New constraints on the Messinian sealevel drawdown from 3D seismic data of the Ebro Margin, western Mediterranean

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ABSTRACT

We present new 3D seismic and well data from the Ebro Margin, NW Mediterranean Sea, to shed new light on the processes that formed the Messinian Erosion Surfaces (MES) of the Valencia Trough (Mediterranean Sea). We combine these data with backstripping techniques to provide a minimum estimate of the Messinian sea level fall in the Ebro Margin, as well as coupled isostasy and river incision and transport modeling to offer new constraints on the evolution of the adjacent subaerial Ebro Basin. Four major seismic units are identified on the Cenozoic Ebro Margin, based on the seismic data, including two major prograding megasequences that are separated by a major unconformity: the MES. The 3D seismic data provide an unprecedented view of the MES and display characteristic features of subaerial incision, including a drainage network with tributaries of at least five different orders, terraces and meandering rivers. The Messinian landscape presents a characteristic stepped-like profile that allows the margin to be subdivided in three different regions roughly parallel to the coastline. No major tectonic control exists on the boundaries between these regions. The boundary between the two most distal regions marks the location of a relatively stable base level, and this is used in backstripping analysis to estimate the magnitude of sea level drop associated with the Messinian Salinity Crisis on the Ebro Margin. The MES on the Ebro Margin is dominated by a major fluvial system, that we identify here as the Messinian Ebro River. The 3D seismic data, onshore geology and modeling results indicate that the Ebro River drained the Ebro Basin well in advance of the Messinian.

INTRODUCTION

The recognition of subaerial erosional surfaces at the rims of the Mediterranean basin has been one of the first and strongest evidences supporting the hypothesis of an exceptional Messinian sea level low stand that substantially exceeded any possible glacio-eustatic change (Chumakov, 1973; Clauzón, 1973, 1978, 1982; Rizzini et al., 1978; Barber, 1981; see also a recent synthesis in Ryan, 2008). Messinian Erosional Surfaces (MES) are essentially of two types: 1) overdeepening of fluvial incisions induced by the lowering of the basin base level. These are deep V-shaped valleys cut at times into bedrock on continental areas close to the Messinian coastline, such as below the Southern Alpine lakes (Bini et al., 1978; Finckh, 1978; Rizzini & Dondi, 1978), the Var Canyon (Clauzón, 1978) and the Rhône canyon (Clauzón, 1973, 1982) in Southern France, the Aqif Canyon in Israel (Druckman et al., 1995) and the Nile Delta (Chumakov, 1973; Rizzini et al., 1978; Barber, 1981); 2) dendritic systems of Messinian erosional valleys, interpreted as ‘fluvial’, in seismic reflection profiles on modern submerged Mediterranean continental margins (Ryan & Cita, 1978), such as in the Valencia Trough (Frey-Martinez et al., 2004; Maillard & Mauffret, 2006; Maillard et al., 2006), in
the Gulf of Lions (Lofi, 2002; Lofi & Berné, 2008), in the Tyrrhenian Sea (Malinverno et al., 1981), in the Sirte Gulf and in the Levantine Mediterranean margin in general (Ryan, 1978). In addition to these, there are several evidences of erosional surfaces from outcrop analysis in Spain, Italy and Cyprus (e.g., Rouchy & Caruso, 2006). It must be noted that in many basins (Apennine foredeep, Tyrrhenian basins, Tertiary Piedmont Basin, Sicily, Eastern Mediterranean, Western Mediterranean) this erosional surface is clearly associated with an angular unconformity suggesting a link to tectonic phases besides the Messinian exceptional sea level low stand (e.g., Roveri & Manzi, 2006).

It is also now well established that several phases of erosion existed during the Messinian (e.g Escutia & Maldonado, 1992; Maillard et al., 2006; Rouchy & Caruso, 2006; Fig. 1). The MES appears as a single erosional surface where no major Messinian deposits exist, but recent work has shown that this surface separates into a Basal Erosion Surface (BES), beneath the evaporitic sequence, and a Top Erosion Surface (TES) above it (Maillard et al., 2006).

In this study we provide a high-resolution 3D image of the MES and the subsequent evolution of the Ebro Margin, in the Valencia Trough (Western Mediterranean). These images present unprecedented detail on the processes that took place on the margins of the Mediterranean Sea during the Messinian and therefore offer new evidence for the origin of the MES. The area of the MES that is imaged in this study can be located in the stratigraphic scheme of Rouchy & Caruso (2006) as indicated in Fig. 1. Therefore, over most of the study area, the data presented display a composite subaerial erosional surface that includes all Messinian phases recognized at a basin wide scale. The pinch out of Messinian marine deposits (onlap on the Basal Erosional Surface of Maillard et al., 2006) provides an indication of the position of the Messinian low stand coast line and, therefore, of minimum amount of Messinian sea level drop. The aim of this article is to provide a minimum estimate of the sea level fall in the Ebro Margin and offer new constraints on the evolution of the adjacent subaerial Ebro Basin, particularly with regard to the time when full open drainage conditions towards the Mediterranean of this formerly endorheic basin were reached.

GEOLOGICAL FRAMEWORK

The Ebro Margin is part of the northwestern boundary of the Valencia Trough. Major sediment supply to the Ebro Margin comes from the Ebro River, which drains the subaerial Ebro Basin, surrounded by the Pyrenees, Iberian Ranges and Catalan Coastal Ranges (CCR) on the Iberian Peninsula (Fig. 1). Abundant sediment supply from the Ebro River has resulted in the relatively wide continental shelf of the Ebro Margin, extending up to 80 km offshore compared to the 40 km of its northern and southern counterparts (Fig. 2). The Catalan-Balearic basin (Valencia Trough) formed on the stretched continental crust that extends between the northeastern margin of the Iberian Plate and the Balearic promontory, the northeastern extension of the Betic thrust-belt (Roca et al., 1999). The Valencia Trough evolved as the southwestern branch of the oceanic Ligure-Provençal Basin, a late Oligocene-Miocene backarc basin associated with the eastward roll-back of the west directed Apennines-Maghrebian subduction front (Doggioni et al., 1997; Gueguen et al., 1998). The associated extension affecting the western Mediterranean started in the early Oligocene in southern France, and migrated progressively southwestward, affecting the northeast Valencia Trough during the latest Oligocene–Aquitanian (Roca et al., 1999). The age of transition from a compressive to an extensional regime in the Valencia Trough is well bracketed between a latest Oligocene age for the youngest syn-compressive sediments and the Aquitanian age of the oldest synrift deposits filling the onshore half-grabens (Parcerisa et al., 2007).

Rift-related tectonic extension of the Iberian continental margin (including the Ebro Margin) was controlled by major seaward-dipping listric faults that resulted from tectonic inversion of previous Paleogene reverse faults of the CCR (Sábat et al., 1997; Roca et al., 1999; Fig. 2). The syn-rift stage resulted in the present horst-and-graben structure of the Catalan margin and it is recorded by thick (up to 2 km) Lower Miocene units restricted to the graben troughs (Fig. 2). The post-rift stage started in the Langhian and is characterized by the attenuation of the tectonic activity, grabens overfilling, and increased sediment bypass towards the Mediterranean continental shelf, leading to the progradation of the shelf and talus sediments of the Castellon Group during the Serravallian-Tortonian (Evans & Arche, 2002). Conglomerates accumulated in the onshore grabens were sourced locally from the Mesozoic and Palaeozoic basement rocks of the uplifted footwall block of the CCR.

