IS THE YOUNGER DRYAS RECORDED IN THE ESTUARINE SEDIMENTS FROM SOUTHWESTERN IBERIA PENINSULA?

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Introduction: the Younger Dryas in the North Atlantic area

The Polar Front shifting during the different phases of the Last Glaciation, directly influenced the palaeoclimatic conditions of the Iberian Peninsula during such period. Specially, the different southward breakthrough of the Polar Front determined the great or smaller influence of the Gulf Stream in the Iberian coasts (Mörner, 1993; Zazo *et al.*, 1996a).

During most of the Last Glaciation, broad areas of the North Atlantic was occupied by cold surficial waters and ice caps, and the Polar Front maximum advance reached a nearby latitude of 40oN (Ruddiman and McIntrye, 1981a), although data from Duprat (1983) indicate that there is record of the Polar Front influence towards suthernmost latitudes of 38oN. Approximately, at ca.13 Ka the Polar Front retreated towrads a nearby position to Iceland, underwent a northward extension through the North European coasts. Between 12 and 11,8 Ka a sharp cooling episode took place in Europe (Older Dryas stadial) followed by a fresh period, but warmer than the previous one (Interstadial Allerød), at approximately 11 Ka when the Polar Front moved back towards northernmost positions (Ruddiman and McIntrye, 1981b).

Data from Bard *et al.* (1987) indicate that this last shifting of the Polar Front was an extremely rapid event, suggesting that during the first setback of the Polar Front (ca.12500-12000 yrBP) it moves back from 35oN to 55oN. On the contrary, data from the atlantic coasts of the Iberian Peninsula (Fatela *et al.*, 1994) indicate that during the Last Glaciation the Polar Front did not reach southern latitudes than 42oN. Even though these data indicate a controversy about the maximum southern advance of the Polar Front on the Iberian coast, seem to be evident that, aside of its controversial position, this advance was sufficient for to trigger substantial modifications on the dynamics of the Gulf Stream.

The aforementioned Polar Front advance promoted a large scale cooling in Europe between 11 and 10 Ka pointing out the so called Younger Dryas Estadial (YD), which is also recorded in the Iberian Peninsula (Pons and Reille, 1988 Turner and Hannon, 1988; Holes, 1994a, 1994b; Allen *et al.*, 1996; Wansard, 1996; Zazo *et al.*, 1996a) and possibly in the North of Africa (Lamb *et al.*, 1989). In ice cores from Greenland this event is recorded as a sharp and short climatic deterioration (about 1000 years), but of great amplitude, developed between 11,5 and 10,5 Ka (Dansgaard *et al.*, 1982 ; Mayewski, 1994). Data from North America indicate also an abrupt climatic change at ca.11-10 Ka equivalent to the European YD. Therefore it seems to be that this cooling was clearly linked to major oceanographic changes in the North Atlantic, which seem to be confirmed by the occurrence of any other evidences of comparable climatic changes in those zones not influenced by the North Atlantic dynamics (Bell and Walker, 1992).

In the Atlantic-Mediterranean linkage area these changes and his effects in the surficial sea water temperature have been recorded in different studies. Weaver and Pujol (1988), based on *Gephyrocapsa oceanica/G.muellerae* rates (Coccolits) from sea cores in the Alboran sea, suggest that the Atlantic and the Alboran sea began to warming during the Bolling-Allerod interstadial showing an increase in *G.oceanica*. By contrast, the absence of this specie during the YD suggests a marked cooling in the Alboran sea at this event. The occurrence of *G.oceanica* in the Atlantic area during this same event suggest that the southatlantic peninsular sea water was warmer than the Alboran sea water at that moment.

