Holocene earthquake record offshore Portugal (SW Iberia): testing turbidite paleoseismology in a slow-convergence margin

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ABSTRACT

The SW margin of the Iberian Peninsula hosts the present-day boundary between the Eurasian and African Plates. Convergence (4–5 mm/yr) is accommodated through a wide deformation zone characterized by moderate magnitude seismic activity. This zone has also been the source of the most important seismic events in Western Europe, such as the 1755 Lisbon Earthquake and Tsunami and 1969 Horseshoe Earthquake. Despite efforts to identify active seismogenic structures in the Gulf of Cadiz in the last ten years, little is known about its paleoseismic history. The turbidite paleoseismology approach was applied for the first time in a low-rate convergent margin to determine the recurrence interval of large earthquake events that occurred in SW Iberia during the Holocene. Four sediment cores collected at strategically located sites offshore Portugal (i.e. Tagus Abyssal Plain, Infante Don Henrique Basin and Horseshoe Abyssal Plain) reveal that these deep-sea basins preserve a record of episodic deposition of turbidites. In the SW Iberian Margin excluding special climatic events, earthquakes are the most likely triggering mechanism for synchronous, widely-spaced distributed turbidites during the Holocene, when the sea level was relatively stable. Age correlation together with textural, mineralogical, physical properties and geochemical signatures of the new cores complemented by pre-existing multicores and gravity cores reveals a total of 7 widespread turbidite events for the Holocene. Precise dating of the most recent turbidite event (E1) based on 210Pb and 137Cs geochronology provides an age of AD 1971 ± 3. This age corresponds to a high-magnitude instrumental earthquake in the region: the 1969 Horseshoe Earthquake (Mw 8.0). Calibrated 14C ages of subsequent widespread turbidite events (E3 and E5) correlate with the dates of important historical earthquakes and paleotsunami deposits in the Gulf of Cadiz area, such as AD 1755 and 218 BC, respectively. If older synchronous events (E6, E8, and E10) with ages ranging from 4960–5510 yr BP to 8715–9015 yr BP are also taken into account, a great earthquake recurrence interval of about 1800 years is obtained for the Holocene. Our correlations suggest that the turbidite record may be considered as a proxy for paleoseismic activity in low-convergence rate margins, and a valuable complementary tool in earthquake and tsunami hazard assessment along the coasts of the Iberian Peninsula and North Africa.

1. Introduction

Crustal deformation in the SW margin of the Iberian Peninsula is controlled by the NW–SE convergence of the African and Eurasian Plates (4.5–5.6 mm/yr) at the eastern end of the Azores–Gibraltar zone (e.g. Argus et al., 1989; McClusky et al., 2003). This convergence is accommodated through a wide active deformation zone (e.g. Sartori et al., 1994; Hayward et al., 1999) characterized by low moderate magnitude seismicity (Udías et al., 1976; Grimison and

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However, great earthquakes (Mw ≥ 8.0) such as the 1755 Lisbon Earthquake and Tsunami and the 1969 Horseshoe Earthquake have also occurred in this region (Fukao, 1973; Buforn et al., 1995, 2004; Baptista et al., 1998; Martínez Solares and López Arroyo, 2004) (Fig. 1). A multidisciplinary marine geological and geophysical dataset acquired during the last ten years offshore southern Portugal revealed a number of active NE–SW trending west-verging folds and thrusts (Zitellini et al., 2001, 2004; Grácia et al., 2003a; Terrinha et al., 2003) and WNW–ESE strike-slip faults (Rosas et al., 2009; Zitellini et al., 2009; Terrinha et al., 2009) (Fig. 1). Deformed Quaternary units together with swarms of earthquakes associated with seafloor surface ruptures suggest that these faults are active and that they may represent an earthquake and tsunami hazard to the coasts of Portugal, Spain and North Africa (Grácia et al., 2003a,b; Terrinha et al., 2003, 2009; Zitellini et al., 2004).

Assessment of seismic hazard in SW Iberia is largely based on the relatively short period of instrumental (<50 years) and historical (<2000 years) earthquake catalogues (e.g. Peláez and López Casado, 2002). This may not be sufficient to assess seismic hazard models in the Iberian Peninsula, especially when considering high-magnitude earthquakes and long recurrence intervals (>103 years) (e.g. Masana et al., 2004; Grácia et al., 2006; Vizcaino et al., 2006). A submarine paleoseismic approach may allow us to determine past seismic activity and to obtain a recurrence rate for great magnitude earthquakes (Mw ≥ 8.0).

In order to investigate the recurrence rate of large Holocene events, such as the Lisbon Earthquake, we tested for the first time the “turbidite paleoseismology” concept (Adams, 1990; Nelson et al., 1996; Goldfinger et al., 2003, 2007) in a low-convergence margin. Data from four sediment cores collected in slope basins and abyssal plains off SW Iberia are presented (Fig. 2). The objectives of this study are fourfold: 1) to characterize turbidite events on the basis of sedimentary facies, texture, physical properties, geochemical composition and radiocarbon ages; 2) to establish a regional correlation of turbidite events integrating results from 210Pb dating of the recentmost events (Garcia-Orellana et al., 2006) and from a local study at the Marques de Pombal Fault area (Vizcaino et al., 2006); 3) to propose a correlation of widespread turbidite events with instrumental and historical earthquakes, and tsunami deposits from the Gulf of Cadiz; and 4) to determine a recurrence interval of large earthquakes that occurred during the Holocene, highlighting the potential of the turbidite record as a marine paleoseismic indicator in low-convergence rate margins.

2. Regional setting

Active deformation in the SW Iberian Margin is revealed by earthquake mechanisms covering both onshore and offshore regions...
Focal mechanisms are heterogeneous and faulting style ranges from reverse to strike-slip (Stich et al., 2005a), as corroborated by recently acquired marine geological and geophysical data (Gutscher et al., 2002; Gra`cia et al., 2003a,b; Terrinha et al., 2003; Gutscher, 2004; Zitellini et al., 2004, 2009)(Fig. 1). Regional seismicity in the Ibero-Maghrebian region is diffuse and does not clearly define the present-day European–African plate boundary (e.g. Sartori et al., 1994; Buforn et al., 1995; Hayward et al., 1999). In the SW Iberian Margin, located at the eastern end of the Azores–Gibraltar zone, seismicity is characterized by shallow to deep earthquakes of low to moderate magnitude (Mw < 5.5) (Udías et al., 1976; Buforn et al., 1995, 2004; Stich et al., 2003, 2005a). However, this region is also the source of the largest and most destructive earthquakes that have affected Western Europe (AD 1531, 1722, 1755 and 1969). The 1755 Lisbon Earthquake (estimated Mw > 8.5) destroyed Lisbon (intensity X–XI MSK) and was accompanied by tsunamis that devastated the SW Iberian and NW African coasts (more than 60,000 casualties in Portugal, Baptista et al., 1998). On the basis of geological evidence, geophysical data and tsunami modelling, different geodynamic models and mechanisms have been proposed as the source of the Lisbon Earthquake (Gutscher et al., 2002; Baptista et al., 2003; Gracia et al., 2003a; Terrinha et al., 2003; Gutscher, 2004; Zitellini et al., 2004, 2009; Stich et al., 2007). However, none of these models satisfactorily accounts for the estimated magnitude of the earthquake and tsunami arrival times at the different localities onshore.