Rifting of the NE Iberian margin was coeval with the late stages of the Ebro Basin fill (e.g., Roca et al., 1999; Per-
The Ebro Basin (Fig. 2) is the southern Pyrenean foreland basin that resulted from the collision of the Iberian and European plates since the Late Cretaceous. Connection of the Ebro Basin with the open ocean initially occurred through the Atlantic Ocean and was interrupted during the late Eocene as a result of tectonic uplift of the western Pyrenees (e.g., Plaziat, 1981; Costa et al., 2009). Since then, interruption of sediment bypass towards the oceanic domain enhanced sediment accumulation in the foreland. Continuous continental sedimentation since the late Eocene to the late Middle Miocene (Barberá et al., 2001; Pérez-Rivarés et al., 2002) resulted in the rising of the basin base level from 0 to about 1000 masl, the approximate altitude at which other closed basins of northern Spain (Calatayud, Teruel and Duero) were filled (e.g., Garcia-Castellanos et al., 2003).

The late Oligocene-Miocene last stages of the Ebro Basin infill record a westward migration of the subsidence as it is observed from the distribution of Oligocene and Miocene lacustrine depocenters in the Eastern Ebro Basin (Anadón et al., 1989). A composite thickness of about 1000 m of endorheic alluvial-lacustrine sediments in the central areas of the basin during the early-Middle Miocene (Barberá et al., 2001; Pérez-Rivarés et al., 2002) resulted in the rising of the basin base level from 0 to about 1000 masl, the approximate altitude at which other closed basins of northern Spain (Calatayud, Teruel and Duero) were filled (e.g., Garcia-Castellanos et al., 2003).

 METHODS

The present study is based on a 3D seismic survey that covers 2700 km² of the Ebro Margin, and a series of well logs, including p-wave velocity, shear wave velocity and gamma ray, that were acquired along the FORNAX-1 well (Figs 3 and 4). Seismic and well data were acquired by British Gas BV in 2002 and 2005 respectively. Seismic data is processed to near zero phase and migrated with a single pass 3D pre-stack time migration resulting in a seismic cube with horizontal grid cells of 12.5 m and a sampling interval of 4 ms. Data used in this study was subsampled by a factor of 2, resulting in seismic grid cells of 25 m. Seismic data are SEG normal polarity, i.e. an increase in impedance is a positive amplitude. The stratigraphic interval of study represents the upper 2.5–3.5 s of the seismic data. The dominant frequency content of the data varies with depth from about 70 Hz in the upper 0.7 s
two-way travel time (twtt) below seafloor to about 30 Hz from 3 to 3.5 s twtt. On average, it is approximately 50 Hz for the first 3.5 s twtt. The average vertical and lateral resolution for this interval is thus estimated to be respectively 14 m and 28 m, using an average velocity value of 2800 m s\(^{-1}\), which is estimated from interval velocities derived from p-wave velocity logs in the FORNAX-1 well. Seismic p-wave velocity in sediments immediately above the Messinian Unconformity is about 4250 m s\(^{-1}\) with a peak at 4500 m s\(^{-1}\) at the unconformity. The dominant frequency in sediments near the Messinian Unconformity is 30 Hz, giving a vertical resolution of about 35 m.

In order to investigate the nature and distribution of the Messinian Erosion Surface (MES) and the infill of its erosional relief, amplitude maps from (or near to) that surface were extracted. The most consistent results were obtained with extraction of the most negative amplitudes. To generate the MES amplitude map the interval of the seismic traces comprised in between 0.05 s twtt below and 0.1 s twtt above the MES was scanned and the trace in the time interval that had the largest negative amplitude was stored. A seismic coherency cube was also calculated, which provided a measure of lateral changes in the seismic response caused by variations in structure, stratigraphy, lithology and/or porosity.

Wireline logs, published stratigraphic reports (Lanaja et al., 1987) and correlation to published seismic data (Bertoni & Cartwright, 2005) have been used for stratigraphic and lithological analysis, correlation of the depositional units and time to depth conversion. Throughout this paper, seismic sections are displayed in two-way travel time while maps are displayed in depth (m or km). Time–depth conversion for the MES was performed using an exponential time depth function calibrated from the sonic log in the FORNAX-1 well. The time–depth function takes the form:

\[
D = 1135.13 T^{1.3643}
\]

Where \(D\) is depth below seafloor (m) and \(T\) is two-way travel time below seafloor (s). Similar relationships are found in Faust (1950), Stagg et al. (1990), Petkov (2004). Since the upper 700 m of the sonic log in the FORNAX-1 well are missing, a synthetic trace was generated so as to correctly place the upper part of the log with respect to time. To generate the synthetic trace, a wavelet was extracted from the seismic traces using the Frequency Matching method. Traces 3.5 s long from an area 500 m in radius around the well were used for wavelet generation. Using one single well to produce time–depth conversions did not produce the best result in areas relatively far from the control well, especially where large post-Messinian hiatuses exist, such as within the canyons in the continental slope.

Two dimensional backstripping of the Ebro Margin was performed with the program Flex-Decomp 2D™ (Kusznir et al., 1995). Backstripping was performed to reconstruct
the post rift evolution of the Ebro Margin taking into account sediment compaction, crustal flexure, and thermal subsidence. Input parameters include a cross section in depth, the rift age, the elastic crustal thickness, the lithospheric stretching, and the age and porosity of sedimentary packages. The cross section was constructed from a seismic section in two-way travel time extracted from the 3D seismic volume and tied to the Fornax-1 well (Fig. 4). The interpreted seismic section was converted to depth (see above) using the Fornax-1 logging information. Seven horizons were picked from the seafloor to the Mesozoic Basement (Table 1). The criteria to assign an age to each horizon are described in Table 1. In order to achieve an effective length of the modeled profile of 156 km, the topography and the stratigraphy of the seismic profile were extrapolated uniformly 60 km at the two ends of the seismic section. The model resolution is 0.5 km spacing between points.

In backstripping, thermal and flexure-induced subsidence are constrained primarily by the rift age. Another relevant parameter is represented by sediment compaction. Sediment decompaction was accomplished using the following relationship:

$$F_z = F_0 e^{-bz}$$

where $\Phi$ is porosity (dimensionless), $z$ is subbottom depth (km) and $b$ is compaction coefficient (km$^{-1}$). In each layer, porosity at the seafloor ($\Phi_0$) is 0.479 and $b = 0.42$. The above porosity relationship was determined from p-wave velocity data in the FORNAX-1 well. Velocity data was converted to density using Gardner’s et al. (1974) equation.
Hillslope diffusion is 0.1 m² yr⁻¹ exposed during the Messinian drawdown. The constant of the unconsolidated character of the deltaic sediments is 

$$\alpha$$

$$\beta$$

et al. (1992) and Garcia-Castellanos et al. (2003), though focusing here on the evolution of the delta since the beginning of its formation. In TISC, water flows following the maximum slope on a dynamic topography. For our purpose here, topography is modified only by river incision, sedimentation, and regional isostasy (fault tectonics is not incorporated). The stream power law formulation is a modified version of the undercapacity model (Beaumont et al., 1992) as modified by van der Bek & Bishop (2003). While the erodability $$K$$ and the exponents $$m = n = 1$$ are adopted from Beaumont et al. (1992) and Garcia-Castellanos et al. (2003), the characteristic length-scale for equilibrium $$L_f$$ is required to be lower than in those works ($$L_f = 4 \text{ km}$$, similar to Loget et al., 2006), probably reflecting the unconsolidated character of the deltaic sediments exposed during the Messinian drawdown. The constant of hillslope diffusion is 0.1 m² yr⁻¹. For marine sedimentation, we assume that the sediment load delivered by the river is deposited at an exponentially-decaying function of the distance to the river mouth. The constant for this exponential decay is $$L_f = 25 \text{ km}$$, which permits reproducing approximately the present extent of the sedimentary body of the Ebro Delta (e.g., García-Castelanos et al., 2003). For the isostatic calculations in TISC we use an elastic thin plate approach, with thickness ranging from 5 to 30 km (Gaspar-Escribano et al., 2004). For further details on the technique used we refer the reader to García-Castelanos et al. (2003).

