SIXTH INTERNATIONAL CONFERENCE ON GEOMORPHOLOGY

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PORTUGAL: COASTAL DYNAMICS

A. Ramos-Pereira; J. Trindade and M. Neves

FIELD TRIP GUIDE

A-1

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PORTUGAL: COASTAL DYNAMICS

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The western coast of Portugal has great geomorphological contrasts. Although the cliff littoral prevails, there are some sandy systems associated with lagoons that are worth visiting, both because of their extension and because of the unbalanced situation and environmental vulnerability. The dynamics of the coastal systems in this exposed littoral and the management and land-use problems are the main reasons for the present environmental conflicts: (i) coastal erosion (natural and man-induced), first reported in the XIX century, which led to heavy defence structures and partial or total destruction of coastal systems, (ii) the infilling of the estuaries and lagoons, promoting their eutrophism and risking the local aquaculture communities. The present-day dynamics and Quaternary evolution are the main topics, as well as coastal zone management and conservation.

Previous note:
This field trip, along 500km of the western littoral of Portugal's mainland abridges a large range of coastal dynamic topics. The diversity of the littoral as well as the different approaches to present-day and quaternary dynamics is the aims of this field trip. Therefore, the leaders decided to invite several colleagues from the different visited areas to expose their research. We hope it will be an opportunity to promote a forum to discuss littoral research in the field. At the end of the guide you can find a Table with the list of all the participants in this field-guide and their contacts.

THE PORTUGUESE LITTORAL SETTING

Ana Ramos Pereira, Jorge Trindade, Mário Neves

1. Geological and geomorphological framework

The area visited in this field trip is developed in five main geotectonic domains - Central Iberian Zone, the Ossa–Morena Zone, the South Portuguese Zone, the Lusitanian Basin and the Tagus-Sado Cenozoïc Basin (Fig.1):

> (i) The Iberian Massif consisting of the Central Iberian Zone, the Ossa–Morena Zone and the South Portuguese Zone. Two major fault zones separate them – the Porto-Coimbra-Tomar shear-zone and the Ficalho up thrust (Chaminé et al, 2003; Ribeiro e Silva, 1983). The latter is considered a Paleozoic plate tectonic boundary between a continental (the Ossa-Morena Zone) and an oceanic plate (South Portuguese Zone) where the deposition of an accretionary prism of sediments during Paleozoic age occurred.Granites and metasediments are the main rocks.
>
> The three geotectonic units cluster the so called Iberian Massif, Hesperian Massif or Ancient Massif, which is in fact "the Iberian Hercynian Chain, razed at the end of the Paleozoic" (Ferreira, 1996, p.15-25). The Ancient Massif has been submitted to several episodes of planation during the Cenozoic. The planation surfaces are frequently tilted and uplifted along the alpine and reactivated late-Hercynian faults. The Ancient Massif occupies 70% of the Portuguese mainland.

(ii) The Lusitanian Basin (and its equivalent – the Algarve Basin) is developed like an aulacogen during Mezo-Cenozoic times. It is a tectonic depression related to the opening of the Atlantic Ocean. Sandstones, clays, marls and limestones are the result of changing continental and deep see sedimentation environment in the basin where the subsidence reached 5km deep during Mesozoic times. However, since the Upper Cretaceous, the geotectonic inversion of the African plate drift creates a convergence boundary and the right movement of the Iberian microplate is responsible for the uplift of the previous basin and the reactivation the late-Hercynian and Alpine faults.

(iii) The Tagus-Sado Cenozoïc Basin is a depression created in the last 50My. Induced by the collision between Africa and Eurasia during the Paleogene, and because of the Iberian drift to the North, a reactivation in distension of the NNE-SSW fault system (Messejana and lower Tagus fault; Fig. 1) was produced generating this basin (Ribeiro et al, 1990). Since then and until the Pliocene this area was filled up with detritic sediments (mainly sand and sandstones, clays and some marls and limestones). Since then, a compressive regime prevails and the upper erosion planation surface cut into the detritic sediments of the basin is considered to be upper Pliocene (Martins, 1999).

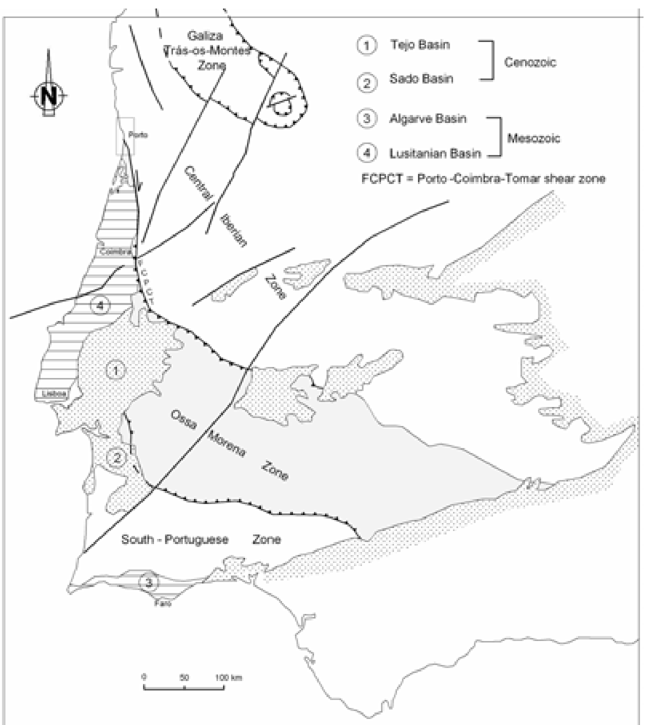

Figure 1. The geotectonic domains.

This compressive regime is responsible for the major landforms (the Serras – small mountains) not only in the Ancient Massif but also in the Lusitanian Basin. Within this regime, the planation surfaces were faulted and uplifted between 0.13 to 0.3mm/y inland and 0.1 to 0.2mm/y in the littoral during the Quaternary (Cabral, 1995) and entrenched by the Quaternary river network. Cuesta and mesa landforms were developed in the sedimentary basins.

2. The littoral

In the Portugal mainland, the landscape is characterized by a heritage landform – the so called coastal platform (equivalent to the Spanish rasa). This platform is related to relative sea level changes and their influence on coastal landforms and deposits (Fig. 2). The continuous flattened landform is slightly sloping towards the sea and can be found at different heights. Near Aveiro, the littoral platform is almost at sea level while, in the South sector of the western front, it can reach 150m of altitude (Pereira, 1990). Landward limit isn't always clear; however, the regular presence of tectonic scarps makes the transition to the continental reliefs often abrupt.

The littoral platform can be erosion or accumulation dominant (Fig. 2). In the first case, the levelling took place independently of the local or regional lithostructural framework. The flat morphology and deposits are more or less preserved depending on the intensity of the posterior local and regional tectonic activity and the density and deepness entrenching the of fluvial network. Correlative deposits locally change in facies so genesis generalizations are not to be made. Littoral and continental sediments, remixed by Pliocene and Quaternary shoreline fluctuations, regularly have no fauna or flora and show the complex evolution of this polygenic littoral feature. The continental shelf has had a Quaternary evolution similar to the littoral platform (Fig. 2) and its western boundary is a structural one – the continental slope - in the transition from the continental to the marine lithosphere.

The nearly 940km of Portuguese West and South coastline are characterised by a semi-diurnal mesotidal regime, with 12.30h tidal cycles which can reach more than 3.5m of amplitude. The western littoral has a NW dominant wave climate with 2-2,5m mean wave height, which contrasts with the 1m in the sheltered South Algarve coast (Pires, 1989). This wave climate is highly seasonal and February is the month with more extreme wave heights frequency, with maximum wave height exceeding 12m.

Coastal systems are diversified but cliffs are predominant. In the western front, granites cliffs are more frequent in the northern sector, sedimentary rock cliffs can be found between Figueira da Foz and Sines and metamorphic rock cliffs in the southern sector. Seasonal wave climate, abrasion resistance and rock permeability are important factors in cliff dynamics and can distinguish predominant processes acting in different sectors of the littoral.

Beach and beach – dune systems are usually of small dimensions conditioned by cliff morphology or small estuary dynamics. However, the Esmoriz (south of Oporto) - Figueira da Foz coastal sector is a 80km continuous sandy system composed by two spits that almost close the Aveiro haff-delta (Ria de Aveiro). This coastal lagoon system is the result of the infilling of a graben. This infilling regime prevails in the present-day, affecting all Portuguese coastal estuaries. West and southern front sandy systems are in great sediment dependence of the coastal drift. In the western front, southwards coastal drift prevails but it can be reoriented seawards by natural promontories, submerged canyons or heavy coastal structures. This results in regional differentiation of sediment transport dynamics along the coast (Fig. 2). Associated with Douro, Tejo and Guadiana Rivers, regions A1, A2, A4 and the Eastern sector of region B (Fig. 2) have a positive sedimentary balance (Pereira, 1992). On the contrary, regions A3 and A5 have a negative one. These regions are in the southern side of two important morphologic structures, the Nazaré and Lisboa Canyons, which stop coastal drift sediment bypassing.

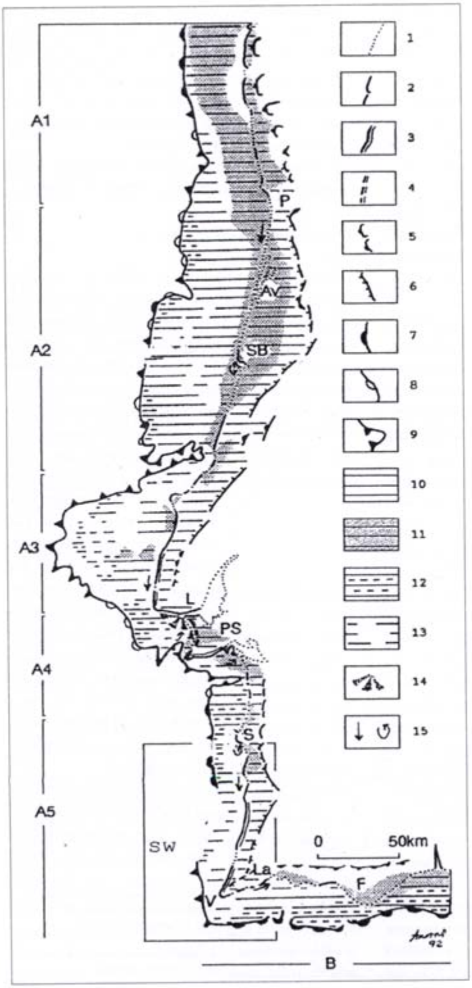

Figure 2. Littoral platform and continental shelf geomorphology. 1 – sandy littoral; 2 – cliff < 50m; 3 – cliff > 50m; 4 – paleocliff; 5 – erosion edge; 6 – tectonic edge; 7 – progradation edge; 8 – aggradation edge; 9 – regradation edge; 10 – erosion dominant coastal platform and continental shelf; 11 – accumulation dominant coastal platform and continental shelf; 12 – progradation dominant continental shelf; 13 – prominent relief's in 10; 14 – profluvial delta; 15 – coastal drift direction. Av – Aveiro; F – Faro; La – Lagos; L – Lisboa; P – Porto; PS – Península de Setúbal; S – Sines; SB – Serra da Boa Viagem. (after Pereira, 2004).

The coastal orientation and the availability of sediments in the continental drift are the main natural factors responsible for the diversity of littoral systems. But the anthropogenic pressure over the littoral has an important role.

Since always the Portuguese have been linked to the sea and see the littoral landscape as a way of subsistence, expansion, development and leisure. During the last decades there has been a population movement towards the littoral (Fig. 3), not only in the metropolitan areas of Lisbon and Oporto, but along the littoral between than and the Algarve. In the 1970's the unplanned management allowed the construction of tourist complexes in sensitive areas, namely on dune fields and instable cliffs. The result was not only the destruction of the dunes but posteriorly the absence of the natural beach, as well as the quick cliffs' retreat. The construction of groins to protect coastal areas submitted to erosion as well as to prevent the infilling of estuaries where the main ports are installed led to a lack of sediments and more erosion leeward (Fig. 3). The critical areas submitted to erosion are the sandy coasts and cliffs cut into sandstones and marls (Fig. 3).

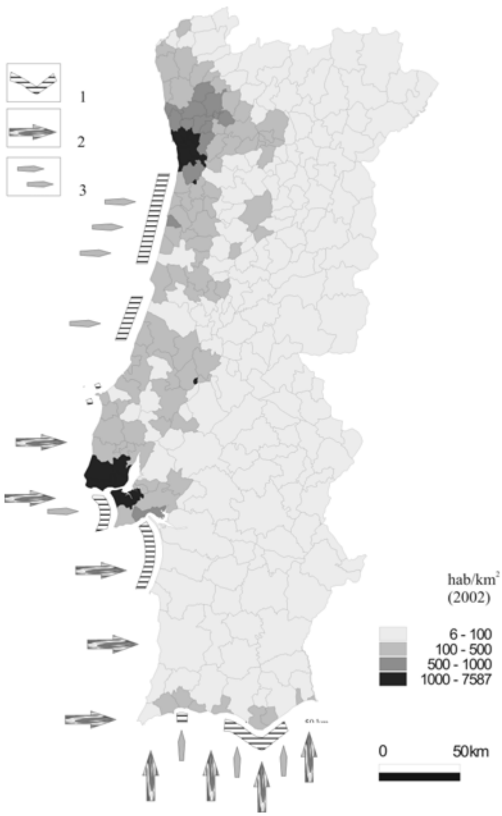

Figure 3. Vulnerability of Portuguese mainland littoral. 1 – coastline submitted to strong erosion; 2 – sensitive coastline to tsunamis; 3 – frequent overwashes (Pereira, 2004a).

The anthropogenic pressure is still growing, related mainly to tourism industry, which is one of the most important of the country. It benefits from a Mediterranean climate with a dry summer and more than 2390 sunlight hours per year (Fig. 4).

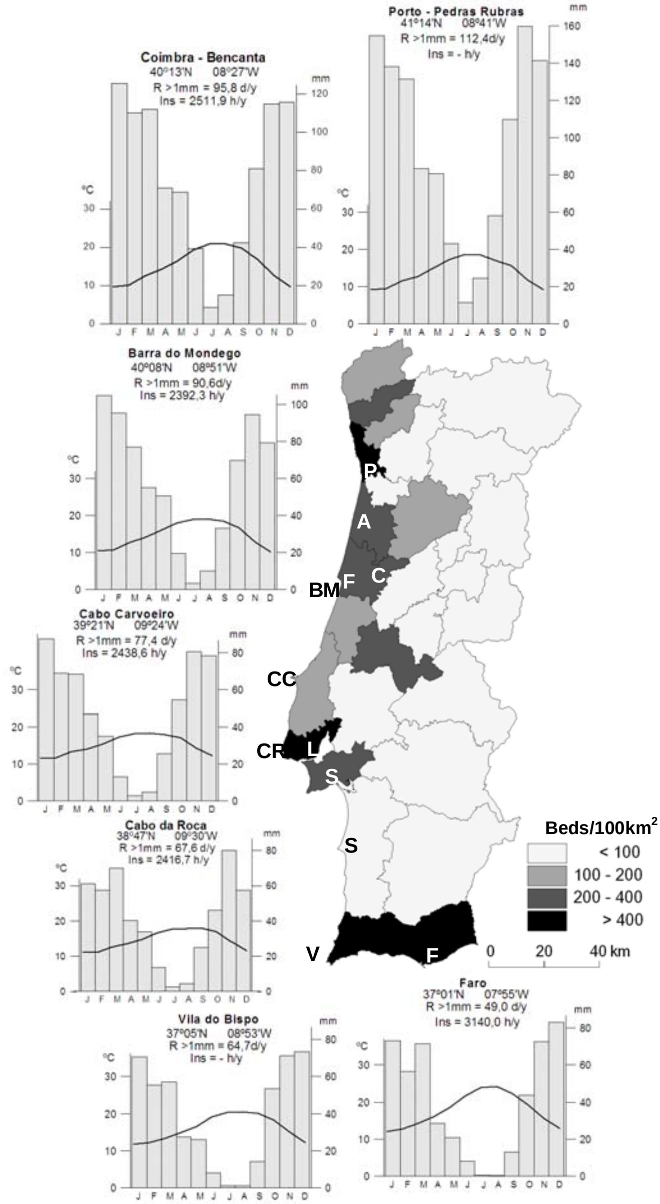

Figure 4. Climate and hotel capacity. P – Oporto; A – Aveiro; C – Coimbra; FF – Figueira da Foz; BM – Barra do Mondego; CC – Cabo Carvoeiro; CR – Cabo da Roca; L – Lisbon; S – Setúbal; SN – Sines; VB – Vila do Bispo; F – Faro; R>1mm – days per year with at least 1mm of precipitation; Ins – Insolation in hours per year. Source: Normais Climatológicas 1941-1970, Instituto de Meteorologia and Estatísticas Gerais (2002), Instituto Nacional de Estatística.

THE LITTORAL OF OPORTO REGION (NORTHWEST PORTUGAL)[1]

António Alberto Gomes; Maria Assunção Araújo; Andreia Sousa

1. Geomorphological and geological setting

One of the most common characteristics of the Portuguese littoral is the so-called “littoral platform”, comprehending different altitudes and bordered from the inland by a contrasting straight relief.

The littoral platform is at around 75m (on the North, close to Ave River) till 130m (at the South) and goes down towards the sea like a staircase (Ferreira, 1983). This platform is bounded to the east by a step relief – the “marginal relief” (Araújo, 1991) and has the shape of a fault scarp related with the Porto-Coimbra-Tomar shear zone.

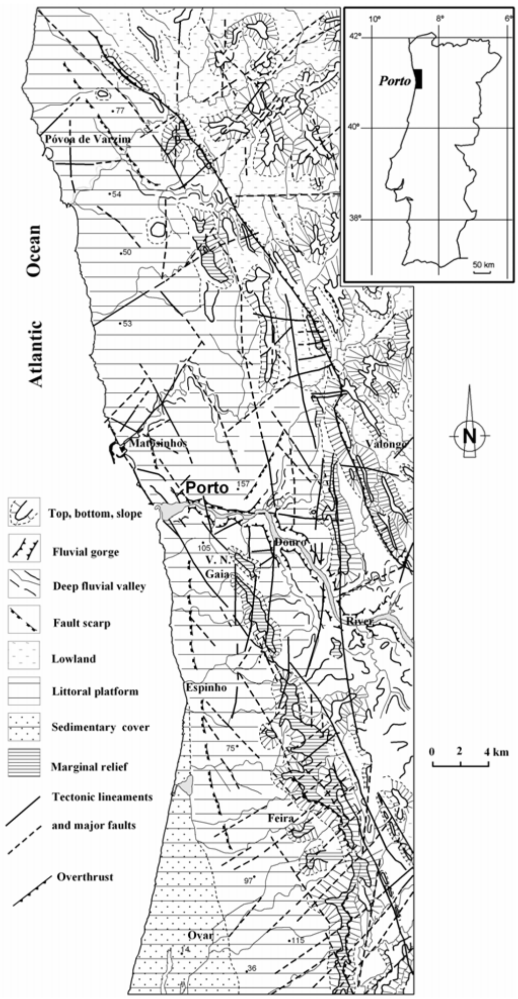

Figure 5. Geomorphological setting of Oporto region.

[1] **1st Stop** – S. Felix de Launde.

This planation surface is generally covered with several outcrops of the so-called Plio-Pleistocene deposits (Fig. 5). Till the eighties this platform has been considered as stable staircase of old marine levels, registering in a passive way the eustatic fluctuations. The rigid step easterly bordering it should be a fossil cliff. However, new studies have proved that many of these deposits have a continental origin.

Douro is the most plentiful river of the Iberian Peninsula and its unusual entrenched valley even close to its mouth, is generally understood as evidence of antecedence (Rebelo, 1975; Daveau, 1977). It seems that the river existed already before the uplift of the tectonic compartment, creating this strongly entrenched valley.

The main lithological units are (Fig. 6):

i) granitic rocks (Variscan and/or pre-Variscan), including two mica granites, medium to coarse grained, with mega crystals; biotitic granites fine to medium grained; gneisses, migmatites and gneissic granites;

ii) metasedimentary rocks (upper Proterozoic-Palaeozoic), which include schists, greywackes, quartz-phyllites and quartzites;

iii) sedimentary cover (post-Miocene), including alluvium and fluvial deposits.
The Crystalline basement, which is strongly deformed, and overthrusted metasedimentary rocks and granites are the regional geotectonic framework of Porto region (Chaminé et al. 2003a, b).

Porto–Albergaria-a-Velha shear zone (s. str.) corresponds to a major NNW-SSE dextral strike-slip fault of sigmoidal multiscale geometry, stretching from Porto to Tomar (Chaminé, 2000). The dextral faulting system is associated with transpressive kinematics triggered by the post-orogenic collapse of the structure along the ancient Porto–Coimbra–Tomar thrust planes. These processes generated a multitude of discrete ENE-WSW to NE-SW regional brittle fault systems.

2. The littoral platform

The littoral platform has been understood simply as a staircase of marine terraces facing the sea, as mentioned before. However, that idea applies only to the lower and western part of this platform. The older deposits have a fluvial origin. A step, between the marine and fluvial deposits, establishes a very sharp separation and seems to have a tectonic origin (Fig. 7).

In the fluvial deposits cluster it is possible to make a distinction between the higher and lower deposits. The first ones are related to low energy environment, probably deposited inside a littoral plain and the lower and newer deposits are much coarser and have a typical alluvial fan facies. There is a clear evidence of the existence of two phases of geomorphological development: the younger, reddish alluvial fan deposits contains, in several places whitish blocks that are remains of the older deposits. The destruction of an old sedimentary cover suggests the change of a low slope sedimentary environment into a landscape where increased slopes furnished some boulders (more than 50cm wide). It is possibly to correlate this change to a climate crisis (Ferreira, 1983), because the kaolinite, which was very abundant in the older deposits (more than 90%), is more scarce in the alluvial fan deposits.

The fluvial deposits reveals another difference: the strong iron cuirasses disappears from the older ones to the newer ones, which may represent the change from tropical climate conditions prevailing during the Miocene and lower Pliocene into more temperate ones at the end of Pliocene and during the Quaternary.