The study area is located in the outer Gulf of Cadiz from 36°N to 37°30’N, in the region between the Gorringe Bank and Cape São Vicente comprising the basins from the Tagus Abyssal Plain (TAP) to the Horseshoe Abyssal Plain (HAP) (Figs. 1 and 2). This region is characterized by an abrupt, irregular physiography dominated by massive ridges and large seamounts, highly incised narrow canyons, and deep, extensive abyssal plains (Fig. 2). As for the infill of the TAP and HAP plains, Lebreiro et al. (1997) corroborated their alternating turbidite and hemipelagic layer composition. These authors also suggested that the emplacement of the Late Quaternary age turbidites, at least for the HAP, was not directly linked to sea-level changes, but probably related to seismic activity. In fact, the most recent turbidite layer identified in the HAP has an emplacement time of 140 ± 120 years BP, coeval with the AD 1755 Lisbon event (Thomson and Weaver, 1994; Lebreiro et al., 1997), suggesting the presence of earthquake-triggered turbidites in the area.

Fig. 2. Detailed bathymetric map of the external part of the SW Iberian Margin including the Horseshoe and Tagus abyssal plains, extracted from the ESF EuroMargins SWIM multibeam compilation (Zitellini et al., 2009). Red dots locate the CALYPSO piston cores studied in the present work and the multicore collected from the same sites presented by Vizcaino et al. (2006) and multicore MC1 and SW37 by Garcia-Orellana et al. (2006). IDHB: Infante Don Henrique Basin; MPF: Marquês de Pombal Fault; HF: Horseshoe Fault; SVC: São Vicente Canyon; LC: Lagos Canyon; PC Portimão Canyon; PB: Portimão Bank.

Apart from the two abyssal plains, the other study area corresponds to the Infante Don Henrique Basin (IDHB), a slope basin bounded by the foothills of the Gorringe Bank to the west and the Marquês de Pombal Fault (MPF) escarpment to the east (Fig. 2). Associated with active faulting, mass transport deposits have also
been recognized in the IDHB at the foot of the Marquês de Pombal fault block (Gracia et al., 2003a; Vizcaíno et al., 2006). Based on radiocarbon dating of the Holocene debris flows and turbidite records, the authors estimated a recurrence rate of mass movement deposits of less than 2000 years (Vizcaíno et al., 2006).

The first turbidite paleoseismology study in the SW Iberian Margin was carried out by Garcia-Orellana et al. (2006). These authors investigated six short sediment cores (50 cm long) collected by a multicorer in the area stretching from the Tagus to the Horseshoe Abyssal plains for radiometric ($^{210}$Pb and $^{137}$Cs) and sedimentological analyses (Fig. 2). Garcia-Orellana et al. (2006) dated the two most recentmost detrital layers and turbidite events (~150 years), the ages of which correspond to two large (Mw > 6.0) historical and instrumental earthquakes that occurred in the SW Portuguese Margin: the 1909 Benavente and 1969 Horseshoe earthquakes.

3. Data and methods

3.1. Core site location and core quality

This study is based on the Holocene sections of four giant CALYPSO piston cores, termed MD03 2701 to MD03 2704, which were acquired during the PICABIA–PRIME cruise (July 2003) on board the French RV Marion Dufresne (Table 1, Fig. 2). These cores are strategically located for characterizing the main depositional areas in the SW Iberian Margin (i.e. TAP, HAP and IDHB), where seismic activity is generated.

Core site MD03 2701 is located in the Tagus Abyssal Plain at 5000 m depth on the NW flank of the Gorgonne Bank (Fig. 2). Core site MD03 2702 is located at 3900 m depth in the IDHB, NW of the Marquês de Pombal Fault area. This core differs from the other cores in that it records fewer mass transport deposits. The two last piston cores, MD03 2703 and MD03 2704, were sampled in the Horseshoe Abyssal Plain around 4900 m depth. They are 12 km apart and separated by an active fault. Four multicores (GeoB 9095-2, GeoB 9095-1, GeoB 9089-1 and GeoB 9099-1) were acquired during the GAP 2003 cruise (RV Sonne) at the same sites as the four CALYPSO piston cores (Fig. 2). These multicores together with multicore MC1 provide information on the uppermost sediment record (Garcia-Orellana et al., 2006), which is often lost during piston coring.

The CALYPSO piston cores were designed to acquire long sedimentary records. However, the quality of the upper core sections is inhomogeneous owing to piston stretching. For instance, the top of the core may be lost during coring and the upper sections are often deformed showing a bending of laminae in which the apex of bent layer is thickened due to sediment flow (Skinner and McCave, 2003). Bending especially affects the sampling of hemipelagic sediment for dating, which instead of being acquired just below turbidite bases needed to be sampled further downcore (up to 6 cm).

The methodology followed to study the piston cores included description, imaging, physical properties and geochemical measurements on half core sections. Sediment composition and grain-size analyses, smear-slide description and radiocarbon dating were carried out on selected samples.

3.2. Sediment measurements and analytical procedures

Immediately after core splitting and cleaning on board the R/V Marion Dufresne, all core sections were imaged with digital colour photo and logged for physical properties at 2 cm intervals using the multisensor core logger from GEOTEK. Sediment physical property measurements included magnetic susceptibility, P-wave velocity and gamma-ray attenuation from which density is calculated. Lightness ($L^*$) and colour parameters ($a^*$ and $b^*$) were manually measured every 2 cm using a spectrophotometer on board. Detailed core description has been performed on based on changes observed in the colour, lithology, texture and structure of the sediments (Fig. 3).