**RESULTS**

**Seismic stratigraphy of the Ebro margin**

The seismic stratigraphy of the Ebro Margin is illustrated on Figs 4 and 5. Figure 4 is a crossline extracted from the 3D seismic data cube. Due to the acquisition geometry, crosslines are almost perpendicular to the margin strike (or parallel to the direction of progradation). Figure 5 is an inline, i.e. parallel to the margin strike. Overall, the margin’s seismic stratigraphy is dominated by two thick prograding megasequences (Fig. 5), in agreement with Frey-Martínez et al. (2004), Bertoni & Cartwright (2005) and Kertznus & Kneller (2009) separated by a deeply incised erosional surface. Analyzing the seismic stratigraphy in further detail shows that four major seismic units are found in the Ebro Margin. From bottom to top these are:

**Unit A**

A lower sequence of chaotic seismic facies, sometimes displaying high amplitude reflectors onlapping the acoustic basement (Figs 4 and 5), which most probably corresponds to the Lower Miocene infill of the grabens (Maufret et al., 1981; Lanaja et al., 1987; Dan/C236 obeitia et al., 1990; Fig. 4). It has a maximum thickness of 0.3 stwtt and is bounded to the bottom by the acoustic basement, an irregular surface, attributed to the Mesozoic/Paleozoic Basement. The irregular relief of the acoustic basement results in great part from horst and grabens formed during the late Oligocene to early Miocene rifting (Mau¡ret et al., 1981; Dan/C236 obeitia et al., 1990; Escutia & Maldonado, 1992) and later erosion.

**Unit B**

The lowermost prograding megasequence, Megasequence A according to the terminology of Frey-Martínez et al. (2004) and Bertoni & Cartwright (2005), corresponds to the ~1 km thick Castellon Group as identified from industry well data (Lanaja et al., 1987). It is bounded by an onlap/downlap surface on its lower boundary and a major erosional unconformity on its upper boundary (Figs 4 and 5). The latter evolves seaward into a paraconformity. Internal reflectors display clinoform geometries basinwards and a more aggrading character in the direction of sediment inputs. The toplap reflections present higher absolute amplitude than their corresponding foresets (Fig. 5). Based on the morphological character (Fig. 6) and amplitude data (Fig. 7), the erosional surface truncating the Castellon Group is believed to be the MES. This unconformity has also been drilled in numerous wells in the Ebro Margin (Lanaja et al., 1987; Fig. 3). Available age data in these wells confirms that the deeply incised erosion surface corresponds to the MES. Over some parts of the studied area, especially where relief is more subdued, the MES is not a
Continuous reflection of unique phase and polarity. It is rather a conspicuous discontinuity on which reflectors are terminated. However, the MES may also appear as a continuous negative high amplitude reflection. This occurs at the bottom of major valleys, but also in regions of relative rough relief. The MES is described in further detail in section 4.2.

Unit C

Onlapping the deepest valleys and the distal most areas of the MES (Figs 8-10) there is a relatively thin unit, maximum 100 ms twtt thick, that displays variable internal facies. In the deepest depressions of the MES, in a relatively more proximal environment, the unit presents a relatively stratified character with internal reflections displaying weaker amplitudes (Figs 4 and 5). Within these major valleys, the upper boundary presents local evidence of erosion (Fig. 8). In contrast, on the most distal part of the studied area this unit presents partially stratified to chaotic reflections and is topped by a high amplitude reflection (Fig. 9). In this area, its seismic character is quite similar to that reported for the Messinian detrital bodies of the Gulf of Lions (see Lofi et al., 2005; their Fig. 17). Where this unit presents this latter seismic character, it pinches out at a boundary where slope gradient appears to increase offshore.

Unit D

The upper prograding megasequence, Megasequence B according to Frey-Martinez et al. (2004) and Bertoni & Cartwright (2005), downlaps on the MES (Fig. 4) and it is bounded to the top by the present seafloor. Internal reflec-
tors display prograding geometries at the base of the unit and a more aggrading character at its top (Fig. 4). According to industry well data, this upper megasequence corresponds to the Ebro Group (Lanaja et al., 1987). As with Megasequence A, topset reflectors display higher amplitude than foresets. The most distal part of the deepest foresets immediately above the MES, displays a quasi-transparent character (Fig. 9), probably attesting to the fine
and homogeneous character of the sediment, resulting in a low-density contrast. Major incisions in Megasequence B are relatively young. They start to develop at the end of the Pliocene but are mostly of Pleistocene age (Fig. 5: crosslines 6100–6600). These incisions are attributed to submarine canyons such as the ones present nowadays on the Ebro continental margin (see also Kertznus & Kneller, 2009). They have a relatively chaotic/transparent infill, despite some events have been recorded as coherent reflections throughout individual canyons. Below the

Fig. 8. Inline seismic section extracted from the 3D seismic cube showing the Messinian Erosion Surface (MES) and different levels of fluvial terrace formation (C. Channel, T1-T4: terrace levels from older to younger, see also Fig. II). Also shown on this seismic line is evidence of a second phase of incision on top of the deposits onlapping the MES (yellow line). Horizontal red lines show location of time slices in Fig. 13. For legend to colors see Fig. 4. See Fig. 3 for location.

Fig. 9. Structure of Messinian Erosion Surface (MES) in the transition from Region II to Region III showing seismic character of deposits onlapping the MES. For legend to colors see Fig. 4. See Fig. 3 for location.

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Pliocene-Pleistocene boundary only minor channels develop on the margin, which present no evidences of major erosional unconformities.

The MES in the Ebro margin

Seismic geomorphology of the MES

The MES in the Ebro Margin is found between 1.25 s twtt and 2.8 s twtt (1.4 and 3.6 km deep) and is structured as a relatively complex surface. It presents three major morphological regions arranged roughly parallel to the present coastline (Figs 6 and 11). Overall, from proximal to distal areas, the MES morphology displays a flat-steep (Region I) - flat (Region II) - steep (Region III) profile in a similar way to that reported in the Nile delta (Barber, 1981), the Gulf of Lions (Lofi et al., 2005; Bache et al., 2009) or the Levant margin (Bertoni & Cartwright, 2006).

