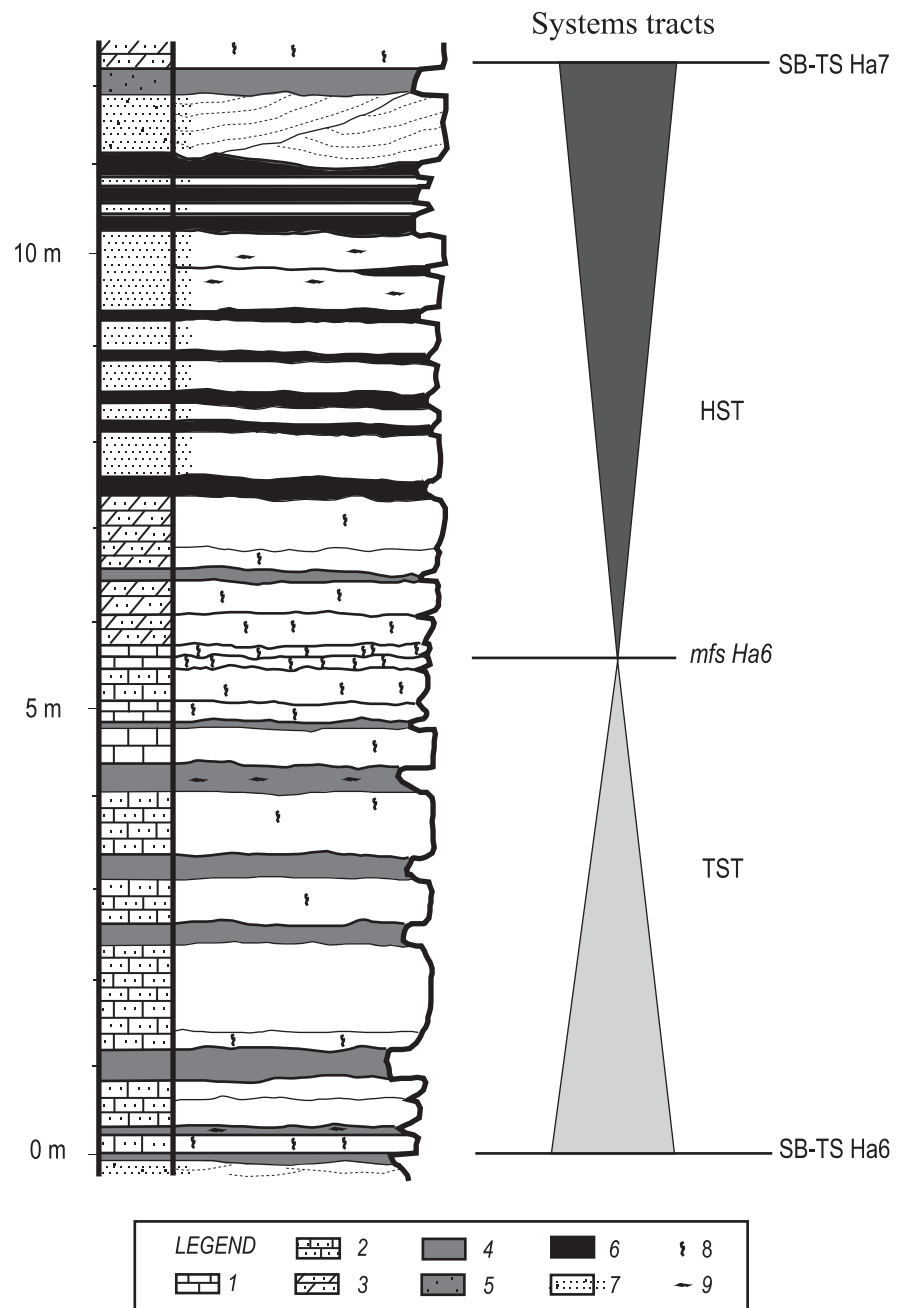


Fig. 15 - The Ha 6 sequence at Ribamar.

Lithology, stratonomy and sequential arrangement. 1: limestone; 2: sandy limestone; 3: sandy dolomite; 4: marl; 5: sandy marl; 6: black mudstone; 7: sandstone; 8: bioturbations; 9: lignite.

4. DESCRIPTION AND INTERPRETATION OF THE GALIOTA SYNCLINE

4.1. Local stratigraphic series

4.1.1. Setting and location

The Lower Cretaceous of this area is poorly studied, both in stratigraphic and sedimentological aspects, and the data and interpretations presented here are preliminary approaches. The area (Figs. 16 to 18) was

selected for this excursion owing to its excellent exposures, allowing an almost continuous observation of the series and its depositional architecture, and the rich and diverse sedimentary structures at several scales.

The few biostratigraphical data collected in this succession include exclusively continental macroflora (Teixeira, 1948) and palynoflora (Rey, 1972; Hasenboehler, 1981) and suggest an Early Cretaceous age. The most probable range for the studied assemblages is Barremian to early Albian (Dinis, 1999). The fact that no angiosperms pollen were found points to the earlier



Fig. 16 - Panoramic view to the north of the Galiota syncline northern flank.



Fig. 17 - Road map and location of stops of the Galiota syncline.

Cretaceous stages, but it is noteworthy that in samples with an abundant and diverse angiosperms mesoflora of nearby Lower Cretaceous outcrops it is common the apparent absence of angiosperms palynomorphs (Friis *et al.*, 1999).

The presented profile (Fig. 19) corresponds to the southward tilted succession along the beach, starting just south of the Óbidos lagoon (the Gronho cliff). The series plunge decreases, becoming horizontal along the beach (N45E, 5NW) about 3 km south, near the Barroco da Adega location. The total thickness is close to 210 m, contrasting with the estimated 75-80 m in the southern flank of the syncline. This difference is probably related with withdrawal to the adjacent Caldas da Rainha diapir of the underlying upper Triassic to lower Jurassic evaporitic marls, either creating an important differential syndepositional subsidence or a pre-depositional relief.

4.1.2. Field description

The succession is composed of siliciclastic facies with sand fraction dominated by angular quartz, slightly arkosic; largest clasts are subrounded vein-quartz and metaquartzite.

The total thickness can be subdivided in several informal lithostratigraphic units based on its lithology, maximum particle size, facies proportion, depositional architecture and peculiar content, such as palaeosols, cemented levels and coalified remains (up to metre wide tree trunks). The correlation of levels and some

interpretations presented here are based in photomosaics of laterally extensive outcrops.

Unit I - 0 to 45 (?) m. This basal unit has the highest average of maximum particle size (MPS), reaching 18 cm, and a maximum individual clast size of 25 cm (Fig. 20), displaying a clear fining upward trend. No floodplain bodies are preserved, but muddy intraclasts up to 50 cm exist. Among the vegetal macro-remains, some well preserved tissues are notable. The architecture is dominated by conglomeratic lags (lower bar segments), cross-bedded (channel) pebbly sandstones and multi-storey and laterally amalgamated low order channels (up to 40 x 3 m), sometimes draped or filled with mudstones. Palaeocurrent data has a low dispersion pattern around a W to WNW palaeoflow.

Unit II - 45 (?) to 83 m. A recent dune field covers the lower half of this stratigraphic extension, but after unpublished cartography (by J. Rey), previous description (Rey, 1972) and the erosional incision on the beach we deduce a composition dominated by relatively fine and friable deposits, probably sandstone with some lignitic mudstone levels. The exposed part is essentially trough cross-bedded sandstones with conglomeratic lags (MPS up to 8, max. 12 cm; Fig. 21) and isolated minor channel mudstone filling. Architectural details are few, due to limited exposure, but no major changes are obvious compared with the adjacent levels. Palaeocurrents also looks much the same, but with dispersion higher than Unit I.

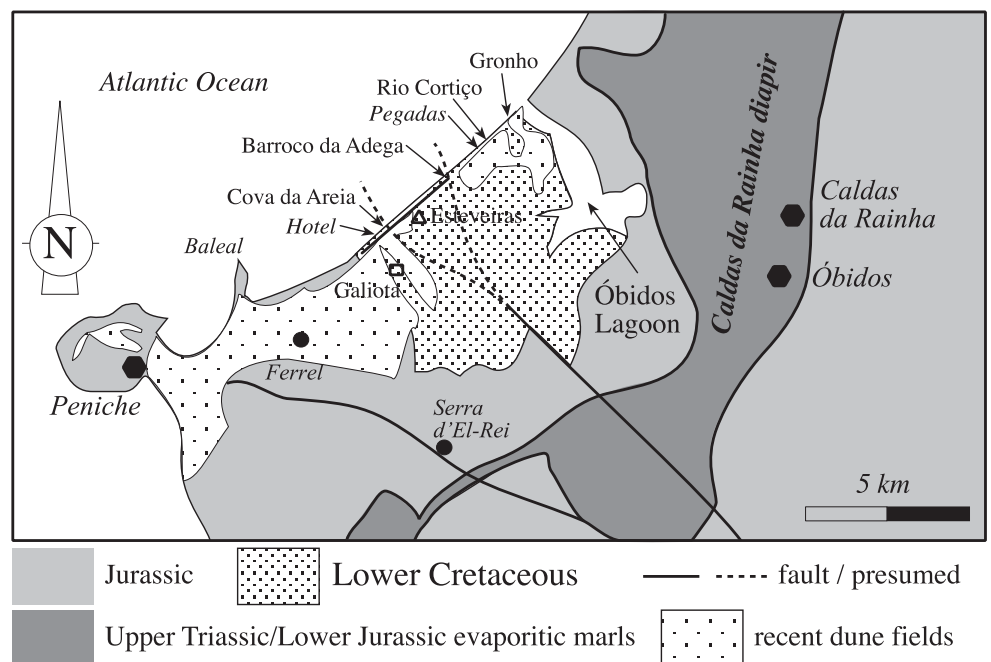


Fig. 18 - Simplified geological map, including the Galiota syncline and locations referred in the text.

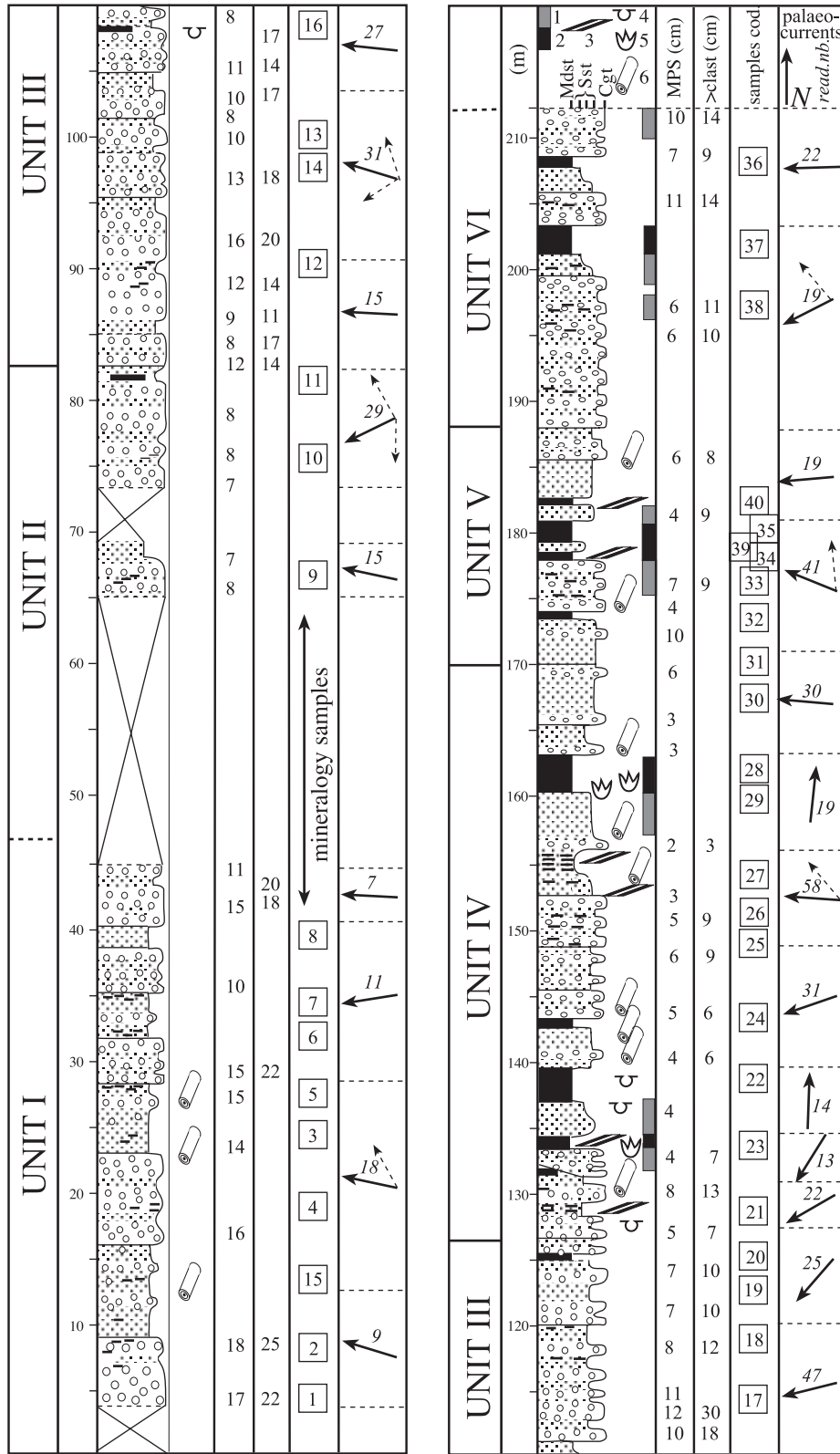


Fig. 19 - The Lower Cretaceous succession of the northern flank of the Galiota syncline. 1: Cemented level; 2: palaeosol; 3: inclined heterolithic stratification; 4: convolute level; 5: dinosaur footprints; 6: coalified drifted tree trunks.

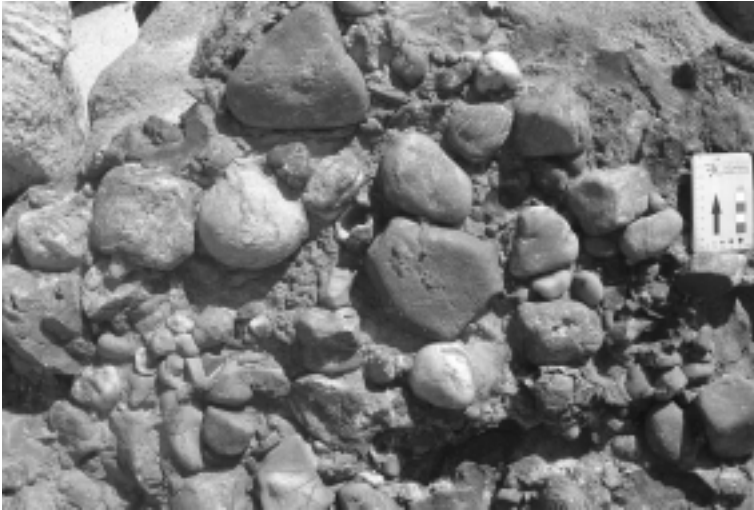


Fig. 20 - Strata parallel view of highest MPS (18 cm) of the Galiota syncline lower Cretaceous succession (Gm lithofacies, lower part of Unit I).

Unit III – 83 to 127 m. The lower limit is defined by a sudden increase in grain-size, and is organised in an increasing to decreasing-size cycle. The coarsest clasts reach 30 cm, and the MPS attains 16 cm. One of the peculiar aspects of this level is the abundance of black metaquartzites as well as other exotic pebbles (meta-graywacke, large orthoclase, pegmatite and reticulated metaquartzite). The facies association is dominated by an intimate alternation between conglomeratic lags or thin lenses (Gm, frequently displaying imbrications) and St/Gt facies. Cross-bedded whitish fine to medium sandstones (Stf) are common and mudstones, other than metre-sized channels and intraclasts, are quite rare. Lignite is rare, comparing with other units of the profile. The architectural surfaces have a clear flat trend, interrupted locally by lenses of fine lithofacies (Stf, Fm and Fh) thinner than 1 m. However, structures like

convolute lenses of Fm and very steep St and Stf sets points to syndepositional deformation. Most of the stratigraphic interval follows the previous westward drainage, but it appears that palaeocurrents shifts to a more SW trend in the upper 10 to 12 m.

Unit IV – 127 to 170 m. Initiated by a drop in grain-size, from conglomeratic to sandstone dominated, it is the unit of the series with the lower grain-size average. St and pebbly St are the majority of accumulated thickness, with important proportions of Fm, Gt, Stf, Gm and Sh. MPS reaches 8 cm at the base, but most of the unit has a MPS around 3/4 cm. The most striking aspect is the large content in lignite, and in particular the frequent occurrence of large tree logs (diameter up to 1 m; Fig. 22). Also remarkable are three cemented levels, quoted as “Rio Cortiço” (I and II) and “Pegadas”.



Fig. 21 - Natural rock arch in conglomeratic sandstones (upper part of Unit II).



Fig. 22 - Coalified drifted tree trunk infilled with sandstone (like a channel) in a complex of amalgamated channels; abundant lignite debris in the fine facies (middle part of Unit IV).

All these have associated metre scale mudstone sheets with palaeosols bioturbated by roots and infauna, showing characteristics of hydromorphic palaeosols, like high Fe oxides and hydroxides content, colour mottling and slickensides. Dinosaur footprints were recently found in the “Rio Cortiço II” and “Pegadas” levels.

In the “Pegadas” level the 128 tridactyl dinosaur footprints found are organized in 17 trackways within an area of around 80 m² of a thin level of rippled and horizontally laminated fine sandstone (Fig. 23). Most of and the best preserved footprints were produced by Iguanodontids, large and medium size theropods and some appear to be from ornithopods (Mateus & Antunes, 2003).

Units IV to VI are almost horizontal, and the bodies and surfaces can be followed in hundreds of metres to a few kilometres along the cliff, allowing a detailed study of its depositional architecture. Comparing to Unit III, clear changes occur and the architecture is characterised by metre scale (2 to 10 m) sheets of sandstones or conglomeratic sandstones separated by decimetre (up to 2 m) sheets or very large lenses (channel like) of mudstones and fine sandstones. Within the main bodies,

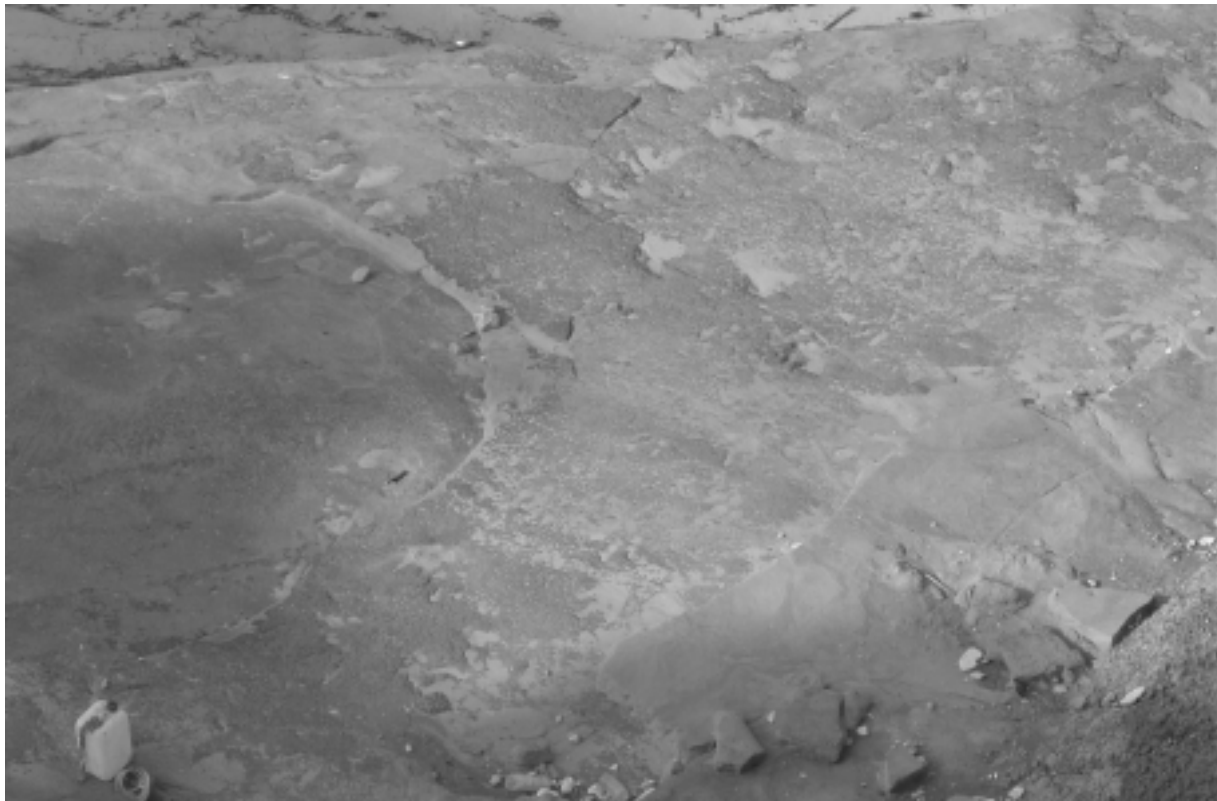


Fig. 23 - Tridactyl dinosaur tracks in the “Pegadas” cemented level (see helmet for scale, upper part of Unit IV).



Fig. 24 - Lateral accretion heterolithic body within a St dominated outcrop (upper half of Unit IV).

decametre-wide single-storey channels appear, sometimes draped or filled with mudstone.

In particular, Unit IV includes heterolithic sheet bodies interpreted as lateral accretion (sigmoidal mudstone/sandstone; Fig. 24), more frequent close to the cemented levels, and multi-storey metre scale channels with clusters of logs. Palaeocurrents are widespread but individual sandstone bodies (cosets) have moderate dispersion; most show drainage to W but others clearly to N and to SW.

Unit V – 170 to 188 m. The lower limit of this unit corresponds to an enlargement of grain-size, with a MPS up to 10 cm, measured in isolated conglomeratic lags. However, sandstones are still the main component and Fm bodies are important (around 20% of the total thickness); it includes also Sm and Stf bodies. Comparing to Unit IV conglomeratic lags are frequent elements and channel deposits constitutes thicker lithosomes. Cementation is more intense in two levels closely spaced (named “Fincha” and “Arco”) affecting medium sandstones. As in Unit IV, these levels are associated with sheets or very wide channels of mudstones with pedogenetic features. Coalified logs are scattered but reach up to 1 m in diameter. The drainage pattern is similar to the underlying unit, trending to the West.

Unit VI – 188 to 212 m. Contrasting with Unit V it has relatively coarse-grained conglomerates (MPS up

to 11 cm and single clasts up to 14 cm) and more rare lignite debris (including logs). Cementation is more or less pervasive in this stratigraphic interval. Many levels of coarse sandstones and conglomerates of this unit are cemented, in particular near the top of the studied succession (the “Covão” level), but one is similar to the above described (the “Adega” level). Palaeocurrents point to WSW or West.

4.1.3. Mineralogical data

Clay assemblages (Fig. 25) are dominated by kaolinite over illite, with frequent occurrence of swelling minerals (mostly illite/smectite and chlorite/smectite mixed-layer clays), all considered as essentially inherited (detrital) material. Despite the facies control on clay mineralogy (as shown by the changes in closely located samples), illite proportion rise vertically, opposing to the kaolinite trend. Smectite and expanding mixed-layer clays have low proportion and an almost random distribution.

In detail, units I, II and III are characterised by a dominance of kaolinite over illite, Unit IV is transitional and in units V and VI contents of illite and kaolinite are similar. Also noteworthy is the increase of kaolinite upwards in Unit I, opposing to illite (if the facies effect is considered). The values of swelling minerals are relatively high in some samples of Unit I (in particular

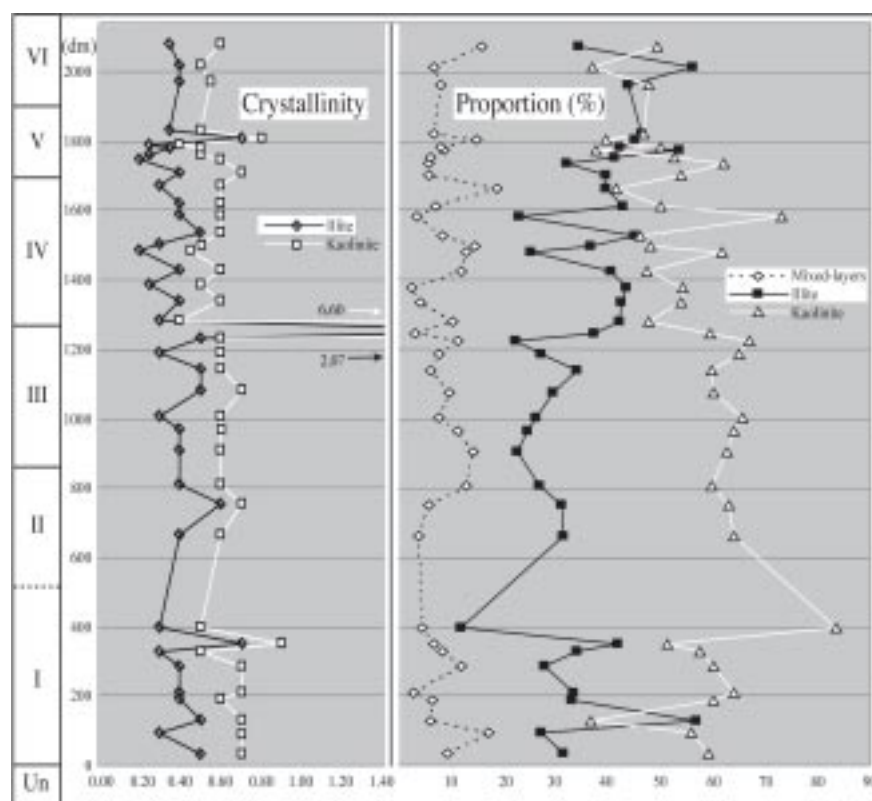


Fig. 25 - Clay minerals (proportion and crystallinity index) of the Lower Cretaceous succession of the Galiota syncline northern flank.
Un - informal lithostratigraphic units.

illite/smectite mixed-layer), Unit III (both illite/smectite and chlorite/smectite mixed-layer) and Unit IV (almost exclusively chlorite/smectite mixed-layer, up to 19% of clays).

Within a slight increase in crystallinity of illite and kaolinite to the top (Fig. 25), two intervals of more crystalline illite and kaolinite (lower index) were identified in the lower half of each unit IV (samples 22 and 25) and V (samples 32, 33 and 39) and interpreted as revealing more intense diagenesis and/or lower energetic systems. Exotic low crystallinity of both minerals in samples 20 and 35 is presumed as resulting from meteoric degradation by long-lasting residence near the floodplain surface.

Phyllosilicates are abundant at the base but decreasing in favour of quartz (Fig. 26). A medial section includes the upper part of Unit III and Unit IV, where quartz declines in disadvantage of the phyllosilicates content. The transition from units III to IV, as well as the unit V (and VI in a lower degree) are marked by high contents of fine quartz. Feldspars are ubiquitous with a general tendency to an upward grow; plagioclase dominates in most of the series, but in units V and VI K-feldspar becomes more frequent.

Siderite is almost ubiquitous, usually in minor amounts (1-3% of the fine fraction), but achieving 3-5% in samples of the top of Unit III and base of Unit IV. It attains maximum content in the uppermost sample. Opal is present in the lower half of Unit I, the top of

Unit II and many samples of Unit IV. Considering field evidences we believe that opal corresponds to incipient cementation, and the fine quartz increase is either detrital, reflecting an overall fining-upward trend, or neoformed and related to the cemented levels of units IV to VI. The total content of siderite, anhydrite, dolomite, halite and pyrite, probably of pedogenetic and/or early diagenetic origin seems to increase upwards (Fig. 26), reaching values up to 16% in units III to V and a maximum of 19% in Unit VI.

4.2. Interpretation

4.2.1. Depositional systems

The above exposed data points to a fluvial braidplain draining the inland Hercynian basement towards the Atlantic coast. The clays and presumed early diagenetic minerals assemblages are interpreted as reflecting a wet and warm climate during the source area weathering and the deposition of the series. A system with a network of channels with several orders is deduced, with transition along the stratigraphic levels from more proximal to more distal settings. The more flat and wide surfaces include minor channel scours or fills (Fig. 27) and can be assigned to regional changes in active channel belts, either by wandering or avulsion.

An important part of Units I and III conglomerates corresponds to longitudinal gravel bars and hollow-fill

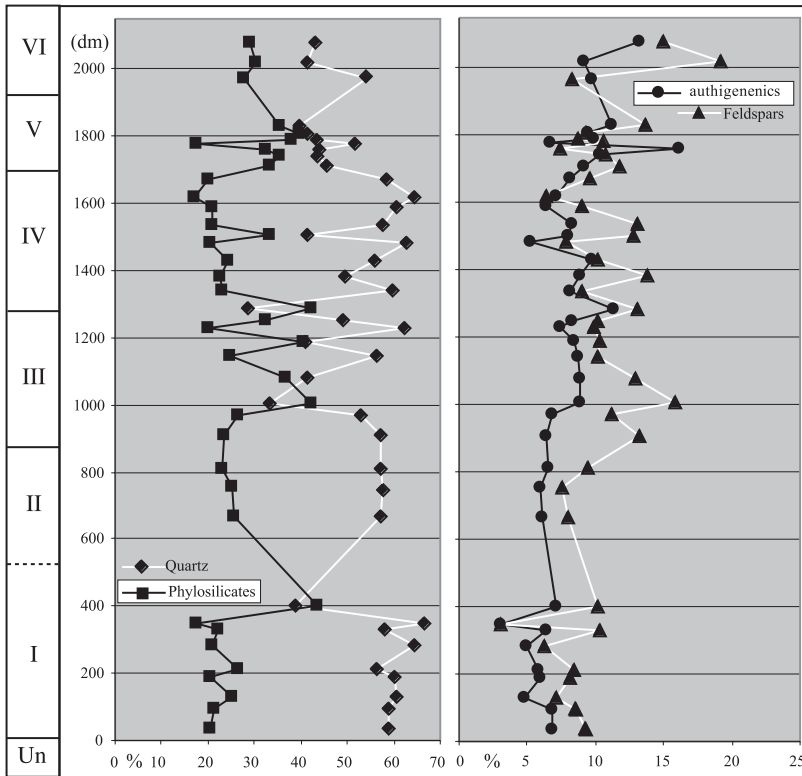


Fig. 26 - Fine fraction minerals of the Lower Cretaceous succession of the Galiota syncline northern flank. Authigenics is the sum of siderite, anhydrite, dolomite, halite and pyrite contents. Un - informal lithostratigraphic units.

deposits, but some are organised in relatively isolated gravel lags of sandy bars base or top. Sandstones are laterally amalgamated channel facies (dune fields and gravely bars). The overall architecture, including the frequent meter-scale channel surfaces and absence of floodplain deposits, points to a high ratio between sediment flux and accommodation.

The large tree trunks are drifted, as revealed by its horizontal position oriented either parallel or perpendicular to the enclosing palaeocurrents, and some exception to this disposition is probably related to fast burial during high energy floods. It is also remarkable the fact that many logs are filled with material similar to the enclosing deposits.

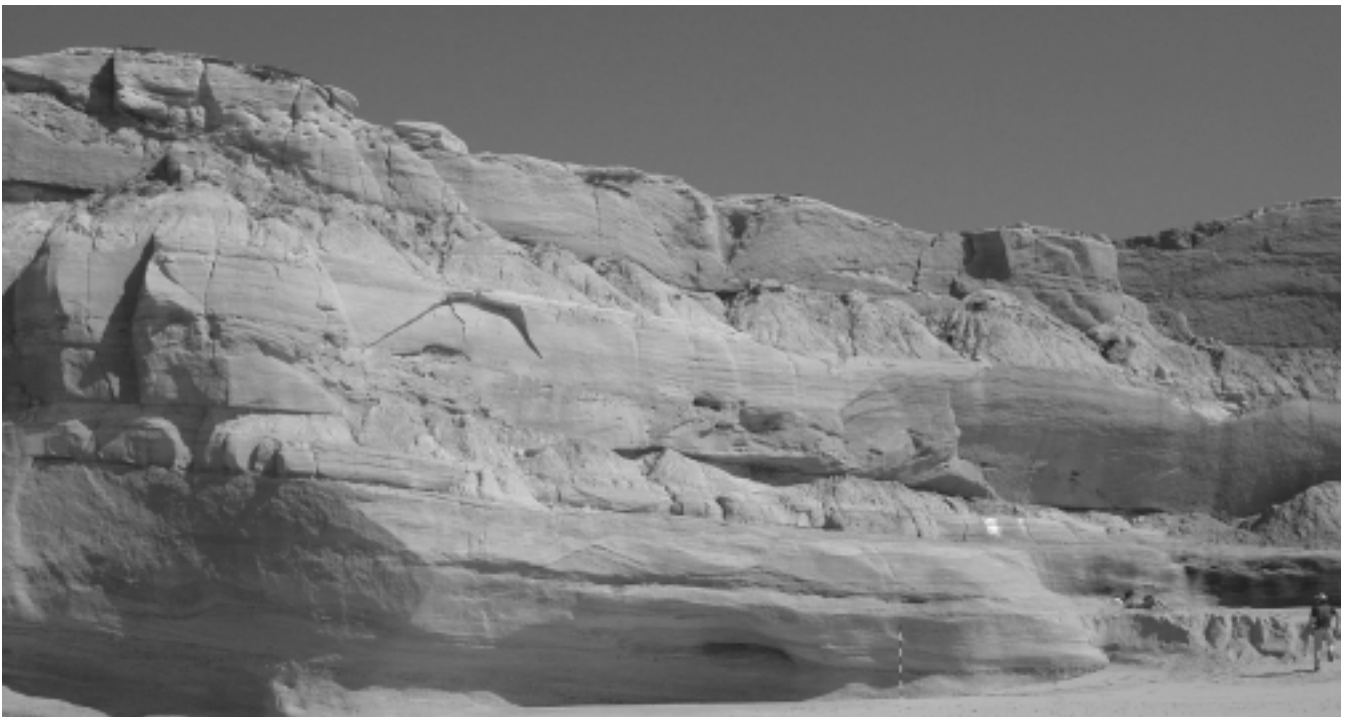


Fig. 27 - Boundaries followed in outcrop for hundreds of metres to kilometres include many scour and fill of minor channels and are interpreted as channel belt wandering or avulsion. The base of the outcrop is the Rio Cortiço I cemented level (lower part of Unit IV).