Geochemical composition was measured on archive sections using the non-destructive X-Ray Fluorescence (XRF) scanner from

<table>
<thead>
<tr>
<th>Core #</th>
<th>Lat (%)</th>
<th>Lon (%)</th>
<th>Water depth (m)</th>
<th>Total core length (m)</th>
<th>AMS lab reference</th>
<th>Core depth (cm)</th>
<th>Foraminifera sampled</th>
<th>Radiocarbon age (Cal yr BP ± 1σ)</th>
<th>ΔR (ppm)</th>
<th>1σ calibrated age (Cal yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MD03 2701</td>
<td>37°30.01</td>
<td>11°02.01</td>
<td>5052</td>
<td>8.5</td>
<td>49063</td>
<td>25–27</td>
<td>Mixed</td>
<td>2530 ± 100</td>
<td>95 ± 15</td>
<td>1940–2210</td>
</tr>
<tr>
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<td>37°30.01</td>
<td>11°02.01</td>
<td>5052</td>
<td>8.5</td>
<td>49064</td>
<td>65–67</td>
<td>O. uni., G. sac.</td>
<td>5200 ± 70</td>
<td>95 ± 15</td>
<td>5395–5565</td>
</tr>
<tr>
<td>MD03 2701</td>
<td>37°30.01</td>
<td>11°02.01</td>
<td>5052</td>
<td>8.5</td>
<td>49065</td>
<td>130–132</td>
<td>O. uni.</td>
<td>6770 ± 40</td>
<td>95 ± 15</td>
<td>7160–7250</td>
</tr>
<tr>
<td>MD03 2701</td>
<td>37°30.01</td>
<td>11°02.01</td>
<td>5052</td>
<td>8.5</td>
<td>49066</td>
<td>218–220</td>
<td>O. uni.</td>
<td>8650 ± 60</td>
<td>95 ± 15</td>
<td>9100–9290</td>
</tr>
<tr>
<td>MD03 2701</td>
<td>37°30.01</td>
<td>11°02.01</td>
<td>5052</td>
<td>8.5</td>
<td>52940</td>
<td>230–232</td>
<td>Mixed</td>
<td>8760 ± 55</td>
<td>95 ± 15</td>
<td>9275–9415</td>
</tr>
<tr>
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<td>11°02.01</td>
<td>5052</td>
<td>8.5</td>
<td>49067</td>
<td>289–291</td>
<td>G. rub.</td>
<td>9700 ± 45</td>
<td>95 ± 15</td>
<td>10425–10535</td>
</tr>
<tr>
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<td>11°02.01</td>
<td>5052</td>
<td>8.5</td>
<td>49068</td>
<td>370–372</td>
<td>O. uni.</td>
<td>12100 ± 60</td>
<td>95 ± 15</td>
<td>13350–13505</td>
</tr>
</tbody>
</table>

Table 1

Location of the studied cores, AMS radiocarbon data and sample age calibrations based on the Marine04 curve (Hughen et al., 2004) included in OxCal 4.0 calibration software.

O. uni.: Orbulina universa; G. sac.: Globigerinoides sacculifer; G. rub.: Globigerinoides ruber; G. con.: Globigerinoides conglobatus; N. pach.: Neogloboquadrina pachyderma.

* Local reservoir correction (ΔR) for the Portuguese margin based on Monte Soares and Alveirinho Dias (2006).

a Ages in brackets are not used in the present study: Sample obtained within a debrite.

b Sample rejuvenated by mixing from overlying sediment.
the University of Bremen (Germany). We measured the following elements: K, Ca, Ti, Fe, Mn, Cu, Sr, V, Cr, Co, Ni, Zn and Pb, at 2 cm interval for cores MD03 2701, MD03 2703 and MD03 2704, and at 1 cm resolution for core MD03 2702. The data obtained correspond to the values of element intensity in counts per second (cps), providing information on the relative element concentration. In the present study we selected K, Ca and Ti, which are the elements that better characterize the relationship between detrital and biogenic sedimentation (e.g. Adegbie et al., 2003) (Fig. 3). These data help us to characterize different types of turbidites as well as to define the hemipelagic intervals between turbidites that are essential to calculate the age models.

Grain-size analyses were systematically carried out every 10 cm except in the homogeneous core MD03 2702 with a sample interval of 20 cm. In addition, other samples were obtained in turbidite bases or debrites adding to make 205 samples. We used the Coulter

![Image](image_url)

Fig. 3. Image, lithological description, grain-size distribution, magnetic susceptibility, density and geochemical composition (K/Ti and Ca/Ti) of the studied sections of the CALYPSO piston cores: a) MD03 2701 from the Tagus Abyssal Plain, b) MD03 2702 from the Infante Don Henrique Basin, c) MD03 2703 and d) MD03 2704, both from the Horseshoe Abyssal Plain. Turbidite numbers are depicted for each core.
LS200 from the GRC Geociències Marines group of the University of Barcelona, which provides the grain-size data as a volume percentage for all the textural distribution (<4 μm to 2 mm). Sand fraction and very coarse silt components (>50 μm) were identified using a binocular microscope. Relative mineral abundance was estimated by counting a minimum of 300 grains per sample. In addition, to define the boundary depths between turbidite tails and hemipelagites, 105 smear slides were qualitatively analyzed.

3.3. Radiocarbon dating, calibration and age models

Radiocarbon dating was performed using 24 samples of hemipelagic sediment mainly located about 0.5–2 cm below the turbidite bases, locally reaching up to 6 cm. For AMS 14C dating, we hand-picked between 7 and 10 mg individual foraminifera of the same species with a diameter larger than 250 μm. *Orbulina universa* was preferentially used because it was the most common species, although we also selected *Globigerinoides ruber* (var. alba), *Globigerinoides sacculifer*, *Globigerinoides conglobatus*, *Neo-globoquadrina pachydermu* and mixed samples. Foraminifera were prepared and dated at the National Ocean Sciences Accelerator Mass Spectrometry Facility (NOSAMS-WHOI laboratory, Woods Hole, USA) (Table 1).

To obtain an accurate turbidite event chronology, the first step was to calibrate the 14C ages of the hemipelagic sediment samples using the Marine04 curve (Hughen et al., 2004) included in the OxCal 4.0 software. To this end, it is necessary to know the value of $ΔR$, the site-specific offset from the global ocean reservoir. In the SW Portuguese Margin, 14C dating of marine shells and associated charcoal or bones from archaeological sites with ages spanning the Holocene yielded a wide range of reservoir ages suggesting fluctuations in the intensity of coastal upwelling (e.g. Monge Soares and Alveirinho Dias, 2006). However, most of the values obtained for the Holocene period are lower than the modern value (250 ± 25 14C yr). To find a suitable $ΔR$ value, we considered the weighted mean of the $ΔR$ obtained for the period from 3000 to ~600 yr BP, which was sampled at a greater resolution including more than 30 $ΔR$ values. This period, with a weighted mean of 95 ± 15 14C yr, seems to have lower upwelling than at present (Monge Soares and Alveirinho Dias, 2006). The resulting calibrated ages using this $ΔR$ are given in Table 1.