Region I

The MES most proximal region, on the northwestern sector of the seismic survey, presents a highly rugged relief with an intricate drainage network, in which v-shaped lower-order minor streams coalesce to form the high-order flat-bottomed major valleys. Tributaries of at least 5 different orders can be identified in the MES. The relief in Region I can be followed further north on 3D seismic surveys acquired by Shell and Repsol (Stampfli & Hocker, 1989; Frey-Martinez et al., 2004; Bertoni & Cartwright, 2005; see Fig. 3 for location of the block). Some authors have defined this relief as suggestive of ‘badland topography’ (Frey-Martinez et al., 2004). Within this region, elevation differences between valleys of order 3 and below and adjacent highs may attain more than 600 m. In some instances, especially in the south-western corner of the 3D seismic survey, elevation differences along the MES result from post-Messinian extensional tectonics (Figs 5 and 6). However, the major contribution to the relief along that surface results from erosional processes. It is remarkable that northwest of Region I, where the MES shows the higher elevations with respect to adjacent valleys, post-Messinian extensional tectonics is more subdued (Fig. 6).

Region II

South-east of Region I, an area of relatively low and smooth relief develops (Fig. 6). The MES on Region II dips gently to the SE and is found at 2.1 km depth at its boundary with Region I and 2.6 km depth at its boundary with Region III. Despite the lack of incisions belonging to lower order tributaries, a drainage network is also well developed as can be seen from amplitude data (Figs 7 and 11). On the depth/structure map of Fig. 6 the only incisions that can be seen to cut through that region are fourth or fifth order streams. Elevation differences between major valleys and adjacent areas within Region II do not exceed 200 m. The limit between Regions I and II is rather steep and rectilinear, but seismic data does not suggest tectonic control on that boundary (Fig. 4).

Region III

Time and depth (Fig. 6) maps suggest the presence of a third region in the southwestern most distal fringe of the 3D seismic survey, where the seafloor smoothly plunges...
basinward. The partially stratified to chaotic deposits onlapping the MES develop preferentially in that sector (Figs 9 and 10). Nevertheless, the drop in elevation of the MES within that third region cannot be correctly assessed since the change in slope also coincides with the approximate position of the Present shelf break. Therefore, time maps are strongly affected by the ‘pull down’ effect exerted by the slower travel time of sound waves within the increasingly thicker water column. Ideally, this ‘pull-down’ effect should be accounted for during time to depth conversion. However, depth maps are also affected by imprecise time-depth conversion due to thinning of the main seismic units in this region and limited well control.

All regions, but specially Regions I and II, are bisected by a major NW-SE oriented flat-floored valley, which appears to originate from the area of the Present Ebro Delta (Figs 2 and 3). Most identified streams within Region I coalesce in this flat-floored valley (Figs 6 and 11). Elevation difference from the bottom of this main valley to the topmost reflector in Unit D (Megasequence A or Castellon Group) along the strike of the different regions is ~1300 m in Region I and ~400 m for Region II. The MES along the valley floor is characterized by a relatively high amplitude reflector. This reflector is further truncated or stepped in a similar fashion to that of subaerial river terraces (Fig. 8). In the study area the MES is associated with only one major erosional event at the top of Unit B. Nevertheless, evidences of minor erosion are also identified at the upper boundary of Unit C (onlapping the MES). The polyphase character of the MES is further developed at more distal locations (Maillard et al., 2006).

Amplitude and coherency characteristics of the MES

Seismic amplitude data (Figs 7 and 12) show probably the most enlightening results in terms of mechanisms at the origin of the MES. The map of Fig. 7 shows that the three regions identified in the isochron map have slightly different minimum amplitude data. Region I presents slightly lower minimum amplitude values, which is in agreement with the fact that here the MES cuts mostly in the topsets of the Castellon Group. Most of the valleys that appear in that region, but especially those that have the lowest longitudinal gradient have low minimum amplitudes (Fig. 11). On the MES amplitude map several small streams that are not apparent on the Depth/structure map (Fig. 6) can now be traced onto the smoother relief of Region II. It can also be observed how the course of these minor Messinian streams changes from a relatively rectilinear to a more meandering pattern as they move from the rough relief of Region I to that of more even relief of Region II (Figs 7 and 11). Sinuosity of these streams changes from values ~1.1 to values close to 2 in Region II. Most Messinian streams disappear when entering Region III or their boundaries appear more diffuse. At the boundary between Regions II

Fig. 11. Interpretation of the Messinian erosion Surface (MES) based on depth structure (Fig. 6) and amplitude data (Fig. 7). Major channels are shown in grey. Terraces (T1-T5) in color are labeled. Major normal faults affecting the MES are also shown. Bold line with semicircles separates major morphologic regions. Bottom right inset shows long-profiles of the main Messinian valley and Present day Oropesa Canyon (see Fig. 2 for location). The approximate backstripped long-profile of the main Messinian Valley (paleo-Ebro River) is shown in red. Note that the Oropesa Canyon is only a proper canyon where gradients are steeper, and evolves into a mostly depositional channel-levee system were gradients are lower (Fig. 2).
and III some channels also split into several minor channels with a morphology reminiscent of a terminal distributary system (see example in Fig. 7 near x: 344 000, y: 444 500). The most striking features are however associated with the major valley dissecting the MES from NW to SE and its third/fourth order tributaries. These show a channel with well developed meanders (sinuosity ~1.3) that is characterized by low minimum amplitudes and adjacent regions of similar or even lower minimum amplitude. Contrasting the amplitude map (Figs 7 and 12) with selected inline seismic sections (Fig. 8) shows that the regions surrounding the main channel that share similar amplitude levels represent a different step in the same reflector (Figs 7, 11 and 12). The features that are identified in the MES amplitude map, including meanders (some abandoned) and zones of even minimum amplitude surrounding the main channel are recognized in different coherency time slices depending on the MES structure (see a coherency time slice example in Fig. 13). The channel has a minimum width of 300 m and a maximum width of almost 1 km, while third and lower order tributaries have channel widths not exceeding 300 m. The largest channel width occurs near the boundary between Regions II and III. The width of the valley floor (channel including the surrounding areas that display low minimum amplitudes) ranges from that of the main channel course to ~3.5 km. The channel long profile (Fig. 11, inset), extracted using a combination of elevation data from the depth/structure map and amplitude data to derive the valley thalweg, shows a relatively low mean channel gradient of ~1%.

Physical properties and nature of the MES

Well log data is available for the Ebro Margin at the For-nax-1 well. With respect to the MES that well is located on a fourth order tributary of the major valley system in the 3D seismic data survey (Figs 3 and 11). Logs of P-wave velocity, Gamma ray and electrical resistivity are available from 678 m below seafloor (bsf) down to the top of the Mesozoic/Paleozoic basement at 3233 mbsf and through the MES, which is found at 2579 mbsf (Fig. 4). The P-wave velocity log shows a gradual increase with depth from about 2550 m s\(^{-1}\) to about 4050 m s\(^{-1}\) at 2150 mbsf. This gradual increase is consistent with a muddy sedimentary column were density increases due to consolidation from increasing overburden. The following 400 mbsf show no trend with depth, and velocity appears fixed to a background value of 4050 m s\(^{-1}\) with positive excursions showing amplitudes of about 500 m s\(^{-1}\). Lower Gamma ray values at these excursions suggest coarser grained sediment. The largest excursion, both in terms of amplitude (~1000 m s\(^{-1}\) more than the background 4050 m s\(^{-1}\)) and thickness (about 50 m) is found associated with the MES. Such velocity increase may reflect coarser grained sediments and/or a densification of the sediment due to cementation. The latter might be especially the case for the MES. Suspected evidence of hard grounds and/or lag deposits on the MES was found in DSDP Leg 13 (Ryan et al., 1973). Presence of hard grounds would help explaining why the MES shows a distinct reflection over large portions of Region I.