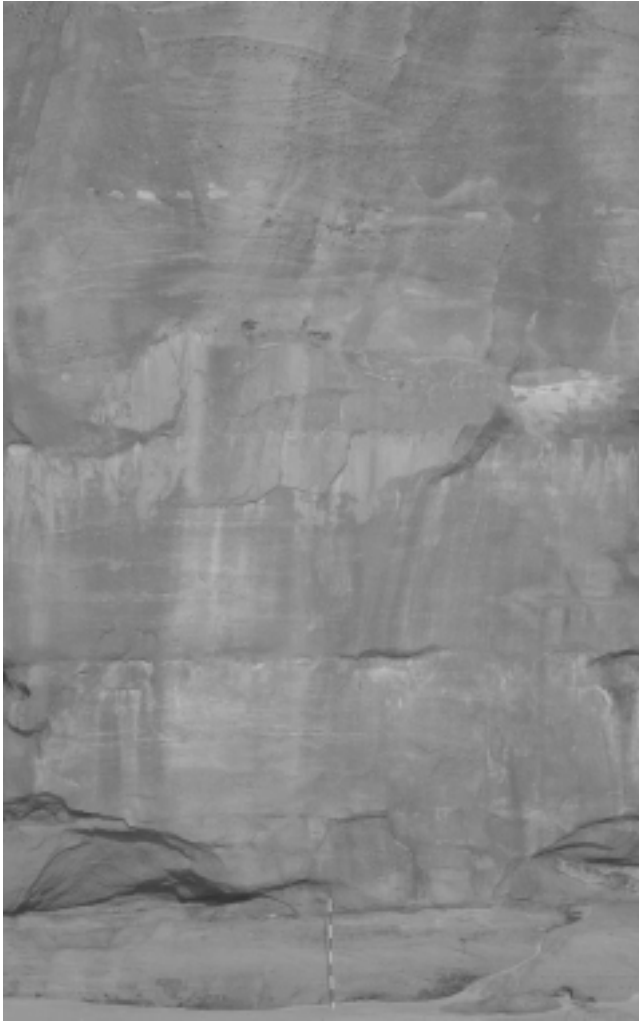


Fig. 28 – Cemented levels of Unit V. The one at the base of the cliff (Arco), affecting medium sandstones, is probably early diagenetic, contrasting with the upper cemented coarse sandstones to conglomerates at the base of Unit VI. Notice also the coarsening upward evolution.

The medium sandstone cemented levels have a laterally extensive sheet-like geometry and are overlaid by floodplain deposits. Coupled with the clay and diagenetic minerals, and the dinosaur footprints, these levels are interpreted as the result of relatively long-lasting conditions of phreatic level close to the topographic surface, or even very shallow water in lakes or ponds. In a broader scale, an extensive plain with low depositional slope can be envisaged, favouring the concentration of solutions. By propitiating also the wandering of channel belts, this can explain the great dispersion of palaeocurrents of these levels when compared to the prevailing tendency.

Unit IV (and II ?) is peculiar, owing to its floodplain extension and/or preservation and levels with clear lateral accretion (inclined heterolithic stratification) in large bars within sinuous channels. The combination of abundant coalified debris, siderite and pyrite/marcassite

is compatible with a high water table in poorly-drained floodplains, and low-oxidising conditions during deposition and/or in the early diagenesis. Comparing with adjacent units, this stratigraphic range evidences a relatively high accommodation to sediment flux ratio, corresponding to lower transport capability within the system, lower input from the source area or higher accommodation (by subsidence increase or base-level rise).

Near the top of Unit III the conglomerates laminae becomes more and more isolated, situation clearer in Unit IV and also applicable to Unit V. These are interpreted as isolated lags of sandy bars base or top, or as scour-and-fill pavements. This tendency is inverted from Unit V to VI. Most of the cementation in Unit VI appears to be of a different kind, mostly affecting coarse sandstone to conglomerates (Fig. 28). Considering the present knowledge, the timing of such cementation is not clear but it can be a late(r) diagenesis.

4.2.2. Sequential approach

Modifications in the proximal-distal character and in the slope of the depositional system can be deduced from vertical changes in grain-size, lithofacies association, depositional architecture, early diagenesis and composition. The application of sequence stratigraphy approaches implies the recognition of high-order unconformities potentially corresponding to sequence boundaries and the evaluation of accommodation changes (e.g. Aitken & Flint, 1995; Boyd *et al.*, 2000; Miall & Arush, 2001). However, the evaluation of the origin and dimension of an unconformity is very difficult in this sedimentological context, not only due to the large number of internal unconformities but also to the similar facies and architecture resulting from progradational and low accommodation phases. In most cases the more obvious unconformities are channel belt shifts related to autogenic mechanisms, but some can be assigned to allogenic changes and form cryptic depositional sequences (Miall & Arush, 2001). We present a possible response to the challenge of applying such concepts to the studied outcrops.

The studied succession records a general positive sequence following a major event, possibly of tectonic nature considering the stratigraphic and geodynamic setting. The kaolinite initial abundance and subsequent decrease, coupled with inverse tendency regarding feldspars, is attributed to the dismantling of palaeo-weathered soils in the hinterland (mainly in granitoids and metapelites) formed under a moist and warm climate.

Three cyclic sequences can be identified (Fig. 29), each starting with a lower coarsest unit, a broad fining-upward drift and including again coarse levels at the top. Such organisation is the most usual in continental clastic deposits and the palaeogeographic changes can be envisaged whatever were the controlling mechanisms (Dinis, 1999). The sequence boundaries are located in major augmentations in grain size but also coincide with changes in other parameters such as depositional architecture and composition. Each sequence has the coarsest levels some meters above the limit, and the lower part is considered as a progradational phase of adaptation of the system to the event that produced the renewed energy and sediment transport capability. The subsequent fining-upward phase is related to a general decline in slope and, hence, with a probable retrogradational pattern – it can correspond to changes in allogenic forcing mechanisms such as sea-level, subsidence or climate, or simply to the combined result of erosion and basin deposition. We attribute the slight grow in grain-size at the top of the sequences to a fluvial profile extension when the landscape converges to a steady state or a base level still-stand.

Sequence Ga1: units I and II. The lowermost part is covered, but it seems that maximum grain-size is reached several meters above the presumed base of the Cretaceous deposits. Accommodation is very low along the sequence, as proved by the wide dominance of preserved lower geomorphic elements of the depositional system and amalgamated (minor) channels. Palaeocurrents have a maximal dispersion in the last 10 m when a rotation to SW emerges. Opal occurring in samples 1 to 3 and 9 to 11 also points to a topographic stability at the base and at the top.

Sequence Ga2: units III to V. Maximum grain-size is located near 10 m above the sequence boundary. Most of the thickness shows a gradual growth in intensity and proximity of symptoms of lower energy and larger accommodation: floodplain deposits, soft-sediment deformation, inclined heterolithic stratification, palaeosols, coalified debris, palaeocurrent dispersion, early cemented levels and authigenic minerals. This trend reaches its maximum in the Pegadas level, consequently considered as a maximum distal position/retrogradation.

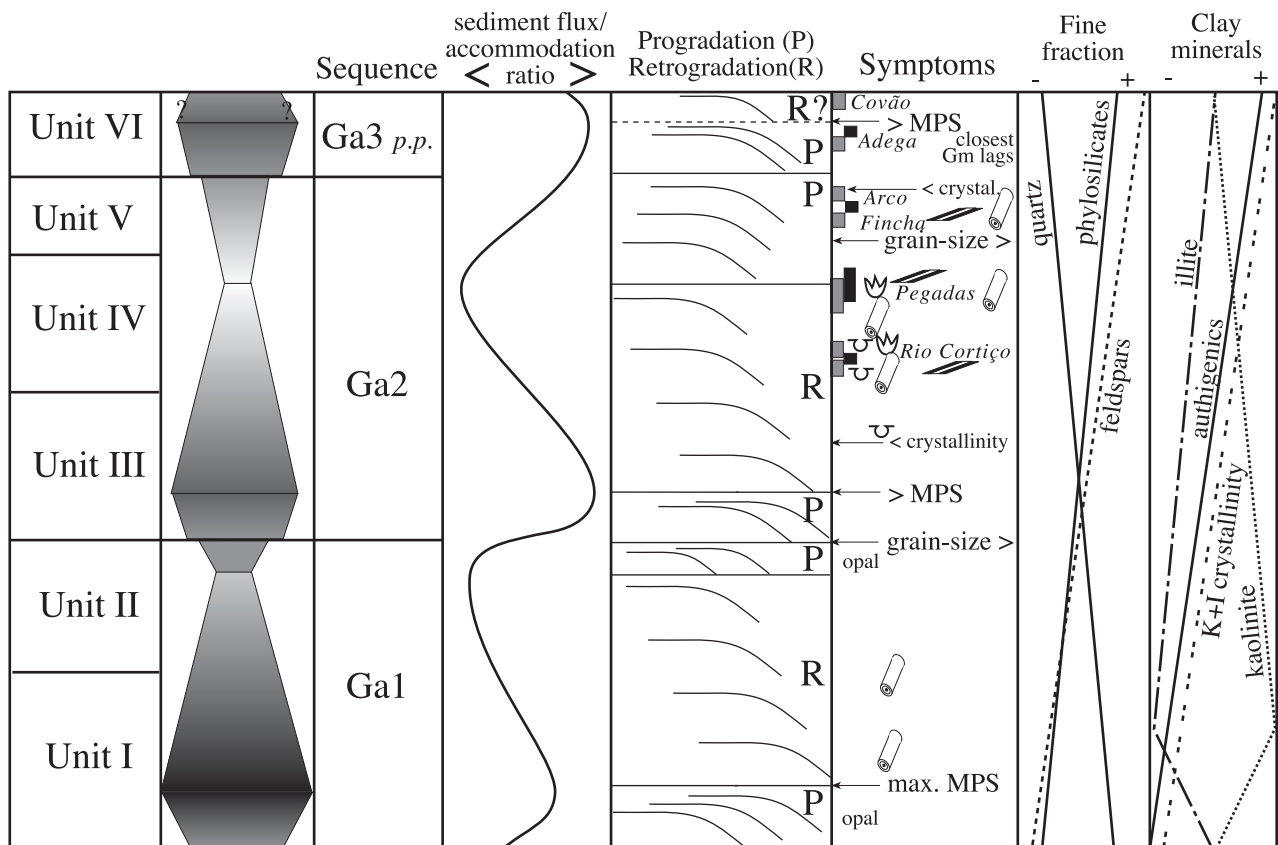


Fig. 29 - Lower Cretaceous succession of the Galiota syncline northern flank: Sequences, deduced depositional architecture, selected diagnostic criteria and synthetic evolution of mineralogy. See legend of Fig. 19.

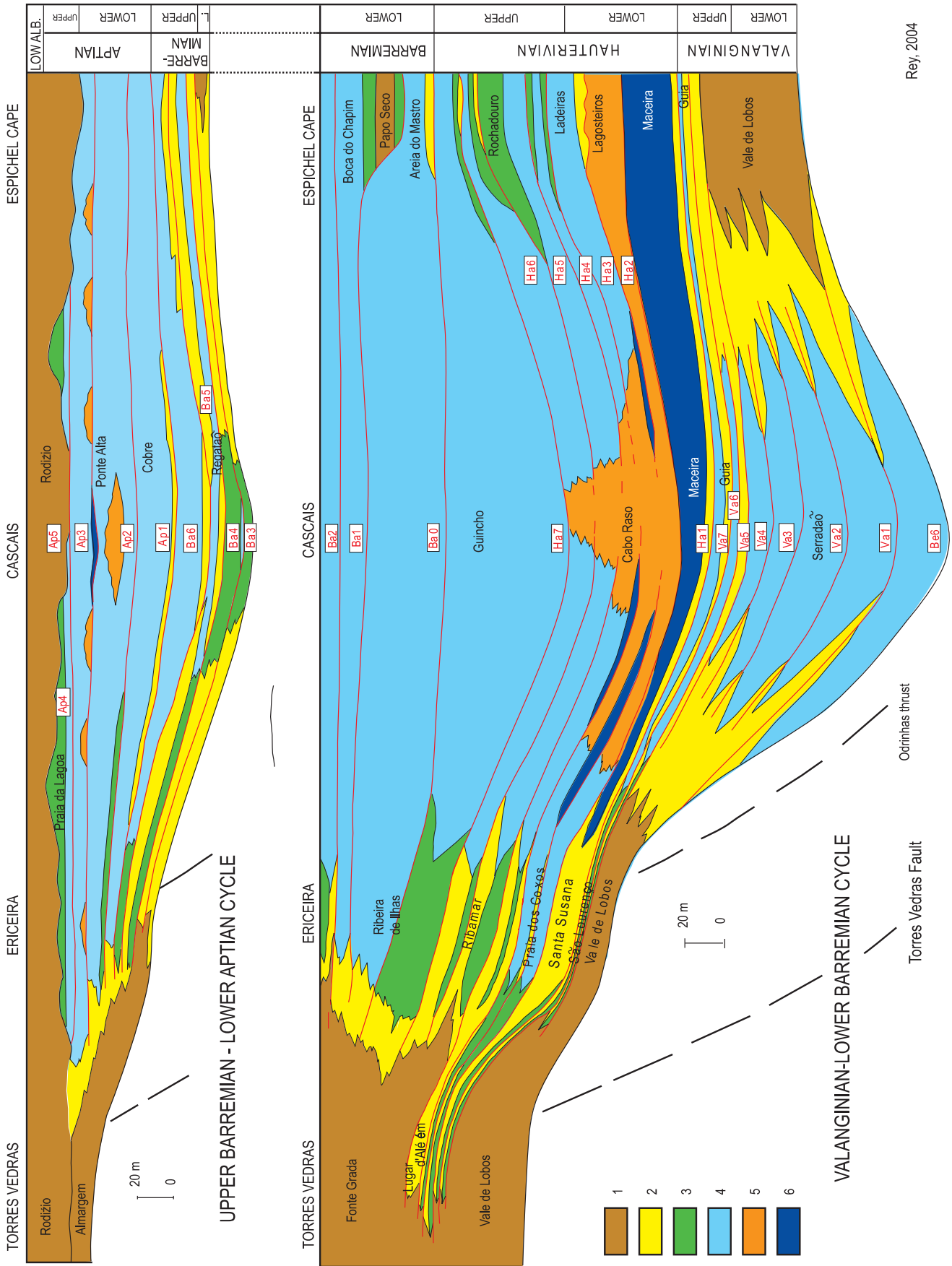
Sequence Ga3 p.p.: unit VI. After a significant increase of grain-size, the referred signs of lower energy and larger accommodation become widely spaced. On the other hand, the space between conglomeratic laminae has a minimum in the base of VI and is considered as directly related with the total energy and/or inversely related to the accommodation in the system. So, this partially preserved sequence has probably its maximum lower progradation in unit VI (eventually including the beginning of the subsequent retrogradation).

Acknowledgements

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Rey, 2004

Plate 1 - The sedimentary infilling of the southern Lusitanian Basin between the Valanginian and the Albian (lower) stages

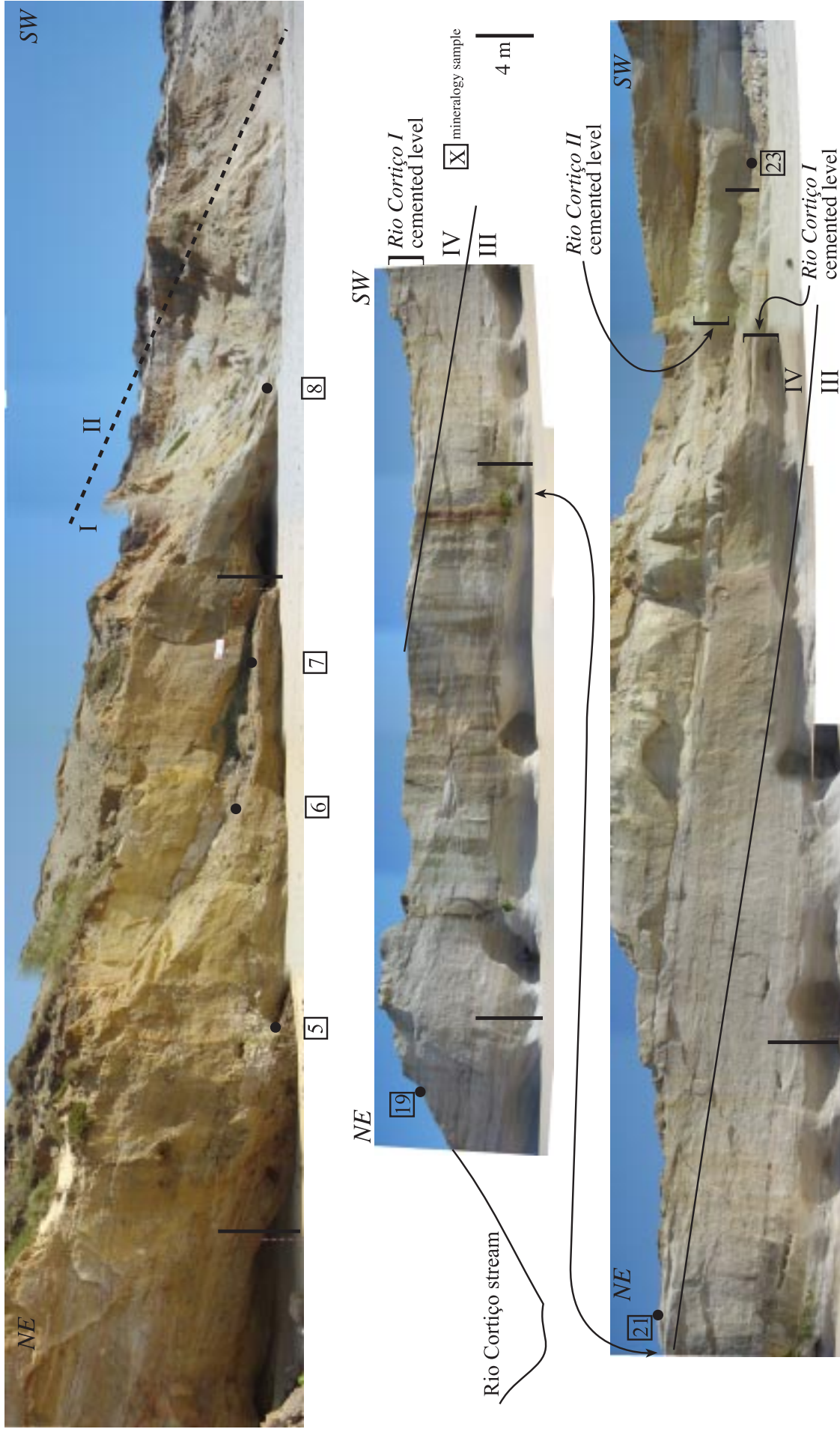


Plate II - Photomosaics of the Galiota Syncline northern flank, indicating units boundaries, cemented levels referred in text and location of samples for mineralogical data.

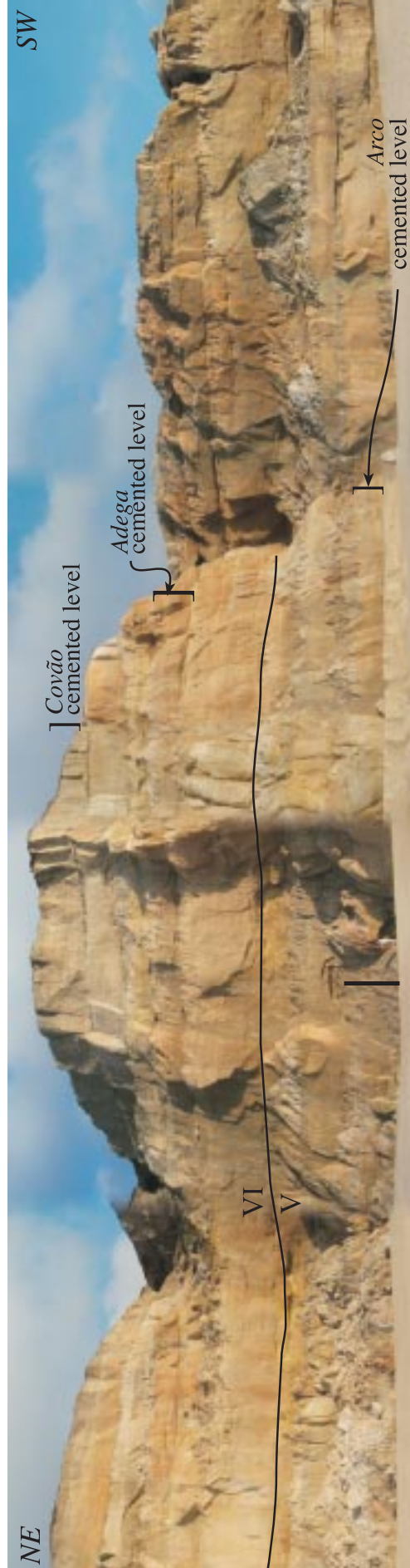
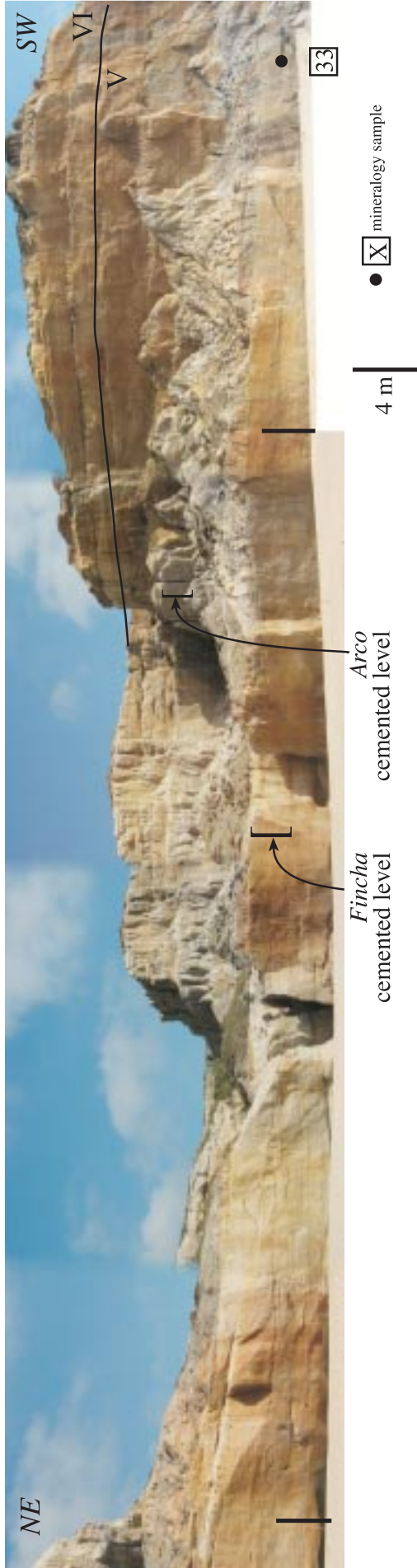


Plate III - Photomosaics of the Galiota Syncline northern flank, indicating units boundaries, cemented levels referred in text and location of samples for mineralogical data.

Post-Meeting Field Trip P5

**The Cenomanian-Turonian central
West Portuguese carbonate platform**

PEDRO CALLAPEZ

The Cenomanian-Turonian central West Portuguese carbonate platform

PEDRO CALLAPEZ¹

1. INTRODUCTION

The West Portuguese or Lusitanian carbonate platform is a typical example of the European and North-African shallow-water rimmed shelves that bordered the northern branch of the Tethyan Realm, during the Cenomanian-Turonian sea-level rise. With a depositional setting related with the Cretaceous evolution of the Western Iberian continental margin, between North and Central Atlantic and North Africa, this platform has long been a subject of special interest concerning the development of rudist reefs, and biotic changes across the Tethyan and Temperate domains.

Researches on the Portuguese Cenomanian-Turonian carbonate units were initiated by Daniel Sharpe during the decades of 1830-40, and continued by Paul Choffat and collaborators, from 1880 until the beginning of XX century. Most studies have been focussed on the successions exposed across the onshore regions of Estremadura and Beira Litoral, from Lisbon to Figueira da Foz and Nazaré-Leiria-Coimbra (Fig. 1). Within these areas, the Cenomanian-Turonian carbonates overlie almost 5000 meters of Upper Triassic, Jurassic and Lower Cretaceous series, recording the sedimentary evolution and main depositional episodes of the Lusitanian Basin.

Together with other contemporaneous basins of Western Europe, the Lusitanian basin was formed as a result of the Mesozoic intracontinental rifting that forced

the break-up of North America and Europe after the Late Triassic (Hiscott *et al.*, 1990). The basinal area is elongated after a meridian direction, and bounded westwards by basement horsts like those of Berlengas and Farilhões Islands (Ribeiro *et al.*, 1979). During the Mesozoic, this alignment acted as a tectonic ridge that isolated in part the West Portuguese Basin from the adjacent Atlantic active margin (Cunha & Reis, 1995).

The stratigraphy of the Lusitanian Basin reveals a complex geotectonic history, controlled by late Hercynian faults reactivated during the extensional events, and by halokinetic structures associated with Early Jurassic evaporites. Three main rifting phases have been recognized (Wilson, 1979, 1988; Wilson *et al.*, 1989; Pinheiro *et al.*, 1996), which were followed by intervals of post-rift thermal subsidence, especially during the Lower and Middle Jurassic, and after the Aptian. The latest of these episodes culminated with the emplacement and expansion of the Cenomanian-Turonian West Portuguese carbonate platform, overlying the earlier rift and post-rift sequences, before the compression and inversion of the basinal area during Late Cretaceous and Cenozoic times.

2. GENERAL STRATIGRAPHY OF THE CENOMANIAN-TURONIAN

The Cenomanian-Turonian carbonate succession of West Central Portugal is part of a megasequence bounded by unconformities, with a record ranging from Late

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Aptian to Early Campanian (Wilson, 1988; Cunha, 1992; Pinheiro *et al.*, 1996). This megasequence is formed by the succession of: (1) alluvial coarse siliciclastic sediments carried out from the Hercynian Massif of Iberia; (2) platform carbonates; (3) marine to alluvial micaceous sandstones, especially preserved on the NE part of the basin; and (4) alluvial sandstones and conglomerates. Its lower boundary is a main unconformity, recorded as discordance over the Triassic and Jurassic series, and the Hercynian basement. The upper boundary consists of a thick silcrete, indicating a prolonged time of non-sedimentation and sub-aerial weathering (Cunha & Reis, 1995).

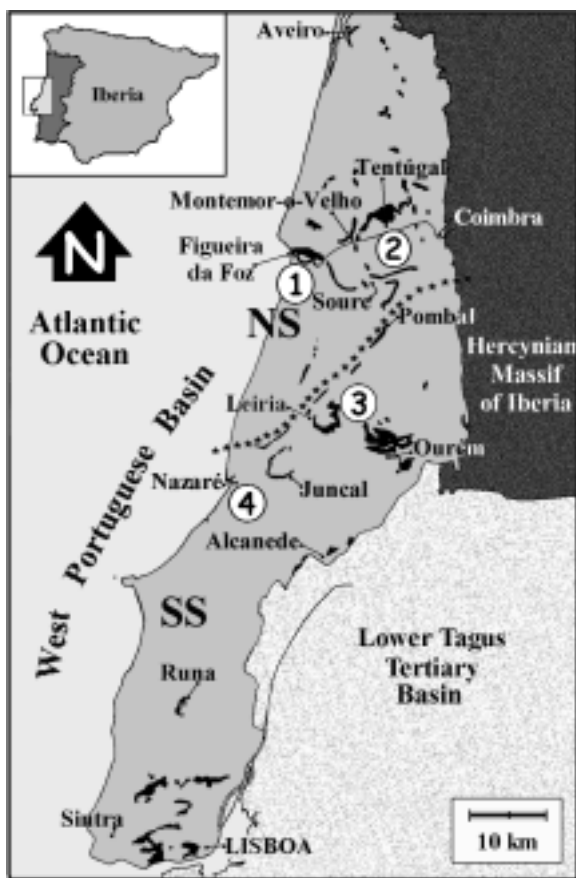


Fig. 1 - The West Portuguese or Lusitanian Basin, and the Cenomanian-Turonian onshore exposures and localities mentioned in the text. 1-2-3-4 - Location of field-trip stops; NS - Northern Sector; SS - Southern Sector; *** boundary between both sectors (Cós-Leiria-Pombal fault and associated diapiric structures).

2.1. The Aptian-Cenomanian siliciclastics and the “Bellasian”

The lower part of the megasequence is a succession of continental conglomerates and coarse sandstones with interbedded red mudstones, currently designated as Figueira da Foz Formation (Dinis, 1999). The source areas of these siliciclastic sediments were the Hercynian Massif of Iberia and, locally, the Berlengas and Farilhões

horsts (Rey, 1972, 1979). Biostratigraphic data suggest that the lower part of this unit is of Late Aptian to lowermost Albian age. The upper limit is diachronous, depending on the progression of the transgression northwards: Middle Albian near the region of Lisbon-Sintra, and Upper Cenomanian near Coimbra and Aveiro.

The siliciclastics of the *Figueira da Foz Formation* are overlain by a thick succession of marine carbonates, related to the Albian-Turonian sea-level rise events and usually designated as *Carbonate Formation* (Soares, 1966, 1980) (Figs. 2-3). Across the region of Lisbon-Sintra the lower members of this unit have been designated as “Bellasian”, a local chronostratigraphic unit created by Choffat (1885, 1900) to group the succession lying below the first beds with Upper Cenomanian ammonites (*Beds with Neolobites vibraeyanus*). This succession nearly 400 m thick is subdivided in four units of limestones and marls, ranging from the Lower Albian to the Middle Cenomanian (Rey, 1979; Berthou, 1973): 1. *Beds with Knemiceras uhligi*; 2. *Beds with Polyconites subvernuili*; 3. *Beds with Exogyra pseudaficana*; 4. *Beds with Pterocera incerta*. The two oldest units were redefined by Rey (1992) as the *Galé Formation*, and yielded faunas assigned to the biozones of *Mesorbitolina minuta*, *Neorbitolinopsis conulus*, *Orbitulina concave*, and *Orbitulina duranddelgai* (Lower Albian-Vraconian). The Early Cenomanian is registered as a sequence of marly limestones with biostromes of *Ilymatogyra pseudaficana* and rich microfaunas of orbitolinids and alveolinids. This unit is exposed across southern Estremadura, and near Nazaré (Callapez, 1998).

Northwards Lisbon, the thickness of the “Bellasian” series decreases, and the carbonate facies change laterally to alluvial siliciclastics. The Middle Cenomanian *Beds with Pterocera incerta* is the single “Bellasian” unit that persists over a large area, reaching the central and northern sectors of the basin. This unit is characterised by mixed facies of sandy and marly limestones and sandy marls, with abundant oysters (biostromes with *Gyrostrea ouremensis* Choffat, 1886) and occasional ammonites (*Turrilites costatus*), suggesting a shallow water sedimentation and a substantial siliciclastic influx. These beds are organized as a deepening-upwards mesosequence with a regressive tendency expressed at the top (Soares, 1966, 1980).

2.2. The Upper Cenomanian

The basal Upper Cenomanian stratigraphic succession starts with a transgressive surface, observed across the

whole sectors of the onshore. The carbonate content of facies increases and ammonites become common, together with a diverse and abundant benthic fauna. At this time the area occupied by the platform was considerably enlarged eastwards, not far from the emerged relieves of the Hesperian Massif.