The next step was to determine the age of each single turbidite with sufficient precision and accuracy. Since the samples of hemipelagic sediment are located a few cm below the base of the turbidites or tens of cm above the turbidite tails, it was necessary to interpolate or extrapolate the ages of the calibrated samples as far as the base or top of the neighbouring turbiditic intervals. For this purpose we used the *P_Sequence*, a Bayesian model of deposition implemented in the computer program OxCal 4.0 (Bronk Ramsey, 2008). Given that the hemipelagic sedimentation rate cannot be regarded as perfectly constant, the *P_Sequence* depositional model takes into account the uncertainties in the variation of the hemipelagic sedimentation rate by regarding sedimentation as an inherently random process. The resulting age model, referring to the calibrated age scale, reflects the increasing uncertainties with distance from the calibrated sample ages (Fig. 4).

To run the *P_Sequence* model, apart from $ΔR$, several input parameters are needed. First, the uncalibrated 14C ages and their corresponding sample depths are provided as the main dataset. Second, the uncalibrated ages of the top and bottom boundaries of each core are estimated. These are determined with ample margins, only constrained by the age of the seafloor at the time of sampling and by the age of the shallowest sample for top boundaries, and by the age of the deepest sample for bottom boundaries. Finally, the regularity of the sedimentation process is determined by factor $k$, with the higher values of $k$ reflecting smaller variations in sedimentation rate (Bronk Ramsey, 2008). For each core, we chose the highest possible values of $k$, on condition that the modelled age fitted each individual calibrated age with $X^2 < 1$ (all agreements >68.2% in *P_Sequence* model output). Applying this criterion, we used $k = 1.5$ for cores MD03 2703 and MD03 2704, and $k = 1$ and $k = 0.8$ for cores MD03 2701 and MD03 2702, respectively. For each core we obtained the 68% and 95% probability ranges which are plotted in the calibrated age vs. hemipelagic depth model (Fig. 4). Based on the hemipelagic depth of every single turbidite, the age models yield the calibrated ages which are reported as maximum probability values, 1σ and 2σ age ranges in Table 2. In the following sections, turbidite calibrated ages are reported with 1σ ranges based on the most likely turbidite correlations that characterize the occurrence of events.

Owing to the limited number of cores available for this study, it was not possible to quantify basal erosion in the calculation of the time interval between the dated sample and the turbidite base (e.g. Nelson et al., 1996; Gutiérrez-Pastor et al., 2009). Consequently, turbidite erosion was assumed to be negligible for modelling the 14C dates to obtain the ages of the turbidites.

3.4. Criteria for turbidite correlation

In this study we use the fundamental concept of “event” (E). This may be constituted by one or more mass transport episodes (turbidite or debrite) correlated across different depositional areas. The correlation between turbidite deposits may be based on a number of factors of which the most important is chronology. Different turbidite deposits located far apart from one another will be regarded as the same turbidite event if their calibrated ages overlap at 1σ uncertainties. As part of the same event, these turbidites will be known hereinafter as “synchronous” or “coeval”. In addition, when synchronous turbidites are found in at least two of these widely separated depositional areas, these events will be referred to as “widespread”. As previously mentioned, radiocarbon dating was performed in the hemipelagic material found below each turbidite layer. Since hemipelagic thickness might have been partially reduced by erosion from the overlying turbidite, we decided to use the youngest age from each event to characterize its occurrence. This geological criterion allows us to disregard the older turbidite ages of a given event which are probably affected by the ageing effect of erosion.

Other criteria can also be used to correlate turbidite events. Sedimentological aspects, such as hemipelagic thickness between different turbidite sequences (e.g. Gutiérrez-Pastor et al., 2009), are also important factors that strengthen the correlation. In addition, physical parameters (magnetic susceptibility and density) and geochemical composition (K/Ti and Ca/Ti) may also reinforce the correlation between different turbidites (e.g. Goldfinger et al., 2007), especially if they are from the same source area. In summary, if turbidite deposits from different cores (T) meet the aforementioned correlation criteria, these turbidites will be considered to be synchronous and will form part of the same turbidite event (E).

4. Results

### 4.1. Sedimentary facies

Three main sedimentary facies were distinguished in the cores: 1) hemipelagites, 2) debrites and 3) turbidites. Hemipelagite was mainly described as homogeneous olive (Munsell notation 5Y 5/4) bioclastic silty-clay (52.4% clay, 43.6% silt and 3.9% sand and a mean diameter of 6.2 μm) highly bioturbated. The thickness of hemipelagic deposits ranges from few millimetres up to 60 cm, with exception
Fig. 4. Age-depth model output for cores a) MD03 2701, b) MD03 2702, c) MD03 2703, and d) MD03 2704 assuming the deposition is a Poisson process using P_Sequence from OxCal 4.0 software (Bronk Ramsey, 2008). 1σ and 2σ probability distributions are depicted for the calibrated sample ages (in black), sequence boundary ages (in white) and modelled turbidite ages (in grey). The value of \( k \) used to run each age model is indicated.
of core MD03 2702, where a single hemipelagic interval exceeds 220 cm in thickness. Debrite only appears in core MD03 2702 and is characterized by homogeneous greyish brown silt-clay (46.7% clay, 48.1% silt and 5.2% sand and a mean diameter of 6.2 μm). Despite textural similarities with hemipelagite, debrite tends to be darker (brownish) with an absence of bioturbation, and may show mud clasts and soft sediment deformation.

Turbidite facies was characterized as a fining-upward sequence with a sharp erosive base, a coarse interval (18.7% clay, 30.8% silt and 50.5% sand and mean diameter of 2.5 μm) and a turbidite tail (48.5% clay, 51.3% silt and 0.2% sand and mean diameter of 6.5 μm), representing well-distinguished sedimentary layers described in the Bouma sequence. In core MD03 2701 from the TAP, we identified 8 grey to olive grey (5Y 5/1 to 5Y 4/2) turbidite layers (T1–T8) with a thickness ranging from 3 to 60 cm. The sand fraction is composed of quartz, feldspar, carbonates, micas and biogenic components (mainly foraminifera) reaching 45% of the total of this fraction. Turbidites were classified as poorly sorted sand to silt with a mean diameter of 1.9 μm (T7) to 5.4 μm (T6) (Fig. 3a). However, two silt-clay turbidites (mean diameter of 6.7 μm) were identified in the upper section (T1 and T3). Few sand layers showing a disrupted sedimentary structure appear anomalously intercalated in the sedimentary sequence (from the sides of the core (Fig. 3a). They resulted from disturbance during piston coring. However, it was possible to accurately identify these artefacts and exclude them from the dataset.