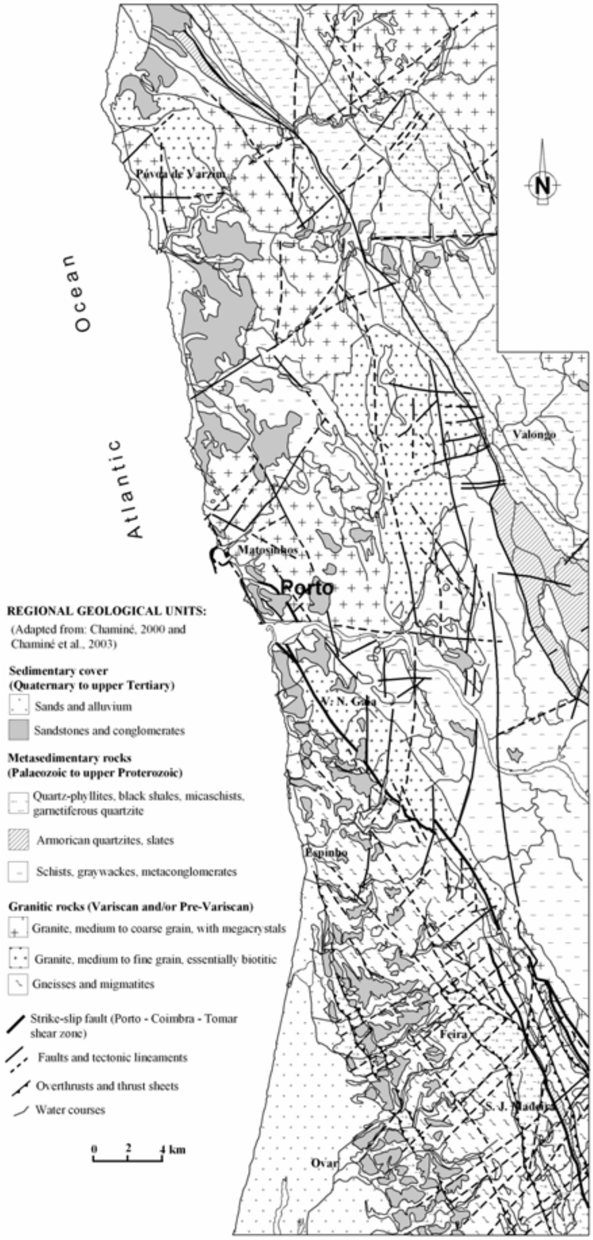

Figure 6. Geological framework of Porto region.

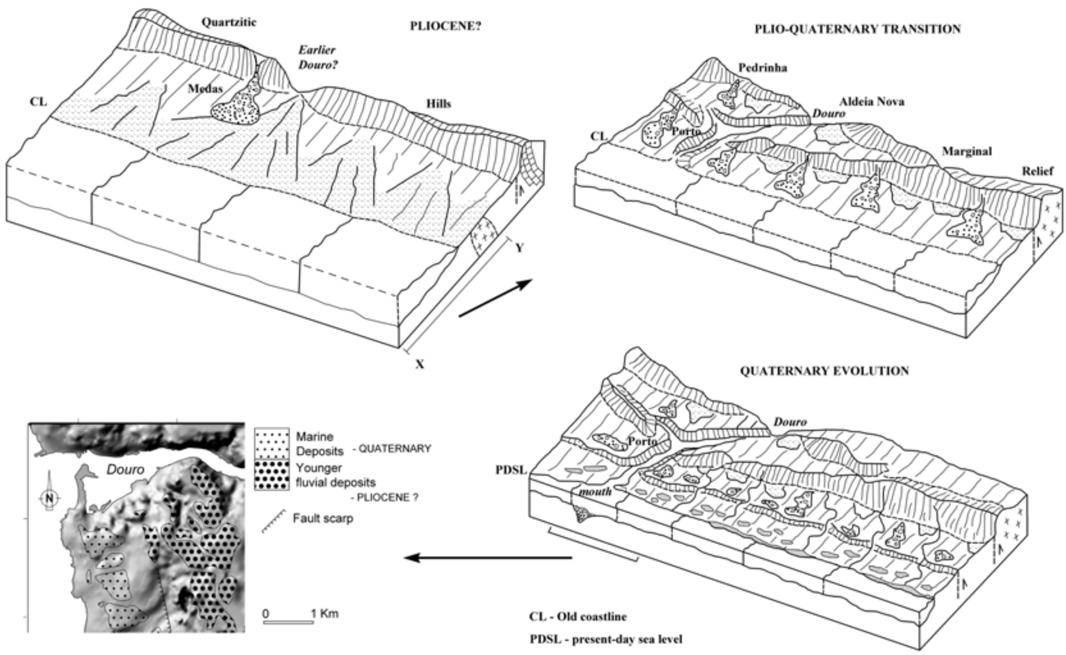

Figure 7. Geomorphological evolution of Porto littoral platform (Araújo et al, 2003).

Since the forties, Ribeiro et al (1943) highlighted that the higher fluvial deposits (Rasa de Baixo) where balanced to the east, towards marginal relief, in the opposite direction of the sea. That implies a presumption of tectonic movements acting after the deposit's formation. The creation of a marginal horst, in NNW-SSE direction, by a post-Pliocene movement on Porto-Coimbra-Tomar shear zone (marginal relief) could be the origin of the alluvial fan deposits (Araújo, 1991; Araújo et al, 2003). Those alluvial fan deposits where also affected by compressive movements, attested by several mostly inverse faults (Araújo et al., 2003).

Between fluvial and marine deposits exists a quite straight step that can reach 30m (Fig. 7). This step suggests that, during Quaternary, the western part of littoral platform must have subsided along a sub-meridian fault (Araújo et al, 2003). This process allowed marine erosion and sedimentation over the subsided tectonic compartment. This fault scarp was reworked, afterwards, by the sea during the higher marine level and may be considered a fossil cliff.

In conclusion, there were at least two phases of neotectonic movements affecting the fluvial deposits. Presumably, the most recent of these movements created also the straight slope between fluvial and marine sediments.

Douro's Estuary Dynamics[2]

António Alberto Gomes; Ana Ramos Pereira; Maria Assunção Araújo; Andreia Sousa; Francisco Veloso Gomes

The Douro River has the greatest drainage basin in Iberia (97, 682km^2), 81% in Spain and 19% in Portugal (INAG, 2003). The Douro's River average natural outflow is 22, 458hm^3 (INAG, 2001), which corresponds to a 710m^3/s average discharge.

The estuary's topography has been modelled throughout the Quaternary, especially during low sea levels. However, insufficient research has been developed on this subject. We can point out that at S. Paio bay (Fig. 8), the Upper Pleistocene thalweg was near the southern bank, deeper than 50m bellow present-day sea level (Carvalho and Rosa, 1988). Nowadays, the estuary's depth reaches -31m about 3km upstream (INAG, 2001).

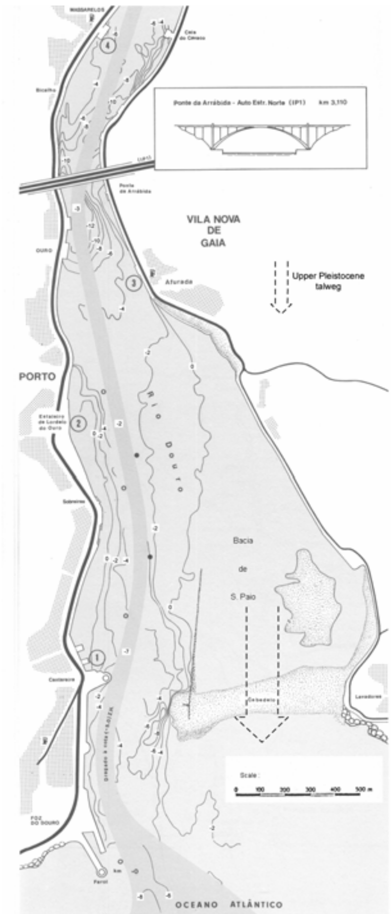

Figure 8. Estuary map showing the bathymetry and the dredge navigation channel; equidistance equal to 2m (IND, 1991)

[2] **2nd Stop** – Vila Nova de Gaia – Cabedelo.

The estuary's width varies between 130m and 1, 310m near the Douro River's mouth. Its entrance is nowadays partially blocked by a changing sandy spit. In a mesotidal environment that reaches 4m, the estuary tidal range varies between 2.4 – 2.7m (spring tides) and 1.2 – 1.5m (neap tides). The tide could penetrate, until 1985, 37km inland, when the Crestuma dam was built. Nowadays, this dam is the new estuary boundary, which extension has been reduced to 21.6km.

During dry weather the tidal currents prevail while during floods it's the opposite. Though the entrenched valley and its irregular bottom topography are conditioning factors, the river discharge and the tidal range are the triggering ones.

The river's discharge has changed in the last half century with the construction of 36 dams in the Portuguese counterpart. The lowering of the water discharge and the velocity curves as well as the retention of sediments in the dams were not the only consequences. The reduction in the flow of solids creates a lack of sediment flowing into the estuary and offshore. This reduction is also related to sediment extraction and dredging of large amounts of sand along the river and inside the estuary (to ensure navigation security), its quantification is still unknown, but it has subtracted a very large amount of sediments. They are now missing in the long shore drift. The sediment dredging estimative in the downstream part of the river, between 1982 and 1986, was $3x10^6m^3$, almost the amount calculated for littoral drift.

The potential transport capacity of oblique waves (between 1 and 2 million m^3/yr) is now greater than the annual sediment volume supplied by the rivers (at present between 0 and 0.2 million m^3/yr). In present-day some tens or hundreds of thousands of cubic meters per year reach the long shore drift, very much depending on climate wave and fluvial discharge (Veloso-Gomes et al, 2002).

Vidinha et al (2002) highlighted the importance of the Douro River to the long shore sediment supply, mainly to the southern sandy beaches and dunes. Studying the mineralogy of the fine fraction of beach and dune sediments, these authors show the importance of this source till the Aveiro inlet, 52km down drift. According to Oliveira et al (1982), the Douro River represented 90% of the sediments drifting along the shore.

2. THE MOUTH SPIT

Cabedelo is the Portuguese name for the sandy spit that almost encloses the river mouth. It is a very changeable feature (in position and width), responding to climate wave and river discharge.

The long shore drift act mainly in the N-S direction although some singular events of SE currents can be found as a consequence of specific hydrodynamic processes (refraction and diffraction) or SW storm episodes. The wave climate has medium significant wave heights from 2 to 3m, with periods ranging from 8 to 12s and storm significant wave heights exceeding 8m (maximum wave heights up to 1.7 times the significant wave heights), with periods reaching 16 to 18s (Fig. 9). The local wave conditions are different from the offshore ones due to the effect of the bathymetry and local phenomena, especially refraction, diffraction and shoaling (Veloso-Gomes et al, 2002).

Sediments travelling S-N, near Cabedelo, can reach 1,5 millions m^3/yr (INAG, 2001) and could close the river mouth. However, the river's discharge and the tidal currents balance and special dredging keep it open.

During the December 1909 flood (19, 500m^3/s and 100 year return period), Cabedelo almost disappeared but it gradually grew up, attaching itself to the southern margin in June 1910.

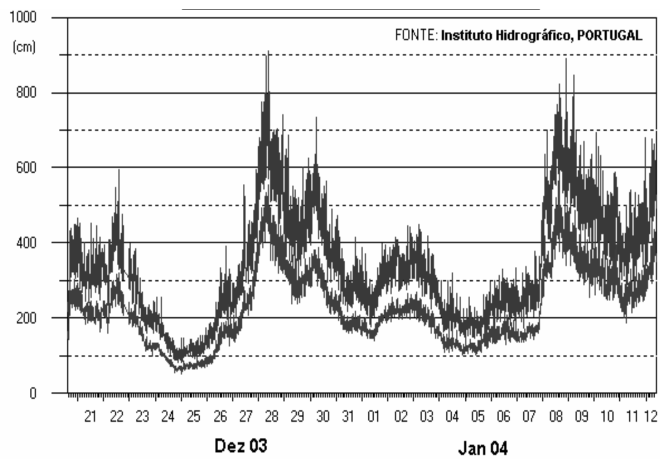

Figure 9. An example of significant and maximum wave high during a storm period at Leixões buoy (www.hidrográfico.pt).

In the nineties the wash-over of the spit, bringing oceanic waves upon S. Paio bay salt marsh got more and more frequent, even during a period with less severe floods while the spit migrates upstream (Fig. 10, 11 and 12).

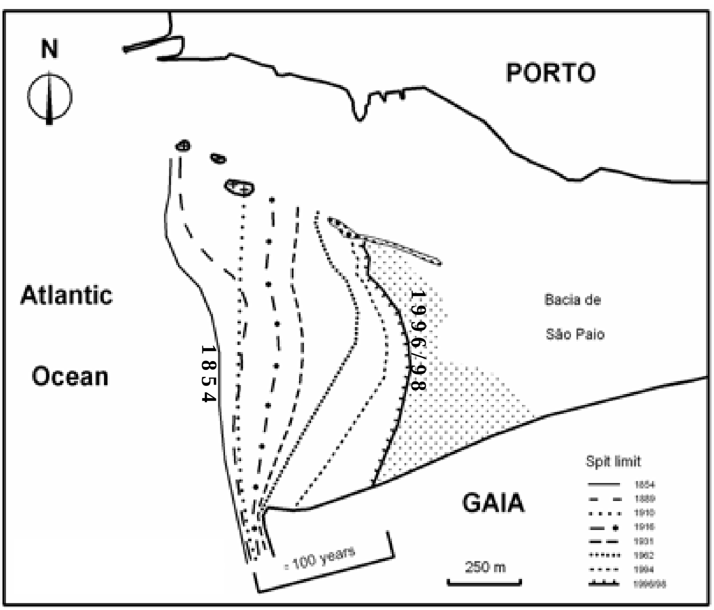

Figure 10. The retreat of the sand spit since 1854 to 1996-98. Administração do Porto Douro – Leixões (1999).

Floods can be controlled by dams only up to amounts of 7,000 a 9,000m^3/s. The extraordinary floods are the ones that go beyond +6.00m (above hydrographic zero), measured at the right margin, near D. Luiz's bridge. These floods can go over Ribeira embankment. Douro's extraordinary floods have a great water flow, a rapid propagation, a strong water level elevation and a short duration, because the drop of water level is relatively fast (IHRH, 2003).

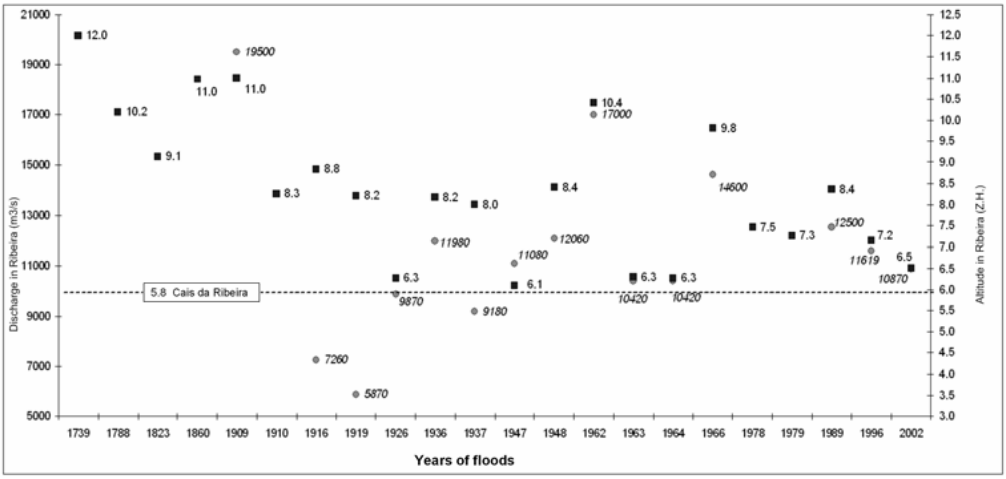

Figure 11. Extraordinaire floods in Ribeira – Porto (IHRH, 2003). The altitudes are referred to hidrographic zero (1,8m under topographic zero).

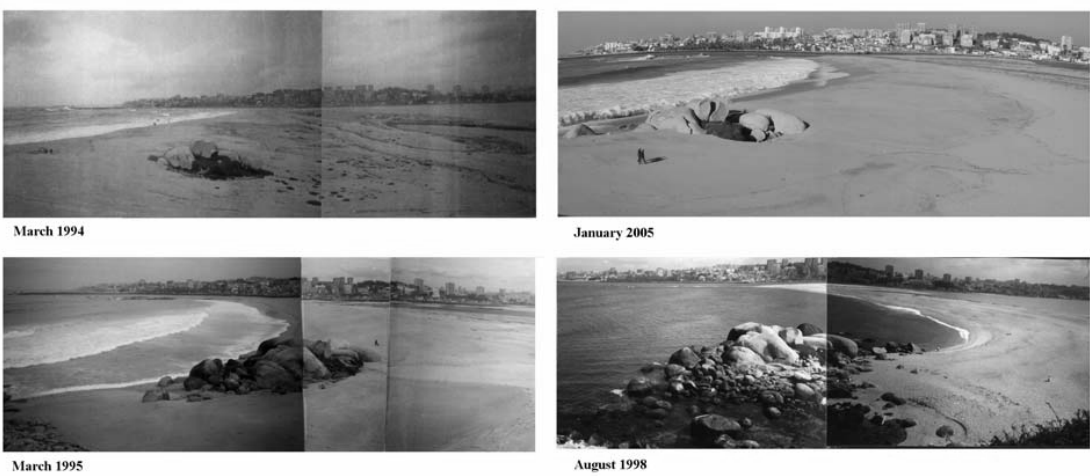

Figure 12. Panoramic views showing recent evolution of the Cabedelo sand spit.

The spit evolution tendency (thickness and upstream migration) changed during the winter 2000/2001, when a big flood occured. The triggering factors were: (i) the intense winter and spring rainfall, that was above the 1961-90 average and doubling the annual average in some areas of the drainage basin and (ii) great discharge of an upstream dam, after a quite long water river retention related to an accident with a bridge. The enormous amount of water and sediments provided the supply of the spit that grows and migrates towards the sea.

The present-day situation shows the previous tendency as no big floods have been registered since then. To stabilize the river mouth and allow safe navigation, a new project is going to be implemented during 2005 and 2006. The layout is present in figure 13.

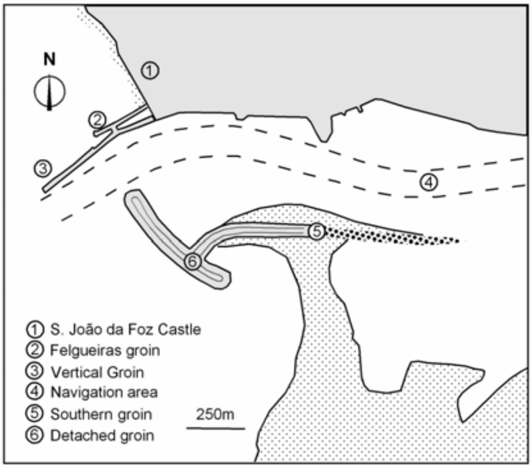

Figure 13. The new project to allow bank protection and safe navigation on the Douro´s mouth (Jornal de Notícias, 30 de Abril de 2004).

THE EVOLUTION OF THE NORTHERN END OF THE AVEIRO LAGOON SYSTEM[3]

Helena Granja

1. Geological and geomorphological setting

The Esmoriz lagoon is located south of Espinho, in a flat area, slightly sloping towards the sea. It has a smooth NNW-SSE relief corresponding to the boundary between the littoral and the Douro valley. This nearly flat surface (the littoral platform) has its greatest width at Cantanhede (20km).

Most of the littoral is covered by different dune systems and Quaternary deposits forming the northern end of the Aveiro lagoon, the border of the brackish coastal lagoons (Esmoriz and Aveiro) and of the fresh-water lagoons (more inland).

The strongly metamorphic ante-Ordovician schist-greywacke complex can be observed in two bands, separated by a NW-SE oriented porfiroid granite, inland from the littoral border.

2. Palaeoenvironmental reconstruction of the lagoon

The present-day Esmoriz lagoon is the remnant of a larger lagoon system, probably identical to the one still existing today at Aveiro (Fig. 14). Based on the description of outcrops and some cores, the

[3] **3rd Stop** – Esmoriz Lagoon.

sedimentary bed succession, overlaying the old bedrock, and the palaeoenvironmental reconstruction of the lagoon were possible.

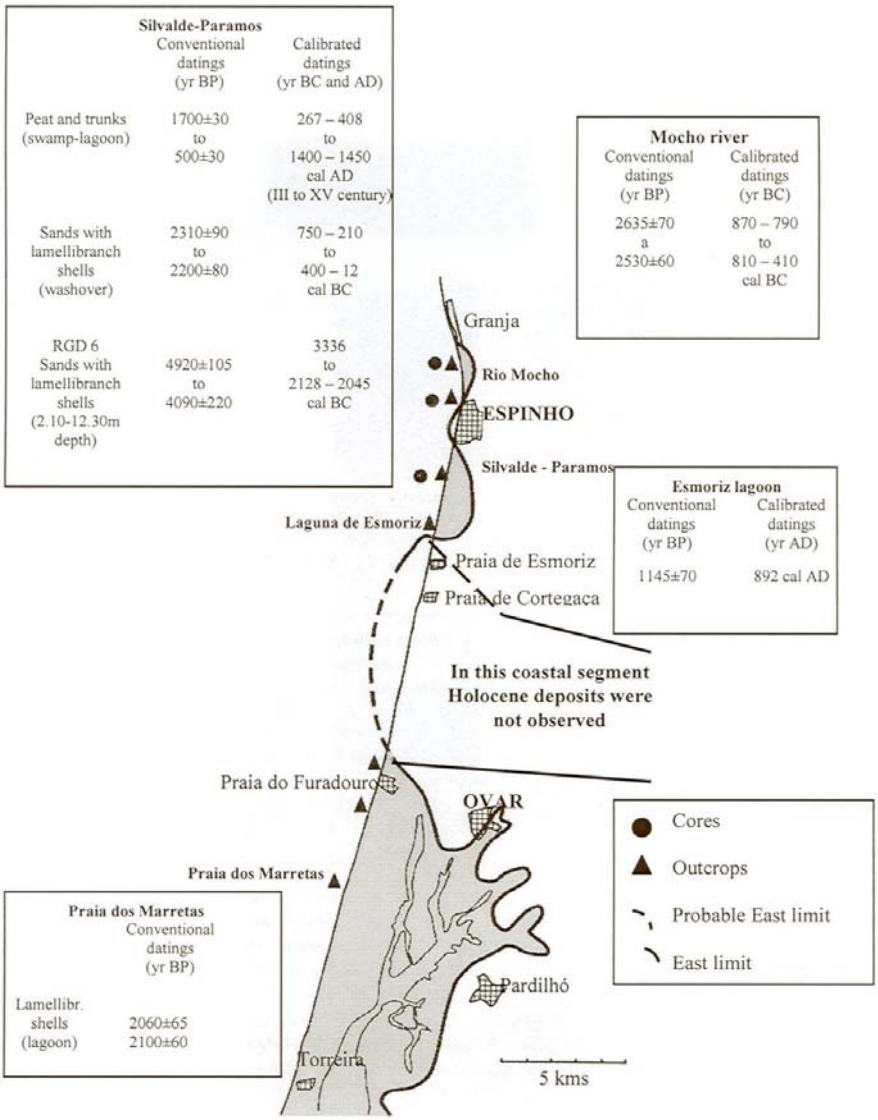

Figure 14. Probable limits of the Holocene lagoonal system between Espinho and Torreira, including the Ovar barrier of the present Aveiro lagoon.