Three depositional sequences can be recognised:

Sequence C/D - The basal Upper Cenomanian is recorded as a set of nodular and marly limestones, known from the region of Lisbon northwards, and associated with the maximal flooding and enlargement of the sedimentation area. The stratigraphic record of this sequence matches with the classic units of Choffat (1900) designated as *Beds with Neolobites vibrayeanus* and *Level with Anorthopygus michelini*. The association of *Neolobites* and *Calycoceras* indicates the standard European *guerangeri* ammonite zone. The same strata yielded a rich fauna of alveolinids (*Praealveolina simplex*, *P. tenuis*, *P. cretacea*, *Ovalveolina ovum*, *Cisalveolina fraasi*), first described by Berthou (1973) and Lauverjat (1982).

Sequence E/I and Sequence J - The middle and upper parts of the Upper Cenomanian succession record an effective differentiation of facies, with individualization of marly limestone domains with ammonites on

the northern sector of the platform. The correlative facies of the central sectors of Estremadura are mainly calcarenitic, with interbedded rudist and coral buildups.

The key area to study the Upper Cenomanian ammonite facies is situated near Salmanha, Figueira da Foz, 40 km West of Coimbra. The carbonate succession forms a set of 14 units, named with the capital letters “B” to “O” (Choffat, 1897). Many of these units are bounded by unconformities and have a large lateral extension, despite the significant lateral changes of facies occurring within the inner sectors of the carbonate platform. According to the biostratigraphic data (Berthou, 1984a; 1984b; Callapez, 1998), only the succession “C” to “J” is of Upper Cenomanian age.

In this area the sequence E/I starts with an abrupt shift from coarse bioclastic to marly facies, followed by important changes on the composition of fossil faunas, including the first occurrence of vascoceratids, and rudists (*Caprinula*, *Sauvagesia*). The faunas of *Vascoceras* first appear within unit “E”, and become widespread across the northern sector of the platform. Two associations have been recognised, indicating the ammonite zones of *septemseriatum* and *pseudonodosoides*, correlative to the standard zones of *geslinianum* and *juddii*.

Southward of Nazaré, Juncal and Leiria, the sequences E/I and J consist mainly of coarse bioclastic carbonates with rudist and coral bioherms, and interbedded rudist

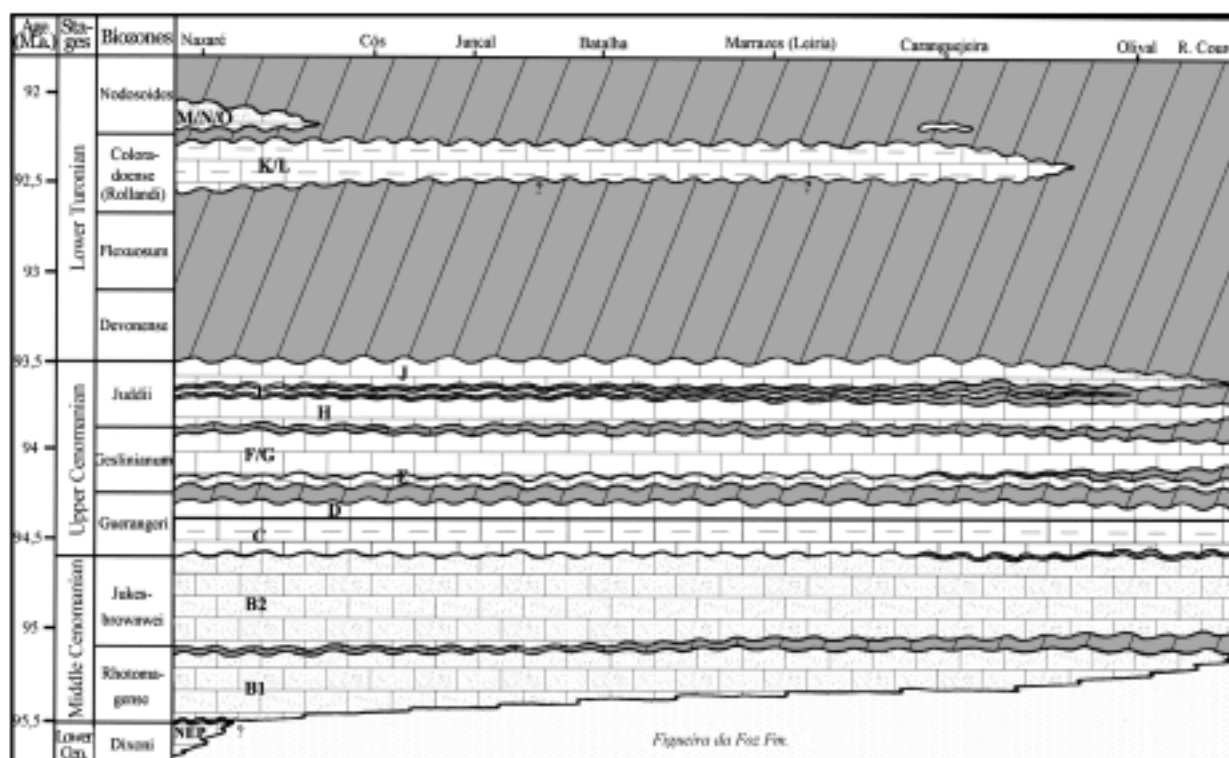


Fig. 2 - General stratigraphy of the Cenomanian Turonian from the onshore sectors of the West Portuguese Margin situated eastwards of Nazaré.

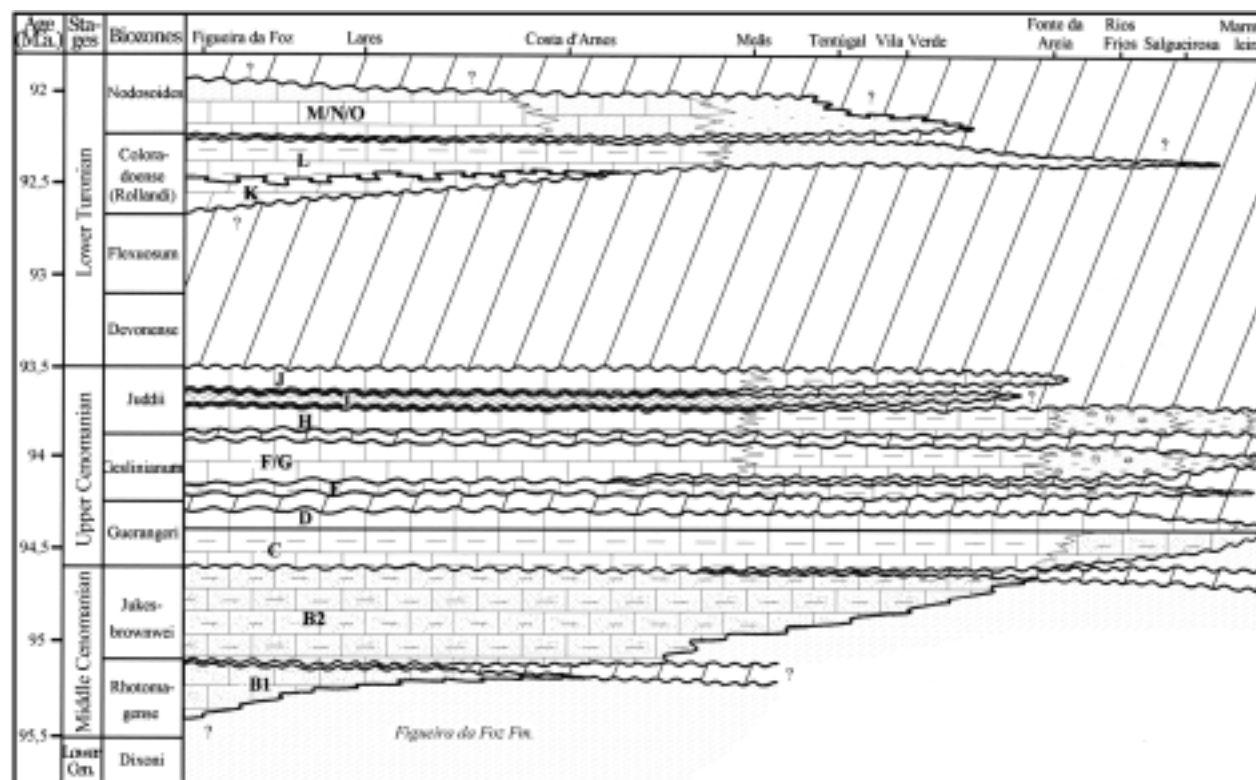


Fig. 3 - General stratigraphy of the Cenomanian Turonian from the onshore sectors of the West Portuguese Margin situated eastwards of Figueira da Foz (Baixo Mondego region).

biostratigraphy. These domains developed over the southern block of the “Cós-Leiria-Pombal fault”, a late-Hercynian NE-SW tectonic alignment associated to diapiric activity since Jurassic times, and reactivated during the Cenomanian-Turonian (Berthou, 1973; Lauverjat, 1982). The tectonic activity of this and other regional structures were fundamental factors on the differentiation of facies across the northern and southern domains of the Upper Cenomanian platform. The stratigraphic succession of Estremadura shows: (1) biocalcarens with bioherms of *Caprinula* and *Sauvagesia*, (2) limestones and marls with *Durania* and *Radiolites*, and (3) limestones and marls with *Durania*, nerineids and *Trochactaeon*. The caprinulids, one of the most typical groups of the Upper Cenomanian faunas from the Tethyan realm, are abundant in the first unit. Together with the radiolitid *Sauvagesia sharpei*, they are the framework builders of a large number of bioherms concentrated on the regions of Lisbon and Runa. Northward of these areas the density of buildups decrease gradually and the facies become essentially biotrititic (Callapez, 1998).

Across the region of Leiria, rudist facies are replaced by a reef complex with coalescent bioherms of massive platy and dendroid corals. This complex is connected to an extensive back-reef lagoon with hemiasterids and exogyrinids, developed in the inner platform sectors of Ourém.

The upper boundary of the mesosequence C/J is characterised by a major break in the sequential evolution of the Cenomanian-Turonian transition. This unconformity is contemporary to the installation of palaeokarsts in the Baixo Mondego and Nazaré, and to important changes in the nature of facies and sedimentation axis.

2.3. The Lower Turonian

The occurrence of marine carbonates with Early Turonian faunas is restricted to the northern sector of the sedimentation area, and to a narrow band of exposures situated over the southeast block of the “Cós-Leiria-Pombal fault” (Fig. 4). Units “K” to “O” have a maximal thickness of 18 meters (Figueira da Foz) and are organised as a mesosequence transgressive at the base, but strongly regressive upwards (Mesosequence K/O, Callapez, 1998). The first carbonate beds are dolomitic and overlie the Upper Cenomanian paleokarst surface over unit “J”. The basal sequence K/L records a positive evolution from dolomitic marls to platy marly limestones with abundant concentrations of oriented turritellids and structures suggesting the importance of storm events on the depositional environment. The upper part of the mesosequence (sequence M/O) consists of calcarenitic beds with fragmented corals, radiolitids and actaeonellids, gradually enriched upwards in coarse

Stage	Units of Choffat	SECTOR OF FIGUEIRA DA FOZ	SECTOR OF MONTEMOR-O-VELHO	SECTOR OF TENTÚGAL	SECTOR OF ANÇÃ - COIMBRA
Lower Turonian	O	Biocalcarenites with concentrations of debris of <i>Radiolites</i> , <i>Actaeonella</i> and corals. Micaceous limestones on the top	Micaceous limestones and sandstones with concentrations of branched bryozoans	Micaceous sandstones and biocalcarenites with cross-bedding	
	N				
	M				
	L	Plated limestones with concentrations of <i>Turritella</i> , <i>Mytiloides</i> , <i>Vascoceras</i> and <i>Tylostoma</i>	Micaceous sandstones with <i>Vascoceras</i> , <i>Mytiloides</i> and <i>Tylostoma</i>		
	K	Dolomitic limestones and marls with <i>Mytiloides</i>			
Stage	Units of Choffat	SECTOR OF NAZARÉ	SECTOR OF CÓS-JUNCAL	SECTOR OF LEIRIA	SECTOR OF OURÉM
Lower Turonian	M	Coarse sandstones with <i>Radiolites</i> + <i>Actaeonella</i>			
	L	Micaceous limestones with debris of corals, <i>Radiolites</i> and <i>Apicardis</i>	Marls with biostromes of <i>Radiolites peroni</i>		
	K				

Fig. 4 - Lower Turonian marine facies and macrofauna of West Portugal.

micaceous sediments, and overlaid by laminated pink sandstones. Ammonites occur on beds “K”/“L”, as an association dominated by vascoceratids from the middle Lower Turonian biozone of *Thomasites rollandi* (Chancellor *et al.*, 1994). *Mytiloides* spp. are also common elements of fossil faunas.

On the inner sectors of the carbonate platform, the correlative Lower Turonian record is entirely siliciclastic, with dominant facies of micaceous sandstone (Furadouro Sandstone, Barbosa, 1981). These beds yielded ammonoids and moulds of *Mytiloides*, correlative with the faunas of units “L” and “M”. The increase of micaceous facies was the result of a steeped progradation of the alluvial siliciclastic plain placed eastwards the carbonate platform.

Over the southeast block of the “Cós-Leiria-Pombal fault” the Lower Turonian record is a set of 2-3 meters of marls with abundant biostromes of the rudist *Radiolites peroni*. These beds (Unit 4 of Berthou, 1984) can be correlated with the sequence K/L of Figueira da Foz (Callapez, 1998). The upper part of the mesosequence K/O is recorded near the top of the exposures of Nazaré, but it is entirely built of coarse sandstones with debris of radiolitids and actaeonellids.

The strongly negative tendency of sequence M/O in both sectors of the sedimentation areas express the last and regressive steps of the Albian-Turonian cycle in West Central Portugal. This tendency has been interpreted as a consequence of the regional uplift and inversion of the central and southern sectors of Estremadura, since the end of the Cenomanian. However, from the “Cós-Leiria-Pombal fault” northwards, continental sedimentation

persisted until the end of late Cretaceous, with sporadic incursions of mixed facies with marine palaeofaunas.

3. DESCRIPTION OF THE STOPS

3.1. The Baixo Mondego region

The Cretaceous units from the northern sector of the platform are widely exposed in the region of Baixo Mondego, across fluvial valleys of the Mondego river drainage basin. The Cenomanian-Turonian carbonate beds are known from a set of discontinuous outcrops, lying from the localities of Figueira da Foz and Coimbra, to Soure and Pombal. The thickest and most expanded sections of the region are those of Figueira da Foz and Montemor-o-Velho, while the exposures of Tentúgal and Coimbra include some of the best outcrops to observe the complex variations of the inner platform facies.

The first comprehensive accounts on the Cretaceous stratigraphy of Baixo Mondego are those of Choffat (1897, 1900), and Soares (1966). On further works (Soares, 1980; Rocha *et al.*, 1981; Soares & Reis, 1982, 84; Soares *et al.*, 1985; Barbosa *et al.*, 1988) new lithostratigraphic units have been introduced, most of them with local significance.

The carbonate units (“B” to “O”) with ammonites and abundant fossil faunas are more expanded in the sections of Figueira da Foz and Montemor-o-Velho (50-65 meters thick), where the main sequences and unconformities are also easier to observe. Eastward of these areas the thickness of the Carbonate Formation

decreases and the facies content become richer on fine siliciclastics. These variations are more effective near Coimbra, where the whole carbonate succession is reduced to 10-15 meters of Upper Cenomanian sandy marls and limestones. The Lower Turonian is recorded as a few meters of fine micaceous sandstones (Furadouro Formation), and “C”/“D” are the single units that can be identified.

1st stop – The Cenomanian-Turonian and the Tethyan *Vascoceras* assemblages of Figueira da Foz

Location: Quarries of Salmanha, Figueira da Foz (U.T.M. 29 S NE 14500 44750) (Fig. 5).

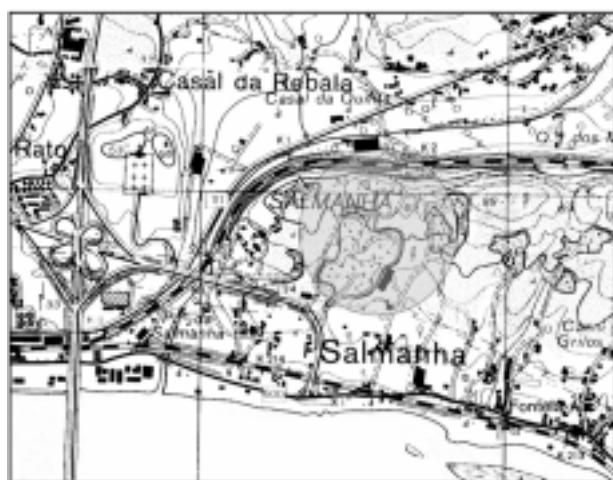


Fig. 5 - Location map of stop 1 (1: 25,000).

Purposes: After a panoramic view of the *cuesta* with the quarry of Salmanha, observation of the Upper Cenomanian and Lower Turonian ammonite facies, with an overview of the biostratigraphy and main palaeobiological features.

Description: The quarry of Salmanha shows a continuous exposure of the Upper Cenomanian units “C” to “L” from the type area of Figueira da Foz (Fig. 6, Pl. IA). The bottom of the quarry was opened according to stratification surfaces of Middle Cenomanian age (upper part of unit “B”). On the vertical faces of the northern front, the succession begins with nodular marly limestones (unit “C”, basal upper Cenomanian), and finishes with blue grey nodular limestones (unit “G”). On the lateral faces of the southern front, the succession also includes units “H”, “I” and “J”. The Lower Turonian units “K” to “N” can be observed just outside the quarry, next to the road to Coimbra.

Unit “C” is a set of six beds of marls and grey nodular limestones and marly limestones, containing abundant

shells of *Pycnodonte*, *Neithea*, *Pinna*, *Hemiaster*, as well many other benthic invertebrates. This unit marks the base of the Upper Cenomanian at the whole basinal area, being widely used as a marker bed. The ammonite fauna includes *Neolobites vibraeyanus* (d’Orbigny, 1851) and *Calycoceras (Calycoceras) naviculare* (Mantell, 1822). This association indicates the Upper Cenomanian standard zone of *Calycoceras guerangeri*, but the abundance of *Neolobites* is an indicator of the Tethyan affinities of the fauna.

Unit “D” is a bed of white cream limestone with fragments of massive or branching corals, disarticulated valves of *Rhynchostreon* and *Neithea*, and rare specimens of the echinoid *Anorthopygus*. The transition to units “E” and “F” is abrupt and involves a change into marls and marly nodular limestones. These beds yield an association of ammonites with *Vascoceras gamai* Choffat, 1898, one of the oldest known vascoceratids, but also *Pseudocalyoceras* sp. and *Euomphaloceras septemseriatum*. The occurrence of *Euomphaloceras* suggests a correlation with the Western European standard zone of *geslinianum*.

The thickest units “G” and “H” are the central part of the quarry fronts. They are formed by beds of grey and white compact limestone with abundant fragments of branched corals (*Dactylosmia*), *Pycnodonte*, *Tylostoma* and *Hemiaster*, together with *Thalassinoides* burrows. The upper limit of unit “G” is an hardground surface that yielded the younger *Euomphaloceras* of the succession. The ammonite fauna of unit “H” forms a new association, with the occurrence of new vascoceratid species (*Vascoceras douvillei* Choffat, 1898 and *V. kossmati* Choffat, 1898), together with *Spathites (Jeanrogericeras) subconciliatus* (Choffat, 1898) and *Pseudaspidoceras pseudonodosoides* (Choffat, 1898). This fauna indicates a zone of *pseudonodosoides* that can be compared with similar faunas from Tunisia (Chancellor *et al.*, 1994), and also correlates with the standard zone of *juddii*.

Unit “I” is a single bed of grey calcareous marl with abundant *Hemiaster scutigera*. This and the following beds can be sampled in the southern fronts of the quarry.

Unit “J” is a set of compact white limestone with an association identical to that of unit “H”, with the addition of *Puzosia* sp. and *Fagesia catinus* (Mantell, 1822). On the upper part of this unit, limestone beds show structures of dissolution and weathering, interpreted as an evidence of subaerial exposure and paleokarst development.

The first Lower Turonian beds (unit “K”) can be observed near the access to the quarry. They consist of

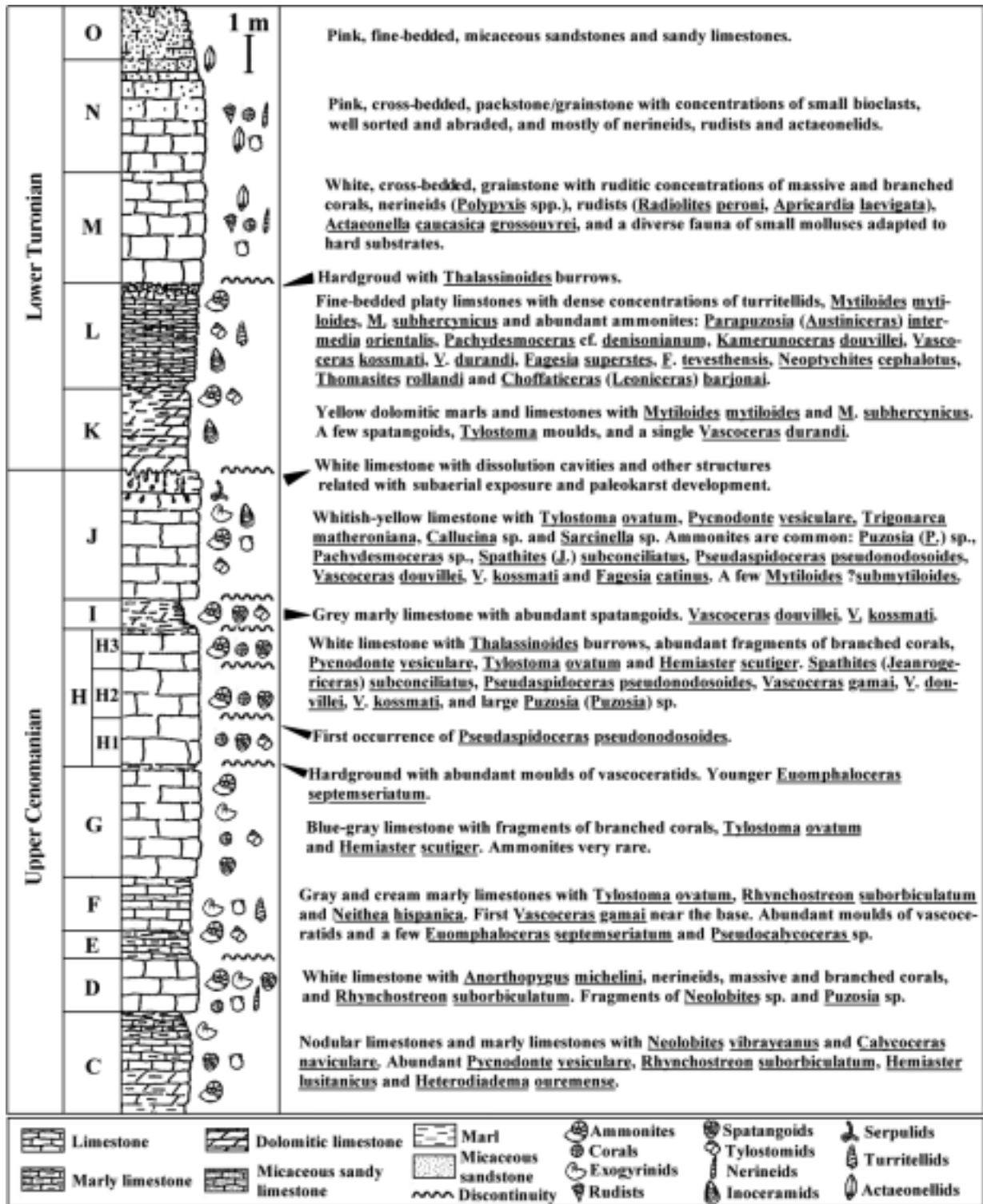


Fig. 6 - The section of Figueira da Foz. Stratigraphy and faunas.

dolomitic marls and limestones with irregular stratification and a fossil content reduced to a few moulds of *Mytiloides*, *Vascoceras* and *Tylostoma*. These dolomitic strata change to a set of platy limestones with storm concentrations of *Turritella* (unit “L”). These beds and the younger Turonian units (“M”, “N” and “O”) are exposed outside the quarry, near the village of Fontela

and the A-14 highway. The ammonite association of unit “L” contain a few vascoceratid species (*V. kossmati*, *V. durandi*, *Fagesia superstes*, *F. tevesthensis*), but also *Kamerunoceras douvillei* (Pervinquière, 1907), *Neoptychites cephalotus* (Courty, 1860), *Thomasites rollandi* (Thomas & Peron, 1889) and *Choffaticeras (Leoniceras) barjonai* (Choffat, 1898). This association

occurs together with *Mytiloides*, indicating the Tethyan biozone of *Thomasites rollandi* (Chancellor *et al.*, 1994).

Suggestions: Attempt to recognize the most important units and unconformities, comparing them with the illustrated section. At the lower part of the succession, you will find a set of fossiliferous beds where you can sample benthic fossils and Vascoceratid ammonites.

2nd stop – Stratigraphy of the inner platform: the example of Tentúgal

Location: Vicinal road near the chapel of Nossa Senhora dos Olivais, Tentúgal, 12 km westward of Coimbra, and 180 km to the North of Lisbon (U.T.M. 29 S NE 34025 54325) (Fig. 7).



Fig. 7 - Location map of stop 2 (1: 25,000).

Purposes: Observation of the Carbonate Formation (Middle Cenomanian to Lower Turonian), with the emphasis on the condensed facies with *Vascoceras* and *Tylostoma*, recording the inner sectors of the carbonate platform.

Description: The section of Senhora dos Olivais was chosen by Choffat (1900) and further workers (Soares, 1966, 1980; Lauerjat, 1982; Callapez, 1998) to illustrate the succession and faunas within the inner sectors of the carbonate platform, northward of the “Nazaré fault”. The mixed and condensed facies of these domains contain rich assemblages of benthic molluscs (*Tylostoma*, *Neithea*, *Exogyra*) as well as many Tethyan ammonites. The exposures along the vicinal road allow the

observation of the Middle and Upper Cenomanian units “B” to “J”, with a total thickness of 19 metres. The Lower Turonian sequence is exposed 1 km away, near the village of Tentúgal.

Unit “B” consists of 4 meters of mixed sandstone-limestone beds organized as positive or oscillating elementary sequences. The faunal content is dominated by the bivalves *Gyrostrea ouremensis*, *Anisocardia orientalis* and *Septifer lineatus*, frequently found as allochthonous concentrations in storm beds.

The basal Upper Cenomanian sequence C/D is registered as a succession of nodular marly limestones (unit “C”) and white calcarenites (unit “D”), with a thickness of 5 metres. The nodular facies yield the typical fauna of unit “C”, including *Neolobites vibrayanus*, *Pycnodonte vesiculare*, *Rhynchostreon columbum*, *Harpagodes incertus*, *Hemiaster scutiger*, and many other invertebrates. The fossil content of unit “D” also includes corals and nerineids (*Polyptyxis*, *Plesioplocus*).

The record of sequence E/I is a succession of nodular grey limestones and marls with interbedded layers of “rognons”. On the top there is a bed of grey marls with small *Hemiaster scutiger* (unit “I”). The identification of units “F”, “G” and “H” can’t be done due to condensation and lateral changes of facies. Vascoceratid ammonites are common in the upper 6 meters of the succession, but in many cases they occur as reworked moulds with erosional facets of hydraulic origin. The association has *Vascoceras gamai* Choffat, 1898, *V. adonense* Choffat, 1898, *V. barcoicense* Choffat, 1898, *Vascoceras* sp., *Rubroceras* cf. *alatum* Cobban, Hook & Kennedy, 1989, *Rubroceras* sp. and *Euomphaloceras septemseriatum* (Cragin, 1893), suggesting a mixing of faunas from two adjacent biozones.

The chapel of Nossa Senhora dos Olivais is placed upon 2 metres of nodular grey limestones with *Tylostoma* and pincers of crustaceans, which seems to be equivalent to the late Upper Cenomanian unit “J”. These beds are overlaid by a discontinuous succession of coarse micaceous sands and cross-bedded platy micaceous limestones, containing Early Turonian faunas that allow a correlation with units “L” and “M/N/O” of Figueira da Foz (mesosequence K/O).

Suggestions: The Upper Cenomanian marly facies with *rognons* of limestone are widely exposed on cultivated fields situated around the chapel. These areas are adequate to collect fossils samples with emphasis to *Vascoceras* and the gastropod *Tylostoma ovatum*. Many of these fossils are reworked moulds that exhibit erosional facets, related to the work of hydraulic agents.

The external surfaces of some *rognons* show also abraded sections of fossils, suggesting the hydraulic remobilization and abrasion of limestone fragments associated to the deposition of condensed sediments.

3.2. The Nazaré-Leiria-Ourém region

The Cenomanian-Turonian is widely exposed over the southeast block of the “Cós-Leiria-Pombal fault”, from the coastal village of Nazaré to the areas of Leiria and Ourém. Four geographic and structural domains can be considered: (1) Nazaré, (2) Alpedriz syncline, (3) Pousos syncline and (4) Ourém (Choffat, 1900; Berthou, 1973; Lauverjat, 1982).

Nazaré is the most external of these domains and presents the most expanded and complete section of the central sector. It is situated over a structural block bounded at the east by the diapiric anticline of Caldas da Rainha.

The Alpedriz syncline comprises a large number of exposures, including the classic sites of Juncal and Cós. This structure occupies a large structural block, bounded by the diapiric anticlines of Caldas da Rainha, Pataias, Maceira and Leiria-Porto de Mós.

Eastward of Leiria, the Cenomanian limestones can be observed in several sections exposed at the flanks of the Pousos syncline, with emphasis to those of Caranguejeira and Padrão. The inner domains of the carbonate platform are recorded in the region of Ourém, with emphasis on the sections of Olival-Boieiro (Crosaz-Galletti, 1979).

3rd stop – The Upper-Cenomanian rimmed shelf of Leiria-Ourém

Location: National road of Leiria-Caranguejeira, close to the villages of Carrasqueira and Souto de Cima, 120 km to North of Lisbon (U.T.M. 29 S NE 23750 01525) (Fig. 8).

Purposes: Observation of the Upper Cenomanian and Lower Turonian stratigraphic succession of the domains with calcarenitic facies, and coral and rudist buildups; contrast and correlations with the ammonite facies from the northern domains of the carbonate platform.

Description: The slopes of the road from Carrasqueira to Caranguejeira are one of the best exposures to observe the Upper Cenomanian units with corals and rudists. Together with some small inactive quarries opened in the vicinity, these exposures allow a continuous

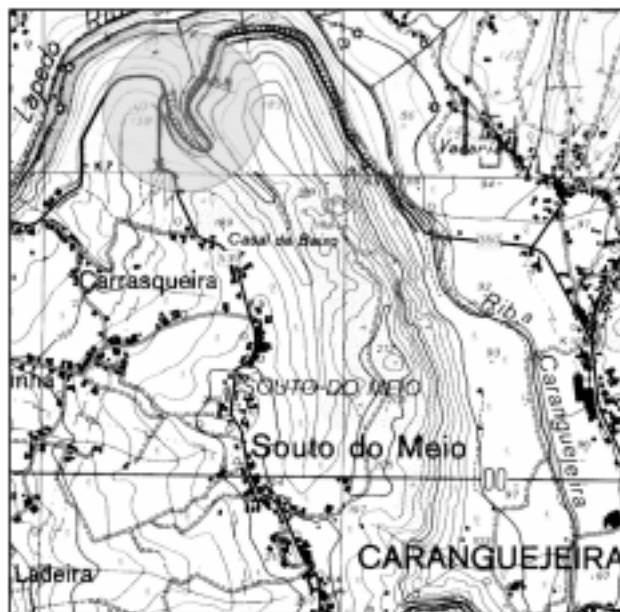


Fig. 8 - Location map of stop 3 (1: 25,000).

observation of the succession lying above unit “C”.

The basal Upper Cenomanian (sequence C/D) consists of beds of nodular marly limestone and coarse biotrititic limestone, with abundant alveolinids, echinoids and other benthic fossils, including *Neolobites vibrayanus* and *Anorthopygus michelini*.

Unit “E” is a bed of nodular grey limestone with abundant *Rhynchostreon columbum*. The first reef unit of the Upper Cenomanian succession (unit 1 of Berthou, 1984b) lies above this bed, and is composed of bioclastic limestone interbedded with red marls and marly limestone, with a total thickness of 7-8 meters. The dominant components of the biofacies are massive corals (*Dimorpharia* sp.), appearing as the frame builders of a large set of coalescent bioherms, with dimensions that reach 15-20 metres width and 5-10 high. Caprinulids are rare. This succession is correlative to the units “F” and “G” of the northern sector.

The upper boundary of the reef unit is a distinct surface, equivalent to the unconformity G/H of Figueira da Foz. Over this surface, there are 12 meters of white bioclastic limestone with cross stratification (unit 2 of Berthou, 1984d), very rich in fragments and disarticulated valves of radiolitids (*Durania arnaudi* and *Radiolites peroni*). These limestone beds can be seen in a small quarry situated close to the road.

