Characterisation of the recent BIG’95 debris flow deposit on the Ebro margin, Western Mediterranean Sea, after a variety of seismic reflection data

G. Lastras a, M. Canals a,*, R. Urgeles a, M. De Batist b, A.M. Calafat a, J.L. Casamor a

a GRC Geociències Marines, Universitat de Barcelona, Barcelona E-08028, Spain
b Renard Centre of Marine Geology, University of Gent, Gent B-9000, Belgium

Accepted 30 September 2004

Abstract

Swath bathymetry and backscatter data, side-scan sonographs, high-resolution (HR) and very high resolution (VHR) reflection seismic profiles, and sediment cores reveal the complexity of the seafloor and subseafloor features resulting from a large debris flow event that affected the Ebro continental slope and base-of-slope at the beginning of the Holocene. The BIG’95 debris flow, as it has been named, disturbed more than 2200 km², including a 26-km³ deposit of remobilised sediment covering ~2000 km². Swath bathymetry and backscatter imagery allow to distinguish four main areas within the debris flow: the source area, the proximal depositional area, the intermediate blocky depositional area and the distal depositional area. In the source area, a sinuous headwall scar, 20 km long and up to 200 m high, and several other secondary scars have been identified. Sediment released from the source area flowed southeastwards to the proximal depositional area, which is the main depocentre of the BIG’95 debris flow deposit, with accumulations over 90 m thick. Large slabs of sediment detached from this area and from the headwall scar and moved southeastwards embedded in a more mobile matrix, thus forming the intermediate blocky depositional area. Only the looser matrix reached the distal depositional area, finally freezing in and partially burying the Valencia Channel, a mid-ocean type channel. Digitisation of VHR seismic reflection profiles, where the debris flow deposit is shown as a body of mainly transparent seismic facies, has allowed the construction of isobath and isopach maps showing the thickness distribution of the deposit and its interplay with preexisting canyons and channels. DATING OF two sediment cores gives a minimum age of 11,000 calendar years BP for the BIG’95 debris flow. A set of triggering mechanisms, including seismicity and oversteepening of the slope due to the existence of a volcanic structure underneath the main headwall, is invoked.

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* Corresponding author. Tel.: +34 934021360; fax: +34 934021340.
E-mail address: miquel@natura.geo.ub.es (M. Canals).

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1. Introduction

The study of submarine landslide events and deposits has grown enormously since the pioneer studies of the Grand Banks landslide and tsunami in 1929 (Heezen and Ewing, 1952; Heezen et al., 1954). Landsliding plays an important role in shaping continental margins and in mobilising huge quantities of sediment downslope, both on active and passive margins (Hampton et al., 1996). It represents a major geo-hazard for offshore infrastructures (Campbell, 1999) and for nearby coastal areas, which could suffer severe damage by landslide-induced tsunami waves (Synloakis et al., 1997). Landsliding modifies the distribution pattern of canyon–channel systems and channel–levee complexes and thus the sedimentary architecture of continental margins. This could result in changes on the sedimentological properties of possible hydrocarbon reservoirs. As a consequence, major efforts are underway to understand the dynamics, triggering mechanisms and products of submarine slides, in parallel with the oil industry increasing exploration and exploitation of deep-sea hydrocarbon resources (Barley, 1999).

Both the economic and geo-hazard implications of submarine slump and slide events are illustrated by the Storegga Slide and by the Sissano slump and tsunami. The giant, 7200 yr BP Storegga Slide disturbed an area of about 112,500 km² on the mid-Norwegian continental margin from where about 3500 km³ of sediment were redistributed into the deep sea (Bugge, 1983; Bugge et al., 1987; Bryn et al., 2002; Evans et al., 1996). Evaluating margin stability in the Storegga region, where gas has been discovered, is one of the main targets of the soil mechanics departments in the oil companies involved. The tsunami consequences of the Storegga slide have been investigated through field work and modelling (e.g., Harbitz, 1992; Henry and Murty, 1992). The Sissano tsunami struck the north coast of Papua New Guinea on July 17th, 1998, damaging seven villages and killing more than 2000 people (Kawata et al., 1999). Although its source mechanism has been controversial, recent studies demonstrate that it was probably triggered by a relatively small (<10 km³) submarine sediment slump 25 km offshore the affected region, triggered in turn by an earthquake (Tappin et al., 1999, 2001). Many other examples can be found in the literature proving the destructive capacity of submarine landslides and related tsunamis. Two of them are the landslide-generated tsunami that devastated Valdez and Seaward, Alaska, during the great 1964 Prince William Sound earthquake (Coulter and Migliaccio, 1966), and the landslides triggered by Hurricane Camille in 1969 that severely damaged three offshore oil platforms on the Mississippi delta (Bea, 1971).

In the Western Mediterranean Basin, where our study area is located, several large mass-wasting deposits dating from the last glacial maximum onwards are known. These include a 22,000-year-old, 500-km³ large mega-turbidite in the Balearic Abyssal Plain (Rothwell et al., 1998, 2000), a 21,000-year-old debris flow on the Western Gulf of Lions (Berné et al., 1999; Canals, 1985; Canals and Got, 1986; Droz et al., 2001) and a 170-km³ large debris flow in the eastern levee of the Rhône deep-sea fan (Droz, 1983; Gaullier et al., 1998; Méar, 1984), amongst other. There are also recent failure events that occurred during the last decades. In 1954 and 1980, two earthquake-triggered turbidity currents cut phone cables offshore El Asnam, Algeria (El-Robrini et al., 1985; Heezen and Ewing, 1955). In 1979, a series of submarine landslides affected onshore structures and triggered a series of tsunamis offshore Nice before evolving into a turbidity current that broke two phone cables in the Ligurian Sea (Genneaux et al., 1980; Malinverno et al., 1988).