In core MD03 2702 (IDHB), 2 turbidites (T1 and T2), 30 and 25 cm thick respectively, are characterized as greyish brown (2.5Y 5/2) sandy silt. These two turbidites are mainly composed of biogenic components. A debrite was identified under turbidite T1 between 51 and 66 cm depth (Fig. 3b). The HAP cores, MD03 2703 and MD03 2704, are formed by a succession of 8 (T1–T8) and 9 (T1–T9) turbidite layers, respectively. The colour of turbidites varies from olive (5Y 4/4) to dark grey (5Y 4/1), and their thickness ranges from 4 to 140 cm (Fig. 3c,d). The sand fraction is composed of quartz, feldspar, carbonates, micas, heavy minerals and biogenic components (50% foraminifera), which may reach 72% of the total components of this fraction. The grain-size distribution of the turbidite bases ranges from silty-clay (mean diameter of 5.8 μm) to medium sand (mean diameter of 3.6 μm).

4.2. Physical properties

Downcore variability of the magnetic susceptibility (MS), density and lightness (L′) allowed us to complete a detailed characterization of the sedimentary facies (Fig. 3). In core MD03 2701, hemipelagites are characterized by low MS values around 10 SI and density values of about 1.9 g cm⁻³. Turbidites were divided into two groups based on MS and density. The first group corresponds to the turbidite deposits from T1 to T6, with MS ranging between 9 and 14 SI and density from 1.8 to 2.0 g cm⁻³. The second group including T7 and T8 presents higher MS values reaching up to 40 SI whereas density ranges from 1.9 to 2.4 g cm⁻³ (Fig. 3a). Turbidite layers are darker (L′ = 42) than the hemipelagites (L′ = 55). This suggests that the lightness shows a good correspondence with the sediment composition and texture. In core MD03 2702, hemipelagites, debrites and turbidites facies show fairly constant MS values (8–18 SI). This is not the case for density and L′, which increase from 1.5 to 2.4 g cm⁻³ and from 45 to 53, respectively (Fig. 3b).

In core MD03 2703, two groups of turbidites were distinguished based on their physical properties (Fig. 3c). In the upper interval (from T1 to T5), magnetic susceptibility ranges between 5 and 24 SI, density oscillates from 1.4 to 2.1 g cm⁻³. In the lower interval (T6–T8), MS and density vary between 8 and 34 SI and 2–2.4 g cm⁻³, respectively. L′ is characterized by a successive increase in darker colours, from 47 in T1 to 31 in T7. In core MD03 2704, there are also two groups of turbidites depending on the physical parameters. In the upper interval (from T1 to T7), magnetic susceptibility ranges between 6 and 32 SI, density oscillates from 0.8 to 2.2 g cm⁻³, and L′ from 35 to 52. T4 constitutes an exception with low values of lightness (L′ = 29).

4.3. Geochemical composition

In light of XRF-scan data, it was possible to recognize relative abundances of terrigenous and biogenic sediment supply, based on the variability of K/Ti (detrital proxy) and Ca/Ti (biogenic calcareous proxy) ratios (e.g. Adegbie et al., 2003). Turbidites of core MD03 2701 are characterized by K/Ti measurements ranging between 1.5 and 3 whereas Ca/Ti is roughly constant (13), locally increasing up to 20 in T8 (Fig. 3a). Hemipelagites are characterized by K/Ti varying from 1.5 up to 2.5 and Ca/Ti from 10 to 35 (Fig. 3a). In core MD03 2702, the uppermost turbidite (T1) and debrite could not be geochemically analyzed because of the watery conditions of the sediment that were unsuitable for XRF scanning. In general, K/Ti and Ca/Ti do not show large oscillations in core MD03 2702 (Fig. 3b). In turbidite T2, the detrital ratio was lower than expected with values reaching 1.5 whereas the calcareous ratio ranged between 14 and 26. This might be explained by the relative abundance of transported biogenic components (mainly foraminifera) with respect to siliciclastic components. Hemipelagites show K/Ti values of 1.7–2.9 and of Ca/Ti between 10 and 36.

Cores MD03 2703 and MD03 2704 also show similar trends (Fig. 3c,d). In core MD03 2703, in the range of turbidites show K/Ti and Ca/Ti values fluctuating between 1.3 and 2, and 10 and 35, respectively. T6 and T7 in core MD03 2703 present the highest values of K/Ti (up to 2.5) and Ca/Ti (up to 43) respectively. In core MD03 2704, turbidites show higher values of K/Ti (between 1.3 and 3.4) and Ca/Ti (between 17 and 65) with marked peaks of both ratios observed in T4, T6 and T7 (Fig. 3d). In both cores, hemipelagites are characterized by detrital values ranging between 1.4 and 2.5 and calcareous values from 14 to 30.

4.4. Chronology of turbidite deposits

The 6 uppermost turbidites (T1, T2, T3, T4, T5 and T6) recorded in core MD03 2701 were deposited during the Holocene at around 1980–2280 yr BP, 4960–5510 yr BP, 7105–7250 yr BP, 8540–8985 yr BP, 9230–9370 yr BP, 10 175–10 425 yr BP, respectively (Fig. 4a, Table 2). The remaining two turbidites (T7 and T8) occurred during the Last Glacial–Interglacial Transition period at around 13 310–13 515 yr BP and 15 020–16 630 yr BP, respectively (Figs. 4a, 5; Table 2). In core MD03 2702, two turbidites and a debrite flow deposit were identified in the Holocene section. The uppermost turbidite (T1 and debrite was deposited at around 385–545 yr BP while turbidite (T2) was deposited at 7880–8145 yr BP (Fig. 4b, Table 2).

Five Holocene turbidites (T1–T5) and three pre-Holocene turbidites (T6–T8) were distinguished in core MD03 2703. In the Holocene section they occurred at around 2080–2620 yr BP, 6340–6505 yr BP, 6690–6985 yr BP, 8185–8425 yr BP, and 8715–9015 yr BP, respectively; and during the Last Glacial–Interglacial Transition period at 12 950–13 325 yr BP, 15 695–16 905 yr BP, and 16 160–16 635 yr BP, respectively (Fig. 4c, Table 2).

In core MD03 2704 the 7 uppermost turbidites (T1, T2, T3, T4, T5, T6 and T7) were deposited during the Holocene, at around 300–560 yr
BP, 855–1110 yr BP, 2285–2415 yr BP, 6110–6365 yr BP, 6745–7020 yr BP, 7930–8240 yr BP and 8880–9090 yr BP, respectively. Turbidites T8 and T9 occurred during the Last Glacial–Interglacial Transition period at around 13 430–13 715 yr BP and 14 275–14 780 yr BP, respectively (Figs. 4d and 5; Table 2).

