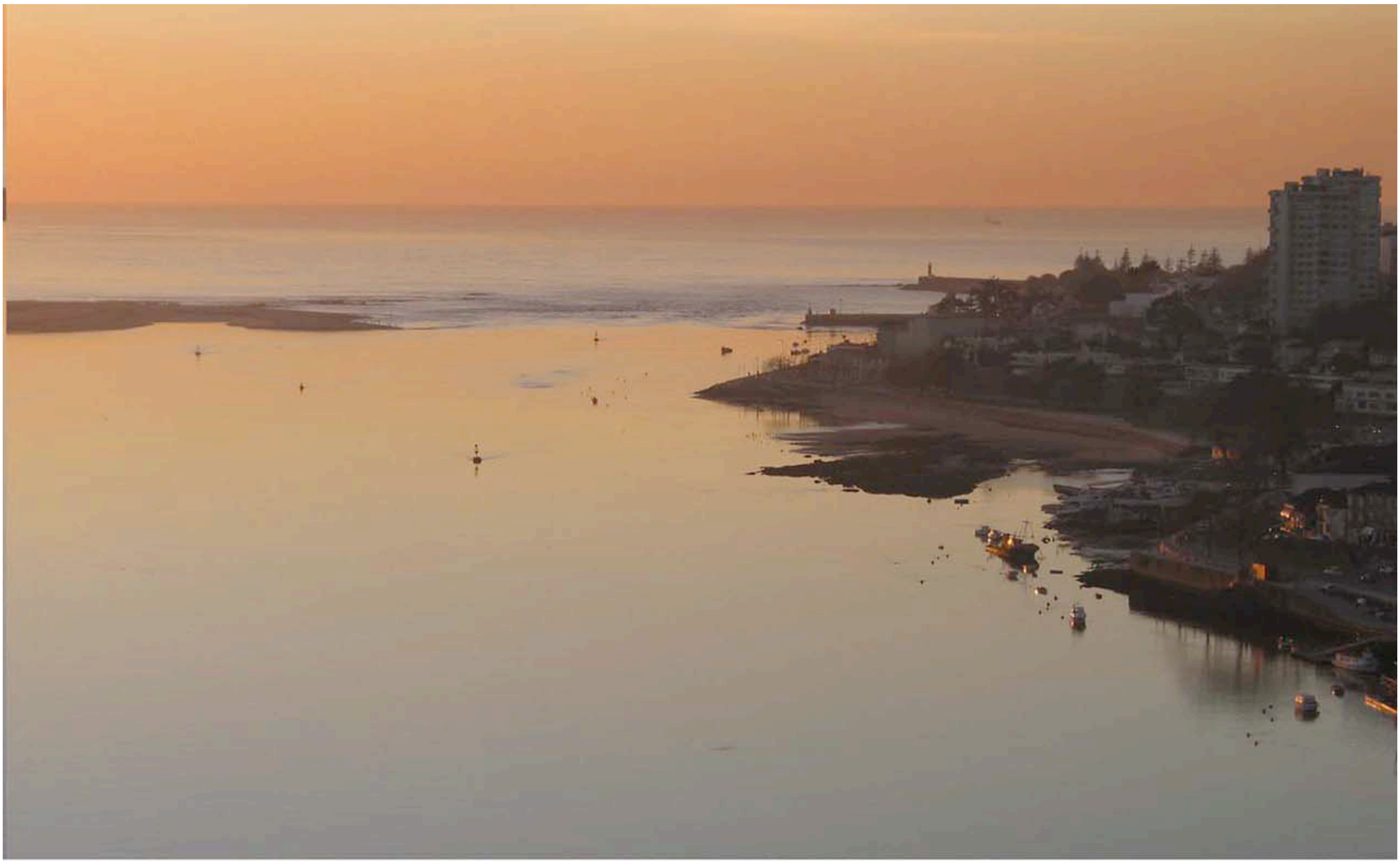

Porto, auditório da Reitoria da UP
19 Setembro 2005

Seminário

Sea level changes
Mudanças globais, variações do nível do mar e dinâmica costeira

PROGRAMA
10:00 Sessão de Abertura
10:30 **Nils-Axel Mörner** (Universidade de Estocolmo) - Sea level changes and coastal evolution; Past records & Future Perspectives.
11:30 Coffee break
12:00 **Ramon Blanco Chao** (U. Santiago Compostela) - Coastal change in the NW Spain since the last interglacial: influencing factors in long to short timescales.
12:30 **Helena Granja** (U. Minho) - Late Pleistocene-Holocene environmental changes (NW coastal zone of Portugal).
13:00 Almoço
15:00 **Assunção Araújo** (U. Porto) – Porto littoral: the influence of tectonics in sea level changes and coastal morphology.
15:30 **Pedro Proença Cunha** (U. Coimbra) - Tsunamis generated by island landslides - the example of La Palma.
16:00 **Ana Ramos Pereira** (U. Lisboa) - Sea level changes and neotectonics - some examples in Portugal (Arrábida and Southwest).
16:30 **Alveirinho Dias** (U. Algarve) – Holocene mean sea-level variations: data precision and compatibility.
17:00 **Mesa redonda**: Global Change – the IPCC and the scientists.

Apoio

FCT Fundação para a Ciência e a Tecnologia
MINISTÉRIO DA CIÊNCIA, TECNOLOGIA E ENSINO SUPERIOR

U. PORTO

Dep. Geografia FLUP
APEQ (Assoc. Port. Estudo Quaternário)
APGeom (Assoc. Port. Geomorfólogos)
APG (Assoc. Port. Geógrafos)
APG (Assoc. Port. Geólogos)

Índice

Sea Level Changes and Coastal Evolution Past Records and Future Perspectives

Nils-Axel Mörner

Paper presented at the Sea Level Seminarium in Porto, September 19, 2005

About the speaker: He has worked with sea level problems for 40 years. As head of the unit on Paleogeophysics & Geodynamics at Stockholm University Sweden, he worked with many different geological problems in different parts of the globe. He was president for the INQUA Commision on Neotectonics 1981-1989 and president of the Commission on Sea Level Changes and Coastal Evolution 1999-2003. In 2000, he launched an international research project in the Maldives.

Below follow some notes as to his contribution sea level research

1969 presented a new eustatic curve (by "solving" the inter-relationship between glacial isostasy and eustas).
1971 comparing sites all over the globe with respect to eustasy and local tectonics
(Paleo3, 9; 153-181).
1971 presenting the observation that sea level cannot have been parallel during the last 8000 years with respect to the present ocean surface (Geol. en Mijnbouw, 50: 699-702; GSA Bull., 82: 787-788).
1976 presenting the novel paradigm that the geoid must have changed with time (J. Geol., 84: 123-151) implying that there is no global eustatic curve and that every region must present its own "regional eustatic curve" such as his NW European regional eustatic curve (Paleo3, 19: 63-85; 29: 281-300).
1986 redefining the concept of eustasy (J. Coastal Res., SI(1): 49-51).
1988 presenting the theory of redistribution of ocean water due to surface current changes in response to variations in the Earth's rate of rotation (in: "Long Term Changes in Marine Fish Population", Vigo, 3-19; in: "Secular Solar and Geomagnetic Variations", Klüwer, 455-478).
1995 further development of the rotation theory (GeoJournal, 37: 419-430).
1996 summarising available data (Z.Geomorphology N.F., 102: 223-232).
1996 observing that Solar Minima corresponds to increased rates of Earth's rotation affecting sea level (An. Brazilian Acad. Si., 68,Supl.1: 77-94).
1999 presenting the special Baltic concept of changes in run-off gradient in sea level changes (Quaternary International, 60: 75-82).
2000 starting the Maldives Sea Level Project indicating that sea level is not in a rising mode today as generally claimed (Global Planet. Change, 40: 49-54 and 177-182) with a prediction for year 2100 of +5 cm $\pm$15 cm.
2003 reporting (INTAS project) at EGU-AGU-EGS meeting in Nice that we around 2040-2050 will be in a new Solar Minimum.
2004 summarising Mediterranean (and global) sea level problems (Paper 1 attached) and debating future sea level changes (Paper 2 attached).

Note also, his books on *Earth Rheology, Isostasy and Eustasy* (1980), *Climatic changes on a Yearly to Millennial basis* (1984) and *Paleoseismicity of Sweden – a novel paradigm* (2003).

Paper 1: from Z. Geomorph. N.F., Suppl-Vol. 137, 91-102, 2005.

Sea level changes and crustal movements with special aspects on the Eastern Mediterranean

Nils-Axel Mörner
Paleogeophysics & Geodynamics, Stockholm University, Sweden

with 8 figurs

Abstract
Eustasy, defined as "changes in oceanic level", is driven by five main factors controlling; water volume, basin volume, geoid topography, sea surface topography and rotational distribution of water. In the last 5000 years, the redistribution of oceanic water masses has been the dominant factor. No signs of any significant on-going rise in sea level are found, on the contrary, the flooding-scenario of IPCC is strongly contradicted. The changes in sea level in the Mediterranean are predominantly caused by local and regional tectonic factors and regional eustatic changes. The eustatic component in the Eastern Mediterranean is sensitively responding to variations in regional evaporation and precipitation; at present as well as in the past.

1. Sea level changes
The word "eustasy" has, by tradition, been used to denote changes in the oceanic level as opposed to changes in the crustal level. Originally, it was assumed that changes in the ocean level were identical all over the globe. Hence, eustasy was defined as "simultaneous changes in global sea level" (e.g. Fairbridge, 1961). Later, it was understood that changes in the ocean level are not simultaneous and similar over the globe, but differential and sometimes even opposed (Mörner, 1976a, 2000). Therefore, Mörner (1986) proposed a redefinition of the term "eustasy" to denote "changes in ocean level (regardless of causation)".

Fig. 1 illustrates the five main eustatic factors: (1) glacial eustasy, (2) tectono eustasy, (3) geoidal eustasy, (4) dynamic changes, and (5) rotation eustasy. Each parameter may be quantified as to rate, amount and time-application (Mörner, 1996a).

1.1. The loading models
It has been proposed that global sea level can be both reconstructed and predicted by means of a geophysical global loading model (e.g. Peltier, 1998; Lambeck, 1996). The basic assumption of these models is that the loading and deloading by the waxing and vanning ice caps of the Ice Ages generate a global isostatic adjustment of coasts and sea-floors all around globe (Fig. 2A). This is only possible, however, if the Earth has a linear viscosity profile in the mantle. The reality and efficiency of the model can be easily tested in (1) the near-field, and (2) in the far-field with respect to the ice caps. If actual observational data are used, the test fails in the near-field (at least in Fennoscandia) and in the far field (the Indian Ocean and the Pacific).

In Fennoscandia and surrounding areas, all available facts speak for the existence of a low-viscosity asthenospheric channel (Fig. 2B), where the loading and deloading was fully compensated by regional horizontal flow (e.g. Mörner, 1979; Fjeldskaar & Cathles, 1991). Besides, the input data used in the models do not concur with available field evidence. This lead Mörner (2003a) to conclude: "As long as the global modelling does not consider these facts, they are bound to be unrealistic".

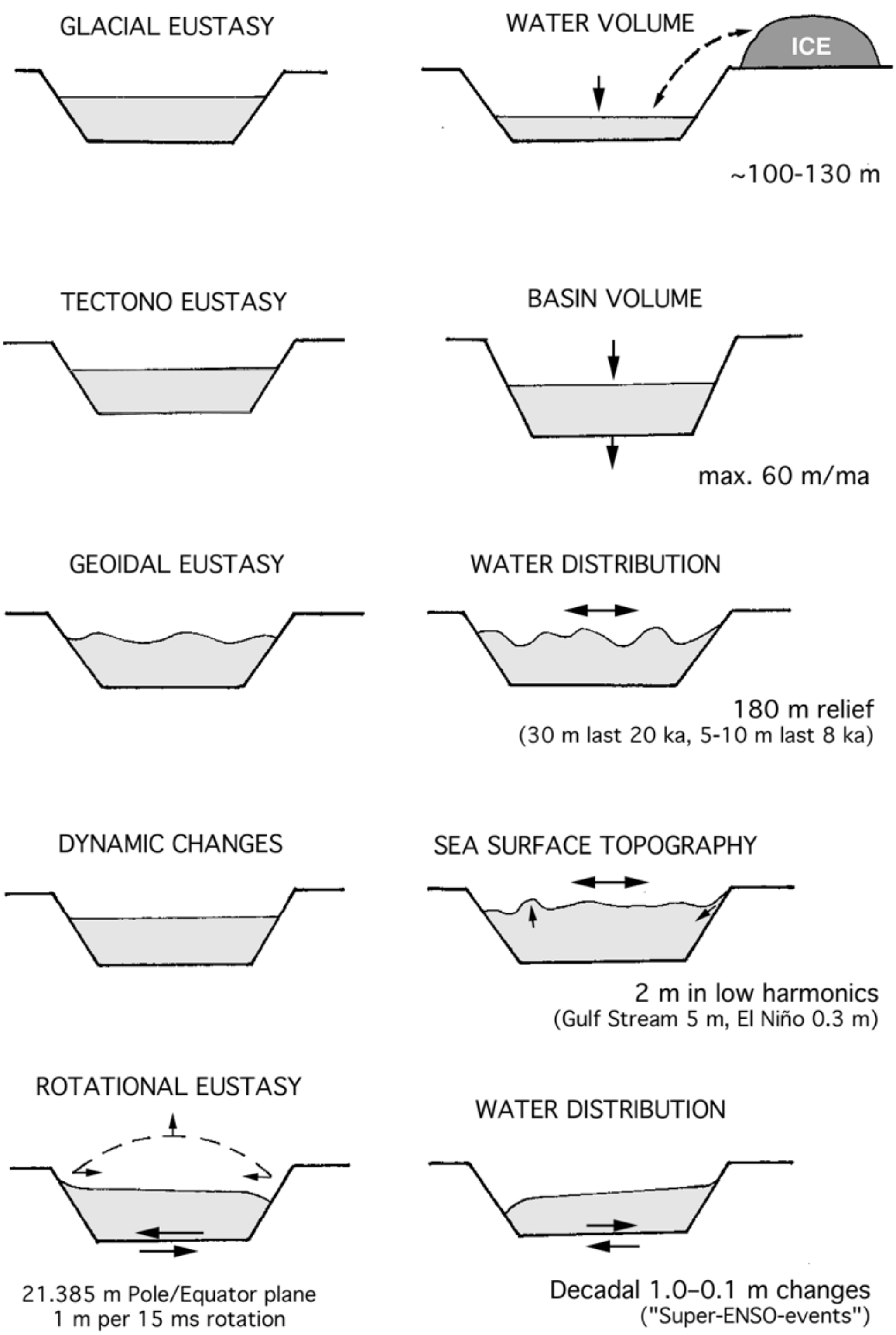

Fig. 1. Main eustatic variables with some quantification added. The glacial eustasy in the Quaternary has a sea level rang of about 100-130 m. Tectono eustasy is a very slow process, of negligible significance in the Holocene, with a maximum rate of 0.06 mm/year. The geoid has a maximum present topographic relief of 180 m. The changes in the geoid relief seem to have amounted to about $\pm$30 m at 20 ka and some 5-10 in the last 8000 years. The sea surface topography has, in the low harmonics, a relief of about 2 m. At major currents, like the Gulf Stream, it may amount to a few to 5 m. The El Niño signal is typically $\pm$0.3 m. Rotation causes a very large bulge difference between the polar and equatorial plane of 21,385 m. The relation between spin rate and sea level height is about 15 ms spin rate to 1 m sea level. Decadal changes in the Earth's rate of rotation have a potential to redistribute oceanic water masses rising and lowering regional sea level in to order of 1.0 to 0.1 m (known as "Super-ENSO events").

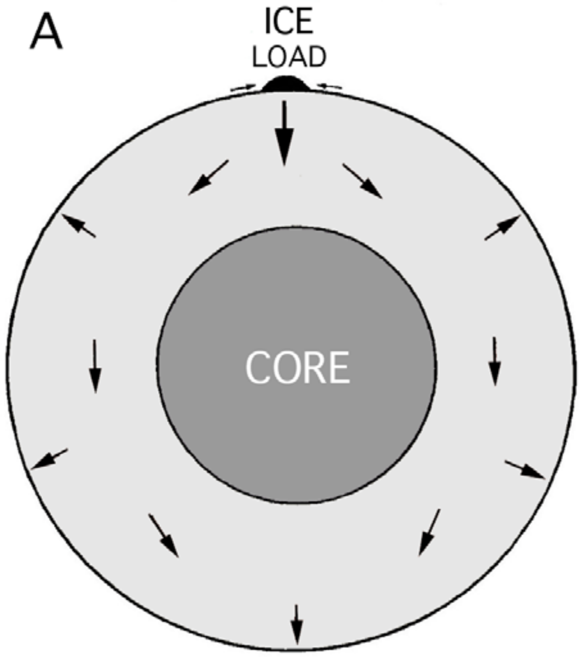

Fig. 2. Global loading versus regional loading. A: In the global loading models, the glacial loading/deloading will be transferred through the mantle and affect the coasts and sea floors all around the globe. This precludes a linear viscosity in the mantle. B: Alternatively the glacial loading/deloading is fully compensated regionally via a low-viscosity asthenosphere. Observational data from Fennoscandia and surrounding areas are consistent only with a regional loading model where compensation takes place via a low-viscosity channel flow (Mörner, 1979; Fjeldskaar & Cathles, 1991).

The loading models predict high Mid-Holocene sea levels in the Pacific and Indian Ocean. This does not concur with observational facts, either in the Indian Ocean or in the Pacific. The new sea level curve of the Maldives (Fig. 3), exhibits a long term base-curve not above present sea level and a number of rapid oscillations caused by dynamic forces (factors 4 and 5 in Fig. 1). In the Pacific, observed short and rapid fluctuations in sea level (Pirazzoli et al, 1988; Nunn, 1995, 1999) do not concur with the loading model but represent high-frequency dynamic sea surface changes. Grossman et al. (1998) reconstructed the spatial distribution of Mid to Late Holocene sea level changes in the Pacific. Their reconstruction does not concur with the prediction from the loading models, but with geoid deformation and/or changes in sea surface topography.

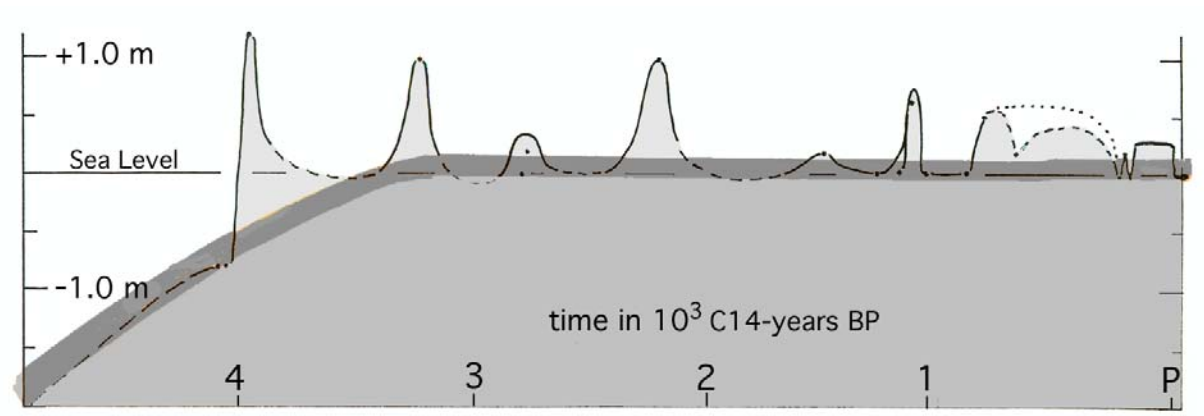

Fig. 3. The new sea level curve of the Maldives consists of short and rapid sea level peaks caused by dynamic variables of Super-ENSO-type, superimposed on a mean base-curve (dark grey line) that differs significantly from all predictions by the global loading models. This implies that the far-filed test fails for the Indian Ocean. The same applies for most of the Pacific records.

Therefore, one should be very careful in the application of model reconstruction and prediction. This is, of course, especially true in an area like the Mediterranean dominated by tectonics and orogenic processes.

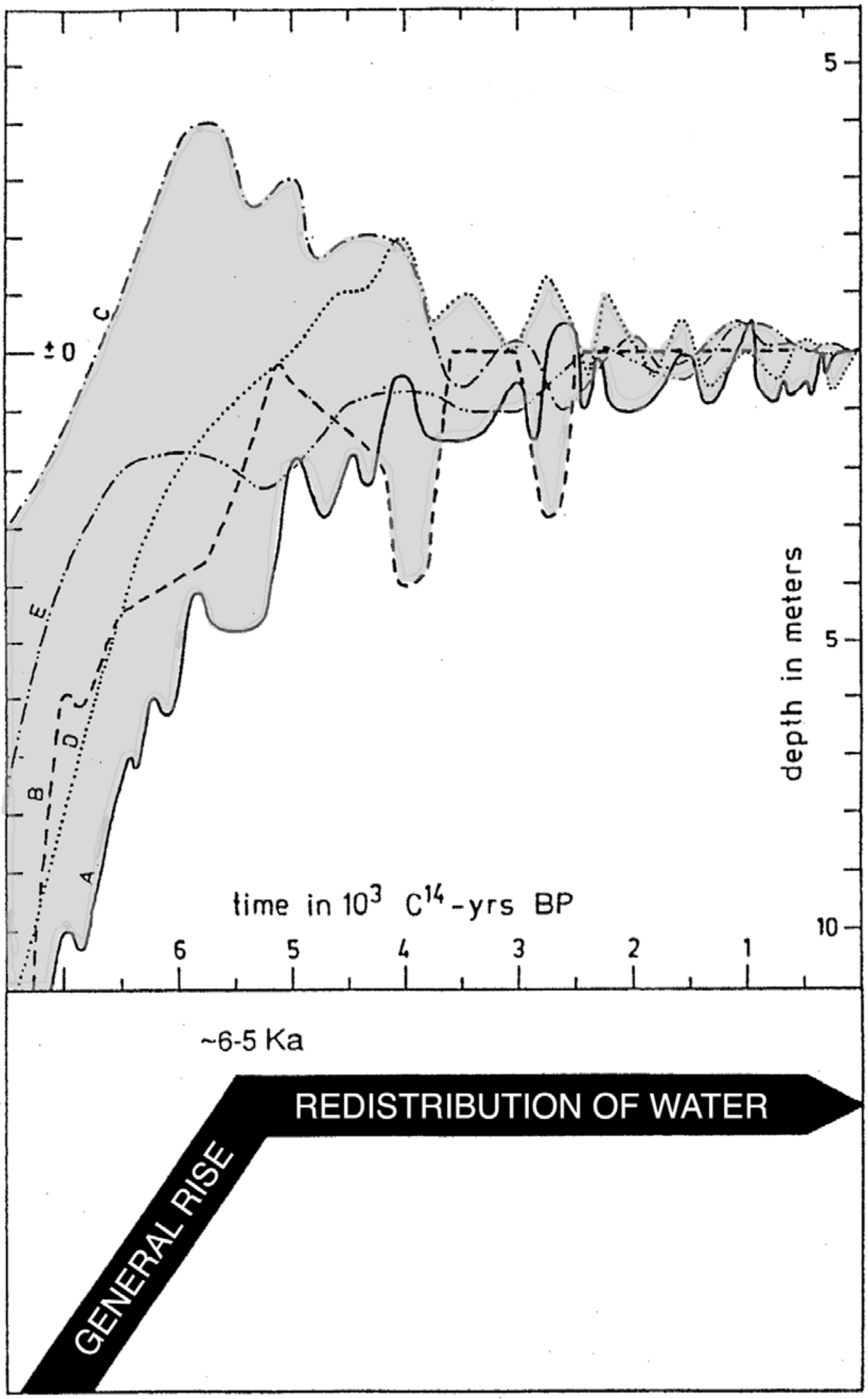

Fig. 4. Five sea level curves from different parts of the globe (Mörner, 1995, 1996b). Up to 5000–6000 C14-years BP, they are all dominated by a general rise of glacial eustatic origin. Thereafter, however, they are all dominated by the redistribution of oceanic water masses.

1.2. The last 5000 years

Up to 5000-6000 C14-years BP, the sea level changes were dominated by a general rise of glacial eustatic origin. After some 5000 C14-years BP (when the glacial eustatic factor had ceased), the situation changed to a dominance of redistribution of the water masses over the globe. This is illustrated in Fig. 4 (Mörner, 1996b). The main driving forces for this redistribution of water over the globe seem to be the interchange of angular momentum between the solid Earth and the hydrosphere (Mörner, 1995) driving and controlling ocean surface circulation in some sort of Super-ENSO/Super-non-ENSO variability (rotational eustasy in Fig. 1).