This paper presents the results of the study of a large data set on the largest and youngest submarine landslide on the Ebro margin, Northwestern Mediterranean Sea: the BIG’95 debris flow. It was first discovered in a survey in 1995 and, since then, it has been the target of a number of surveys on which a variety of complementary techniques have been used. This landslide is located offshore on a heavily populated coastal area near the Ebro shelf-Tarragona oil and gas fields and its deposit covers more than 2000 km² (Lastras et al., 2002, 2003). The location of this particular landslide points to the need to evaluate the potential of the Ebro margin for future geohazards.

2. Geological setting

The geodynamics of the Western Mediterranean are strongly controlled by an eastward migration of
the Apenninic arc system. The opening of the basin took place mainly during the last 30 Ma, with the generation of a set of subbasins that migrated in age from west to east (Gueguen et al., 1998). The Valencia Trough, an extensional basin lying in between the Balearic Islands and the Iberian Peninsula (Fig. 1), developed simultaneously with the Provençal and Alboran basins, during Late Oligocene–Early Miocene, and was almost completely opened at 10 Ma BP (Fernández et al., 1995; Gueguen et al., 1998). It is bounded by the Balearic Islands to the southeast, the Ibiza Promontory to the...
south and the Catalan and Ebro passive continental margins to the northwest, and it opens to the Provençal Basin northeastward. The Valencia Trough is incised along its axis by the Valencia Channel, a mid-ocean channel type submarine valley according to Canals et al. (2000) (Fig. 2).

The terrigenous and strongly subsiding Ebro continental margin displays a 70-km-wide shelf and a 10-km-narrow slope. The shelf break is located at a mean depth of 180 m (Fig. 2). Plio–Pleistocene deposits in the Ebro margin form a thick progradational sequence known as the Ebro Group. It includes the lower Ebro Clays, a prograding Pliocene clayey unit, and the upper Ebro Sands, a Pleistocene clastic shelf complex (Soler et al., 1983). This sequence overlies the Messinian unconformity (Clavell and Berastegui, 1991; Maillard et al., 1992). The outbuilding of the Ebro continental shelf and margin during the Late Quaternary was mainly controlled by glacioeustatic sea-level oscillations, subsidence and

![Swath bathymetry map of the Balearic sea from a compilation of BIG’95 and CALMAR surveys data, and data from Instituto Español de Oceanografía (IEO) and Smith and Sandwell (1997), with contours every 50 m. In dotted pattern, location of the BIG’95 debris flow deposit. Location of two seamounts (SM), the Valencia Channel (VCH), the main canyon heads (arrows), the Ebro Delta and the Columbretes Islets is also shown. Strong lines are HR and VHR seismic reflection profiles illustrated in Figs. 6–12.](image-url)

Fig. 2. Swath bathymetry map of the Balearic sea from a compilation of BIG’95 and CALMAR surveys data, and data from Instituto Español de Oceanografía (IEO) and Smith and Sandwell (1997), with contours every 50 m. In dotted pattern, location of the BIG’95 debris flow deposit. Location of two seamounts (SM), the Valencia Channel (VCH), the main canyon heads (arrows), the Ebro Delta and the Columbretes Islets is also shown. Strong lines are HR and VHR seismic reflection profiles illustrated in Figs. 6–12.
changes in sediment supply (Farran and Maldonado, 1990).

The Ebro continental slope and rise are cut by several, <15-km-long submarine canyons and gullies that occasionally reach up to the continental shelf. The base-of-slope is occupied by channel–levee complexes and interchannel areas covered by debris flow and apron deposits, which form the Ebro Turbidite Systems (Fig. 2; Alonso and Maldonado, 1990; Nelson and Maldonado, 1988). The channels south of 40°N vanish after reaching the base-of-slope while those to the north open into the Valencia Channel (Canals et al., 2000).

Several seamounts and buried volcanic edifices have been recognized in the study area by several authors using a broad spectrum of techniques (i.e., Maldonado et al., 1985; O’Connell et al., 1985; Ryan et al., 1973; Fig. 2). The most prominent are the Columbretes Islets (Fig. 2), a small volcanic archipelago in the Ebro outer shelf, which is the topographic expression of a large, mostly buried volcanic field measuring ca. 90×40 km (Maillard and Mauffret, 1993).

3. Data set

Four surveys have been carried out to study the BIG’95 debris flow (Fig. 1): the ‘BIG-95’ survey (R/V Hesperides in 1995), the ‘CALMAR’ survey (R/V L’Atalante in 1997), the ‘MATER-2’ survey (R/V Hesperides in 1999) and the ‘TTR-11 BIGIMAGES’ survey (R/V Professor Logachev in 2001).