INTERPRETATION: ORIGIN OF THE MESSINIAN UNCONFORMITY IN THE EBRO MARGIN

It is now mostly accepted that the MSC was accompanied by a drastic sea level fall of several hundred meters (Ryan 2008), despite some controversy with regard to the amplitude of that sea-level fall (Roveri et al., 2001). Here we provide additional evidence that large portions of the once submarine margins of the Western Mediterranean Sea were exposed to subaerial erosion. Indicative features of subaerial exposure of the Ebro Margin include the dendritic character of the drainage network (Figs 6, 7 and 11), fluvial terraces (Figs 7, 8, 11 and 12), meandering channels (Figs 7, 8, 11 and 12) and the low gradient of the thalweg when compared to the gradient of submarine canyons (e.g. Pirmez & Imran, 2003; see inset in Fig. 11). The first three features have also been reported on the Ebro Margin further north (Stampfl & Hocker, 1989; Frey-Martinez et al., 2004), but there, they are less developed. Most prob-
ably, the drainage network observed in these previous studies constitutes a tributary of the main valley reported here, or these valleys drained relatively small catchment areas with streams originating on the Castellon sandstones forming Region I. These streams would be equivalent to those observed north and south of the main valley course in the study area. The length, drainage hierarchy pattern and channel characteristics of the major valley on the MES in the Ebro Margin are unprecedented for the NE Iberian margin, and comparable only to valleys produced during the Messinian by major rivers such as the Rhone (Clauzon, 1982) and the Nile (Rizzini et al., 1978; Barber, 1981). The seismic data analyzed in this work suggests also that this major valley extends into the continent from a similar position to the present Ebro River mouth (Fig. 11). Therefore all evidence suggests that this major valley corresponds to the Messinian Ebro River valley.

The presence of a distinct MES reflector, not only in regions of valley infill but also on the rough relief areas of Region I (Figs 4, 5 and 8), provides evidence for pedoge-netic processes, including cementation as reported by Stampelli & Hocker (1989). A 1000 m s⁻¹ positive kick in p-wave velocity, accompanied by lower gamma ray at the MES appears to confirm the cemented and/or coarser grained character of the unconformity.

Additional evidence for the subaerial character of the MES in the Ebro Margin comes from its stepped character (Figs 4 and 6) with a flat-steep-flat-steep profile similar to the one reported by Lofi et al. (2005) in the Gulf of Lions. Further similarity between the Ebro Margin and the Gulf of Lions is the development of a detrital fan in the most distal steep region (Region III in this study, Sector I in the Gulf of Lions; Lofi et al., 2005) (Figs 9 and 10). It is to be noted, however, that Region II on the Ebro Margin (equivalent to sector II in the Gulf of Lions) is located at 2200–2500 m below present sea level, while it is only at 1600–1800 mbsl in the Gulf of Lions. Nevertheless, to properly compare the depth of the different regions in both margins, the backstripped depth should be taken into account, i.e. the depth at which these different regions
occurred during the MSC. In the Ebro Margin Region II occurred at 1300 below present sealevel (see section 6.1).

Various factors could be at the origin of this stepped character of the MES. The boundary between Regions I and II appears almost as a scar over most of its length (Fig. 6). Occurrence of that scar with a platform at its foot (Region II) is evocative of a sea cliff and marine abrasion platform. Similar scarps and platforms have been reported by Bertoni & Cartwright (2006) in the Levant margin and by Bache et al. (2009) in the Gulf of Lions. The extensive character of that platform suggests, however, that it might also result from progradation of a falling stage systems tract and continuous adjustment of the rivers’ base level, as it is also reproduced by modeling (see section 6.2 and Fig. 15; 5 Ma stage). It has also been suggested that the origin of the scarps results from preferential erosion of the sandier topsets of the Castellon Group with respect to the more clayey foresets (see for instance Frey-Martinez et al., 2004). According to our data, the basin base level did not descend much deeper than the base of the topsets in the Castellon Group. Consequently, there was little capacity for the Ebro River to erode in the lower more clayey foresets. This suggests that this step records the top of a falling stage systems tract.

Despite the ‘coastal’ look of the scarp and platform on isobath maps (Fig. 6), the basin base level appears to occur ~30 km further offshore and ~500 m deeper than the base of that scarp (Figs 6, 7 and 11), at the boundary between Regions II and III. Here, most of the streams vanish, some forming a distributary channel system before disappearing (Fig. 7). This suggests that this boundary may mark the location of a relatively stable base level in agreement with results from Bache et al. (2009). The flat bottom valley, low gradient and meandering nature of the Messinian Ebro River indicate also that the boundary between Regions II and III is probably the base level of that major Messinian fluvial system. The location of a relatively stable base level is also confirmed by the pinch out of Unit C (shallow water detrital fan) along large portions of that boundary (Fig. 9). It must be noted that many streams that are evident on the amplitude map originate on the scarp at the boundary between Regions I and II (Figs 7 and 11), while they create little erosion on Region II (and in the scarp itself). This suggests predominance of transport/deposition processes or emergence of Region II for a relatively short time span.

Over most of the seismic cube we observe evidence for only one major erosion event, here generally described as the MES. However minor erosion is also observed on top of the deposits that onlap the MES (Unit C; Fig. 8). Recent work in the Valencia Trough by Maillard et al. (2006) has shown the polyphase character of the MES. They found that the MES extends beneath the Messinian salts into the so-called Basal Erosion Surface (BES). Maillard et al. (2006) interpret that, by analogy to the MES, the BES formed under subaerial conditions. They also found a second event of erosion on top of the evaporitic sequence that they named the Top Erosion Surface (TES). Maillard et al. (2006) attribute the TES to the Lago Mare event, a period of increased precipitation and runoff (see also Bertini et al., 1998; Londeix et al., 2007), and indicate that the origin of the TES is also subaerial. We suggest that the most important erosional surface observed at the base of Unit C, the onlapping deposits, corresponds to the MES, while the erosional surface at the top of Unit C, displaying minor evidences of erosion, corresponds to the TES. In the Nile fan, 3D seismic data shows that sinuous valleys have also been carved on the topmost Messinian anhydrite (Wescott & Boucher, 2000), but in this case the TES has been interpreted to be formed under submarine conditions. In a similar way to the Ebro Margin (Figs 12 and 13), the Messinian channel systems of the Nile fan display a relatively deep, broad and generally sinuous valleys at the base of the Messinian sequence. Higher in the stratigraphic section, single valleys divide into multiple channel courses which are narrower, shallower, and less sinuous (Wescott & Boucher, 2000). In the Nile fan, these changes have been interpreted to reflect the transition from more proximal to more distal submarine channel facies. The similar succession of channel characteristics (Fig. 13) in the Ebro Margin may also reflect a similar transition.

DISCUSSION
Implications of the seismic mapping for the Messinian sealevel fall and margin backstripping

As mentioned earlier, the elevation difference from the bottom of the main valley to the topmost reflector of Unit B (Megasequence A or Castellon Group) amounts to a maximum of ~1300 m and this difference should be close to the depth of incision of the fluvial network in the Ebro Margin during the MSC. This number could also be seen as a minimum estimate for the associated sealevel drawdown. Identification of the Messinian basin base level at the boundary between Regions II and III is the key for a meaningful estimate of the Messinian sea level fall through backstripping.