1.3. The present

I have recently discussed the present general sea level trend with respect to past records and future expectations (Mörner, 2003b, 2004). It will not be repeated here again. For the last decade, satellite altimetry has become an important new tool. In my previous papers (op. cit.), I showed the observational raw-data for the period 1992–2000. The extension up to 2003 can be found on internet (NASA/CNES), but with a significant change; now the original data set has been "corrected" so that it has assumed a tilting trend. When the new curve is calibrated back to its original ("uncorrected") observational values, a curve is given (Fig. 5) that lack sings of any rising trend. A number of ENSO-events are recorded super-imposed on a mean trend around zero. This implies a total lack of signs of an on-going sea level rise, and certainly not any recent sea level acceleration, as claimed in the IPCC scenario (IPCC, 2001).

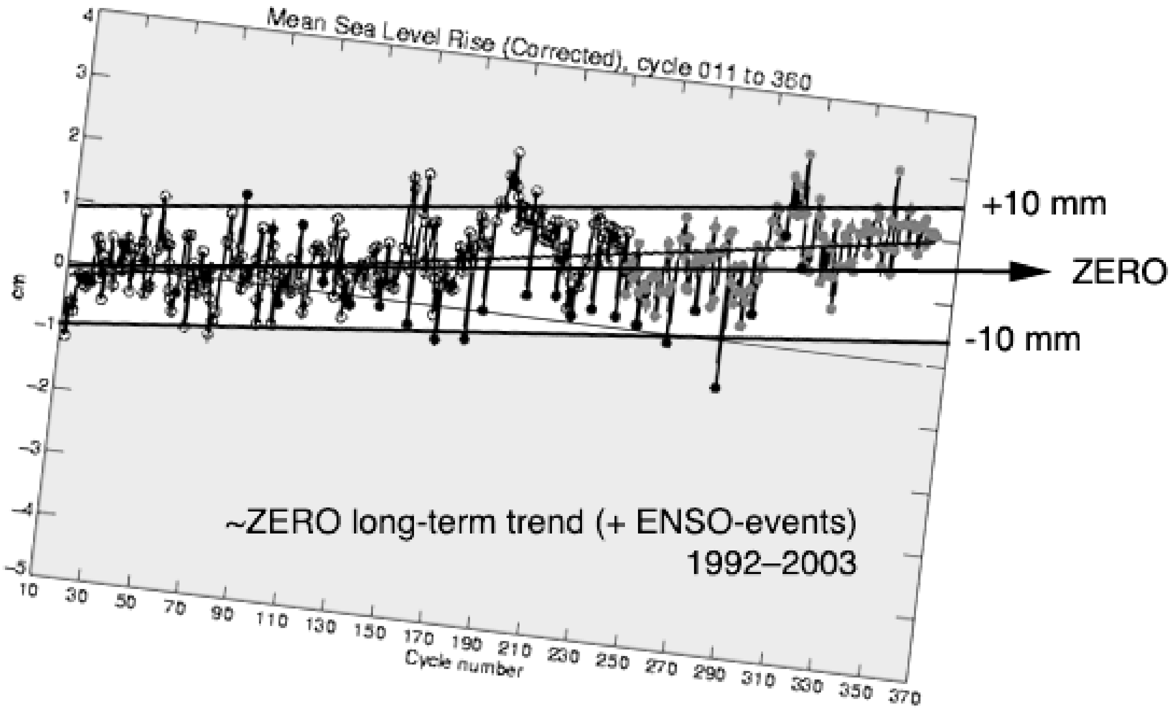

Fig. 5. The original satellite altimetry data up to early 2000 were presented on the web-site of NASA/CNES as given in Fig. 2 of Mörner (2004). In early 2003, the extended web-site data had assumed new tilt, however (and the data set was labelled "corrected"). In this figure, the 2003 data-set has been tilted back to the original, early 2000, level (i.e. the position of the raw data). Now the sea level rise is gone. The curve show variations around a zero-line (possibly there could be slight rise of maximum 0.5 mm/year as shown by a thin line). A number of ENSO-events are seen as annual peaks.

2. The Mediterranean

Geodynamically, the Mediterranean is a high-active region. It is crossed by a highly active seismic zone. It is bounded to the north by the Alpine orogenic belt. It is crossed by the active plate boundary between Africa and Europe. Major shear zones cross the area. The Nile delta induces subsidence. The Tyrrenian Sea is an area of rapid Pliocene-Quaternary subsidence.

Some of these structures lead their origin at great depths in the lower lithosphere and upper mantle. Therefore, it seems very unlikely that these long-term forces and processes would suddenly become over-printed and even reversed by loading forces transferred via the mantle from Fennoscandia (and North America) as claimed in the global loading models (e.g. Peltier, 1998; Lambeck, 1996).

No doubts, the changes in sea level within the Mediterranean region are predominantly affected by local to regional tectonism in combination with regional eustatic changes in sea level.

2.1. The eustatic component

A regional eustatic curve has been presented for Northwestern Europe (Mörner, 1976b, 1980). It has been tested for the North Sea region by Shennan (1987) and for the Baltic by Harff et al. 2001). For the Mediterranean, there does not exist any regional eustatic solution. Sea level curves from the north-western Mediterranean have been presented by Labeyrie et al. (1976) and Dubar (1987). For Eastern Mediterranean, the situation is still quite confused, and it seems fair to conclude that we are still lacking a reliable eustatic solution for the Eastern Mediterranean region.

2.2. The tectonic component

In the main parts of eastern Mediterranean, the tectonic factor seems to be the dominant factor behind recorded sea level changes in the last 5000-6000 C14-years BP (e.g. Pirazzoli, 1991). Besides seismo-tectonics, there are down-warping from sediment loading of the Nile delta (sedimento-isostasy), plate-boundary uplift, orogenic uplift, coastal warping, basin subsidence and related phenomena. Many authors have tried to relate their observations to predictions from the loading models of Lambeck (1996) and Peltier (1998). Understanding the basic problems with these models both in the near-field and in the far-field (above), and the exceptionally heterogeneous conditions of the lithosphere and upper mantle in this region, one must be very sceptical of any such model predictions for the Mediterranean region.

It seems, at least to me, much more realistic to work with the interaction of eustasy and tectonics. Below follow a few examples.

At Heraion on the Perachora Peninsula, there are a number of elevated notches explained in terms of co-seismic uplift (Priazzoli et al., 1994; Stiros & Pirazzoli, 1998). In Fig. 6, I have tried to test how this can be reconstructed by interaction of eustasy and seismotectonics. As eustatic component I have used the regional NW European eustatic curve of mine (Mörner, 1976b, 1980). The Fig. 6 test shows that all four notches can be perfectly well simulated, if the eustatic curve is cut by three major co-seismic uplift events; viz. ~4.5 m at ~5500 BP, ~2.0 m at ~3500 BP and ~1.0 m at ~1300 BP. The picture can be improved with a eustatic factor better fitted to the Mediterranean conditions. Still, it provides a more realistic picture than previous simple box-like diagrams (Stiros & Pirazzoli, 1998, Fig. 20).

From Mavra Litharia at the southern coast of the Gulf of Corinth, there are 11 dates of elevated marine species (Stiros & Pirazzoli, 1998, Table 1). The dates range all the way back to 10,000 BP. The Holocene Marine Limit (ML) is at +9.3 m. It seems primarily to have been cut by the sea at the transgression peak at 7000 C14-years BP. The "normal" eustatic depth of the 7000-level is at –10 m. One might advocate an uplift of 19 m in 7000 C14-years or 2.7 mm per year (~2.4 mm/yr in calibrated age). Pirazzoli talks about "avarage uplift rates between 1.7 and 2.5 mm/yr" (Stiros & Pirazzoli, 1998, p. 32) whilst Stewart & Vita-Finzi (1996) give a mean rate of 1.5 mm/yr. I think the available data can only be understood in terms of multiple co-seismic uplift events. Fig. 7 provides a test of the interaction between eustasy (same curve used as in Fig. 6) and seismotectonics. In order to explain available data in a meaningful way, the eustatic component has to be cut up by 9 co-seismic uplift events of ~12 m, ~10 m, ~7 m, ~9 m, ~1.5 m, ~3 m, ~2.5 m, ~2 m and ~1 m, respectively. According to this simulation, the sea level data are a function of the interaction between eustasy and multiple co-seismic uplift events.

Figs. 6 and 7 illustrate that it is much more fruitful to work with the interaction of eustasy and tectonics than referring to loading model predictions. It is the regional eustatic factor that is in urgent need of definition, and this can only be done by an extended observational data-base.

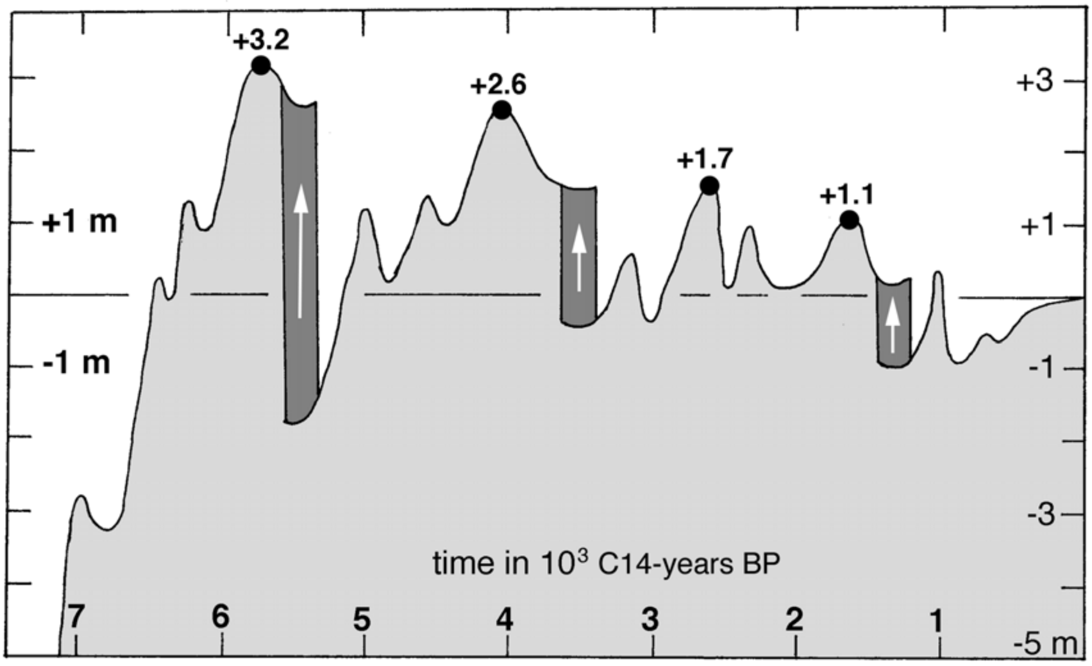

Fig. 6. The Heraion site with its four notches (black dots with elevation heights) simulated by the interchange of a eustatic curve cut by three co-seismic uplift events.

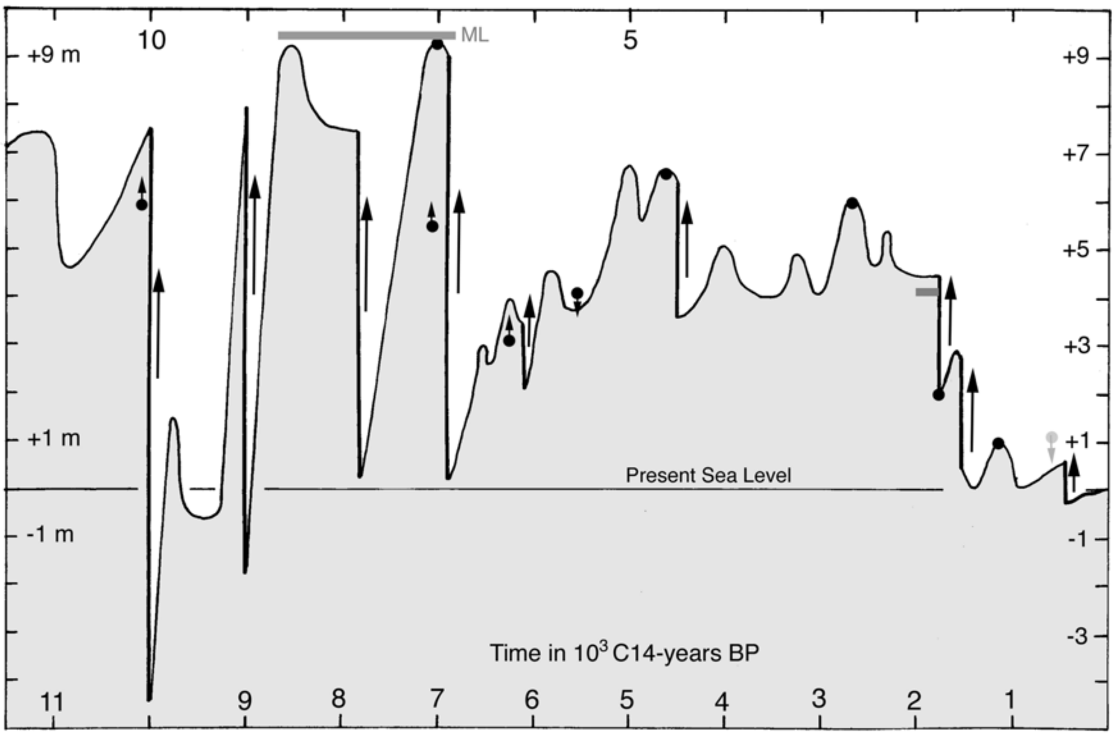

Fig. 7. Mavra Litharia site with its Marine Limit (ML) at +9.3 m and 11 dates of elevated marine species (black dots) simulated by the interchange of eustasy and nine co-seismic uplift events (black lines with arrows at the side).

2.3. The climatic components

Climatic factors is another major component behind recorded sea level changes in the last 5000 years and at present. Increased evaporation during warm periods and draughts will lower the sea surface, whilst periods of high precipitation will rise the sea level.

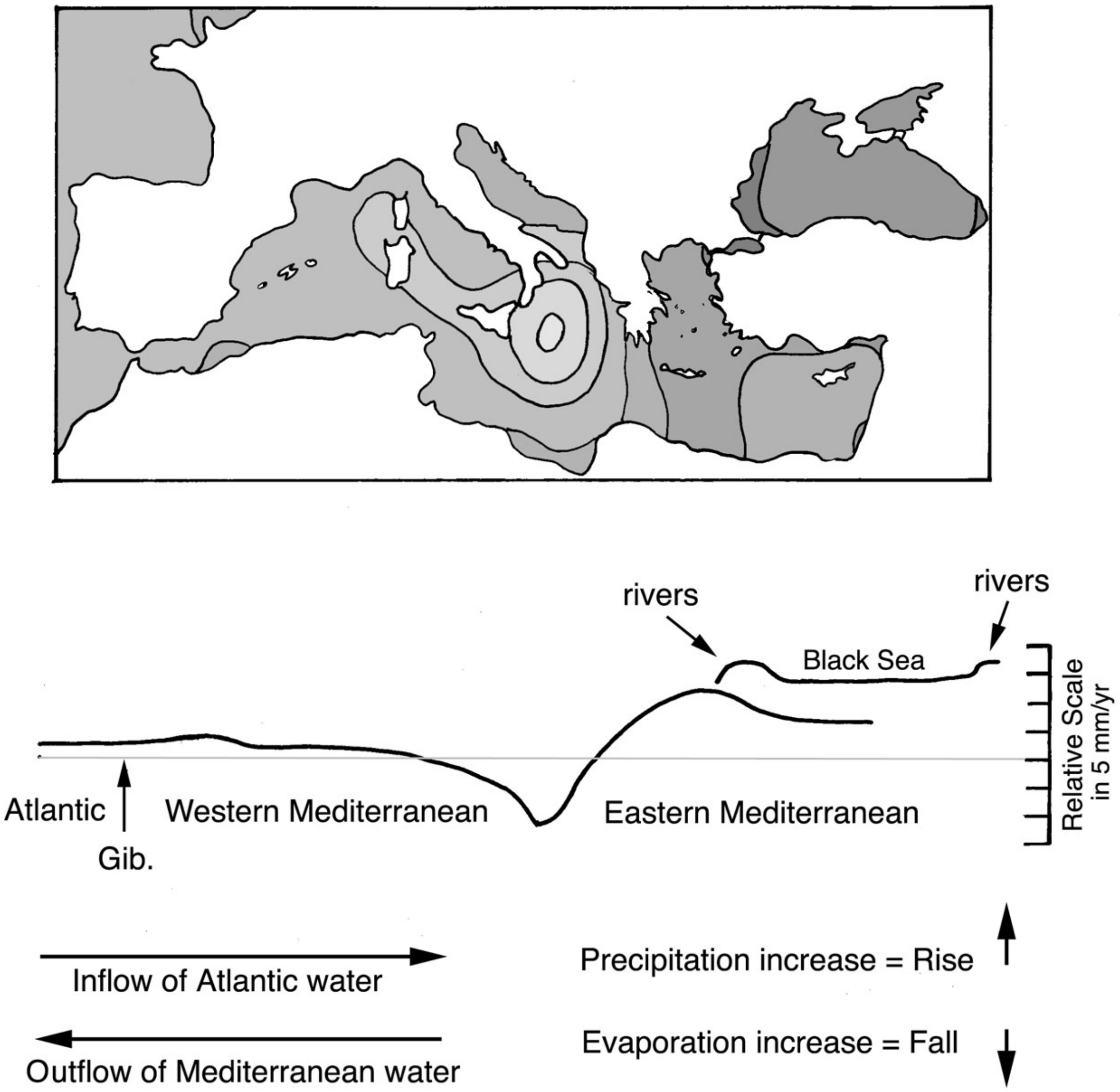

Fig. 8. Present sea level changes in the Mediterranean from satellite altimetry. A: spatial variations with high values in the Black Sea and down in a band over the Aegean Sea to the north African coast, and a marked low in the central Ionian Sea. B: East–West profile (on a relative scale) recording the high values in the east and the low point in the Ionian Sea. At the base, the main parameters controlling the changes in sea level. Whilst lateral interchange of water seems to de dominant for the Western Mediterranean, changes in precipitation and evaporation are the main controlling factors in the Eastern Mediterranean.

This can be observed in our present-day distribution of water in the Mediterranean and the Black Sea as recorded by satellite altimetry (Fig. 8). In the Black Sea and in Eastern Mediterranean, sea level is presently rising. In the central Ionian Sea, on the other hand, there is a point of significant sea level lowering. The present changes in sea level in the Western Mediterranean are driven by the interchanges of inflow of Atlantic water and outflow of Mediterranean water, and other regional dynamic factors. In the Eastern Mediterranean, the situation seems more variable. Obviously, the main controlling factor is the balance between evaporation and precipitation. In the satellite altimetry data of Fig. 8, the highest values are in the Black Sea and in the Aegean Sea. This seems to indicate that increased precipitation over

East Europe is now causing sea level to rise in a broad band over the Aegean Sea via Crete down to coast around Tobruk. An increase in evaporation over the Eastern Mediterranean and surrounding coasts would cause a lowering of the sea surface in the Eastern Mediterranean. If we are able to detect this in the micro-scale of the last decade, it should be identifiable in our past records, too.

Because there have been significant changes in precipitation and evaporation in the past, one would expect to see related changes in sea level in our sea level records. It seems urgent to focus future sea level research on this topic. It is interesting that the Eastern Mediterranean, via the Black Sea and the Bosporus-Dardanelles, is affected by changes in precipitation over the East European area surrounding the Black Sea, i.e. an area not directly bordering the Eastern Mediterranean. Changes in evaporation, on the other hand, are much more restricted to the Eastern Mediterranean area itself.

3. Summary and conclusions

The changes in oceanic sea level are driven by five main factors as illustrated in Fig. 1. Global loading models are seriously questioned as a tool for understanding local and regional changes in sea level. In the last 5000 years, global sea level has been dominated by the redistribution of oceanic water masses as a function of changes in rotation and dynamic climatic-oceanographic variables. The last centuries record a significant rise in the period 1850-1930. In the last decade, there are no traces of any significant rise, however, and certainly not of any alarming "acceleration" as claimed in the IPCC-scenario.

The sea level changes in the Mediterranean, a region of very complex crustal-dynamics, is generally dominated by tectonic factors. No reliable regional eustati solution has yet been presented. The application of loading model simulations (Lambeck, 1996) is considered untenable. The Heraion and Mavra Litharia sites at the Gulf of Corinth are re-visited with respect to a test of the simple interaction of eustasy and tectonics. The field data seem well expressed in terms of eustatic changes in sea level cut by 3 and 9 co-seismic events of uplift, respectively.

The Eastern Mediterranean changes in eustatic sea level seem sensitively affected by changes in evaporation and precipitation; at present as well as in the past. Whilst increased evaporation and general draught will lead to a lowering of regional eustatic sea level, increased precipitation will lead to a rise in sea level.

References

Dubar, M. (1987): Données novelles sur la transgression holocène dans la région de Nice (France). Bull. Soc. Géol. Fr., (8) 3, 195-198.

Fairbridge, R.W. (1961): Eustatic changes in sea level. Physics & Chemistry of the Earth, 4, 99-185.

Fjeldskaar, W. & Cathles, L. (1991): The present rate of uplift of Fennoscandia implies a low-voscosity asthenosphere. Terra Nova, 3, 393-400.

Grossman, E.E., Fletcher, C.H. & Richmond, B.M. (1998): The Holocene sea-level highstand in the equatorial Pacific: analysis of the insular paleosea-level database. Coral Reefs, 17, 309-327.

Harff, J., Frischbutter, A., Lampe, R. & Meyer, M. (2001): Sea level changes in the Baltic: Interaction of climate and geological processes. In: L.C. Gerhard, W.E. Harrison & B.M. Hanson (eds.): Geological perspectives of global climate change: 231-250.

IPCC (2001): Climate Change. Cambridge Univ. Press.

Labeyrie, J., Lalou, C., Monaco, A. & Thommeret, J. (1976): Chronologie des niveaux eustatiques sur la cöte du Roullillon de –33.000 ans BO à nos jours. C. R. Acad. Sci., D-282, 349-352.

Lambeck, K. (1996): Sea-level changes and shoreline evolution in Aegean Greece since Upper Paleolithic time. Antiquity, 70: 269, 588-611.

Mörner, N.-A. (1976a): Eustasy and geoid changes. J. Geology, 84, 123-151.

Mörner, N.-A. (1976b): Eustatic changes during the last 8,000 years in view of radiocarbon calibration and new information from the Kattegatt region and orther northwestern European coastal areas. Palaeogeogr. Palaeoclim. Palaeoecol., 19, 123-151.
Mörner, N.-A. (1979): The Fennoscandian upliftt and Late Cenozoic geodynamics: Geological evidence. GeoJournal, 3, 287-318.
Mörner, N.-A. (1980): The northwest European "sea level laboratory" and regional Holocene eustasy. Palaeogeogr. Palaeoclim. Palaeoecol., 29, 281-300.
Mörner, N.-A. (1986): The concept of eustasy. A redefinition. J. Coastal Res., SI-1, 49-51.
Mörner, N.-A. (1995): Earth rotation, ocean circulation and paleoclimate. GeoJournal, 37, 419-430.
Mörner, N.-A. (1996a): Rapid changes in coastal sea level. J. Coastal Res., 12, 797-800.
Mörner, N.-A. (1996b): Sea level variability. Z. Geomorph. N.F., Suppl-Bd. 102, 223-232.
Mörner, N.-A. (2000): Sea level changes and coastal dynamics in the Indian Ocean. Integrated Coastal Zone Management, Launch Issue, 17-20.
Mörner, N.-A. (2003a): Paleoseismicity of Sweden – a novel paradigm. ISBN-91-631-4072-1.
Mörner, N.-A. (2003b): Sea level changes in the Past, at Present and in the Near-Future. Global aspects. Observations versus Models. GI2S Coast, Research Publication, 4, 5-9. (IGCP-437).
Mörner, N.-A. (2004): Estimating future sea level changes. Global Planet. Change, Global Planetary Change, 40, 49-54.
Peltier, W.R. (1998): Postglacial variations in the level of the sea: implications for climate dynamics and soild-earth geophysics. Reviews of Geophysics, 36, 603-689.
Pirazzoli, P.A. (1991): World Atlas of Holocene Sea-Level Changes. Elsevier Oceanogr. Ser. 58, 300 pp.
Pirazzoli, P.A., Montaggioni, L.F., Salvat, B. & Faure, G. (1988): Late Holocene sea level indicators from twelve atolls in the central and eastern Tuamotus (Pacific Ocean). Coaral Reefs, 7, 57-68.
Pirazzoli, P.A., Stiros, S.C., Arnold, M., Laborel, J. Laborel-Deguen, F. & Papageorgiou, S. (1994): Episodic uplift deduced from Holocene shorelines in the Perachora Peninsula, Corinth area, Greece. Tectonophysics, 229, 201-209.
NASA/CNES: Observing the oceans by altimetry, www-aviso.cis.cnes.fr
Nunn, P. (1995). Holocene sea-level changes in the south and west Pacific. J. Coastal Res., SI 17, 311-319.
Nunn, P. (1999). Sea level in the last 1200 years. Paper at the PAGES-LOICZ meeting in Hawaii, 1999.
Stewart, I. S & Vita-Finzi, C. (1996): Coastal uplift on active normal faults: the Eliki Fault, Greece. Geophys. Res. Lett., 23, 1853-1856.
Stiros, S.C. & Pirazzoli, P.A. (1998): Late Quaternary coastal changes in the Gulf of Corinth, Greece. IGCP-367, Guidebook , Patras. Univ.