Swath bathymetry data (Figs. 1, 2, 3 and 4) were obtained in the first three surveys using Simrad’s Merlin/Mermaid and IFREMER’s Caraibes acquisition software for the following systems: EM-12S (13 kHz, 81 beams) in 1995, EM12-Dual (13 kHz, 162 beams) in 1997, and EM-1002 (95 kHz, 60 beams) in 1999. The data acquired were first processed with Simrad’s Neptune and IFREMER’s Caraibes and later refined at the Ocean Mapping Group (University of New Brunswick, Fredericton, Canada) using SwathEd processing and mapping tools.

Seismic reflection data were obtained by means of state-of-the-art tools, including very high resolution profilers (4590 km of 3.5 kHz Simrad TOPAS PS 018 profiler, and 560 km of 3.5 kHz mud penetrator subbottom profiler) and high-resolution sleeve airguns (385 km at 0.05–0.1 kHz). In addition, ca. 850 km of air-gun and water-gun seismic reflection profiles and ca. 400 km of sparker profiles, obtained in 1979 and 1984 during the MCB-79 and GC-84-2 surveys, were supplied by ICM. The total length of seismic lines considered in this study is thus over 7500 km. The base of the debris flow can be recognized in most of the seismic reflection profiles, although they vary in penetration and resolution.

Towed deep-sea side-scan sonar imagery was obtained using TOBI (699 km of 30 kHz side-scan sonar and 7 kHz chirp subbottom profiler, covering a total area of ca. 4000 km²) and MAK-1M (38 km of Professor Logachev’s 30 kHz high-resolution side-scan sonar and 5 kHz chirp subbottom profiler, covering a total area of ca. 70 km²). Positioning was obtained with GPS ASHTECH 3DF in 1995 and 1999, GPS LORAN-C in 1997, and GPS Navstar Magnavox MX 4400 in 2001, with a position error below 1 m.

Seven piston cores obtained during the CALMAR cruise and two gravity cores during the TTR-11 cruise (Table 1), seafloor video recording and photography provide ground-truthing for the geophysical data.

4. Seafloor expression of the BIG’95 debris flow

The BIG’95 debris flow source area and deposit are located offshore the coastal town of Castelló and off the Columbretes volcanic archipelago, from 39°30’ to 40°10’N, and from 0°55’ to 1°55’E (Figs. 1 and 2). The Ebro continental slope and base-of-slope in the neighbourhood of the BIG’95 debris flow source area is cut by various canyon–channel systems and several short, steep gullies. Channel–levee complexes as those observed further north (Canals et al., 2000) are lacking in this specific margin segment of 50 km in length. The debris flow affected an area of ca. 2200 km² of the Ebro continental slope, base-of-slope and the uppermost course of the Valencia Channel, at water depths from 200 to 1800 m. The deposit itself occupies ca. 2000 km², which is four times the surface of the neighbouring Ibiza Island, in the Balearic Archipelago, opposite to the Ebro margin (Fig. 1). The BIG’95 is the largest debris flow deposit identified in the Ebro margin, compared to the other
seismically transparent units in the area, which are attributed to debris flows and, more generally, to nonchannelled unsorted deposits (Alonso et al., 1990; Field and Gardner, 1990). Within the seafloor region affected by the BIG'95 debris flow, four main segments can be distinguished based on seafloor morphology and seismic reflection profiles (Fig. 3):

4.1. Source area (and scars)

The source area extends approx. from 39°45' to 39°58'N and from 0°52' to 1°07'E (Fig. 3), at water depths in excess of 200 m, with slope gradients varying between 4° and 17°. The BIG’95 headwall scar and several secondary scars are present in this area (Figs. 4 and 5; Table 2).

The sinuous, irregular main headwall opens to the SE (Fig. 4). It lies between ~600 and 1230 m water depth, and has a total length of about 20 km. It is up to 200 m high and has a maximum slope angle of 17°. In MAK-1M sonographs, different geometries are displayed along the headwall scar (Fig. 5): a staircase geometry with 6 to 12 small steps ~10 m high in the easternmost section, a single smooth and regular slope in the central, steepest section, and a single step along its westernmost segment, where the headwall height decreases. Locally, the main headwall is partially buried by material released from secondary scars,

Fig. 3. Swath bathymetry map of the BIG’95 debris flow region. The area affected by the BIG’95 debris flow is bounded by a thick dashed line. Location of secondary scar E is shown. The different erosional–depositional areas within the debris flow (SA, source area; PDA, proximal depositional area; IDA, intermediate depositional area; DDA, distal depositional area) are limited by thick lines. In the IDA, blocks observed in bathymetry and backscattering maps are represented in grey. Also labelled are the Ebro shelf and the Balearic slope, the axis of the Valencia Channel (VCA), a seamount (SM), and the sector where the debris flow deposits displays a climbing behaviour (CL), as well as all sediment cores available. Arrows show pathways of channels affected by the debris flow. Note the rejuvenation of the slope in the source area compared to the rest of the Ebro slope. F4 and F5 boxes limit areas shown on Figs. 4 and 5.
further upslope. Degradation of the headwall itself can be observed as well (Fig. 5).