5. Discussion

5.1. Turbidite events

Detailed turbidite characterization and accurate chronology allowed us to conduct a synchronicity test between turbidite deposits from the four piston cores described in this paper (Figs. 5–8). In addition, we also included the age results from four gravity cores of the MPF area (Vizcaino et al., 2006) and six multicores, two from the MPF area and four from the same locations as the CALYPSO piston cores (García-Orellana et al., 2006) (Fig. 2). The most recent turbidite events (E1 and E2) in the SW Iberian Margin are characterized by very thin (<2 cm thick) silty-clay deposits identified and dated in multicore MC1 (García-Orellana et al., 2006). E1 and E2 were not detected in the CALYPSO piston cores. Their turbidite deposits reflect different sediment characteristics and distribution. E1 is characterized by thicker turbidites that were identified in all the basins studied (TAP, IDHB and HAP) whereas E2 was locally observed only in the IDHB and MPF area. Precise dating based on 210Pb and 137Cs geochronology provides ages of AD 1971 ± 3 and AD 1908 ± 8 for the turbidite events E1 and E2, respectively (García-Orellana et al., 2006) (Fig. 7). The uppermost turbidite event recognized in the CALYPSO piston cores (T1 in both MD03 2702 and MD03 2704) is E3, which occurred at around 300–560 yr BP (Figs. 5 and 7, Table 2). This event can also be correlated with the most recent turbidite identified by Thomson and Weaver (1994) in the TAP at 300 ± 120 yr BP, and in the HAP at 140 ± 120 yr BP. E4 was only identified in core MD03 2704 and is characterized by a muddy turbidite with MS and Ca/Ti values lower than those detected in E3 (Fig. 6). This event occurred around 855–1110 yr BP.

The next event (E5), detected in the cores from the TAP and HAP, is widespread. This event includes the uppermost turbidite T1 from cores MD03 2701 and MD03 2703, and turbidite T3 from MD03 2704 (Figs. 3 and 6). This suggests that the top of cores MD03 2701 and MD03 2703 suffered a sediment loss with respect to core MD03 2704, where the base of the turbidite characterizing E5 in this core is located around 1 m downcore. E5 is characterized by muddy to silty turbidites with similar values of MS and Ca/Ti in all three cores (Fig. 6), and occurred at around 1980–2280 yr BP (Figs. 5 and 7, Table 2). Vizcaino et al. (2006) identified a turbidite deposit in the MPF area with a similar calibrated age (1940 ± 55 yr BP) (Fig. 8). Event E6 includes turbidite T2 of core MD03 2701 from the TAP (Fig. 5). E6 is the most biogenic turbidite deposit of core MD03 2704 and this could account for the similar values in physical properties and geochemical composition with respect to the hemipelagites (Fig. 6). E6 occurred around 4960–5510 yr BP (Figs. 5 and 7, Table 2). A coeval turbidite was identified by Vizcaino et al. (2006) in the MPF area (Fig. 8).

Event E7 was locally detected in the deep-sea cores from the HAP and includes turbidites T2 from core MD03 2703 and T4 from MD03 2704. This event is characterized by the thickest Holocene turbidites (58 cm thick in core MD03 2704), coarse turbidite bases, the highest values of MS, density, K/Ti and Ca/Ti, and the lowest values of lightness of all Holocene sections (Figs. 3 and 6). E7 occurred around 6110–6365 yr BP (Figs. 5, 7, 8, Table 2). Event E8 is widespread and was detected in cores MD03 2701 (T3) from the TAP and MD03 2703 (T3) and MD03 2704.
2704 (T5) from the HAP. Their turbidite deposits are characterized by low amplitudes in physical properties and geochemical composition (Fig. 6). E8 occurred at around 6690–6985 yr BP (Fig. 7; Table 2).

E9 is a widespread event and was detected in cores MD03 2701 (T4) from the TAP, MD03 2702 (T2) from the IDHB, and MD03 2703 (T4) and MD03 2704 (T6) from the HAP. Turbidite T4 of core MD03 2701 from the TAP is thicker, coarser grained, and has an older age than the rest of the E9 turbidites (Figs. 3, 5, 7; Table 2). This suggests a relatively high erosional potential, which would explain the older age of this turbidite. Given the stratigraphic position of T4 and T5 from this core, we assigned T4 to E9 and T5 to the immediately older event (E10) (Fig. 5). E9 occurred around 7880–8145 yr BP. E10 is the oldest widespread event in the Holocene and was detected in cores MD03 2701 (T5), MD03 2703 (T5) and MD03 2704 (T7). E10 occurred around 8715–9015 yr BP (Fig. 8; Table 2). E9 and E10 are characterized by high amplitude geochemical signatures detected in the two cores from the HAP (Fig. 6). The first event recorded during the Holocene in the SW Iberian Margin was the event E11, only detected in core MD03 2701 (T6), which occurred at around 10175–10425 yr BP (Table 2).

As regards the Last Glacial–Interglacial Transition period, the last event is E12, a widespread turbidite event identified in cores MD03 2701 (T7) from the TAP, and MD03 2703 (T6) and MD03 2704 (T8) from the HAP (Figs. 5–7). The three turbidite deposits characterizing E12 show differences in texture and in physical and geochemical properties. T8 of core MD03 2704 from the HAP is the thickest turbidite identified (>1 m thick) characterized by a coarse grain-size base showing large pulses in the MS (Fig. 6). In addition, the slightly older age of T8 of core MD03 2704 suggests that it may have eroded the underlying sediment (Figs. 6 and 7). E12 occurred around 12950–13325 yr BP (Fig. 7; Table 2). E13 is the oldest widespread turbidite event and was detected in core MD03 2701 (T8) from the TAP and cores MD03 2703 (T7) and MD03 2704 (T9) from the HAP. Its turbidite ages were extrapolated from the overlying deposits characterizing E12 (Fig. 7). E13 occurred at around 14275–14780 yr BP (Table 2). The oldest event is E14 only detected in core MD03 2703 (T8) from the HAP (Figs. 6 and 7), which occurred at around 16160–16635 yr BP (Table 2).

Summarizing, 11 Holocene turbidite events (E1–E11) and a total of 14 events for the last 16.5 ka (E1–E14) were identified in the SW Iberian Margin including the TAP, the IDHB, MPF area and the HAP. In the Holocene, only 7 of the total 11 turbidite events (E1, E3, E5, E6, E8, E9 and E10) are widespread, i.e. detected in cores from at least two of the different depositional areas (Figs. 5 and 6). The remaining four Holocene turbidite events (E2, E4, E7 and E11) were only identified in sediment cores from one of the basins (Fig. 7).