The succession continues with a single bed of marl, containing a fauna dominated by *Tylostoma ovatum* and small *Hemiaster scutiger* (unit “I”). Over these marls, there are 5 metres of white limestone with debris of *Radiolites* and *Durania*, but also disarticulated valves of *Trigonarca matheroniana* (d’Orbigny, 1844) and oriented concentrations of *Polyptyxis* sp. and

Trochactaeon (Tr.) giganteus subglobosus (Münster, 1844). These beds have been grouped in the unit 3 of Berthou (1984d) and are characterized by regressive micaceous facies on its upper part. The faunal content and facies suggest an obvious correlation with the Upper Cenomanian sequence “J” of the northern sector of the platform.

The Lower Turonian is restricted to a few meters of dolomitic marls and laminated marls with biostromes of *Radiolites peroni* Choffat, 1886 and *Apricardia* (unit 4 of Berthou, 1984d). This succession is poorly exposed, but some beds can be sampled in cultivated fields near the road. According to Callapez (1998), these beds correlate with the sequence K/L of Figueira da Foz.

Suggestions: In order to observe the most interesting points of the succession, you can walk throughout the vicinal road to Caranguejeira. Take also a look across the magnificent landscape of the deep ravines incised in the Cenomanian reef limestone of Carrasqueira and Lapedo.

4th stop – The Cenomanian-Turonian transition on rudist and actaeonellid facies: the example of Nazaré

Location: cliffs and slopes of the promontory of Nazaré, a fishing harbour situated in the West coast, 100 km to North of Lisbon (U.T.M. 29 S MD 92850 84100 to 93450 83990) (Fig. 9).



Fig. 9 - Location map of stop 4 (1: 25,000).

Purposes: After a panoramic view of the promontory, with a regard to the diapiric anticline of Caldas da Rainha and to the Berlengas Islands (exposed Hercynian basement), observation of the middle and upper parts of the section, including the Upper Cenomanian and Lower Turonian beds with rudist and siliciclastic facies.

Description: The cliffs of the Nazaré promontory show one of the most interesting Cenomanian-Turonian sections of West Central Portugal (Pl. IB). Unfortunately,

the local stratigraphy is of difficult interpretation, due to the specificity of facies and to the inaccessibility of the southern slopes.

Choffat (1900) and Berthou (1973) made the first descriptions of these exposures. After, Lauerjat (1982), Reis *et al.* (1997), Callapez (1998) and Corrochano *et al.* (1998) described new details of the section, including the macro and microfauna, the palaeoecology and some relevant sedimentological features associated to palaeo-karst development.

Easier to observe at low tide, the section of the northern beach ranges from the top of the bioclastic limestone beds with rudists (unit “H”) to the Lower Turonian sequences K/L and M/O. The succession is cut by an intrusion of weathered olivinic basalts. Near the base, the Upper Cenomanian units consist of bioclastic packstones and grainstones with cross bedding and concentrations of radiolitid debris. Unit “I” is a single bed of pink marly limestones with dense *Thalassinoides*.

Unit “J” is registered as a succession of marls and limestone with a thickness of 7 metres. The lower 4 metres are cross-bedded bioclastic limestones, containing a few rudstone layers with abraded fragments of *Durania arnaudi*, *Apricardia* sp., corals, nerineids and *Thochactaeon*. This association has affinities with those from the region of Cós-Juncal and Leiria. Callapez (1998) suggests that the majority of these fragments are bioclasts transported from the *Durania arnaudi* biostromes placed eastwards, over the shallow domains of the carbonate platform. This carbonate sequence finishes with a bed of nodular marly limestone densely bioturbated.

The upper part of unit “J” is exposed near two small caves opened at the beach, and consists of white or pink laminated calcarenites, containing concentrations of small bivalves, nerineid and actaeonellid gastropods.

The Lower Turonian sequences can be observed with detail on the northern corner of the cliffs, where they rest unconformably on Upper Cenomanian limestones (sequence J). Sequence K/L is composed of 5-6 metres of white and pink limestone interbedded with thin beds of cross-bedded, micaceous sandy limestone. These strata are organized as fining-upwards elemental sequences, initiated with coarse bioclastic beds with erosive bases. The fossil content consists of pavements with highly abraded fragments of *Radiolites peroni*, *Apricardia* sp., massive and branched hexacorallia, and *Actaeonella caucasica grossouvrei* (Callapez, 1998). The faunal similarity with the *Radiolites peroni* biostromes is evident, suggesting that the concentrations

of Nazaré were storm accumulations of shell debris, transported from biostromic fringes placed eastwards, over the diapiric shoals of Caldas da Rainha and Leiria. In recent studies by Reis *et al.* (1997) and Corrochano *et al.* (1998), the depositional setting of these beds was interpreted as a shallow inner lagoon, marginal to rudist buildups and affected by storm-dominated events.

The boundary between sequences K/L and M/O is marked by an abrupt change of facies associated to an erosional surface. Over the micaceous limestones, there are 4.5 metres of yellow, cross-bedded, coarse sandstones with rounded fragments of *Acataeonella caucasica grossouvrei* and *Radiolites peroni*, two typical species from the middle Lower Turonian assemblages of unit "M". The nature of these facies indicates a shallow environment, controlled by wave action and tidal currents.

The Cenomanian-Turonian succession is affected by a deep palaeokarst evidenced by coarse breccias with clasts of limestone, large dissolution cavities filled with cross-bedded coarse sandstones, and collapse structures with chaotic accumulations of blocks. Most karst structures are concentrated in the Upper Cenomanian limestones of sequence J, suggesting that the early stages of subaerial weathering were contemporary to the palaeokarst of Figueira da Foz (unconformity J/K) and to the Cenomanian-Turonian transition. However, there is also evidence of a later episode, previous to the Late Campanian. After Corrochano *et al.* (1998), the paleokarst of Nazaré can be considered as an endokarst controlled by local tectonics and sedimentary evolution.

The upper part of the section is a set of Late Cretaceous units that rest unconformably over the Cenomanian-Turonian. The younger of these units (Upper Campanian-Maastrichtian) is a thick succession of alluvial red mudstones interbedded with conglomerates (Sitio da Nazaré Formation; Antunes, 1979), containing clasts of Upper Cenomanian limestone and a nicely preserved fauna of small land-snails (*Anadromus ribeiroi*, *Punctum basalticum*, *etc.*).

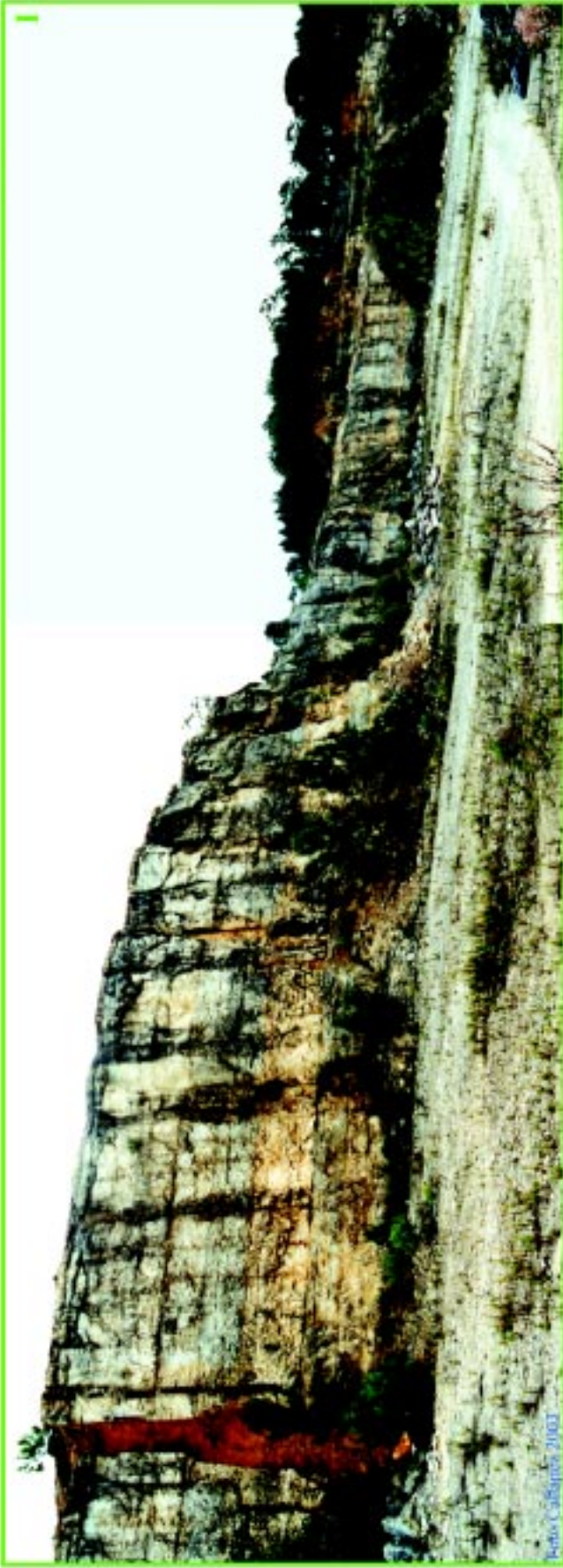
Suggestions: After an observation of the beach section, you can walk across the top of the promontory, in direction to the fort. Close to this track, you have the chance to observe exposures of the Sitio da Nazaré Formation, and olivinic basalts cutting the Turonian sediments. Near the lighthouse, a small stairway can be used to observe the Lower Turonian and uppermost Upper Cenomanian successions. The Lower Turonian beds are affected by small synsedimentary faults, suggesting a local tectonic activity during the last stages of development of the carbonate platform.

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Plate 1A - Panoramic view of the Upper Cenomanian from the quarry of Salmanha (Figueira da Foz)..



Pl. 1B - Panoramic view of the southern cliffs of the Nazaré promontory. 1 - Lower Cenomanian; 2 - Middle Cenomanian; 3 - Upper Cenomanian; 4 - Lower Turonian; 5 - Upper Campanian and Maastrichtian silticlastics.

Pre-Meeting Field Trip A4

**Cycles and sedimentary events in the Cenozoic
of the Setúbal peninsula
(Lower Tagus basin)**

J. PAIS
P. LEGOINHA

Cycles and sedimentary events in the Cenozoic of the Setúbal peninsula (Lower Tagus basin)

J. PAIS¹ & P. LEGOINHA¹

1. FOREWORD

The continental margin of Portugal is quite interesting, considering that the Neogene units are well represented and complete. This is particularly true for the Lower Tagus basin.

At the beginning of the XIX century, research was being conducted by the mineralogist (and future politician) José Bonifácio de Andrada e Silva (1763-1838). In the 30's and 40's, the contributions followed from Baron Von Eschwege, Alexandre Vandelli, Daniel Sharpe and a few others. Meanwhile, there was a considerable development in investigations with the second Geological Commission (1857-1867), mainly due to the field works of Carlos Ribeiro and the important contributions of Francisco Pereira da Costa.

Studies were restarted (towards the end of the 19th century) at the Geological Survey by Jorge Cândido Berkeley Cotter. The lithostratigraphic classification of Lisbon's Miocene, classical and still in use, is due to him.

Some palaeontologists such as Oswald Heer, F. Fontannes, Perceval de Loriol, Frédéric Roman and Gustave Dollfus also made important contributions.

After a long pause, there was a resumption of Neogene studies by Portuguese and foreign geologists at the service of Portugal. G. Zbyszewski deserves a special reference.

Since 1958 new developments have been achieved by the work of M.T. Antunes in coordinating several research projects (with the collaboration of colleagues from other countries) on the palaeontology and geology of the Lower Tagus basin. Knowledge of the chronostratigraphy and micropalaeontology (ostracoda, foraminifera, palynology) has been significantly improved.

2. GEOLOGICAL FRAMEWORK

Since the end of the Cretaceous, the West Iberia has been subjected to compression that has led to the differentiation of grabens, some of which have become large sedimentary basins during the Cenozoic. This is the case of the Douro, Mondego, Lower Tagus, Alvalade, Algarve and Guadalquivir basins. The Lower Tagus Basin has been affected by several marine incursions during the Miocene and the Pliocene; in the Alvalade basin there was a marine incursion in the Messinian and in the Mondego basin there was only one in the Piacenzian. In the Algarve, a temperate marine platform sedimentation type developed during the Early and Middle Miocene and, during the Late Miocene, the region East of Faro became part of the Guadalquivir basin.

As far as is known, it only was in the Mondego basin, immediately to the North of the Pombal-Caldas fault, that sedimentation continued in continental environ-

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ments after the Maastrichtian (e.g. Late Paleocene of Silveirinha). In other regions, there seems to have been a hiatus corresponding to the Paleocene and Early Eocene. Sedimentation only began after the Pre-Pyrenaic phase responsible for the opening of several Tertiary grabens.

2.1. Paleogene

The Paleogene of the Lisbon region is represented by continental deposits overlying volcanic rocks of Late Cretaceous age and underlying fossiliferous marine deposits of Aquitanian age. It consists of the Benfica Formation, over 400 meters thick in Lisbon, but only about 200 meters thick at the Arrábida chain. In Lisbon, different macrosequences have been defined starting with the works of Choffat (1950). The lower one (C1 to C5 of Choffat, 1950; Unit A of Antunes, 1979; Reis *et al.*, 1991) is composed of 100 meters of coarse detrital deposits, containing arkosic layers and mainly Paleozoic clasts. It is capped by a massive carbonate layer (“Calcários de Alfovelos” of Choffat, 1950; Reis *et al.*, 1991). These carbonates represent an important decrease in terrigenous input and the arrival of palustrine conditions. The upper macrosequence (C6 to C15 of Choffat, 1950; Units B + C of Antunes, 1979; Reis *et al.*, 1991) is composed by sandy mudstones and conglomerates which increase in abundance and clast-size towards the top. These conglomerates contain abundant Mesozoic clasts, including Cretaceous limestones and hornfelses from the metamorphic aureole of the Sintra igneous massif. This sedimentary break and change in source area may be attributed to the Pyrenaic phase and the uplift of the Sintra region (Azevêdo & Pimentel, 1995).

2.1.1. Palaeogeographic remarks

The first modern palaeogeographic reconstruction for the Paleogene at the Lower Tagus Basin (Carvalho, 1968) proposed the existence of a large closed tectonic depression filled by coarse materials coming from the surrounding mountains and from hills inside the basin, which provided the feldspars and Mg for the arkosic and palygorskitic deposits (*op. cit.*, p. 197-198). The intercalated carbonate layers correspond to “tectonic quiet periods” and the development along the basin of lacustrine and palustrine conditions (*op. cit.*, p. 198). The climatic conditions were hot and humid, with intense hydrolysis of the silicates, that promoted the authigenesis of palygorskite. More recently, several studies in similar deposits from Central Portugal have suggested hot and sub-arid climatic conditions (Azevêdo

& Carvalho, 1986; Azevêdo, 1991; Cunha, 1992; Barbosa, 1995), producing coarse siliciclastic depositional systems in continental environments, related to active fault-scarps and multiple alluvial-fans. These conditions explain the abundant and characteristic alteration of the Paleogene deposits (Pimentel *et al.*, 1996).

The source of the Paleogene deposits of the Lisbon region (distal sector of the Tagus Basin) is still somewhat controversial. Azevêdo (1991) and Azevêdo & Pimentel (1995) had previously proposed the existence of at least two narrow depressions, bordering a central horst elongated NE-SW. This horst explains the abundance and coarseness (up to 30 cm) of Paleozoic clasts, the abundance of arkosic sands and the presence of unstable minerals in the Paleogene deposits of Alenquer, Lisbon and Arrábida (nowadays more than 50 km away from the nearest Palaeozoic outcrops). In the Arrábida chain, this horst is bordered by the Setúbal-Pinhal Novo fault, active from the Jurassic until the Late Miocene (Azevêdo, 1998).

2.2. Neogene

The Neogene of the Lower Tagus basin is remarkable for the following: its geographical position, at the boundary between the Mediterranean and Atlantic realms; the complete representation of nearly all the stages, from the Lower Miocene to the Pliocene; the interfingering of continental and marine levels; the richness of the data it provides, allowing direct correlations and biostratigraphic scale comparisons; the possibilities of geological correlations; and its economic importance.

The Pliocene, whose palaeogeography is quite different, formed in a band that stretched, along the shoreline, and to fluvial deposits covering large areas inland.

2.2.1. The Miocene in the Western region of the Lower Tagus Basin

In Lisbon, the thickness of the Miocene does not exceed 300 meters, and the Pliocene is hardly represented; to the South, the thickness of the whole Neogene increases to about 1200 meters whereas the outcropping Pliocene reaches a thickness of 320 meters. Subsidence was balanced by sedimentation. Continental debris arrived in large quantities during orogenic episodes.

Several transgressions and regressions are recognised in the distal part of the basin and strongly influenced the palaeogeography (Pls. I and IIA) (Pais, in press).

Ten Miocene depositional sequences (Pl. I) (Antunes *et al.*, 1999; 2000b; Legoinha, 2001) have been characterised in the distal region of the Lower Tagus Basin.

The positions of the levels containing land mammals have been recognised. This allows direct correlation between marine and continental horizons for the Lower Miocene and the early part of the Middle Miocene.

Climatic changes have been recognised (Pl. IIB). As far as marine environments are concerned, tropical conditions prevailed, at least since the genesis of the Aquitanian coralline barrier reefs. A temperature maximum (comparable to present day conditions in the Gulf of Guinea), was reached in the Upper Burdigalian and Langhian. Later, the prevailing thermal conditions were closer to those off the today's Moroccan coast. The continental faunas and vegetation clearly show an alternation between humid events (that correspond to forest, rather than savannah or steppe environments) and dry events.

2.2.2. Miocene of Setúbal Peninsula

The Miocene of the Setúbal Peninsula comprises a succession of mostly marine beds spanning the lowermost Miocene (Aquitanian) to Upper Miocene (Tortonian) (Zbyszewski *et al.*, 1965; Zbyszewski, 1967; Antunes & Pais, 1983; Pais, 1986; Nascimento, 1988; Antunes *et al.*, 1992; Sen *et al.*, 1992; Lauriat-Rage *et al.*, 1993; Manuppella *et al.*, 1994; Antunes *et al.*, 1992; 1995a, b, c; 1997, b; 1998a, b; 1999; 2000b; Legoinha 2001). The stratigraphic (Pl. I) and environmental framework (Pl.IIB) is mainly based on foraminifera, ostracoda, vertebrates and palynomorphs from the Foz da Fonte, Penedo, Penedo Norte and Ribeira da Lage sections (Antunes *et al.*, 1995a; 1997; 1998a; 2000b).

Most of these beds are rich in fossils. Hence it has been possible to obtain a fairly accurate local time-scale based on marine and continental fossils, K-Ar glauconite ages, ⁸⁷Sr/⁸⁶Sr dating, and palaeomagnetism.

2.3. Pliocene of Setúbal Peninsula

The Pliocene sequence starts in the Setúbal Peninsula (S.P.) with the first fluvial deposits carried out by the pre-Tagus river (alternation of reddish coarse layers of conglomerates and arkosic sands outcropping on the N side of the Albufeira syncline). They are followed by fluvial sands (Santa Marta sands) reaching about 320 m thickness near the depocentre (Pinhal Novo) (Pl.IIIA).

2.3.1. Santa Marta Sands

These sandy deposits are exploited for the construction industry in several large sand-pits on the Setúbal

Península. The depositional environment was fluvial. Some marine characters were observed at the present coastal cliffs. The coastline should have been more to the W than the present one (Azevêdo, 1983).

White sands outcrop in several localities on the Setúbal Peninsula. They are interpreted as a consequence of their position in the migrating main channel of the braided system. In contrast, the reddish and yellow colours are due to the deposition in more oxygenated sites such as the lateral bars and banks of the river. In these sandy deposits clasts only occur as pavement of the channels. The coarse clasts include basalts, granites or metaquartzites coming from Sintra and Lisbon areas through the left bank drainage system of the pre-Tagus.

Lignite and diatomite were only deposited in the central axis of the Setúbal Peninsula (Azevêdo, 1983).

2.3.2. Belverde Conglomerate

The Santa Marta Sands are overlain by a 5 m thickness of alternating conglomerates and arkosic sands. The Belverde Conglomerate clasts are almost exclusively of quartzite (73%) and quartz (27%). The clasts are white because of their demineralised cortices, and are sub-rounded. They can reach 15 cm, in diameter, and outcrop only to the W of Ribeira de Coia. They extend to the coastal cliffs and to the hills of the Cabo Espichel area and are also found at the same altitude in the N part of the Setúbal Peninsula (Almada-Caparica), attesting the existence of an old alluvial plain; now they are completely dissected by the stream network (Azevêdo, 1983, 1986, 1997a, b, c).

The Belverde Conglomerate yielded Pre-Acheulian lithic artefacts (Azevêdo *et al.*, 1979b, Azevedo & Cardoso, 1986).

2.3.3. Marco Furado Formation

Above the Belverde Conglomerate there is a "raña" formation that allows the dating of the path change of the Tagus river between these two events. It represents an alluvial fan depositional system with its origin in the Arrábida Chain extending to the N into the Barreiro area. The Marco Furado Formation is a debris-flow deposit with very coarse angular clasts in a red clay-sandy matrix. The biggest clasts, 15 cm long, are enveloped by red iron oxide coating due to their mineralised cortex. They are exclusively of Paleozoic rocks, such as quartz, metaquartzite, jasper and schist (Azevêdo 1979; 1982; 1983; 1993; 1997a,b; 1998; Azevêdo & Pimentel 1993).

3. STOPS

3.1. The Paleogene of the Arrábida Chain

Stop 1 - São Caetano (Azeitão)

(Pimentel *in* Antunes *et al.*, 2000a) (Figs. 1 and 2)

This stop is located along the road between Azeitão and São Caetano.

The Paleogene deposits are in contact with Cretaceous fluvial sandstones, are covered by Lower Miocene marine marls, and have a thickness of *ca.* 200 meters. Coarse reddish sandstones constitute the basal Paleogene deposits. The strata dip at around 20° to NNW and a small fault (N30°, 80° W) has cut these deposits and brought them into contact with the Cretaceous. The clasts are exclusively composed of Paleozoic lithologies, mainly vein-quartz and minor lydite, metaquartzite, jasper, etc. In the first 40 meters of the sequence there are several conglomeratic intercalations, becoming gradually finer. This trend is accompanied by a change from mainly red to orange, yellow and finally whitish colours, with increasing carbonate accumulations. At around 35 meters from the base a 5 meters thick massive carbonate layer suggests post-depositional cementation (Pimentel, *in* Antunes *et al.*, 2000a).

Higher deposits are composed of coarse mottled sands, organised in 3 to 5 meters thick units, dipping around 15° to NNW. The units have a conglomeratic base, followed by pinkish sandy clays with carbonate concretions. The most important aspect in this part of

Fig. 1 - Itinerary and geographical situation of the stops.

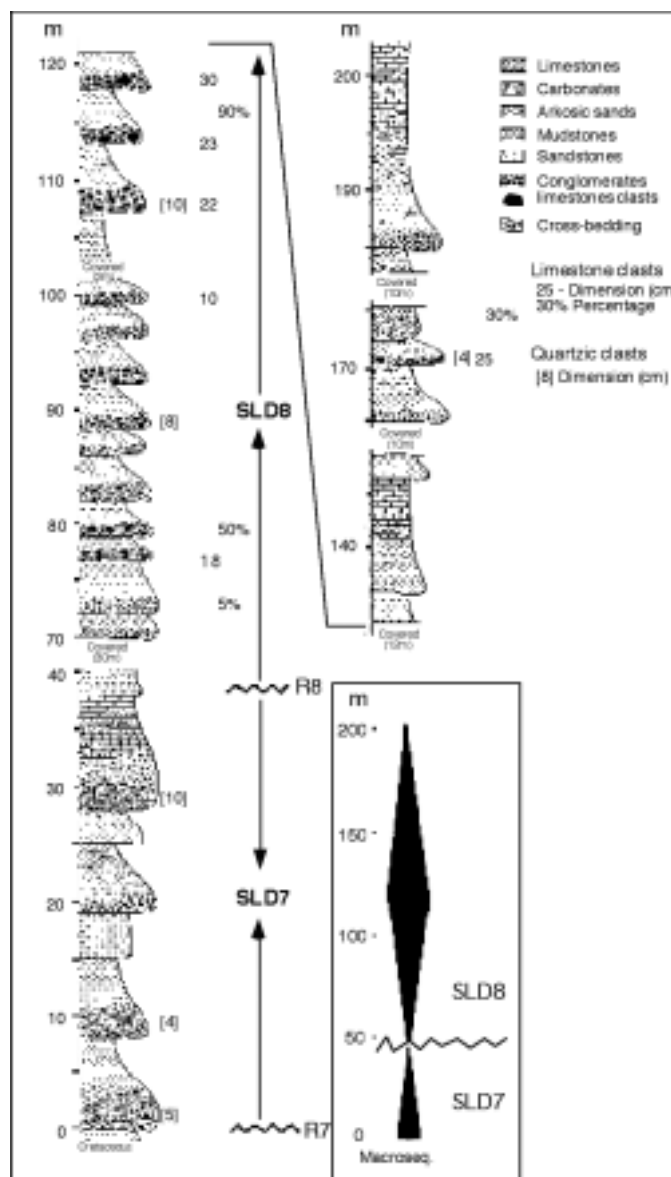


Fig. 2 - The Paleogene of São Caetano (Azeitão).
Log by Nuno Pimentel
(adapt. from Report SC 4, 1992;
in Antunes *et al.*, 2000a).

the log is the presence of small Cretaceous limestone clasts (Seifert, 1963) totally absent in the basal layers and therefore indicating a new source-area. The relative abundance and size of these clasts clearly increase towards the top, attaining 80 % of the gravel and 30 cm long about 2/3 of the way up the Paleogene macrosequence.

At the top of the sequences, and increasingly towards the top of the section, carbonate accumulations show irregular concretionary layers. As a consequence, the predominant colour gradually becomes lighter, and the upper layers are composed mainly of yellow and white sandy clays. The capping 5 meters consists of a massive carbonate layer, dipping around 10° to NNW (Pimentel, *in Antunes et al.*, 2000a). Zbyzsewski (1964) and Zbyszewski *et al.* (1965) reported continental gastropods from these carbonates. Azerêdo & Carvalho (1986) recognised fragments of ostracoda and charophytes.

Sedimentological laboratory studies were made in 1992 and 1993 (supervision by T.M. Azerêdo – Reports SC 4 to SC 7, 1992-93). The gravels are composed mainly by quartz and limestones, the later ones increasing towards the top.

The grains are mainly quartzitic, with sub-angular (sometimes angular or even sharp) shapes. The heavy minerals are mainly tourmaline, zircon and andalusite, with minor moscovite and garnet. Iron oxides and opaque grains are abundant or even predominant. Exoscopic observation of the quartz grains showed the presence of physical features indicating medial to high-energy aquatic transport, followed by an immobilisation in silica supersaturated environments. The clays are composed by an average of 30% smectite, 25% palygorskite, 25% kaolinite and 20 % illite, and all these minerals are present in all samples. Facies control on clay mineralogy (detected in closely located samples) makes difficult the identification of a clear vertical macrosequencial trend.

The predominance of massive sandy and conglomeratic facies (Gms, Sc, Sm and Fm) in the Paleogene of São Caetano reflects the transport and deposition in sediment gravity-flows (SG) and laminated sand-sheets (LS). Multiple debris-flows and sheet-flood events accumulated coarse siliciclastic layers, without significant erosional surfaces. An alluvial-fan environment was responsible for the accumulation of these deposits, not far away from an up-lifted area with metasedimentary low-grade Paleozoic lithologies. The deposits represent proximal, medial and distal alluvial-fan facies, reflecting both progradation and retrogradation of the depositional system. Climatic conditions are considered to have been generally hot and sub-arid, with strong seasonality (Pimentel, *in Antunes et al.*, 2000a).

Sedimentary pauses between successive depositional events allowed the development of pedogenic processes related to significant phreatic level fluctuations. During these pauses, hydromorphic features and rhizogenic carbonate accumulations developed as a consequence of seasonality. Mobilisation of iron and calcium during wet phases was followed by precipitation of iron oxides and calcium carbonates during dry phases (Pimentel, *in Antunes et al.*, 2000a).

Studies of other areas of the Lower Tagus Basin indicate that post-depositional modification is thought to have continued even after pedogenesis and burial, during early diagenesis (Cunha, 1992, 2000; Barbosa, 1995). No specific studies have been made on these materials, but detailed petrographic and mineralogical analysis in very similar deposits from southern Portugal allows a confident extrapolation: “Pedogenesis and diagenesis were highly active processes in the modification of the primary features of the deposits, specially of their carbonate and clay mineralogy. Both processes inter-acted in space and time, in a continuum from the moment of deposition until very early sub-surface diagenesis. Palygorskite neoformation was developed all through that time-interval, as a result of meteoric, soil and ground water circulation and mixing. Dolomitization started later but was partially contemporary with the clay transformations, probably inter-acting with them” (Pimentel *et al.*, 2000; Pimentel, *in Antunes et al.*, 2000a).

The vertical evolution of the sequence reflects the existence of two main macrosequences: the first with 40 to 50 meters thickness (SLD 7) and the second one with more than 150 meters (SLD 8). The first sequence represents the basal filling of the basin, supplied by up-lifted Paleozoic areas, some kilometers away from São Caetano. The positive trend towards finer and more reduced deposits reflects the gradual filling of the basin and increasing clay and carbonate neoformation in low-gradient flood-plains or even palustrine areas. The second macrosequence has a negative trend in the first half, followed by a positive in the second half. This trend is particularly clear in the abundance and dimension of the Mesozoic clasts, reflecting the up-lifting behaviour of the Cretaceous areas. These areas must have corresponded to some part of the Arrábida chain, possibly the São Luís anticline, affected by an intra-Paleogene movement. The upper part of the Paleogene indicates generalised flat areas and no meaningful tectonic activity. Palustrine conditions gradually developed and, locally, lacustrine conditions resulted in thick carbonates (Azerêdo & Carvalho, 1986; Pimentel, *in Antunes et al.*, 2000a).

3.2. The Miocene of the Arrábida Southern limb

The Arrábida Mountain Range is the best example in Portugal of a structure of Alpine age. During the Miocene it acquired its present structure: overthrust reverse faults striking ENE-WSW and N-S or NNE-SSW sinistral lateral ramps. In the Southern limb of the Arrábida the Miocene forms a narrow discontinuous band striking E-W.

The Southern limb of the Arrábida is interesting as far as the Quaternary is concerned. At Forte da Baralha, in the western part of the Arrábida chain, marine terraces are well developed. The deposits have yielded mollusca (Dollfus *in* Choffat & Dollfus, 1904-07; Zbyszewski, 1943; 1958). Calcareous conglomerate and sandstones with abundant shells constitutes the 5-8 m terrace of Forte da Baralha (Zbyszewski & Teixeira, 1949). It yielded two Languedocian implements *in situ*, one at Forte da Baralha, another one at Lapa de Santa Margarida (Breuil & Zbyszewski, 1945). Daveau & Azevedo (1980/81) published on the geomorphology of southwestern Arrábida. Cardoso (1994) produced a synthesis of the Quaternary coastal geology and geomorphology of Sesimbra Municipality with emphasis on the marine terraces, the faunas and the lithic implements.

Stop 2 and 3 - The Miocene of the Arrábida Southern limb (Pl.IIIB, Figs. 1 and 3)

Marine sedimentation started in the beginning of the Middle Burdigalian (Sr dating \approx 18.8 Ma) probably in the early stages of the Burdigalian transgression represented in Lisbon by local lithostratigraphic divisions III and IVa (Pl. I). The deposits consist of biocalcarenites as well as yellowish fine grained sandstones poor in fossils overlaying the Paleogene (unit **a**, Fig. 3). At Portinho da Arrábida its average thickness is 30 meters, striking N75°W and dipping 40°N.

The unit **a** is followed by algal concretion rich whitish and yellowish biocalcarenites, coarse-grained sandstones, and conglomerates with carbonate cement (unit **b**). In the westernmost part of Chã da Anixa they overlie unit **a** with an angular unconformity whereas to the East the contact is a paraconformity. Elsewhere, there is an angular unconformity over the Middle Jurassic. The average thickness is 35 meters. Sr date (\approx 16.5Ma) is still Burdigalian.

Over this assemblage there are fossil poor whitish and/or yellowish siltites (unit **c**). The thickness is approximately 8-9 meters. They may date from the late Burdigalian.