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Paper 2: from " To the House of Lords Economic Affairs Committee", 6 pp, 2005

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Facts and Fiction about Sea Level Change

May low-lying islands and coastal areas be freed from the condemnation to become flooded in the near-future

Nils-Axel Mörner

Head of Paleogeophysics & Geodynamics, Stockholm University, Sweden
President (1999-2003) of the INQUA Commission on Sea Level Changes and Coastal Evolution
Leader of the Maldives Sea Level Project

Climate is becoming increasingly warmer we hear almost every day. This is what has become known as Global Warming. The driving idea is that there is a linear relationship between CO_2 increase in the atmosphere and global temperature. The fact, however, is that temperature has constantly gone up and down. From 1850 to 1970, we see an almost linear relationship with Solar variability; not CO_2. For the last 30 years, our data sets are so contaminated by personal interpretations and personal choices that it is almost impossible to sort up the mess in reliable and unreliable data.

Most remarkable in the record of climatic changes during the last 600 years are the cold periods around 1450, 1690 and 1815 and their correlation with periods of Solar Minima (the Spörer, Maunder and Dalton Solar Minima). The driving cyclic solar forces can easily be extrapolated into the future. This would call for a new cold period or "Little Ice Age" to occur at around 2040-2050. Still, we hear nothing about this. It is as if IPCC and the Kyoto Protocol enthusiasts want to "switch off the Sun itself". Let us take this, at least, as a piece of information to rise our awareness and curiosity.

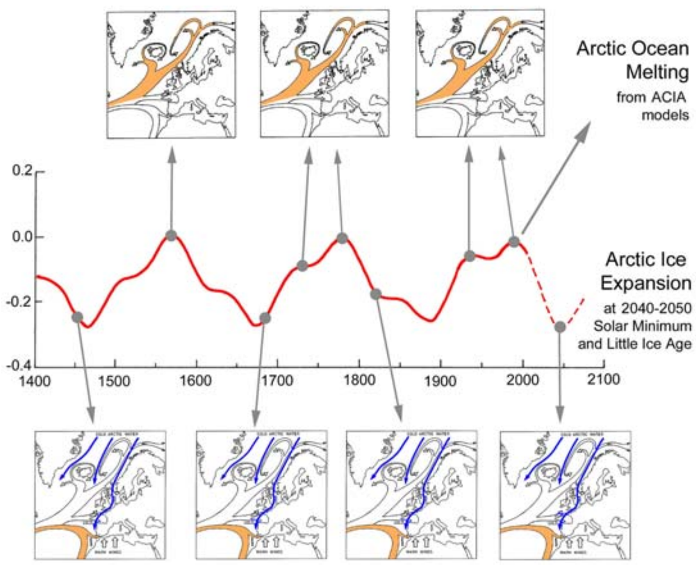

Fig. 1. The main Solar cycle in the last 600 years and its expected extension into the future.
At Solar minima NW Europe, the North Atlantic and the Arctic have experienced cold phases known as "Little Ice Ages".
By year 20040-2050, a new Solar Minimum is to be expected, and hence a new cold phase.

In the global warming concept, it has been constantly claimed that there will be a causal rise in sea level; a rise that already is in the accelerating mode, in the near future to cause extensive and disastrous flooding of low-lying coastal areas and islands. *"It will be the death of our nation"*, says the President of the Maldives, and the people of Tuvalu in the Pacific claim that the flooding has already commenced.

Is this facts or fiction? It is true that we are flooded by this information. But what lies behind this idea? And, especially, what do the true international specialists think?

The recording and understanding of past changes in sea level, and its relation to other changes (climate, glacial volume, gravity potential variations, rotational changes, ocean current variability, evaporation/precipitation changes, etc.) is the key to sound estimates of future changes in sea level.

The international organisations hosting the true specialists on sea level changes are to be found with the INQUA commission on sea level changes and the IGCP special projects on sea level changes. When I was president of the INQUA Commission on *Sea Level Changes and Coastal Evolution*, 1999-2003, we paid special attention just to this question; i.e. proposed rise in sea level and its relation to observational reality. We discussed the issue at five international meetings and by Webb-networking. Our opinion is illustrated in Fig. 2. In view of the Fig. 1 prediction, I have later revised the estimate for year 2100 to: **+5 cm ±15 cm**.

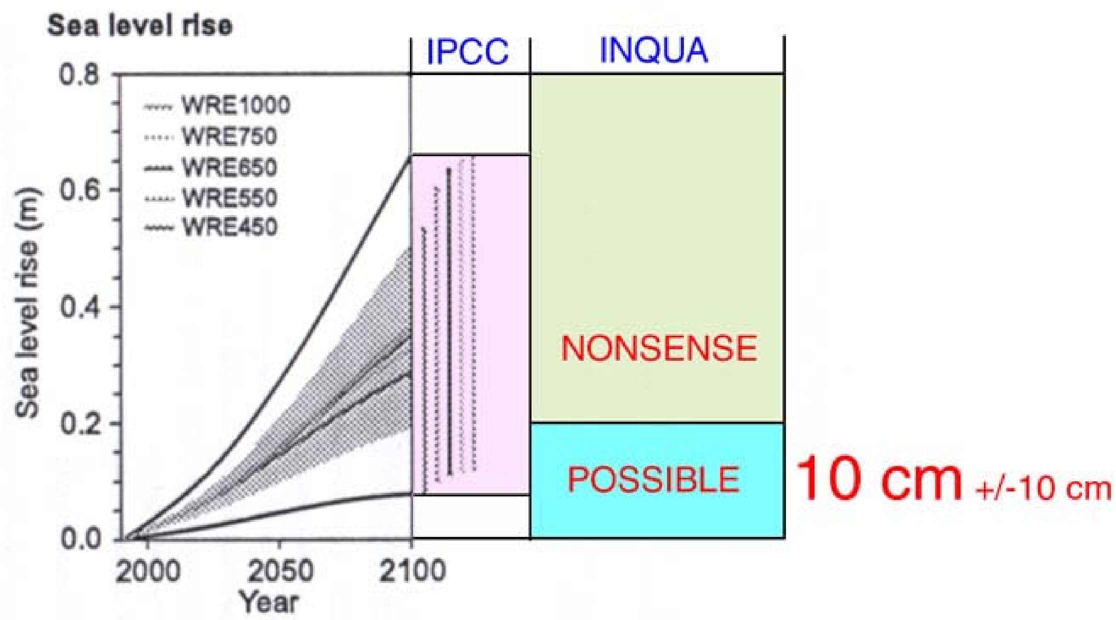

Fig. 2. The sea level rise by year 2100 according to IPCC and its evaluation by INQUA.

Prior to 5000–6000 BP, all sea level curves are dominated by a general rise in sea level in true glacial eustatic response to the melting of continental ice caps. In the last 5000 years, global mean sea level has been dominated by the redistribution of water masses over the globe. In the last 300 years, sea level has been oscillating close to the present level, with peak rates in the period 1890–1930 (Fig. 3).

It is true that sea level rose in the order of **10-11 cm** from 1850 to 1940 as a function of Solar variability and related changes in global temperature and glacial volume. From 1940 to 1970, it **stopped** rising, maybe even fell a little. In the last 10-15 years, we see no true signs of any rise or, especially, accelerating rise (as claimed by IPCC), only a variability around **zero**. This is illustrated in Fig. 3.

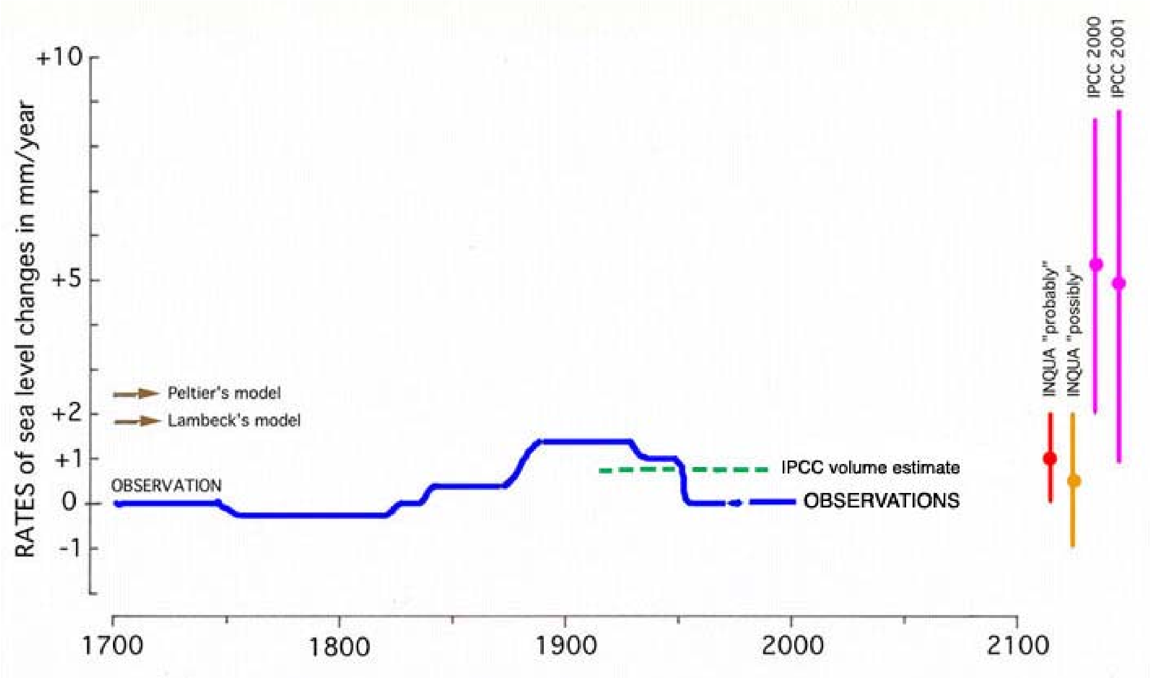

Fig. 3. Observed sea level changes in the past 300 years and estimated changes by year 2100 (from Mörner 2004a).

With the TOPEX/POSEIDON satellite mission in 1992, we now have new means of recording actual sea level changes. The record from 1992 to early 2000 (Fig. 4) lacks any sign of a sea level rise; it records variability around zero plus a major ENSO even in year 1997.

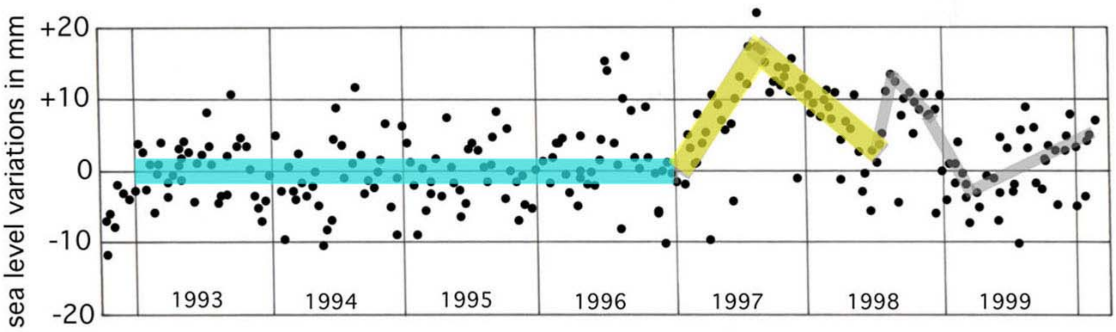

Fig. 4. Satellite altimetry of TOPEX/POSEIDON (from Mörner, 2004a).

When we three years later have the same record extended into year 2003 on the Webb, a tilt has been introduced. This tilt does not originate from the satellite altimetry readings, however, but represents an inferred factor from tide-gauge interpretations. In order to get back to true satellite data, we have to tilt the whole record back to its original data of Fig. 4. When this is done, there is no sea level rise to be seen – only a variability around zero plus a number of high-amplitude ENSO oscillations (Fig. 5). This is why I in Fig. 3 conclude that the sea level remained stationary at around zero for the last 10-15 years (as further discussed in Mörner, 2004a and 2005).

The tide-gauge introduced into the satellite data on the Webb seems to violate observational facts at sites spread all over the globe; not least our NW European data covering both uplifted areas (Fennoscandia, Scotland) as subsiding areas (the North Sea).

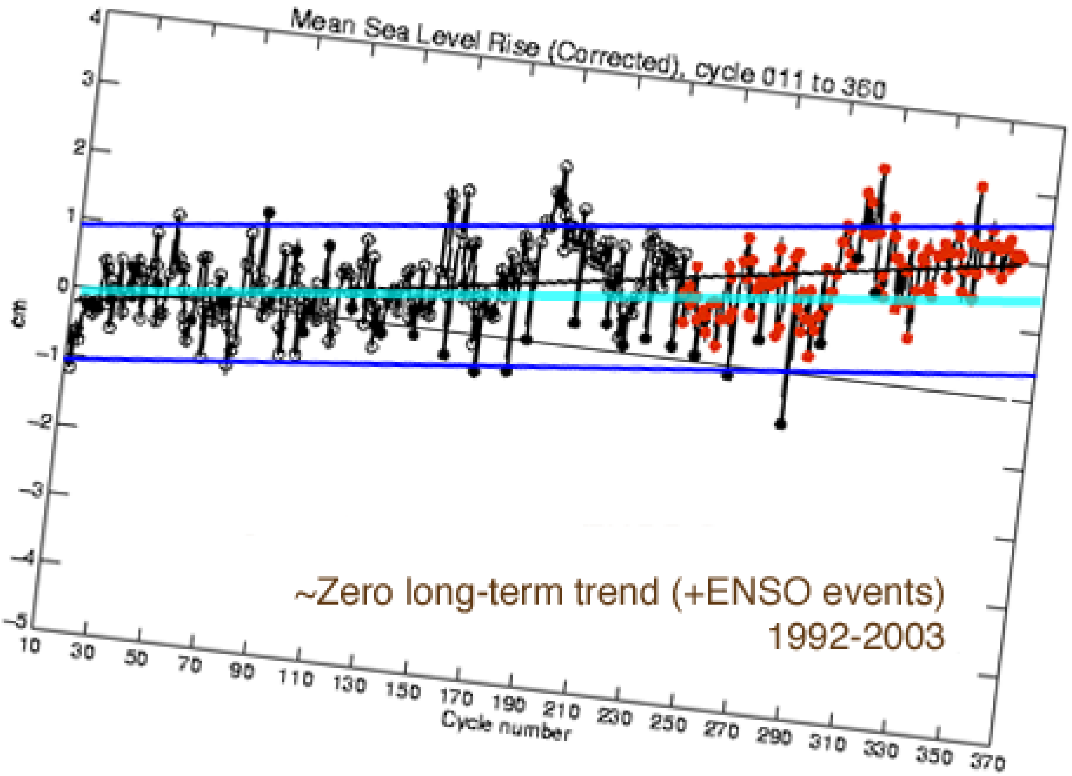

Fig. 5. Satellite altimetry data of TOPEX/POSEIDON tilted back to original level (excluding the tide-gauge factor) providing a variability around zero plus ENSO events (from Mörner, 2005).

From 2000 to the present, we have run a special international sea level project in the Maldives including six field sessions and numerous radiocarbon dates. Our record for the last 1200 years is given in Fig. 6. There are no signs of any on-going sea level rise. It seems all to be a myth.

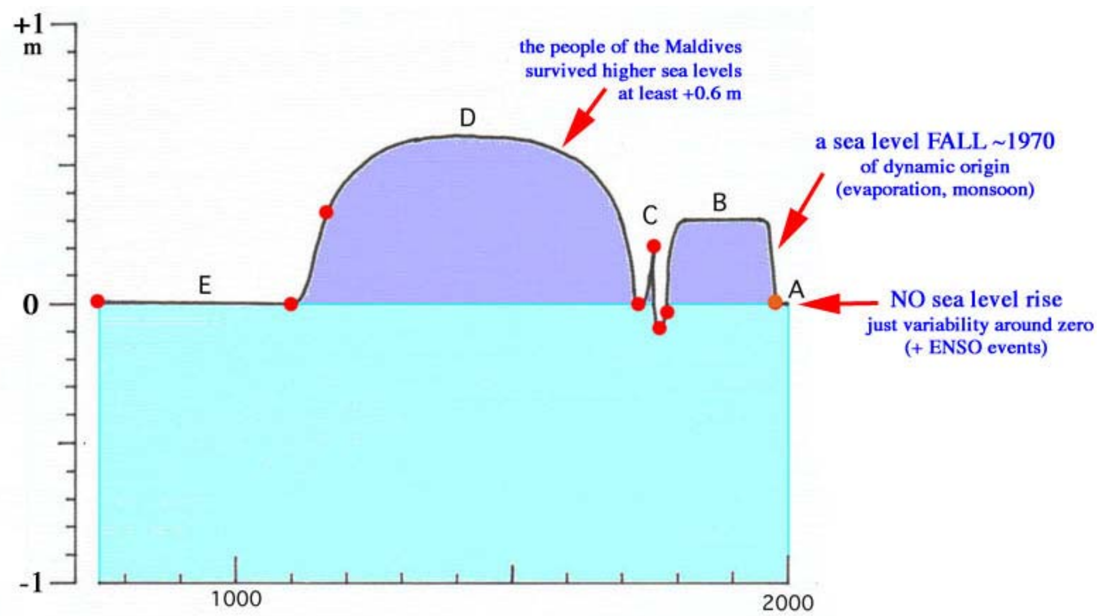

Fig. 6. Our sea level record from the Maldives (see Mörner et al., 2004).

Tuvalu in the Pacific is often said already to be in the flooding mode. The tide-gauge record (Fig. 7) for the last 25 years does not show any rise, however. The truth seems to be that a Japanese pineapple industry had subtracted too much freshwater by that forcing saltwater to invade the subsurface.

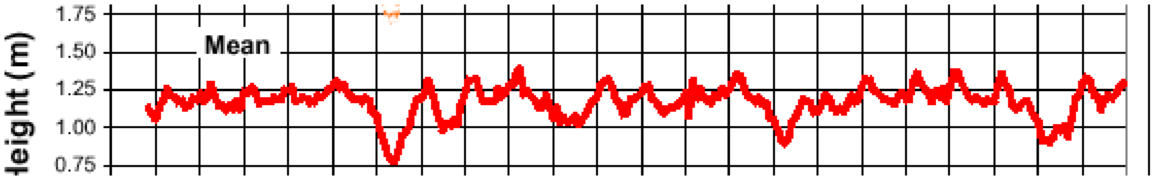

Fig. 7. The Tuvalu tide-gauge record 1978–2003 showing stability around a zero level plus three negative ENSO events (from Mörner, 2004c).

Venice is notorious for its flooding problems. It lies on a delta area subjected to subsidence. Therefore, the sea level variations are superposed on a long-term subsidence trend (Fig. 8). Any rise in sea level would immediately worsen the situation. The last 30 years lack signs of any rise or accelerated rise, on the contrary sea level fell (partly as a function of engineering work).

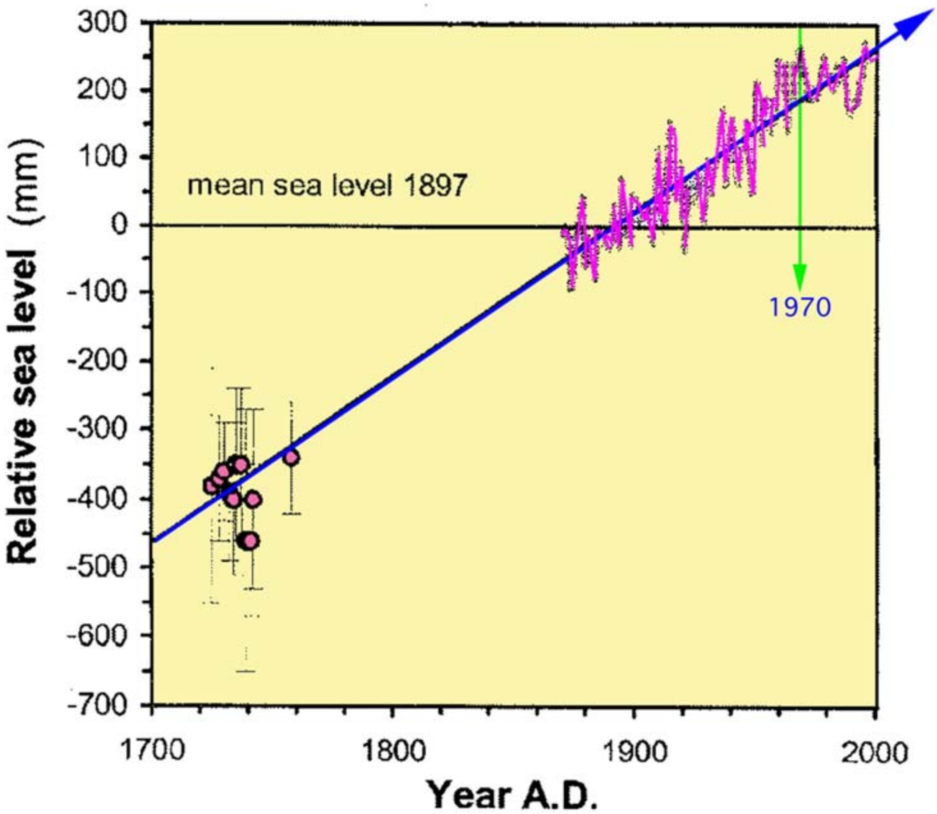

Fig. 8. Observes sea level changes (purple) superposed on a long-term subsidence trend (blue). At 1970 (green arrow), there is a marked change in tendency, partly due to engineering work, but certainly seriously contradicting a sea level rise and especially an accelerated sea level rise.

In conclusion; observational data do not support the sea level rise scenario. On the contrary, they seriously contradict it. Therefore, we should free the world from the condemnation of becoming extensively flooded in the near future.

There are more urgent natural problems to consider on Planet Earth like tsunamis, earthquakes, volcanic eruptions, etc.

Some recent scientific papers by the author

The INQUA Commission – www.pog.su.se/sea
Authors homepage – www.pog.su.se.

Mörner, N.-A., 2005. Sea level changes and crustal movements with special aspects on the eastern Mediterranean. *Z. Geomorph. N.F.*, Suppl.Vol. 137, p. 91-102.

Mörner, N.-A., 2004d. Changing Sea Levels. In: *Encyclopedia of Coastal Science* (M. Schwartz, Ed.), p. 284-288.

Mörner, N.-A., 2004c. Sea level change: Are low-lying islands and coastal areas are under threat? In: "The impacts of climate changes. An appraisal for the future", p. 29-35. International Policy Press.

Mörner, N.-A., 2004b. The Maldives Project: a future free from sea level flooding. *Contemprary South Asia*, 13 (2), p. 149-155.

Mörner, N.-A., 2004a. Estimating future sea level changes. *Global Planet. Change*, 40, 49-54.

Mörner, N.-A., Tooley, M. & Possnert, G., 2004. New perspectives for the future of the Maldives. *Global Planet. Change*, 40, 177-182.

Mörner, N.-A., 2002. Livello dei mari e clima (Sea Level Changes and Climate). *Nuova Secondaria*, 10/2002, p. 43-45.

Mörner, N.-A., 2001. Global and local sea level changes: the interaction of multiple-parametres (hydrosphre, cryosphere, lithosphere, ocean dynamics and climate). *Schr. Deutschen Geol. Gesellschaft*, 14, 3-4.

Mörner, N.-A., 2000b. Sea level changes in western Europe. *Integrated Coastal Zone Management*, Autumn 2000 Ed., p. 31-36, ICG Publ. Ltd.

Mörner, N.-A., 2000a. Sea level changes and coastal dynamics in the Indian Ocean. *Integrated Coastal Zone Management*, Spring 2000 Ed., p. 17-20, ICG Publ. Ltd.

Mörner, N.-A., 1999. Sea level and climate. Rapid regressions at local warm phases. *Quaternary International*, 60, 75-82.

Mörner, N.-A., 1996b. Rapid changes in coastal sea level. *J. Coastal Res.,* 12, 797-800.

Mörner, N.-A., 1996a. Sea Level Variability. *Z. Geomorphology N.S.,* 102, p. 223-232.

Mörner, N.-A., 1995. Earth rotation, ocean circulation and paleoclimate. *GeoJournal*, 37, 419-430.

Mörner, N.-A., 1995b. Recorded sea level variability in the Holocene and expected future changes. In: *Climatic Change: Impacts on Coastal Habitation* (D. Eisma, Ed.), pp. 17-28.