Several secondary scars in the source area, less than 10 m high and with limited lateral continuity, are partially or totally buried by the debris flow deposit and thus only visible in the very high resolution seismic reflection profiles. Two of them are, however, prominent enough to be worth describing. The first of them (B in Fig. 4) is situated upslope from the main headwall, between 800 and 1025 m water depth. It is up to 100 m high and has an irregular horseshoe shape, with both ends connecting to the main headwall. The

<table>
<thead>
<tr>
<th>Core</th>
<th>Type</th>
<th>Area</th>
<th>Location</th>
<th>Water depth (m)</th>
<th>Core length (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CLKS-01</td>
<td>Piston</td>
<td>Source area</td>
<td>39°52.50’ N, 1°04.44’ E</td>
<td>1230</td>
<td>4.58</td>
</tr>
<tr>
<td>CLKS-02</td>
<td>Piston</td>
<td>Off source area</td>
<td>39°53.30’ N, 1°06.50’ E</td>
<td>1194</td>
<td>8.40</td>
</tr>
<tr>
<td>CLKS-03</td>
<td>Piston</td>
<td>Off source area</td>
<td>39°51.74’ N, 0°58.26’ E</td>
<td>882</td>
<td>8.60</td>
</tr>
<tr>
<td>CLKS-04</td>
<td>Piston</td>
<td>Source area</td>
<td>39°50.28’ N, 1°01.86’ E</td>
<td>1127</td>
<td>7.20</td>
</tr>
<tr>
<td>CLKS-05</td>
<td>Piston</td>
<td>Proximal area</td>
<td>39°43.82’ N, 1°03.75’ E</td>
<td>1308</td>
<td>5.90</td>
</tr>
<tr>
<td>CLKS-06</td>
<td>Piston</td>
<td>Distal area</td>
<td>39°46.72’ N, 1°33.39’ E</td>
<td>1579</td>
<td>5.61</td>
</tr>
<tr>
<td>CLKS-07</td>
<td>Piston</td>
<td>Distal area</td>
<td>39°56.24’ N, 1°43.13’ E</td>
<td>1681</td>
<td>3.10</td>
</tr>
<tr>
<td>277G</td>
<td>Gravity</td>
<td>Intermediate area</td>
<td>39°46.94’ N, 1°21.17’ E</td>
<td>1475</td>
<td>1.10</td>
</tr>
<tr>
<td>278G</td>
<td>Gravity</td>
<td>Intermediate area</td>
<td>39°47.12’ N, 1°21.43’ E</td>
<td>1485</td>
<td>2.00</td>
</tr>
</tbody>
</table>
second one (C in Fig. 4) is located north of the main headwall, at about 1050 m water depth. It is up to 50 m high and has a roughly N–S-oriented bow-like shape. While material released from scar B partially buries the main headwall, material from scar C joins the BIG’95 debris flow deposit east of the seafloor headwall scarp (Fig. 4), at a water depth of 1350 m. Debris released from scar C and other less important scars (e.g., scar D) fill in the lower course of the canyon–channel system limiting this area to the north (F in Fig. 4).

4.2. Proximal depositional area

The proximal depositional area, which constitutes the principal depocentre of the BIG’95 debris flow, is immediately below the main headwall, at water depths ranging from 1200 to 1400 m (Fig. 3). In this area, the sediments mainly flowed in a SE direction.

The seafloor is relatively flat and smooth, with slopes below 1°. The meandering canyon–channel system that limits the source area to the south is truncated abruptly because of the debris flow deposit (G in Fig. 4). A ghost of this meandering channel with a much more subdued relief can still be identified on the swath bathymetry data within the area where the debris flow accumulated (Fig. 4). A secondary scar is located in this area (scar E in Table 2 and Fig. 3), towards the southern limit of the debris flow, at water depths ranging from 1350 to 1400 m. The scar displays a crescent shape opened to the SE, and although it is partially buried, indicating that it presumably formed synchronously to the rest of the slide, its seafloor expression is up to 40 m high.

4.3. Intermediate depositional area

The intermediate depositional area, comprising about 50% of the total surface occupied by the debris flow deposit, extends at water depths ranging from 1400 to 1550 m, mostly between 1°10’ and 1°30’E (Fig. 3). The uneven seafloor displays a pattern of topographically elevated blocks separated by linear depressions, which contrasts with the rest of the debris flow area and with the generally smooth Ebro margin base-of-slope. Mean slopes in this area are less than 1.2° and never exceed 4°.

<table>
<thead>
<tr>
<th>Scar</th>
<th>Water depth range (m)</th>
<th>Height (m)</th>
<th>Length (km)</th>
<th>Slope angle (°)</th>
<th>Shape</th>
<th>Class</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>600–1230</td>
<td>25–200</td>
<td>20</td>
<td>5–17</td>
<td>sinuous</td>
<td>main headwall</td>
</tr>
<tr>
<td>B</td>
<td>800–1025</td>
<td>10–100</td>
<td>10</td>
<td>11</td>
<td>horseshoe</td>
<td>secondary</td>
</tr>
<tr>
<td>C</td>
<td>1050–1150</td>
<td>10–50</td>
<td>10</td>
<td>6–9</td>
<td>bow</td>
<td>secondary</td>
</tr>
<tr>
<td>D</td>
<td>600–900</td>
<td>10–30</td>
<td>6</td>
<td>9</td>
<td>linear</td>
<td>secondary</td>
</tr>
<tr>
<td>E</td>
<td>1350–1400</td>
<td>5–40</td>
<td>15</td>
<td>1–9</td>
<td>crescent</td>
<td>secondary</td>
</tr>
</tbody>
</table>

Location of A, B, C and D in Fig. 4; location of E in Fig. 3.
Up to 200 individual blocks, mostly between 1 and 5 km$^2$ large and less than 10 m high, can be counted. However, there are also some blocks up to 25 km$^2$ and 35 m high. The surface of the blocks is topographically irregular. Elongated depressions separating the blocks are generally less than 1 km wide, but some of them are up to 3 km wide, and are flat-bottomed. Blocks are not distributed regularly in all the area, but concentrate in block clusters, each of them made of 4–5 large blocks and a number of smaller blocks. The sides of some adjacent blocks have similar geometries suggesting that they previously were attached. Larger blocks within clusters are separated by the narrower depressions, while wider depressions separate the block clusters.