Later Miocene deposits consist of about 76 meters whitish fossiliferous biocalcarenites, striking N30°E,

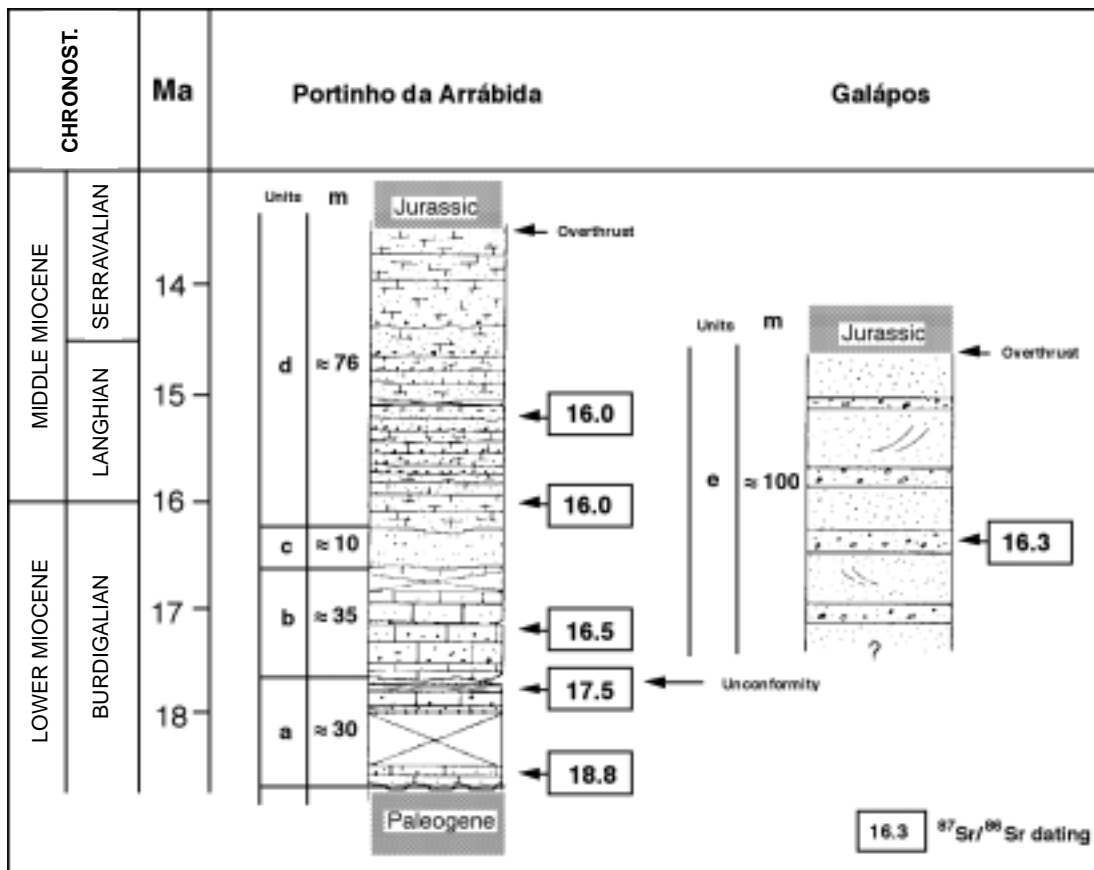


Fig. 3 - Lithological columns and dating for the Miocene of the Southern limb of the Arrábida Chain (Antunes *et al.*, 1995a).

dipping 25° SE; the dip increases to the East up to 50° SE. Sr dating is 16 Ma (beginning of the Middle Miocene).

Between Galapos and the Figueirinha beaches, there are outcrops of coarse-grained sandstones interlayered with conglomerates with fossiliferous beds (unit **e**); Cross-bedding structures indicate currents oriented to the East at the bottom and to the West at the top. The thickness is around 100 meters; dipping is about 25° N. They are overthrust by the Lower Jurassic. Sr analysis indicates an age of 16.3 Ma in the middle part (end of Burdigalian). They may be a lateral equivalent of units **b** and/or **c** (Pl. IIIB, Fig. 3).

Unit **b** is folded, locally overthrusting units **c** and **d** and laying unconformably over **a**. Unit **d** contacts unit **c** through an irregular surface. It was not possible to correlate unit **e** in the field with the other units, their attitude and lithology being completely different (Antunes *et al.*, 1995a).

Unit **a** of the Portinho da Arrábida Neogene, locally folded into a closed syncline with SE vergence, is in sequence with the Paleogene, the Cretaceous and the Upper Jurassic. These inverted units are part of the overturned limb of the Formosinho anticline. The latter is limited in the continental shelf to the South by the main overthrust of the Arrábida Chain (Ribeiro *in* Ribeiro *et al.*, 1990); and to the North by another overthrust which has placed the Dogger in contact with the Miocene. This assemblage forms a duplex structure probably extending for several kilometers. The tectonic phase, that gave rise to the folding of the lower unit, is dated around 17 Ma. The units **a** to **d** were folded in a sequence of anticline, syncline and anticline. The axis trends approximately ENE-WSW, dips to ENE and passes respectively between Pedra da Anixa e Chã da Anixa, at Coelhos beach, and to the North of this beach (Pais *et al.*, 1991; Antunes *et al.*, 1995a) (Pl. IIIB). Folding occurred less than 16 Ma ago and is compatible with the Neocastillian phase well known in Iberia during the Langhian.

Quaternary deposits are also known near Portinho da Arrábida. Antunes *et al.* (1992) described quartz and silex Mousterian implements collected *in situ* at Creiro in the southern limb of the Arrábida Chain. The source rocks of the lithic material are slope deposits with quartz and Miocene biocalcarenite clasts up to 50 cm in size, in a fine sandy or reddish silty matrix. Particularly important are the infillings of the karst surfaces as well as the coastal deposits over the marine abrasion surface eroded in the Miocene calcarenites and related caves such as those at Lapa de Santa Margarida and Gruta da Figueira Brava.

3.3. Miocene of the Arrábida northern limb

3.3.1. Stop 4 - Foz da Fonte (Figs. 1, 4 and 5)

At Foz da Fonte, Lower Miocene directly overlies Lower Cretaceous (Albian/Cenomanian) limestones with a low angle unconformity. Sediments are mainly fossiliferous biocalcarenites and marls. The lowermost beds include cobbles and pebbles of Cretaceous volcanic rocks and limestones. The middle section consists of detrital shallow facies with concentrations of *Turritella* and erosion surfaces. One of these is overlain by a remarkable oyster shell concentration. These detrital, shallow marine facies mark the beginning of the Burdigalian transgression. The upper part of the section has a marly cyclic character.

Planktic foraminifera are abundant at the lower levels and in the upper levels. *G. altiapertura* occurs from near the base to the top of the section. Scarce *Catapsydrax unicavus* at high levels precludes correlation with younger zones than N5/N6. In the lowest beds, the ostracods *Hemiciprudeis helvetica* and *Pokornyyella lusitanica* (typical Aquitanian species) occur for the last time. Upper beds contain typical Burdigalian forms (Nascimento, 1988).

Benthic foraminifera (*Amphistegina*, *Elphidium*, *Quinqueloculina*, *Asterigerina*) in the lower beds indicate an infralittoral environment. A greater depth may have been attained at the bed corresponding to the sample 4; planktic forms become more abundant and *Cibicides*, *Cancris*, *Pleurostomella* and *Uvigerina* are present in the benthic assemblage showing circalittoral influence.

In the lowermost levels, the presence of the thermophile ostracoda *Cnestocythere truncata* and *Pokornyyella lusitanica* indicate warm waters.

Disconformities can be observed in the middle part of the section; variability of diversity and composition of benthic assemblages, as well as the scarcity or absence of planktic forms suggests environmental instability (between the littoral and the infralittoral with little circalittoral influence).

Towards the upper beds, environmental conditions seem to have been more stable. An increase of planktic forms, greater diversity of benthic foraminifera (*Lenticulina*, *Cassidulina*, *Brizalina*) and ostracoda suggest infralittoral with circalittoral influence. Ostracods point to a decrease in water temperature.

3.3.2. Stop 5 - Penedo Norte (Figs. 1, 6 and 7)

The outcrop at the northern part of the Praia das Bicas cliff shows some levels higher than at Penedo.

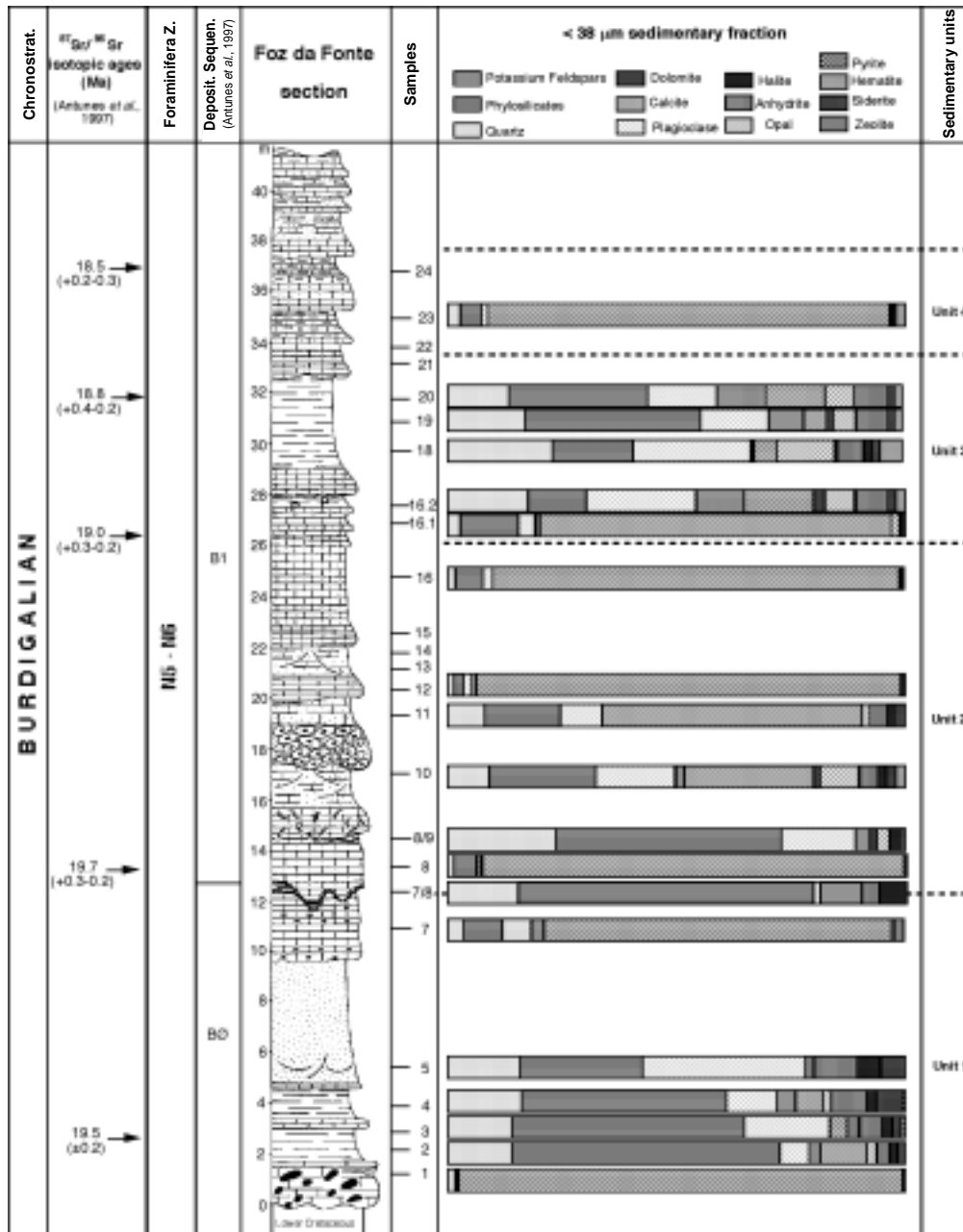


Fig. 4 - Foz da Fonte section. Mineralogy of the <38 μm sedimentary fraction (Antunes *et al.*, 1998a).

The lower beds (1 to 5) yielded *Globigerinoides* cf. *altiapertura*, *G. subquadratus*, *G. triloba*, *Globoquadrina dehiscens*, *Globorotalia mayeri* and *G. praescitula* (N7/N8, Upper Burdigalian).

No typical Lower Miocene ostracod species were found. In bed 5 *Bosquetina carinella* is plentiful; the ostracod association suggests 50 to 100 meters depth (circalittoral); *Olimfalunia costata*, common in the Middle Miocene, occurs.

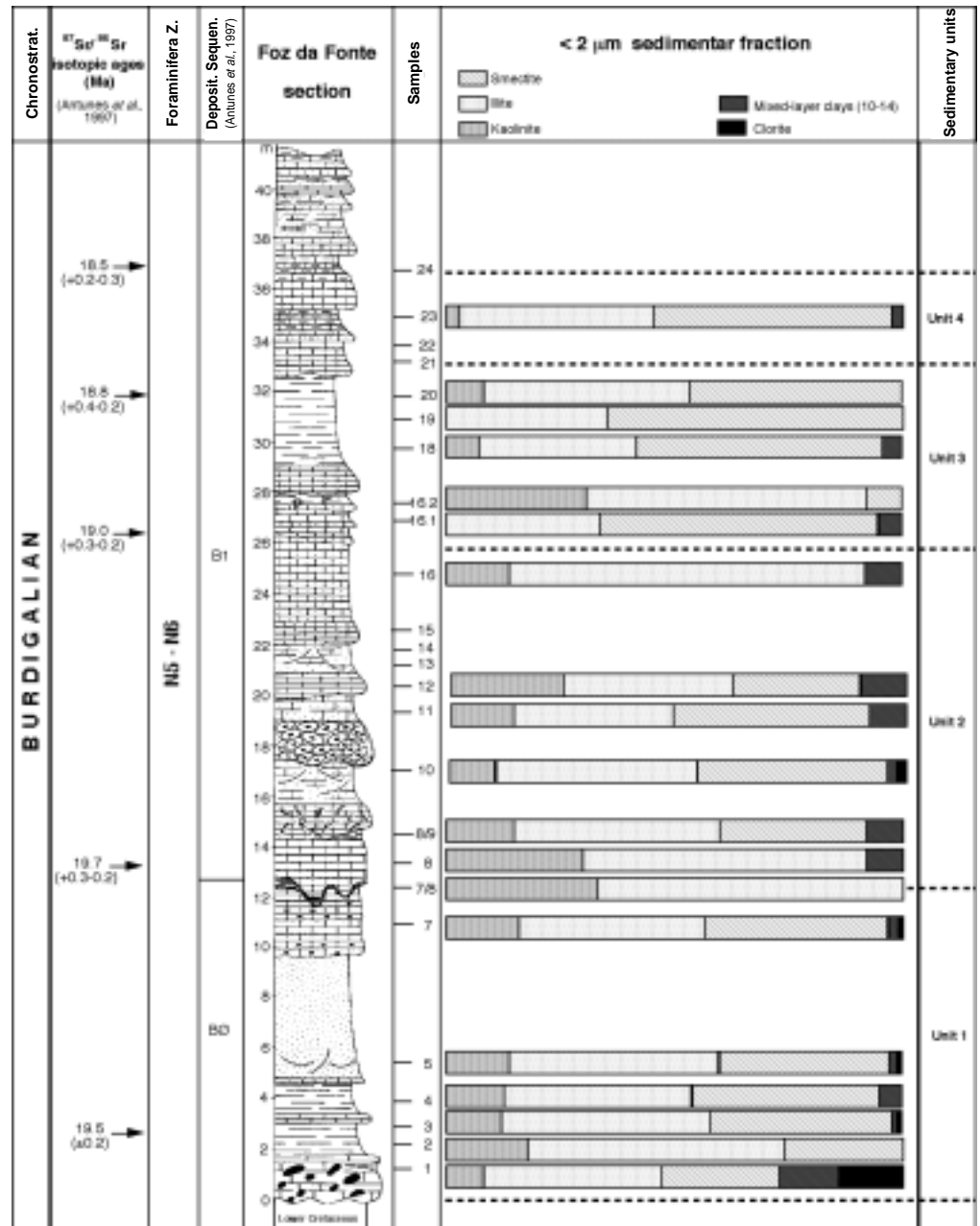
There is an erosion surface between beds 5 and 6. On this surface there is a conglomerate whose elements are mainly very abraded bivalve casts with a black, phosphate-rich patina. *Orbulina suturalis*, *Praeorbulina* cf. *glomerosa* and *P. transitoria* (N9, Langhian) occurs,

however, absence of forms from the evolutionary lineage *G. sicanus* to *P. glomerosa* suggests that there is a gap corresponding to part of N8.

A rich association from the lowermost part of bed 7 indicates N10 (*Globorotalia peripheroronda*, *Orbulina suturalis*, *O. universalis*, *Praeorbulina transitoria*). *Globigerinoides subquadratus* and frequent *Globorotalia* cf. *menardii* (N11, Lower Serravallian) occur at the top of bed 7.

The overlying conglomerate (bed 8) is unsuitable for the study of microfauna. It contains glauconite and fragments of phosphate crusts. Pectinid shells from bed 8 were $^{87}\text{Sr}/^{86}\text{Sr}$ dated (H. Elderfield, Cambridge University): 11.5 to 13Ma.

Fig. 5 - Foz da Fonte section. Mineralogy of the <math>< 2 \mu\text{m}</math> sedimentary fraction (Antunes *et al.*, 1998a).



Bed 9 is middle-grained, glauconite rich sandstone. K-Ar age (C. Regêncio Macedo, Coimbra University) is 10.97 ± 0.25 Ma, Lower Tortonian.

Vertebrate remnants are common (mainly in bed 8). Among sharks, the most frequent are Carcharhinids (*Hemipristis*, *Galeocerdo* and *Negaprion*) stenotherm, warm water forms. Lamniform sharks are also very common, especially *Isurus*. There is a contrasting variation among marine mammals, Sirenians being frequent until the early Middle Miocene (but nearly disappearing afterwards), while cetaceans (*i. e.* delphinids) and even a seal occur in later units. The seal means non tropical environments and a post-14Ma age. A land mammal (Cervid) was also recognised. These

Vertebrate remnants whose ages range from Late Burdigalian-Langhian to Serravallian accumulated in marine environments that changed from tropical to temperate (Antunes *et al.*, 1997).

As far as bed 10 is concerned, an association of *Globorotalia cf. menardii*, *G. mayeri*, *G. scitula*, *Neogloboquadrina continua* and *Globigerina druryi*, as well as the absence of *Globigerinoides subquadratus* indicates N14 (Late Serravallian).

Ostracods suggest a Serravallian age for beds 6 to 10. *Loxocochoa (L.) ducasseae*, *Pterigocythereis (P.) jonesi*, *Ruggieria nuda*, *Ruggieria tetraptera tetraptera* and *Bythocythere sp.* from beds 7 and 10 show that circalittoral stage environments were attained (Nascimento, 1988).

The commonest dinoflagellata (*Lingulodinium*, *Operculodinium*, *Polysphaeridium zoharyi*, *Spiniferites* spp.) point to warm, high salinity waters. On other hand, *Homotryblium*, *Lingulodinium machaerophorum*, *Operculodinium israelianum* and *Tuberculodinium vancampoe* (beds 8 and 9) suggest a depth decrease. The richest (bed 5) corresponds to open sea waters (Antunes *et al.*, 1998a).

Spores are scarce. The Polypodiaceae are the best represented family (beds 8 and 5). Gymnosperms predominate. Angiosperms have been found only in bed 5: Oleaceae in association with *Ephedra*, Compositae and grasses may point to dry environments (Antunes *et al.*, 1998a).

Plio-Quaternary sands cover all the Miocene deposits.

3.3.3. Stop 6 - Ribeira da Lage (Figs. 1 and 8)

The outcrops, 2 km North of Penedo Norte section, show medium to fine grained, micaceous sands with frequent roughly decimeter thick beds. *Chlamys macrotis* is very common towards the top.

The planktic foraminifera association is marked by *Globorotalia cf. menardii*, *Neogloboquadrina conti-nuosa*, *Globigerina apertura*, *G. druryi*, *Globigerinopsis aguasayensis*, *Orbulina suturalis* and *O. universalis*. Neither *Globorotalia mayeri* nor *Neogloboquadrina acostaensis* were observed (N15 zone).

Ostracoda suggests infralittoral environments; species regarded as typical of the Tortonian are lacking (Nascimento, 1988).

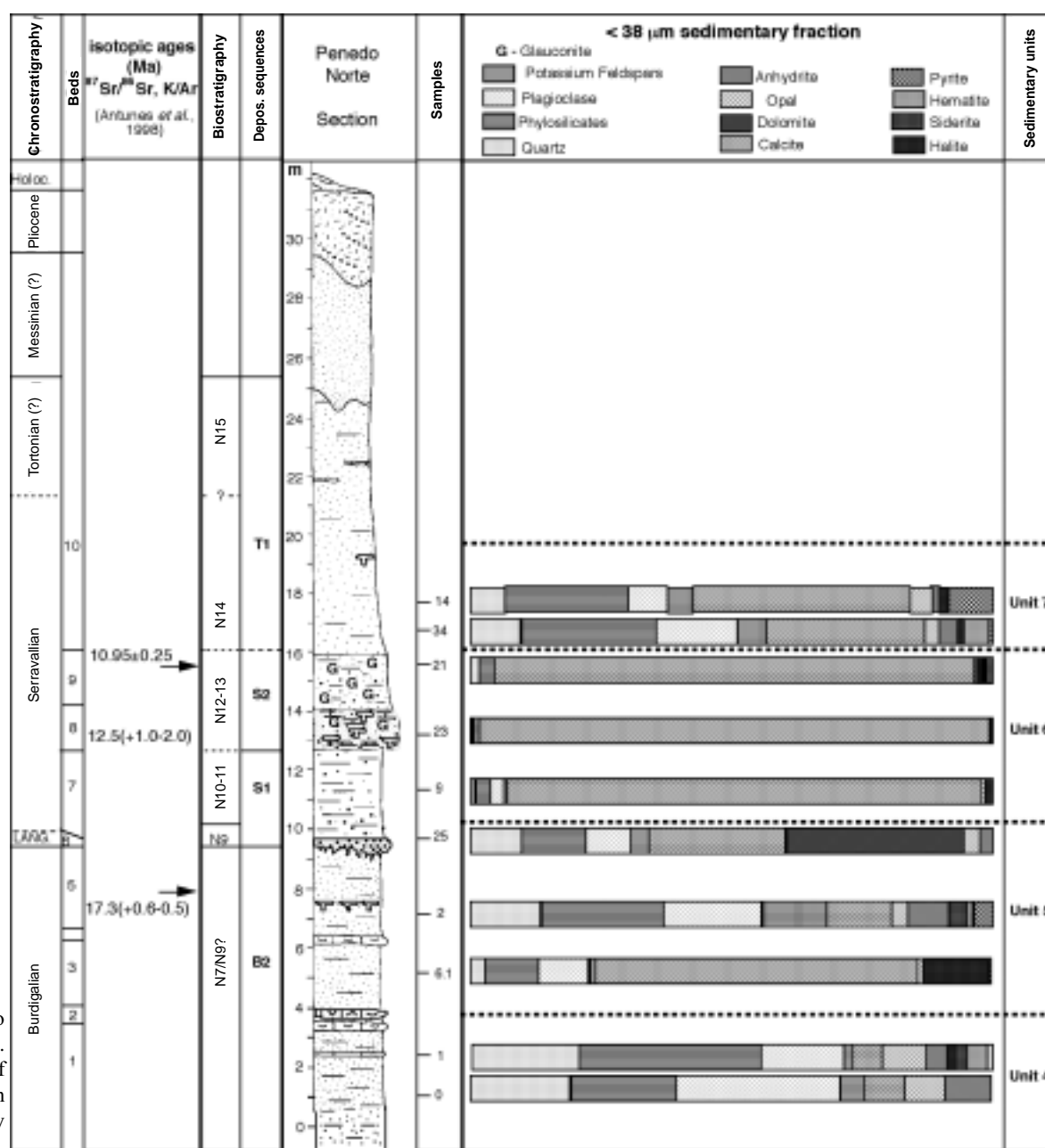


Fig. 6 - Penedo Norte section. Mineralogy of the < 38 μm sedimentary fraction.

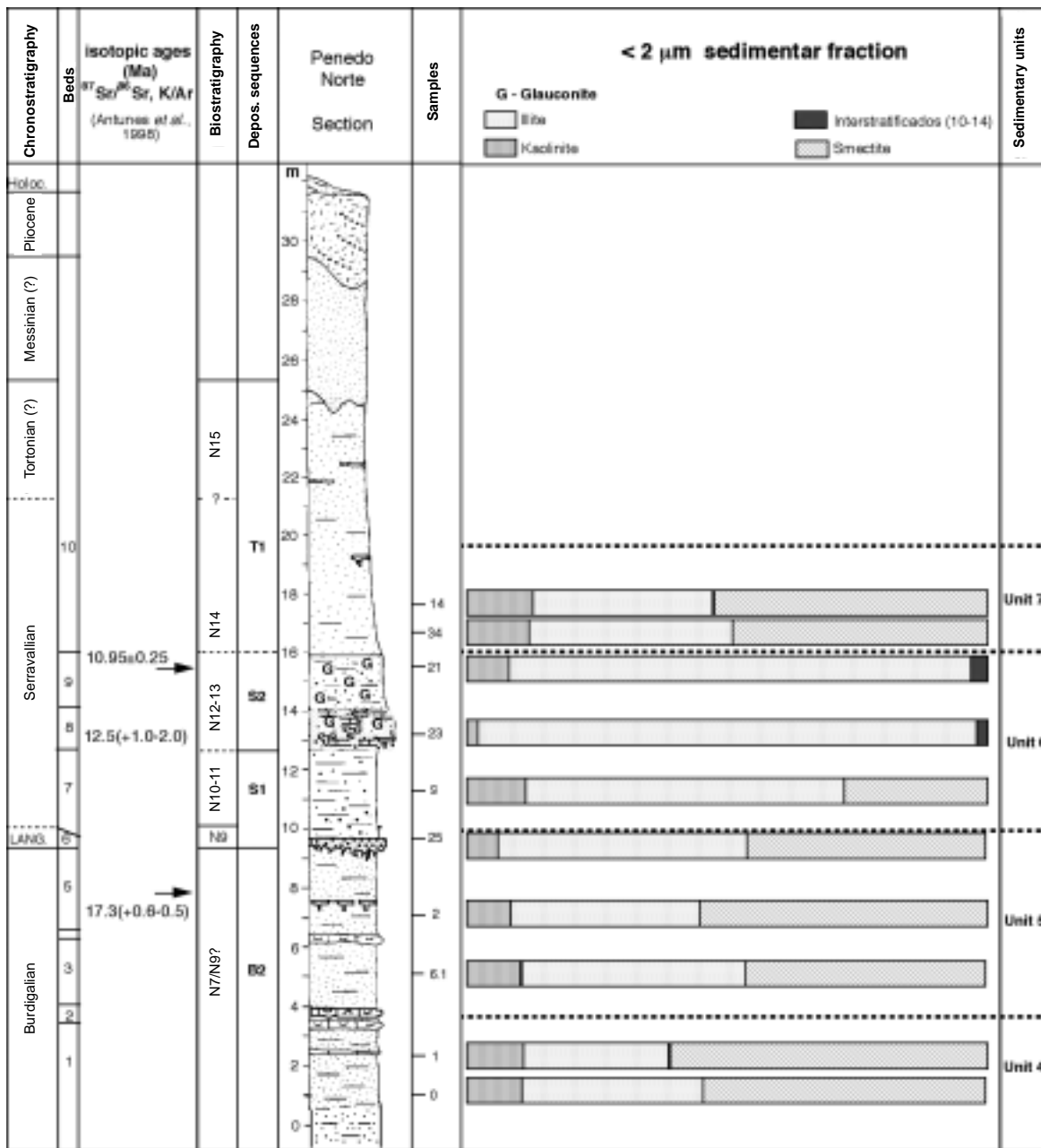


Fig. 7 - Penedo Norte section. Mineralogy of the $< 2 \mu\text{m}$ sedimentary fraction (Antunes *et al.* 1997).

Dinoflagellates are rare and poorly diversified. *Lingulodinium*, *Polysphaeridium* and *Spiriferites pseudofurcatus* predominate.

Spores are always scant. However some hepatics (*Anthoceros*) and ferns (Polypodiaceae) are present. Pollen are commoner than at Penedo Norte. Bialate pollen predominates. In the lower part of Ribeira da Lage section, Compositae, Amaranthaceae/Chaenopodiaceae and *Ephedra* are frequent. This association is related to the close-by littoral. Otherwise, *Ulmus*, *Myrica*, *Castanea* and *Ilex* point to a temperate and humid climate. As for the upper part of the section, the pollen assemblage comprises *Quercus* and Compositae associated with *Cathaya* and *Keteleeria*, which indicates

a rather warm and humid climate (*Cathaya* and *Keteleeria* still live in evergreen Chinese forests).

Sample 16 *Chlamys* shells were dated, $^{87}\text{Sr}/^{86}\text{Sr}$: 11.5 to 15Ma.

4. C AND O ISOTOPE ANALYSES

$\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ analysis from Foz da Fonte, Penedo and Ribeira da Lage

The isotopic curves (error: 0.03‰ $\delta^{13}\text{C}$ and 0.11‰ for $\delta^{18}\text{O}$) show strong oscillations, mainly in respect of Pectinid $\delta^{13}\text{C}$ contents which may attain 5δ (Foz da Fonte samples FF7 and FF8). These oscillations are in

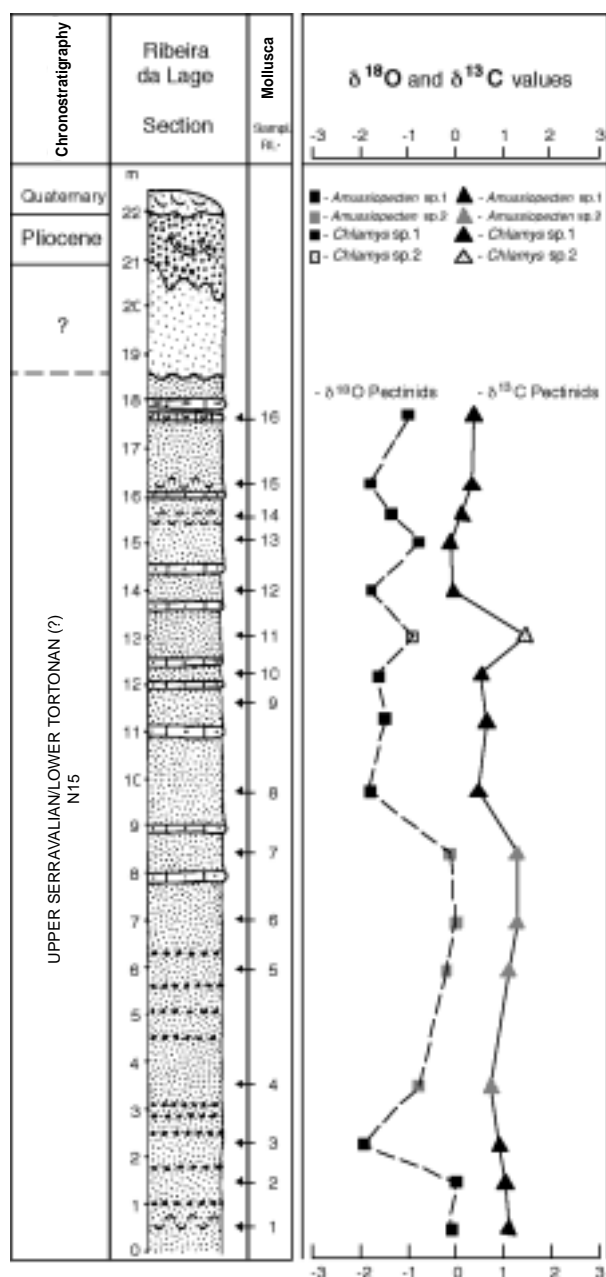


Fig. 8 - Ribeira da Lage section. C and O isotope data (Antunes *et al.*, 1997).

agreement with local sea level fluctuations as suggested by ostracoda, benthic foraminifera and sedimentation features.

During the Burdigalian, a $\delta^{18}\text{O}$ increase has been registered, mainly from the Pectinids; this can be related to water cooling in the top of the section. The water temperature, should be higher than today. Extant *Chlamys varia* specimens from the Algarve region give $\delta^{18}\text{O}$ values about 1.13‰ heavier than any of the studied Pectinids. A maximum of temperature seems to have been attained around 19.6 Ma (Early Burdigalian).

There is no covariation of C and O in the studied samples. Some anomalous isotopic values between Pectinids and Oysters are registered.

The $\delta^{13}\text{C}$ values from Foz da Fonte are much lower than those from the Middle Miocene sections of Penedo and Ribeira da Lage (Antunes *et al.*, 1997).

All the isotopic values obtained from mollusks of Penedo and Ribeira da Lage are in close agreement to the central field of “shallow-water mollusks and foraminifera” of Milliman (1974).

Oxygen values are more changeable (more variable conditions) at Penedo Norte (2,30‰ range for pectinids, Fig. 10) than at Ribeira da Lage (1,21‰ range for *Chlamys* and 0,26‰ for *Amusiopecten*, Fig. 8). At the Penedo Norte section the isotopic values for pectinids show more variation than those for *Ostrea*. The same for Carbon isotopes. At the Penedo Norte section depth variation cannot be clearly correlated to the isotope distribution. Contradictory data from pectinids and oysters (PN-2 and PN-5 samples) may be the result of re-sedimentation.