Mörner, N.-A., 1995a. Sea Level and Climate – The decadal-to-century signals. *J. Coastal Res.*, Sp. I. 17, 261-268.

Plus numerous sea level papers in the period 1969-1995.
See also:

Lars Mortensen, 2004; *Doomsday Called Off*, TV-documentary, Danish TV, Copenhagen.

Stockholm, March 30, 2005

Nils-Axel Mörner

Coastal change in the NW Spain: long to short timescales factors since the last interglacial.

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1. Introduction. The coast of the NW of Spain

The coast of NW Spain, mainly that corresponding with Galicia (Spain), can be characterized by its lithological diversity and marked tectonic structure, giving a very irregular planform, with numerous indentations at various scales, from the great rias of the south-atlantic coast, to small inlets. Many rocky sectors alternate with sedimentary complex, usually not too long, but comprising a variety of environments: beaches, lagoons, intertidal flats, dune complex, etc. This irregular form and the presence of coastal mountains, up to 600 m high, helped to create a very special morphogenic conditions in the southwestern atlantic Europe. The intensity of cold processes during the last glacial sea-level regresion generated an extensive sedimentation over abandoned coastal landforms. The sequence of fossilization and exhumation of coastal landforms established a paraglacial or para-periglacial dynamics in which inheritance it is one of the major factors in recent, present and future behaviour.

2.- Inheritance in slow rate coastal changes

The differences in rates of coastal change between sedimentary and rocky coasts implies that, to try to understand the whole behaviour of this type of coastal systems, long and medium time-scales factors must be considered. Coasts are very complex environments, the interface within seas, continents and atmosphere, in which the memory of the system arises as a basic element (Cowell and Thom, 1994). In rocky coastal sectors many landforms are in part inherited, relict or poligenetic, it is difficult to acquire cuantitative data on them, given the wide range of time and space scales in which processes operates (Trenhaile, 1997, 2002). To know more about the morphodynamics of slow changing coasts, and their response to environmental changes in the past it's of great importance to forecast what will happen in the future, and how we can response to coatal changes.

3.- The last Interglacial in the NW Iberian Peninsula: stability *vs* unstability

It is well know that interglacial sea-level was lower than in the present for most of the Pleistocene, whereas was similar to today during Isotopic Stages 11, 9 and at least the substage 5e (Shackleton and Opdyke, 1973). Recently, these relative sea-levels are being revised, and probably there were more interglacial periods with sea-level close or higher than today in the Early and Middle Pleistocene, as well as during substages of the last interglacial always assumed as lower sea-levels as the 5c (Zao, 1999).

The rocky sectors studied, most of them associated to regressive continental deposits, revealed that landforms shaped during a sea-level 2-3 m higher than today and prior to the glacial period are at their original position, as well as high energy coastal deposits as boulder beaches (Blanco Chao *et al*, 2002; 2003). Present evidences suggest a stable tectonic behaviour in the coast of Galicia, at least since the last interglacial. The references to the relative position of coastal landforms and deposits dated as corresponding to the last interglacial suggests that there have been no significant tectonic or isostatic activity at least since the last interglacial. Figure 2 shows the occurrence of uplifting movements on the spanish north coast and Portuguese coast after the last interglacial, which diminishes in intensity westwards and probably, but in a more complex way, northwards. Thus, the Galician coast would have remained stable, being the NW of the Iberian Peninsula a sort of pivotal axis, wich allowed the coastal landforms to remain in their original place (Pérez Alberti *et al.*, 2000).

During the last interglacial highstands, coasts of Galicia were probably subjected to intense erosion, given the existence of inherited erosive landforms: shore platforms, cliffs, sea caves and coarse-grained beaches with abundance of angular clasts.

4.- Cold processes during the marine regression

During the last glaciation, three main cold phases, stand out as responsible for a significant slope unstability (Valcarcel Díaz, 1999). The first phase, prior to 31ky BP was very humid, with a decrease in the mean annual air temperature (MAAT) of about 6ºC as to compare with present average annual temperature values. The second cold period coinciding with the maximum glacial stage, recorded a descend in MAAT of about –12º. The last phase presents two episodes, the first around 16-13 ky BP, with a temperature decrease of up to 6-7ºC and the second in 11-10 ky BP, with a 4ºC decrease in MAAT.

The indented coast and the existence of coastal mountains, sets up a very favourable context for the creation of a morphogenetic environment dominated by cold processes, which resulted in a high unstability of the slopes. From 38 ky BP to the Late-Glacial (10 ky BP) thick deposits were formed on the abandoned Eemian coastlines. The thickness and facies of such deposits are controlled by their position, source materials and distance to the source area (Pérez Alberti *et* al, 1998; Costa Casais *et al.*, 1996).

5.- The Holocene transgression and paraglacial dynamics

The Holocene transgression represents the beginning of a paraglacial dynamic, in which the continental deposits that fossilized the Eemian coastline were eroded. Given the diffrences in extension, thickness, facies and settings of the deposits, the retreat was irregular both in space and time, and was controlled by the volume and nature of the sediments, the energetic environment and the exhumation of fossilized landforms.

Unfortunately, there is no a detailed sea-level curve for the Holocene for the coast of Galicia. Nevertheless, many authors suggest that sea-level reached its present position around 3000 BP, although many coastal systems began to develop before, around 5000 BP, coinciding with the deceleration of the sea-level rise (Dias et al, 2000; Delgado *et al*, 2003; Freitas et al, 2003, Leorri and Cearreta, 2004;). The erosion of the deposits conditioned the recent evolution of the coastal dynamics, and there are evidences of phases of coastal retreat until very recent times (CITAS). These phases of coastal retreat are not directly caused by sea-level oscilations, but by the availability of sediments in cliff deposits. The role of human activity also arises as a very important factor in sedimentation, especially in the last 1000-500 yr (Santos *et* al, 2001; Pérez-Arlucea et al, 2004).

6. Present environment and responses to future environmental changes

Today, many processes are still controlled by the combination of inherited landforms and the disposal of sediments in periglacial cliff deposits. Many cliff profiles, shore platforms and beaches (especially coarse-grained) attained equilibrium with present sea-level and wave conditions recently. This equilibrium is still fragile, and a change in present conditions could lead to significant environmental changes. In some places sea-level changes of a few centimetres could have less effects than an enhanced wave energy caused by a increase in storm frequency and intensity.

References

Blanco Chao, R.; Costa Casais, M.; Martínez Cortizas, A.; Pérez Alberti, A. ; Vázquez Paz, M. (2002) Holocene evolution in Galician coast (NW Spain): an example of paraglacial dynamics. *Quaternary International*. 93-94: 149-159.

Blanco Chao, R.; Costa Casais, M.; Martínez Cortizas, A.; Pérez Alberti, A.; Trenhaile, A.S. (2003) Evolution and inheritance of a rock coast: western Galicia, northwestern Spain. *Earth Surface Processes and Landforms*, 28: 757-775

Costa Casais, M., Martínez Cortizas, A. & Pérez Alberti, A. (1996a) Tipos de depósitos costeiros antiguos entre o Cabo de Fisterra e o de Corrubedo. In A. Pérez Alberti, P. Martini, W.Chesworth, A.Martínez Cortizas (eds.): *Dinámica y evolución de medios Cuaternarios*. Xunta de Galicia. Santiago.

Cowell, P.J. and Thom, B.G. (1994) Morphodynamics of coastal evolution. En R.W.G. Carter and C.D. Woodroffe (eds.) *Coastal evolution. Late Quaternary shoreline morphodynamics.* Cambridge University Press

Delgado, C.; Bao, R.; Alonso, A. and Pagés, J.L. (2003) Evolución paleoambiental inferida por diatomeas de la laguna costera de Traba (A Coruña, NO España) durante los últimos 5000 años In G. Flor (ed). *Actas de la XI Reunión Nacional del Cuaternario.* Principado de Asturias, Consejo de Candamo, Cajastur.

Dias, J.M.A.; Boski, T.; Rodrigues, A. and Magalhaes, F. (2000) Coast line evolution in Portugal since the Last Glacial Maximum until present – a synthesis. *Marine Geology*, 170: 177-186

Freitas, M.; Andrade, C.; Rocha, F.; Tassinari, C.; Munhá, J.M.; Cruces, A.; Vidinha, J. and da Silva, C.M. (2003) Lateglacial and Holocene environmental changes in Portuguese coastal lagoons 1: the sedimentological and geochemical records of the Santo André coastal area. *The Holocene*, 13,3: 433-446

Pérez-Alberti, A.; Blanco-Chao, R.; Costa-Casais, M. (1997) La importancia de los balances de acumulación/erosión y fosilización/exhumación en la evolución geomorfológica de la costa deGalicia. *Dinámica Litoral-Interior. Actas del XV Congreso de Geógrafos Españoles.* Vol. I. pp 205-219

Pérez-Arlucea, M.; Méndez, G.; Clemente, F.; Nombela, M.; Rubio, B. and Filgueira, M. (2004) Hydrology, sediment yield, erosion and sedimentation rates in the estuarine environment of the Ria de Vigo, Galicia, Spain. *Journal of Marine Systems*. 54: 209-226.

Santos, L.; Bao, R. and Sánchez Goñi, M.F. (2001) Pollen record of the last 500 years from the Doniños coatal lagoon (NW Iberian Peninsula): changes in the pollinic catchment size versus paleoecological interpretation. *Journal of Coastal Research*, 17, 3:705-713

Shackleton, N.J., and Opdyke, N.D. (1973) Oxygen isotope and palaeomagnetic straitgraphy of Equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 105 year and 106 year scale. *Quaternary Research,* 3.

Trenhaile, A.S. (1997) *Coastal dynamics and landforms*. Clarendon Press. Oxford.

Trenhaile, A.S. (2002) Rock coasts, with particular emphasis on shore platforms. *Geomorphology*, 48:7-22

Trenhaile, A.S., Pérez Alberti, A., Martínez Cortizas, A., Costa Casais, M., Blanco Chao, R. (1999) Rock coast inheritance: an example from Galicia, Northwestern Spain. *Earth Surface Processes and Landforms*, 24: 605-621.

Valcárcel Díaz M. 1999. Evolución geomorfológica dinámica de las vertientes en el NE de Galicia: importancia de los procesos de origen frio en un sector de las montañas lucenses. Tesis Doctoral, Universidade de Santiago, Spain (unpublished).

Littoral 2002, The Changing Coast. EUROCOAST / EUCC, Porto – Portugal
Ed. EUROCOAST – Portugal, ISBN 972-8558-09-0

Stability of Sedimentary Cliffs in the Coast of Galicia (NW Spain): Long Term Inheritance Influence in Rocky Coastal Systems

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Abstract

This work is focused in the factors controlling the stability of sedimentary cliffs in the Galician coast, specially those related with inherited processes. The analysis of three sectors allows to conclude that the main factors controlling their stability are the facies, extents and thickness of the deposits, the exposition to wave action, the existence of exhumated Eemian landforms and the existence of boulder beaches developed during the Holocene erosion of the deposits. Knowing the basis of the present dynamics of this settings, and the strong influence of the inherited processes, is possible to predict their response in a scenery of rising sea level.

1. INTRODUCTION

Given their relative fast rates of erosion, the most of the research carried out in sedimentary cliffs are constrained to time-scales in which the use of monitoring systems are possible. In this paper we analyze the stability of sedimentary cliffs in the coast of the NW Iberian Peninsula, defining them as those developed on non consolidated sediments deposited during the middle and Late Weichselian. The presence of that cliff deposits leads to consider their morphodynamics in medium to long timescales.

One of the most important characteristics of the Galician coast is the existence of thick deposits sedimented over the abandoned Eemian coastline during the glacial regression (Pérez Alberti *et al.* 1998, 2000). The very indented outline of the coastline and the presence of mountains close to the coast, of up 600 m of altitude, were important factors for the formation of such deposits. Periglacial and snow-melt processes were the dominant, resulting in a great variety of facies, with a high amount of coarse debris.

The altitude, gradient, aspect and distance from the source area have had a strong control in the facies, thickness and extent of the deposits (Costa Casais *et al*. 1996; Pérez Alberti *et al*. 1998; Trenhaile *et al*., 1999). The Holocene transgression established a paraglaciar dynamics, characterized by the erosion of the ancient deposits and the exhumation of the Eemian coastal landforms (Blanco Chao *et al.* 2002; Pérez Alberti *et al.* 2000).

The remnants of that deposits are today coastal cliffs, being their morphodynamics strongly controlled by the inheritances of the evolution during the Holocene marine transgression.

2. STUDY AREA

We have selected three different places: Area Longa beach in the north coast, Arnela de Lourido in the Atlantic coast and Oia in the southwest (Figure 1). The three settings represents different coastal environments, in which the balance between subaerial and marine processes, the distribution of wave energy, the variety of sedimentary facies and the exhumation of ancient landforms are the main factors for their stability.

Waves of Galicia are between 1 and 2.5 m in height for about 80 % of the year, coming from NW, W and SW, being the most of the waves higher than 3 m generated by Atlantic low pressure in winter. The tidal environment is semidiurnal with a mean tidal range of 2.5 m, and spring tidal range of between 3.75 and 4 m.

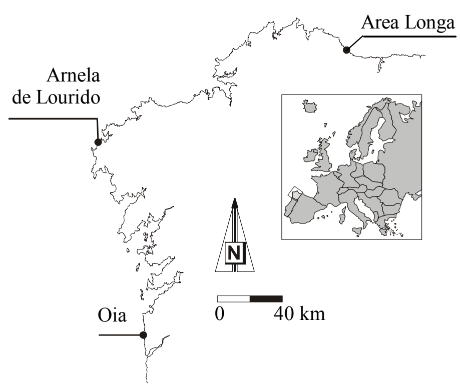

Figure 1. Location of the three study areas.

2.1 Area Longa beach

The beach plan runs SE-NW with a slightly concave plan, about 850 m in length and 60 m wide at low tides. A rocky point divides the beach in two sectors: the northwestern, sandy and with a dissipative profile, and the southeastern, more reflective given the higher amount of shingle and cobbles.

The response to high energy wave action is mainly the sweep of sand to subtidal levels, which is reposed to high tidal levels during fair weather periods. The geology of the area are stratified schists and quartzite dipping 20º to the SE.

The sedimentary cliffs at the back of the beach diminishes their elevation from 8 m in the southeast to 5 m in the northwest. In the southeast beach the cliffs are composed of sands, with some stone and gravel layers, and, at a few places, coarse clastic periglacial facies.

At the south side of the rocky point an ancient shore platform and a cobble beach is fossilized by sands and gravels. In the northwestern sector, the base of the deposit are composed of a cohesive rich-organic layer and a gray clay layer. Over the clay, the sediments are poorly ordered sands with angular gravels. In the decade of 70's, a protection wall of concrete was built trying to stop the retreat of the cliff. The wall was destroyed after several severe storm and the fragments are today at a mean distance of 2 m from the cliff (Figure 2).

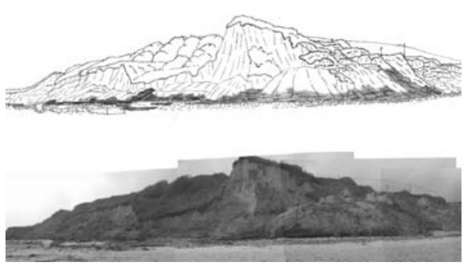

Figure 2. Northwest sector of Area Longa cliffs.

2.2 Arnela de Lourido

This is a narrow funnel-shaped embayment, associated to a north-south running fracture, in which the valley of a short river was opened. The embayment is around 800 m in length and between 600 and 50 m wide in the mouth and the bottom respectively. In plan, it shows an asymmetric shape, being the western flank more straight. The surrounding topography has altitudes of 174 and 162 m in the east and west slopes, being the gradient of the first steeper than the western one. Both slopes are regularized by coarse periglacial slope deposits with abundant blocks of up 1 m of diameter.

The geology of the area are granitic rocks with a dense joint pattern. In the mouth of the embayment the deposits that fossilizes the slopes remains hanged over a rock sea cliff, in a typical slope-over-wall profile. In the internal section, the sediments has been eroded developing cliffs fronted by a irregular intertidal rocky surface of about 5 m wide, covered at the upper levels by an accumulation of blocks. In the flanks the cliffs deposits are between 2 and 4 m of poorly ordered, heterometric material, with a great volume of blocks, gravels and cobbles, with some sand layers. The cliffs have a vertical profile with a dense cover of terrestrial vegetation. At the bottom of the embayment the cliff has 8 m of thickness; the base of the deposit is a rich-organic clay layer, covered by 5 m of sand and angular gravels and 2 m of angular sands with gravel and pebble lenses (Figure 3).

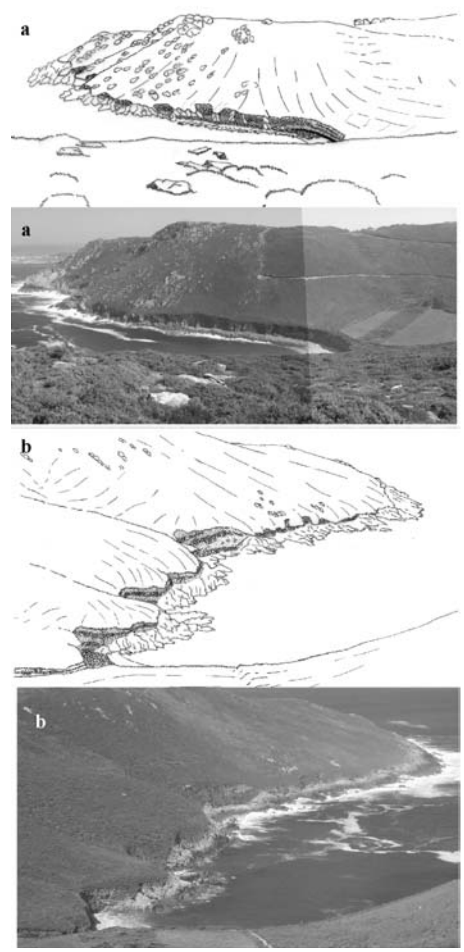

Figure 3. East (a) and west (b) sides of the Arnela embayment. The flat surface with meadows are the top of the sediments infilling the valley.

At both ends of the beach the sediments are coarser, given the proximity to the slopes and the dominance of periglacial and slope processes during their deposition.

The bottom cliff section is fronted by a narrow sand beach, with well rounded boulders at both sides. At the back of the cliff in the bottom of the embayment, there is a gentle surface extending 200 m landwards, which corresponds with the fluvial, fine sediments that infilled the valley.

2.3 Oia

This is a nearly straight, low rocky coast running north to south in the southwestern coast of Galicia, at the foot of the A Groba mountains (663 m). Between the steep slopes of the mountains and the coastline a wide and gentle surface extends. The substrate are fractured granitic rocks, with many quartz dyques and a few outcrops of metamorphic rocks in the slopes. The selected place is a small and shallow inlet, in which a sedimentary cliff was developed, fronted by a irregular shore platform with a maximum width of 50 m and composed of two segments, with 3º and 2.5º of gradient in the upper and lower respectively. The upper segment of the platform is covered by a boulder beach 20 wide with a mean gradient of 9º. The cliff deposit are the remnants of the extensive sediments that fossilized the coastline during the glacial regression. The cliff has between 8 and 10 m of cobbles and blocks alternating with sand and gravel lenses, result of snow-melt processes during the last glacial period (Costa Casais 2001). The basal layer is a rich-organic layer, with an erosive contact with the upper gravel and cobble sediments. The upper section are sand layers alternating with subrounded cobbles and block layers (Figure 4).

Figure 4. The cliff deposit, shore platform and boulder beach at Oia.

3. STABILITY OF SEDIMENTARY CLIFFS

Given the differences in the three morphological settings, the type and rate of cliff recession has significative variations, ranging from very active mass movements to a complete stability.

In Area Longa Beach the cliffs shows evidences of previous rotational and planar slides along the cliff face. In the southeast sector the front of the cliffs are stabilized by terrestrial vegetation, except at the north end beside the rocky point dividing the beach, where the cliff maintains a steep profile and is common the occurrence of slumps. In the northwest sector the existence of rotational slides can be identified by the bowl-shaped scar, and the tilting of the peat and clay layers. The wide sandy beach has a dissipative profile, with the beach-cliff junction at the spring high tide level, in a way that the waves only attacks the base of the cliffs during the maximum high tides, and only with significant energy during the most severe storms. The existence of a platform developed on the organic and clay layers, normally underlying the sand, but usually exposed at the back of the broken wall, reveals that even with the low energy of the waves reaching the cliff, the sand act as an abrasive tool to erode the peat and clay layers, a process that has been confirmed as an important factor in the erosion of cohesive sediments (Kamphuis 1990; Kamphuis and Asce 1987; Davidson and Ollerhead 1995). The wave action is not the only process responsible of the instability of the sedimentary cliffs in Area Longa. The impermeability of the clay layers at the base is a favorable condition to develop a shear plane for rotational and planar slides induced by rainfall and seepage. The alternance of periods of active retreat by waves are interrupted by periods during which debris protect the cliff base, in a cyclical behavior common on tall cliffs (Quigley *et al.* 1977).

In Arnela, the degree of stability changes from the sides to the bottom of the embayment, as a result of differences in sedimentary facies and wave action. The incoming waves entering the embayment are refracted maintaining a higher velocity and wave energy at the center, where the deep of water allows them to reach the sandy beach and the fine-sediment cliff. At the flanks, the intertidal rocky surface and the block accumulation dissipates the most of the energy of waves before they reach the cliff. Consequently, the retreat due to subaerial processes is the dominant in the coarser sediments of the flanks, and the wave action at the bottom, where the sand gives an abrasive tool for erode the cohesive organic layers. During periods of strong wave action the surficial sediment of the beach may be swept, lowering the beach profile and exhumating the platform shaped on the organic layer. This reduction of the beach sediment thickness results in the reactivation of abrasion on the plastic platform (Davidson and Ollerhead 1995). Given the reduced volume of the beach the sand swept don't develops a wider dissipative profile, allowing the waves to reach the cliff base more frequently.

At the east side of the beach the erosion of the coarser sediments supplied the material for a small boulder beach, which protects the cliff from wave action. At the west end, the presence of a rocky point causes a higher turbulence and erosive capacity of the waves on the periglacial facies.

In the sector of Oia the Holocene erosion of the continental sediments that fossilized the Eemian coastline exhumated the rocky shore platform, and supplied the material for the development of the boulder beach (Blanco Chao 1999; Blanco Chao *et al.* 2002). The high wave energy of the coastal sector allows an intense abrasive work on the shore platform at the seaward end of the boulder beach. Nevertheless, the dissipation of the wave energy over the platform and specially over the boulder beach prevents the cliff to be eroded by waves. The dense terrestrial vegetation that covers the cliff face and the upper segment of the boulder beach are the evidence of the high degree of stability, result of an state of equilibrium between the present wave regime and a boulder beach without nourishment.

In the three settings, the ancient deposits are the only source of sediments, being almost exhausted during the Holocene transgression. In a sea level rising scenario, the behavior of each system will be different: in Oia, the reactivation of the cliff erosion could release a high amount of coarse debris, meanwhile the shore platform and the existent boulder beach will continue with their protective role. In Arnela, the volume of sediment available at the flanks could exhumate the rock, becoming to a slope-over-wall profile, meanwhile at the bottom there is still a huge volume of fine sediments. Since the Arnela embayment is a closed sedimentary cell, the increase in sand and gravel supply will increase the volume of sand at the beach, countering the cliff retreat. In Area Longa, the scarce capability of the embayment to store sand would lead to an intense phase of cliff retreat, more severe since they are today the most unstable of the three places studied. Today, the retreat of the cliff by active rotational slides had affected a secondary road, but if the rate of erosion increase a railway and a first order road would be affected.

4. CONCLUSIONS

The sedimentary cliffs in the coast of Galicia are the result of the erosion of the sediments deposited on the abandoned Eemian coastline between the Middle and Late Weichselian.

In the coast of Galicia the factors accepted as major controls in the erosion of sedimentary cliffs, i.e. wave action, strength of the sediments, and the presence of beaches fronting the cliffs are consequence of inherited landforms and processes.

The most significative is the existence of exhumated ancient landforms as the shore platforms, in which the energy of waves are expended.

The beaches and their sedimentary characteristics are also determined by the characteristics of the ancient deposits.

The possibility of a sea level rise will lead to the almost exhaustion of sediment supplies to the systems since the ancient deposits are the only sedimentary source.

REFERENCES

Blanco Chao R. 1999. Formas y procesos geomorfologicos diferenciados en las costas de Galicia: morfodinámica y evolución de un sector de costa rocosa: Cabo Silleiro - A Garda (Pontevedra). Tesis Doctoral, Universidade de Santiago, Spain (unpublished).