New data of the sea surface water temperatures obtained from lipids, also from Alboran sea cores (Cacho *et al.*, 1996), show a nearly 4°C variation in the mean surface temperature between the Last Glacial and the present. A general warming took place between 15000 yrBP and 8500 yrBP and the Optimum was reached at ca. 7000 yrBP. Similar records have been founded from the study of subtropical dinoflagelates from the same core samples (Cacho *et al.*, 1996). Also, from this core, is observed that the thermal increase that took place during the Last Deglaciation was not continuous. During a first warming stage the surficial sea water temperature was stabilised during the Bolling-Allerod interstadial, falling between 11000-10000 yrBP (YD). Similarly, Cacho *et al.* (1996) observed from the percentage of organic carbon an increase in the marine productivity in this area during the first stage of the Last Deglaciation. This was observed also in diatoms, foraminiferes and isotopes studies (Abrantes, 1988; Vergnaud-Grazzini y Pierre, 1991) associated with a greater upwelling activity due to and acceleration in the western anticiclonic turn originated by the input of oceanic water trough the Gibraltar strait (Abrantes, 1988).

The record of the estuarine deposition sequences of the south-western Atlantic during the Upper Pleistocene - Holocene transition

Based on the studies studies of the depositional sequences of the estuaries of the Guadalete, Guadalquivir and Tinto-Odiel rivers (Dabrio *et al.* 1995; Goy *et al.*, 1996; Lario, 1996; Zazo *et al.*, 1996; Dabrio *et al.*, 1997), it has been proposed a sea level curve during the Upper Pleistocene - Holocene (Lario, 1996). The segment most interesting is the one corresponding to the beginning of the Holocene (Fig.1).

Around ca.18000 yrBP the sea level was located at -125/-120 m of the current level (Hernandez Molina, 1993), beginning a rapid rise approximately ca.16000-14000 yrBP, when it was displaced of about -100 m until about - 20/25 m of depth, level that reaches before 9600 yrBP. This depth is labelled in the base of the organic and peats levels founded from cores from the Guadalete and Odiel estuaries (Dabrio *et al.* 1995; Goy *et al.*, 1996; Lario, 1996; Zazo *et al.*, 1996b; Dabrio *et al.*, 1997). Nevertheless, we should emphasise that it exists a mistake range, so much in age as in-depth, due to the radiocarbon data accomplished in that levels and to the subsidence and compactation that it has affected to the sediments.

From 9600 yrBP and until at least ca.8000 yrBP is produced a deceleration in the sea level rise rate, labelled by the organic and peats deposits and lags of shells found in the cited cored (Dabrio *et al.*, 1995). A change in the environmental magnetic properties and grain size paramethers between this layers and the upper and lower ones has been related with changes in the depositional dynamic (Lario, 1996). This deceleration, that is recorded at least to 9600 yrBP, it is the one which must be related to the YD event and to the sea level rise stop that has been registered during this event at least in the Noth Atlantic (Mayewski, 1994; Stanley, 1995).

After 8000 yrBP a new acceleration was produced in the sea level rise rate and reach approximately - 12 m of depth at ca.6500-6000 yrBP. In that moment was produced a new deceleration in the sea level rise rate, or even fall, recorded by the input of river sediments, suported by the environmental magnetic features and the gain size analyses (Lario, 1996), as well as by the freshwater contributions shown by the isotopic values (Lario, 1996; Dabrio *et al.*, 1997).

In the estuary zones from the Atlantic area the maximum marine conditions were reached , even though the level of the sea were located to some 12 m. for below of the current level. This moment has been fixed between 7600 and 7100 yrBP (Goy *et al.*, 1996; Lario, 1996; Dabrio *et al.*, 1997) , even though seems that it can be extended until ca.6800-6500 yrBP (Lario, 1996).

A new acceleration in the sea level rise took place subsequent at ca.5800 yrBP reaching the transgressive maximum ca.5000 yrBP (5800 CalBP), being maintained this level until ca.4500-4200 yrBP. It is remarkable to verify that it is to ca.5000 yrBP when generally it is assumed that the glacio-eustatic sea level rise was completed (Pirazzoli, 1991; Mörner, 1994, 1996). Equally, in the study of Pirazzoli (1991) of 594 relative sea level rise curves, the author finds a sea level rise at 5000 yrBP in , at least, half of the studied localities.