The sedimentary sequence of the Esmoriz lagoon constitutes the Silvalde-Paramos Formation, a Holocene formation in unconformity with the Cortegaça (Holocene) and the Maceda (Late Pleistocene) formations.

The core RGD 6 drilled near Paramos beach reached the schist bedrock at 17m depth (11m below sea level) over which lays the Silvalde-Paramos Formation. This formation shows several sedimentary units (Fig. 15). The coarsest pebbly sands may correspond to overwashes during storm periods, like those dated 440±50yr BP (Granja, 1999, 2002). The upper part of this formation was observed at Silvalde Beach and contains two main units (Fig. 14).

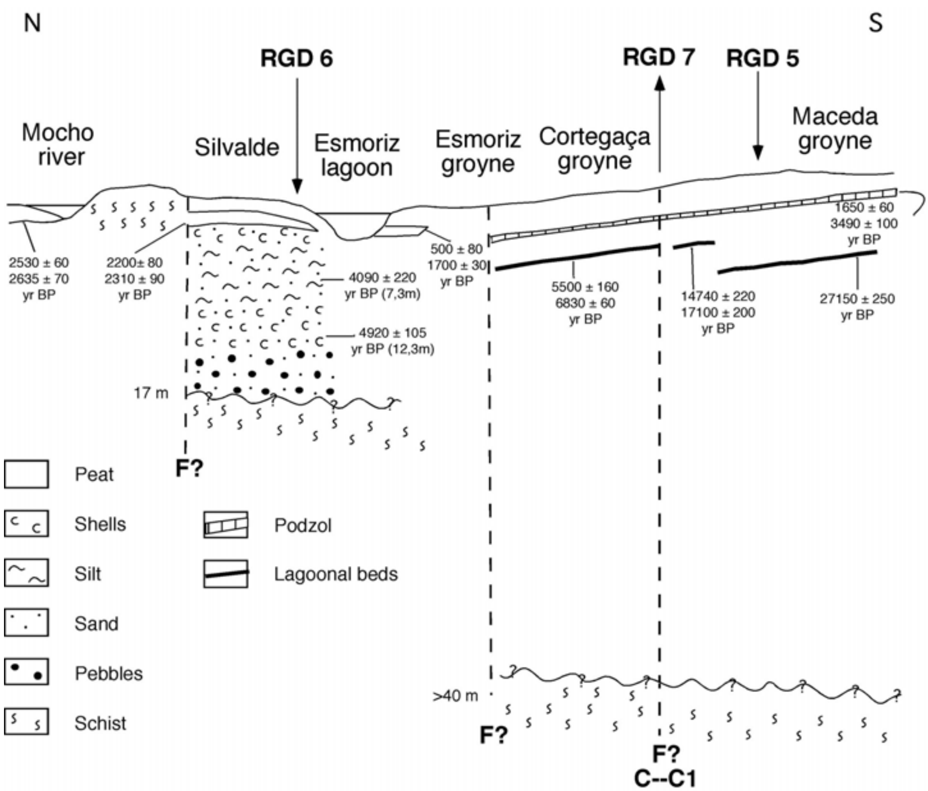

Figure 15. Schematic cross-section along the coast showing the Esmoriz lagoon emplacement

It can be observed south of Paramos, where outcrops have been uncovered during a few short periods. Some of these outcrops contained oyster shells (Ostrea edule) and bivalves from quiet waters with low salinity (e.g. Cerastoderma edule, Scrobicularia plana, Tapes decussatus), indicating a brackish and confined environment (Fig. 16). That is a wider lagoon environment with a seaward barrier more westwards than the present day Aveiro lagoon barrier.

Though data is still not enough to reconstruct the Aveiro lagoon evolution, there are some geo-indicators that point to one that differs from that of Girão (1941) and other authors. The lagoon barrier of Roman times should have gradually migrated eastwards. Now, due to strong erosion, inland beach migration and cliff retreat, the older lagoon deposits are being exposed.

According to the available datings, soil development over beach and aeolian sands coexisted with the silting up of the lagoon during Roman times. The soil was later overwashed as can be seen in the cliffs of the Cortegaça and Maceda beaches.

3. Forcing factors

The existence, during the Late Holocene, of lagoon systems in the NW coastal zone of Portugal is the consequence of sea level rise (major controlling factor), sedimentary budget, climate and (neo) tectonics. The area where the Esmoriz lagoon is situated has several faults, some of which were probably active during the Pleistocene and are geo-indicators of deformation.

The lagoon itself seems to be controlled by faults (Granja et al, 1999a), the southern one putting in contact the Silvalde-Paramos Formation and the Cortegaça Formation, whose facies are different (Fig. 15).

Besides the macroscale forcing by sea level and neotectonics, climate certainly played a role on the mesoscale; at this point this subject is still poorly understood. Soil development and overwashing are probably related to climate forcing, mainly to shifts during the Holocene.

Figure 16. Outcrop of Holocene lagoonal deposit with Cerastoderma edule aged of 2060±65yr BP, Marretas Beach, 2001.

GEOMORPHOLOGY AND COASTAL DYNAMICS OF THE FIGUEIRA DA FOZ REGION[4]

Pedro Proença e Cunha; António Campar; A. Ramos; Lúcio Cunha; Jorge Dinis

1. Introduction

In the Figueira da Foz region the main geomorphological units are (Fig. 17):

(i) a large dune field as the main feature of the littoral plain, including several generations;

(ii) a limestone massif tilted southwards – Serra da Boa Viagem, with the top cut by a marine incursion, which deposited quartz-rich sands;

(iii) the Mondego River estuary was practically pristine until the mid 20th century, except the mouth area fixed by embankment; the recent urban growth of Figueira da Foz and its harbour development have changed the local landscape;

(iv) a beach-dune system over 42km long (Figueira da Foz – São Pedro de Moel), currently under erosion.

2. The Serra da Boa Viagem and the northern field-dune

The Serra da Boa Viagem is mainly a Jurassic limestones massif, with marls and sandstones, dipping towards S. Its north face is a fault scarp (Fig. 17). The top is a platform reaching 258m, dipping E, probably an abrasion ramp of marine genesis (early Piacenzian ?), dominating the staircase of strath terraces that can be documented in the estuary area. The terrace deposits are conglomeratic, with minor sandstones and mudstones. On the west face of the Mondego cape (near the lighthouse), there is a wave-cut terrace around 100m a.s.l. The well-sorted marine coarse sands and gravels include abundant quartz and some carbonate bioclasts. It is planar cross-bedded and partially cemented. Conglomeratic lags are mainly composed by quartz and quartzite. In this outcrop the deposit is located in the palaeo-shoreline angle, and its composition points to significant palaeo-slope stability. Further north, the same beach deposit also includes bioclasts of molluscs and crustaceous representative of seawaters colder than nowadays and it is covered by a colluvium (including angular limestone clasts). The age of this marine terrace, could be early to middle Pleistocene (?) (Soares et al, in press).

Near the base of the slope, another marine terrace (Murtinheira deposit) can be observed, at 2 to 8m a.s.l., assigned to the last interglacial period.

In the limestone massif two episodes of karstification have been recognized. One is superficial and the other developing dolines that reach 15m deep, funnel-shaped and sometimes carved in dish-shaped and swallow-holes and caves (Fig. 18). The older one probably occurs after the Pliocene marine incursion and is under a sand cover. The second phase took place most likely during and after the uplift of the massif (Pleistocene ?).

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[4] **4th, 5th and 6th Stops -** Bandeira sightseeing, Montego Cape lighthouse and Cova beach.

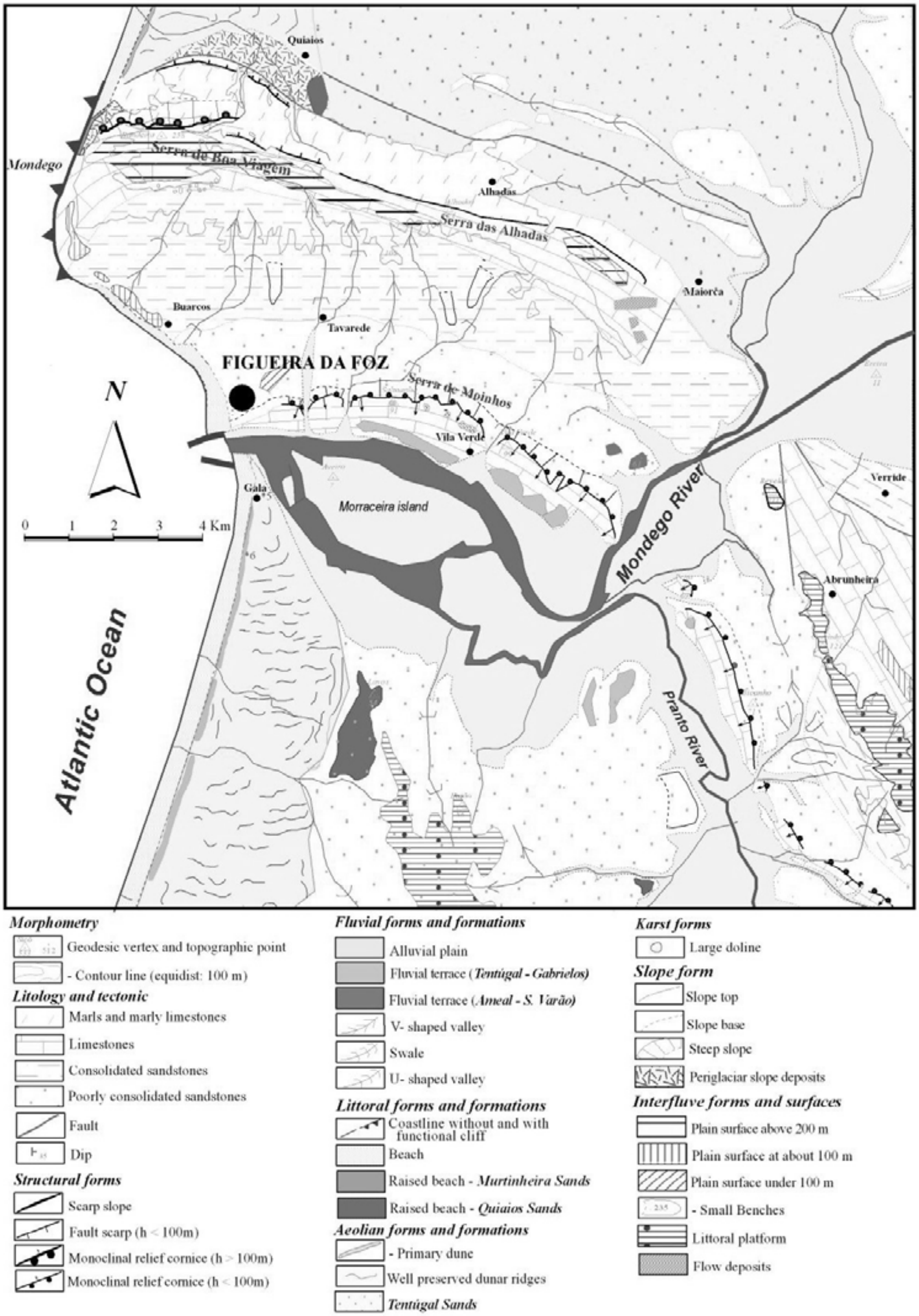

Figure 17. Geomorphological map of the Figueira da Foz region. (A. Ramos and L. Cunha).

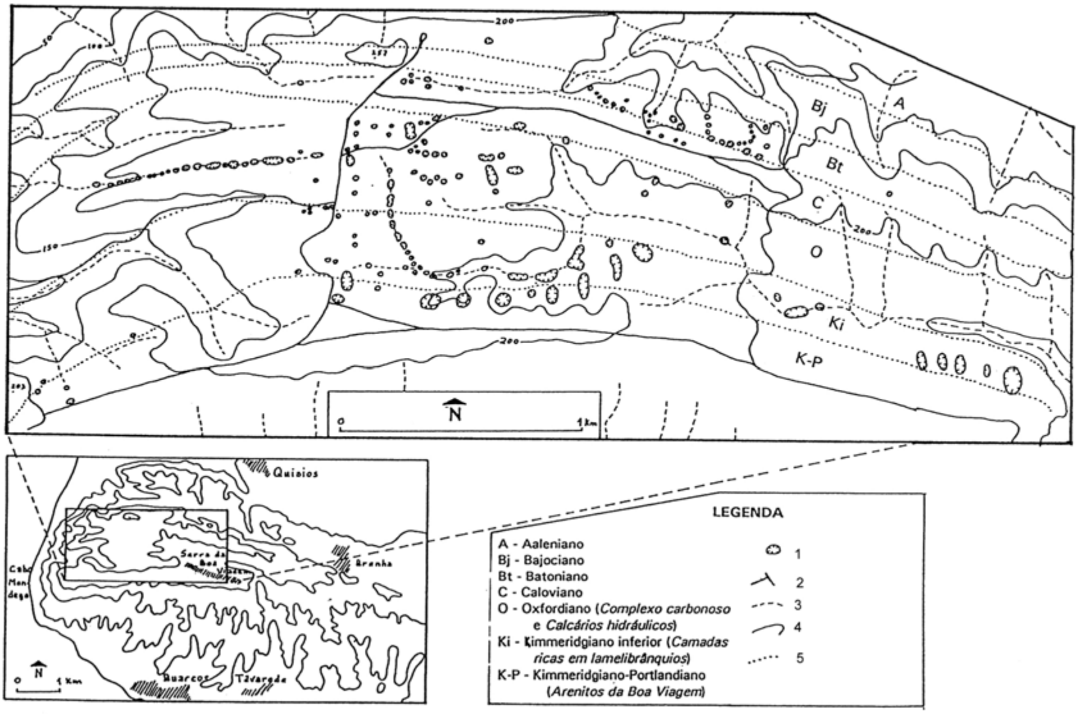

Figure 18. Dolines of the Boa Viagem hill (Almeida, 2001). A - Aalenian; Bj - Bajocian; Bt - Batonian; C - Calovian; O - Oxfordian; Ki - Early Kimmeridgian; K-P - Kimmeridgian-Tithonian; 1 - doline; 2 - road; 3 - stream; 4 - contour line; 5 - geological contact.

The oldest dunes constitute an eastern belt and are intensely farmed. Their morphology is quite smoothed and they have an organized hydrological network. A podzol with a hardened B-horizon has developed. Most of the dunes have a parabolic shape and are oriented NW-SE. Dating performed by Carvalho and Granja (1997) in similar soils at Cortegaça (around 50-70 km to the North) has given an age of 1500 years BP, at least.

The dunes’ second generation still well preserved have parabolic aeolian morphology, related to W-E dominant winds. They have soils evolving to a podzol but without a hardened horizon yet. Its larger outcrop is located in a triangle close to Quiaios (Fig. 19). There are no datings but we presume they were built during the Little Ice Age or even before.

The latest dune generation was transgressive with an average speed of 20m/yr till it was covered with a maritime pinewood seeded between 1921 and 1940. It is built by linear dunes (transverse and oblique) oriented W-E. A regosoil with a pH from weakly acid to basic has developed in these sands. It represents most of the dune surface and its eastern boundary, with the oldest dunes, is marked by a linear sequence of inter dune lakes.

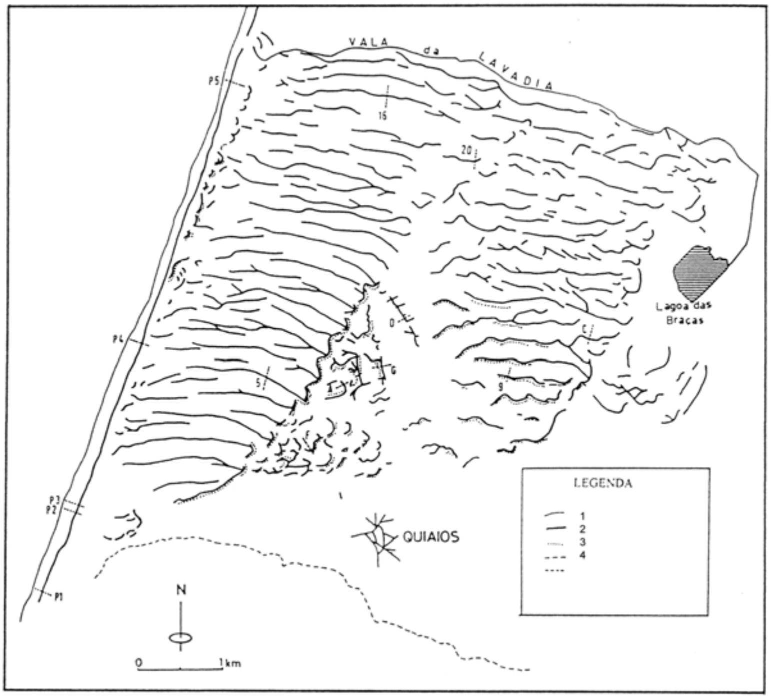

Figure 19. Dune field (Almeida, 1995). 1 - Coastline; 2 - Dune crests; 3 - Sharp lee faces; 4 - Serra da Boa Viagem.

3. The Mondego River estuary

The Mondego estuary is 26km long, between Figueira da Foz and Montemor-o-Velho (Fig. 20).

Upstream of Coimbra the Mondego River has a deep V-shaped valley incised in the Hercynian basement, but downstream the river flows in a floodplain up to 4km wide, where the estuary is developed (Fig. 20). The Serra da Boa Viagem relief and headland (Mondego cape) controlled the position of its distal sector.

The Mondego Estuary morpho-sedimentary evolution was strongly controlled by fluvial and tidal dynamics, being its mouth dominated by waves. Historically, the Mondego River mouth was an unstable inlet; during the 19th century. Several attempts were made to assure stabilization and the region had little environmental changes. Since 1960, this coastal zone, near the town of Figueira da Foz, experienced a very quick morphological change caused by intense human activities in the Mondego drainage basin and in the highly dynamic coastal zone.

From about 6.5km upstream of the mouth, the river divides into two branches (northern and southern). They converge downstream defining the Morraceira island (Fig. 21 and 22), an old tidal mudflat. Despite the significant fluvial and tidal flows, the strong southwards longshore currents, generated by the main NW wave trains, built an important spit in the northern side, such that the estuary can be morphologically classified, after Fairbridge (1980), as a bar-built estuary. The area has a mesotidal range (3.6m), with semi-diurnal tides and a small diurnal inequality.

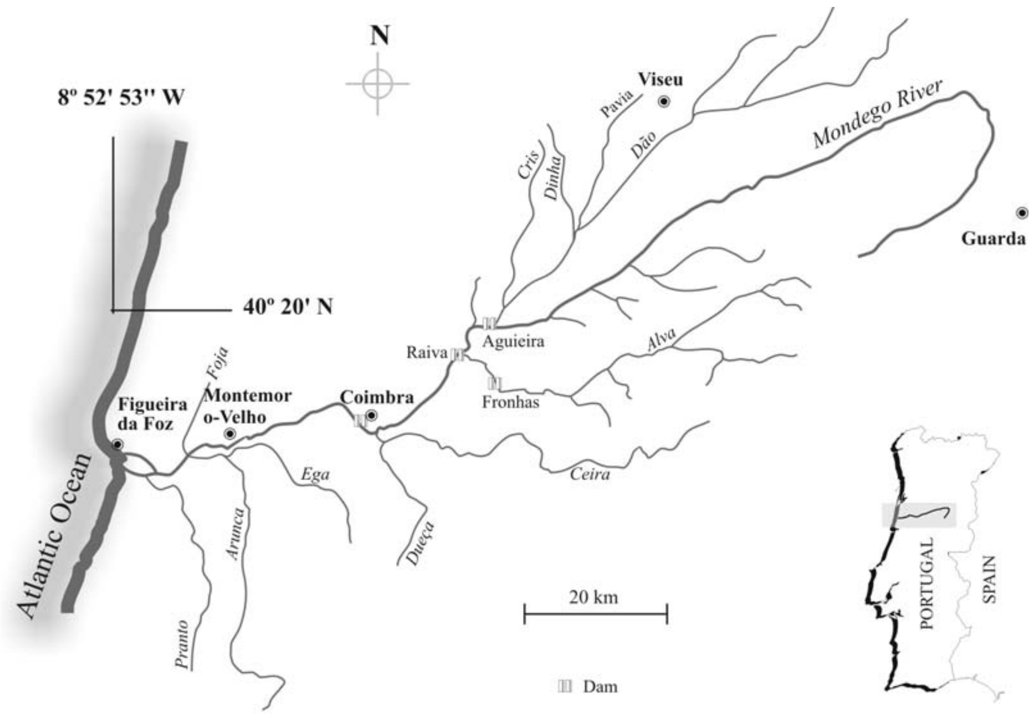

Figure 20. Map of mainland central Portugal, identifying the Mondego River drainage basin, main localities and dams. The estuary area is comprised between Montemor-o-Velho and Figueira da Foz (Cunha *et al,* in press).

In the Figueira da Foz offshore the most frequent wave height is 1-2m, dominantly from the NW quadrant (Carvalho and Barceló, 1966). Storm events are important: the return period for extreme annual significant wave height of 9.5-10.0m is 5-years and for waves with 11.5-13.6m it was estimated in 50-years (Carvalho, 1992).

The Mondego River estuary (Fig. 17, 20, 21 and 22) is composed by two subsystems — the Mondego and the Pranto subsystems — with contrasting morphological, sedimentary, hydrodynamic, physical and chemical characteristics (Cunha and Dinis, 2002). In the Mondego subsystem the river inflow is more important than the tidal control, but the Pranto subsystem is clearly dominated by tides. The Mondego subsystem is the main unit directly connected to the trunk river, currently including a long navigation channel, and exclusively bounded by artificial banks in the outer side; the Pranto subsystem is shallower and less affected by anthropogenic interventions.

The Mondego subsystem is well mixed with reduced fluvial flow and stratified during seasonal floods. The hydrodynamic pathways (tidal vs. fluvial, this prevailing in the northern margin) causes clear contrast in sediment transport. The Pranto subsystem is mainly brackish, well mixed, with strong tidal hydrodynamic and large physical and chemical fluctuations; fluvial floods rarely have high discharges. In general, a natural silting is deduced as a centennial trend for the entire estuary, but modified by recent heavy engineering works: opposed by dredging in the Mondego subsystem yet reinforced by a reduction of both circulation and connections in the Pranto subsystem.