Blanco Chao, R.; Costa Casais, M.; Martínez Cortizas, A.; Pérez Alberti, A.; Vázquez Paz, M. 2002. Holocene evolution on Galician coast (NW Spain): an example of paraglacial Dynamics. *Quaternary International* (In press)

Costa Casais M. 2001. Análise sedimentaria e reconstrucción paleoambiental da costa Atlántica de Galicia. Tesis Doctoral, Universidade de Santiago, Spain (unpublished).

Costa Casais, M., Martínez Cortizas, A. & Pérez Alberti, A. 1996. Tipos de depósitos costeiros antiguos entre o Cabo de Fisterra e o de Corrubedo. In (A. Pérez Alberti, P. Martini, W.Chesworth, A. Martínez Cortizas eds.) *Dinámica y evolución de medios Cuaternarios*. Xunta de Galicia, Santiago; pp. 417-430. ISBN 84-453-1627-3

Davidson-Arnott, R.G.D. & Ollerhead, J. 1995. Nearshore erosion on a cohesive shoreline. *Marine Geology,* 122, pp. 349-365.

Forbes, D.L. & Syvitski, J.P.M. 1994. Paraglacial coasts. In R.W.G. Carter & C.D. Woodroffe (eds.) *Coastal evolution. Late Quaternary shoreline morphodynamics.* Cambridge University Press, Cambridge; pp. 373-424. ISBN 052141976X

Kamphuis, J.W. 1990. Influence of sand or gravel on the erosion of cohesive sediment. *Journal of Hydraulic Research*, Vol 28, nº 1, pp. 34-53.

Kamphuis, J.W. & ASCE, M. 1987. Recession rate of glacial till bluffs. *Journal of Waterway, Port, Coastal and Ocean Engineering*. Vol 113, nº 1, pp. 60-73.

Pérez Alberti, A., Blanco Chao, R. & Costa Casais, M. 1998. Depósitos sedimentarios antiguos en la costa atlántica gallega: tipología, localización y área fuente. Un análisis comparativo. In (A. Gómez Ortiz & F. Salvador Franch eds.) *Investigaciones Recientes de la Geomorfología Español.* Universitat de Barcelona, Huesca; pp. 341-346. ISBN 84-87779-33-6.

Pérez Alberti, A., Blanco Chao, R., Costa Casais, M., Vázquez Paz, M. & Valcárcel Díaz, M. 2000. Dinámica y evolución de las costas rocosas. El ejemplo de Galicia. In (J.R. de Andrés; F.J. Gracia eds.) *Geomorfología litoral. Procesos Activos*. Instituto Tecnológico Geominero de España, Madrid; pp. 175-185. ISBN 84-7840-395-7

Quigley, R.M., Gelinas, P.J., Bou, W.T. & Packer, R.W. 1977. Ciclic erosion-instability relationships: Lake Erie north shore bluffs. *Canadian Geotechnical Journal*, 14, pp. 301-323.

Trenhaile, A.S., Pérez Alberti, A., Costa Casais, M., Martínez Cortizas, A. & Blanco Chao, R. 1999. Rock coast inheritance: an example from Galicia, Northern Spain. *Earth Surface Processes and Landforms,* 24, pp. 605-621.

Late Pleistocene-Holocene environmental changes
(NW coastal zone of Portugal)

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Abstract

The coastal zone of northwest Portugal can be subdivided into two main geomorphological sectors, Espinho being the boundary between both (Granja 1999).
In the first sector, between the Minho River and the town of Espinho, the coastal segments are narrow and consist of estuaries, sand- and shingle beaches with rocky outcrops, and generally poorly developed dune systems (foredunes, parabolic and some migrating dunes associated to blow-outs). The estuaries and the foredunes in particular are very degraded by human activities. Although today no lagoons are present, with the exception of the residual lagoon of Apúlia, a large palaeolagoon system was present in the sector during the Holocene.
The second sector, between Espinho and the Mondego Cape, consists of wider segments, coastal lagoons, and Holocene dune systems (foredunes, parabolic and transverse dunes).
The two sectors have had different histories at the macro- and mesoscale, conditioned by their main geomorphological features and to their relative exposition to the main forcing factors, leading to different responses.
At the macroscale, i.e. 100-1000 years, sea-level changes and neotectonic activity were the main forcing factors, while at the mesoscale, i.e. 1-100 years, climate fluctuations were.
Macroscale sea-level changes and neotectonic activity seem to have played a dominant role in the (palaeo-)environmental evolution of the coastal zone since the Late Pleistocene. The mesoscale role of climate is still difficult to assess at the present stage of knowledge. However, dispersed field data point to climatic changes, some of them also referred to in historical documents. This is the case of e.g. the environmental changes on estuarine saltmarsh (Cávado) and lagoon (Torrão do Lameiro) evolution that point to changes in storminess, as does also the burying of 20th century human settlements by aeolian sands (Bonança, Esmoriz). The development of Medieval dune systems in both sectors is attributed to the Little Ice Age, and thus to climatic changes, as does the overwashing of some palaeolagoons during high medieval times (at Esposende, Torreira, and Tocha).
Sediment starvation on the shoreface is postulated to be one of the major causes for coastal erosion since at least the 15th century. Investigations in the coastal area of Northwest Portugal show evidence of changing patterns of sediment supply during the Quaternary.

Sector north of Espinho

In this segment several geomorphological units and correlative deposits were found. From east to west: 1) an older abandoned cliff of uncertain Quaternary age, associated with 2) a higher platform, 40-0 m high, with relict Pleistocene (?) marine deposits, 3) a younger abandoned cliff cut into this higher platform, associated with 4) a lower platform, consisting of two sub-units, one between 30 and 10m high with Pleistocene (52500±640 OSL) deposits on the landward side of the platform, and one less than 10m high, with Holocene lagoonal deposits (3250±100 to 360±40 yr BP) on the seaward side, extending till the sub-tidal area. The lower platform in turn, is covered to seaward by 5) dune systems of Medieval to sub-recent ages.

Sector south of Espinho

Three main outcrops of different ages and history, characterise this sector: the S. Pedro da Maceda Beach, the Cortegaça Beach, and the Silvalde-Paramos Beach.

S. Pedro da Maceda Beach has a history of alternating dry and wet aeolian Pleniglacial environments (Granja et al., in prep) sometimes draped by thin soils, one containing trunk remains of *Pinus sylvestris nigrita* (27150±250 to 19910±260 yr BP). The base of this sequence, on the bedrock, can have still another origin (fluvio-lacustrine).

At Cortegaça beach, another history is revealed. The outcrops point to a Holocene barrier environment in which top a podzol was formed (3490±100 to 950±80 yr BP). This podzol would have been flooded (15th century) before the initiation of the dune systems (19th century) that are the main feature of the present landscape. The Cortegaça deposits are lying over those of Maceda, through an apparent (temporal) unconformity.

At Silvalde-Paramos Beach, outcrops have shown a Holocene environment over the bedrock, a lagoonal brackish environment that silted up till a wet environment where trees (species still not determined) grew (1180±45 to 1020±80 yr BP). A core near Paramos, yielded the whole sedimentary succession (4920±105 to 440±60 yr BP). This lagoonal deposit is in unconformity with the Cortegaça and Maceda deposits, being partially contemporaneous with the formation of the podzol.

To reconstruct the sedimentary environments and distribution patterns of the second sector (south of Espinho), cored boreholes, supplemented by a counter-flushed extension, and geophysical data, have been analysed in order to fill the knowledge gap concerning the subsurface.

Grain-size data obtained from sediment samples in the cored parts of the boreholes, were statistically analysed, confirming the interpretation of the coastal outcrops and pointing to a wet aeolian dune and interdune environmental setting. OSL and ^{14}C dates obtained from the eroded cliff faces on the beach, indicate Pleniglacial to Late Glacial ages. The counter-flushed extensions of the boreholes show more pebble-rich beds, which may correspond to interglacial marine highstands (Granja et al, in prep).

Some of these pebble-rich beds could be contemporaneous with those found north of the Cávado River (in the first sector) in deposits over the bedrock, on the lower platform (sub-unit of 30-10 m) (e.g. Cepães).

The wet aeolian palaeo-environments of the sector south of Espinho do not have equivalents (or at least, not found till now) in the northern sector.

The Holocene lagoonal deposits of Silvalde-Paramos could correspond to a similar

palaeoenvironment set in in the northern sector. Meanwhile, in the northern sector the lagoonal beds do not contain shells as the southern deposits do.
Ongoing research could answer some of the main questions that still arose.

References

García-Amorena I., Granja H. M., Pissarra J., Estela Silva R., Rubiales J. M. & Soares de Carvalho G. (accepted). Late Quaternary palaeoenvironmental reconstruction in the northwest coast of Portugal: an example of a geo-site archive exploration. IV *International Symposium Progeo on the Conservation of Geological Heritage,* Braga, September 2005

Gomez-Orellana, R. L., Ramil-Rego, P. & Sanches, S. M., 2001. Landscape modifications during the Upper Pleistocene – Holocene in the NW Atlantic margin of Iberian Peninsula. *Estudos do Quaternário*, Journal of the Portuguese Association for Quaternary Research 4, 79-96

Granja H. M., 1999. Evidence for Late Pleistocene and Holocene sea-level, neotectonic and climatic indicators in the northwest coastal zone of Portugal. *Geologie en Mijnbouw,* Kluwer Academic Publishers, 77 (3-4): 233-245

Granja H. M., 2002. Reconstituição paleoambiental da zona costeira, a norte da laguna de Aveiro, desde a Idade Média até à actualidade. In: *O Litoral em Perspectiva Histórica (Séc. XVI a XVIII).* Instituto de História Moderna, Universidade do Porto, 93-109

Granja H. M., Ribeiro I., Carvalho G. S. & Senos Matias M., 1999. Some neotectonic indicators in Quaternary formations of the northwest coastal zone of Portugal. *Physics and Chemistry of the Earth*, Elsevier Science, Ltd., 24(4): 323-336

Granja, H.M., de Groot, T.A.M. & Costa, A.L. (in prep.) Evidence for Pleistocene wet aeolian dune and interdune accumulation, S. Pedro de Maceda, NW Portugal.

López Cancelo L., 2004. *Cambios paleoambientales en el NW peninsular durante el Holoceno determinados a partir del estudio de foraminíferos bentónicos*. PhD thesis, Universidade de Coruña, 293 p

Ribeiro I., 2003. *Deformação neo-tectónica pós-plistocénica na zona costeira entre os rios Minho e Ave*. PhD thesis, Universidade do Minho, 355 p

Soares de Carvalho G. & Granja H. (accepted). *As mudanças da zona costeira pela interpretação dos sedimentos plistocénicos e holocénicos (metodologia aplicada à zona costeira do noroeste de Portugal).* In: Livro de homenagem à Professora Doutora Rosa Fernanda Moreira da Silva, Faculdade de Letras, Universidade do Porto

Thomas P. J., Murray A. S., Granja H. M. & Jain M. (in revision). Optical dating of Late Quaternary coastal deposits in Northwest Portugal

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Porto littoral: the influence of tectonics in sea level changes and coastal morphology.

Abstract

The plotting of relative sea level variations for several stations belonging to the Iberian Peninsula shows different trends. These trends seem to be related with the diastrophism affecting more intensely the Northern and Southern façade of the Peninsula.

Porto is located on the riverside of Douro, the most plentiful river in the Iberian Peninsula, which is deeply entrenched on the littoral platform, close to its mouth. This littoral platform is a quite common feature at Portuguese coastline, surrounding it almost in all its length.

At Porto area the littoral platform contains several outcrops of cenozoic deposits and it is limited to the interior by a generally step relief (marginal relief) which is probably a fault scarp acting mostly after the earlier deposits had been formed

A careful study of those deposits showed up that they are not at all primarily marine as its situation, facing the Atlantic could make us suppose: the marine deposits are disposed only in a narrow fringe lower than 40 m high. The upper deposits have a clear fluvial origin, they go up until 130m and they are clearly disturbed by tectonics.

Apparently, a sub-meridian accident produced the subsidence of the narrow fringe (1-2km maximum width) where the marine deposits are lying. This seems to indicate that the sea retouched this lower block when it subsided along that sub-meridian fault (fig. 1).

The newer, marine deposits can be assigned to at least three levels (around 30m, 20m and from 10 to recent sea level) distinguished by sedimentary criteria. They are not everywhere at the same altitude, but they are disposed in a irregular up and down pattern, with a general trend indicating a subsidence towards the meso-cenozoic basin that evolved like an aulacogen during meso-cenozoic times, and begins at Espinho, 15km south of Porto (Lusitanian basin).

We will focus on the tectonic style and regional tectonic framework that created the differences between 2 places, one of them appears at the north and the other at the south of Porto area.

A aeolian sandstone lying upon a marine deposit (Labruge beach, 15 km north of Douro river mouth, 5m above mean sea level) was TL dated, with a result of 84kaBP. Therefore, the underlying marine deposit must be from last interglacial. In addition, it is possible to correlate other iron-cemented sandstones covering old marine platforms with a similar position, found at several places in this coastline, with the same interglacial.

Admitting the sedimentology-based correlation of marine deposits at the North of Espinho they seem to be balanced to south, in the direction of Lusitanian basin (fig. 2).

At Aguda beach, some 12 km south of Porto the last interglacial marine deposit is laying at a lower altitude, 1m above mean sea, level fossilizing a wide marine platform. Upon it, we found some lagoon deposits. The upper part of them (around 4-5m above msl) was TL dated with a result of ca. 8ka BP.

So, this lagoon deposits are a testimony of continental conditions during last glaciation and/or flandrian transgression.

The lagoon deposit is covered by **another** marine sandstone, about 5 m above msl. Therefore, in this area, there are clear evidences of two marine sea levels with about 120 ka difference of age lying at quite similar altitudes (fig. 3).
In this coastal area, generally the marine deposits dispose themselves in a staircase fashion, the older ones at higher altitudes and the more recent at lower altitudes - which is typical of a slow uplifting area (Cabral, 1995). However, at Aguda, near Espinho, the last interglacial marine deposit is superposed by a flandrian one – and this disposition suggests that the uplifting trend prevailing in the north is replaced by a subsiding one.
At the south of Espinho the marine deposits seem to vanished – may be they are simply buried under the fini-Würmian and Holocene aeolian sands that cover the western part of the littoral platform in that area.
These aeolian sands include a podzol cemented old dune, which seems balanced to the north towards Esmoriz lagoon (see a development at Araújo, 2002 in the next pages).
In conclusion, the marine deposits at the North of Espinho and the würmian/holocene deposits at the South of this city show opposite tectonic trends that seems to define a tectonic depression corresponding approximately to the localisation of Esmoriz lagoon.
At Espinho, coastal erosion began in the middle of nineteen century. We think that the rising of sea level that began after the end of Little Ice Age, together with a possible subsiding trend, may be responsible for the severe erosion endured by this area.

Some bibliography

Araújo, M. A., 1985. O estudo do litoral norte de Portugal (região do Porto); alguns problemas metodológicos. Actas da I Reunião do Quaternário Ibérico, Vol. I, Lisboa, Grupo de Trabalho Português para o Estudo do Quaternário, 269-279.

Araújo, M. A., 1991. Evolução Geomorfológica da Plataforma Litoral da Região do Porto. Unpublished PhD thesis, Faculdade de Letras da Universidade do Porto, 534 p.

Araújo, M. A., 1997. A plataforma litoral da região do Porto: Dados adquiridos e perplexidades. Estudos do Quaternário, Lisboa, APEQ, 1, 3-12.

Araújo, M. A., 2000. Depósitos de fácies continental e marinho na plataforma litoral da região do Porto: importância da tectónica na sua organização espacial. Ciências da Terra, Lisboa, Universidade Nova de Lisboa, 14, 111-122.

Araújo, M. A., 2002. Relative sea level, diastrophism and coastal erosion: the case of Espinho (Portuguese NW coast). Littoral 2002 proceedings, Associação Eurocoast-Portugal, Instituto de Hidráulica e Recursos Hídricos, Faculdade de Engenharia da Universidade do Porto, 2, 125-132.

Araújo, M. A., 2004. O final do Cenozóico na plataforma litoral da região do Porto. Geomorfologia do NW da Península Ibérica, Porto, FLUP-GEDES, 117-138.

Araújo, M. A., Gomes, A., Chaminé, H. I., Fonseca, P. E., Gama Pereira, L.C., Pinto de Jesus, A., 2003. Geomorfologia e geologia regional do sector de Porto-Espinho (W de Portugal): implicações morfoestruturais na cobertura sedimentar cenozóica. Cadernos Laboratorio Xeolóxico de Laxe, Coruña, 28, 79-105

Cabral, J., Ribeiro, A., 1988. Carta Neotectónica de Portugal (escala 1:1000000), Serviços Geológicos de Portugal, Lisboa.

Cabral, J. M. L. C., 1995. Neotectónica de Portugal Continental, Memórias do Instituto Geológico e Mineiro, Lisboa, 31, 265.

Chaminé, H.I., 2000. Estratigrafia e estrutura da faixa metamórfica de Espinho-Albergaria-a-Velha (Zona de Ossa-Morena): implicações geodinâmicas. Faculdade de Ciências da Universidade do Porto. 497 pp. (unpublished PhD Thesis).

Chaminé, H. I., Gama Pereira, L. C., Fonseca, P. E., Noronha, F., Lemos de Sousa, M. J., 2003a. Tectonoestratigrafia da faixa de cisalhamento de Porto-Albergaria-a-Velha-Coimbra-Tomar. Cadernos Laboratorio Xeolóxico de Laxe, Coruña, 28, 37-78.

Costa, J. C. & Teixeira, C., 1957. Carta Geológica de Portugal na escala de 1:50000, Notícia explicativa da folha 9-C (Porto), Serviços Geológicos de Portugal, Lisboa, 38 p.

Daveau, S., 1977. L'évolution géomorphologique quaternaire au Portugal, Recherches Françaises sur le Quaternaire, Paris, INQUA, 50, 11-21

Ferreira, A. B., 1983. Problemas da evolução geomorfológica quaternária do noroeste de Portugal. Cuadernos do Laboratorio Xeolóxico de Laxe, Coruña, 5, 311-329.

Ferreira, A. B., 1993. As Rañas em Portugal: significado geomorfológico e estratigráfico. O Quaternário em Portugal, balanço e perspectivas, Lisboa, Colibri, 7-15.

Granja, H. M., Soares de Carvalho, G., 1995. Sea-Level Changes During the Pleistocene-Holocene in the NW Coastal Zone of Portugal. Terra Nova, Terra Research, 7, 60-67.

Mörner, N. A., 1973. Eustatic changes in last 300 years. Paleogeograp., Paleoclimat., Paleoecol., 9, 153-181.

Ribeiro, A., 1984. Néotectonique du Portugal, Livro de Homenagem a Orlando Ribeiro, Lisboa, Centro de Estudos Geográficos, 1, 173-182.

Ribeiro, A., 2002. Soft Plate and impact tectonics, Berlin, Springer Verlag, 324 p.

Ribeiro, O., Cotelo Neiva, J. & Teixeira, C., 1943. Depósitos e níveis pliocénicos e quaternários dos arredores do Porto (Nota preliminar). Boletim da Sociedade Geológica de Portugal, 3, 95-101.

Ribeiro, O., Lautensach, H., Daveau, S., 1987. Geografia de Portugal. I. A posição geográfica e o território, Lisboa, 334 p.

Somoza, L., Zazo, C., Bardaji, T., Goy, J. L., Dabrio, C. J., 1987. Recent Quaternary Sea Level Changes and Tectonic Movements in SE Spanish Coast, Trabajos sobre Neógeno-Cuaternario, 10, 49-77.

Teixeira, C., 1962. Carta Geológica de Portugal na escala 1/50000: Notícia explicativa da folha 13-A (Espinho), Serviços Geológicos de Portugal, Lisboa, 28 p.

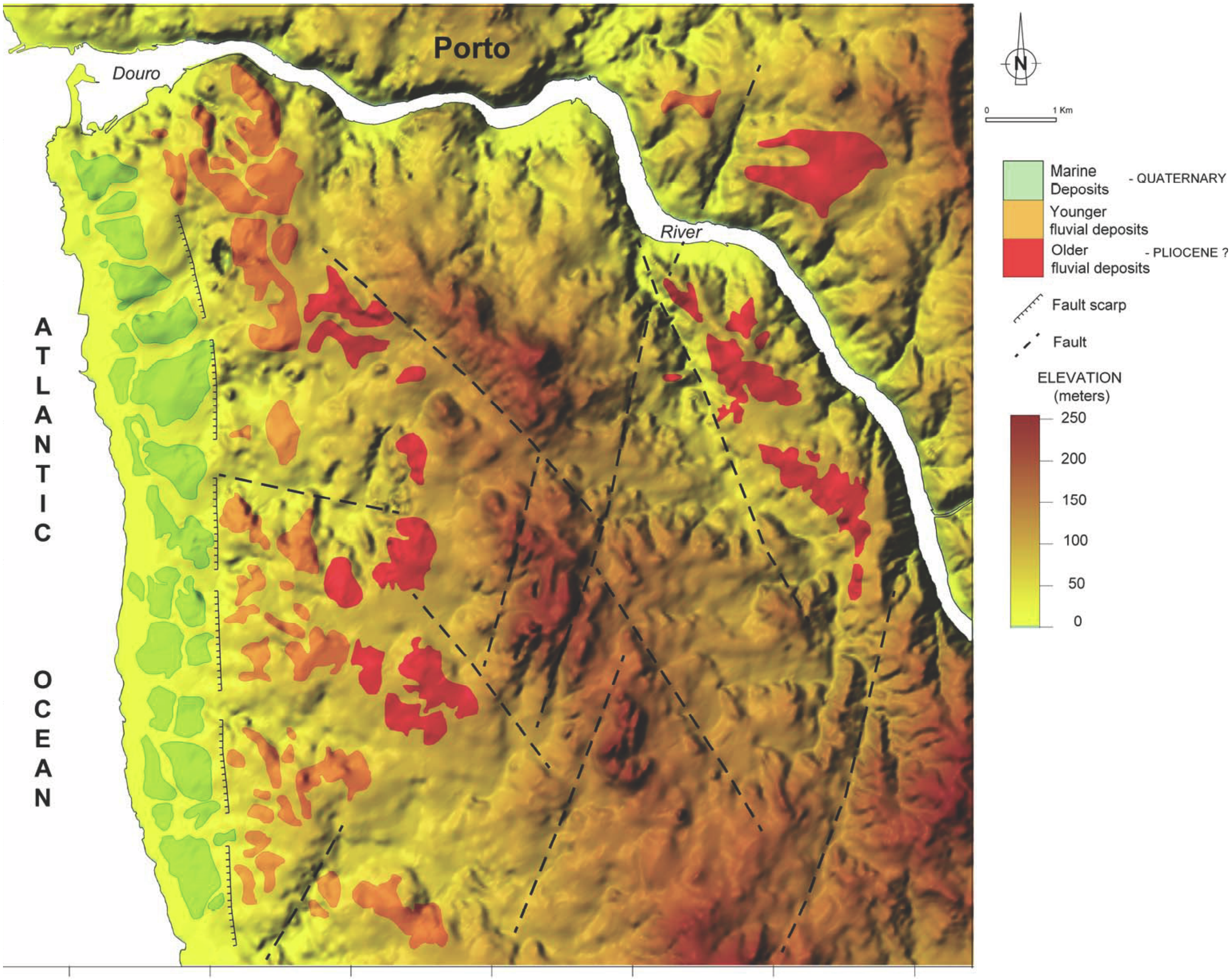


Fig. 1 – Digital Elevation Model (illumination from SW, angle of 45º).