Backscattering imagery (Lastras et al., 2002) shows that blocks display a relatively low backscatter compared to the overall echo-character of the BIG’95 debris flow deposit. In turn, linear depressions show high backscatter, thus appearing as sediment pathways through where coarse material flowed in a ESE direction before reaching the distal depositional area.

4.4. Distal depositional area

In water depths below 1550 m, the blocks observed in the intermediate depositional area are no longer present, and the smoother distal depositional area, approximately east of 1°30’E, extends down until ca. 1700 m (Fig. 3). In this area, with slopes under 1°, flow was forced to turn from SE to NE because of the presence of the Balearic slope, before burying the uppermost course of the Valencia Channel. In addition, two small channels proceeding from the Ebro continental slope are buried by the debris flow deposit at their junction with the Valencia Channel (Fig. 3).

Although neither blocks nor elongated depressions appear in this area, backscatter imagery shows the presence of high backscatter lineations throughout (Lastras et al., 2002), with some of these paths originating in the wider depressions between the block clusters upslope. Low backscatter patches with no topographical expression are present in between the high backscatter lineations. Some of the high backscatter material appears as plastered on the very base of the Balearic slope, east of the Valencia Channel, thus suggesting a climbing behaviour during transport and subsequent deposition in this particular area.

5. Acoustic facies of the BIG’95 deposit and underlying features

5.1. Very high resolution seismic (VHR) reflection profiles

The BIG’95 debris flow deposit appears as a huge, lens-shaped sediment body deposited on top of the Plio–Quaternary sequence of the Ebro slope and base-of-slope, thus indicating that it corresponds to the latest event on the margin. It is made of mainly transparent acoustic facies with local chaotic to hyperbolic patches, almost lacking internal reflectors. It overlies the high-frequency undisturbed continuous parallel late Pleistocene reflectors (Fig. 6). These characteristics allow an easy identification of the BIG’95 deposit along the profiles that run along or across the debris flow.

VHR seismic profiles show that the debris flow deposit is present both upslope, derived from secondary scars, and downslope from the main headwall (Fig. 6). Upslope from the main headwall, the debris flow deposit is generally less than 20 ms thick (18 m assuming a mean sound speed of $1800 \text{ m} \cdot \text{s}^{-1}$ into the sediment measured in cores). Its chaotic seismic facies turns into a more hyperbolic facies in the neighbourhood of secondary scars (Fig. 6). At the rim of the main headwall, the debris deposit becomes thinner, less than 10 ms (~9 m) thick, and shows exclusively hyperbolic facies. The stratified acoustic facies underlying the deposit become weak, discontinuous and broken. The main headwall constitutes a jump of more than 100 ms (~90 m), both in the seafloor morphology and in the underlying deposits (Fig. 6).

In the area north of the main headwall eastern termination, where sediment is released from secondary scar C, the deposit shows transparent acoustic facies while sediment underlying the deposit displays continuous parallel facies (Fig. 7). No debris deposit can be identified in the area among secondary scars C and D (Fig. 4). The underlying sediments are clearly faulted below secondary scar C, with an offset of more than 50 ms (~45 m; Fig. 7). Overall, sediment
left on the source area represents a small percentage of the total BIG’95 debris flow deposit.

The proximal depositional area is the main depocentre of the debris flow. It shows a mean total thickness over 100 ms (~90 m) as observed in VHR seismic profiles (Fig. 6). There, the flow deposit displays a transparent seismic acoustic facies with no internal reflectors or structures. The seafloor reflector is stronger and of higher amplitude than upslope from the main headwall. The base of the debris flow corresponds in this area with a single reflector (Fig. 6), that could represent a weak layer as those observed by Lykousis et al. (2002) or Lastras et al. (2004). Underlying sediment is undisturbed, continuous and stratified, but reflections are fainter than in the source area. Although the proximal depositional area represents only the 15% of the total affected area, it contains more than 40% of the transported sediment.

Within the intermediate depositional area, the deposit mostly displays transparent facies, although strong internal reflectors are identified on some profiles. Blocks and elongate depressions in this area are evidenced by abrupt changes in topography, but no changes occur in seismic acoustic facies underneath (Fig. 8). The base of the debris flow is irregular...
and does not correspond to any specific reflector. Underlying stratified sediment layers display parallel reflectors, locally truncated and less continuous in comparison with the proximal depositional area (Fig. 3B in Lastras et al., 2002).

In the distal depositional area, the debris flow deposit appears on VHR seismic profiles as a homogeneous, thin layer of transparent acoustic facies with no internal structure (Fig. 9). A N–S profile across this area shows that the northern limit of the debris flow deposit is located at deeper water depths than the southern limit, which climbs the base of the Balearic slope (cf. Section 4.4; Fig. 9).