At the Ribeira da Lage section (Fig. 8), the $\delta^{18}\text{O}$ curve points to open marine environments and, from the RL-8 sample upwards, to higher temperatures. The $\delta^{13}\text{C}$ decrease (also from RL-8 sample upwards), may be related to growing continental influence or to sea bottom more oxidising environments.

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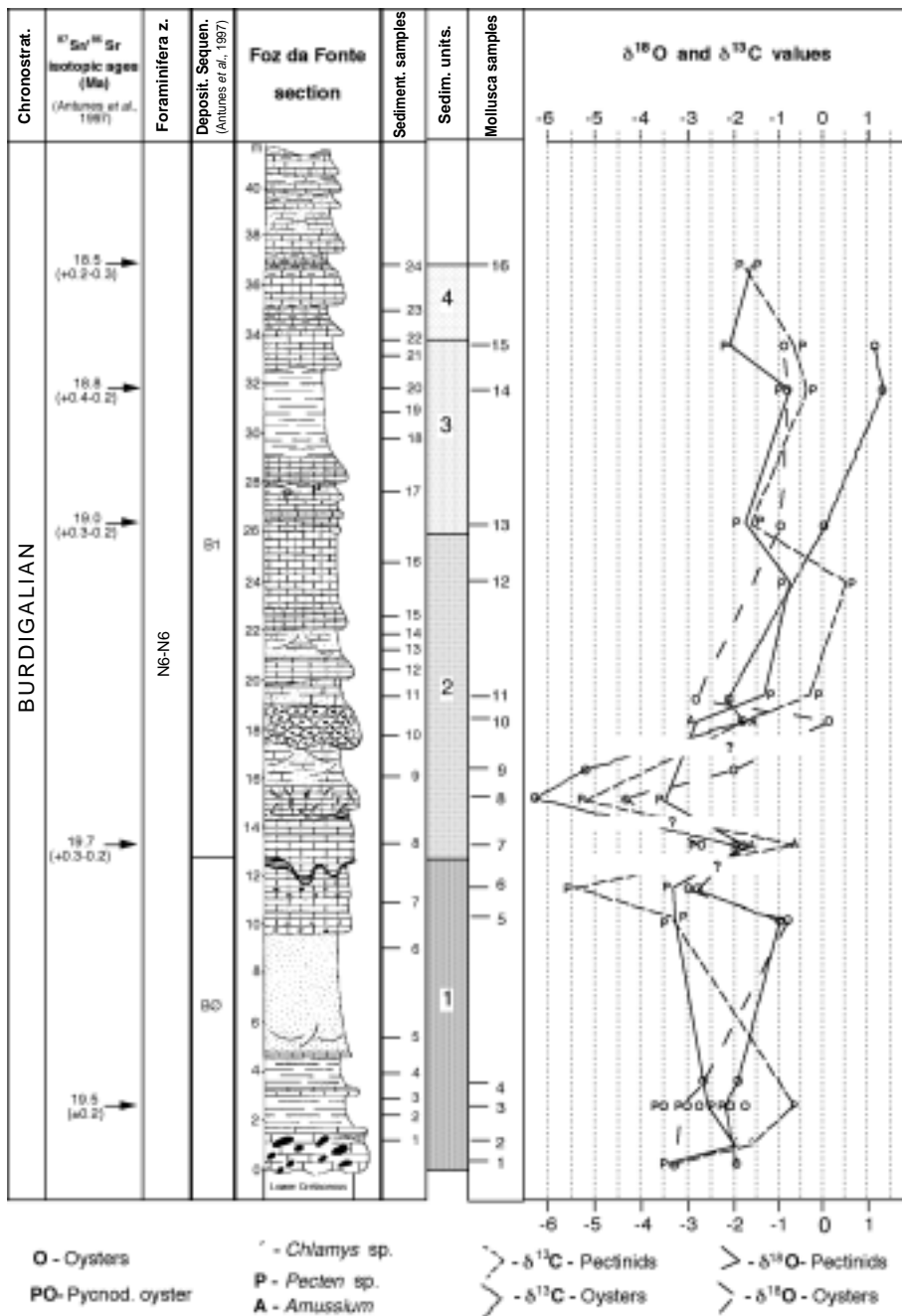


Fig. 9 - Foz da Fonte section. C and O isotopes and sedimentary units (Antunes *et al.*, 1998a).

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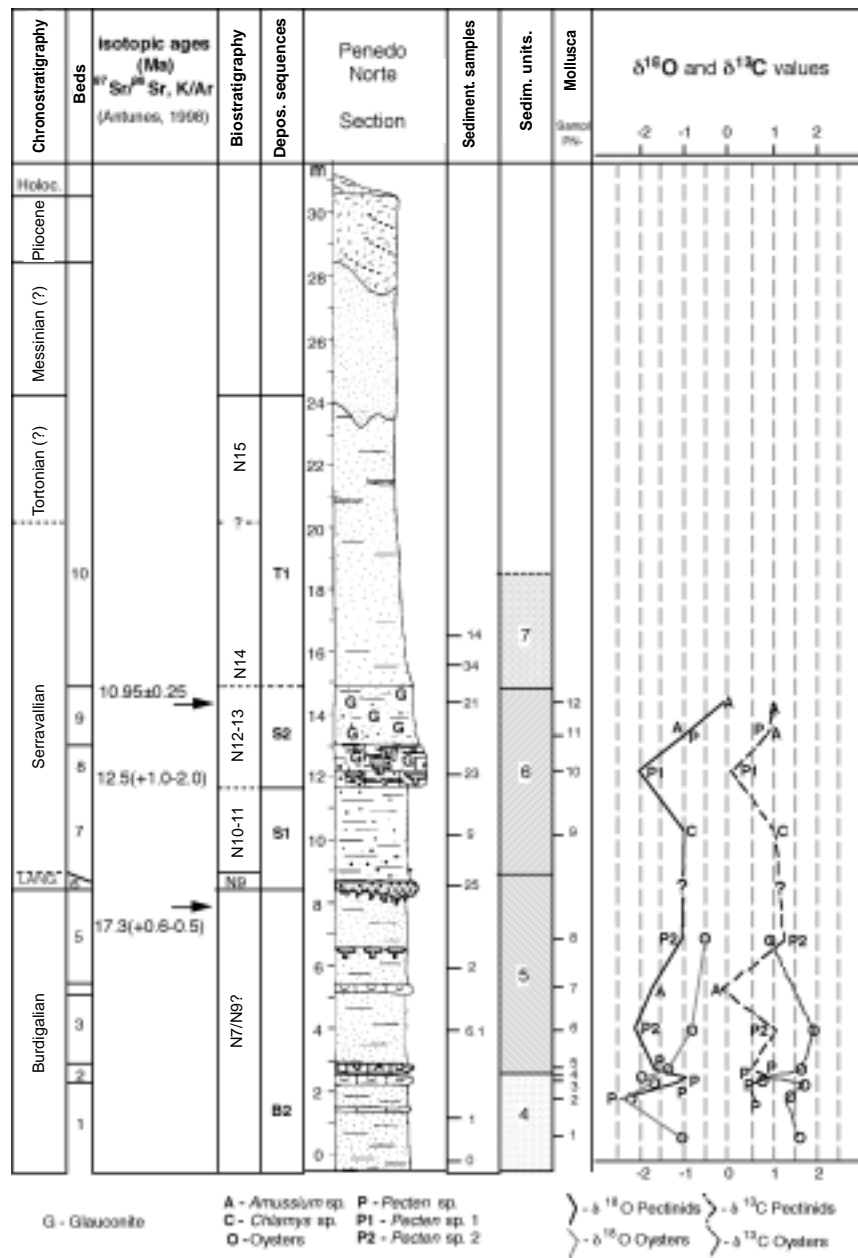


Fig. 10 - Penedo Norte section. C and O isotopes and sedimentary units (Antunes *et al.*, 1997).

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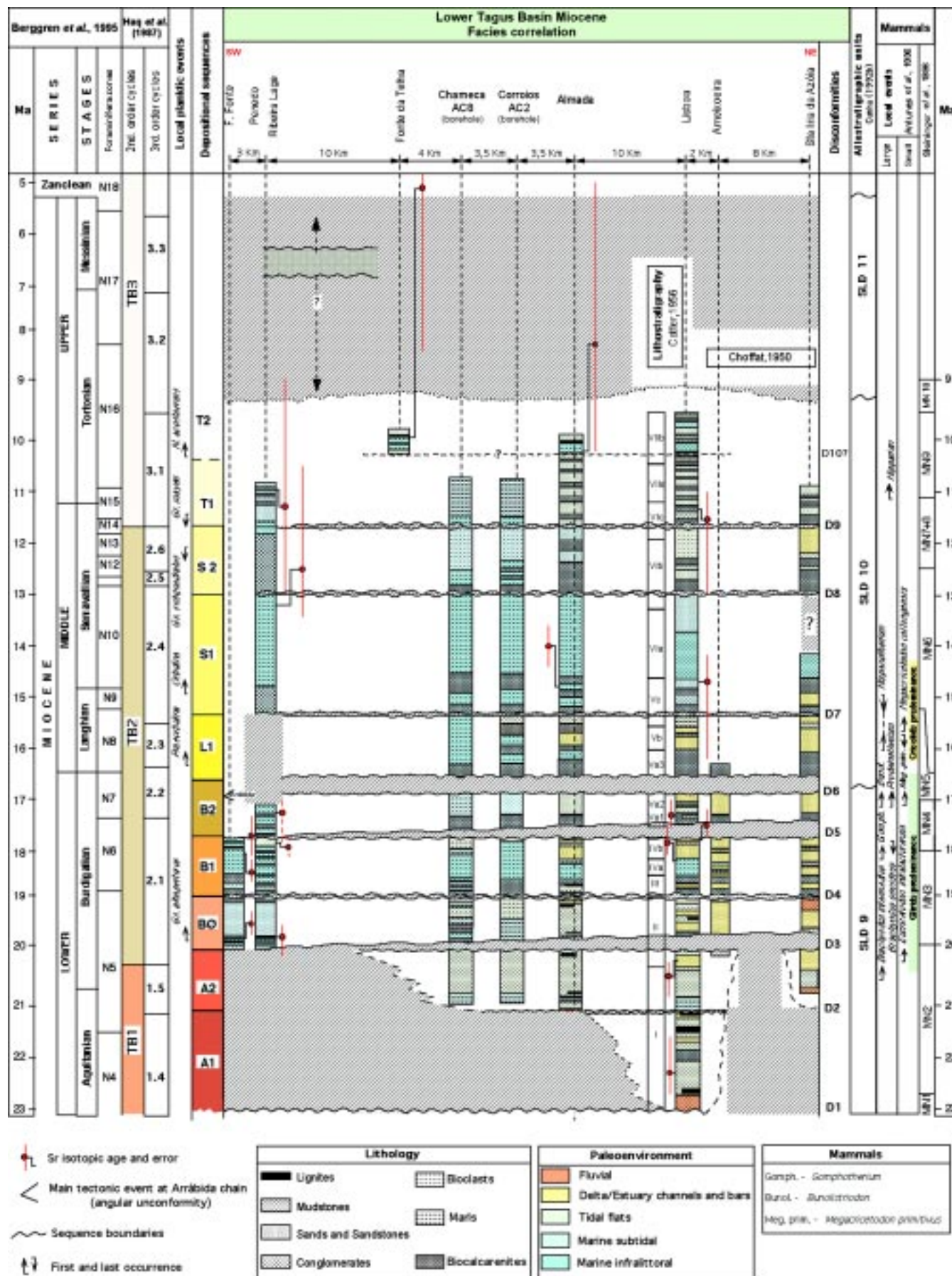
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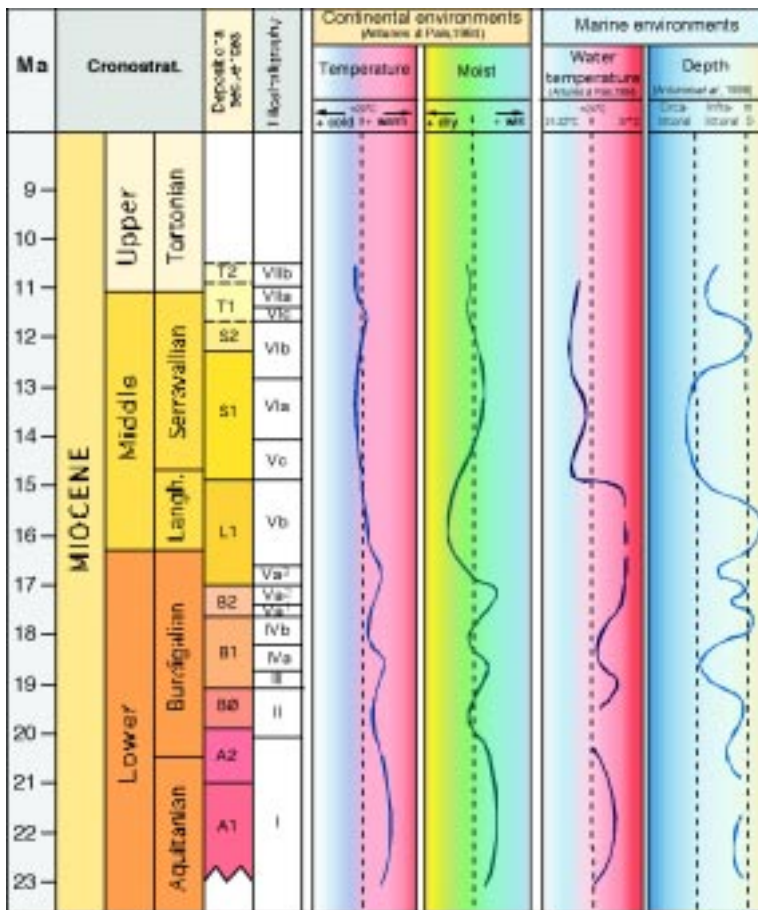
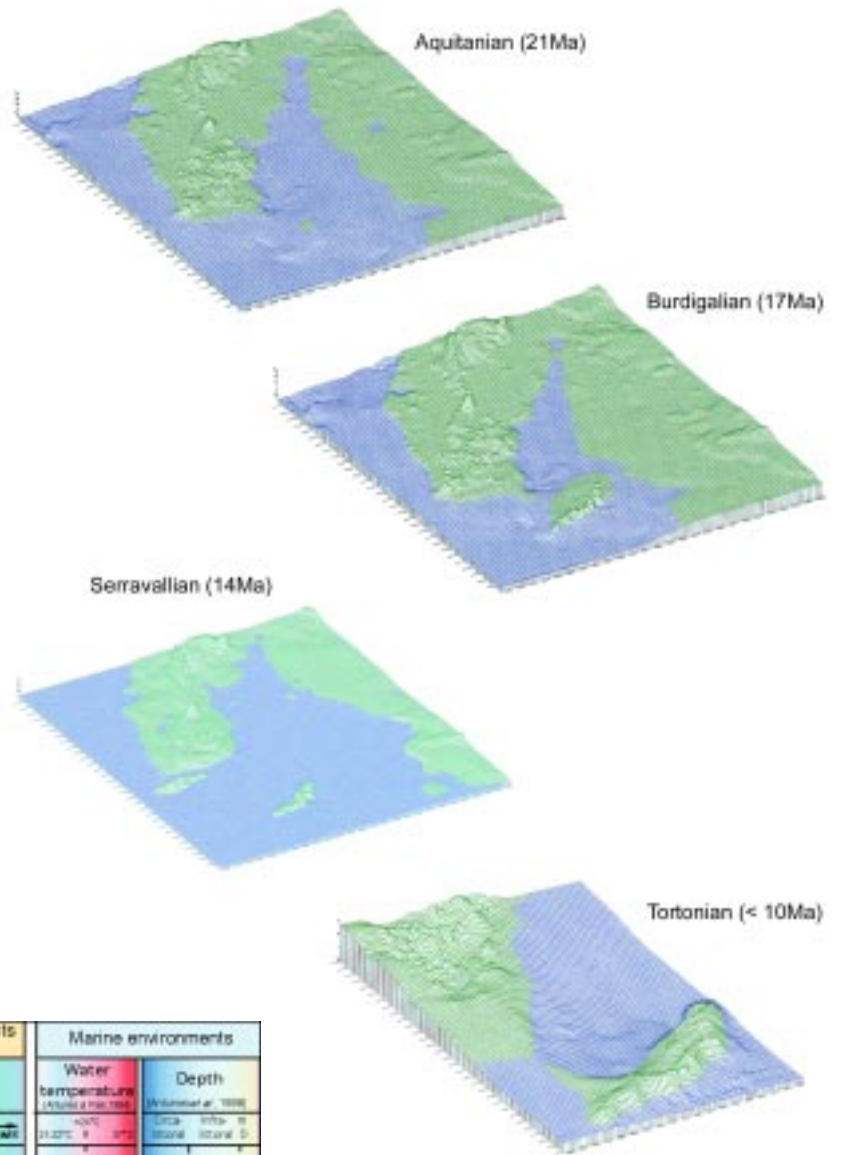
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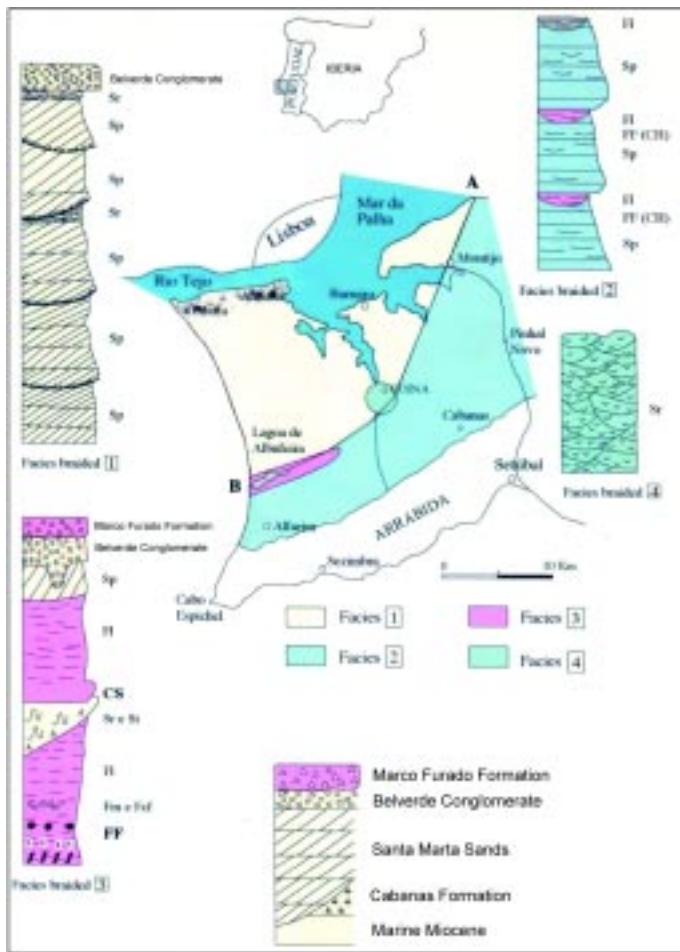


Pl. I - Stratigraphic framework for the Miocene of the distal part of the Lower Tagus Basin (Antunes *et al.*, 2000b).

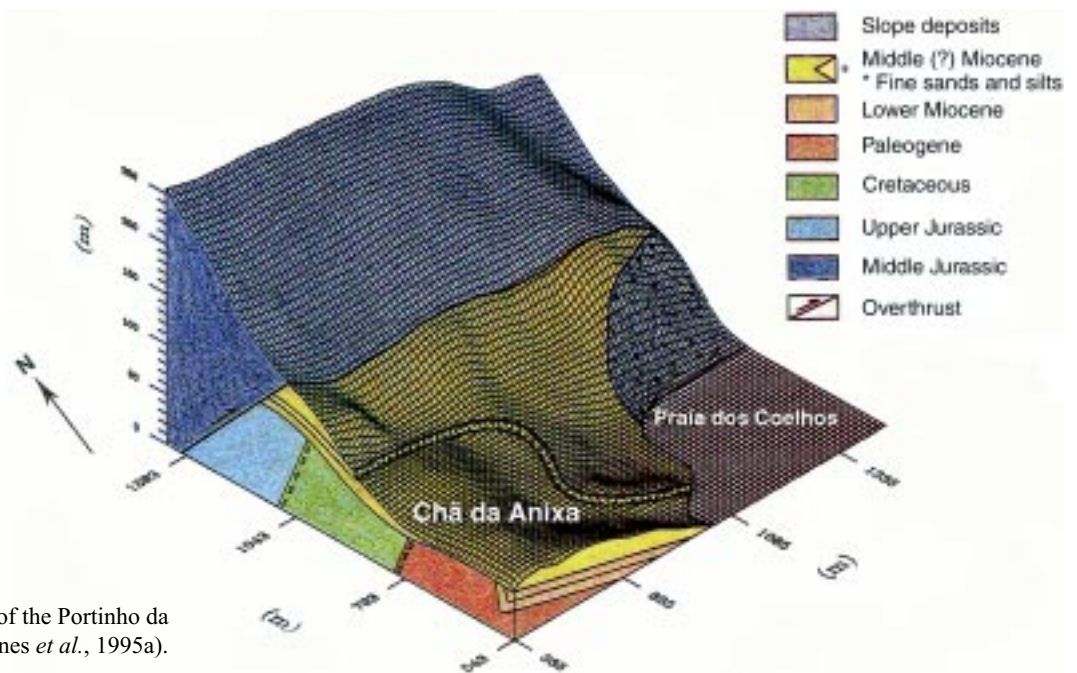
Pl. II.A - Palaeogeographic maps from the Miocene of the Lower Tagus Basin (Antunes *et al.*, 1999).



Pl. II.B - Evolution of the environments during the Miocene, in the distal part of the Lower Tagus Basin (Pais, 1999).



Pl. III.A - Main stratigraphic units and facies at the Setúbal Peninsula (Azevedo in Antunes *et al.*, 2000a).



Pl. III.B - Block diagram of the Portinho da Arrábida (Antunes *et al.*, 1995a).

Post-Meeting Field Trip P6

Palaeoseismites and structures related to karst evolution in the Algarve region

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Palaeoseismites and structures related to karst evolution in the Algarve region

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1. INTRODUCTION

The Algarve region, the southernmost area of the Portuguese mainland, is located close to the Azores-Gibraltar plate boundary, which extends from the Azores islands to the Gulf of Cadiz. It is also located near the West-Iberia continental margin, which probably is in a transitional state to a convergent plate boundary (Cabral, 1995; Ribeiro *et al.*, 1996; Ribeiro, 2002; Fig. 1).

This tectonic setting explains the regional seismicity and the neotectonic activity, evidenced by numerous earthquakes and by Pliocene to Pleistocene deformation (Dias & Cabral, 1995a, 2000b, 2002b; Dias, 2001).

The earthquake activity comprises important historical and instrumental events located at the northern edge of a wide belt of seismicity that extends approximately from the Gorringe submarine ridge (SW of Algarve) to the Straits of Gibraltar (Fig. 2). In fact, large historical earthquakes have occurred in the Atlantic SW of Algarve, as the 1755 “Lisbon earthquake” (estimated ML = 8.5), in an area that also experiences important instrumental seismicity (Buforn *et al.*, 1988; Zitellini *et al.*, 1999, 2001). Significant seismic activity also occurs near the littoral and onshore, including some historical events that have caused substantial damage, as in Portimão (1719, IMM max. IX), Tavira (1722, IMM max. X) and Loulé (1856, IMM max. VIII), and a

scattered low magnitude instrumental seismicity (Carrilho *et al.*, 1997).

The regional neotectonic activity – intended as the tectonic activity from the Upper Pliocene to the Present, is evidenced by vertical crustal movements at the regional scale (Dias & Cabral, 1999; Dias, 2001), and by structures at the outcrop scale affecting Pliocene-Pleistocene sediments (Fig. 3 and 4). These comprise brittle deformation structures (including several macroscale and mesoscale faults and a large number of joints), ductile deformation structures (represented by folds) and soft sediment deformation structures.

Vertical crustal movements are evidenced by the presence of an E-W elongated relief, 100 km long (the “Serra Algarvia”), which decomposes into two bulges: the Monchique mountain (903 m in the Foia) and the Caldeirão mountain (589 m in Pelados), separated by a NW-SE elongated fault controlled depression (S. Marcos-Quarteira fault). This relief evolved from an uplifted polygenetic erosion surface cut mainly on the Variscan basement, which was last retouched in the (Late?) Pliocene to Quaternary (Feio, 1951; Pimentel, 1989; Dias & Cabral, 1997, 2000b, 2002b; Dias, 2001).

Brittle deformation structures that have been observed affecting the Pliocene to Quaternary cover sediments are characterized by a great number of fractures at the mesoscopic scale. The faults have diversified orientations and tectonic styles: reverse, normal and strike-slip faults, although the reverse fault geometry is largely predominant (Fig. 5). In addition to these faults, an

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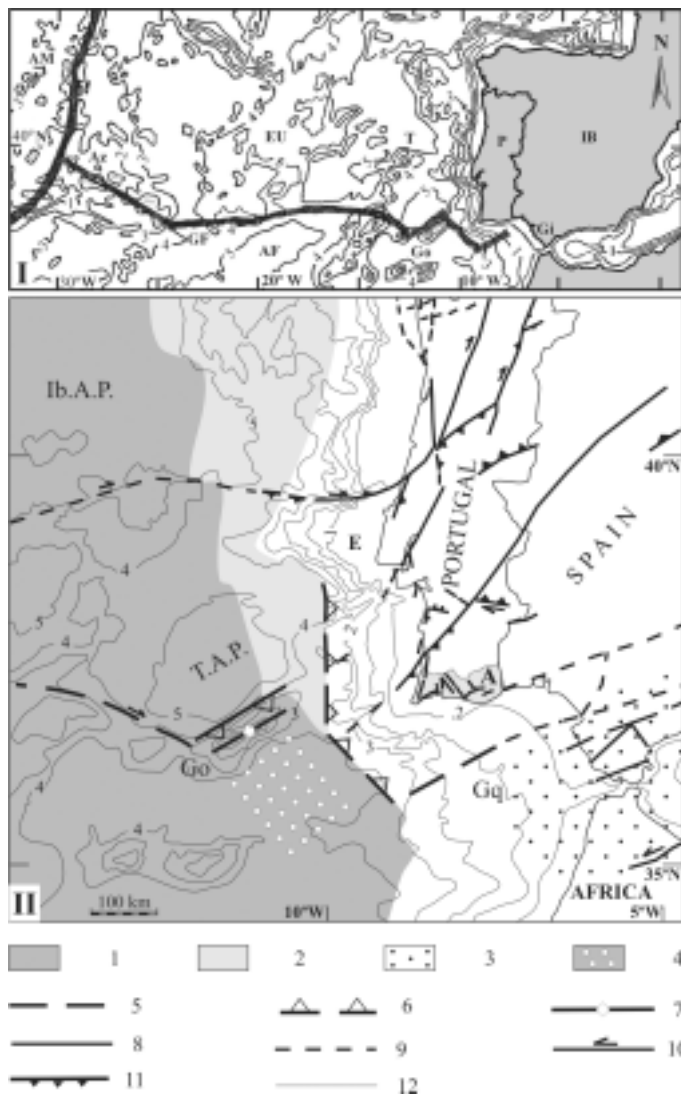


Fig. 1 - Regional geodynamic framework for the studied area (modified from Ribeiro *et al.*, 1996; adapted from Dias & Cabral, 2002a). I: regional lithosphere plates setting; II: eastern sector of the Eurasia-Africa plate boundary in the Atlantic Ocean (Azores-Gibraltar fracture zone). 1: oceanic crust; 2: thinned continental crust; 3: diffuse plate boundary; 4: zone of distributed plate deformation by buckling and thrusting; 5: plate boundary; 6: incipient subduction along the Southwestern Iberian continental margin; 7: active antiformal fold; 8: active fault; 9: probable active fault; 10: active fault with significant strike-slip movement; 11: reverse active fault; 12: bathymetric curve (in kilometres with first curve at 0.2 km); A: Algarve; AF: African plate; AM: American plate; Az: Azores islands; E: Estremadura high; EU: Eurasian plate; GF: Gloria fault; Gi: Gibraltar; Go: Goringe bank; Gq: Guadalquivir bank; IB: Iberian Peninsula; Ib.A.P.: Iberia Abyssal Plain; P: Portugal; T.A.P.: Tagus Abyssal Plain; T: Tore submarine mountain.

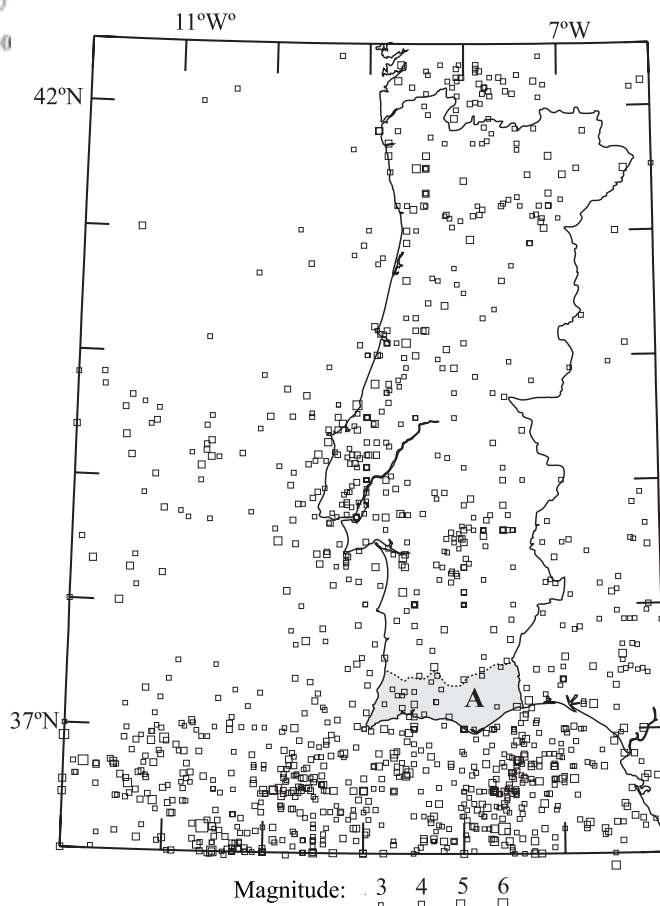


Fig. 2 - Epicentre distribution (period 63 B.C. to 1997 D.C.) of historical and instrumental earthquakes of magnitude ≥ 3 in the Portuguese Mainland and nearby areas, based on the earthquake catalogue by Martins & Mendes-Victor (1990) updated to December 1997 (Martins & Mendes-Victor, unpublished data). Symbol dimension is proportional to magnitude. Shaded area (A): Algarve region (Dias & Cabral, 2002a, b).

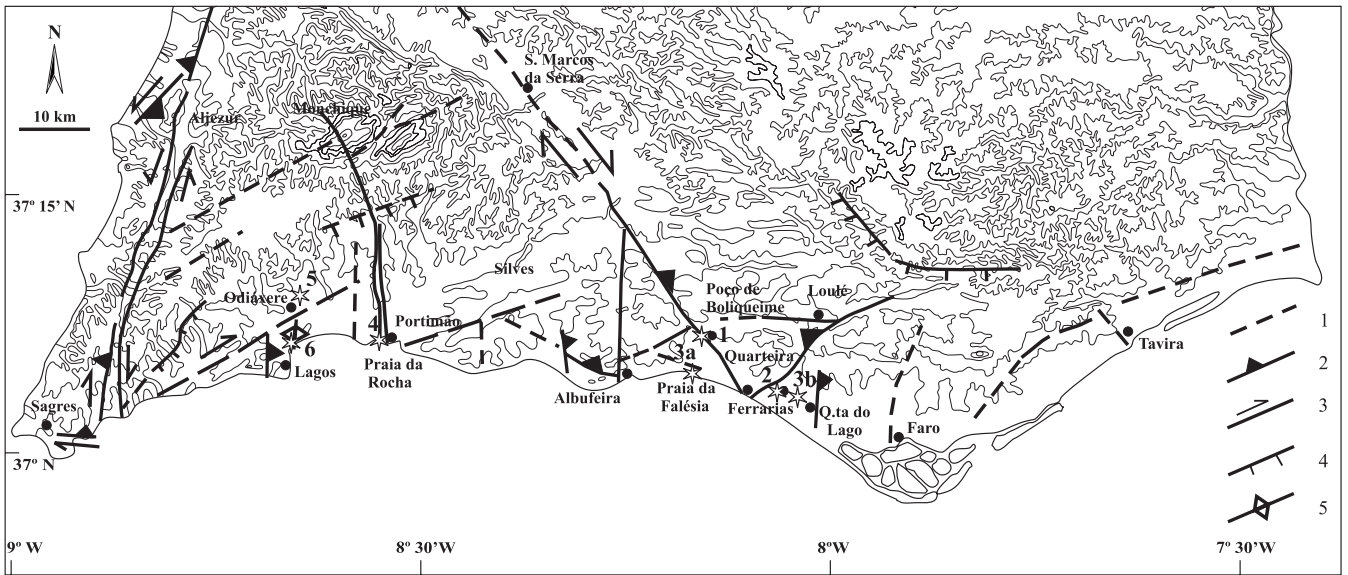


Fig. 3 - Map of the main active faults identified in the Algarve Region. 1: probable active fault; 2: reverse fault (teeth on hanging wall); 3: strike-slip fault; 4: fault with unknown dip (ticks on downthrown side); 5: anticline fold; stars: stops. Relief shown by contour lines every 100 m except the first at 50 m. (Dias, 2001; Dias and Cabral, 2002a, b).

intense fracturing of indeterminate kinematics (probably corresponding mostly to joints) was observed in distinct areas, showing a heterogeneous spatial distribution. These fractures are generally almost vertical, with a large dispersion in trend (Fig. 6).