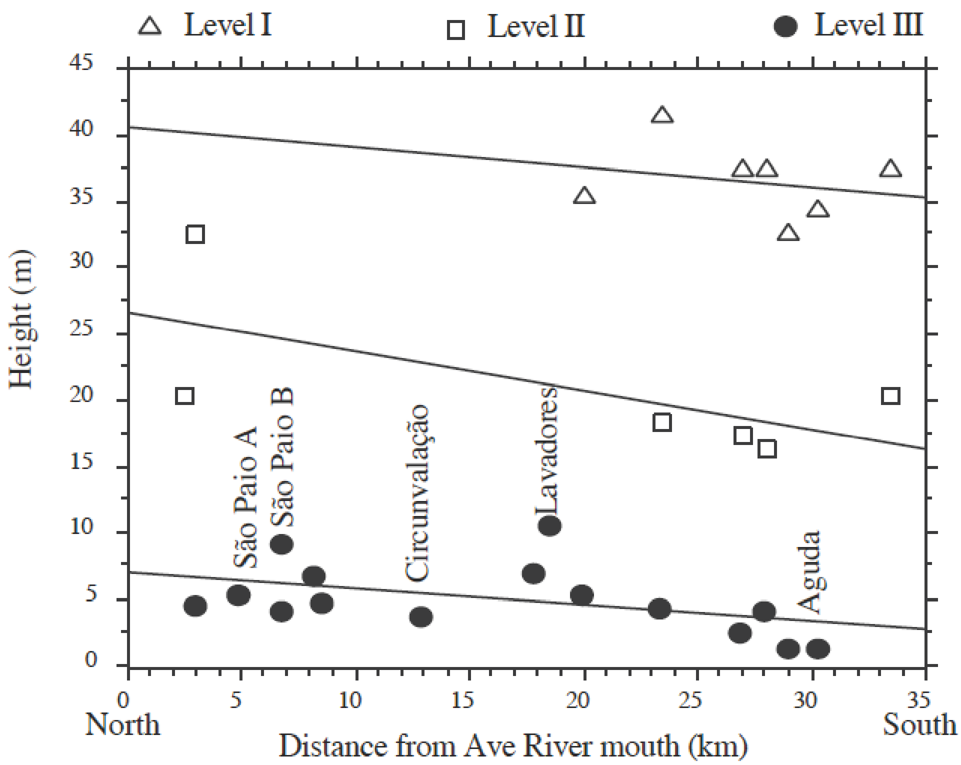
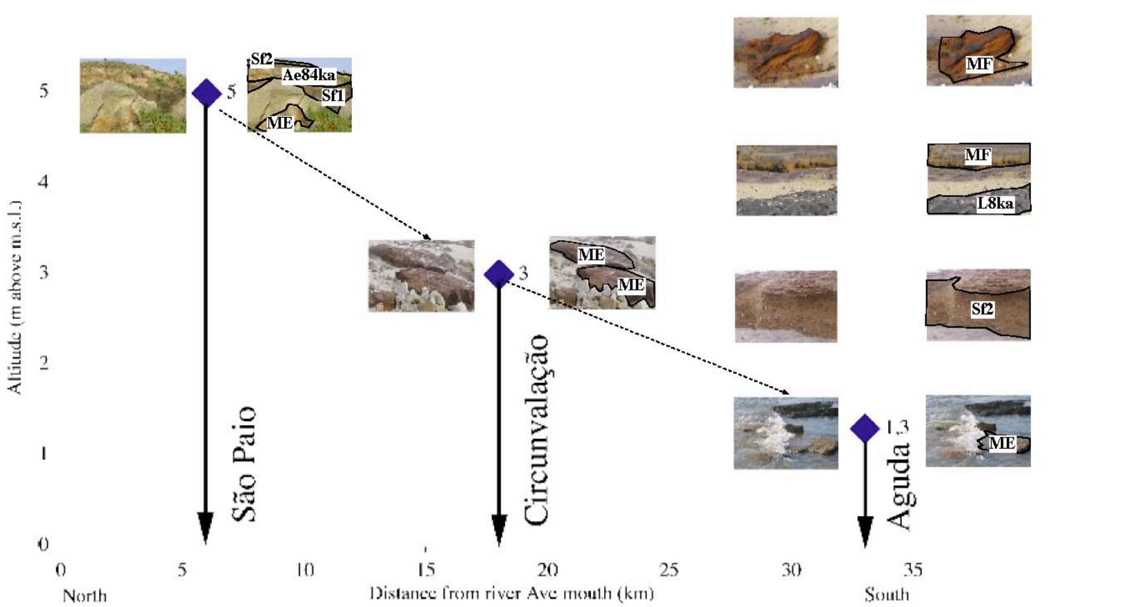


Figure 2 – The height of marine deposits outcrops and its evolution along the studied coastline.



ME - Eemien marine deposit (ca. 125.000 BP); **Sf1** - Solifluidal deposit (<125.000; >84.000 BP); **Ae 84 ka** - Aeolian deposit (ca. 84.000 BP); **Sf2** - Solifluidal deposit <84.000>8.000 BP; **L8ka** - Lagoon deposit (ca. 8.000 BP); **MF**- Flandrian marine deposit <8.000 BP

Deduced balancement of Eemien marine deposits

Fig. 3 - Essay of correlation of the different outcrops referred in the text.

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Relative Sea Level, Diastrophism and Coastal Erosion: the Case of Espinho (Portuguese NW coast)

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Abstract

The plotting of relative sea level for several stations belonging to the Iberian Peninsula shows different trends. These trends seem to be related with the diastrophism affecting more intensely the Northern and Southern façade of the Peninsula. The marine terraces (Pleistocene) at the North of Espinho (NW coast of Portugal) and the würmian/holocene deposits at the South of this city show opposite tectonic trends that seems to define a tectonic depression corresponding approximately to the localisation of Esmoriz lagoon. At Espinho, coastal erosion began in the middle of nineteen century. We think that the contemporary rising of sea level that began after the end of Little Ice Age, together with a possible subsiding trend, may be responsible for the severe erosion endured by this area.

1. INTRODUCTION

The understanding of coastal dynamics lies, ultimately, on the relationship between relative sea level and sediment supply to the coastline.

Relative sea level depends on several kinds of data that can be resumed in the simple diagram of R. Paskoff (1985).

Eustatic variations have a global character. On the contrary, the movements that take place in the continent are spatially localised. Generally, land movements have a slower rate than the eustatic variations. However, they must not be neglected, as eustatic variations can be reduced or amplified by the land movements, which have a much bigger duration in time. So, the net sea level variations must be understood as a resultant of the interference between sea-level changes and land movements. That's why we must always speak about "relative sea-level". Obviously this resultant can be different in adjacent areas if its diastrophism has a different rate or a different sense.

Diastrophic eustatism Glacio-eustatism Geöidal eustatism Sedimentary eustatism Thermo-eustatism Halo-eustatism Hidro-isostasy	OCEANS	CONTINENTS	Glacial isostasy Erosional isostasy Tectonic movments Compactation

RELATIVE VARIATIONS OF SEA LEVEL

Figure 1: Phenomena interfering with long time variation of sea level (R. Paskoff, 1985)

2. SOME DATA AND ITS DISCUSSION

The permanent service for mean sea level (PSMSL) presents the data of more than 1000 sea level stations. The complete data set can be found at:

http://www.pol.ac.uk/psmsl/psmsl_individual_stations.html

Figure 2: Localisation of the stations discussed in the text

This data allows us to plot the recent sea level variations for 31 places within the Iberian Peninsula. The length of the series varies greatly. Only 20 stations have series of, at least, 10 years. So, the study of its variation is not equally reliable for all the 31 stations. That's why we only present the ones that seem more reliable. The chosen stations are localised in figure 2.

Within these 20 stations, we can see different kinds of relative sea-level plots: in most of the cases, the relative sea level is rising. However, sometimes, like in the case of Santander, the 2 stations (Santander I and Santander II, figures 3 and 4) have opposite trends, even if they are geographically quite close (only 2' of distance in longitude).

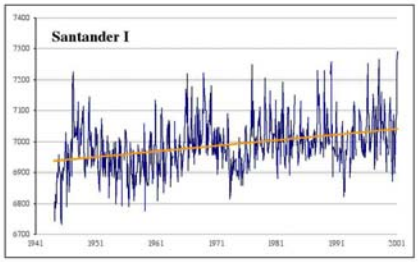


Figure 3: Monthly sea level variations at Santander I (1944-2000)

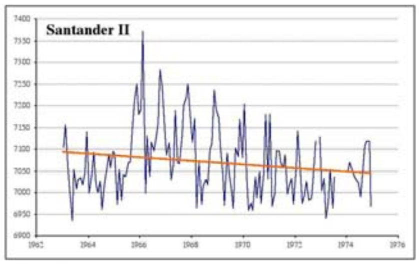


Figure 4: Monthly sea level variations at Santander II (1963-1974)

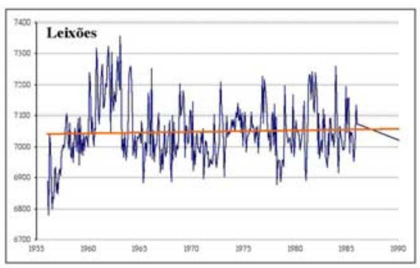


Figure 5: Monthly sea level variations at Leixões (1956-1985)

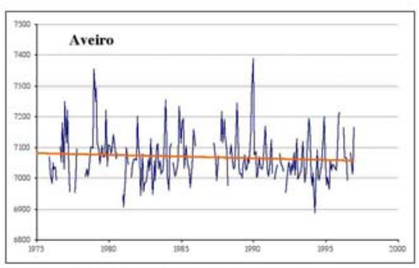


Figure 6: Monthly sea level variations at Aveiro (1975-1996)

To the South of the Peninsula, at Cádiz, we find again 2 stations quite closely situated (Cádiz II and Cádiz III) with opposite trends.

Most of the plots we made for the Portuguese coast show some sea-level rise. However, at Aveiro, there is a slight trend to a sea level drop.

Cascais has the longer series of the Iberian Peninsula (95 years) and one of the longest in Europe. Another long series correspond to Lagos (63 years), at Algarve.

At figure 11 (at the end of the paper) we plotted the trends of all the PSMSL stations in the Iberian Peninsula. Of course, the reliability of sea-level curves depends very much on the length of the series. That's why we also plot the number of years used to calculate the trends.

We can see that most of the stations have a positive trend. This means that at those stations the sea level is going up. However, the amount of sea level rise can be quite different from one place to another. And there are also some stations where the sea level is descending.

At figure 12 we can see the global sea level rise that took place after the end of Little Ice Age. The subsequent climate warming is the main cause of a slight a sea level of about 12 cm in 140 years (according to Mörner, 1973). This should mean a "global" trend of about 1mm/year.

Recent data is slightly higher. According to J. M. A. Dias (1990), the global sea level rise should be around 1,5mm/year.

Of course, the sea level variations that are clearly out of this "medium" range may be assigned to other phenomena (see figure 1).

The figure 11 shows that the trends at the western coast of Iberian Peninsula have a low variation.

However, when we approach the Northern and Southern coast, we have stronger variations and several cases of dropping sea level. We think that the only explication for this phenomena is the tectonic background of this areas, which represent the newer orogenic belts (Pirinéus at the North, Béticas at the SE coast) or the collision front between Eurasia and Africa (Southern coast west of Gibraltar).

3. TECTONIC BACKGROUND OF THE LITTORAL PLATFORM NEAR PORTO

We have made some attempts to analyse the distribution of marine terraces in the area between Vila do Conde and Espinho (fig. 13; Araújo, 1991, 1997, 2000 and 2001). Based on sedimentological criteria, we ranged the marine deposits in 3 different levels (the referred altitudes were find at Lavadores, where our staircase model has been defined):

- Level I (the highest, around 30m and the oldest); level II (between 18-15m); level III (from 10m till 5m; Araújo, 1991).

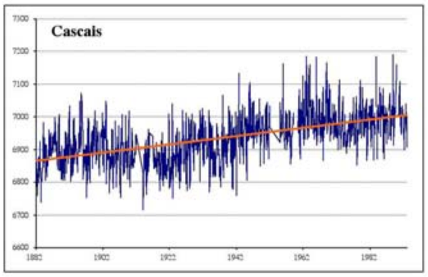


Figure 7: Monthly sea level variation for Cascais (1882-1992)

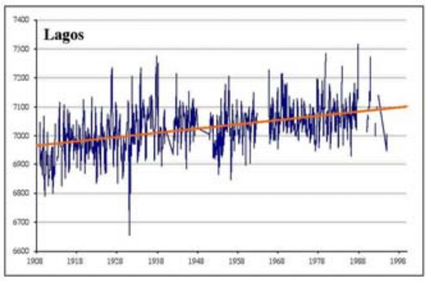


Figure 8: Monthly sea level variation for Lagos (1909-1985)

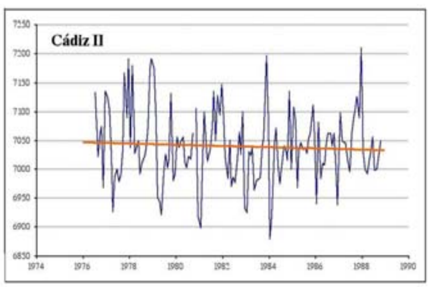


Figure 9: Monthly sea level variation for Cádiz II (1976-1988)

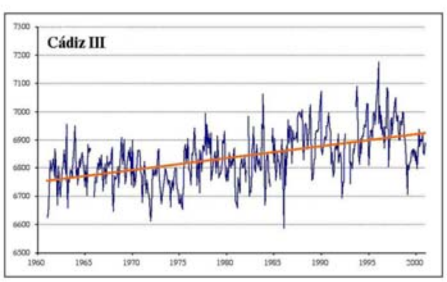


Figure 10: Monthly sea level variation for Cádiz III (1961-2000)

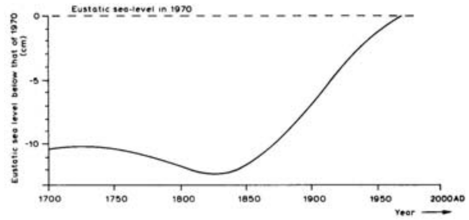


Figure 12: Eustatic sea level changes during the past 250 years (after Mörner, 1973).

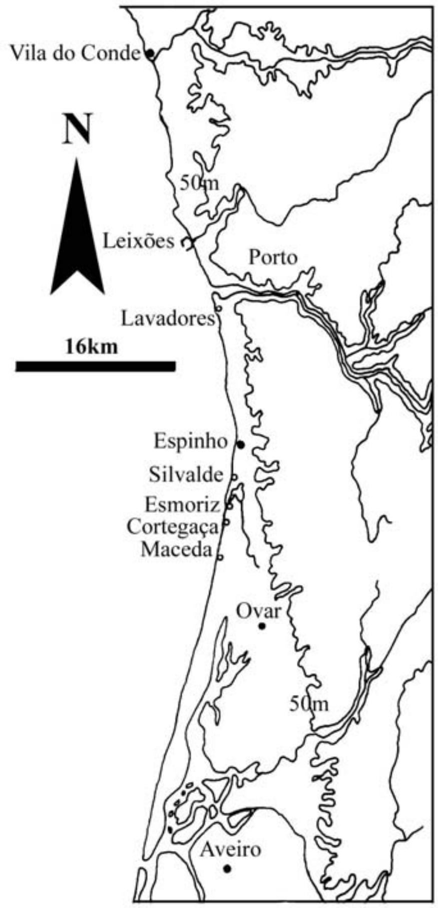


Figure 13: Localisation of studied area

The trendlines for marine deposits (fig. 14) show they have lower altitudes at the South of the studied area (in the direction of Espinho).

These general trends don't mean a regular subsidence: the height variation of marine terraces along the coastline is quite irregular and seems to define a puzzle of small blocks which may undergo different tectonic movements.

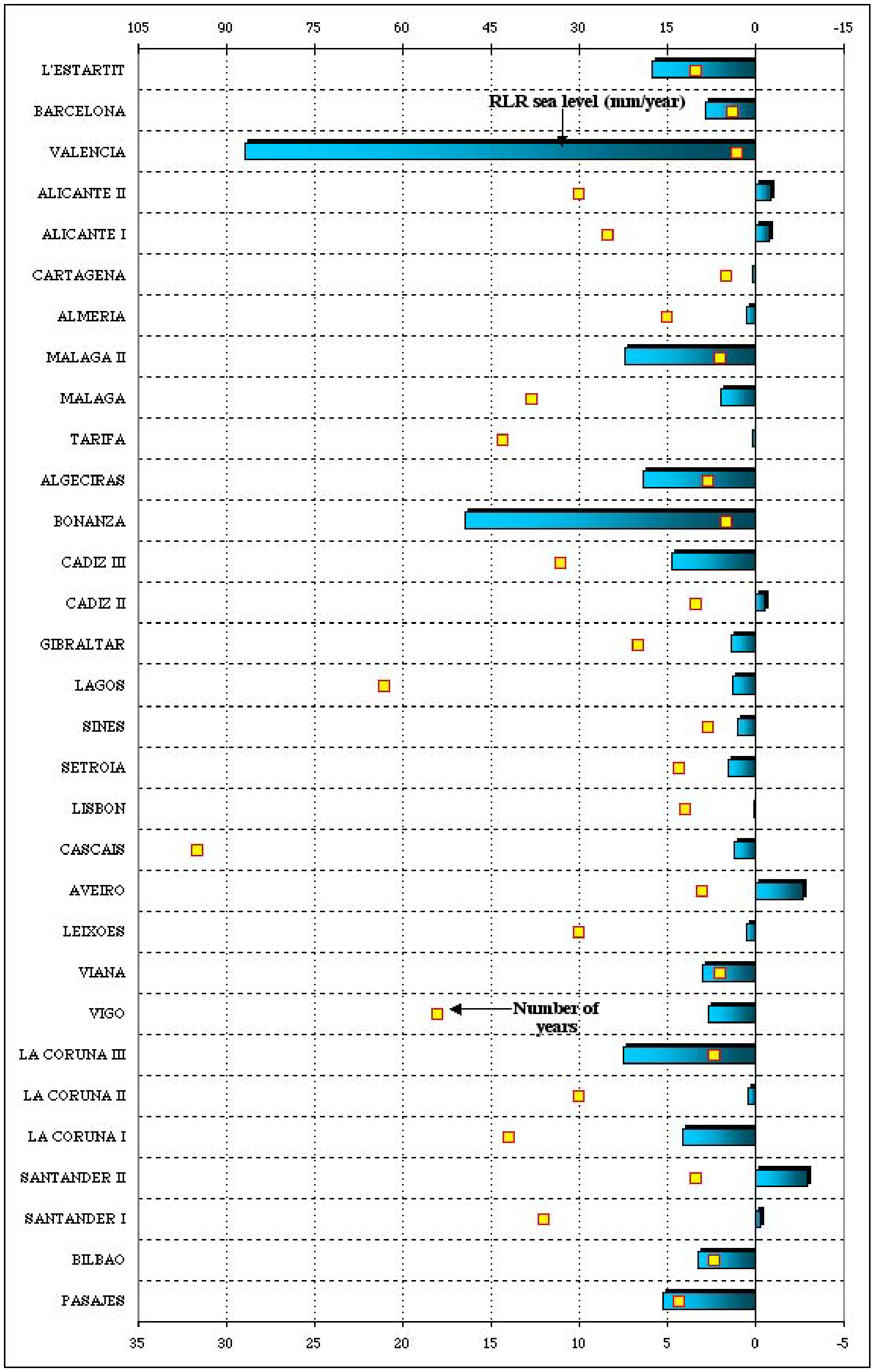


Figure 11: The trends of sea level variation at the Iberian Peninsula (data from Permanent Service for Mean Sea level: PSMSL).

4. POSSIBLE NEOTECTONICS AT THE SOUTH OF ESPINHO.

The area to the South of Espinho is a low area, covered with several dune systems.

The retreat of the coastline has carved cliffs upon the dune systems, allowing us to see the fini-pleistocene and holocene deposits that lay beneath recent sand dunes and beaches.

Since 1986 (Araújo, 1986) these deposits have been monitored and the first ideas we had about them have been reviewed (Araújo, 1991).

After that, several papers have been produced on this subject (Carvalho & Granja, 1995, Granja, & Groot, 1996, Granja *et. al*., 1999).

At Esmoriz, very close to its breakwater and because of the beach erosion due to its construction, we could see (around 1980), an aeolian sandstone with an irregular surface, forming ridges and grooves (fig.15), covered by a black layer of peat. This deposit appeared around mean sea level and, to our knowledge, was never seen again.

This sandstone with the curious ridges and grooves is quite conspicuous in that area and it appears 1,75km to the South, at Cortegaça beach as a part (Bsh, spodic horizon) of a podzolic soil.

The aeolian sandstone is about 1,5m thick. On the top of it we found some pieces of coal, for which we obtained a datation of 5885±75 BP (Niedersächsisches Landesamt für Bodenforschung, Hannover).

Beneath it there is a greenish-gray silty layer for which, at the same Hannover lab, we obtained a datation of 13810±380 BP. These silty layers are recurrent. Recently (May 2002), at Maceda beach, we count 4 of these layers (fig. 16). They seem to be the remain of ponds installed in the troughs between dune ridges, and probably each one represent a moment of a wetter climate during Würm and Tardiglaciar.

At Cortegaça and Maceda beaches the spodic horizon that appeared at mean sea level near Esmoriz lagoon (fig. 15) was considerably higher (respectively 5m and 7m: fig. 17).

Figure 17 represents the difference in altitudes of the brown spodic horizon and the lower greenish layer from Esmoriz beach till S. Pedro de Maceda beach.

On the contrary of the dipping to the South trend we assumed for the marine pleistocene deposits at the North of Espinho, the spodic horizon and the lower greenish layer seem to have a North dipping trend.

So, it seems that the depressed character of the area between Espinho and Esmoriz may have something to do with this tectonic trends.

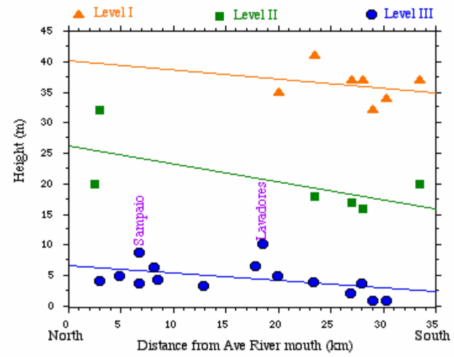


Figure 14: The height of marine deposits outcrops and its evolution along the studied coastline

5. SOME CONCLUSIONS: POSSIBLE CAUSES OF COASTLINE EROSION AT ESPINHO

Espinho has suffered from severe erosion since 1869 (Ferreira Diniz, 1909). The occidental part of the city of Espinho was destroyed around 1909 (fig. 18).

The construction of Leixões harbour began in 1884. The fundamental issues are two large "L" shaped breakwaters. The Northern breakwater is 1.579m long. The Southern breakwater is 1.147m long. They were ready at February 1895 (APDL site: http://www.apdl.pt).

So, at 1889 when a part of the city of Espinho was already destroyed (figure 18) Leixões breakwaters were not ready yet.

A contemporary testimony (Ferreira Diniz, 1909) concludes that the cause for the erosion at Espinho can not be the building of Leixões harbour because the coastal retreat began <u>before</u> that its construction was ready.

The same conclusion can be see at G. Soares de Carvalho (1999), where much older "sea invasions" are referred to have happened at 1834, 1869, 1871 and 1874.

For the 1889 "sea invasion" we don't know if the breakwaters were enough long at that time to produce a serious influence on littoral drift, which is prevalent from the North. But we can imagine that, as Leixões stays at the North of Douro mouth, its breakwaters will stop the sand coming from the Northern (and less important rivers, Mota-Oliveira, 1990) and its influence is not much relevant at Espinho sedimentary budget. The hypothesis (G. Soares de Carvalho (1999) of an inversion of littoral drift at the area of Leixões harbour (from the general direction North to South to South-North) goes in the same direction.

But even if there is some influence of Leixões harbour to the Espinho erosion, the non-coincidence of dates seems quite impressing.

That brings we back to the Mörner curve for sea level evolution. Apparently, the lowest point of sea level was reached at 1830. Than, the sea level began to rise, slowly but continuously.

According to Brunn principle (Paskoff, 1985) for each mm of sea level rise the coastal variation in the horizontal direction will be 100 times bigger.

According to the same principle, the rate assumed in the Mörner curve (around 1mm/year) should produce a coastal retreat of 10cm/year.

Even if it is a slow rate, the cumulative effect from 1830 to 2000 (170 years) could mean 170m of coastal erosion, and this is already meaningful. This issue is surely responsible for some of the coastal retreat that happens at 70% of the coastlines in the world (Bird, 1993).

However, the sea level rise is only one of the causes of beach erosion. According to A. Dias (1990) sea level rise is responsible for only 10% of the beach erosion problem.

Other reasons can be pointed out:

- The sedimentary deficit of the rivers, due to dam construction (the sand supply to the coast by the Douro river is actually only 20% of its “natural” conditions (Mota-Oliveira, 1990);
- The impact of coastal constructions (the breakwaters of Leixões, for instance);
- Another interesting idea pointed out by Paskoff (1985) and G. Soares de Carvalho (1999) is the hypothesis that because of the end of flandriana transgression the sediment supply to the coastline has been slowed down. This, together with the other referred facts, could create a sedimentary deficit and the consequent coastal erosion.

Now we must come back to the tectonic trends we inferred from the distribution of marine terraces and the heights of podzolic soils.

If those trends are confirmed, the area between Espinho and Esmoriz corresponds to a tectonic depression. This could explain why we have a small lagoon in Esmoriz (fig. 13).

This could also explain why the coastal erosion began before the construction of the breakwaters at Leixões: the small rising of sea level that began after the end of Little Ice Age, acting upon soft pleistocene and holocene deposits was increased by a possible subsidence of Espinho-Esmoriz area.

Although no movement is objectively referred at that area in Neotectonic map of Portugal (Ribeiro and Cabral, 1988), there is a “lineament” that can be prolonged from Gerês Mountain till the area of Ovar. Like the other strike-slip faults of NNE direction, this one is probably an active fault, with neotectonic movement.

Another meaningful lineament of NNE-WSW direction crosses the coastline just where its direction changes from NNW to NNE, precisely at the place of Espinho.