Other lens-shaped bodies of transparent seismic facies are embedded in the Plio–Quaternary sequence of the Ebro continental slope and base-of-slope. VHR seismic reflection profiles show the presence of up to three of these bodies beneath the debris flow material released from secondary scar C (Fig. 4), thus providing evidence that this is an area prone to long term failure.

5.2. High-resolution (HR) seismic reflection profiles

The BIG’95 debris flow deposit appears in HR seismic reflection profiles as the uppermost sedimentary unit in the Ebro slope and base-of-slope. It displays a chaotic and locally hyperbolic seismic facies (Fig. 10). Because of the lower resolution of the HR seismic reflection profiles compared with VHR
profiles, identifying the debris flow deposit may pose some difficulties where it is particularly thin, e.g., in the source and distal depositional areas. Instead, it is clearly identifiable in the proximal depositional area. Overall, the debris flow deposit conformably overlies the continuous high-frequency stratified reflectors of the Plio–Quaternary sequence. In specific locations, the base of the debris flow deposit truncates some of these reflectors (Fig. 11).

Other sedimentary deposits that seem to have their origin in other landsliding events that are older than those seen on VHR lines can be observed by means of HR seismic reflection profiles (Figs. 10, 11 and 12). They are distributed regularly within the Plio–Quaternary sequence, the most prominent of them being located just overlying the reflector G, which limits the Pliocene from the Quaternary sequence (Alla et al., 1972). Overall, this landslide deposit shows the same location and spatial distribution as the BIG'95 debris flow deposit (Fig. 12), pointing again to the recurrent character of sediment failures at this specific location.

HR seismic profiles show, in addition, an acoustically chaotic, dome-like W–E-oriented structure located right under the main headwall of the BIG'95 debris flow (Figs. 10, 11 and 12). This structure is linked to the BIG'95 debris flow main headwall by a south-dipping normal fault and a marked increase in the mean gradient of the Ebro slope. It is interpreted as a volcanic dome that corresponds to the subseaﬂoor expression of the inactive Columbretes volcanic field. Similar subbottom structures have been identified in HR seismic reflection profiles at various locations along the base of the Ebro margin (Maillard and Mauffret, 1993).
Fig. 11. Along slope HR seismic reflection profile across the BIG’95 debris flow source area, main headwall and proximal depositional area. Location in Fig. 2. Note the en-echelon geometry of the base of the BIG’95 debris flow deposit, the presence of the volcanic dome and also the buried debris flow deposit overlying reflector G (black arrows in the profile, dotted line in the interpretation). Vertical scale in seconds TWTT.

Fig. 12. HR seismic reflection profile across the BIG’95 debris flow deposit from the source area until the distal depositional area. Location in Fig. 2. Note the volcanic dome, reflectors G and M, and the existence of buried debris flows in the same location and with almost the same extension as the BIG’95. Vertical scale in seconds TWTT.
6. Geometry of the BIG’95 debris flow base and deposit

Most of the VHR seismic reflection profiles have been digitised using Dynamic Graphics’ Earthvision software to obtain isobath and isopach maps. The isobath map of the base of the debris flow deposit shows how it affected the previous seafloor morphology and how it interplays with canyon–channel systems (Fig. 13). The isopach map illustrates the thickness distribution and the location of the depocentres (Fig. 14). From both isobath and isopach maps, synthetic longitudinal and transverse cross sections have been constructed showing specific features of the debris flow deposit (Fig. 15).

The base of the debris flow deposit is smoother than the present seafloor morphology, mainly due to the existence of the blocks within the deposit. It has a mean slope of 3°. Basal roughness in the source and the proximal depositional area is produced by the presence of two canyon–channel systems that have been affected by the passage of the debris flow (see above; Fig. 13). The channel bounding the deposit to the north of the source area gets buried eastward of 1°05′E. Its outline in the isobath map runs from there parallel to the border of the deposit for almost 10 km before finally disappearing at ca. 1°10′E. In turn, the channel limiting the deposit to the south of the proximal depositional area turns counterclockwise before getting buried by the mobilised sediment. Then, its outline runs meandering, while following a mean direction perpendicular to the border of the deposit. The buried channel crosses the deposit from SW to NE for more than 10 km before fading out. From swath bathymetry images, it can be inferred that there are more buried channels in the intermedi-
ate and distal depositional areas that are not represented in the isobath map due to its insufficient resolution.

The BIG’95 debris flow deposit has a mean thickness of 15 ms (~13 m). Thickness distribution (Fig. 14) is rather homogeneous in the intermediate and distal depositional areas. On the contrary, the thickness of the deposit is much more variable in the source and proximal depositional areas. The depocentre of the deposit, with sediment accumulations of up to 170 ms (~153 m), is located at the foot of the main headwall (Fig. 14). Local depocentres appear north of the main headwall, where sediment released by secondary scar C accumulates, and southeast of the main headwall, underneath a large block close to secondary scar E. Sediment accumulation decreases to an average thickness of 25 ms (~22 m) in the intermediate depositional area and less than 15 m in the distal depositional area. Locally, where the debris flow filled the uppermost course of the Valencia Channel, the deposit thickens as it fills the previous seafloor morphology. This is clearly imaged by the transverse cross sections in the distal depositional area (Fig. 15). The volume of the debris flow deposit is estimated to be at least 26 km$^3$, approx. $40 \times 10^9$ tons of sediment; the total area about 2000 km$^2$.