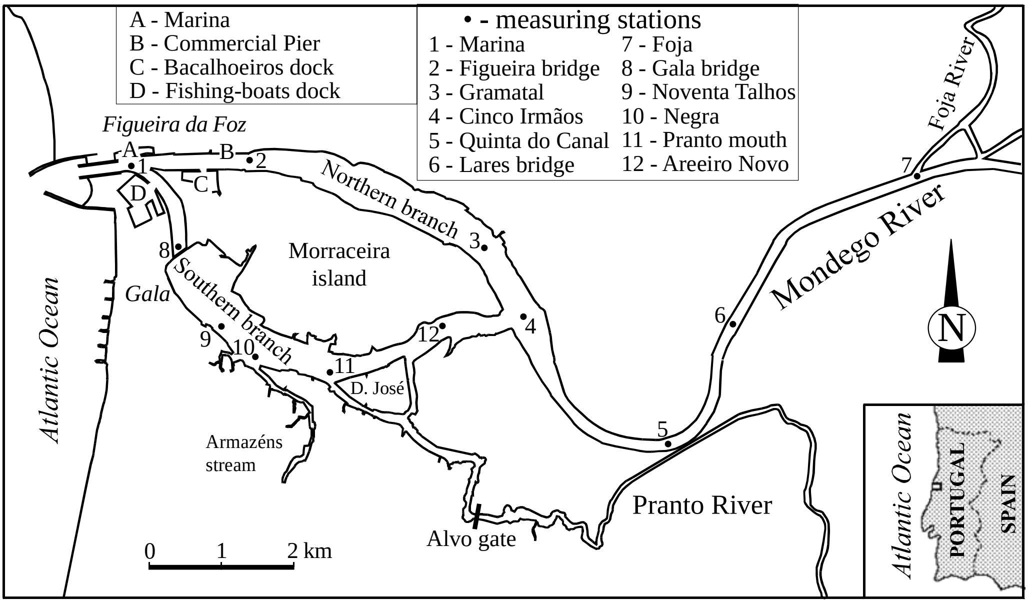

Figure 21. Map of the Mondego estuary (except a small portion of the upstream sector) (Cunha *et al*, in press).

In the Mondego subsystem, four major geomorphological zones are distinguished (Fig. 23); (Cunha et al, in press): Mouth Complex, Sandy Bay, Tidal Flats and Upper Channel. These zones have been extensively dredged and modified (construction of docks and artificial banks). The deepening by dredging and narrowing by embankment improved the upstream penetration of saline water and marine sands. In the channel the fluvial sediment transport is expressed by a grain-size decrease towards the mouth (gravel to fine sand), but an inverse variation results from the tidal flood currents. Fine sediments, like mud and muddy very fine sand, are accumulated on areas of reduced hydrodynamics of the channel margins and in the tidal flats (Fig. 24).

As in the previous subsystem, the Pranto one can be segmented into four major geomorphological zones: Mouth Complex, as an inlet channelled by training walls; Sandy Bay with a large flood-tide delta, sand and mud flats, marshes, and tidal creeks; well represented Tidal Flats and Upper Channel. It is typically fully mixed but during high fluvial discharges, rare nowadays, it probably grades to stratified. In a tidal cycle the salinity variation is high, mainly in the central sandy bay, but local damming of waters by muddy sand bars was documented. The seaward area of marine influence is dominated by sand with some shell gravel, whereas the upstream area is mainly muddy (Fig. 24). The freshwater and sediment inflow from the Mondego River to this subsystem occurs only during seasonal floods.

Figure 22. Mosaic of vertical aerial photos of the Mondego estuary (comprising the 9km downstream) in a spring low tide of 1947 (summer). The human interventions were reduced and limited to the mouth area. Notice the general silting of both estuarine branches.

The estuary supports fishing and commercial harbours, industrial activities, aquaculture farms and salt-works. It receives nutrients and chemical pollutants from the drainage of cultivated areas and until recently, urban wastewaters were discharged into the estuary without treatment.

Successive anthropogenic interventions led to an important artificialisation of the landscape, including rapid expansion of urban and industrial areas over the estuary and adjacent coast, development of aquaculture and rice culture, but decrease of traditional salt pans exploitation (“salinas”) and other agricultural types. In particular, embankment and reclamation greatly reduced the natural areas, and changes on the hydrodynamics were also promoted by landfills of intertidal flats and dredging.

The estuarine area of the Mondego River is an important tidal wetland of the Portuguese coast, and constituting an ecosystem characterised by high levels of organic productivity and biodiversity. The estuarine southern channel (Pranto subsystem) and its vicinities (including the Morraceira Island) are one of the last areas of the lower sector of the almost “natural condition” Mondego river, or having traditional human activities with low environmental impact, such as fishing and salt-works (“salinas”). However, in the Morraceira Island, large areas of “salinas” are being transformed into “fish-farms”. The sedimentary sub-systems, having particular hydrodynamics and sediments, are an important natural heritage, a beautiful landscape and constitute biological specific substrata. The “salinas”, salt marshes and intertidal mud flats are extremely valuable for breeding water birds, shorebirds and other migrants.

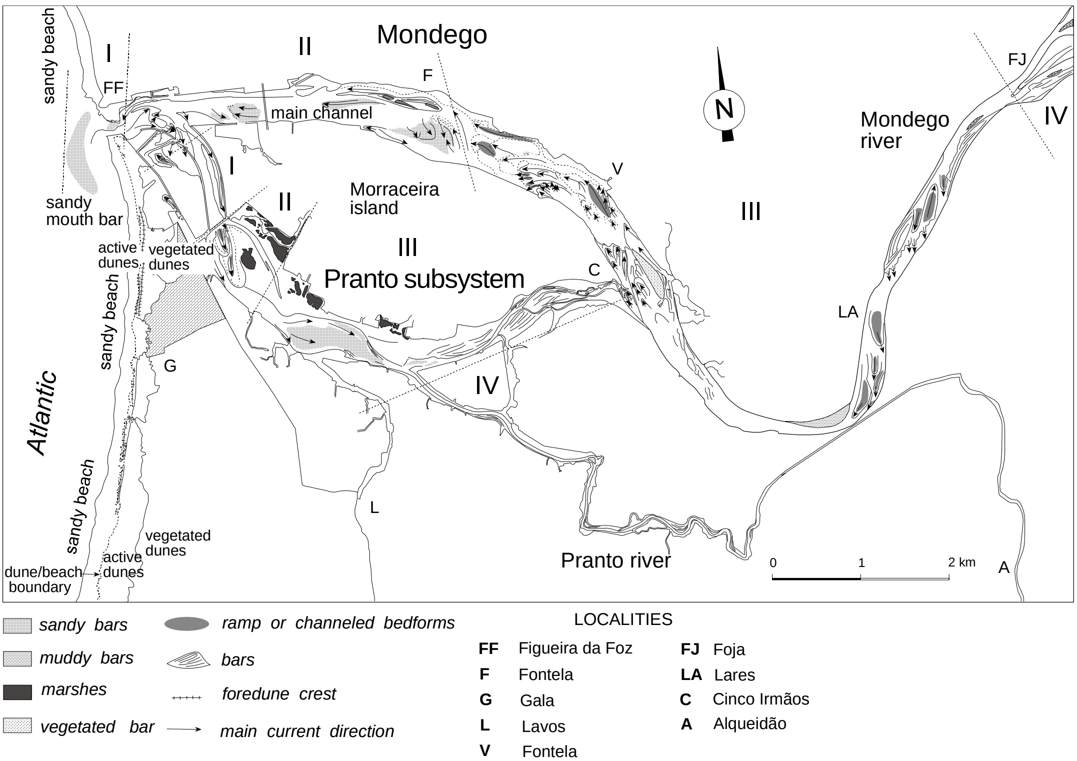

Figure 23. Longitudinal geomorphological zonation of the Mondego and Pranto estuarine subsystems in 1947, comprising four major segments (authors: German Flor and P.P. Cunha; Cunha *et al*, in press): I) Mouth Complex; II) Sandy Bay; II) Mudflats; and IV) Upper Channel. Arrows show the main currents deduced in the spillover lobes, responsible for the sediment distribution by tidal or fluvial flows. The Mondego subsystem is dominated by fluvial discharge, while the Pranto subsystem is tide dominated.

Synthesising the evolution of this estuarine system during the last decades, several major consequences must be stressed. Contrasting with the intense silting mainly with fluvial supply, still obvious in 1947 (Fig. 21), the later severe reduction of sediments in the estuary resulted from the upstream retention by dams and the removal by sand mining and dredging. Moreover, the dynamics of the system was reduced by stabilisation of the fluvial discharge, as well as tidal prism decrease due to successive landfills of intertidal flats. The inlet narrowing and reduction of the upstream connection with the Mondego subsystem accelerated the silting in the Pranto subsystem. For harbour maintenance and enlargement, large volumes are removed from the sedimentary system (see data from the Port Authority of Figueira da Foz and environmental agencies, in Cunha and Dinis, 1998). The trap of sandy sediments with marine and fluvial provenance in the Mondego subsystem is a feedback effect of the intense dredging and energy reduction, but it does not balance the fluvial sediment starvation and the removal of sediments. The aforementioned fluvial and coastal heavy engineering works had major environmental impacts, namely the drastic reduction of the sedimentary inflow from the Mondego River to the littoral. This inverted the coastal progradation caused by man-driven high discharge of river sediment during the last centuries.

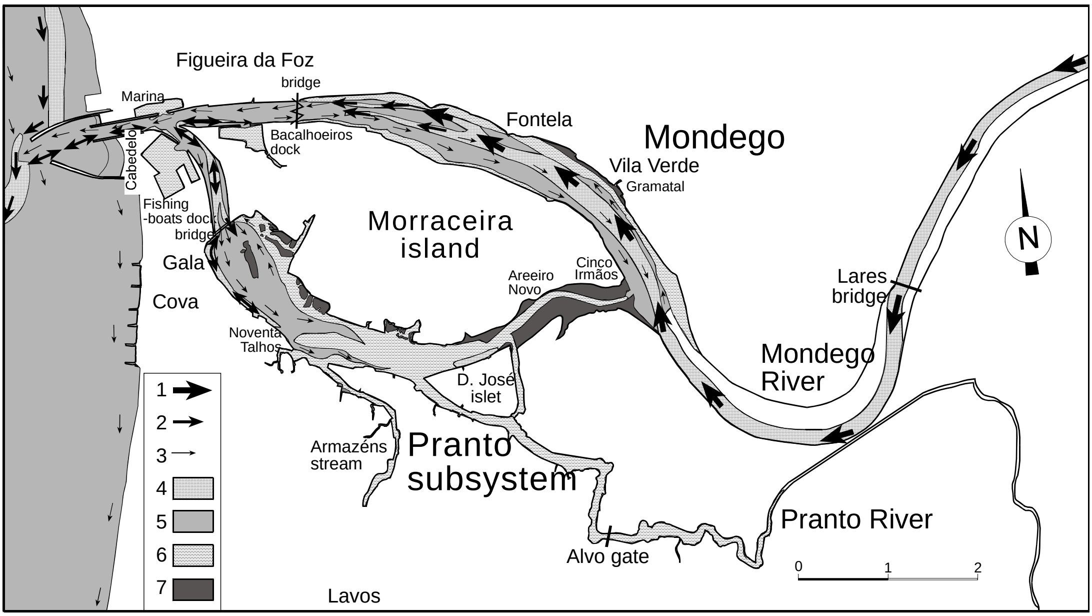

Figure 24. Grain-size distribution and sediment circulation scheme in the Mondego estuary and adjacent beaches (Cunha et al., in press): maximal transport capability of traction currents (1 - up to granules, 2 - up to very coarse sands, 3 - up to medium sands); superficial sediments mean grain-size (4 - granules to coarse sand, 5 - medium to fine sand); tidal mudflats, muddy and channel bottoms (6); saltmarshes (7).

3. Cova beach

In the beaches adjacent to the estuary the longshore drift is reduced and persists towards the South from May to October, but during the remaining months the transport is important in both North and South directions (Vicente, 1990), the annual resulting is towards South. The jetties built in 1965-67, to stabilise port access, produced huge effects on the coastal morphology and sand drift: large accumulation of sand updrift against the northern jetty led to sand mining, while the erosion hit the beaches located southward. The Cova beach documents intense erosion, already reaching the foredune, compelling to the construction of groins and seawalls (Cunha et al., 1997; Cunha and Dinis, 1998).

The beaches consist mainly of coarse to medium sand and the associated dune fields have dominantly fine sands. The influence of wave dynamics is the main control on beaches and estuary mouth grain-size, morphology and evolution (Carvalho and Barceló 1966; Cunha and Dinis, 1998). The narrow beach south of the river mouth is mainly medium sand and nourished by both the longshore current and the local erosion of the aeolian dune field (Fig. 22, 23 and 24).

Comparing with the dune field located North of the Serra da Boa Viagem relief, the dune system extending South of the Mondego estuary has a quite similar stratigraphy and morphology, allowing the presumption of an identical timing and nature (Bernardes et al, 2001). Over the oldest and deeply podsolised generation, the second unit of dune sands includes a thin level of estuarine lagoonal mud and muddy sands rich in organic matter and shells. Radiocarbon dating of samples collected around 10km of Figueira da Foz yielded an age of 2950±100 BP for a peat sample and 2060±90 BP for lamellibranchs and gastropods (both calibrated ages; Bernardes et al., 2001).

The Figueira da Foz – Nazaré dune field has a maximum width of 9km and includes several dune bedforms. Its evolution results from several natural controls, but recently mainly from anthropogenic procedures. The successive phases of the pine forest expansion were the most important factor to avoid dune migration (André, 1996; André et al., 2001). Currently, the deterioration of the primary dune is expressed by frequent blowouts, initiated mainly by pedestrian access, and overwashes. The dune field in this area experiences a rapid reduction due to urban/industrial expansion and is also affected by waste disposal, sand mining, etc. (Cunha, 1998).

THE PONTA DO FACHO LANDSLIDE EVENTS AND THE PRESENT EVOLUTION OF ROCK COAST BETWEEN LAGOA DE ÓBIDOS AND NAZARÉ[5]

Mário Neves

The geomorphological study of a 2km long coastal stretch, between Ponta do Facho and Praia da Gralha (Fig. 25), allowed the identification of the different phases of the erosive cycle that characterizes the present evolution of the rock coast sectors that stretch between Lagoa de Óbidos and Nazaré.

A big slope movement event, the main cliff retreat process, was chosen as a mark, simultaneously of the end of a cycle and the beginning of another one.

Geologically, the study area is carved in an Upper Jurassic alternation of layers with different permeability and resistance, mainly composed of yellowish or reddish fine grain sandstones and greyish or reddish marls. This ensemble is inserted also by limestone and clay layers. All the outcrops dip between 22º and 24º to N330 - N333º and are crossed by a set of faults with azimuth N132º, near Ponta do Facho. As the coastal line between Peniche and Nazaré is mainly exposed to the NW, rock coast sectors in this stretch commonly present a seaward-dipping attitude.

Between January and March 1954, at Ponta do Facho, a slope movement displaced a calculated 171 000m^3 volume of material and caused a top of the cliff retreat between 120 and 200m (Fig. 26). This

[5] **7th stop** – Salgados-Facho -Concha de S. Martinho.

event, reported at that time in the newspapers with some emphasis, was only studied in detail by Neves (2004).

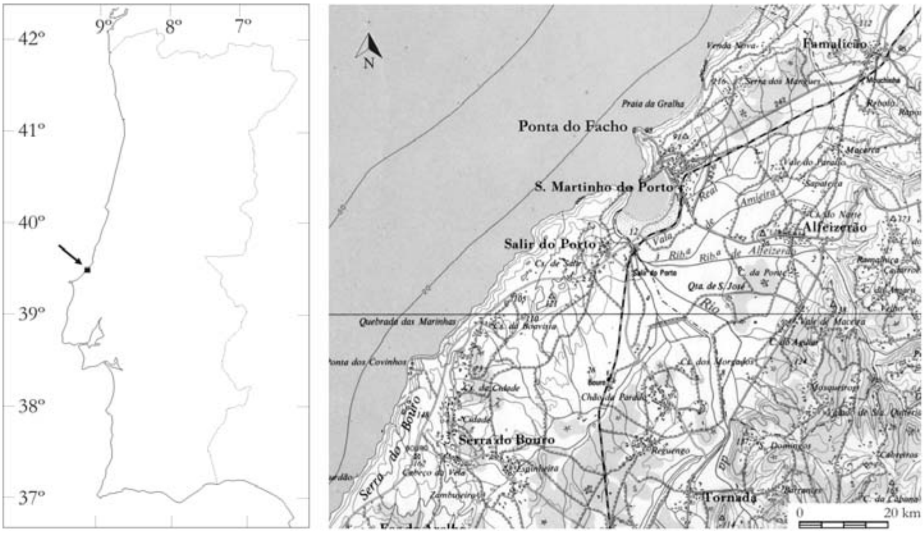

Figure 25. Localization of Ponta do Facho-Praia da Gralha studied area.

The movement reconstitution has allowed the identification of two intercepting failure surfaces composed by a marl-clay layer (23ºN330º dip), located at a maximum depth of 23m and a sub-vertical N132º azimuth fault (Fig. 27). According to the Hutchinson (1988) classification, described in Dikau et al (1996), it is possible to label this movement as a translational slide with wedge failure. Associating the geological characteristics with the topography observed in the photos, taken prior to the event, allowed the individualization of the morpho-structural conditioning factors that are favourable to the occurrence of translational landslides, as was recognized, among others, by Varnes (1984), Sorriso-Valvo et al (1996), Zêzere et al (1999) e Tsige et al (2001):

- outcrops of permeable materials over impermeable layers, stratigraphic sequence that improves the rain waters' infiltration, allowing the accumulation of these waters in the impermeable level;

- sedimentary structure dips towards the slope;

- high slope degree near the base, favouring there a shear stress increase;

- sub-vertical fault that promotes the rain waters infiltration and, therefore, an increase in water pressure and a shear strength decrease, not only along the discontinuity, but also on the argillaceous layers;

- the existence of two different sets of discontinuities, which interception dip toward the slope, is favourable to the occurrence of translation
al slides with wedge failure.

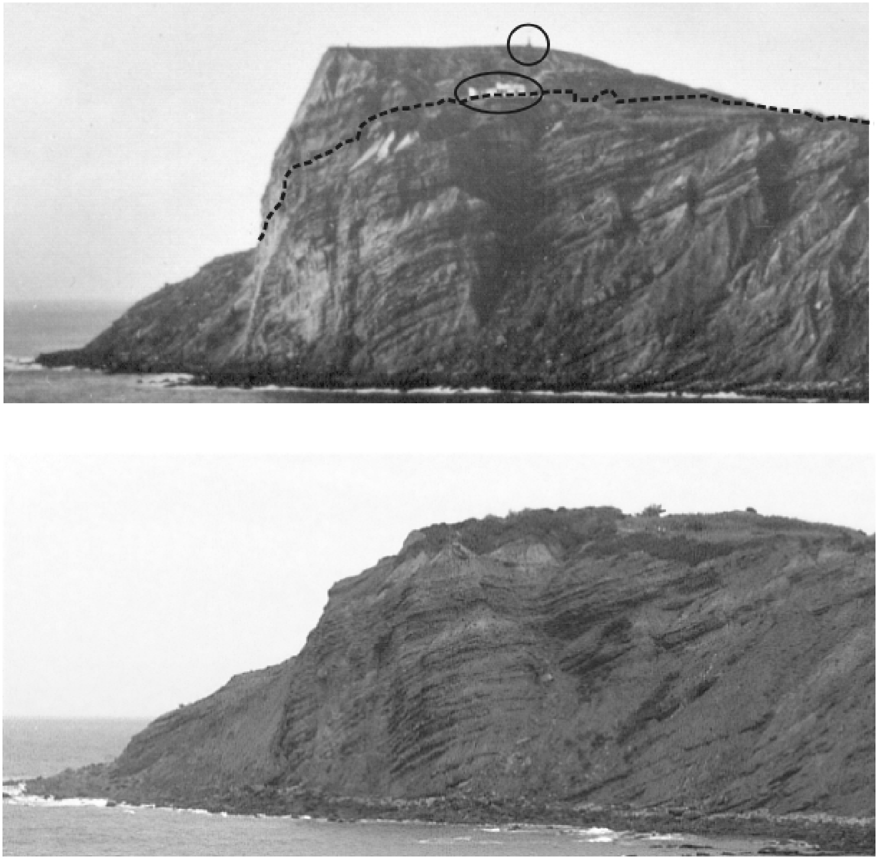

Figure 26. Ponta do Facho in: 1–1952/53 (©Foto Beleza); 2–September 2003 (©MNeves). In photo 1, the dashed line indicates the present topography, the circle identifies the geodesic landmark of Facho 2º and the ellipse points out the position of the Facho Café.

The daily rainfall data[6] points out that the beginning of this landslide took place in a period with low or no records of rainfall. Nevertheless, as Zaruba and Mencl (1982), Gostelow (1991) and Van Asch and Nieuwenhuis (1994) concluded, in deep landslides, there are two main factors that control the gap between the rainfall episodes and the event initiation: the permeability of the more superficial layers and the failure surfaces' depth. This gap may take weeks. According to this, the 40 days accumulation rainfall is the parameter that best adjusts to the beginning of the instability period at Ponta do Facho cliff (Fig. 28).

[6] Recorded at the udometric station of *Alfeizerão,* located 4km ESE from *Ponta do Facho* (fig. 25).

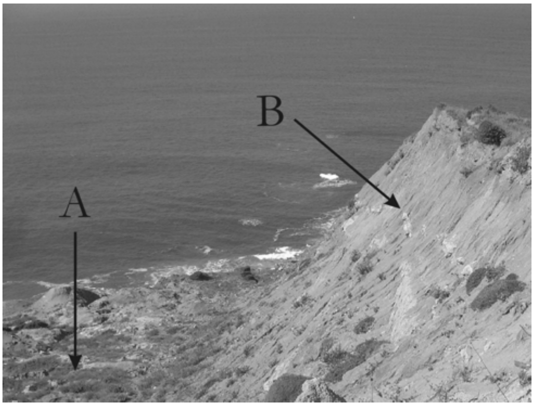

Figure 27. Failure surfaces of 1954 Ponta do Facho landslide. A – marl-clay layer; B – fault with azimuth N132º. Photos taken in September 2003.

Moreover, it must be stressed that the previous climatological year was the dryer of the studied period 1951-2001, with a total less than half the average rainfall in the Alfeizerão station. The limited rainfall records joined with the normal high summer temperatures led to a pronounced water loss in the soil. In those circumstances, according to Mariolakos (1991), the drying of the clayey materials causes a retraction that creates multiple micro fissures, which ultimately improves the water pressure in the following wet period.

Landslide reactivations were identified subsequently, in 1981 and 1996 (Neves, 2004). They unstabilize different sectors at Ponta do Facho (Fig. 29), highlighting the crown of the 1954 movement, that was affected by a retrogressive landslide.