Backstripping analysis was performed on a seismic section that is representative of the seismic stratigraphic architecture of the margin: it is nearly perpendicular to the slope, it is located very close to the main Messinian erosional valley and it includes the pinch out of the Messinian marine deposits (Unit C; Fig. 4).

In this backstripping exercice, the end of rifting in the Valencia Trough was considered 10 Ma. This age takes into consideration the identification of the post-Langhian post-rift subsidence, expressed in the deposition of the Serravallian-Tortonian Castellon Group and the Plio-Pleistocene Ebro Group (Negredo et al., 1999; Gaspar-Escribano et al., 2001), and the evidences of prolonged tectonic extension up to the end of the Tortonian (8 Ma; Watts & Torné, 1992) and even to the Pliocene and Quaternary (Maillard & Mauffret, 1999).
Due to the young post-rift age, the effect of varying the lithospheric stretching factor (β) and crustal elastic thickness (T) was minimum on the backstripping results. According to Negredo (1996), during the rifting β is spatially and temporally variable in the Valencia Trough and decreasing from 2.13 in the lower Miocene to 1.27 in the middle Miocene (Serravallian) for the central Valencia Trough. Maillard & Maufrret (1999) report β = 1.45 on the Catalán margin based on restoration of normal fault throw, although the thinning induced by lower crust extension is probably higher (1.62 for upper crust and 2.02 for lower crust across the Valencia Trough; Vergés & Sábat, 1999). Gaspar-Escribano et al. (2004) report values of T, that gradually change between 25 km in the Ebro Basin to 5 km in the Valencia Trough. In this work, five test models were run with variable β and T, values, which were the least constrained parameters. Different model results were compared looking at the evolution of the continental shelf morphology from the Present to the Messinian. We discarded the models in which the Pliocene and Pleistocene continental shelves became excessively deep (more than 200 m) or where the topographic profile attained landward slope. The model test that best represents a continental shelf profile evolution according to glacio-eustatic sea-level change was that with T = 15 and a variable β profile: from 1 approximately at the coastline to 3 in the deepest part of the Valencia Trough (Fig. 14).

During backstripping, the age of the MES was allowed to vary from 6.00 to 5.33 Ma (onset and end of the MSC, respectively according to Ryan, 2008) in two different models. This age difference of the MES provides an average difference of the Messinian. We discarded the models in which the Pliocene and Pleistocene continental shelves became excessively deep (more than 200 m) or where the topographic profile attained landward slope. The model test that best represents a continental shelf profile evolution according to glacio-eustatic sea-level change was that with T = 15 and a variable β profile: from 1 approximately at the coastline to 3 in the deepest part of the Valencia Trough. (Fig. 14).

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The isostatic effect resulting from removal of the load of seawater during the Messinian desiccation between 6.00 and 5.33 Ma was taken into account by backstripping in 20 kyr an uncompressible layer of density 1 g cm$^{-3}$, transport charge (close to the present-day mean Ebro value) of 285 kg s$^{-1}$ of dry sediment (Nelson, 1990) is imposed to simulate the entrance of the Ebro River. An additional limitation to take into account when comparing the results with those coming from the backstripping is that compaction has not been accounted for in this simulation. Another limitation to take into account when comparing the results with those coming from the backstripping is that compaction has not been taken into account and a constant mean density is adopted for the sediments. On a central location at the left boundary of the model domain, with elevation decreasing towards the right so that the coastline is initially defined at x = 70 km (30 km away from the left model boundary). Note that this coast is oriented NE-SW in the real Ebro Margin. Further to the right, and in order to reproduce the present depth of the Messinian unconformity as displayed in the depth-converted data, we need to impose an initial basin depth at ~3000 m, which must be regarded as the sum of the bathymetry at 8 Ma deduced from the backstripping analysis and the subsequent thermal subsidence (unaccounted for in this simulation). Another limitation to take into account when comparing the results with those coming from the backstripping is that compaction has not been accounted for in this simulation. Another limitation to take into account when comparing the results with those coming from the backstripping is that compaction has not been accounted for in this simulation. Another limitation to take into account when comparing the results with those coming from the backstripping is that compaction has not been accounted for in this simulation.

Because of the favorable tectonic environment, a subsiding passive margin without intervening tectonic phases, this figure represents a reliable quantitative constraint on the minimum extent of the sea-level drop during the Messinian. The Mediterranean sea-level likely decreased more, at times, during the deposition of the Lower Evaporites because it is known that the erosional surfaces have cut deeper than the lower evaporite pinch out that we image in our seismic data set. (e.g., Escutia & Maldonado, 1992; Maillard et al., 2006; Rouchy & Caruso, 2006).

Models of Messinian relief to discriminate the margin’s catchment area

Numerical simulations to determine drainage area presented here are inspired on the topography of the NE Iberian margin: A drainage divide at 1500 m altitude (equivalent to the paleo-CCR) is prescribed at t = −8 Ma at the left boundary of the model domain, with elevation decreasing towards the right so that the coastline is initially defined at x = 70 km (30 km away from the left model boundary). Note that this coast is oriented NE-SW in the real Ebro Margin. Further to the right, and in order to reproduce the present depth of the Messinian unconformity as displayed in the depth-converted data, we need to impose an initial basin depth at ~3000 m, which must be regarded as the sum of the bathymetry at 8 Ma deduced from the backstripping analysis and the subsequent thermal subsidence (unaccounted for in this simulation). Another limitation to take into account when comparing the results with those coming from the backstripping is that compaction has not been taken into account and a constant mean density is adopted for the sediments. On a central location at the left boundary of the model, an input flow of 500 m$^3$ s$^{-1}$ of water discharge (close to the present-day mean Ebro value) transporting 285 kg s$^{-1}$ of dry sediment (Nelson, 1990) is imposed to simulate the entrance of the Ebro River.

Topography and bathymetry evolution is calculated as a result of flexural isostatic compensation of sediment and water column changes (an elastic flexural model is adopted). We define the sea level between t = −6 and t = −5.33 Ma 1500 m below present sea level, resulting in

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subaerial exposure of the sedimentary body that formed as a product of the Ebro sediment supply. During this period of subaerial exposure, the sedimentary body is incised by a newly developing subaerial drainage network.

An outcome of the modeling is that the direction adopted by the main river within the exposed delta is mostly controlled by the lithospheric elastic thickness ($T_e$). Low values of $T_e$ imply that the exposed continental slope undergoes smaller wavelength-higher amplitude uplift, an isostatic response to the removal of the water column during the Messinian. This in turn implies a higher capability of this uplift to divert the river path. As a result, we find that for $T_e$ values lower than 10 km the main river is usually diverted and reoriented parallel to the coast, whereas for values of 20 km or higher incision along the main river copes with uplift and keeps its direction towards the right side of the model. Regional surveys based on 2-D seismic data (Farrán & Maldonado, 1990; Escutia & Maldonado, 1992; Maillard et al., 2006) allow to infer that the major Messinian valley imaged in the 3D seismic data originates slightly NE of the present delta and descends SSE and then SE. The portion of that latter trend corresponds to the segment of the valley imaged in the studied area. These gentle changes observed in the Messinian paleo-Ebro River course can thus be explained by this mechanism and confirm the selected value of $T_e \sim 15$ km used in the backstripping.