Figure 15: The aeolian sandstone covered by peat at Esmoriz

Figure 16: At Maceda beach the Bhs horizon appears at about 7m above msl (mean sea level).

As we have seen before (figs 3-10) the land movements does have some influence on relative sea level variation. May be this is the reason why coastal erosion has been so fast and devastating at Espinho.

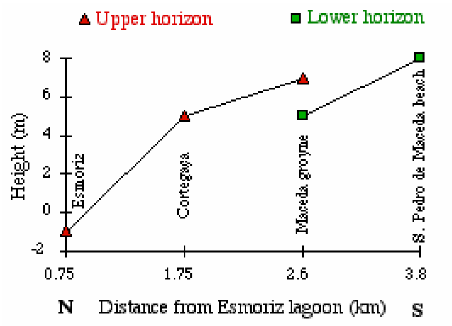
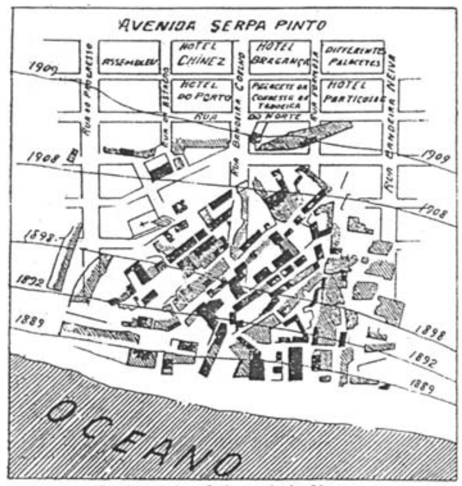
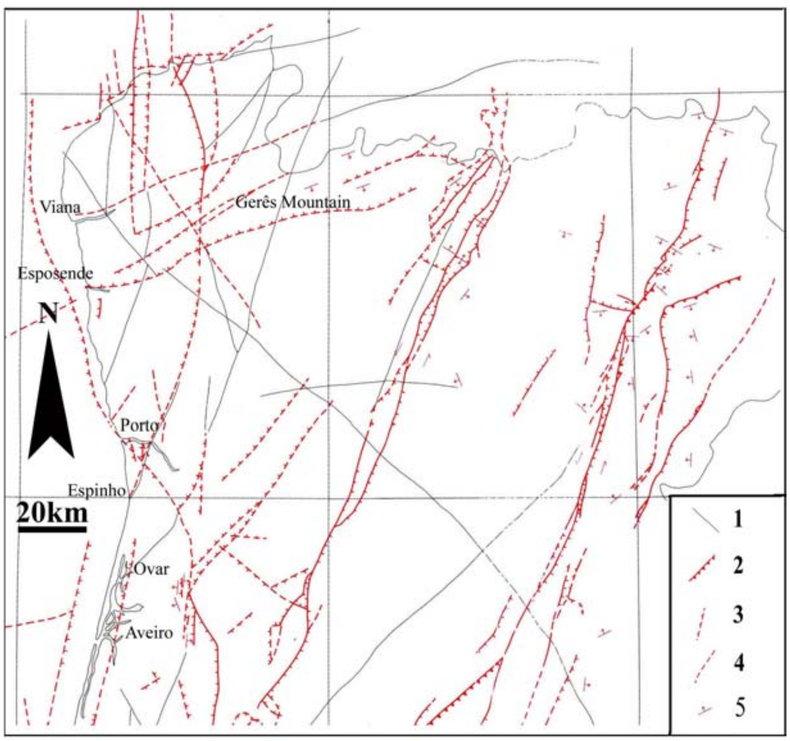


Figure 17: Height variation of podzolic soils and lagunar deposits between Esmoriz and Maceda beaches



Graphico da invasão do mar desde 1889 a 1909

Figure 18: The sea invasion at Espinho from 1889 till 1909 (After Ferreira Diniz, 1909).



1: Lineament
2: Inverse fault
3: Normal fault
4: Fault with unknown movement
5: Tilted surface

Figure 19: Fragment of the Neotectonic map of Portugal (after Cabral & Ribeiro, 1988)

ACKNOWLEDGEMENTS

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REFERENCES

Araújo, M. Assunção (1997) - A plataforma litoral da região do Porto: Dados adquiridos e perplexidades - *Estudos do Quaternário*, nº 1, APEQ, Lisboa, p. 3-12.

Araújo, M. Assunção (2000) - Depósitos de fácies continental e marinho na plataforma litoral da região do Porto: importância da tectónica na sua organização espacial, *Ciências da Terra (UNL), nº 14*, Lisboa, p. 111-122.

Araújo, M. Assunção (1991) - *Evolução geomorfológica da plataforma litoral da região do Porto,* Ph.D thesis, FLUP, Porto, 534 p.

Araújo, M. Assunção (1994) - Geomorphological Evolution of Littoral Platform from Porto Region (North of Portugal), Subcomission on Mediterranean and Black Sea Shorelines, *Newsletter n° 16*, INQUA, pp. 19-22.

Araújo, M. Assunção (2001) - Marine platforms from Porto area (NW portugal): Tectonics versus eustasy - *Actas da V Reunião do Quaternário Ibérico*, GTPEQ and AEQUA, Lisboa, pp. 178-181.

Bird, E. C. F. (1993) - *Submerging Coasts. The Efects of a Rising Sea Level on Coastal Environments*, John Wiley & Sons, Chichester, 184 p.

Cabral, J., Ribeiro, A. (1989) - *Carta Neotectónica de Portugal* (escala 1:1.000.000), Serv. Geol. Portugal, Lisboa.

Carvalho, G. S. (1999) - A responsabilidade das estruturas portuárias na migração das praias para o interior ("erosão costeira"), *1ªs Jornadas de Engenharia Costeira*, International Association of Navigation, Porto, pp. 209-226.

Dias, J. M. A. (1990) - A Evolução Actual do Litoral Português, *Geonovas*, nº 11, Lisboa, pp. 15-28.

Ferreira Diniz (1909) - Praia de Espinho, "*A Construção Moderna*", Ano IX, nº 29, p. 226-228.

Ferreira, O., Dias, J. M. A. (1991) - Evolução recente de alguns troços do litoral entre Espinho e o cabo Mondego, *Actas do 2º Simpósio sobre a Protecção e Revalorização da faixa costeira do Minho ao Liz,* Instituto de Hidráulica e Recursos Hídricos, Porto, pp. 85-95.

Granja, H. M, Soares de Carvalho, G. (1995) - Sea-Level Changes during The Pleistocene-Holocene In The NW Coastal Zone of Portugal. *Terra Research*, Blackwell Science, pp. 60-67.

Granja, H. M, de Groot, T. (1996) – Sea-level rise and tectonism in a holoccne coastal environment at Cortegaça beach (NW Portugal): a case study, *Journal of Coastal Research,* 12, 1, 160-170.

Granja, H. M, Ribeiro, I. C., Soares de Carvalho, G., Matias, M. S. (1999) – Some neotectonic indicators in quaternary formations of the Northwest coastal zone of Portugal, *Phys. Chem. Earth* (A), Vol. 24, nº 4, Elsevier Science, pp. 323-336.

Mota-Oliveira, I. B. (1990) - Erosão costeira no litoral Norte: considerações sobre a sua génese e controlo, *Actas do 1º Simpósio sobre a protecção e revalorização da faixa costeira do Minho ao Liz,* Inst. Hidráulica e Recursos Hídricos, Porto, p. 201-221

Nacional Research Council (1990) - *Sea-Level Changes*, Studies In Geophisics, Nacional Academy Press, Washington D. C., 234 p.

Nummedal, D., Pilkey, O. H., Howard, J. D. (1988) - *Sea-Level Fluctuation And Coastal Evolution*, Based on a Symposium In Honor of William Armstrong Price, Society Of Economic Paleontologists And Mineralogists, Special Publication n° 41, Oklahoma, 267 p.

Paskoff, R. (1985) - *Les littoraux - impact des aménagements sur leur évolution,* Col. Géographie, Masson, Paris, 185 p.

Pugh, D. T. (1987) - *Tides, surges and mean sea level*, John Wiley and Sons, Chichester, 472 p.

Somoza, L., Zazo, C., Bardaji, T., Goy, J. L., Dabrio, C. J. (1987) - Recent Quaternary Sea Level Changes and Tectonic Movements in SE Spanish Coast*, Trabajos Sobre Neógeno-Cuaternario*, pp. 49-77.

Warrick, R. A., Barrow, E. M. & Wigley, T. M. L. (l993) - *Climate and sea level change: observations projections and implications*, Press Syndicate of the University of Cambridge, Cambridge University Press, 424 p.

Tsunamis gerados por deslizamentos insulares — o exemplo de La Palma (Tsunami generated by island landslides — the example of La Palma)

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A teoria de que enormes tsunamis podem ser gerados por deslizamentos em ilhas oceânicas resultou de estudos dos fundos marinhos no Hawaii. Foi Moore (1964) quem primeiro identificou restos de colapsos dos flancos destas ilhas do Pacífico. No registo geológico recente, o estudo destes depósitos permitiu avaliar da sua abundância, magnitude (até 5000 km^3 de material), extensão (até 300 km comprimento) e velocidade (deslocação na água atingindo 140 m/s). O registo histórico descreve tsunamis destruidores gerados a partir de distantes pequenos colapsos de estrato-vulcões insulares (Johnson, 1987; Satake & Kato, 2001). A erupção explosiva do vulcão de St. Helens (Pacífico), ocorrida a 18 de Maio de 1980, permitiu a observação da geração de um deslizamento (3 km^3).

No seguimento do tsunami gerado pelo sismo de 26 de Dezembro de 2004 que devastou o sudeste asiático, cientistas britânicos voltaram a alertar para o perigo de um fenómeno semelhante vir a ocorrer Atlântico. Neste oceano pode ser importante o risco de tsunami por colapso, devido ao grande número de ilhas activas e às recentes propostas (Day *et al.*, 1999a, b) de que alguns destes vulcões (ex. La Palma e Ferro, nas Canárias; Fogo, no arquipélago de Cabo Verde) evidenciam instabilidade.

Nas Canárias, os depósitos de deslizamento documentam que nos vulcões ocorreram cerca de uma dúzia de grandes colapsos nos últimos milhões de anos (Fig. 1). La Palma é uma ilha vulcânica recente, ainda activa. Para sul e a partir dos cerca de 2000 m de altitude, a cumeada montanhosa de Cumbre Nueva torna-se numa cadeia de vulcões alinhada N-S (Cumbre Vieja) (Fig. 2). Os depósitos sedimentares no fundo do mar evidenciam vários colapsos laterais dos flancos da ilha de La Palma. Do mais recente (cerca de 566 mil anos) ainda se observam restos no Cumbre Nueva (Carracedo, 1994; Day *et al.*, 1999a). A possibilidade de um futuro colapso do Cumbre Vieja é favorecida por nos últimos 125 mil anos o Cumbre Vieja (2426 m) ser o vulcão mais activo nas Canárias (Carracedo *et al.*, 1999) e por a parte subaérea do Cumbre Vieja, que forma o terço sul da ilha, atingir 2 km acima do mar com declives médios de 15° a 20°.

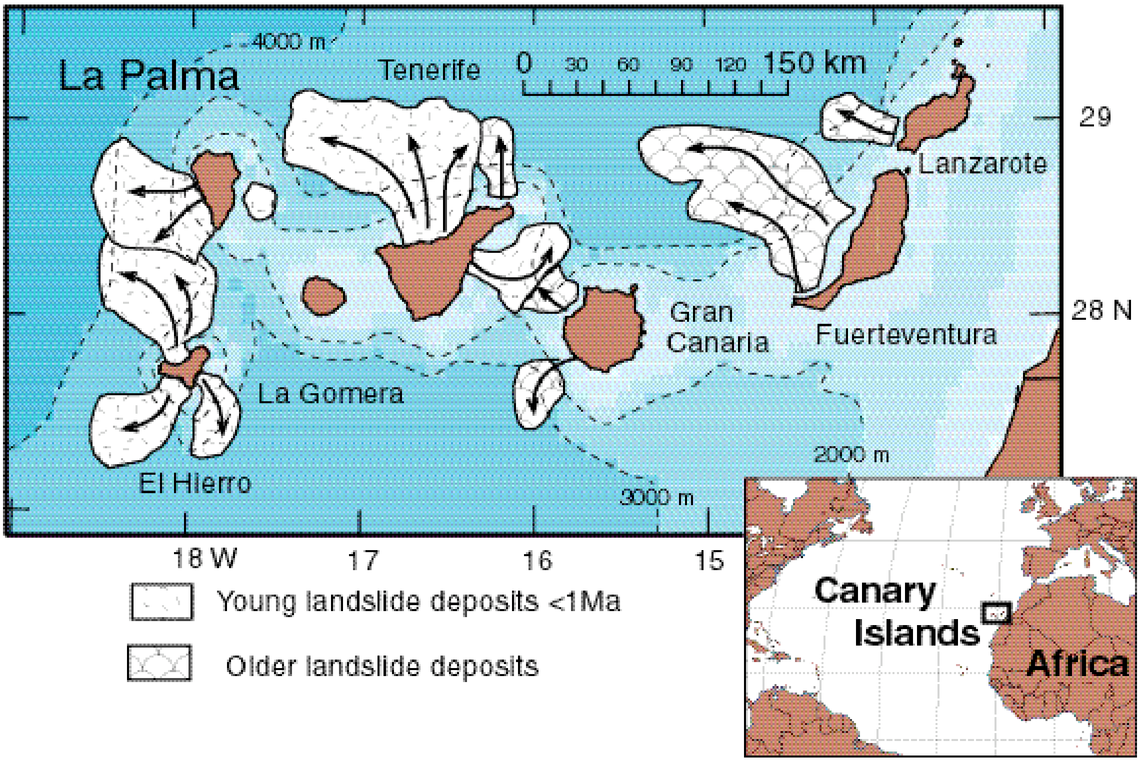

Fig. 1 – Localização da ilha de La Palma, onde se situa o vulcão de Cumbre Vieja (Ward & Day, 2001). Os depósitos submarinos evidenciam que os vulcões tiveram uma dúzia de grandes colapsos nos últimos milhões de anos.

Day *et al.* (1999a) concluiram que nesta montanha, nos últimos milhares de anos, a distribuição e orientação da rede de filões mudou de uma configuração em triplo rifte para uma consistindo de um rifte N-S com uma rede de fendas estendendo-se para oeste. Esta alteração foi por eles interpretada como resultando do desenvolvimento de um descolamento no flanco oeste do vulcão. Na mais recente erupção do Cumbre Vieja, em 1949 (Bonelli Rubio, 1950), gerou-se uma falha normal na crista do vulcão, estendendo-se 4 km, com um rejogo vertical de 4 m, que poderá ser a expressão superficial do descolamento. O exame pormenorizado desta ruptura no período 1994-98 indicou que esta falha tem estado inactiva (Moss *et al.*, 1999). Contudo, esta inactividade não é de estranhar dado que a instabilidade da vertente pode só ser desencadeada pela implantação distensiva de diques ou pela pressurização de água subterrânea confinada (Elsworth & Voight, 1995) numa futura fase eruptiva.

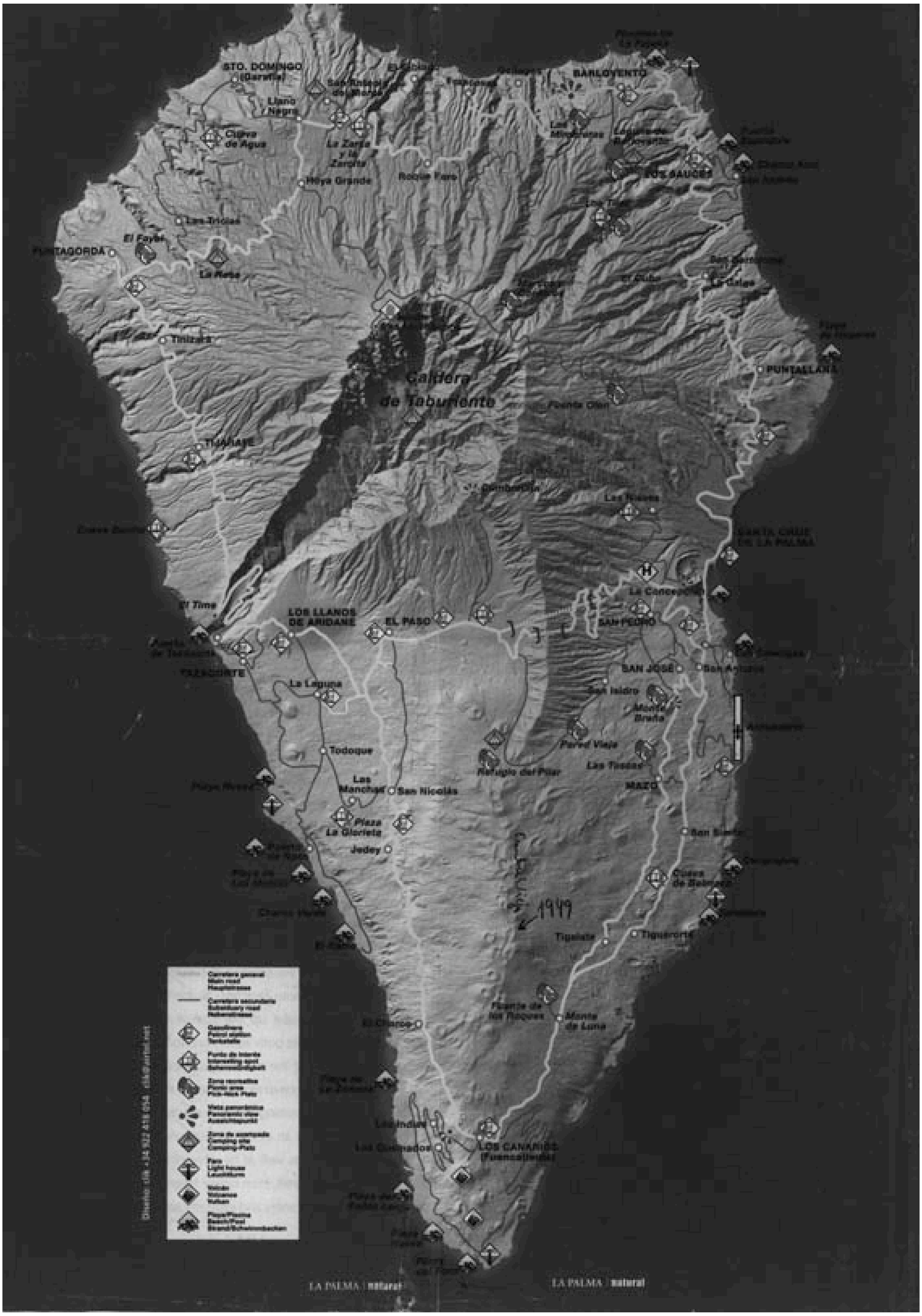

Fig. 2 – MDT de La Palma. O centro da ilha é formado pela Caldeira de Taburiente, com 8 km de largura e 1,5 km de profundidade, ligada ao mar pelo Barranco de Las Angustias. Para sul destaca-se a imponente paleo-escarpa do deslizamento de Cumbre Nueva e a recente cadeia de vulcões de Cumbre Vieja.

Day *et al*. (1999a) estimaram que o bloco instável acima do provável descolamento apresenta 15-20 km de largura, 15-25 km de comprimento e espessura média de 1-2 km, correspondendo a um volume de 150-500 km^3. Num tanque de ondas foi construído um modelo físico da ilha de La Palma e numa câmara de alta velocidade foi registado o colapso e sucessiva onda que à escala real atingiria 650 m de altura. Além disso, Ward & Day (2001) modelaram em computador o pior "cenário" possível: rápido deslizamento (100m/s) em 60 km de um grande bloco rochoso com 500 m^3 (ou 150 km^3), que permanece intacto até atingir a água. O modelo previu que as ondas geradas atinjam as costas atlânticas com 5-25 m (ou 3-8 m) de altura.

Wynn & Masson (2003) estudaram os mais recentes deslizamentos nas Canárias: o El Golfo (cerca de 15 mil anos ou >120 mil anos ?, na ilha El Hierro) e o Icod (cerca de 170 mil anos, em Tenerife). A cartografia submarina indica que ambos os deslizamentos já estavam bastante desagregados no momento em que se depositaram e o estudo dos depósitos turbidíticos com eles relacionados sugeriu que os deslizamentos apresentaram múltiplas fases, o que reduz significativamente o potencial risco de tsunami.

Uma monitorização do vulcão de Cumbre Vieja e da recente falha associada pode providenciar um aviso antecedendo o desastre em algumas semanas. Contudo, muitas erupções podem ocorrer até que se dê o suposto colapso lateral da ilha. Não obstante, os governos devem ter em conta o risco de tsunami e desenvolver estratégias de protecção e aviso das populações costeiras, pois são vários os processos que podem gerar importantes tsunamis.

Portugal está próximo de um limite de placas que apresenta significativa sismicidade — a Zona de factura Açores-Gibraltar. Nesta zona ocorreu a 1 de Novembro de 1755 um sismo com magnitude provável de 8,7-9,0 que gerou um tsunami que atingiu o litoral português com altura talvez atingindo cerca de 15 m (Algarve) a 6 m (Lisboa), também com efeitos muito destruidores no SW de Espanha e W de Marrocos. A ocupação do litoral nas últimas décadas aumentou o risco face a tsunami, principalmente nos troços de litoral arenoso baixo como os do leste Algarvio, Figueira da Foz-Aveiro e nas áreas marginais aos estuários. Por isso, a ocupação da zona costeira deve ser adequadamente ordenada e planeada (zonamento de usos) tendo em conta os vários tipos de variações do nível do mar. Algum investimento na monitorização, em sistemas de aviso e na implementação de procedimentos de evacuação/socorro também seria útil.

Referências bibliográficas:

Bonelli Rubio, J.M. (1950) Contribution al estudio de la erupcion del Nambroque o San Juan. Madrid, Inst. Geog. y Cadastral, 25 p.

Carracedo, J.C. (1994) The Canary Islands: an example of structural control on the grow of large oceanic-island volcanoes. *J. Volcan. Geotherm. Res*. 60, 225-241.

Carracedo, J.C; Day, S.J.; Guillou, H. & Gravestock, P. (1999) The later stages of the volcanic and structural evolution of La Palma, Canary Islands. *Geol. Soc. Am. Bull*. 111, 755-768.

Day, S.J.; Carracedo, J.C.; Guillou, H. & Gravestock, P. (1999a) Recent structural evolution of the Cumbre Vieja volcano, La Palma, Canary Islands. *J. Volcan. Geotherm. Res*. 94, 135-167.

Day, S.J.; Heleno da Silva, S.I.N. & Fonseca, J.F.B.D. (1999b) A past giant lateral collapse and present day flank instability of Fogo, Cape Verde Islands. *J. Volcan. Geotherm. Res*. 94, 191-218.

Elsworth, D. & Voight, B. (1995) Dike intrusion as a trigger for large earthquakes and the failure of volcano flanks. *J. Geophys. Res*. 100, 6005-6024.

Moore, J. (1964) Giant submarine landslides on the Hawaiian ridge. *U.S. Geol. Survey Prof. Paper* 501-D, 95-98.

Moss,J.L.; McGuirre,W.J. & Page, D. (1999) Ground deformation monitoring of a potential landslide at La Palma, Canary Islands. *J. Volcan. Geotherm. Res*. 94, 251-265.

Johnson, W. (1987) Large-scale volcanic cone collapse: the 1888 slope failure of Ritter volcano. *Bull. Volcan*. 49, 669-679.

Satake, K. & Kato, Y. (2001) The 1741 Oshima-Oshima eruption: extent and volume of submarine debris avalance. *Geophy. Res. Lett*. 28, 427-430.

Ward, S.N. & Day, S. (2001) Cumbre Vieja volcano - Potential collapse and tsunami at, La Palma, Canary Islands. *American Geophysical Union*, 1-4.

Wynn, R.B. & Masson, D.G. (2003) Canary Islands landslides and tsunami generation: myth or legend? *Southampton Oceanography Centre*.

Sea level changes and neotectonics: some examples in Portugal (Arrábida and Southwest).

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Introdution

In Portugal mainland, the landscape is characterized by a heritage landform – the so call coastal platform. This platform is related to relative sea level changes and their influence on coastal landforms and deposits (fig.1). The continuous flattened landform is slightly sloping towards the sea and can be found at different heights. Near Aveiro the coastal platform is almost at sea level while in other places of the western front can reach 150m of altitude, showing several steps,. Landward limit isn't always clear, however the regular presence of tectonic scarps makes the transition to continental relief's often abrupt.

Coastal platform can be erosion or accumulation dominant (fig. 1)[1]. In the first case, the razing took place independently of the local or regional lithostructural setting.