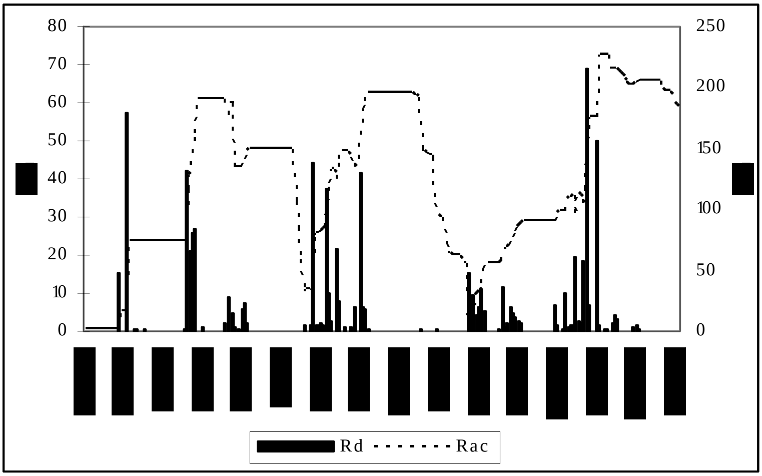

Figure 28. Rainfall, daily (Rd) and accumulated for forty days (Rac), at Alfeizerão between September 1, 1953 and April 15, 1954. The shadowed area indicates the activity period of the 1954 Ponta do Facho landslide.

A landslide event causes an accumulation of debris that, in a first stage, protects the cliff base from the wave's erosive attack. However, in the most exposed sectors, the removal of those materials can be relatively rapid, given its low resistance to wave erosion. The analysis of the 1958 aerial photographs allows the identification that, at Ponta do Facho, four years after the slope movement, the majority of the accumulated material was removed by the waves, remaining only a small accumulation with 70m length, by 25m maximum width. On the contrary, upnorth, at Praia da Gralha, a great part of the debris, resulting from an older similar landslide that displaced more or less the some volume of material, is still in place, because here, the cliff's base is partially protected from the waves attack by the headland of Ponta do Facho. Simultaneously, with the wave work at the cliff base, near the top, erosion carries out the clayey material that composes the new slope surface as a consequence of the landslide.

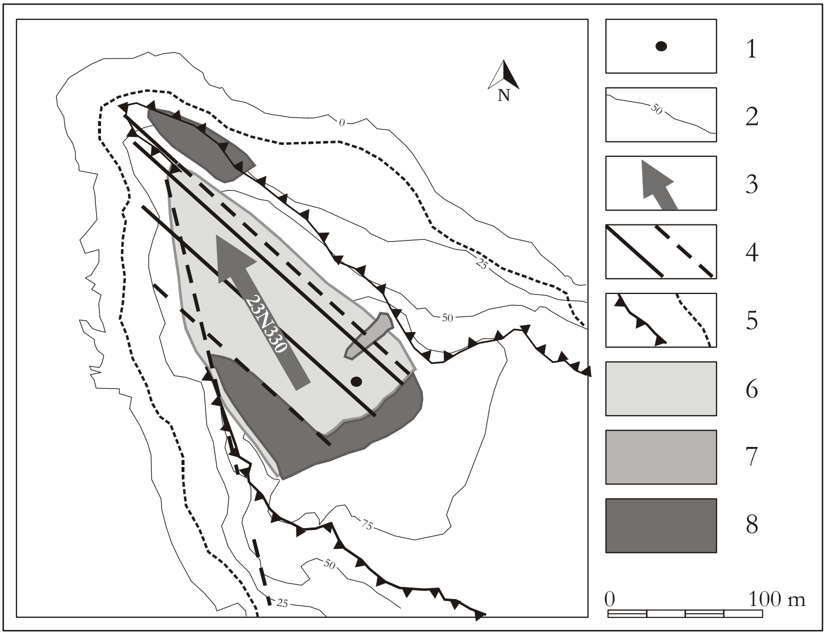

Figure 29. Geomorphological sketch of Ponta do Facho. 1-lost geodesic landmark; 2-contour lines; 3-sedimentary layer's dip; 4-tectonic discontinuity: certain, probable; 5-cliff top and base; 6-1954 landslide; 7-1981 landside; 8-1996 landslide

Those two erosive actions cause the progressive outcrop of the more resistant sandstone or limestone layers. At this stage, the cliff presents a rectilinear sub-structural profile. The continuation of the wave erosion action, widening and deepening the pre-existent discontinuities in the sandstone and limestone layers, and exploring the fragility of the underlying marl and clay layers, contributes to an increasing slope degree at the cliff base. At the some time, it creates caves with variable dimensions that cause an increase on the unstabilized rock block shear stress. On the remnant slope, the discontinuities that affect the sandstone or limestone layers promote the rainwater infiltration. That's to say, there are again morpho-structural conditions that can lead to new landside events.

The extent of a complete cycle is variable according to morpho-structural characteristics and wave exposition of each place. However, the temporal data associated with the dynamics of some spots studied in this coast, point to a more than one hundred years duration for each cycle. This apparent slow evolution, nevertheless, is not reassuring, because what must be emphasized is the fact that landslide events tend to cause very high cliff retreats.

This concern should be present in the development of management plans, with special emphasis in the delimitation of the expansion urban areas. That is not always the case, as some new urbanizations, developed near Pico do Facho, clearly demonstrate.

ESTIMATING THE SHORE PLATFORM DOWNWEARING RATE AT RIBEIRA DE ILHAS USING A TMEM. FIRST RESULTS[7]

Mário Neves

A Traversing Micro-Erosion Meter (TMEM) was conceived and used, aiming to increase the knowledge of the shore platforms present evolution, particularly, the identification of magnitude and frequency of the different processes involved as well as their relative weight.

The entire characteristics of this equipment are described in detail in Neves et al (2001). Its conception follows the main features of the previous TMEM used by Trudgill et al (1981), Spencer (1985) and, specially, Stephenson (1997) and Stephenson and Kirk (1998). However, some modifications were inserted to allow the increase of monitorized points in each measurement Area and the attainment of a higher reliability of the results. In consequence, it is possible to evaluate periodically the topographical position of 255 points spread regularly in a triangular area of 116,7 cm^2, with an accuracy of 0,005mm (Fig. 30).

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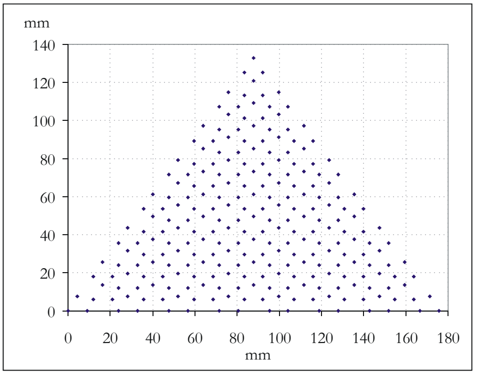


Figure 30. 1-TMEM installed in one of the studied areas; 2-Relative position of the 255 points measured by the TMEM.

[7] **8th Stop** – Ribeira de Ilhas.

Three spots in the Portuguese Estremadura coastal area were monitorized, between May/June 2001 and May/June 2003, with a biannual periodicity NEVES (2004). Here, the results in one of those spots – Ribeira de Ilhas – are analysed.

The most extended shore platform of coastal Estremadura lies NNW of Ribeira de Ilhas beach. It has a length of about 1200m and a maximum width of 120m. This rock coast sector, exposed to WSW (N245), is cut in lower Cretaceous material mainly composed of an alternation between relatively thin layers (< 0,9m) of marly limestones and marls presenting often an internal structure organized in small slabs (thickness <1 cm). These outcrops have a general monoclinal disposal with a weak inclination towards South (4ºN180º). The contact between the cliff and the shore platform occurs less than 2 meters above maximum high water spring level. Sometimes, an accumulation of sands or rock fragments heaps here, or a little lower, being removed in high water periods of storm situations. The shore platform has a superior 1,4º-1,7º slope between the cliff foot and the high water neap level, due to the stepped profile in connection with the geological structure, and a lower gradient (1,1º) downwards.

Four TMEM measurement Areas were set along a profile (RI1) with different exposure periods to marine and sub-aerial processes (Fig. 31). The selected Areas (Fig. 32) have an analogous substratum cut in marly limestone with a sedimentological composition where carbonates prevail (88-93%), presenting also 7 to 10% clays and residual values of sands.

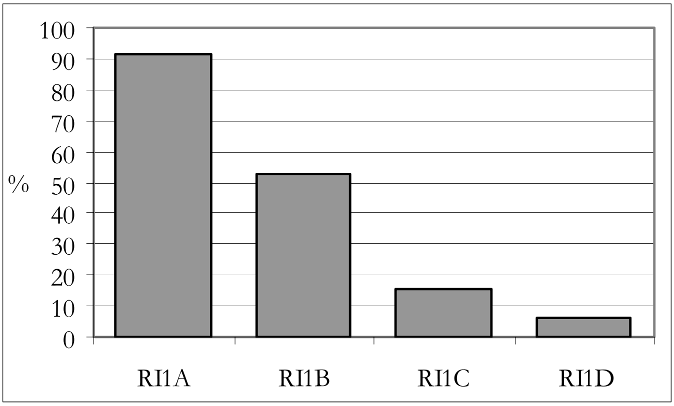
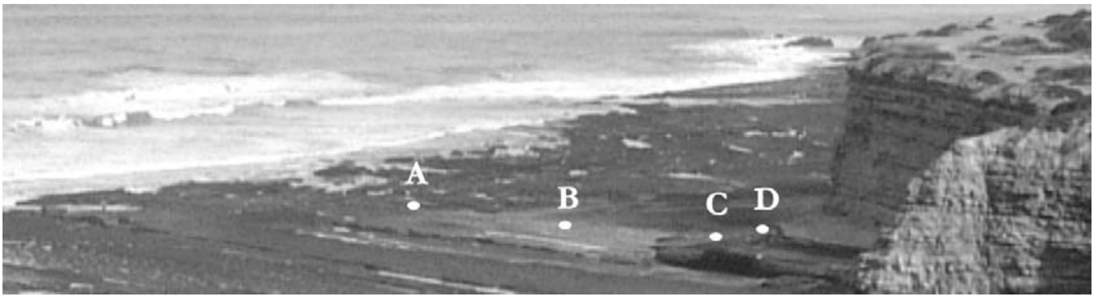


Figure 31. Annual submersion period for each TMEM measurement Areas at Ribeira de Ilhas (in %).

The almost complete biological coverage of the lower intertidal shore platform if, on one hand, didn't allow the TMEM measurement Areas installation, on the other hand, indicates a present geomorphological dynamics clearly influenced by biochemical and biophysical processes.

The comparison between the downwearing rates of the four measurement Areas for the ensemble of the monitorized period[8], show that the average rates are fairly similar (Fig. 33). Three of these Areas - RI1A, RI1B e RI1D – present results with differences inferior to 0,15mm year^{-1}, leaving only RI1C with maximum divergence of 0,257mm year^{-1}.



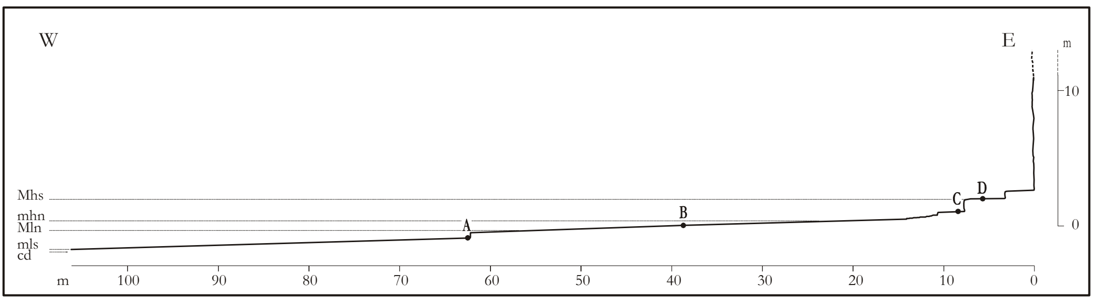


Figure 32. Localization of the 4 TMEM measurement Areas in the Ribeira de Ilhas shore platform: above–on a photo taken the 17/04/03 (tide at 0,3m a.c.d); below-on the Ribeira de Ilhas topographic profile, with reference to the tide levels. Mhs-Maximum high water spring level; mhn-minimum high water neap level; Mln-Maximum low water neap level; mls-minimum low water spring level; cd-chart datum.

On RI1A and RI1B, located respectively at 0,9m and 1,9m a.c.d., that means, at a lower position regarding mean sea level, the regularity of the results, with small dissimilarities Summer-Winter, must be highlighted (Fig. 34), pointing out to a quite equilibrium situation between weathering processes, predominant in summer, and wave erosion, more active in winter months. The rougher wave climate that characterized 2002/03 winter months (period 4-5) reflected mainly in the extreme values recorded, with 5 measurement points surpassing the maximum values of the previous periods and also, on the superior standard deviation, showing a greater data spreading.

[8] Remark that RI1C semester downwearing rates concerns only to the last period (4-5) - Winter 2002/03 - because for the most part of the remaining time it was covered by a sand accumulation that obstructs the measurement procedures.

Swelling values were also registered. In this particular case, they were imputed to the rock coverage by endolithic lichen Verrucaria maura (Fig. 35).

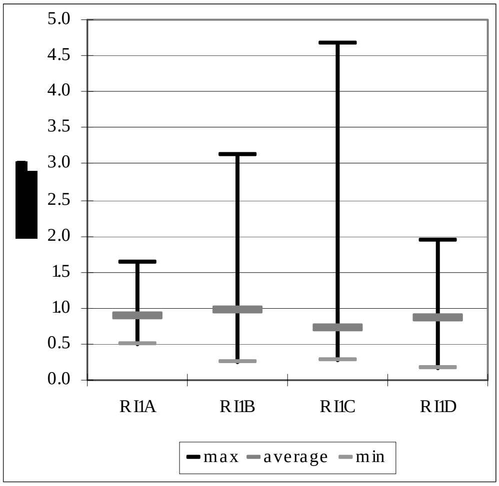


Figure 33. Anual downwearing rates for Ribeira de Ilhas TMEM Areas: absolute maximum (max), average and absolute minimum (min).

The RI1C results emphasize the occurrence, in the monitorized period (winter 2002/03), of two processes with distinguishable magnitude and frequency: one rather continuous downwearing of smaller magnitude, related with wave erosion and the sand presence; and the slabs' detachment that originate a sudden lowering of the shore platform (Fig. 36).

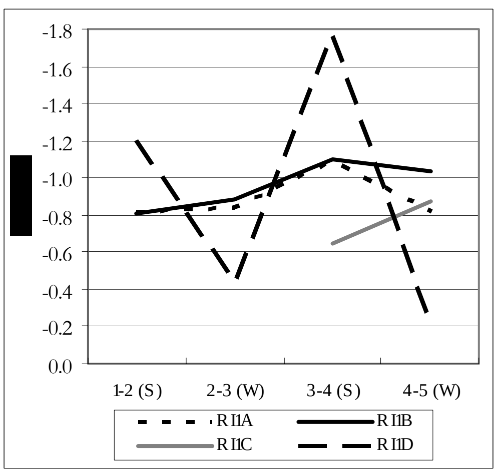


Figure 34. Anual average downwearing rates for Ribeira de Ilhas TMEM Areas in each measurement period: S-summer; W-winter.

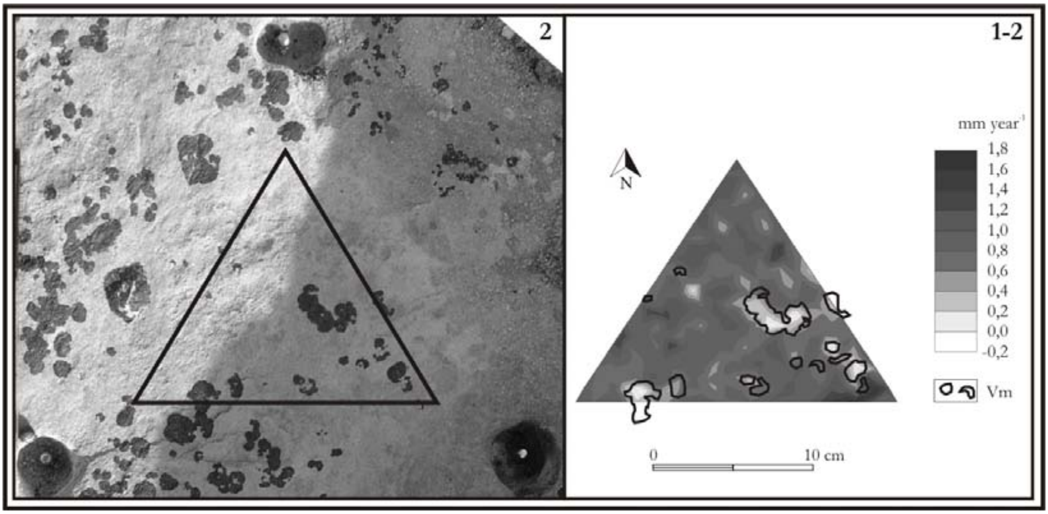


Figure 35. Influence of the Verrucaria maura (Vm) distribution area in the swelling results (measurement 1-2 – summer 2001).

Analysis of measurement Area RI1D points out to a remarkable difference between summer and winter data, with the first one reaching proportions 2,8 to 9,1 times superior than the winter results (Fig. 34). This superior summer lowering of the shore platform at this level occurs also on minimum and maximum results. More reduced differences (average of 2 to 3 times superior, reaching 6x on hotter summers) were also achieved by Mottershead (1989), on his seven years research on a south England supratidal platform. Nevertheless, it must be said that coastal environmental contrasts between summer and winter months are not so striking in that region as they are on the Portuguese coastal area, justifying, in a way, the more pronounced differences registered at RI1D measurement Area. The obtained data indicates a predominance of summer active processes of downwearing, namely weathering processes. Beyond that, the spatial distribution of the downwearing rates of the four periods shows that, regarding the second summer, there is a strong tendency to the occurrence of superior rates at the same points affected during the past summer. This fact can be explained by small differences in the marly limestone composition that can justify the existence of more fragile sectors to the action of the processes indicated or to the slaby marly limestone inner structure that can benefit the concentration, particularly of salt weathering processes, on the increased fragility bands that separate each slab.

The research continuation and the input of new data will permit to complete, with higher precision, the analysis to the first results now produced.

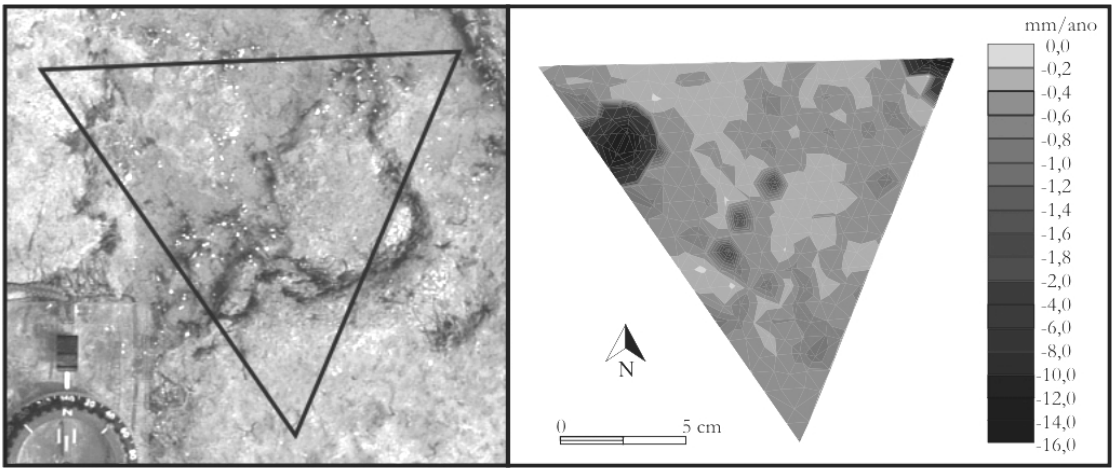


Figure 36. On the left, RI1C measurement Area on 30 June 2003. On the right, RI1C annual downwearing rate registered in winter 2002/03 (period 4-5).

CABO DA ROCA HEADLAND AND ITS SURROUNDINGS[9]

Ana Ramos Pereira

1. Serra de Sintra

Serra de Sintra is the most prominent feature of this littoral area NW of Lisbon (Fig. 37). It is a hipovolcanic massif of granites, syenites, diorites and gabbros. This massif began its uplift under a thick cover of Jurassic sedimentary rocks, mainly limestones and marls, which are still surrounding the massif.

This sedimentary cover with 2200 to 2700m thickness (M.C. Kulberg e J.C. Kulberg, 2000) was inflated by the massif uplift, while the rocks were submitted to metamorphism, and exposed to weathering and fluvial erosion. Since the uplift began and for 30 million years, the erosion destroyed the sedimentary cover in an area circa 150km^2. The crystalline rocks are exposed since 65 million years.
This massif gives birth to a well defined hill – the Serra de Sintra. This feature has characteristic dissymmetric dome shaped, related to the structural domain (Fig. 37). The high slope facing north is related to a thrust, while the southern slope until the Cascais platform is smooth. It has a set of small hills surrounding the Serra and a low plateau (S. Pedro plateau), which corresponds to the metamorphic halo (Ribeiro, 1940).

The fluvial network reaches the sea by suspended hanging valleys, sometimes over 150m a.s.l. This fact, as well as the discovery made by Ferreira (1984) of a beach deposit (sands and rounded small gravels) at 250m, near Ulgueira, points out to the continuous uplift of the Sintra massif (Fig. 38).

The massif uplift rating is estimated to be circa 12.5cm/1000 years.

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[9] **9th Stop** – Cabo da Roca.