The sections displayed in Fig. 15 show horizons, the ages of which coincide with those selected for the backstripping analysis (Fig. 14). The final geometry of these sedimentary units reproduces the basic trends of the backstripping. The unit above the MES (5.33–3.1 Ma) shows a depocenter

Fig. 14. Backstripping along depth converted model of Fig. 4 to illustrate the paleo-Messinian topography and the post-Messinian stratal architecture evolution of the Ebro margin. This model’s crustal parameters are: $T_e = 15$ and a variable $\beta$ profile (from 1.5 at the NW end of the profile to 3.5 at the SE end). The age of the end of the rift is 10 Ma. See text for further details. See Table 1 for key to labels.
that is clearly shifted towards the continent relative to the
preceding unit (8–5.6 Ma) as a result of the erosion of a sig-
nificant part of the deltaic body during the low sealevel
stage.

As for the 3D geometry of the Messinian unconformity,
the model systematically predicts a much larger valley for
the main river imposed at the model boundary than for
other streams coming from the emerged areas, regardless
the values adopted for the erosional parameters. Consis-
tently with the data, the lower part of the main valley is
formed at the low Messinian water level (1300 mbsl), and
is subsequently lowered to the present depth by isostatic
response to the weight imposed by Pliocene sediments.
The relief across the valley is also satisfactorily reproduced
(see section in Fig. 16). The width of the valley is underes-
timated in our models relative to the observations. This
could be improved by using a lower $L_f$ value, but this in
turn would induce a smaller relief across the valley. Using
the exact formulation and erodability values of Garcia-
Castellanos et al. (2003), leads to even narrower valleys,
inconsistent with our observations. This might result from
the less consolidated nature of the sedimentary infill of the
Ebro margin relative to the Ebro Basin (Garcia-Castellanos
et al., 2003), but also from short-wavelength hillslope
transport processes that are very simplified in our model
(Braun, 2006). Concerning the width of the channel, we
adopted a function of water discharge as in the version of
the undercapacity model by van der Beek & Bishop
(2003). The channel widths predicted in the model can be
used to discriminate between a river of comparable size to
the Ebro River and smaller rivers that originate on the
CCR, assuming that precipitation rate was similar in both
catchments during the Messinian and at present.

The results obtained from the simulations show that, as
a consequence of river incision during the sea level draw-
down, the $-2500$ m contour line penetrates about 40 km
into the delta body, in agreement with what is observed in
the seismic cube. This is only attained by the main river,
whereas smaller rivers descending from the CCR do not
significantly alter that contour line. Similarly, the width of
the main valley, though underestimated in our models, is
much closer to the Ebro valley width of $\sim 20$ km observed
in the seismic cube than the width of valleys originating in
the CCR. Formulations different from the described un-
dercapacity model have been tested (e.g., Davy & Crave,
2000; Garcia-Castellanos et al., 2003) obtaining similar re-
sults but producing a narrower valley along the main river.
All formulations, however, predict a significantly larger
canyon (in width and depth) along the main river in com-
parison with rivers draining only the topographic barrier
defined at the left side.

Evidence for a pre-Messinian Ebro river

A hypothesis based on geomorphological and tectonostra-
tigraphic relationships and modeling was developed by
Ribé et al. (1983), Conen et al. (1996) and Babault et al.
(2006) suggesting that the Ebro Basin did not connect to

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Messinian sealevel drawdown from 3D seismic data
the Mediterranean prior to the MSC, and therefore no major fluvial course drained the subaerial Ebro Basin into the nowadays Ebro Margin prior to the MSC. In contrast, the seismic data analyzed in this work demonstrate the presence during the Messinian of a river that drained a large basin beyond the CCR drainage divide, indicating that the Ebro River was at least partially formed at that time. The question that arises is therefore whether the Messinian drawdown, spanning at most 260 kyr (Krijgsman et al., 1999) provided time enough to produce a drainage capture (piracy) of the Ebro Basin by a relatively small river originating in the CCR and then carve the valley of similar characteristics to that observed in the 3D seismic data and attain nearly equilibrium conditions as suggested by the meandering nature and low gradient of the channel (Hickin, 2003).

Onshore data provides some constraints on capture of the Ebro Basin by the paleo-Ebro River. Middle Miocene lacustrine sediments occur over vast regions in the central Ebro Basin indicating endorheic conditions up to, at least, 13.5 Ma based on magnetostratigraphy (Pérez-Rivarés et al., 2002). Even younger lacustrine units (upper Aragonian, 12 Ma, uppermost Serravallian) are still preserved along its southwestern margin (Azanza, 1986). These sediments are overlain by a 60 m thick carbonate unit, which represents the youngest known sediments filling the Ebro Basin, today found at an elevation of 750 masl. These sediments are interpreted as of fluvio-lacustrine origin and, in contrast to underlying lacustrine sediments, suggested to represent an open hydrologic system (Vázquez-Urbez et al., 2002). A late Miocene age was loosely attributed to this top carbonate unit (Pérez et al., 1989), but no direct age constrain can be strictly assigned to it.

The Llobregat River (100 km NE of the Ebro River, Fig. 2) drains the NE portion of the former Ebro Basin. Capture of this region into the catchment area of a paleo-Llobregat started as early as late Miocene, as evidenced by clast composition of late Miocene alluvial conglomerates filling the grabens in the CCR (Valles-Penedes basin). However, despite this early incision, capture of the base level of the Central Ebro Miocene lake by the Llobregat River could not be achieved because of the active rift-shoulder uplift of the northeastern Ebro basin and concomitant south-westward migration of the basin depocenters from late Oligocene onwards. Consequently, the final capture of the Central Ebro lake basin was accomplished by incision of the Ebro River across the southwestern CCR, where uplift rates were lesser.

Fig. 16. Final stage of the drainage model evolution 0 Ma (present). a) Topography, drainage network (thickness of line indicating river is proportional to water discharge), and total sediment thickness (contours every 0.5 km); b) depth of the Messinian unconformity (contours every 500 m); c) and d) sections across the profiles located in A and B (red lines).
(Gaspar-Escribano et al., 2004). Given the similar competence of basement rocks that the Llobregat and Ebro rivers had to cut through in the CCR it is expected that capture of the Ebro Basin base level by the Ebro River did not occur much later.