7. Coring and dating results

A total of seven piston cores, up to 8.6 m long, and two gravity cores have been obtained from the area (Table 1). Two piston cores were obtained from the source area, two from specific locations not affected by the debris flow off the source area, one from the proximal depositional area, two gravity cores in the intermediate depositional area and two piston cores in the distal depositional area (Fig. 3). Distinct signatures for pre-, syn- and postdebris flow deposits in cores were obtained by direct observation and core logs of density, porosity, magnetic susceptibility, sound velocity and impedance. A thin (20–120 cm) hemipelagic oxidized-muddy layer can be recognized at the top of all cores. This layer is in direct contact with the debris flow sediments, where present. The debris flow deposits in the cores consist of sandy and silty mud and sand debris, unbedded or displaying strongly
contorted beds, with clay chunks, mud chips, and other indications of flow.

The hemipelagic layer is the only postdebris flow sediment present in the cores, thus indicating the youth of the instability event. AMS $^{14}$C dating of foraminifer shells from the base of this unit in cores CLKS-01 and CLKS-06 provides a consistent minimum age of ca. 11,000 calendar years BP for the BIG’95 debris flow (Table 3; Lastras et al., 2002). Micro-paleontological analyses from samples taken from the debris flow layer on core CLKS-01 have “colder” dinoflagellates and more steppic spores when compared to samples from the hemipelagic layer. This is an indication that sediments that later were involved
in the debris flow were deposited in a colder and more arid environment.

8. Discussion

8.1. Mass transport dynamics

The formation of the BIG’95 debris flow deposit involved complex processes and mechanisms, and the entrainment of materials with contrasting rheologies. Although details of some of these processes still remain unclear, the large data set from the study area (cf. Section 3) allows constraining the overall mass transport dynamics involved in the BIG’95 event.

The main headwall of the BIG’95 debris flow formed in an area with a long history of sediment failure (cf. Sections 5.1 and 5.2, and also Section 8.2). Large quantities of sediment were released, probably favored by the presence of a weak layer, as commented above. The formation of the main headwall in turn induced instability further upslope, leading to the formation of shallower secondary scars, e.g., secondary scar B, thus illustrating retrogressive failure following the main instability event. Material released from these secondary scars partially buried the main headwall. Additional landslides, such as the events represented by secondary scars C and D, which occurred roughly at the same time as the main event since released sediment did not form individual deposits, added material to the BIG’95 debris flow. Following the formation of the main headwall, large blocks of sediment detached from further downslope at ca. 1350 m water depth, leaving behind additional secondary scars, e.g., secondary scar E. The process of formation of the different scars and the release of material produced a rejuvenation of the slope relief, truncating canyons, burying channels and destroying gullies that were present in the rest of the slope. Similar processes have been described for other submarine landslides (e.g., Driscoll et al., 2000; Prior et al., 1986).

Material released from the main headwall and from the upslope secondary scars was different to that forming the blocks (Lastras et al., 2002). We interpret the blocks as part of the debris flow deposit and not as remnants of the previous seafloor since they show transparent to chaotic acoustic facies in the VHR seismic profiles and not the continuously stratified acoustic facies observed in the in situ sediment (Fig. 8). Furthermore, the blocks display an uneven surface that contrasts with the much smoother Valencia Trough seafloor (Lastras et al., 2002). Having their origin in the proximal depositional area, the blocks would be composed of more distal, probably finer and more cohesive sediment than the coarser, looser sediment released from the main headwall and upslope secondary scars in the source area.

These large blocks moved downslope while essentially keeping their internal coherence, although disturbed enough to obliterate internal structures. During transport, original blocks broke into smaller fragments creating the block clusters described above (Fig. 3). Meanwhile, looser material from the slope moved distinctly and faster than the blocks, firstly partially burying the scars from where the blocks were

| Table 3 |
|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|
| **Location**    | **Water depth** | **Sample**      | **Core depth**  | 14C age         | Calendar age    |
| **(N)**         | **(m)**         | **Core depth**  | **Foraminifers**| **(yr BP)**     | **(yr BP)**     |
| **Lat (N) Long (E)** |                 |                 |                 |                 |                 |
| CLKS-01         | 39°52.50'       | 1°04.44'       | 1230            | 1               | 39              | G. ruber        |
|                 |                 |                 |                 | 39              | 3260±50         | 3159–2996       |
|                 |                 |                 |                 | 2               | 99              | G. ruber        |
|                 |                 |                 |                 | 8201±60         | 8841–8589       |
|                 |                 |                 |                 | 3               | 119             | G. bulloides    |
|                 |                 |                 |                 | 9950±50         | 11 124–10 979   |
|                 |                 |                 |                 | 4               | 119             | G. ruber        |
|                 |                 |                 |                 | 10 430±60       | 11 647–11 129   |
| CLKS-06         | 39°46.72'       | 1°33.39'       | 1579            | 5               | 30              | G. ruber        |
|                 |                 |                 |                 | 10 250±60       | 11 593–11 531   |

Samples 3 to 5 are located less than 5 cm above the top of the debris flow. The 14C ages have been calibrated to calendar years for 1 s ranges for the marine environment with the program Calib3.0 assuming a reservoir age of 402 yr. Location of cores is shown in Fig. 3.
detached, and then flowing through the depressions in between the blocks. The decrease in the slope gradient and the Balearic margin counter-slope slowed down the debris flow, stopping the blocks in the intermediate depositional area, after a run-out in excess of 20 km. Looser material flowed farther on reaching the distal depositional area, slightly climbing over the lowermost Balearic slope, and turning NE before burying part of the Valencia Channel after more than 100 km of run-out from the source area.