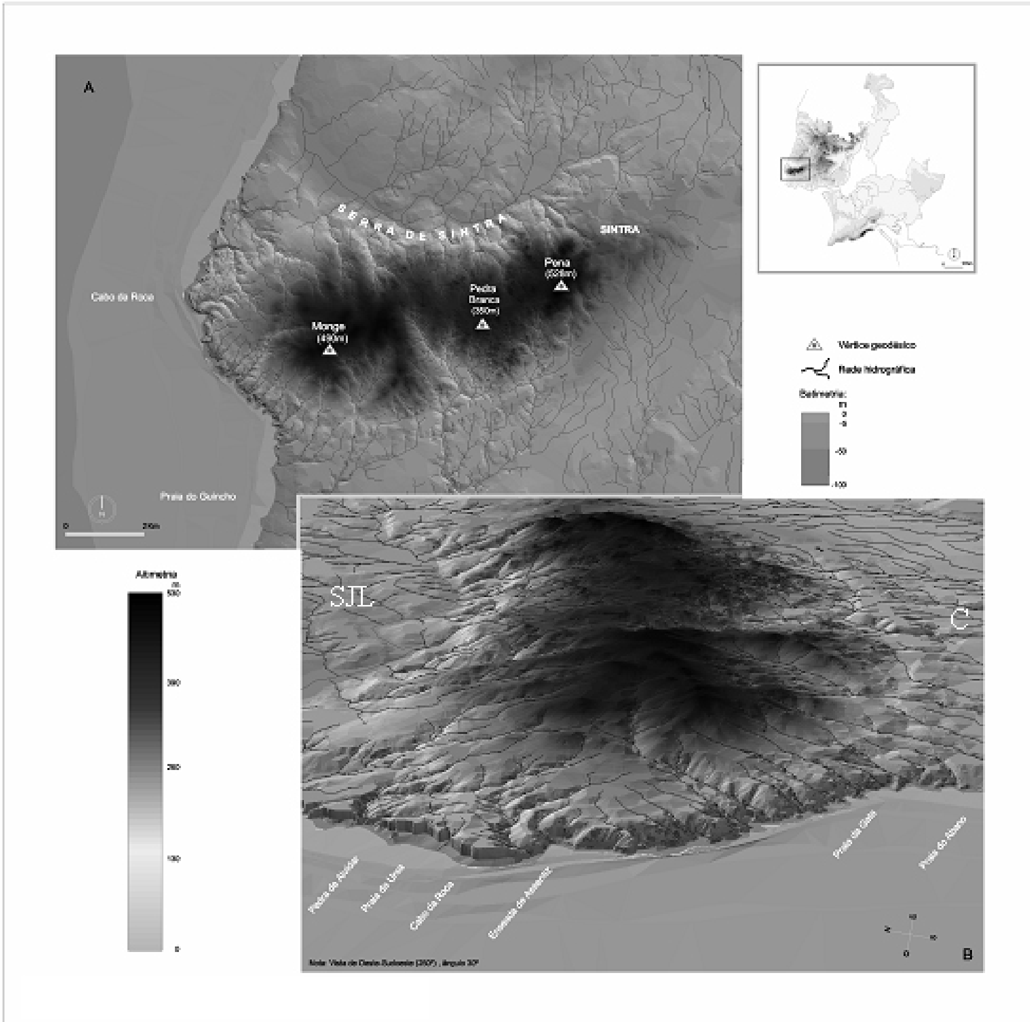

Figure 37. Serra the Sintra and S. João das Lampas (SJL) and Cascais (C). From Pereira, 2003.

2. The Cascais platform

The Serra de Sintra is surrounded by two elements of the littoral platform: (i) the platform of S. João das Lampas in the North, between 100 and 150m a.s.l. and (ii) the platform of Cascais, lower and better preserved between 60 and 80m a.s.l. (Fig. 37).

Such as other elements of the coastal platform, they have complex genesis, probably polygenic, however beach sands and rounded gravel are still preserved, namely in the karst holes of the platform of Cascais, showing a marine ingression. The deposits are much eroded and they have not given any dating elements so far.

Figure 38. Serra de Sintra and its coastline with hanging valleys.

In this cliff coast, the beaches are rare, reduced to very narrow estuary beaches or pocket beaches. This entire coastline has a deficit in sediments because there are no big rivers and the cliffs are mainly cut into marls and limestones. The Guincho beach is the exception. This quite large beach profits from the sheltered bay developed in the contact between the metamorphic and the sedimentary rocks, where some faults have been recognized. This area is down drift to a granular massif that provide sand to the long shore drift, and where refraction promote a shelter bay, where sands deposition are possible. There has been no major environmental changes since, at least, 32 000 years BP, exception made to the width of beach and dunes belt during low sea level. Over the karst developed in Cascais platform, a palaeosoil in aeolian sands has been dated. The ensemble is sealed by a carbonate dune, built up by north-westerly winds, forming a partially eroded hill, 40m above Cascais platform (Pereira and Angelluci, 2004).

THE WESTERN LITTORAL OF THE SETÚBAL PENINSULA[10]

Ana Ramos Pereira

1. The littoral landforms

The coastline south of the Tagus estuary is an arc shaped sandy coast. Despite the predominance of a cliff dominated coast in the Estremadura littoral, this area has a 25km long sandy beach related to the presence of the river's sand supply and the wave refraction of the NW dominant waves promoted by the Lisboa-Cascais promontory (Fig. 39). This promontory refracts the NW dominant waves, which reach the coast from SW and originate the inversion of the longshore drift that here is from the south to the north. This sheltered coastal area, has a wave climate with a significant average wave height

[10] **10th stop** – Capuchos.

between 0.7m and 2.6m and the average maximum wave height between 1.2m and 4.1m, with periods between 5 and 9s, in a spectral direction from WSW to WNW (the Bugio lighthouse). In Cascais tide gauge, the maximum spring tide reaches 3.83m and 0.27m during neap tide.

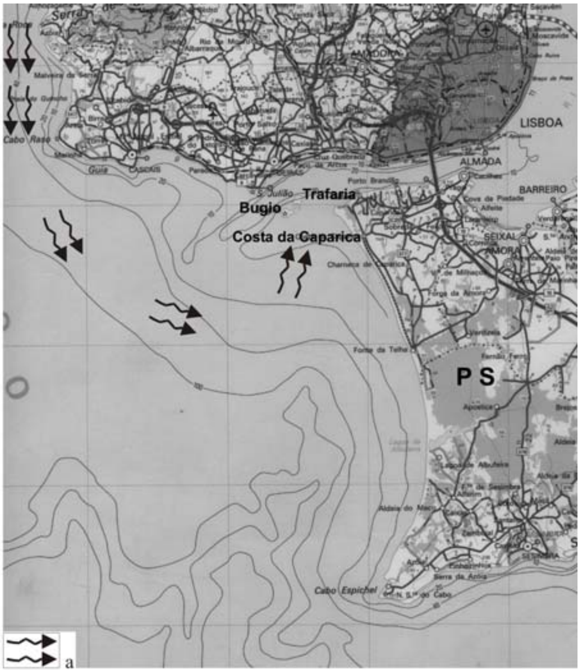

Figure 39. The western littoral of Setúbal Peninsula. a – Longshore drift.

The meso-scale geomorphological units are: the coastal plain of Costa da Caparica; a fossil cliff cut into sub-horizontal layers of Middle Miocene to Quaternary sedimentary rocks (marls and sandstones) and the coastal plain of Costa da Caparica, besides the Tagus river estuary.

2. Tagus estuary

The Tagus is the longest Iberian river and the third in discharge. In the Lower Tagus River, the valley has a sand and gravel plain that reaches 13km width. However, when it approaches the coastline the valley is entrenched and is only 2km width (Fig. 39). This original feature is connected to a structural fault. This faulted area was better known after the works developed for the construction of the distal bridge over the Tagus River (Fig. 40).

A study of the Pliocene sediments in the Setúbal Peninsula has shown the presence of granite and basalt gravels, which came from the Lisbon Peninsula (Azevêdo, 1983). The presence of these lithologies points to the absence of the entrenched valley during the Pliocene. This estuary must be a Quaternary feature.

The Tagus River assures the sediment supply of this littoral area, not only by present-day sediments but also by the sediments accumulated in the continental shelf. In this area it is a prograding accumulated continental shelf.

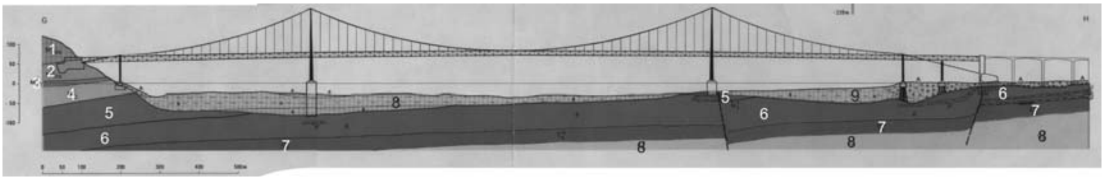

Figure 40. Geologial section of the Tagus estuary. From Almeida, 1986. 1 – Quinta do Bacalhau sandstone; 2 – Forno do Tijolo clays; 3 – Entrecampos limestones; 4 – Estefânia clayey sandstone; 5 – Prazeres clays and limestones; 6 – Lisbon Volcanic complex; 7 – limestones; 8 – Clayey limestones and clays; 9 – Alluvial deposits.

According to INAG (2001), the present-day input of fluvial sediments in the Tagus estuary depends mainly on floods, which are less frequent but with higher magnitude in relation to dam construction and the irregular Mediterranean regime rainfall. The total estuary discharges was estimated to be 1 to $77x10^6$ ton/year, decreasing to 0.4 to $1x10^6$ ton of sediments in dry years. The average sedimentation rates, between 1983 and 1984, points to values between 1.1 to 1.5cm/year. Another fact must be pointed out: the huge amounts of sediments dredged, partly to local artificial sand nourishment.

The velocities of the tidal currents in the Tagus estuary are strong although with low heights. In the spring tide they exceed the 2.0m/s during the flood and 1.8m/s during the ebb.

3. Coastal plain of Costa da Caparica

South of the Tagus inlet, a triangular coastal plain developed. It is the result of: (i) the abundant sand supply, not only from the near mouth river but also from the continental shelf and (ii) the sheltered area where a vortex circulation exists, keeping the sediments from leaving the area. This dynamic equilibrium prevailed till the fifties, when the major erosion problems began.
The sandy coastal plain, narrower to the south, is bound by a straight slope. It has a beach, a foredune and a historical dune field inland (Fig. 41).

This triangular landform is related to the N-S fossil cliff and the refracted wave equilibrium of the arc shape, WNW-ESE. The coastal plain narrows to the South.

The sandy plain was elongated northward by a spit, connecting the coastal plain to the light-island of Bugio (Fig. 39). According to a newspaper mentioned by Freire (1986), in 1910, in the low ebb tide, it was usual to go on foot to the Sunday mass at Bugio chapel (now at about a 3km distance).

Big sea storms took place during the forties and fifties and strong erosion began destroying houses (mainly fisherman houses) built over the foredune and began cutting the spit. A large sandy bar remains in the Bugio lighthouse and a very smaller spit begun its movement towards the interior of the estuary. Meanwhile, the Bugio sand bar moved 700m to the north, between 1939 and 1985 (Oliveira, 1993). A seawall (dike) was built, in 1959, between Trafaria and Costa da Caparica to prevent overwashing and sea flooding (Fig. 39). Afterwards, a groins field was built up to protect this area.

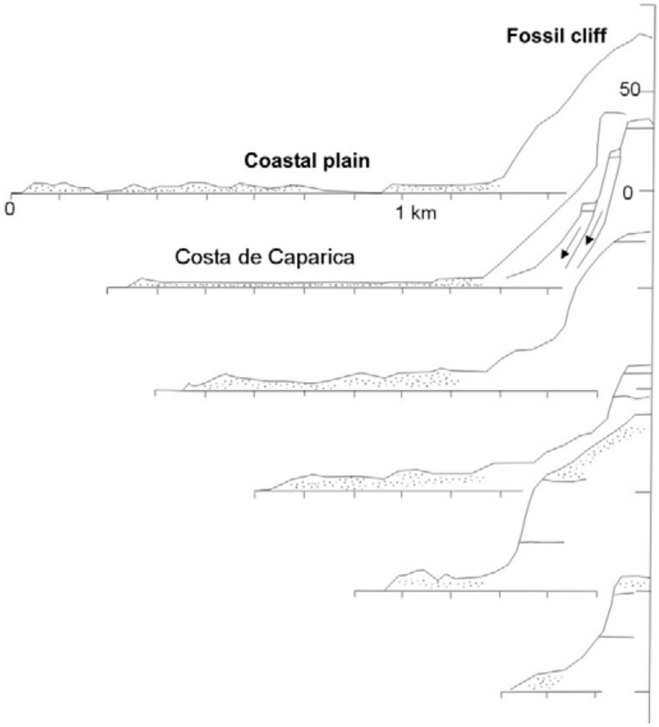

Figure 41. The coastal plain of Costa da Caparica and the fossil cliff (Pereira, 2003).

However, at that time the anthropogenic pressure was still very low. The accessibility to this coast was bad; the only bridge to cross the Tagus River was 30km upstream and the boat was the only way the reach the other bank and not directly to the beach area.

The lack of sediments, the decrease in frequency of the Tagus' big floods and the changes in land usage of this plain modified the complete scenery.

Until the construction of the bridge, in 1966, the river was a frontier difficult to transcend. The land was devoted especially to agriculture, was covered by acacia brushwood to prevent aeolian sands from blowing to the agriculture lands, and had very small fisherman villages and some restaurants near the sea. The improvement of the accessibility, in 1966, promoted the urban growth in this area near Lisbon and the foredune has, since then, been occupied by buildings, restaurants, car parks and pedestrian paths. The groins and the seawall were an essay to stabilize the natural moving coastline (!), when buildings on the dunes (where they still exist) and on the aeolian sands destroyed the transversal equilibrium in the beach-dune system (Fig. 42 and 43).

Figure 42. Costa da Caparica beach before the beginning of the erosion process (Barceló, 1971). View to the SW.

Despite being a sheltered coast, this littoral is exposed to the most morphogenic storms, those coming in from the SW. The erosion proceeds and almost every year the defence structures have to be repaired and the artificial nourishment of the beaches must be done (Fig. 42 and 43).

Figure 43. Costa da Caparica beach during a sea storm in 9/11/2003. View to the N. A. Ramos Pereira.

To access beach erosion in this coastline, Ferreira (1999) studied the storm event of December 1995-January 1996 (21/12/1995 to 15/1/1996). This event was a sea storm from SW, and the estimated[11] wave hight exceeded 8m at 9m deep (there is no buoy in the near sea) (Fig. 44 and 45). This storm was followed by a storm surge however, with low magnitude, 70cm high (Pereira et al, 1997; Fig. 46). Nevertheless, it was enough to approach the wave breaking to the beach and promote several overwashes which enlarged dune breaches, specially created by unmanaged paths.

[11] The wave height was estimated with the model MAR3G_W2 (Pires, 1999).

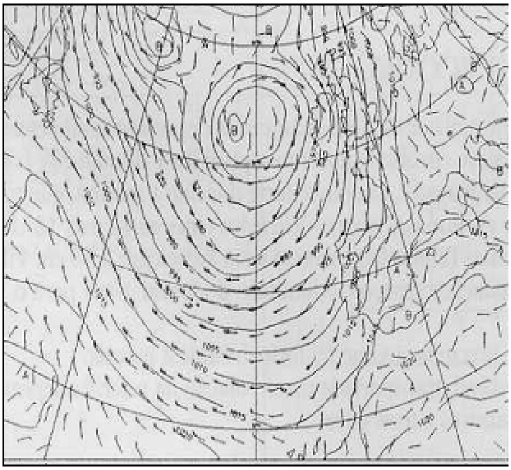

Figure 44. Synoptic chart of 6 January 1996. Daily bulletin of the Meteorological National Institute

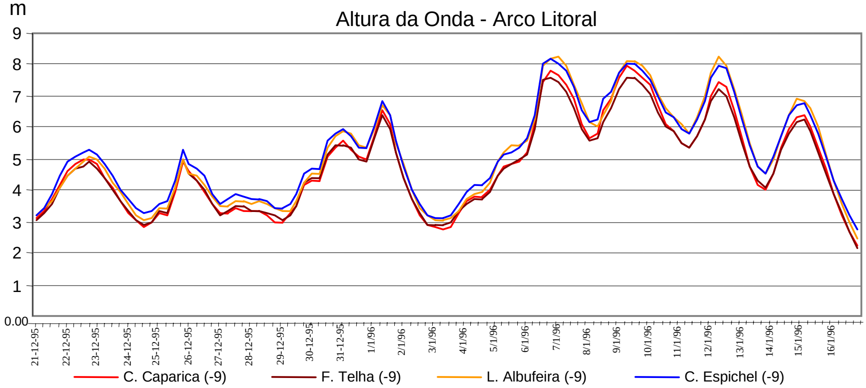

Figure 45. Estimating wave height at 9m deep, from 21/12/1995 to 16 1/1996. (Ferreira, 1999).

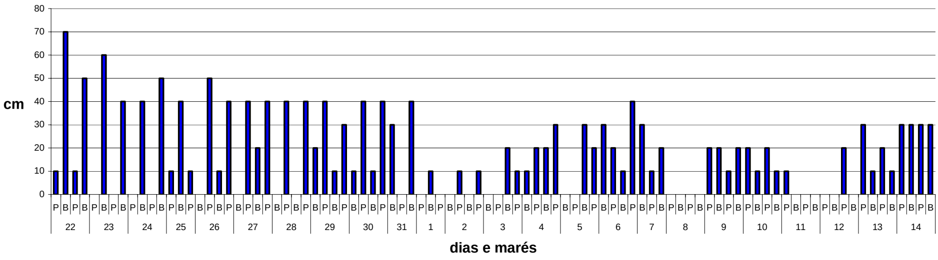

Figure 46. Storm surge during de 22/12/1995 to 14/1/1996. (Pereira *et al*, 1997).

Figure 47 shows what happened during this event in the Northern half of the sandy arc: generalized beach and foredune erosion and several overwashes in the most exposed area, while the heavy defence structures protected, from the overwashes, the beach-foredune system located downstorm, i.e., NE of the groins.

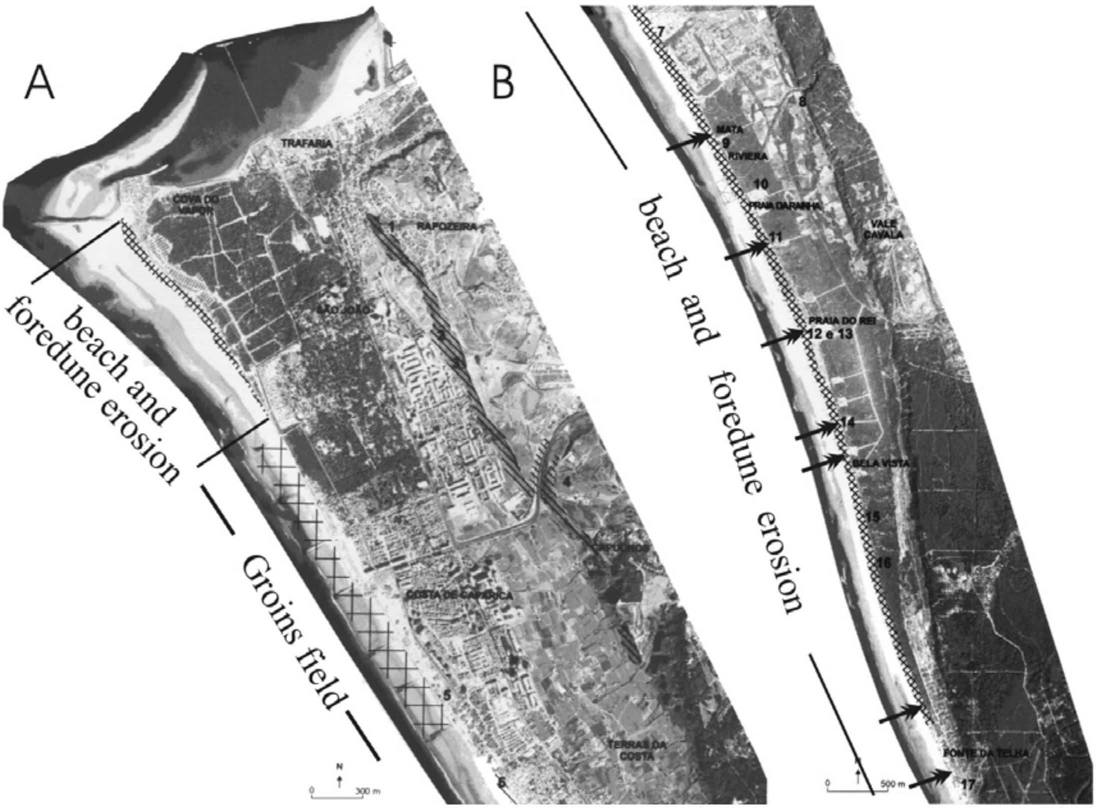

Figure 47. The Northern half of the Caparica arc. The arrows show the overwashes during the December 95-January 96 sea storm event. Adapted from Ferreira, 1999.

We must not forget that this area, now submitted to high human pressure, is subject to another hazard: tsunamis event. However, the agents and those responsible for the land use and management seem to ignore it, and the urban growth is still increasing, in the year we “commemorate” 250 years of the 1755 tsunamis, that killed about 10,000 inhabitants in Lisbon.

Figure 48 synthesises the contrast of the human pressure along this sandy coast.

Figure 48. The coastal plain of Costa da Caparica at the beginning of the 21st century. Photo AML.

3.The fossil cliff of Costa da Caparica

The littoral platform (the local name is Belverde littoral platform) is cut by a scarp slope facing the sea, known as the fossil cliff of Costa da Caparica (Fig. 49).

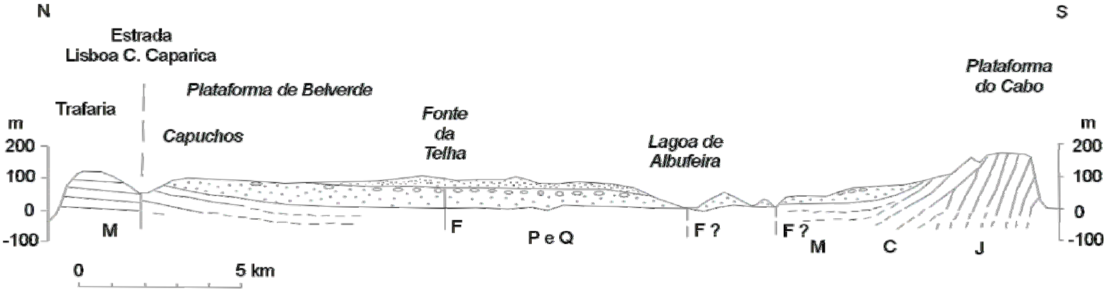

Figure 49. The structure of the rocks cut in the heritage cliff of Costa da Caparica. J – Jurassic; C – Cretaceous; M – Miocene; P – Pliocene; Q - Quaternary (Pereira, 1988).

This slope is and old cliff which is positioned away from the sea by the large accumulation of sediments at its bottom, creating the coastal plain mentioned before. This process of sea sand accumulation was possible by the sediments supply provided by Tagus River. It is not known when the barrier sand system developed (Pereira, 1988), but in the last century, a lagoon opened to the north (Tagus estuary), still existed. The presence of the unhealthy conditions of the marshes as well as the mobility of the aeolian sands generated a disagreeable environment. In the 18th century, King D. João V ordered the stabilization of the aeolian sands by forestation. In the early 19th century it was decided to implement a drainage system to dry these marshes and, at the beginning of the last century, to plant acacia to stabilize the aeolian sands.

The fossil cliff lowers towards the south while the coastal plain narrows (Fig. 41). It is a sensitive area because of the high slope and the low degree of cohesion of the Mio-Pliocene rocks and sediments. The slope has great landslides, as in Capuchos belvedere, and gully erosion (Fig. 42). These two processes – landslide and gully erosion – promote the cliff retreat and it is advisable not to build on the top of the cliff or at its bottom.

In 1984, this area was defined as a protected area (DL 168/84 of 22 of May).