Fig. 5. Age correlation of turbidites from the Tagus Abyssal Plain (MD03 2701) and the Horseshoe Abyssal Plain (MD03 2704). Probability distribution curves of modelled turbidite ages (orange) were obtained using P.Sequence from OxCal 4.0 software (Bronk Ramsey, 2008). Black horizontal lines link turbidite bases and their respective calibrated ages. Modelled turbidite ages (1σ) are projected on the time axis by a purple band, where event numbers are indicated.
Turbidites have been used as a proxy for paleo-earthquakes since 1952, when the study of the 1929 Grand Banks Earthquake and associated turbidity current was published (Heezen and Ewing, 1952). However, they found difficulties in linking the gravity flow deposits with the 1929 Earthquake mainly because turbidity currents may be triggered by different processes. Adams (1990) proposed four triggering mechanisms for turbidite generation: a) sediment loading, b) wave-induced slumping, c) tsunamis and d) large earthquakes. This list was expanded by Goldfinger et al. (2003) adding e) crustal earthquakes, f) slab earthquakes, g) aseismic accretionary slip wedge, h) hyperpycnal flows, and i) gas hydrate dissociation. Tides (Sarı and Çagatay, 2006) and postglacial isostatic rebound (Blumberg et al., 2008) have also been regarded as possible turbidite triggers. A number of these mechanisms are related to local processes but only earthquakes, storm waves and hyperpycnal flows can trigger widespread turbidites in a region (Goldfinger et al., 2003). Hyperpycnal flows and storm waves seem to be the most important turbidite triggering mechanisms during the low-stand glacial periods, when there is a direct sediment supply from rivers to submarine canyons (Goldfinger et al., 2007). However, in an active, mid latitude margin such as the SW Iberian Margin, and during the high-stand Holocene period, earthquakes are the most likely mechanism for turbidite generation.

In turbidite paleoseismology, synchronicity is probably the most accepted criterion for suggesting that turbidity currents are triggered simultaneously by an earthquake. As defined by Adams (1990), Nelson et al. (1996) and Goldfinger et al. (2003, 2007) synchronicity is based on the "confluence channel test", i.e. the same number of turbidite events should be found upstream and downstream from the confluences between tributaries, slope channels and deep-sea channels. In the SW Iberian Margin, sediments are transported from the shelf edge and upper slope directly to the abyssal plains through large, deeply incised canyons. The particular physiography of this margin is unsuitable for applying the channel confluence test. Instead, the synchronicity test in this study is based on the existence of coeval deposits located in widely separated depositional areas (slope basins and abyssal plains) (Figs. 2, 5, 6 and 7).

Although it may be assumed that earthquakes are the most likely triggering mechanism of turbidites in the SW Iberian Margin during the Holocene, we should not exclude other non-seismic causes for the emplacement of widespread turbidite events (Fig. 7). Of the 7 widespread events (E1, E3, E5, E6, E8, E9 and E10), event E9 occurred immediately after the well known 8.2 ka climatic event (e.g. Bond et al., 1997, 2001). Although an earthquake cannot be ruled out as the triggering mechanism for the E9 turbidite event,
the 8.2 ka event may have created a climate-related instability in the SW Iberian Margin slope. Despite being widespread, we do not regard event E9 as seismically triggered (Fig. 8).

The non-widespread Holocene turbidite events (E2, E4, E7 and E11) may have been generated by low to moderate earthquakes, although we cannot exclude other non-seismic causes (Fig. 8). We are unable to assign an earthquake origin to the turbidite events during the Late Pleistocene because their triggering could have also been related to the rise of sea level and associated instability processes.
Fig. 8. a) Turbidite ages separated by study areas for the last 16,000 years. TAP: Tagus Abyssal Plain; IDHB: Infante Don Henrique Basin; MPF: Marquês de Pombal Fault area; HAP: Horseshoe Abyssal Plain. The time interval including all turbidite ages for event is depicted by light orange bands. The 1σ age given to each event (E1–E13) corresponds to the youngest turbidite age and is depicted by a dark orange band. The ages of turbidites presented in this work are depicted in black and grey (extrapolated), and the ages of turbidites from the MPF area are shown in red (Vizcaino et al., 2006). Dashed-line rectangle corresponds to Fig. 9. b) Eustatic sea-level curve (green) and GISP2 δ¹⁸O curve (purple) modified from Hernández-Molina et al. (1994) and Jouzel (1994), respectively.
5.3. Linking Holocene turbidite events with historical earthquakes and tsunami deposits in the Gulf of Cadiz

To lend support to the hypothesis that earthquakes are the main triggering mechanism of the widespread turbidite deposits in the SW Iberian Margin, we correlate these turbidite events with the instrumental and historical seismic records. Numerous seismic events have been instrumentally recorded in the area since the 1960s, the largest being the Horseshoe Earthquake in 1969 (Mw 8.0) (Fukao, 1973) (Table 3). In addition to the instrumental data, historical record allows us to investigate the seismicity for the last two millennia (Table 3) including great earthquakes, such as the 1755 Lisbon Earthquake and Tsunami (Luque et al., 2001, 2002; Ruiz et al., 2005, 2008; Baptista and Miranda, 2009, 2010) (Fig. 9; Table 3). Morphological coastal changes caused by this tsunami together with the magnitude of its sedimentary deposits suggested that this event was probably as destructive as the AD 1755 Earthquake and Tsunami (Luque et al., 2001, 2002; Ruiz et al., 2005, 2008; Baptista and Miranda, 2009, 2010). Two local turbidites (E2 and E4) may also be related to moderate historical earthquakes. Garcia-Orellana et al. (2006) suggested that E2 was coeval with the Benavente Earthquake in 1909 (Mw 6.0) (Moreira, 1984; Mezcua et al., 2004; Ruiz et al., 2005) (Table 3). E4 (855–1110 yr BP) seems to be coeval with the AD 881 historical earthquake and tsunami (Galbis, 1932; Mezcua and Martinez Solares, 1983) (Fig. 9a; Table 3).

Before 2500 yr BP, we should look for the geological record of past earthquakes and tsunamis. Unfortunately, there is no paleoseismological catalogue available for SW Iberia but instead, there is a relatively good record of paleotsunamis. Paleotsunami studies are based on sedimentological fieldwork in lagoons and marshes from the Iberian Peninsula around the Gulf of Cadiz and west Portuguese coast (e.g. Andrade et al., 1994; Dawson et al., 1995; Lario et al., 2001, 2010; Luque et al., 2001, 2002; Buflon et al., 2004; Martinez Solares and Lopez Arroyo, 2004; Mezcua et al., 2004; Ruiz et al., 2005; Stich et al., 2005a; Scheffers and Kelletat, 2005; Stich et al., 2007; Ruiz et al., 2008; Baptista and Miranda, 2009; Lario et al., 2010).