At ca.4200 yrBP was produced a series of negative and positive oscillations that, together with the sedimentary dynamics of the moment, favours the developing of spit bar systems (Zazo *et al.*, 1994; Lario *et al.*, 1995; Lario, 1996). Small positive pulsation's (of the decimetre order) with a ca.350-400 years cyclicity, were responsible for the deposition of the different sets recognised in each morphosedimentary unit. Rises of the decimetre to metric order, generated in moments of climatic instability, produce different erosion gaps that separate the different units. These rises have been also registered as marine terraces (Lario, 1996).

Even though these trends are related to a natural conditions, their influence in the coastal morphology has been affected by the human activity and more intensely from 500 yrBP until the present time at least during the last 2000 yrBP, (Lario *et al.*, 1995; Zazo *et al.*, 1995).

As has been observed, once the transgresive maximum it reached (ca.5000 yrBP), a relax of the glacio-eustatic component is produced and the redistribution processes of the oceanic water is the dominating sea level changes factor. In our case the Gulf Stream is the one which affects our coasts, existing since that date record of decimetre to metric order variations in relationship at the current sea level. The influence of the distribution of the Atlantic oceanic water in the the sea level changes and in the evolution of the coastal morphology is understood if the existing record in other similar characteristics areas to the our is review.

Relation between the sea level rise deceleration at ca.9000 yrBP and the YD: data from nearly records

The Warne and Stanley (1995) work about Upper Pleistoce - Holocene delta and estuarine sequences from subsiding areas all around the world suggest that would be a comun pattern in the sea level changes trend. A minimu 1 level is present ca.20000-18000 yrBP with a rapid rise between 16000 and 9000 yrBP (Fairbanks, 1989). A deceleration in the sea level rise rate took place between

9000 and 5000 yrBP and reach the present day level at ca.5000-4000 yrBP (Lighty *et al.*, 1982). Better resolution of this data (Stanley, 1995) show a sea level nearly -125/-120 m from the present day at ca.18000 yrBP (Fairbanks, 1989). A rapid sea level rise took place near 8000-6500 yrBP. At the end of this phase a significant desaleration in the sea level rise rate occurs. Equally, changes in the sea level changes curve ca.11500 yrBP related with the YD event have been recorded in continental margins and coasts (Mayewski, 1994; Stanley, 1995).

The works of the south-atlantic portuguese shelf (Rodrigues *et al.*, 1991) show a small sea level fall related by the author to an interruption in the general postglacial sea level rise trend and due to the climatic conditions that took place during the YD. This fall is not recorded in the south-atlantic spanish shelf, but a sea level stillstand has been recorded (Hernadez Molina *et al.*, 1994).

In the mediterranean area some deposits of basal peats at the bottom of the holocene sequence have been observed and these would dial, equally, a deceleration or even small drop, in the sea level rise rate. The works of Viñals (1991) and Viñals and Fumanal (1995) of the Gulf of Valencia show the existence of basal peat layers recorded from cores from the Pego laggon, with some ^{14}C ages between 10120±460 yrBP and 7120±90 yrBP as it was deposited progressively landwards and peat layers were drowned and covered by retreating-beach deposits. The works of Stanley (1990) and Stanley and Warne (1993) of the Nile delta fixed the age of these basal peats at 8140±130 yrBP, being interpreted equally as deposits related with a deceleration in the sea level rise rate, and being straddled within the age ranges found in this area.

Conclusions

The record of the estuary data do not allow to accomplish reconstructions of palaeotemperatures. Even though is possible observe that the climatic oscillation recorded during the Allerød-Younger Dryas-Preboreal cycle has remained recorded as a deceleration of the sea level rise that starts to ca.9000 yrBP, subsequent to the YD, after which is recorded all the delta-estuarine holocene sequence.