DYNAMICS OF SANTO ANDRÉ COASTAL LAGOON[12]

Maria Conceição Freitas, Aanabela Cruces and César Andrade

1. Introduction

The Santo André coastal lagoon is located in the southern half of the Tróia-Sines coastal bay (Fig. 50A). Tides are semi-diurnal, ranging between 1.5-3.5m and wave energy is high, predominantly associated with northwesterlies (Hs – 1-2m, T – 5-8s during summer, increasing to 2-4m and 10-14s during winter). Sheltering offered by the Arrábida chain added by the arcuate plan shape of the beach and nearshore bathymetry, reduces the net longshore drift to a small figure in spite of the high wave power at breaking which is essentially spent in cross-shore sediment transfer.

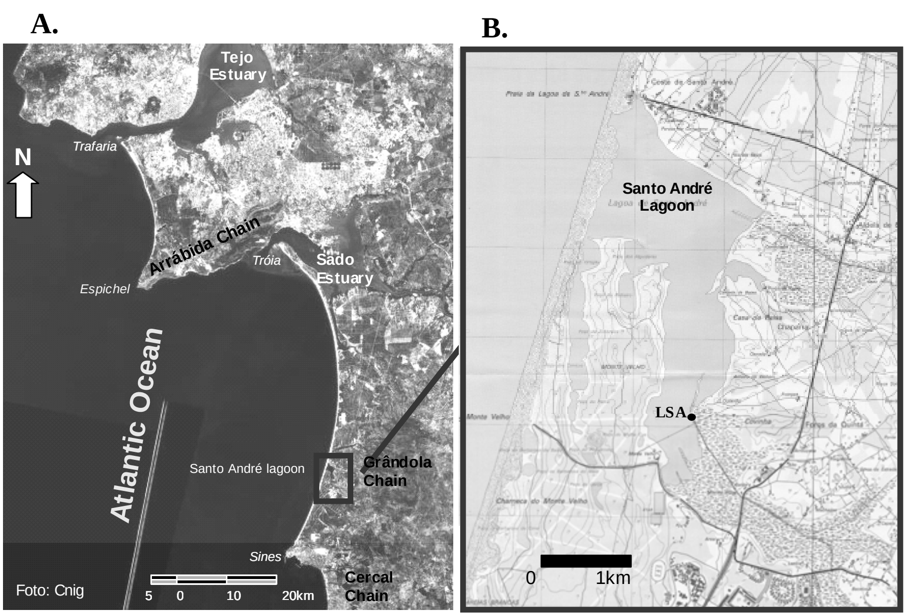

Figure 50. Location of Santo André lagoon.

[12] **11th Stop** – Sto André.

The 2.5km² flooded surface of the Santo André lagoon displays a complex shape including a central basin, which extends southwards through a number of confined N-S elongated swales that are connected to the main body by narrow and shallow channels (Fig. 50B). The lagoonal surface varies seasonally (between 1.7 and 2.5km², exceptionally reaching 3.7km²), as a function of rainfall and fluvial input, intensity of overwash and retention ability of the basin, the latter depending on the tidal regime when an active inlet is present. The average depth and stored water volume vary, with maximum depth reaching 4m in tidal channels. The tributary watersheds extend for 146km² and develop in Paleozoic flysch and Meso-Cenozoic (essentially Miocene to Pleistocene) detrital rocks (Fig. 51).

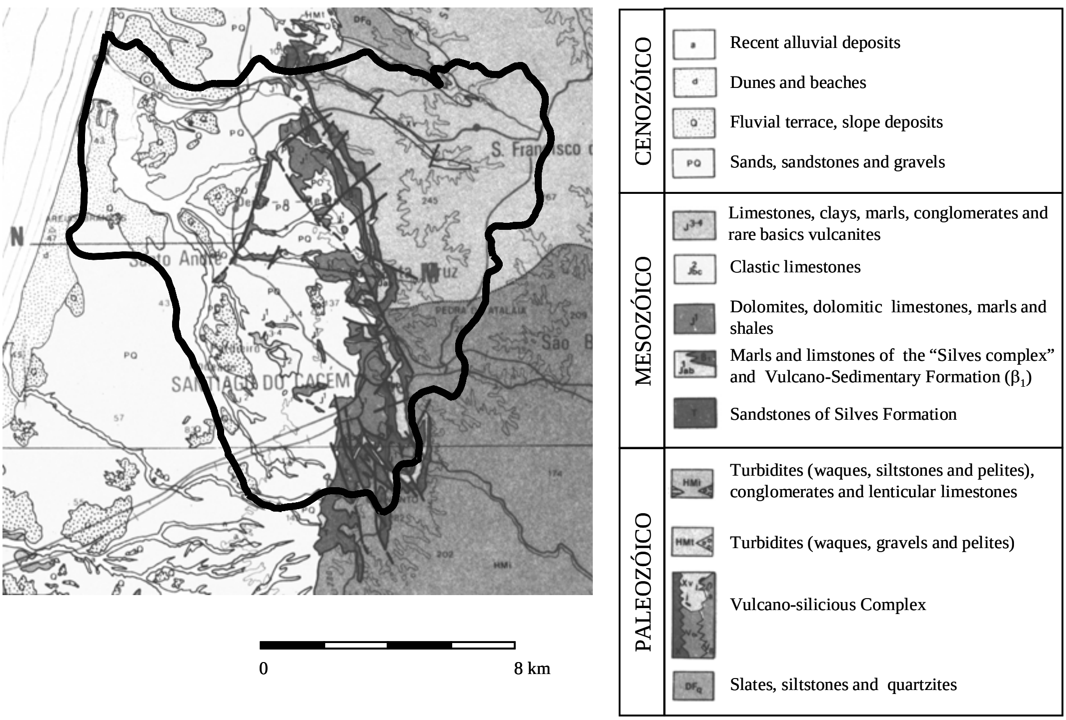

Figure 51. Geological settling of Santo André lagoon watershed.

2. The barrier

The lagoon is separated from the ocean by a continuous 4km-long reflective sand barrier trending N15ºE. It is formed by a beach-active foredune system including numerous washover gaps in the northern tip and its morphological complexity increases southwards where beach and foredune weld and lean against an older and higher, robust and vegetated inner aeolian, ridge, locally scarped. This is one of a complex ridge-and-swale structure trending N-S, rising 25-40m above sea level, which is obliquely intersected by the present-day coastline: Monte Velho W is the most seaward body and further east three other ridges (Caniços-Parral, Barbaroxa and Monte Velho W) (Fig. 52) follow, separated by interdune depressions (slacks); a number of slacks accommodate fresh-water lagoons in close dependence of groundwater. The outermost ridge displays large-scale blowouts, bounded by parabolic and steep (>30º) precipitation ridges, which drown vegetation and slacks (Fig. 53). The ridge crest is in general vegetated grades with the beach by means of steep (occasionally scarped) slopes that may reduce inclination whenever an incipient hummocky foredune accumulated and climbed its toe.

One 46m-long core retrieved from Poço do Pinheirinho slack contains evidence of a buried interdune deposit 3.25m-thick, similar to present day analogues, imbeded in clean coarse minerogenic sand. This deposit, dated > 30 000 BP consists of acidic, fermented organic debris including charcoal (OM content up to 37%), framed between layers of organic mud. The study of the organic infill of neighbour slacks suggest the older aeolian complex expanded westwards by emplacement of successive sand ridges, among which the youngest structure dated so far yielded a minimum age circa 4600 BP.

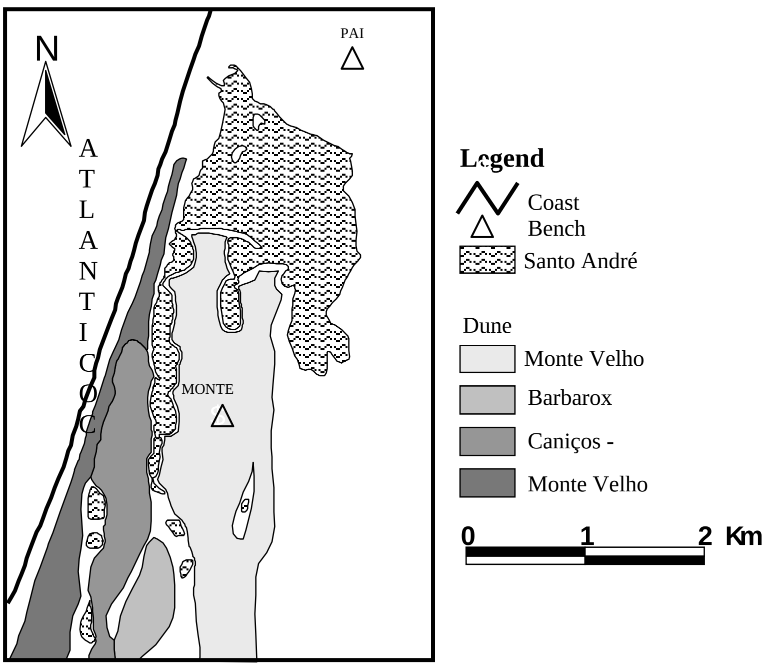

Figure 52. Present-day beach-foredune system cutting older ridge and slack and relation with Santo André lagoon.

3. The inlet

The barrier may occasionally breach during storms and is artificially and regularly opened since at least the 17th century, to promote water exchange, prevent eutrophication and drain the alluvial plains, at present reclaimed for grazing (formerly for rice). This procedure is known in similar environments of the Portuguese coast since the 15th century and might have started earlier.

Figure 53. Image of active blowouts drowning slacks.

This is usually carried out with annual periodicity during Easter in spring tide condition. While active, the tidal inlet evolves naturally throughout a sequential number of stages (Fig. 54) and usually silts up completely in less than one month, without significant drifting. Stage 1 immediately follows the breaching of the barrier and is characterized by very strong ebb currents, which cut a “V” shaped, deep and symmetrical channel through which sediment eroded from the barrier and lagoon is flushed to the near shore and reshaped as an ephemeral ebb delta. Stage 2 begins when significant onshore sediment movement, driven by wave or tidal currents, overwhelms the transport ability of the ebb jet-flow.

In this stage, the ebb delta rapidly migrates onshore and its surface is reworked as a large-scale swash platform. The main inlet channel is shallower but still linear at the inlet gorge and bifurcates over the ebb delta. Sand may be pushed across the inlet to build a flood shoal. During stage 3 onshore creeping of the outer delta proceeds and its shape changes to an ellipsoid with long axis paralleling the shoreline; it almost plugs the inlet gorge, which develops a contorted and meandering shape due to the presence of the outer shoal and fattening of the flood delta. This induces a point bar extension of one barrier tip and escarpment of the opposite slope, the cross-section of the gorge becoming asymmetrical. Stage 4 consists of a gradual change of the outer shoal into a pure elongated large-scale swash bar, which eventually completely welds to the beach and seals the remnants of the inlet channels (stage 5). From this moment onwards, the inlet area is filled by a sand bar constrained between the scarps of the initial channel and that accretes as a beach berm by overwash and deposition of successive sand laminae. Monthly surveying indicates that vertical accretion of the young berm is non-linear, major depositional events associating to storms (September 1998 and January 1999) – Fig. 55 Both the inlet and neighbouring regions are flood-dominated and marine sand may further invade the lagoon confined in tidal channels; this way, sand is trapped in the seaward margin of the lagoon and retained as fragments of flood deltas which may coalesce and accrete to the barrier and washover, contributing to the rapid reduction of the accommodation volume.

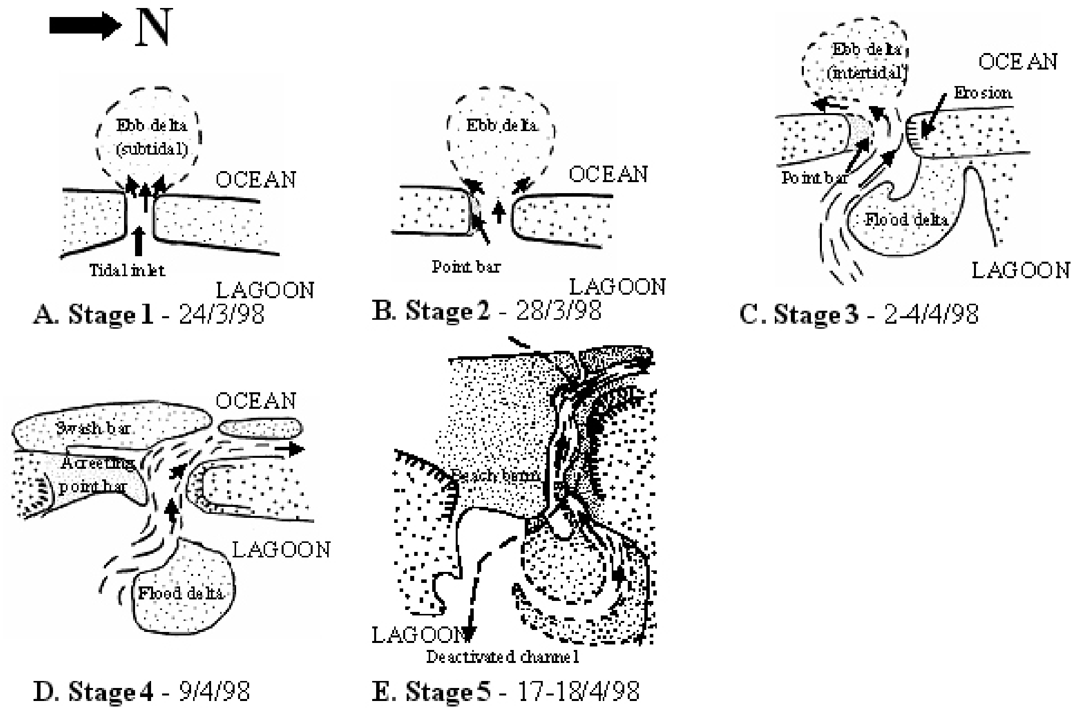

Figure 54. Schematic stage model of inlet morphodynamic changes.

4. The lagoon

When the lagoon is isolated from the ocean, water becomes brackish to fresh and near-bottom anoxic conditions may develop together with vertical stratification in temperature, salinity, pH, dissolved oxygen and turbidity (Fig. 56A, 56B); in this stage the lagoon collects fluvial inputs of water and sediment from two main drainage networks located East and South, through wide alluvial plains extensively choked with sediment. Most of the fluvial bedload is entrapped in alluvial fans while suspended sediment added by autochthonous organic matter eventually settles in the central main basin. Marine inputs restrict to occasional storm-induced overwash affecting mainly the seaward margin. In contrast, while the barrier is breached, marine water is allowed to interact with bottom sediment and partial to complete renewal of the lagoonal water may be accomplished in less than 48h in favourable conditions. Throughout the lagoonal basin the water column becomes homogeneous, salinity increases abruptly to normal marine concentration (35‰) and turbidity drops, while oxidic conditions replace previous near bottom anoxy (Fig. 56A).

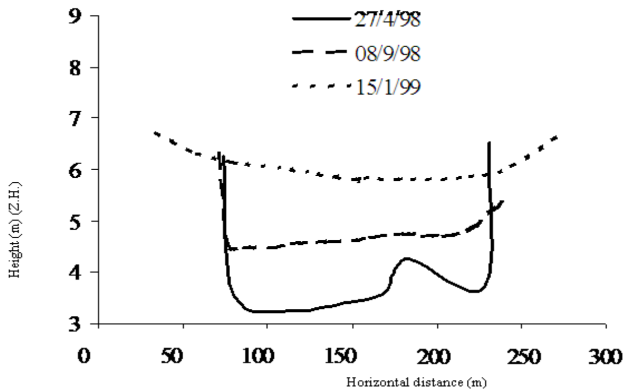

Figure 55. Major vertical accretion episodes recorded within the scar of the Santo André former inlet.

5. Holocene reconstruction

The reconstruction of the regional environmental changes since the Late Glacial in this area relies upon the multidisciplinary study of a number of boreholes, among which core LSA – Fig. 39 - 25m long) sampled the infill accumulated during the last 14ka in a sector of the Santo André lowland:

Pre-Holocene - terrestrial sedimentation in a fluviatile, semi-arid environment, contemporaneous of a low sea level (below circa –14m) and a distal shoreline. Sediment inputs resulted from intensive weathering and high erosion rate in the adjacent watershed and accumulated in a lowland incised in Miocene basement that extended westwards trough a wide coastal plain.

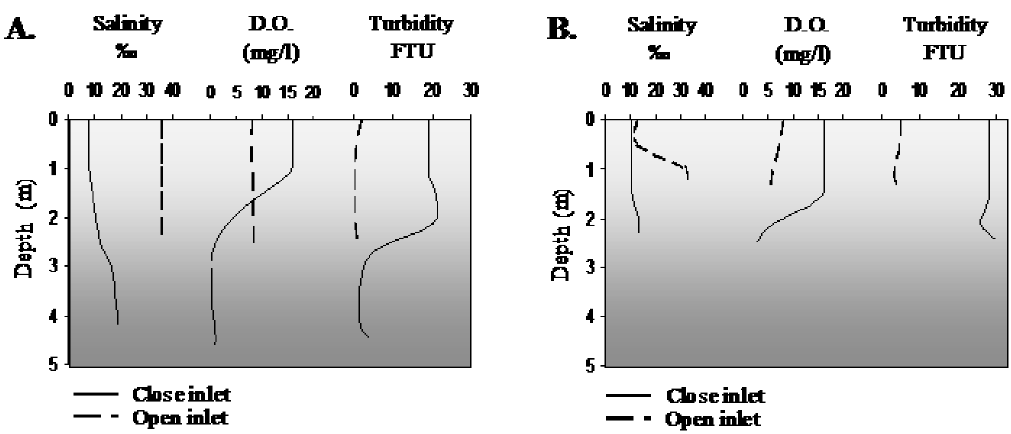

Figure 56. Vertical profiles of physic-chemical parameters of Santo André lagoon in open and closed inlet periods. A - 1998. B – 1999.

Early Holocene (circa 10020 to 5380 BP) – the first signals of marine flooding of this lowland associated with the Holocene transgression were dated about 10020 BP and since this time onwards the eustatic forcing factor increased in importance regarding environmental changes: the sea steadily invaded the area and defined a progressively more open-marine and shallow, wide gulf. This trend was disturbed by short episodes of terrestrial signature and came to an end circa 5380 BP, when an efficient detrital barrier formed; this defined a major threshold expressed by the isolation of the existing bay and change of this area into a coastal lagoon. The development of this barrier and lagoonal environment, which persists until present, is most probably an effect of the marked deceleration of the sea-level rise rate that occurred at that time.

Late Holocene (5380 BP-Present) – the lagoonal environment evolved essentially as a function of local forcing factors among which, the frequency and efficiency of exchanges with the ocean (promoted by overwash and mainly by the tidal inlet) dominated. Until circa 3570 BP the environment was severely restricted with fresh water and terrestrial sediment input added by autochthonous organic matter that altogether promoted rapid silting of the basin. A second sedimentation episode lasted until circa 1620 BP with brackish conditions and increasing marine influence due to a decrease of the efficiency of the barrier. The final stages of filling (since 1620 BP) culminated with the progradation of alluvial sediment over the lagoonal margin. Anthropogenic activity is clearly recorded in this area and the effects of deforestation, grazing and agriculture in the lagoon and watershed translated into contrasts in sediment texture and rate of accumulation.

THE SOUTHWEST OF PORTUGAL'S MAINLAND: A HERITAGE LITTORAL[13]

Ana Ramos Pereira

1. Introduction

The Southwest of Portugal, between Porto Covo (on the western coast) and Burgau beach (on the South coast), a 748 km^2 stretch is a heritage coast since 1995 (Fig. 57).

The major landforms are: the littoral platform bordered by small hills corresponding to a hemi-horst in the northern middle half area and by several grabens developed N-S, in the southern area. The littoral platform has a complex genesis, both fluvial and marine, has been faulted, uplifted and downlifted since the Pliocene (Fig. 57). Nine types of deposits were recognized: marine deposits, beach deposits, aeolian deposits (some of them carbonate) and alluvial fan deposits, from the Miocene to the Holocene (Pereira, 1990). Its evolution was highlighted in the area near Vila Nova de Milfontes, where 6 correlative deposits are better preserved (Pereira, 1990).

The coast is a cliffed one and the beaches are always narrow. They are more often related to the small estuaries but some small pocket beaches are recognized.

2. The littoral platform near Vila Nova de Milfontes

2.1. The preserved deposits

In this section, the platform is a well preserved feature, no major river exists, exception made to the Mira River. However, this monotonous landform hides a levelled and then faulted Paleozoic bedrock (Cambric schist and greywakes), where different types of deposits are still preserve (Fig. 58).

[13] **12th Stop** – Vila Nova de Milfontes

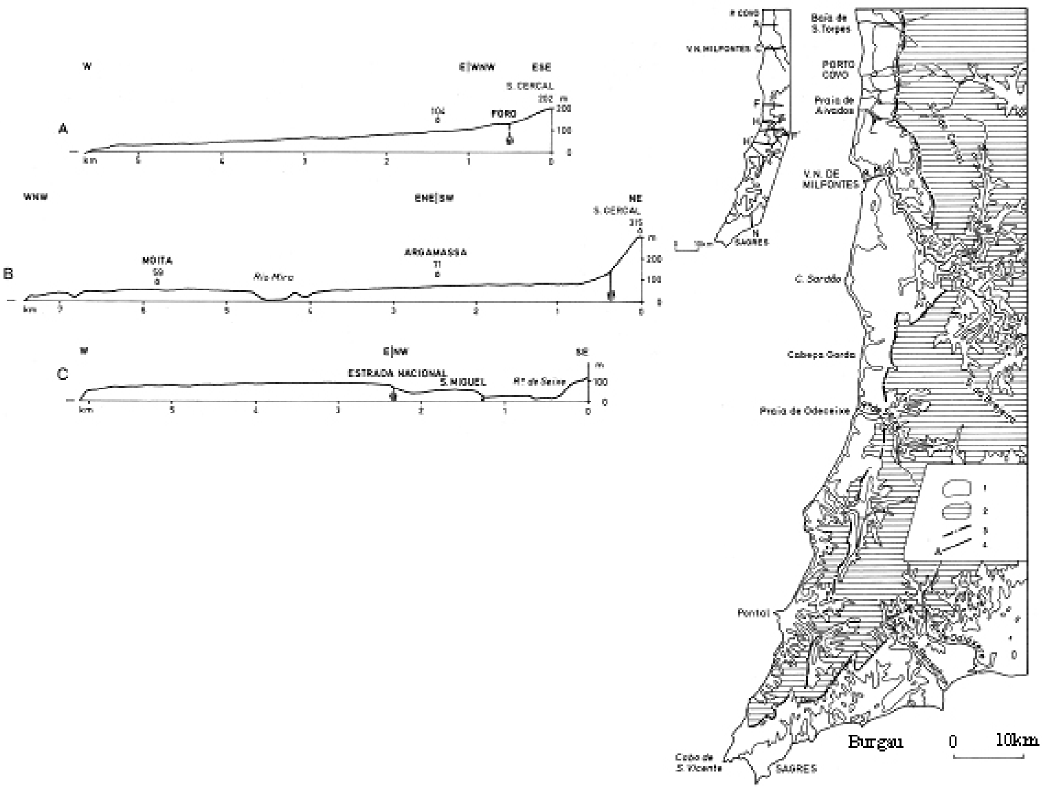

Figure 57. The Southwest of Portugal's mainland. 1 – height < 100 m; 2 – *id*. > 100 m; 3 – inland boundary of the littoral platform; 4 – transversal profiles of the littoral platform (From Pereira, 1990).