Offshore, larger sediment supply from the Ebro River to the Ebro Margin, compared to its northern Catalan Margin and Southern Valencia margin, is suggested by the protruding continental shelf extending up to ~80 km from the present coastline (in comparison to the ~40 km of continental shelf extension in the Catalan and Southern Valencia margins; Fig. 2). Above the MES the prograding character of the Ebro Group attests to a sustained but variable (Farrán & Maldonado, 1990; Nelson, 1990) Ebro River supply during the Plio-Quaternary. As discussed earlier, the large fluvial system that carved the MES indicates that the Ebro River already reached open Mediterranean waters during the Messinian (Figs 4–12). Clioform geometries in Unit B (lowermost prograding megasequence or Castellon Group), deposited during the Serravallian-Tortonian time interval, suggest that the pre-Messinian Ebro Margin prograded by a similar amount to the Plio-Pleistocene margin (see also Bartrina et al., 1992; Roca, 2001; Evans & Arche, 2002). Numerical simulations of river transport and drainage evolution (Figs 15 and 16; see previous section) indicate that a progradation remarkably larger than those produced by smaller streams sourced in the CCR is only possible under the presence of a larger river such as the present Ebro River (see margin progradation off the main river compared to small rivers draining the mountain range at the left boundary of the model in Fig. 15). This indicates that the Ebro River was already draining a significant part of the Ebro Basin towards the Mediterranean, at least since the Tortonian. To seize the drainage area associated to the observed channel width of the Messinian Ebro River valley, channel and drainage basin characteristics are compared with the Present Ebro and Francoli (a nearby river draining the CCR; Fig. 2) in Table 2. The channel width of the Messinian Ebro River valley is within the range of bankful width of the present Ebro River. Using empirical relationships between channel width and water discharge or drainage area obtained from river transport studies in various regions (Smith et al., 1995; Attal et al., 2008; Ames et al., 2009) shows that the ~1000 m-wide channel near the supposed river base level, as observed in the seismic cube, is closer to what would be expected for a catchment with a size closer to the Ebro Basin than to a stream draining the CCR with a catchment area one order of magnitude smaller (Table 2). The slightly broader channel might result from: 1) the larger drainage basin after incorporating the continental shelf and slope due to the sealevel fall, 2) higher runoff due to the more humid conditions that are supposed at the end of the Messinian period (Lago Mare event; Bertini, 2006; Longeix et al., 2007), and 3) the probably higher erodability of unconsolidated marine sands and muds of the Castellon Group. This confirms, together with the simulations shown earlier in this paper, that the Ebro River already drained a significant part of the Ebro Basin during the Messinian.

In this sense, it has also been pointed out that the rate of sedimentation increased by a larger factor at the base of the Serravallian-Tortonian Castellon Unit than at the base of the Plio-Pleistocene Ebro Group (Daño beitia et al., 1990; Martinez del Olmo, 1996). Whereas the Pliocene increase has been attributed to a global increase in sedimentation rates (related to onset of glaciation in the Northern hemisphere; Peizhen et al., 2001), the increase at the base of the Castellon Group can only be the result of a sudden increase in sediment supply from the Ebro Basin. Thus, the results question previous interpretations according to which the internal (endorheic) drainage of the Ebro Basin was captured by a Mediterranean stream only during or after and as a result of the MSC drawdown (Coney et al., 1996; Babault et al., 2006).

The question remaining is why the observed incision did not propagate upstream as far as in other Mediterranean rivers. In other large rivers along the Mediterranean coast this incision is observed today as a thick body of post-Messinian sediments filling a wide Messinian valley that extends 300 (Rhone) to 1000 (Nile) km upstream into the continent (Chumakov, 1973; Clauzon, 1973). Clearly, there is no evidence of a Messinian incision filled with Pliocene deposits where the river presently leaves the Ebro Basin and crosses the CCR. Our calculations and those from Garcia-Castellanos et al. (2003) and Loget et al. (2006) indicate that incision along a large river during the Messinian would propagate upstream into the Ebro sedimentary basin. Along the Ebro, though, the incision preserved today propagated at most ~30 km from the present coastline (~80 km from the Messinian lowstand sealevel – coast). Different reasons may explain this. First, despite being similar in catchment area, the Ebro River has a mean water discharge that is at least 4 times smaller than that of the Rhone. Following Gladstone et al. (2007),

Table 2. Comparison between the present channel width near the mouth of the Ebro River, the Francoli River (chosen as representative of the main rivers draining the CCR), and the valley observed in the seismic data.

<table>
<thead>
<tr>
<th></th>
<th>Present Ebro</th>
<th>Present Francoli</th>
<th>Messinian valley</th>
</tr>
</thead>
<tbody>
<tr>
<td>Channel width near the mouth – m</td>
<td>220–800</td>
<td>55–65</td>
<td>800–1000</td>
</tr>
<tr>
<td>Drainage area A – km²</td>
<td>84 000</td>
<td>838</td>
<td>45 100–65 400 (1)</td>
</tr>
<tr>
<td>Discharge Q (mean/historical maximum) – m³ s⁻¹</td>
<td>430/23 400</td>
<td>1.2/225</td>
<td>30 000–47 200 (2)</td>
</tr>
</tbody>
</table>

Calculated using \( W = 1.29 \times 10^{6} \) (A in km²; Ames et al., 2008) from observed W.
Calculated using \( W = 4.6 \times 10^{6} \) (Q in m³ s⁻¹; Attal et al., 2008) from observed W.
this difference may have been nearly double right before the Messinian, due to different climatic regimes. Second, the Ebro River crosses a structural barrier in the CCR (horst), close to its mouth, while the courses of the Nile and Rhone Rivers run mostly parallel to the main regional structures (Adamson & Williams, 1980; Doglioni et al., 1997). The lithology incised in the CCR (mainly Mesozoic limestones; Fig. 1) is much harder than the sediments filling the Ebro Basin. This could have provided a lithological knockpoint that prevented or slowed down incision further upstream. Third, preferential, fast erosion of the soft material within the Ebro Basin provided the Ebro River with excess sediment load when crossing the CCR, thus reducing incision capacity (i.e., transport-limited incision). Forth, post-Messinian erosional rebound related to erosion of the Ebro Basin and the CCR, and to thermal processes in the Valencia Trough amounts probably more than 500 m of uplift (Negredo et al., 1999; Gaspar-Escribano et al., 2004), thus contributing to partial exhumation of a potential Messinian paleovalley across the CCR. Fifth, the drainage area of the Ebro River upward from the CCR might have been significantly smaller at that time, since the capture was relatively recent. Evidences exist for Quaternary and late Pliocene-Pleistocene Ebro Group) a large erosional surface that this major valley corresponds to the Messinian Ebro River, which should have already attained a drainage area comparable to the present one. Constraints provided by the data set, onshore geology and coupled isostasy and river transport and drainage evolution modeling techniques indicate that drainage of the Ebro River into the Mediterranean Sea occurred prior to the Messinian, most probably during the Serravallian/Tortonian, which is the age of the first significant >1-km thick siliciclastic deltaic megasequence deposited in the Ebro Margin.

CONCLUSIONS

3D seismic and well data offer an unprecedented view of the Messinian Erosional Surfaces (MES) on a large area of the Ebro Margin in front of the Ebro Delta. The data set, in combination with backstripping and coupled isostasy and river transport and drainage evolution modeling techniques, allow to draw the following conclusions:

The Cenozoic Ebro Margin seismic stratigraphy is composed of four main units, including two major prograding megasequences. In between these two megasequences (Serravallian-Tortonian Castellón Group and Pliocene-Pleistocene Ebro Group) a large erosional surface develops that is attributed in this study to the MES. Offshore, the MES evolves into a paraconformity and is onlapped by shallow water detrital bodies.

The MES paleorelief can be separated in three major physiographic regions that are roughly parallel to the Present Day coastline and are dissected by a fluvial subaerial network with tributaries of at least five different orders that cross-cut the three regions almost perpendicularly. These three regions shape a stepped-like profile for the Ebro Messinian margin. No apparent tectonic control exists at the boundaries among the different regions. The boundary between the two most distal regions, where most streams disappear is believed to show the limit of a relatively stable base level.

Identification of this stable base level and backstripping modeling techniques allow to estimate a Messinian scale-
Messinian sealevel drawdown from 3D seismic data


Messinian sealevel drawdown from 3D seismic data


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