This interpretation is compatible with the additional data founded by the cited authors in the atlantic-mediterranean linkage area, where they have recorded a cooling of the surficial water related to the generalized cooling that took place during the YD.

The differences between the ages presented by the different authors from the oceanographic records and from the response founded in the estuarine deposits can be explained from the different precision and calibrations of the radiocarbon data, but equally it can be must to a lag in response to the new dynamic conditions in the estuarine systems, since it has been observed in coastal systems that the necessary relaxing times to indicate morphological changes are of the order of 10^3 years (Cowel and Thom, 1994).

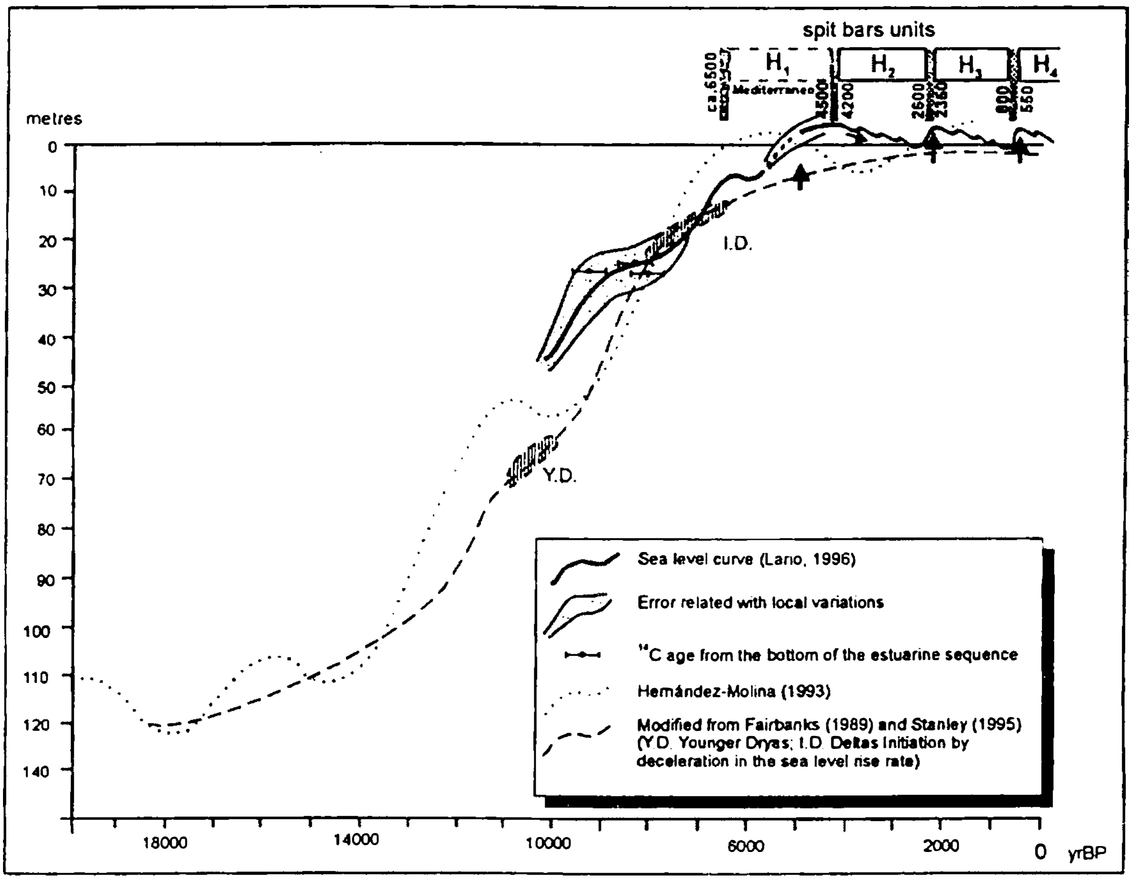

Fig. 1.- Sea-level curve in the Atlantic-Mediterranean linkage area (After. Lario. 1996).

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