The detailed study of the morphology, the outcrops and the sedimentological analysis of the deposits and its lateral variations show:

(i) Over the faulted bedrock, the Red Formation of Foro (FVF) or simply Red Formation, a sandstone formation, with a pebble layer at the bottom. The pebbles, sometimes boulders, and the sands are rounded and bright. This beach deposit changes to a fluvial one and then to a red aeolinite inland, at the bottom of Serra do Cercal (Fig. 58, log D6, D5, G2 and A1). The fluvial facies appears again at the top of the Serra, with iron layers. The landscape was then a large alluvial plain near the sea, where the FVF was deposited. There was a sandy coastline with dunes and non-entrenched rivers drained the plain. The Serra do Cercal did not exist (Fig. 59-1).

(ii) An enormous change in the landscape was produced afterwards with the uplift of the Serra do Cercal. This tectonic episode is probably correlative of a climatic change because the plain was invaded by alluvial fans (LAI). Debris and mud flows were only episodic and locally reached the present day coastline; they are well preserved at the bottom slope of the Serra and also inside the small valleys, near the scarp (Fig. 58, log D3 and D5). The deposits have a sandy-mud matrix and pebbles not only from the bedrock but also of the iron bands of the FVF (Fig. 59-2). During this episode, the so-called littoral platform was created.

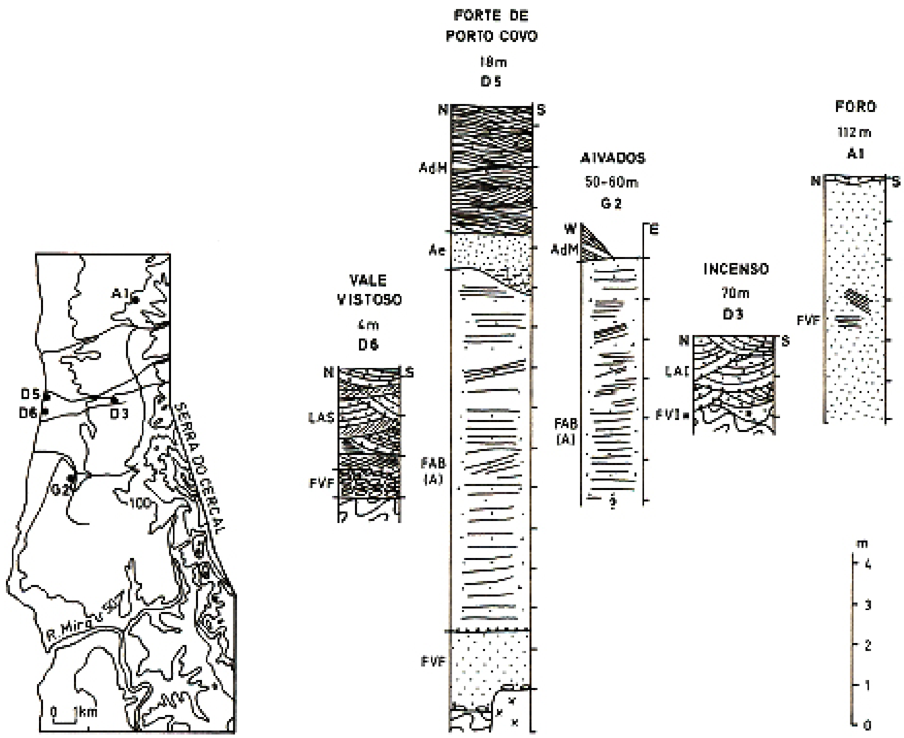

Figure 58. Logs of the deposits preserved on the littoral platform north of Vila Nova de Milfontes (Pereira, 1990).

(iii) The next episode registered in the platform was a marine invasion, leaving well-calibrated sands, with marine shells and Fe-Mg sand layers. This deposit, the Aivados-Bugalheira Formation (FAB, Fig. 58, log G2 and D5) penetrates 11km inland from the present day coastline in a subsided area, where the sea creates a small bay (Fig. 60-3). In this area, this Formation decreases in altitude from 50m to present day sea level in a stretch less than 10km wide. These indications as well as the visible liquefaction marks suggest that they have been submitted to tectonic strain.

(iv) A sea retreat allowed the establishment of small rivers, which reworked the sands of FAB, and the aeolian mobilization. A big sand field - Malhão dune field (M in Fig. 60-4), cutting into cliff today and penetrating 3km inland, was built up by N to WSW winds (it occupies today 20km^2). This dune field is still preserved because it was stabilized by vegetation and then submitted to carbonation (Fig. 58, AdM in log G2 and D5). In the outcrops the $CaCO_3$ can reach 80% and is the result of the shells dissolution (there is no $CaCO_3$ in the bedrock). The aeolianite of this dune field is also faulted near the sea and a scarp is still visible. The western tectonic compartment subsides and was invaded by the sea leaving a characteristic morphology of platforme a vasques and sand and small rounded pebbles (Monte Figueira Formation).

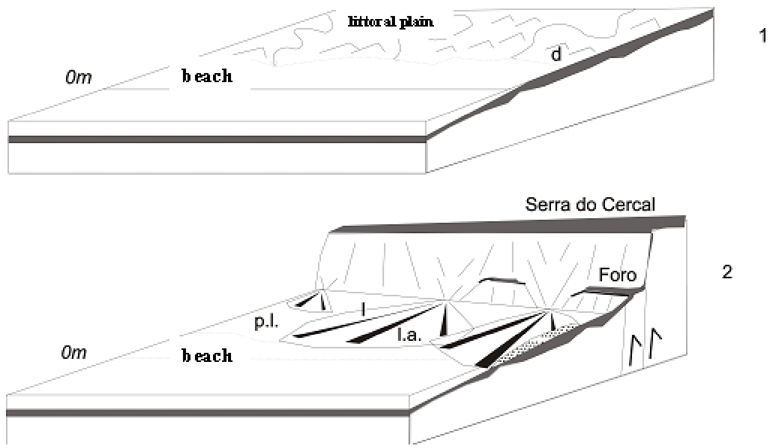

Figure 59. The evolution of the littoral platform dominated by Serra do Cercal during the Pliocene (1) and the Plio-Quaternary transition (2) (Pereira, 1999). 1 - the littoral plain bathed by the sea, with beach and dunes (d); 2 – littoral gravel plain (p.l.), with alluvial fans (l.a.), bathed by the sea and dominated by Serra do Cercal (legend in Fig. 61).

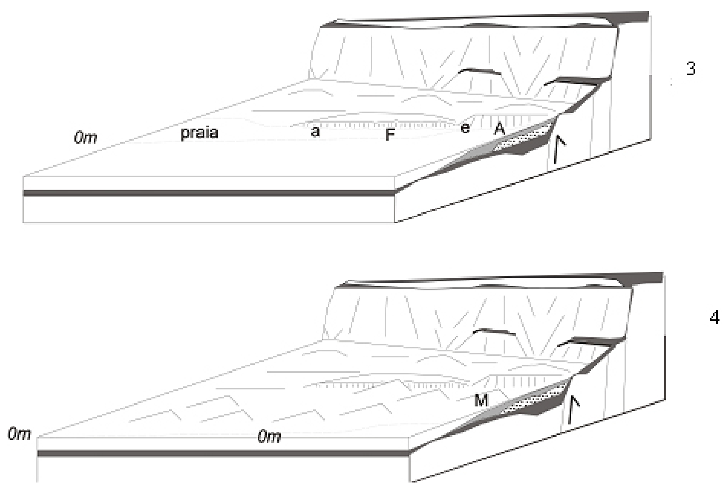

Figure 60. The evolution of the littoral platform dominated by Serra do Cercal during the Lower and Middle Plistocene (?). 3 - e – scarp fault, a – cliff cut into the alluvial fans, A – Aivados, F – Fort). 4 - littoral platform covered by a large dune field (M – Malhão), after a sea retreat. (Pereira, 2000). (Legend in Fig. 61).

(v) Another sea retreat was registered and a new dune field – Aivados dune field, was built and then submitted to carbonation. Only the eastern leeward of the dune field is still preserved. It is cut into the cliff, but it is still recognized in the internal continental shelf, where it creates islands, like the Pessegueiro Island (P in Fig. 61). The dunes were built by NW to SW winds.

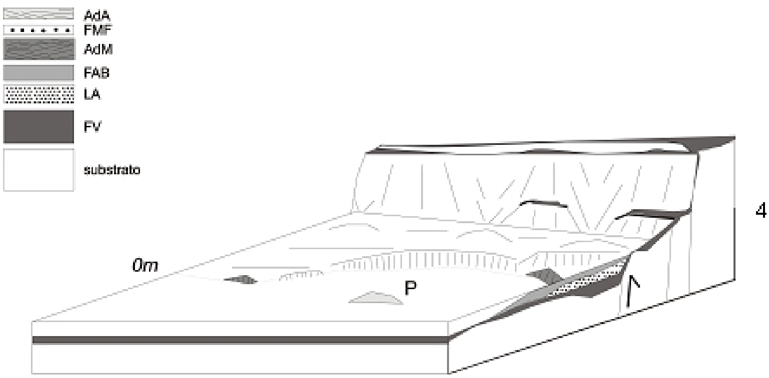

Figure 61. Present day landscape (P – Pessegueiro Island). Deposits from the oldest to the youngest: FV – Red Formation (FVF); LA – alluvial fan; FAB – Aivados-Bugalheira Formation; AdM – Malhão aeolianite; AdA– Aivados aeolinite (Pereira, 1999).

(vi) The latter evolution of the platform is related to the establishment of the drainage network and the transgressive dune field (Fig. 61 and 62).

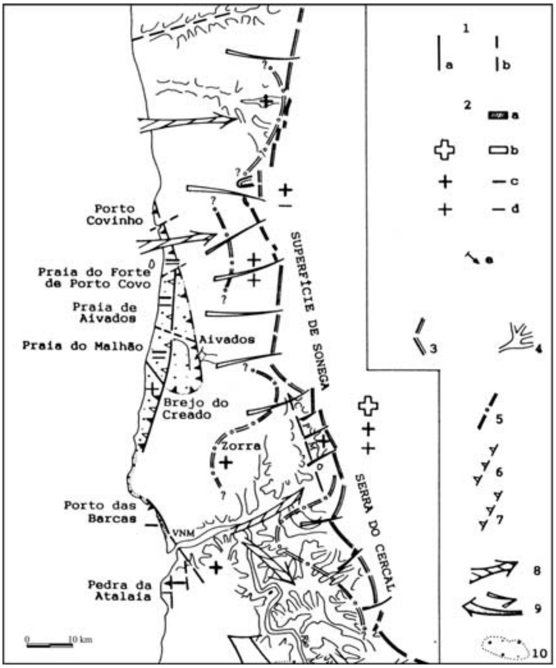

Figure 62. The polygenic evolution of the littoral platform near Vila Nova de Milfontes. 1- tectonic lineations: a – fault; b – probable fault; 2 – relative displacement of the tectonic compartments: a – previous or locally correlative of Red Fm; b – post Red Fm; c – post Aivados-Bugalheira Fm; d – post Malhão aeolianite and direction of tilting; 3 – scarp slope of Serra do Cercal; 4 – main valleys entrenching the platform; 5 – approximate position of the contemporaneous Red Fm coastline; 6 – correlative cliff of the Aivados-Bugalheira Fm; 7 - correlative cliff of the Monte Figueira Fm.; 8 – sea transgression during Red Fm deposition; 9 – alluvial fans advance; 10 – Aivados-Bugalheira Fm extension (Pereira, 1990).

3. Data information

In all the deposits mentioned before there are no fossils. After establishing the succession several correlations have been made not only with this section of the coast but also inland and to the south. Pereira (1990) gives a Pliocene age to The Red Formation (Zanclian or Placencian), in relation to the establishment of the exoreic drainage network. The climate was probably warmer and wetter than in the present day, with contrasting seasons toward conditions of increasing dryness. The aridity increased more during the tectonic episode and the related alluvial fan. This episode took place in the transition between the Pliocene and the Quaternary. It can belong to the Ibero-Manchega II phase.
Therefore, the southwest littoral platform is the result of the morphotectonic differentiation of an old Tertiary surface.

The table 1 synthesises the latter sequence and the uncertainties.

Table 1. Sequence of geomorphological episodes from the Lower Plistocene to Holocene. (based on Pereira, 1990 and Pereira a and Angelluci, 2004)

Geomorphological feature	Sedimentological unit	Environment	Probable Age
Aivados beach (old)	Aivados-Bugalheira Formation	Littoral - sandy coast	Lower Plistocene ?
	Fluvial rework of Aivados-Bugalheira Formation	Fluvial	Middle Plistocene?
Malhão dune field	(i) aeolian sand accumulation (ii) vegetation settling (iii) carbonation (iv) Malhão aeolianite	Aeolian	Middle Plistocene (OIS6?)
Monte Figueira beach/rocky platform	(i) Monte Figueira Formation	Littoral –beach /rocky platform	OIS 5
Malhão fault scarp	*Regional tectonic deformation; faulting of the Malhão dune field and tilting of Monte Figueira rocky platform*		OIS5 – 4 ?
	Aeolian sand	Aeolian	OIS 4 – 3 ?
Valleys	Turf /Psammit palaeosoil	Fluvial/ Lacustrine / Pedogenetic	42 519 ± 1263[14] OIS 3? BP
Aivados dune field	(i) aeolian sand accumulation (ii) vegetation settling (iii) carbonation (iv) Aivados aeolianite	Aeolian	OIS 3 – 2?
Transgressive dune field	Aeolian sands	Aeolian	Holocénico - OIS 1

[14] Calibrated data; 39 490 ± 2340 BP (Schroeder-Lanz, 1971).

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Road Log

Day 1 – 2 September
STOP 1 – S. Félix de Launde – "The littoral of Oporto region (Northern Portugal)"
STOP2 – V. Nova de Gaia-cabedelo – "Douro's estuary dynamics"
STOP 3 - Barrinha de Esmoriz – "The evolution of the northern end of the Aveiro lagoon system"

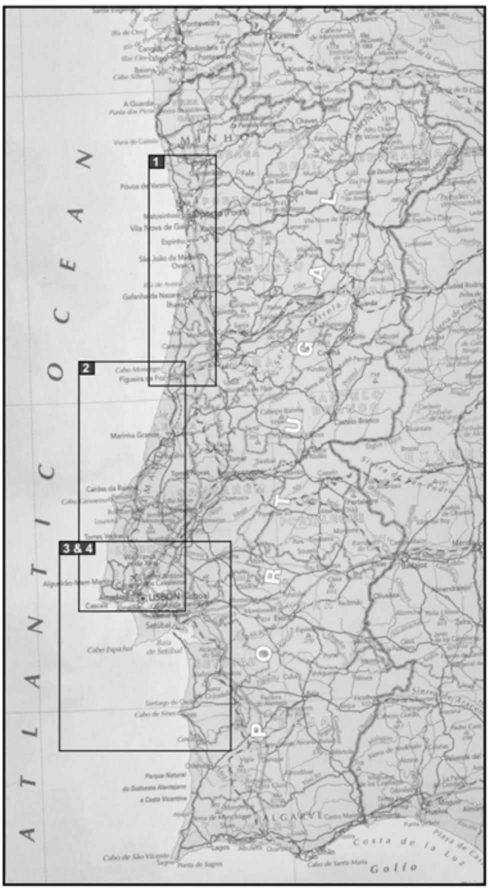
Day 2 - 3 September
STOPS 4, 5 and 6 – "Geomorphology and coastal dynamics of the Figueira da Foz region"
STOP 7 – "The Ponta do Facho landslide event and the present evolution of rock coast between Lagoa de Óbidos and Nazaré".

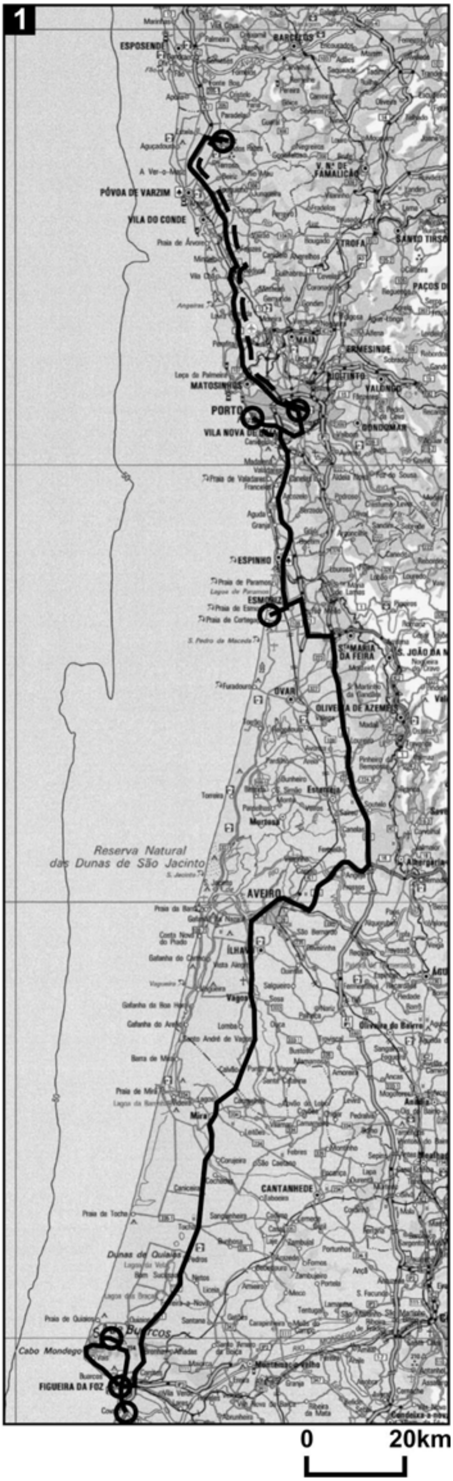
Day 3 – 4 September
STOP 8 – "Estimating the shore platform down wearing rate at Ribeira de Ilhas using a TMEM. First results"
STOP 9 "Cabo da Roca headland and its surrounding".
STOP 10 – "The western littoral of Setúbal Peninsula".

Day 4 - 5 September
STOP 11 – "Dynamics of Santo André coastal lagoon".
STOP 12 – "The Southwest of Portugal's mainland: A heritage littoral**".**

Sixth International Conference on Geomorphology Field Trip Road - log

A1 - Portugal: Coastal Dynamics



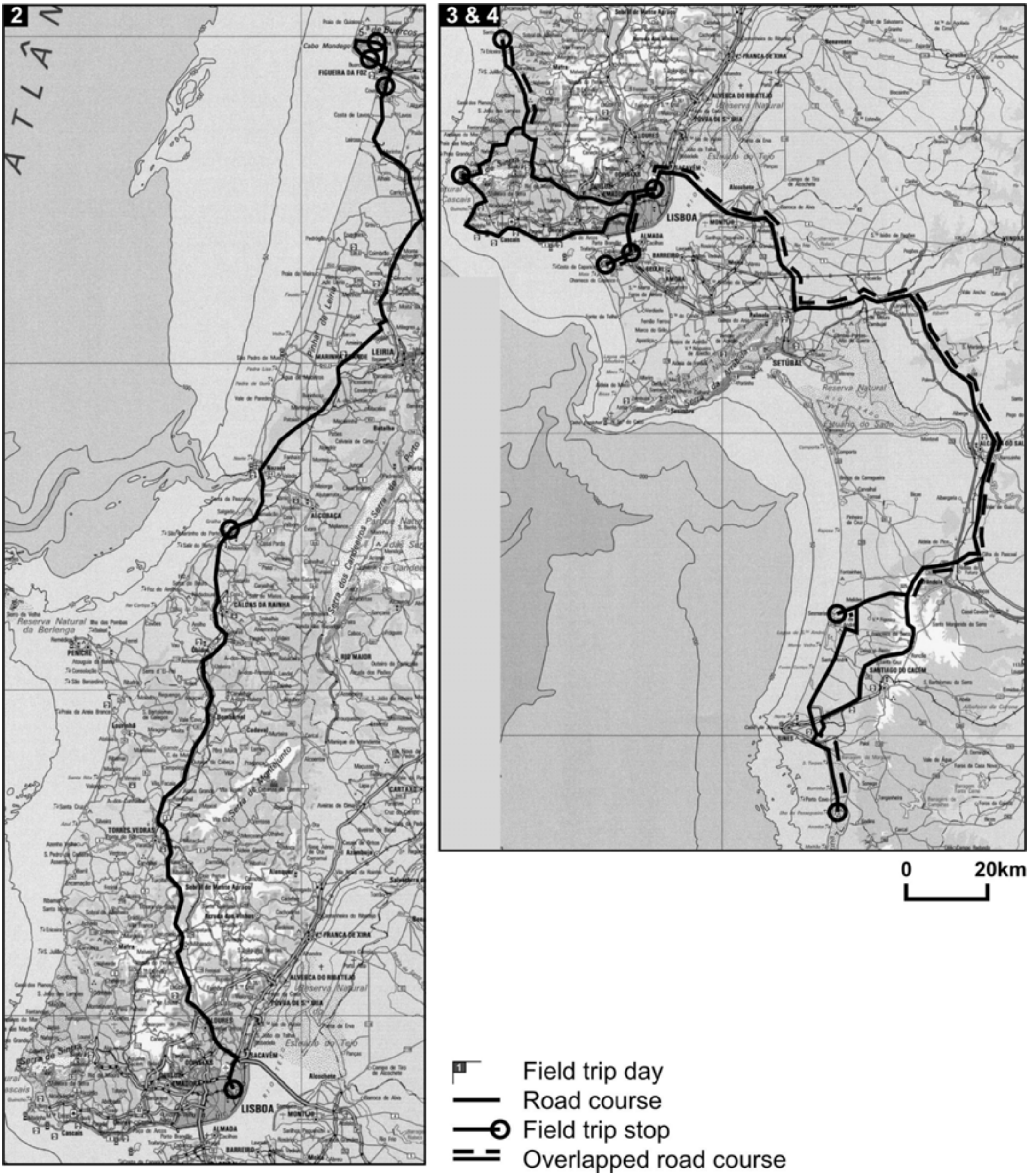


Field trip day

Road course

Field trip stop

Overlapped road course


2
3 & 4
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Field trip day
Road course
Field trip stop
Overlapped road course

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