To lend support to the hypothesis that earthquakes are the main triggering mechanism of the widespread turbidite deposits in the SW Iberian Margin, we correlate these turbidite events with the instrumental and historical seismic records. Numerous seismic events have been instrumentally recorded in the area since the 1960s, the largest being the Horseshoe Earthquake in 1969 (Mw 8.0) (Fukao, 1973) (Table 3). In addition to the instrumental data, historical record allows us to investigate the seismicity for the last two millennia (Table 3) including great earthquakes, such as the 1755 Lisbon Earthquake and Tsunami (e.g. Martins and Mendes Victor, 2001; Martinez Solares and Lopez Arroyo, 2004; Baptista and Miranda, 2009) (Figs. 8 and 9).

Instrumental and historical records of earthquakes and tsunamis enabled us to identify three widespread turbidite events in the SW Iberian Margin for the last 2500 years: E1, E3 and E5 (Fig. 9a). E1, which occurred at around AD 1971 ± 3 dated by 210Pb chronology (Garcia-Orellana et al., 2006), was correlated with the AD 1969 Horseshoe Earthquake (Mw 8.0) (Fukao, 1973). E5, which occurred at 300–560 yr BP, was also identified by Thomson and Weaver (1994), which occurred at AD 1994 (Mw 6.7) (Fukao, 1973). E5 is a thin widespread turbidite event in the SW Iberian Margin. A relationship between the turbidite thickness and earthquake magnitude in this margin cannot be demonstrated because there are other factors to be considered, such as sediment availability and stability conditions in the source area. Coeval with E5 (1980–2280 yr BP) was the large historical earthquake and tsunami that occurred around 218 BC in the Bay of Cadiz in Roman times (e.g. Galbis, 1932; Luque et al., 2002; Ruiz et al., 2008; Lario et al., 2010) (Fig. 9; Table 3). Morphological coastal changes caused by this tsunami together with the magnitude of its sedimentary deposits suggested that this event was probably as destructive as the AD 1755 Earthquake and Tsunami (Luque et al., 2001, 2002; Ruiz et al., 2005, 2008; Baptista and Miranda, 2009, 2010). Two local turbidites (E2 and E4) may also be related to moderate historical earthquakes. Garcia-Orellana et al. (2006) suggested that E2 was coeval with the Benavente Earthquake in 1909 (Mw 6.0) (Moreira, 1984; Mezcua et al., 2004; Stich et al., 2005) (Table 3). E4 (855–1110 yr BP) seems to be coeval with the AD 881 historical earthquake and tsunami (Galbis, 1932; Mezcua and Martinez Solares, 1983) (Fig. 9a; Table 3).
(8715–9015 yr BP) could also have been seismically triggered (Fig. 9b).

Finally, we sought to determine how often great magnitude (Mw ≥ 8.0) seismic events, such as the AD 1755 Lisbon Earthquake and Tsunami, occurred in the SW Iberian Margin. In this regard, Ribeiro et al. (1996) suggested a broad recurrence interval of about 300–1500 yr, and Gutscher (2004) of about 1000–2000 yr, but none of them was based on marine paleoseismic studies. Based on the correlation between turbidite events and historical earthquakes and tsunamis, we established that the three widespread turbidite events (E1, E3 and E5) that occurred during the last 2500 years correspond to great instrumental and historical earthquakes of estimated Mw ≥ 8.0 (Fig. 9a). In contrast, local, non-widespread turbidite events (E2 and E4) that occurred during the same period of time, seem to correlate with historical earthquakes and tsunamis of lower magnitude (around Mw 6.0–7.0) (Fig. 9a). Hence, the historical record suggests that widespread turbidite events require great magnitudes (Mw ≥ 8.0) to be generated. Extrapolating to the pre-historical period, a Mw ≥ 8.0 is suggested for the oldest Holocene widespread events (E6, E8 and E10). Hence, considering the 6 Holocene widespread turbidite events as seismically triggered (Fig. 9b), a recurrence period for great earthquakes (Mw ≥ 8.0) of approximately 1800 years is determined.

6. Conclusions

Correlation between synchronous turbidite layers found in long distance apart basins (Tagus Abyssal Plain, Infante Don Henrique Basin and Horseshoe Abyssal Plain) allowed us to construct a turbidite event chronology for the SW Iberian Margin. In accordance with this chronology, we recognized a total of 14 turbidite events in the last 16.5 ka (E1–E14), of which 11 occurred during the Holocene.

Coeval turbidites from distal depositional areas can be used as a paleo-earthquake proxy during the high-stand Holocene period. This may account for the origin of the 7 widespread Holocene events identified (E1, E3, E5, E6, E8, E9 and E10). In support of this paleoseismic hypothesis, we use a correlation between the deep-sea sediment record, the instrumental and historical seismic record, and tsunami deposits of the SW Iberian Margin. Widespread turbidite events E1 (AD 1971±3), E3 (300–560 yr BP) and E5 (1980–2280 yr BP) can be correlated with instrumental and historical earthquakes that occurred during the last 2500 years, such as the AD 1699 Horseshoe Earthquake, the AD 1755 Lisbon Earthquake and Tsunami, and the 218 BC Earthquake and Tsunami, respectively. Older widespread Holocene events E6 (4960–5510 yr BP) and E8 (6690–6985 yr BP) can be correlated with tsunami deposits that occurred around 5310 yr BP and 6000–7000 yr BP, respectively. On the basis of synchronicity, E10 (8715–9015 yr BP) may also be regarded as a seismically triggered event. In contrast, the fact that E9 occurred immediately after the 8.2 ka climatic event suggests the possibility of a non-seismic origin for its associated turbidite layers. As regards the non-widespread Holocene turbidites E2 (AD 1908±8), E4 (855–1110 yr BP) and E7 (6110–6365 yr BP) and E9 (7880–8145 yr BP), the first two could be related to low-moderate magnitude historical earthquakes, although no further evidence is available to support this interpretation.

Widespread turbidites deposited during the Holocene suggest that they are related to great earthquakes (Mw ≥ 8.0) occurred in the SW Iberian Margin. If we regard the 6 Holocene widespread turbidite events as seismically triggered, the recurrence interval for great earthquakes is determined as approximately 1800 years. The relatively good correlation between deep-sea turbidites and instrumental and historical seismic events and tsunami deposits suggests that turbidite records may be used as a paleoseismic indicator in the slow-convergence SW Iberian Margin, thus constituting a valuable complementary tool for seismic hazard assessment in Western Europe and North Africa.
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