The continental shelf morphology is similar to the coastal platform (fig.1) and

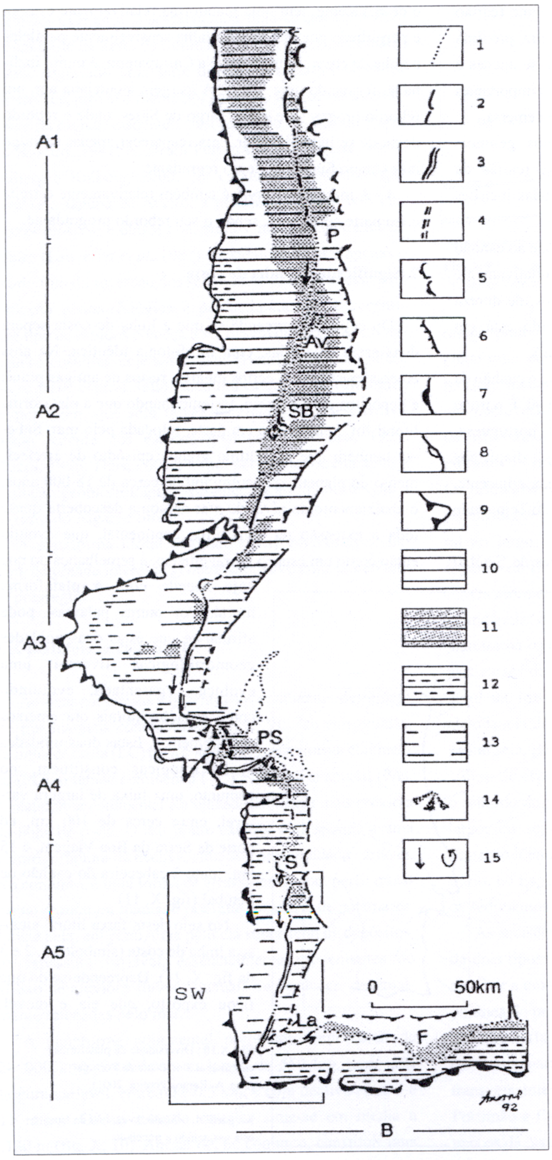

Fig. 1 – Coastal platform and continental shelf geomorphology. 1 – Sandy littoral; 2 – Cliff < 50m; 3 – Cliff > 50m; 4 – Paleocliff; 5 – Erosion edge; 6 – Tectonic edge; 7 – Progradation edge; 8 – Aggradation edge; 9 – Regradation edge; 10 – Erosion dominant coastal platform and continental shelf; 11 – 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).

[1] The expression coastal or littoral platform is use in a morphologic/physiographic sense and never in a genetic one.

its western boundary is a structural one – the continental slope -in the transition from the continental to the marine lithosphere.

1. The sea level change indicators and the main problems to assess them

The analysis of the nearly 940km of Portuguese West and South coastline show different sedimentary balance. Associated with Douro, Tejo and Guadiana Rivers, regions A1, A2, A4 and the Eastern sector of region B (fig. 1) have a positive sedimentary balance (Pereira, 1992). On the contrary, regions A3 and A5 have a negative one. These regions are in the south side of two important morpho-tectonic structures, the Nazaré and Lisboa Canyons, which stop coastal drift sediment bypassing.

The present-day different coastal systems are related to the sediment supply by the main rivers and the longshore drift and its continuity (as well as human management of the coast). Its interruption is related to two major faults that already exist along all the Quaternary. It means that, in the past the coast was as diversify as today with sandy plains with beach-dune systems, barrier systems, cliffs with or without rock platforms at its bottom, estuaries, marine terraces (elements of the coastal platform).

Landforms and sediments are the main indicators of sea level change.

However: (i) correlative deposits locally change in facies as happens today (from beach to dune sediments, for example); (ii) the genetic sediments environment of the Quaternary deposits has change and they have been remixed and eroded by Pliocene and Quaternary shoreline fluctuations; (iii) the neoctectonic activity have been recognized in several places along the Portuguese coastal platform, destroying or tilting the main features and faulting sediments, so **genesis generalizations are not to be made and sometimes chronological correlations are difficult to establis**h.

In what concerns the erosional landforms (ii) the cliffs with notches and the rock platforms have been submitted to Quaternary evolution since they were generated, namely by mass movements, and **their features are difficult to recognize**. Another difficulty must be pointed out: most of the scarps between the coastal platform and the inland are considered to be scarp faults with Quaternary activity.

The chronological correlation between several deposits and landforms present along the coastal platform is not easy. The absence of **recognized fauna or flora in the sediments of Late Quaternary** are also a major problem**.** The lack of this kind of data can be the result of the incipient research carried out about this subject.

2. The balance between tectonic and sea level changes: two examples

2.1.The Arrábida coast

The Arrábida ridge, South of Lisbon, is a small limestone mountain uplifted, faulted and tilted during Quaternary. The south front is a complex structural feature. The major coastal landforms are: a cliff that can reach more than one hundred meters, in some places with three steps related to sea level fluctuations, one of them submerged.

The higher level at 40m has no marine sediments and it is covered by colluvial material. Below it, cliffs with a notch and a rock platform have been recognized. It represents the most continuous level and it is covered in some places by a discontinuous coarse sand beach deposit, sometimes with boulders (Pereira & Regnauld, 1994). This level show the same trend as the mountain tectonic deformation, i. e., faulted and tilted to the West, from 10m to 7m above m.s.l. This episode is understood to be Eemian but no fauna were recorded. The beach deposit is covered by an aeolianite. The cement and the shells have been dated by ^{14}C and give an age of 36 786 cal BP (32 040 {+ 1410 / - 1190 BP}, (36 786 BP, is the position of the interception point of the calibration curve; CalPal, 2004; Pereira & Angelucci, 2004). This date is related to a low sea level (about 30 to 45meters below m.s.l.) but when the coastline was not far from the present one, about 4km south, when Mousterian inhabited an eastern grotto in correlation with this level. The continental fauna recorded in the archaeological horizon show a fresh environment at about 30 930 ± 700 BP (ICEN-387; Antunes, 1991).

A lower level were recognize at –7m (Equipe ERLIDES, 1992), represented by a narrow discontinuous step, without dating elements. The authors proposed then a probable age of 5000 years taking into account what was known in the continental shelf. This interpretation agrees with the new data and conclusions about rising sea level in the Sado estuary (Psuty e Moreira, 2000).

2.2. The Southwest coast.

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. 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).

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 leveled but faulted Paleozoic bedrock (Cambric schist and greywakes), where different types of deposits are still preserved (fig. 1).

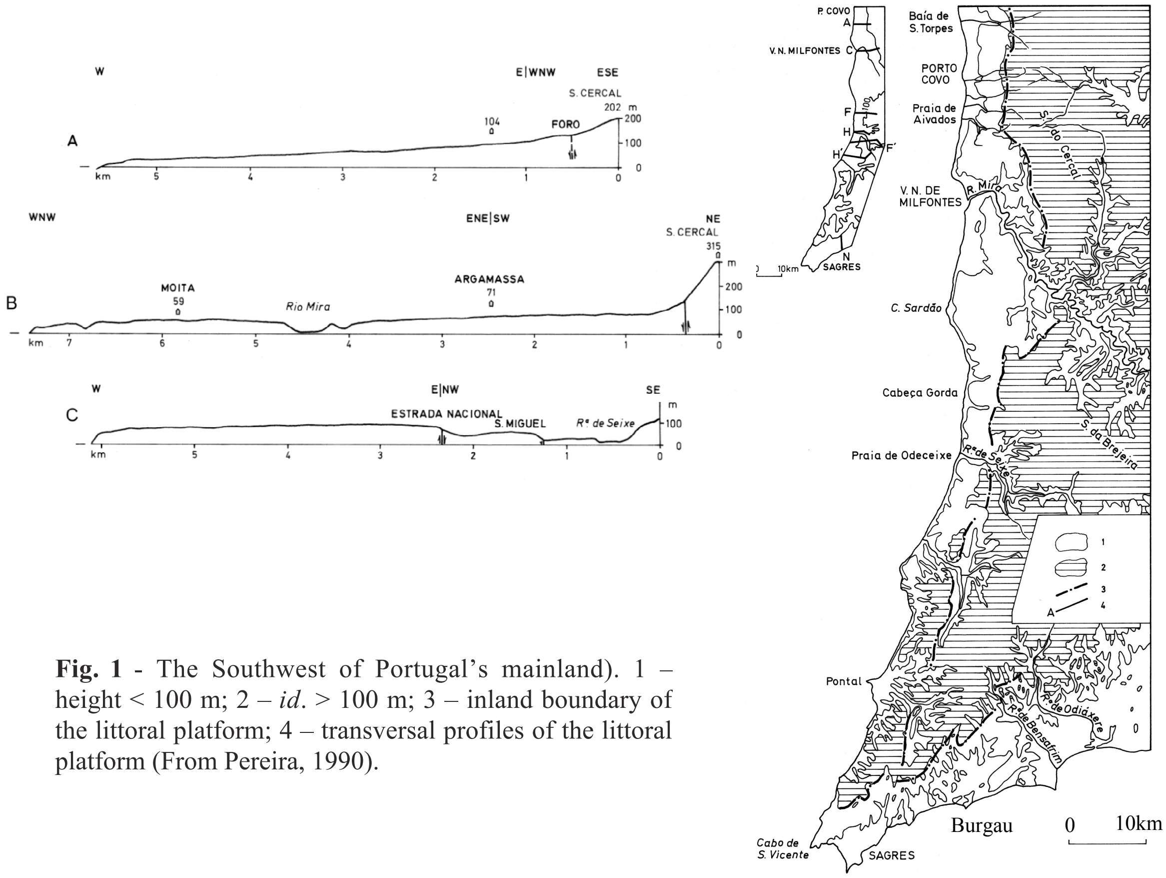

Fig. 1 - The Southwest of Portugal's mainland). 1 – height < 100 m; 2 – *id.* > 100 m; 3 – inland boundary of the littoral platform; 4 – transversal profiles of the littoral platform (From Pereira, 1990).

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. These beach deposits changes to a fluvial one and then to a red aeolinite inland at the bottom of Serra do Cercal (fig. 2, 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. 3-1).

(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 mudflows 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. 2, 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. 3-2). During this episode, the so-called littoral platform was created.

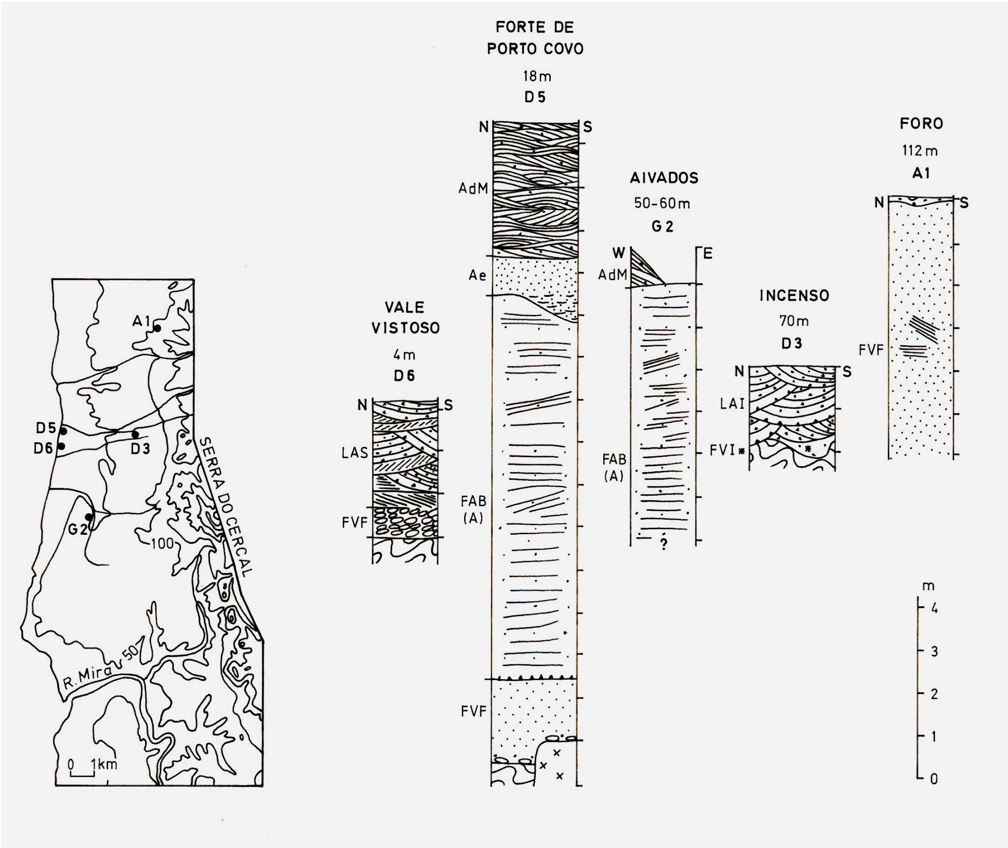

Fig. 2 – Logs of the deposits preserved on the littoral platform north of Vila Nova de Milfontes (Pereira, 1990).

(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. 2, log G2 and D5) penetrates 11km inland from the present day coastline in a subsided area, where the sea creates a small bay (fig. 4-3). 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.

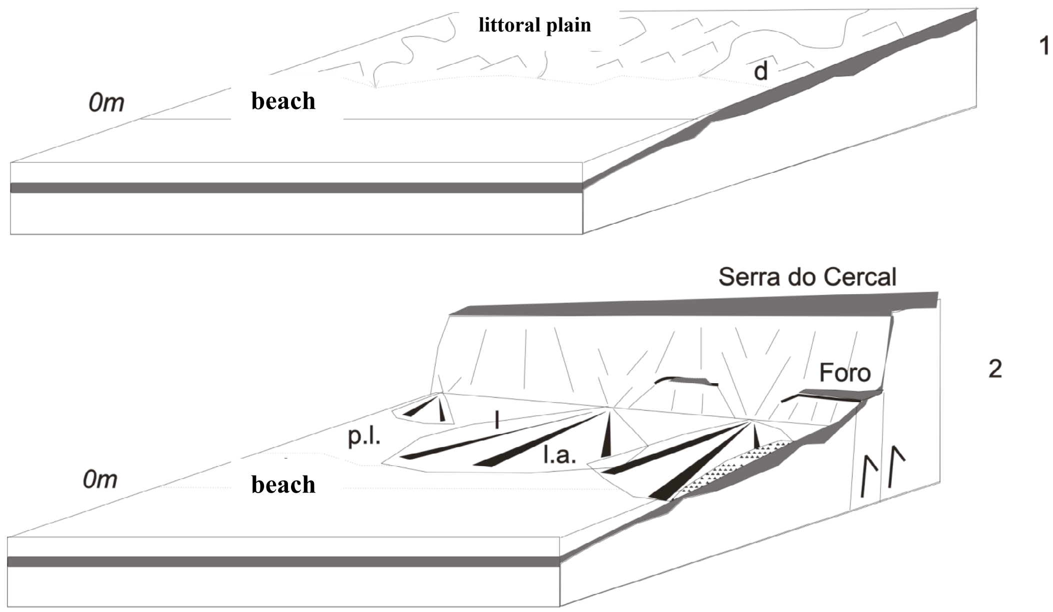

Fig. 3 – The evolution of the littoral platform dominated by Serra do Cercal during the Pliocene (1) and the Plio-Quaternary transition (2) (Pereira, 1999). 1 - the littoral plain bathed by the sea, with beach and dunes (d); 2 – littoral gravel plain (p.l.), with alluvial fans (l.a.), bathed by the sea and dominated by Serra do Cercal (legend in fig. 11.5).

(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..4-4) - 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. 2, 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).

(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.5). The dunes were built by NW till SW winds.

(vi) The latter evolution of the platform is related to the establishment of the drainage network and the transgressive dune field (figs. 5).

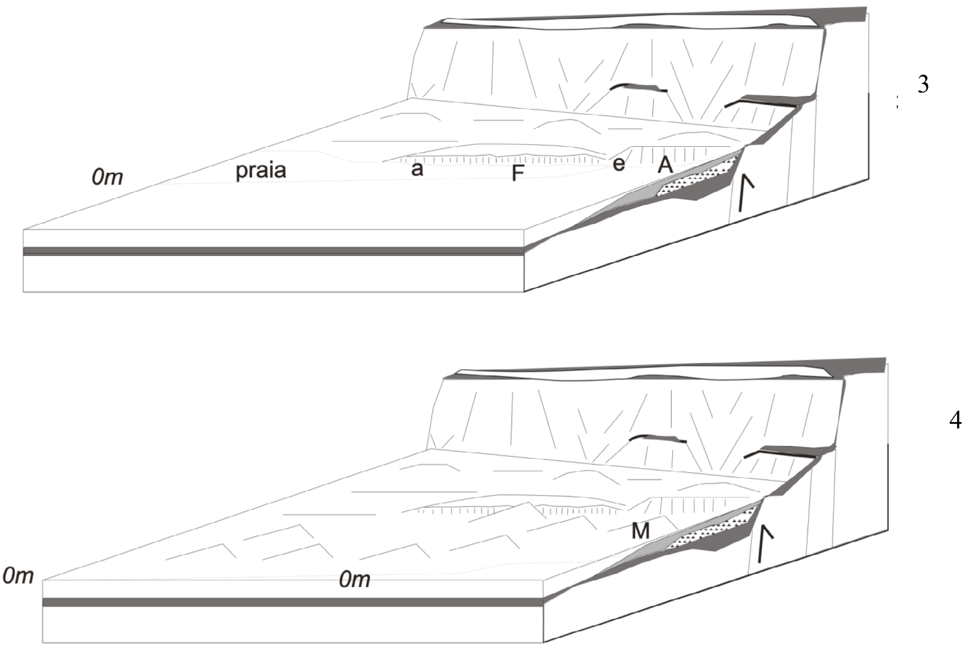

Fig. 4 - The evolution of the littoral platform dominated by Serra do Cercal during the Lower and Middle Plistocene (?). 3 - e – scarp fault, a – cliff cut into the alluvial fans, A – Aivados, F – Fort). 4 - littoral platform covered by a large dune field (M – Malhão), after a sea retreat. (Pereira, 2000). (Legend in fig. 1.5).

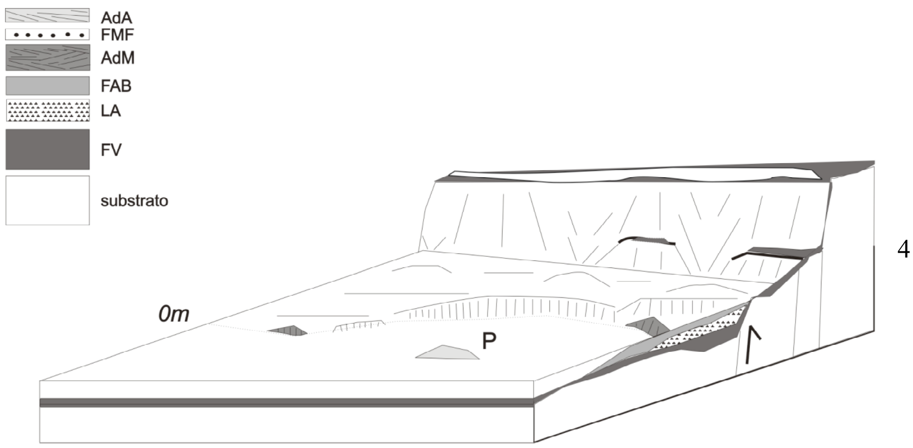

Fig. 11.5 - Present day landscape (P – Pessegueiro Island). Deposits from the oldest to the youngest: FV – Red Formation (FVF); LA – alluvial fan; FAB – Aivados-Bugalheira Formation; AdM – Malhão aeolianite; AdA– Aivados aeolinite (Pereira, 1999).

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 & 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[2] 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

References

Antunes, M. Telles1991 – O Homem da gruta da Figueira Brava (ca 30 000BP), contexto ecológico, alimentação, canibalismo. *Memórias da Academia das Ciências*, Lisboa, Classe de Ciências, XXXI, p.487-538.

Equipe ERLIDES (1992) – Découverte d'un niveau marin submergé le long de la chaîne de l'Arrábida, Portugal. Finisterra, XXVII (53-54), p.183-186.

Pereira, A. Ramos (1990). *A Plataforma Litoral do Alentejo e Algarve o Ocidental. Estudo de geomorfologia*. PhD thesis, Faculdade de Letras, Universidade de Lisboa.

PEREIRA, A. Ramos (2000). A Geomorfologia e o ordenamento da costa alentejana. in G. Soares de Carvalho; F. Veloso Gomes; F. Taveira Pinto (ed.)– *A Zona Costeira do Alentejo*. Associação Eurocoast – Portugal, Porto, p. 9-26.

PEREIRA, A. Ramos (1992) – L'Homme et l'érosion: l'exemple du littoral portugais. *Finisterra*, XXVII (53-54), Lisboa, p.205-225.

Pereira, A. Ramos & Regnauld, H. (1994) – Litorais quaternários (emersos e submersos) na extremidade sudoeste da Arrábida (Portugal). *In* A. Ramos Pereira *et al – Contribuições para a geomorfologia e dinâmicas litorais em Portugal.* Centro de Estudos Geográficos, Linha de Acção de Geografia Física, 35, p. 55-73.

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

PEREIRA, A. Ramos (2004) – A faixa litoral. In Feio, M. & Daveau, S. – *O relevo de Portugal. Grandes unidades*. Publicações da Associação Portuguesa de Geomorfólogos, vol.II. Coinbra, p. 133-145.

Pereira, A. Ramos e Angelucci, D. (2004). Formações dunares no litoral português, do final do Plistocénico e inícios do Holocénico, como indicadores paleoclimáticos e paleogeográficos in A. A. Tavares, M.J. Tavares & J. L. Cardoso (ed.) – *Evolução Geohistórica da Litoral Português e Fenómenos Correlativos*. Universidade Aberta, p.221-256.

Psuty, N.P. & Moreira, M.E. (2000) – Holocene sedimentation and sea level rise in the Sado estuary. *Journal of Coastal Researche*, 18 (1), p. 125-138

Schröder-Lanz, H. (1971). Die ersren ^{14}C datierten Mittelwurmbildungen von der sudlichen Alentejokuste (Portugal). *Eisszeitalter u. Gegenwart*, 22, p. 3-42.

Variações Holocénicas do Nível Médio do Mar: Precisão e Compatibilidade de Dados

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A detecção de variações holocénicas do nível médio do mar é tarefa de grande relevância científica que, entre outras consequências com aplicação sócio-económica, se revela fundamental na determinação da vulnerabilidade das zonas costeiras e do risco a que a ocupação humana dessas zonas está sujeita. Porém, a detecção de variações aludidas não é, normalmente, fácil, sendo imprescindível adoptar posição bastante critica sobre a validade dos resultados.

Verifica-se que, com frequência, surgem na literatura científica resultados cuja precisão é incompatível com a qualidade dos dados utilizados e até, nalguns casos, incoerências ou erros derivados de incorrecta interpretação de dados cronológicos.

Este último aspecto é particularmente importante. Por um lado, é extremamente arriscado misturar métodos de datação diferentes e não devidamente calibrados (por exemplo, utilizar em conjunto e acriticamente determinações de idade obtidas por radiocarbono, por termoluminescência, por racemização de aminoácidos, etc.). Por outro lado, mesmo os dados obtidos pelo método do Carbono 14 carecem de conhecimentos sobre a sua validade e significado, por forma a evitar misturar, como por vezes se verifica, idades calibradas com idades não calibradas, e idades corrigidas com diferentes efeitos de reservatório.

No que se refere aos indicadores do nível médio do mar, é difícil obter materiais para datação que estejam *in situ*, que tenham condições convenientes para o efeito. É principalmente no que se refere a estes indicadores do nível médio do mar que a precaução e o criticismo do investigador devem ser aplicados de forma rigorosa. Como se referiu, é muito difícil encontrar indicadores directos do nível médio do mar. Na realidade, não são muitos os organismos que vivem na faixa entre-marés e que apresentam condições de preservação. Face a esta dificuldade, recorre-se, com frequência, a conchas de animais cuja profundidade normal de distribuição é conhecida. As conchas de ostra têm sido bastante utilizadas, embora haja polémica no meio científico sobre a precisão dos dados assim obtidos.

Com frequência, também, utiliza-se material proveniente de paleo-sapais atravessados por sondagens de prospecção. É, também, uma via de conhecimento bastante útil, embora, como sempre, seja necessário ser bastante critico sobre os resultados. Efectivamente, a distribuição altimétrica dos sapais é significativamente maior do que a variação da maré no litoral oceânico, e é difícil conhecer a compactação a que os sedimentos foram sujeitos.

Provavelmente os melhores indicadores são determinadas formas costeiras (plataformas de abrasão marinha, sapas de arribas, etc.). Todavia, também estes casos não são isentos de problemas, porquanto há características estruturais das rochas e fenómenos de recorrência de ocupação que é preciso considerar.