8.2. Triggering mechanisms

The presence of several acoustically transparent, lens-shaped bodies embedded in the Plio–Quaternary sequence underlying the BIG’95 debris flow deposit observed both in VHR and HR seismic reflection profiles indicates the recurrence of instability events in this segment of the Ebro margin. These bodies, although not studied in depth, often display similar spatial extension and thickness, especially that shown in Figs. 10, 11 and 12 just overlying reflector G.

Although the Ebro continental margin is a passive margin, seismic activity of small magnitude is present (Grüntthal et al., 1999). Instrumental record of earthquake activity in this particular sector is recent, with a maximum historical shock, according to the USGS/NEIC (PDE) 1973–2002 (15th April) data base, on May 15th, 1995 at 40.9°N 1.55°E with a local magnitude of 4.9. Since 1973, six earthquakes have had a 4.0 magnitude or larger. Furthermore, volcanic structures have been described in nearby areas (cf. Sections 2 and 5.2). In addition, the dome-like structure immediately beneath the scar (Fig. 10) is believed to be of volcanic origin.

A lowstand sedimentary depocentre of the paleo-Ebro River has been identified in the outermost shelf and upper slope adjacent to the Columbretes Islets (Farran and Maldonado, 1990). This could have provided significant volumes of rapidly sedimented material, which may have been characterised by a state of underconsolidation and low shear strength. During the present Holocene highstand sedimentation rates calculated from core CLKS-01, immediately upslope the main scar, vary from 8.5 to 12.2 cm kyr⁻¹.

Furthermore, the existence of the volcanic dome could have lead to an oversteepening of the slope due to differential compaction of the sediment and the volcanic rocks. Either seismicity or differential compaction could be responsible for the formation of the normal fault linking the volcanic dome with the seafloor (Fig. 9) and the rest of the faults observed (Fig. 7), which correspond to the main and secondary scars, respectively.

Gas hydrates have not been reported in the Ebro slope, but gassy sediments exist in the modern Ebro prodelta (Díaz et al., 1990), and fluid escape structures, such as pockmarks, have been observed in nearby areas (Acosta et al., 2001; Lastras et al., 2004). The possibility that such prodeltaic gassy sediments have also existed off the paleo-Ebro River mouth mentioned above cannot be ruled out. An increase in near-bottom water temperature during the glacial-Holocene transition, at about the time of the BIG’95 event, could have enhanced fluid release and subsequent shear strength lowering in the sediments. The presence of a “weak layer” favoring failure, as can be inferred from the seismic records, cannot be discarded.

However, no proven triggering mechanism exists for the BIG’95 debris flow. Apparently, the most plausible scenario appears to be a combination of some of the factors above. In this region, prone to failure due to oversteepening, overloading and maybe the presence of gas and a weak layer, a small earthquake or a slight change in water temperature could have led to sediment destabilization. The presence of other debris flow lenses in the area or just below the BIG’95 deposit proves that this region is particularly unstable.

9. Conclusions

(1) The BIG’95 debris flow extends from the upper Ebro continental slope at 200 m water depth to almost 2000 m water depth in the Valencia Channel, affecting an area of more than 2200 km². The resulting 26 km³ deposit covers an area of ca. 2000 km². The debris flow occurred ca. 11,000 calendar years BP.

(2) Four different erosional–depositional areas have been described within the debris flow. The source area includes a complex main headwall up to 200 m high, and several secondary scars with varying geometries, both upslope and downslope from the main headwall. The pro-
ximal depositional area represents the main depocentre of the deposit with accumulations over 100 ms (~90 m) thick. The intermediate depositional area displays a particular pattern of blocks and depressions that disappears in the distal depositional area, where the deposits smoothen before partially burying the uppermost Valencia Channel.

(3) The BIG’95 debris flow deposit shows mainly transparent but also chaotic to hyperbolic acoustic facies on VHR seismic reflection profiles, in contrast with the mainly stratified acoustic facies of the underlying Plio–Quaternary sequence. Other deposits similar to the BIG’95 have been identified on VHR and HR seismic reflection profiles in the region. The latter show, in addition, a dome-like structure, interpreted to be of volcanic origin, just below the main headwall.

(4) The BIG’95 debris flow truncated canyons and obliterated channels, rejuvenating the continental slope. Buried channels as well as thickness distribution among other features, can be observed in the isopach and isobath maps obtained by digitisation of the deposit.

(5) The BIG’95 debris flow is probably the result of several complex processes and mechanisms involving materials with contrasting rheologies and specific source areas. These mechanisms lead to the formation of the different erosional–depositional areas within the debris flow.

(6) Although there is no proven triggering mechanism for the debris flow, a set of factors increasing instability in the Ebro region is suggested. The existence of these factors explains the recurrence of the destabilization processes in the area.

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