Ductile deformation structures that have been observed affecting the Pliocene to Quaternary cover sediments are represented by folds at the mesoscopic scale (Fig. 7).

Probably many of these folds and fractures (faults and joints) result from the evolution of an underlying cryptokarst developed on Miocene and Mesozoic limestone basement rocks, mainly in the areas where

the fracturing shows larger scattering and smaller offsets (Dias & Cabral; 1995a, b, 1998a, b, 2002a; Dias, 2001). In fact, a variety of ductile, semi-brittle and brittle structures developed in the sediments that fill up the karst wells, controlled by different rheological behaviour of the Pliocene-Pleistocene deposits, various strain rates associated with sudden collapse or progressive sinking, and the variable shape of the karst pits walls (Dias & Cabral, 2002a).

Soft sediment deformation structures due to liquefaction resulting from palaeoearthquakes (palaeoseismites) have also been identified in several areas, consisting in

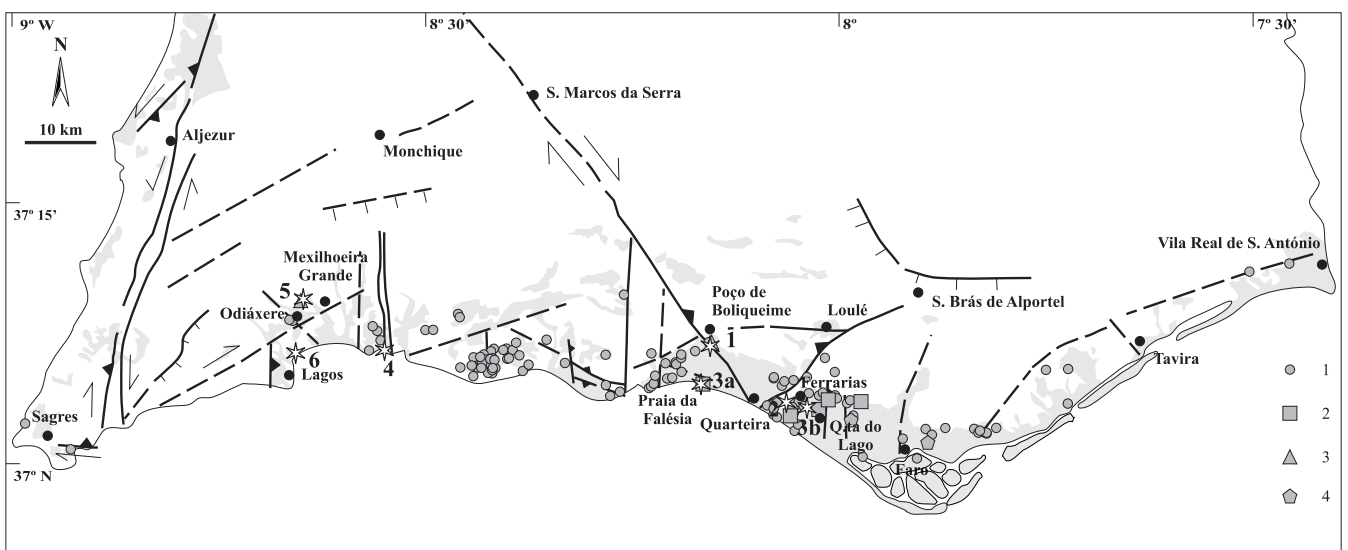


Fig. 4 - Map of the palaeoseismites identified in the Algarve Region affecting Pliocene to Quaternary sediments (in light grey). 1: intrusive detrital dike; 2: neptunian dike; 3: convolute folding; 4: flame structures; stars: stops (Dias, 2001; Dias & Cabral, 2002b).

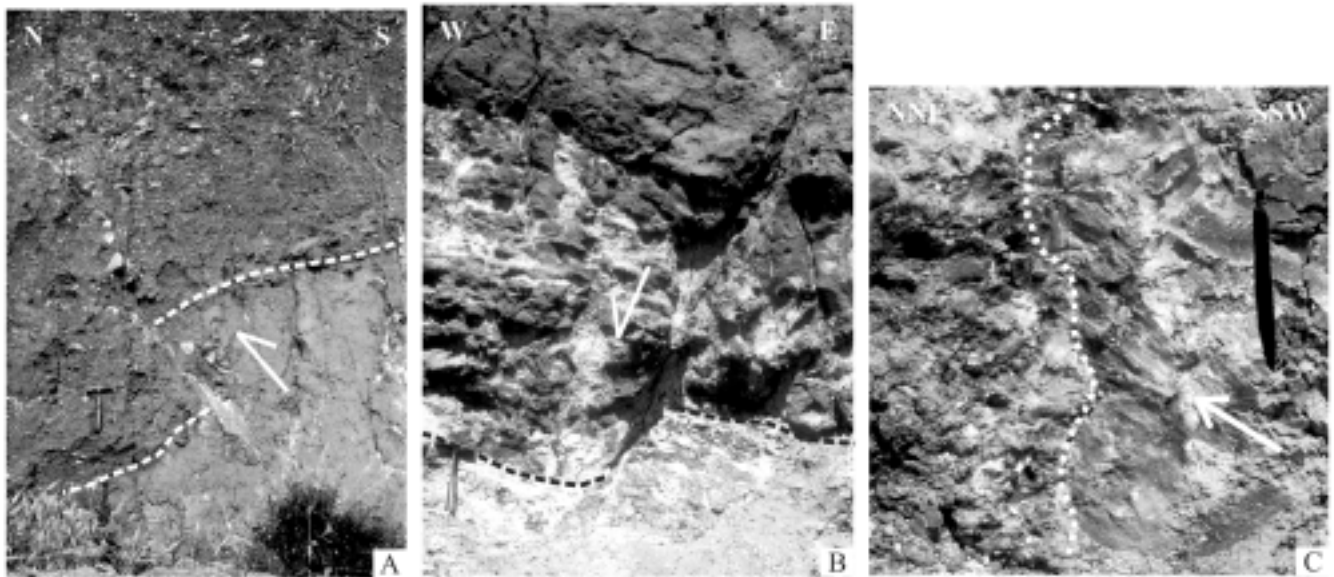


Fig. 5 - Examples of the faults affecting Pliocene to Quaternary sediments with diversified tectonic styles. A - Reverse fault (E of the Portimão). B - m Normal fault (W of Poço de Boliqueime). C - Strike-slip fault (Loulé).

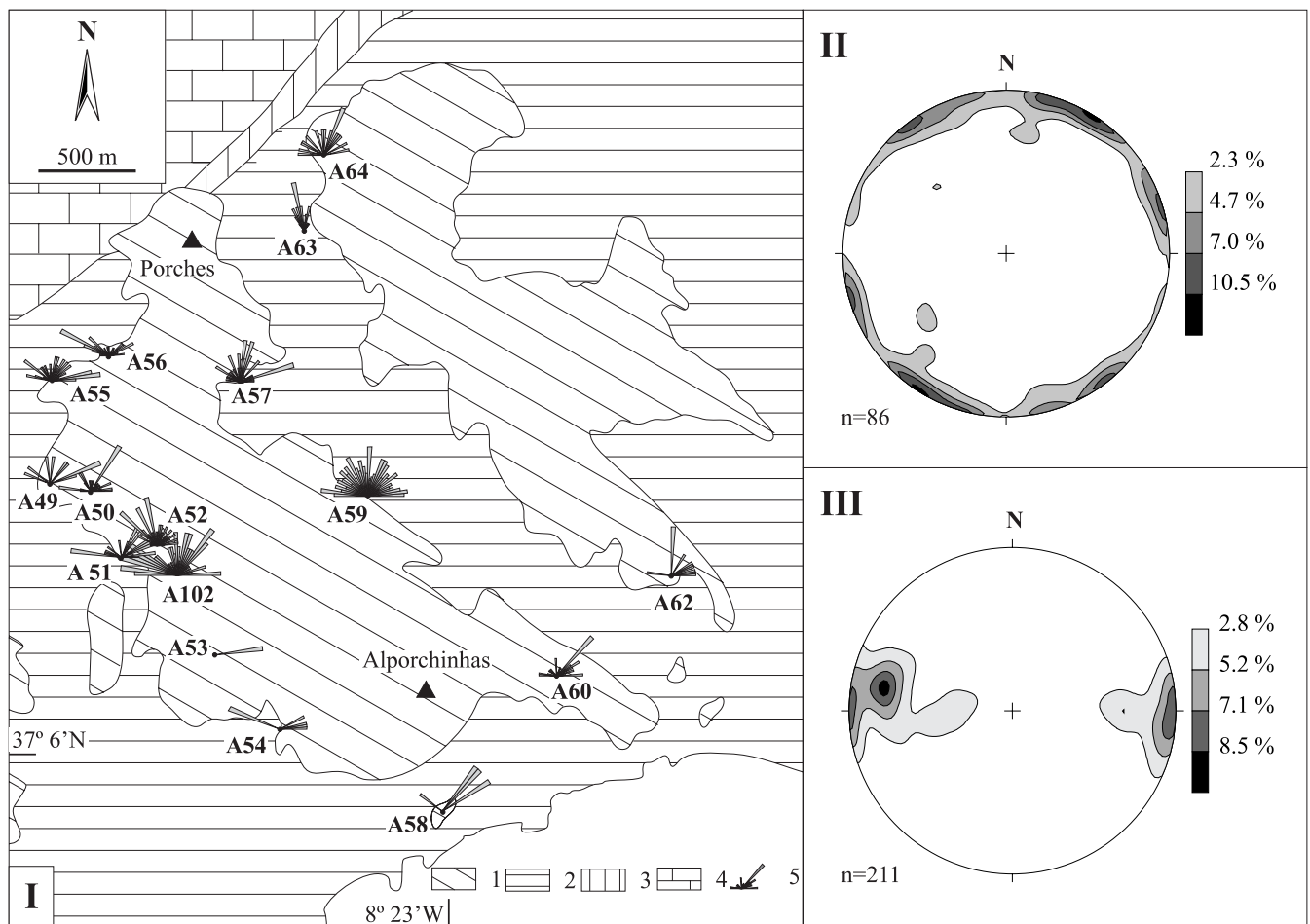


Fig. 6 - I - Synthetic geological map of the Porches area (adapted from Rocha *et al.*, 1981) showing the spatial distribution of fractures (rose diagrams) affecting Pliocene to Quaternary sediments. II and III - density contour diagrams for the poles to the planes of the fractures (Schmidt net, lower hemisphere) measured in the road Poço Partido - Salicos (S of Lagoa; II) and Praia do Forte Novo (near Quarteira; III). 1: Pliocene to Pleistocene sediments; 2: Miocene; 3: Cretaceous; 4: Jurassic; 5: directional frequency rose-diagram of measured fractures (5° intervals) (Dias, 2001; Dias & Cabral, 2002b).

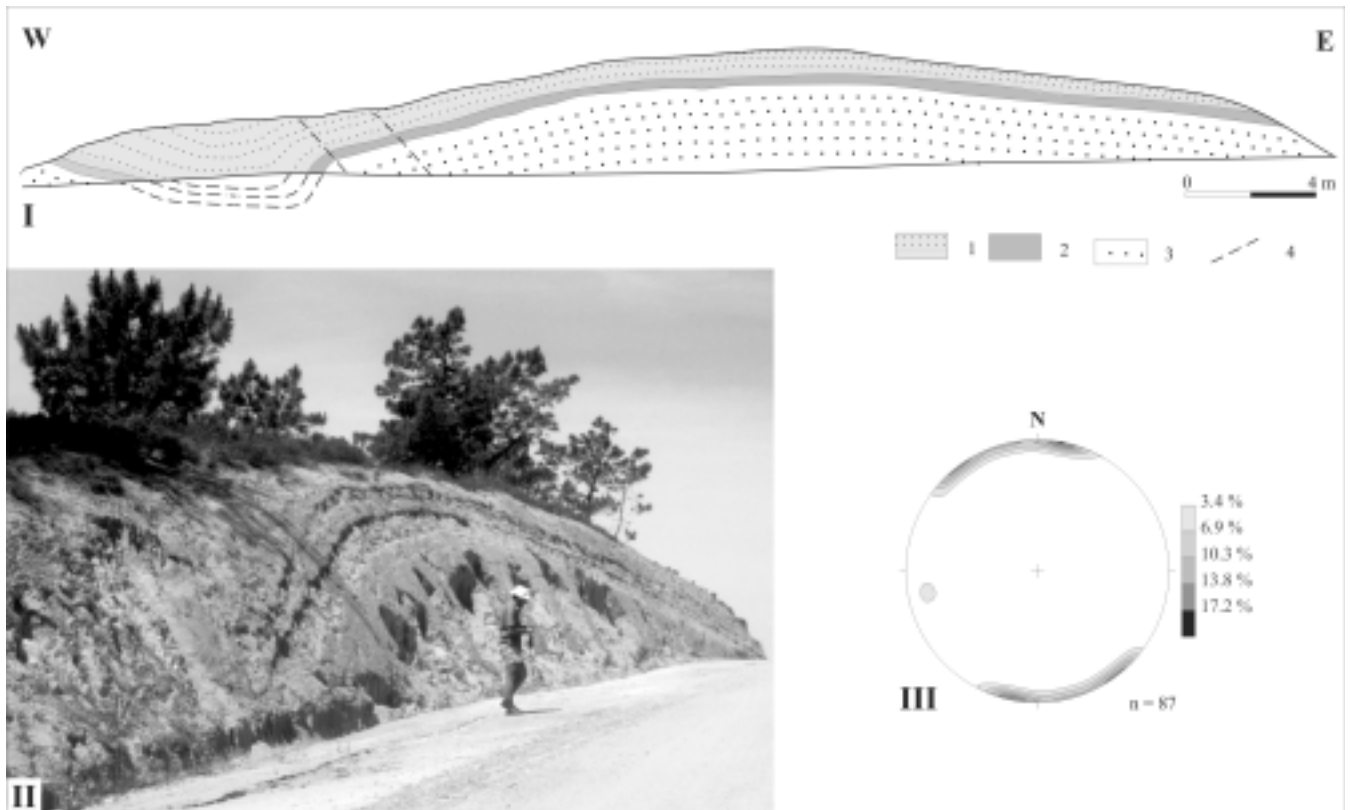


Fig. 7 - Folded Pliocene to Quaternary sediments outcropping at a road cut near Meia-Praia (E of Lagos). Fold axis trends approximately N20°-25°E. I - schematic cross section of the outcrop; 1: red sandstones; 2: yellow mudstones; 3: white sandstones; 4: fracture. II - General sight of the outcrop. III - Density contour diagram for the poles to the planes of fractures affecting the Pliocene to Quaternary sediments at the fold and nearby outcrops (Schmidt net, lower hemisphere) (Dias, 2001; Dias and Cabral, 2002b).

convolute folding of sandstone and conglomerate layers, fractures filled with collapsed sediments due to sudden opening (neptunian dikes), and detrital dikes (Rodríguez-Pascua, 1998; Dias & Cabral, 2000a; Dias, 2001). These structures may result from ground motions related to strong distant earthquakes, but the proximity of many of the seismites to known active faults suggests a relationship to moderate to high magnitude ($M > 5.5$) events generated by these faults, which is compatible with the known regional seismicity although some of these palaeoseismites point to a higher level of local earthquakes magnitude (Dias & Cabral, 2000a, Dias, 2001).

In this field trip we intend to show some aspects of tectonic structures, palaeoseismites and other structures related to karst evolution, that affect Pliocene to Quaternary sediments outcropping in the Algarve region. Accordingly, we will visit several outcrops to see the following structures: (Fig. 8):

- a) Faulting and convolute folding of Pleistocene sandstone and conglomerate layers, in Fonte de Boliqueime (Stop 1);

- b) detrital dikes in Pliocene-Pleistocene sandstones of the Ludo Formation, in Ferrarias (Stop 2);
- c) fractures filled with collapsed sediments due to sudden opening (neptunian dikes) in Pliocene-Pleistocene sandstones of the Ludo Formation, in Falésia beach and/or Quinta do Lago (Stop 3);
- d) convolute folding of conglomerate layers of the “Casalheira de Odiáxere” unit, of probable Pleistocene age, at Arão, near Odiáxere (Stop 5);
- e) folding of Pliocene-Quaternary sediments of the Ludo Formation at Meia-Praia beach, near Lagos (Stop 6).

We will also look at the sea cliffs in Praia da Rocha beach (Portimão, Stop 4) to view the excellent exposures of the karst morphology developed on the Miocene limestones, covered by deformed Upper Miocene and Pliocene-Quaternary sediments. Here we will see some impressive structures (folds and faults) affecting these sediments, which developed as a result of the evolution of the underlying cryptokarst on the Miocene limestones.

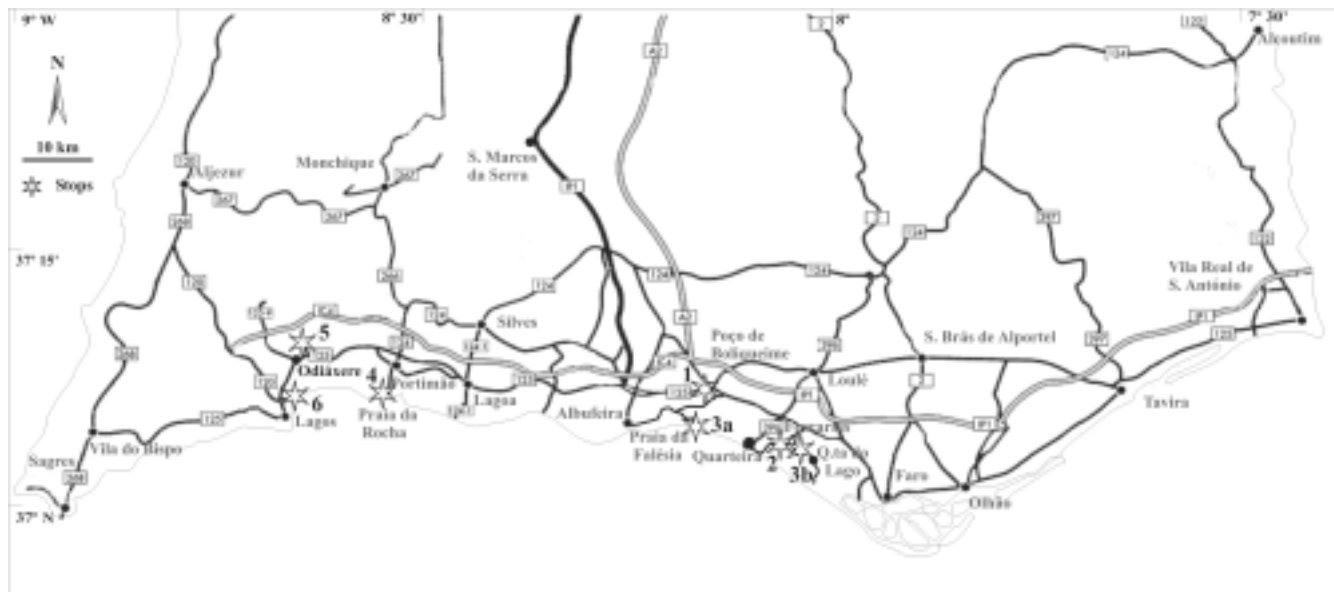


Fig. 8 - Location map of the stops (stars).

2. GEOLOGICAL SETTING

The Algarve region comprises Paleozoic basement rocks of Upper Carboniferous age, that outcrop in the northern area, and Mesozoic and Cenozoic rocks of two superposed sedimentary basins, in the South (Fig. 9).

The Paleozoic basement consists mainly of flysch sequences of slates and metagraywackes that were intensely folded and faulted during the Variscan orogeny, presently trending NW-SE and verging to the SW.

The Mesozoic rocks comprise mainly continental siliciclastic and marine carbonate sediments ranging in age from Late Triassic to Early Cretaceous, with a prevalence of Jurassic limestones and dolomites. These sedi-

ments were deposited in a basin developed in a transtensional regime related to the closing of the Tethys Sea and the opening of the Central and North Atlantic Ocean.

In the Upper Cretaceous there was the emplacement of an igneous intrusive hypabyssal syenitic massif that presently outcrops intruding the Paleozoic basement at Monchique, in northwestern Algarve. It shows an approximately elliptical shape in outcrop, elongated in the E-W direction, with 16 km in length and 6 km in width (Fig. 9). It was installed approximately 70 My ago (González-Clavijo & Valadares, 2003), when the Africa-Iberia relative movement changed from left-lateral transtensive to convergent (Srivastava *et al.*, 1990; Terrinha, 1998; González-Clavijo & Valadares, 2003).

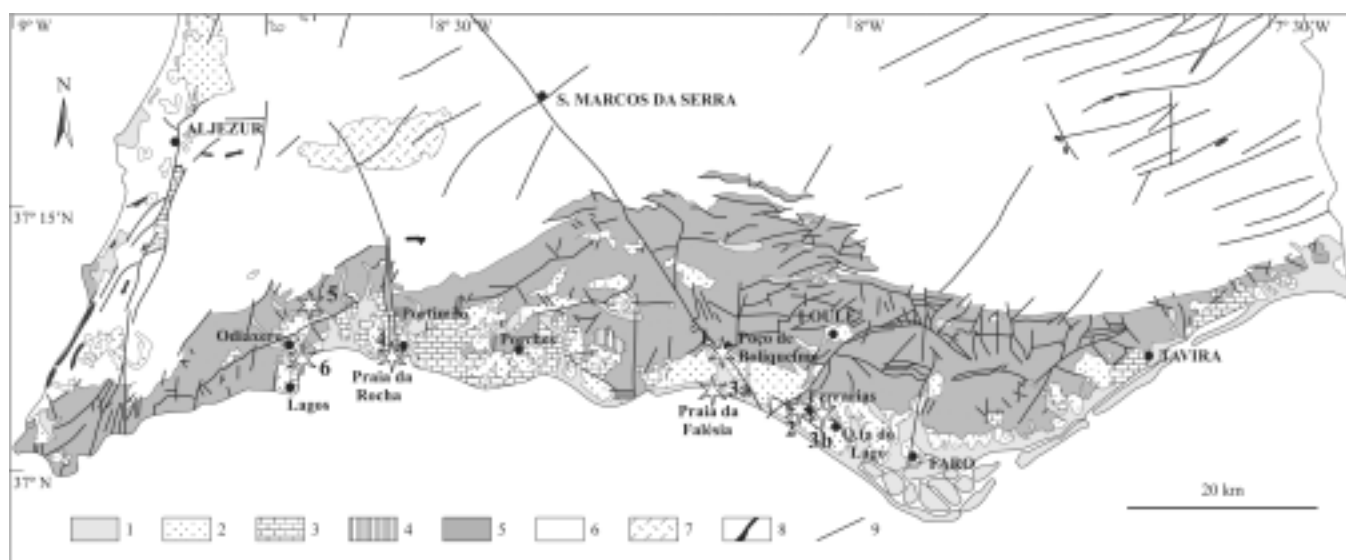


Fig. 9 - Schematic geological map of the Algarve region (adapted from Oliveira *et al.*, 1992); 1: Quaternary; 2: Pliocene-Pleistocene; 3: Miocene; 4: Paleogene; 5: Mesozoic; 6: Paleozoic; 7: Monchique intrusive massif (Late Cretaceous); 8: dyke; 9: fault.

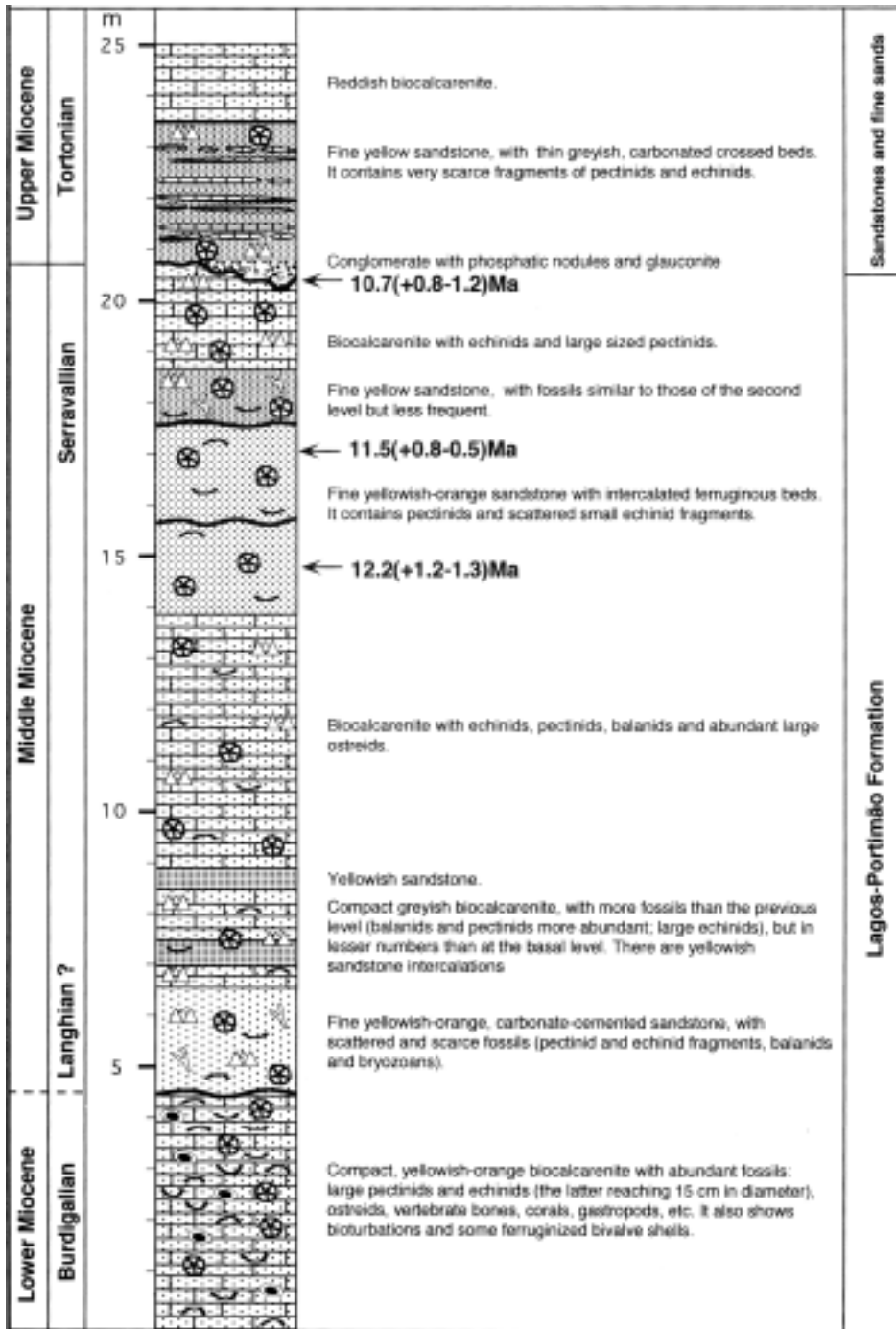


Fig. 10 - Praia da Rocha section (Pais *et al.*, 2000).

The Cenozoic basin was formed by flexural processes associated with the collision of Africa and Iberia (Terrinha, 1998, Terrinha *et al.*, 1998). The convergence between Africa and Eurasia since Early Tertiary produced a polyphasic tectonic inversion of the extensional Mesozoic basin, with an important post-Cretaceous and pre-Miocene inversion phase and another, less intense, Neogene to Quaternary inversion phase (Terrinha, 1998; Lopes & Cunha, 2000; Dias, 2001; Lopes, 2002).

The first inversion phase, post-Cretaceous and pre-Miocene, is testified by important compressive structures that are mostly sealed by the Miocene sediments or

that only slightly affect these sediments. This phase produced a regional crustal uplift along with intense sub-aerial erosion, explaining the absence of Paleogene sediments in the Algarve.

The Miocene sediments are present from W to E along the Algarve region although they are restricted to its southern area, extending to the continental shelf (Fig. 9). Onshore they comprise two formations: a lower one, consisting mainly of shallow marine sandy limestones of Early to Middle Miocene age (the Lagos-Portimão Formation) (Pais *et al.* 2000), which is unconformably overlain by shallow marine mudstones and sandstones

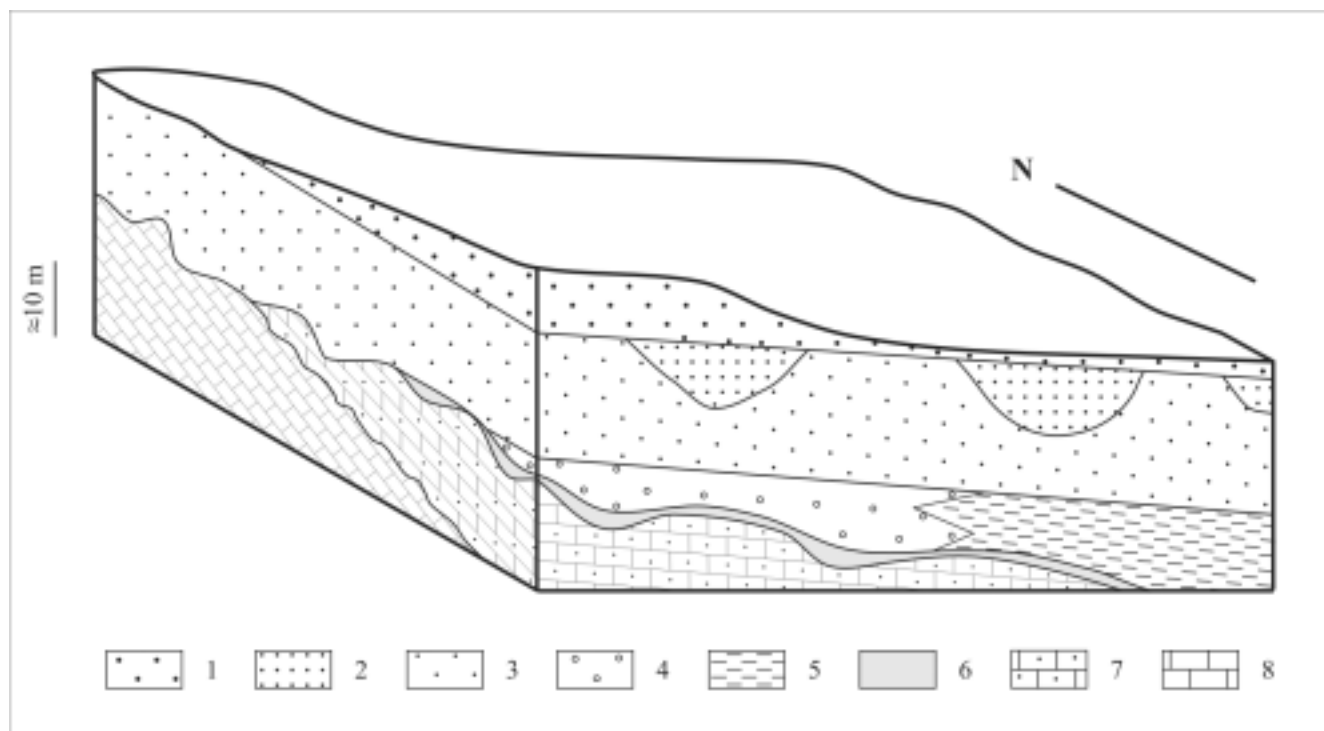


Fig. 11 - Schematic block diagram showing the vertical and lateral distribution of the Pliocene-Pleistocene Ludo Formation, and its relationship with the underlying Mesozoic and Miocene rocks (adapted from Moura & Boski, 1999). 1: Gambelas fluvial sandstones and conglomerates (Upper Pleistocene); 2: Ludo fluvial sandstones (Pleistocene ?); 3: Faro-Quarteira fluvial and marine sandstones (Upper Pliocene); 4: Falésia marine sandstones (Lower Pliocene); 5: Monte Negro marine fine sandstones (Lower Pliocene); 6: Cacela clayey mudstones (Miocene); 7: Lagos-Portimão sandy limestones (Miocene); 8: Mesozoic predominantly carbonate rocks (scale is approximate) (Dias & Cabral, 2002a).

of the Cacela Formation, of Late Miocene age (Pais, 1982; Antunes & Pais, 1988, 1992, 1993; Cachão, 1995; Cachão, *et al.* 1998; Pais *et al.* 2000; Fig. 10).

Fluvial-deltaic mudstones, sandstones and conglomerates of the Ludo Formation, of Pliocene to Pleistocene age (Moura & Boski, 1994, 1999; Moura, 1998; Moura *et al.*, 1998; Figs. 10 and 11), follow in continuity the “Cacela Formation”, or unconformably rest on the Paleozoic basement, the Mesozoic rocks or the Miocene sediments of the Lagos-Portimão Formation (Fig. 9).

The presence of Lower to Middle Miocene shallow marine sediments unconformably overlying the Mesozoic rocks, and even directly overlying the Paleozoic basement in certain locations in western Algarve, evidences a relative sea level rise that produced a marine transgression over a rather regular erosional morphology that extended “inland” beyond the borders of the Mesozoic basin.

This transgression was followed by a new period of emersion and erosion, during which the Miocene limestones were karstified, with the development of dissolution wells. This karst also evolved on the Mesozoic limestones where the lower Miocene unit was absent (by erosion or lack of deposition), and was later

filled by the Upper Miocene sediments or directly by the Pliocene to Quaternary sandstones and conglomerates of the Ludo Formation.

Tectonic activity was continuous after the Miocene. The Pliocene-Pleistocene Ludo Formation, and particularly one of its units, the widespread Faro-Quarteira Sands, were used as stratigraphical references to recognize and characterize the regional neotectonic activity.

The occurrence of Pliocene to Quaternary deformation is mainly testified by structures at the outcrop scale, consisting of a large number of fractures (faults and joints), some soft sediment deformation structures (such as collapse structures and convolute bedding) and a few folds, affecting the younger sedimentary units, mostly the “Faro-Quarteira Sands”. These meso-scale structures are often scattered through large areas and are not easily related to major tectonic features in the underlying rocks.

The predominance of variably trending active faults showing reverse movement component suggests the action of a compressive stress field that produced a regional constrictive finite strain roughly in the last 2 My, although some (short?) extensional (normal and strike-slip faulting) episodes were recognized in some locations (Dias & Cabral, 1995a, 2002b; Dias, 2001).

3. STOPS

In this Field Trip we will visit several outcrops in the Algarve region to see examples of deformation structures due to soft sediment liquefaction resulting from palaeo-seismic activity, as collapse and fluid ejection structures, and structures (folds and faults) related to karst evolution, affecting Pliocene and Quaternary sediments (Fig. 8).

3.1. Fonte de Boliqueime

One of the most interesting fault outcrops can be seen at the N-125 road-cut near Fonte de Boliqueime (km 75.9), ENE of Albufeira, where Pleistocene sediments are intensely deformed, as shown schematically (Pl. I).

The deformed sediments correspond to the “Areias de Boliqueime” unit, with a thickness of approximately 15m, comprising sandstones, pelitic and lacustrine limestone layers, with interbedded conglomerates. These

sediments probably correspond to a lacustrine palaeo-environment, and have been dated as Pleistocene by their content of several ostracode species (Cabral *et al.*, 2003).

Besides being affected by faults, the detrital deposits are strongly folded, apparently by fault drag, and in certain areas show convolute and even chaotic bedding suggesting liquefaction and fluidization of water saturated sediments, which produced fluid escape structures, probably during Magnitude >7 earthquakes (Dias, 2001).

Despite the absence of clear kinematic indicators, the geometry of the main faulted contact (Pl. I: F1, N55°W, 85°SW) suggests the occurrence of strike-slip movement in a transpressive regime. The presence of several low angle reverse faults immediately to the Northeast (Fig. 12) may be explained by strain partitioning.

Attending to the location and strike, these meso-scale faults apparently are correlative to one of the major macrostructures in Algarve I the S. Marcos-Quarteira fault zone (Manuppella *et al.*, 1986; Manuppella, 1988,

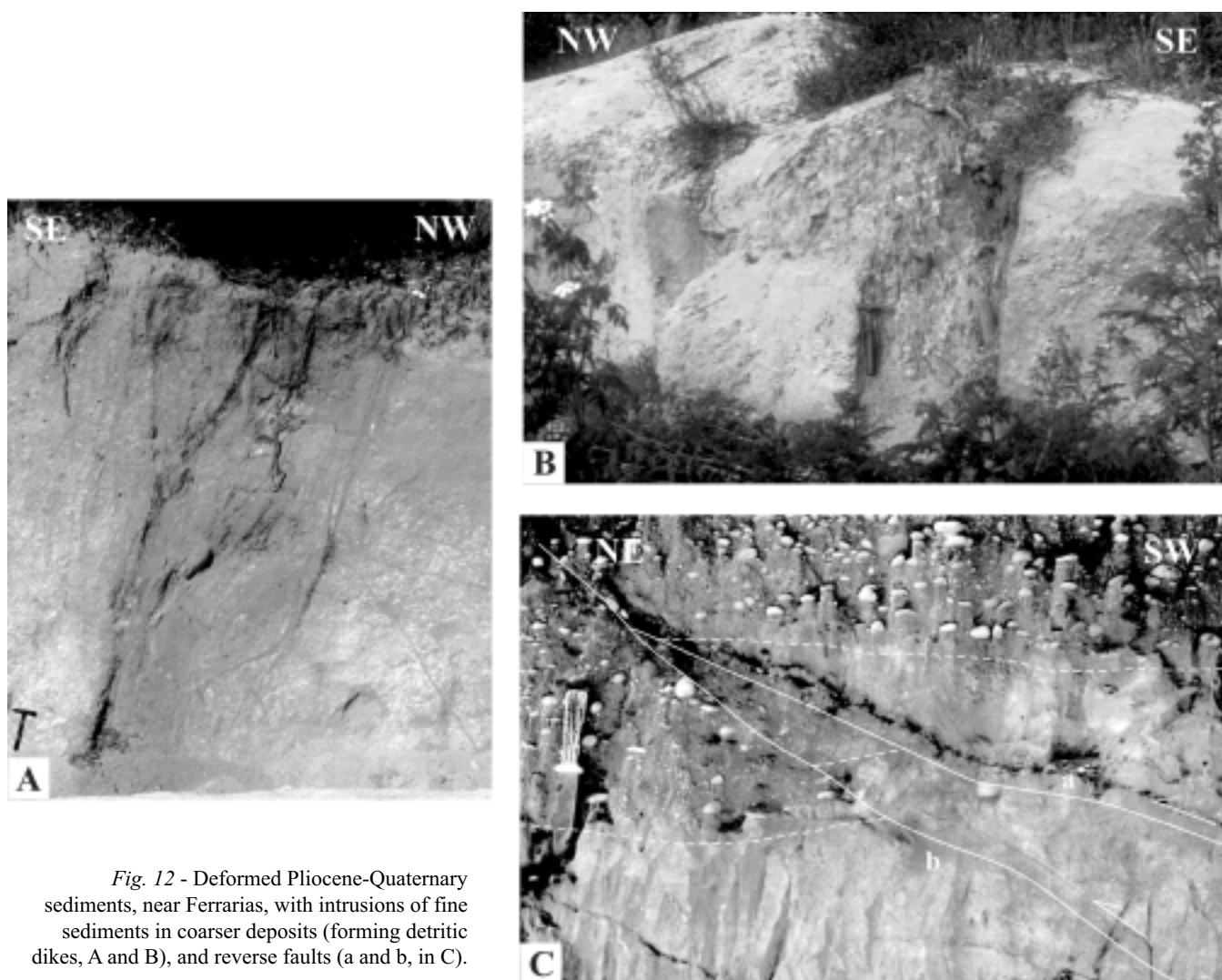


Fig. 12 - Deformed Pliocene-Quaternary sediments, near Ferrarias, with intrusions of fine sediments in coarser deposits (forming detritic dikes, A and B), and reverse faults (a and b, in C).

1992; Kullberg *et al.*, 1992; Terrinha, 1998; Dias & Cabral, 2002b, 2002; Dias, 2001), possibly reflecting local compression on a restraining bend of a NW-SE trending right-lateral strike-slip fault.

The S. Marcos-Quarteira fault (Fig. 3) is a NW-SE trending structure that extends from S. Marcos da Serra, at the N, to Quarteira, at the S, along approximately 40 km, probably extending to the immersed area.

It is a major regional tectonic structure inherited from the Variscan Orogeny, that controlled the evolution of the Algarve transtensional basin during the Mesozoic, separating it in two blocks that evolved differently. It was reactivated during Cenozoic tectonic phases and, since the Neogene, it controlled the evolution of the regional landscape, namely of the Serra de Monchique and Serra do Caldeirão mountains, located West and East of the fault, respectively (Choffat, 1907; Feio, 1951; Manuppella, 1988; Dias & Cabral, 1995b, 1999, 2002b; Dias, 2001).

3.2. Ferrarias

Ferrarias (near Quarteira; Fig. 8) is located near the eastern branch of the NE-SW trending Carcavai fault (Fig. 3). Fault activity in this area is indicated by a zone of disturbed sandstones of the Areias de Quarteira formation, which shows several intrusions of intensely deformed silt and clay.

The Areias de Quarteira formation, of Late Pliocene age, is composed of siliciclastic redish sandstones and conglomerates (Moura & Boski, 1994, 1999; Moura, 1998; Moura *et al.*, 1998).

Intrusions of fine sediments in coarser deposits (Areias de Quarteira), forming detritic dikes, have been identified in some areas of the Algarve region (Fig. 4), most frequently in the neighborhood of active faults. The presence of these detrital dikes is not conclusive as a palaeoseismicity evidence, because they can result from the infilling of opened fractures by sediments (Keller & Pinter, 1996). However, the downwards increase in thickness of the detritic dykes and the presence of striated surfaces in the dyke walls, suggest that the fine materials were intruded through fractures, after a process of liquefaction, associated to seismic events (Dias & Cabral, 2000a; Dias, 2001) of moderate to large magnitude ($M > 5.5$, according to Rodríguez-Pascua, 1998).

At this stop, we can see some detritic dikes, made up of silt, clay, and, sometimes, including material of the surrounding sediments (Areias de Quarteira) (Fig. 12A and B), as well as some reverse faults (Fig. 12C).

3.3. Quarteira

In the Quarteira region (Falésia beach and Quinta do Lago), in both sides of the S. Marcos-Quarteira fault, there are fractures filled with collapsed sediments due to sudden opening (neptunian dikes; Fig. 4).

The neptunian dikes have been observed in the zone between Olhos de Água and Faro, crossed by important active faults, as the S. Marcos-Quarteira, Carcavai and Almancil faults (Fig. 4; Dias & Cabral, 1999; Dias, 2001).

These structures, with a wedge like geometry in cross section, are interpreted as the result of the gravitational collapse of surrounding water saturated sediments towards the inner part of a fracture, that suffered a sudden opening during a seismic event, generated by near faults. The collapsed deposits are close to vertical near the fracture wall and horizontal in the centre (*op. cit.*).

3.3.1. Falésia beach

Along the road to the Falésia beach we can observe several aspects of neptunian dikes in sediments of the Areias de Quarteira formation. The fall of sediments into fractures is evidenced by the geometry of conglomeratic layers (Fig. 13): close to the fracture walls the conglomerates are nearly vertical while in the central area the layers are horizontal.

3.3.2. Quinta do Lago

At Quinta do Lago (400 m NW of the geodesic vertex) some neptunian dikes similar to those seen on stop 3.3.1 outcrop, affecting the Areias de Quarteira sediments. However, in this location, the gravels inside the fractures are not layered. Several pebbles are disposed along the fractures (Fig. 14).

3.4. Praia da Rocha

In the Praia da Rocha beach (near Portimão) we can see, along the cliffs, the excellent exposures of the karst morphology developed on the Miocene limestones, filled in by deformed Upper Miocene (Cacela Formation) and Pliocene-Quaternary sediments (Ludo Formation). These sediments are affected by several structures (folds and faults) developed as a result of the evolution of the underlying cryptokarst on the Miocene limestone rocks (Dias & Cabral, 1998a, b, 2002a; Dias, 2001).

Some mesoscopic folds, small-scale faults and intense shearing affecting the Miocene sediments are exposed in the Praia da Rocha cliffs (Terrinha, 1998; Terrinha *et al.*, 1999, 2003; Dias, 2001) interpreted as

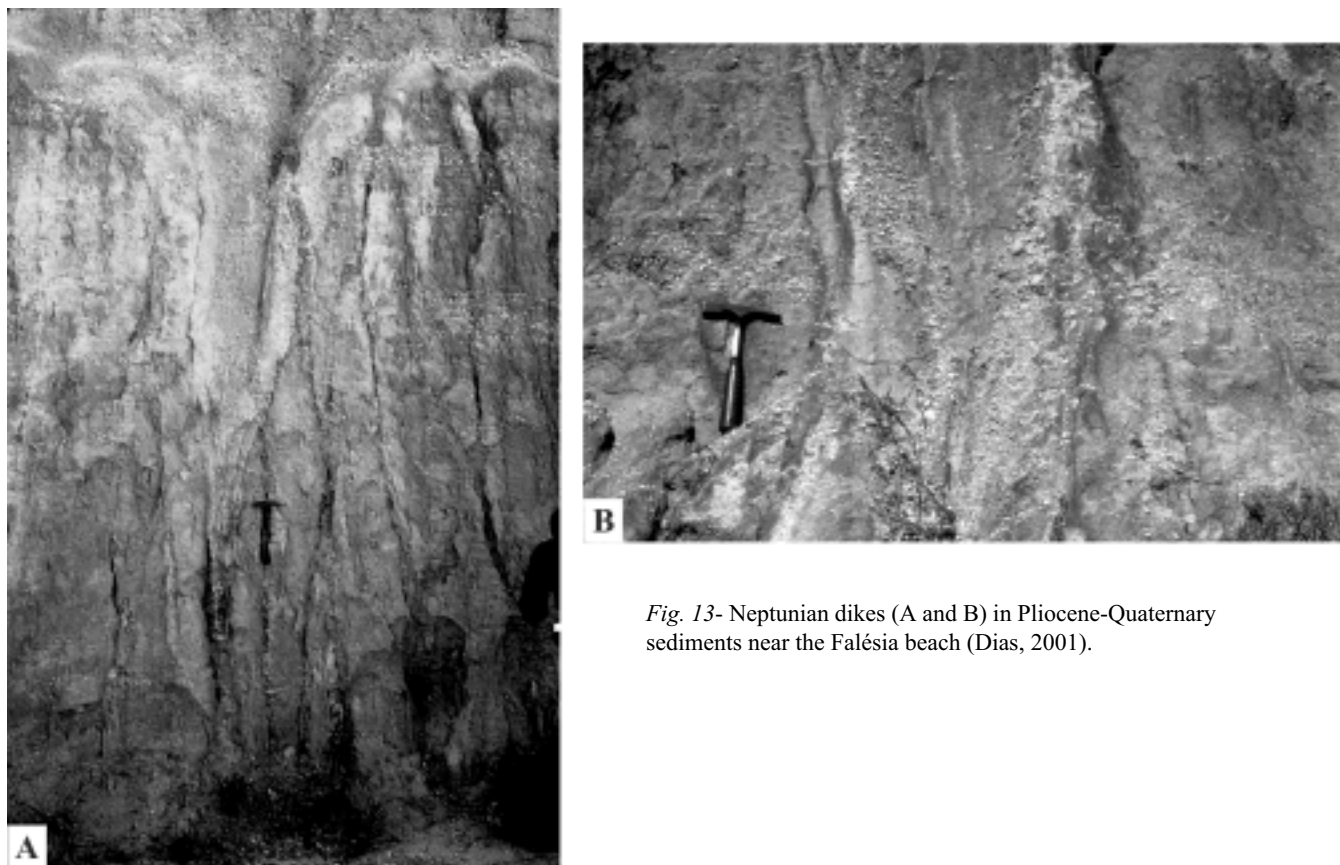


Fig. 13- Neptunian dikes (A and B) in Pliocene-Quaternary sediments near the Falésia beach (Dias, 2001).

the result of the reactivation of the Portimão Fault (*op cit.*).

The Miocene infill includes the Lagos-Portimão Formation, of Burdigallian-Serravallian age (Pais *et al.*, 2000), and the Praia do Castelo Formation (Sandstones and fine sandstones in Pais *et al.*, 2000), of Tortonian age (J. Pais, oral commun.; Fig. 10).

The Lagos-Portimão Formation consists of yellow or pink massive and very fossiliferous shelf biocalcarenes. The unit is bounded at the top by an erosion surface, overlaid by laminated sandstones, scarcely fossiliferous - the Praia do Castelo Formation (Fig. 10; Pais *et al.*, 2000).

In general, the Cacela Formation comprises mudstones and sandstones, while the Ludo Formation includes sandstones and conglomerates (Fig. 11). Both overlie the Middle Miocene Lagos-Portimão Formation limestones, usually through a very irregular karst surface.

This covered karst is well exposed along the coastal cliffs, where the overlying sediments can be seen resting on a rugged corrosion surface and filling deep (> 10 m) solution wells developed on the underlying limestones (Fig. 15; Dias & Cabral, 2002a).

Although some sedimentary filling of previous erosional irregularities occurred as the underlying

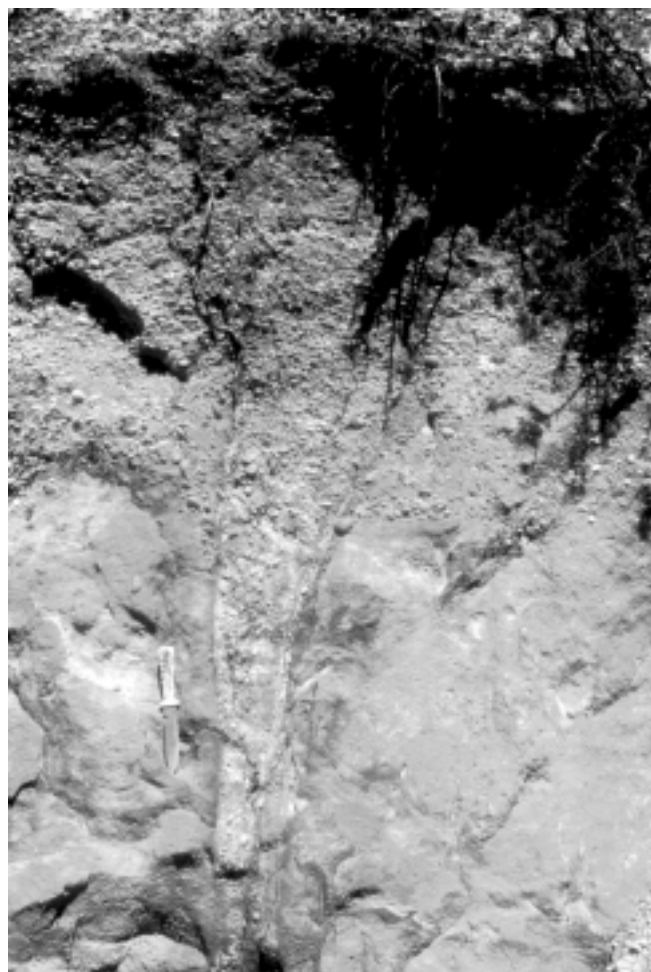


Fig. 14 - Neptunian dikes in Pliocene - Quaternary sediments near Quinta do Lago (Dias, 2001).



Fig. 15 - View of the karst morphology, at Praia da Rocha, developed on Miocene limestones, infilled with deformed Upper Miocene and Pliocene-Quaternary sediments, already exhumed in part by the ongoing erosion (Dias & Cabral, 2002a).

Miocene carbonate rocks were covered by terrigenous deposits, the continued underground evolution of a buried karst morphology is evidenced by the strong deformation that usually affects the sediments inside the karst wells (*op. cit.*).

Limestone solution was triggered, or accelerated, by groundwater lowering related to a relative lowering of the general base level (mean sea level), which probably resulted mostly from regional uplifting during the Quaternary. The fracturing associated to the Portimão Fault created good conditions for the development of water conduits, promoting solution (*op. cit.*).

During the underground evolution of the karst, the overlying sediments were deformed, creating a variety of ductile, semi-brittle and brittle structures (folds, joints and faults), mainly through mechanisms of heterogeneous simple shear apparently controlled by (*op. cit.*):

- different rheological behavior of the cover deposits due to their variable composition, moisture content, or compaction;
- various strain rates associated with sudden collapse or progressive sinking;
- variable geometry and trend of the karst pits walls, that frequently are very steep or even overhanging the sedimentary fill.

Folds were generated by subsidence of the cover sediments due to a progressive amplification, or sudden collapse, of karst wells and caves located in the underlying Miocene (Fig. 16; Dias & Cabral, 1998a, b, 2002a; Dias, 2001). Folds generated in this way are of the bending type, resembling drape folds, and tend to

have a bowl-shaped geometry, separated by areas where the sediments may present a dome-shaped structure. Folds with extremely thinned limbs developed possibly as a result of the sudden collapse of karst cavities (Fig. 16; Dias & Cabral, 2002a).

The fractures usually show high dips and present a large dispersion in trend. Generally, the faults affecting the cover karst deposits were seen to root at a residual clayey veneer or at a thinly laminated clayey silt that sometimes line the karst cavities. These materials accommodate displacements into the evolving karst cavities due to the progressive lowering of the limy cavity floor by solution. The fracturing on the cover sediments that is triggered by the solution of underlying limestones often presents dip-slip normal fault geometry. However, many fractures with reverse fault geometry were also observed, that may be explained by several mechanisms as: fault plane rotation by the continuing sediments subsidence, imposed geometric control by upwards fault propagation from nearly vertical to overhanging cavity walls, extensional faulting of thinned sedimentary layers that line sub-vertical cavity walls, and sediment failure in response to stretching induced by the lowering of the cavity floor (Dias & Cabral, 1998a, b, 2002a; Dias, 2001) (Fig. 17).

3.5. Arão

In Eiras Velhas, on the road to Arão (Odiáxere), near the Espiche-Odiáxere Fault, we can see convolute folding of conglomerate layers, as well as some fractures, affecting sediments of the “Casalheira de Odiáxere”

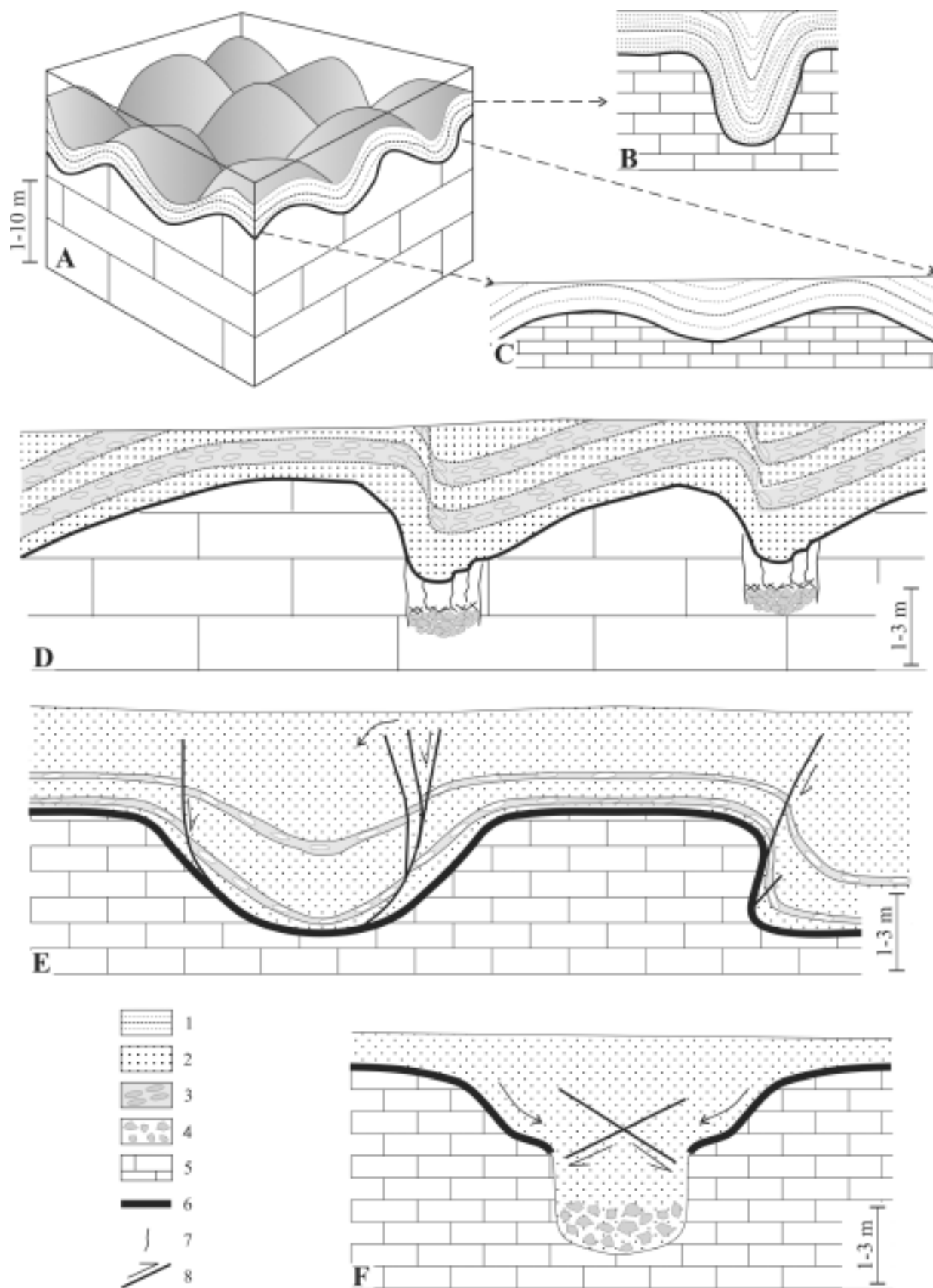


Fig. 16 - Schematic cross-sections showing different processes for generating karst related folding and fractures with reverse fault geometry in Pliocene-Pleistocene sediments (scale is approximate). A: 3D bowl and dome geometry; B: bending fold geometry showing limb thinning, related to the progressive deepening of a karst cavity; C: 2D succession of antiformal and synformal folds produced by moderate karst evolution; D: folds with intensely thinned limbs, related to sudden collapse of underlying caves; E: fault development associated with progressive enlargement of karst cavities; F: faulting in response to sudden cavity collapse. 1: reference bedding surface in sandstones; 2: sandy layer; 3: conglomerate layer; 4: collapse breccia; 5: limestone basement (Mesozoic or Miocene); 6: residual clayey layer; 7: fracture; 8: shear fracture with relative displacement indicated by arrow (adapted from Dias & Cabral, 2002a).

unit, of probable Pleistocene age (Fig. 17A) (Dias & Cabral, 2000; Dias, 2001). These sediments comprise:

- orange coloured mudstones, with sparse grains of quartz and metagreywacke pebbles;
- red coloured coarse sediments containing sub-rounded pebbles of quartz, metagreywacke and quartzite (maximal diameter around 40cm), with a sparse fine sand and silt matrix.

The fine-grained layers are fractured. The fractures trend nearly E-W. Some slip surfaces show radial striations (Dias, 2001).

Coarse layers are folded, showing a very irregular contact geometry with the fine-grained layers. Some pebbles are fractured (Fig. 17B), suggesting hydraulic fracturing, and the convolute folding is probably related with liquefaction and fluidisation during an earthquake of high magnitude ($M \geq 7.5$) (Rodríguez-Pascua, 1998; Dias & Cabral, 2000; Dias, 2001).

3.6. Meia-Praia beach

At a cut of the Meia-Praia to Albardeira road, near the Meia-Praia beach (East of Lagos), gently folded Pliocene-Quaternary sediments of the Faro-Quarteira Sands

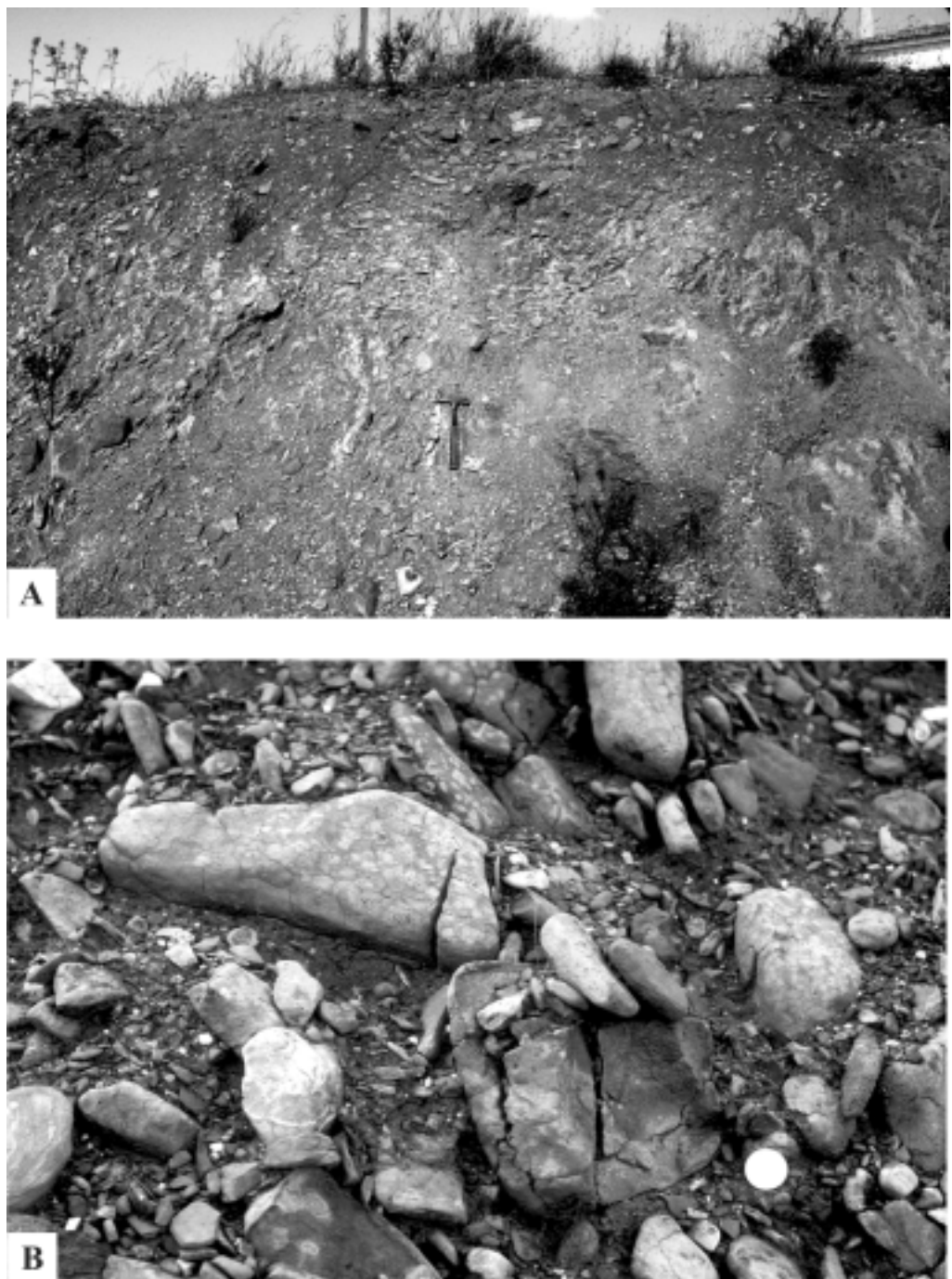


Fig. 17 - Outcrop near Arão, West of Mexilhoeira Grande, showing intense folding affecting coarse sediments (“Cascalheira de Odiáxere”);

A: probably related with liquefaction and fluidization during an earthquake; B: some fractured pebbles, suggesting hydraulic fracturing.

outcrop, with some sparse fracturing (Fig. 7; Dias & Cabral, 1995, 2001; Dias, 2001). The sediments comprise layers of:

- whitish fine sandstone with alternating medium to coarse sand layers, composed of quartz grains, with some feldspar, mica and dark mineral grains, within a silty matrix.
- yellowish mudstone with scarce grains of quartz;
- fine to coarse reddish sandstone with intercalated pebbly layers. The sandstone is made of quartz grains, with some feldspar and mica grains. The coarse layers contain sub-rounded to rounded pebbles of quartz and metagreywacke. These sediments show some oblique lamination. A layer of manganese concentration occurs sometimes at the base.

The Pliocene-Pleistocene sediments are clearly folded into an anticline and a syncline of decameter wavelength. The folds are asymmetrical, apparently verging to the West. At the road-cut outcrop, the fold geometry looks approximately cylindrical, with the fold axis sub-horizontal and trending roughly N20 \pm -25 \pm E. Although the possibility that the folding may be related to underground karst evolution cannot be discarded, a tectonic genesis is also plausible. On the anticline fold there are a few vertical fractures trending perpendicularly to the fold axis (Fig. 7), suggesting that they could have been generated by the same horizontal compressive stress that produced the folding, with a WNW-ESE direction.

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Plate 1 - Deformed Pleistocene sediments at the S. Marcos-Quarteira fault zone, near Boliqueime. Notice the transpressive kinematics inferred from the outcropping structures, with strain partition into (probable) right-lateral strike slip and imbricate thrusts. A: schematic cross section; 1: soils; 2: conglomerates; 3: sandstones and mudstones; 4: fault, with slip sense; B: stereo plot of measured fault planes (Schmidt net, lower hemisphere); C: detail of the outcrop (photo by G. Manuppella) (Dias & Cabral, 2002b).

