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Submarine mass movements around the Iberian Peninsula. The building of continental margins through hazardous processes

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ABSTRACT

Submarine mass movements, such as those which occur in all environments in every ocean of the world, are widely distributed across the Iberian continental margins. A lack of consistent data from various areas around the Iberian Peninsula makes it difficult to precisely understand their role in the sedimentary record. However, all the studies carried out over the past two decades reveal that they are a recurrent and widespread sedimentary process that may represent a significant geohazard. The majority of submarine mass movements observed in both the Mediterranean and Atlantic margins of the Iberian Peninsula have been generically identified as Mass Transport Deposits, but debris flows, slides, slumps and turbidites are common. Only a few remarkable examples involve huge volumes of sediment covering large areas (such as ~500 km^3 and ~6x10^4 km^2), but more moderate deposits (<200 km^2) are frequently found on the seafloor or embedded in the sedimentary sequences, building margins and basins.

Key words: continental margin, geohazard, Iberia, sedimentary architecture, submarine mass movement.

Inestabilidades sedimentarias submarinas alrededor de la Península Ibérica. Construcción de márgenes a través de procesos peligrosos

RESUMEN

Las inestabilidades sedimentarias submarinas, como en todos los océanos del mundo, están ampliamente presentes en los márgenes continentales ibéricos. La disposición irregular o la falta de datos adquiridos alrededor de la Península Ibérica hace difícil tener un conocimiento preciso acerca del papel de los movimientos en masa en la evolución del registro sedimentario submarino. Sin embargo, todos los estudios realizados en las últimas décadas muestran que son un proceso sedimentario esencial que puede representar un riesgo geológico importante. La mayoría de los depósitos observados en ambos márgenes, Mediterráneo y Atlántico, han sido genéricamente definidos como de transporte de masa, pero flujos de derrubios y deslizamientos (rotacionales y translacionales) o turbiditas son comunes. Algunos ejemplos notables involucran grandes volúmenes de sedimentos que afectan grandes áreas (~ 500 km^3 / ~ 6x10^4 km^2) pero depósitos más moderados (<200 km^2) están ampliamente presentes afectando la superficie del fondo del mar o formando parte de las secuencias sedimentarias que constituyen los márgenes y cuencas.

Palabras clave: arquitectura sedimentaria, Iberia, margen continental, movimiento de masa submarino, riesgo geológico.

VERSIÓN ABREVIADA EN CASTELLANO

Inestabilidades sedimentarias submarinas. Procesos y factores de control

Los movimientos en masa representan un proceso sedimentario importante en la evolución de los márgenes continentales, y están presentes en todos los contextos geológicos y océanos del mundo. Se han descrito en todos los ambientes fisiográficos presentando tamaños que varían desde pocos metros a varios quilómetros. Las inestabilidades sedimentarias submarinas presentan características parecidas a sus equivalentes subaéreos, con algunas excepciones exclusivas de medios acuáticos como las corrientes de turbidez, que pueden afectar áreas enormes. El deslizamiento Storegga, con 95000 km^2 (Haflidason et al., 2004), o los flujos de derrubios en las Islas Canarias y margen Sahariano afectando a más de 600 km (Masson et al., 1997) son dos ejemplos de ello. A pesar de su importancia, excepto en zonas costeras, estos procesos nunca se han observado directamente y su conocimiento se basa en el estudio de los productos sedimentarios resultantes. También se han detectado debido a los daños que han provocado sobre infraestructuras submarinas. Este hecho sumado a su potencial capacidad para generar tsunamis o maremotos certifica que estos procesos representan un riesgo geológico tanto en áreas someras como profundas (Gisler et al., 2006; Harbitz et al., 2013).

Existen diferentes tipos de inestabilidad sedimentaria (Locat y Lee, 2000). La clasificación de estos procesos es compleja y puede estar ligada a diferentes criterios como la reología, el mecanismo de soporte o transporte de sedimento, su concentración, etc. (Mutti and Ricci Lucchi, 1975; Mulder and Cochonat, 1996; Locat and Lee, 2000; Shanmugam, 2000; Mulder and Alexander, 2001; Gani, 2004; Masson et al., 2006). Los términos movimiento en masa, transporte en masa, inestabilidad sedimentaria o proceso gravitativo, ampliamente utilizados en este texto, se consideran sinónimos e incluyen indistintamente todos los tipos existentes. Los deslizamientos y la compleja "familia" de los flujos se encuentran entre los tipos de inestabilidades más comunes e importantes del medio marino (Tabla 1).

El estudio de los movimientos en masa submarinos se basa en el análisis de datos sísmicos y acústicos que ofrecen observaciones indirectas del marco tectónico-sedimentario así como de las características de los eventos (geometría, morfología etc.). Pero una compresión global requiere además la integración de datos sedimentológicos, geotécnicos y de propiedades físicas de los sedimentos involucrados. La generación y ocurrencia de una inestabilidad sedimentaria es un problema multivariable expresado como un complejo equilibrio entre fuerzas de resistencia y los esfuerzos aplicados (Hampton et al., 1996; Leynaud et al., 2004; Mulder et al., 2009). El conocimiento de propiedades físicas y geotécnicas de la columna sedimentaria, obtenidas mediante diferentes test de laboratorio sobre muestras de sedimento es crítica para la caracterización del equilibrio existente (Fig. 1).

Muchos factores se han identificado como precursores de inestabilidades submarinas. Estos incluyen los procesos que operan a escalas de tiempo de minutos (terremotos) y procesos geológicos que operan en escalas de tiempo de decenas o cientos de miles de años, como el cambio climático, cambios en la sedimentación, etc. Los principales factores identificados son: altas tasas de sedimentación, presencia de gas o hidratos de gas en el sedimento, erosión, actividad tectónica, terremotos, olas de tormenta, actividad volcánica y la actividad antrópica. Por lo tanto, diferentes sedimentos pueden ser propensos a la inestabilidad dependiendo de su composición, geometría y en última instancia su ubicación. Por ejemplo si forman parte de depósitos que puedan experimentar una disminución de su resistencia a la cizalla (sub-consolidación, exceso de presión intersticial, etc.) y/o están sometidos a procesos que aumentan el esfuerzo aplicado (carga cíclica de terremotos, por olas de tormentas, etc.).

La distribución de todos los factores mencionados puede explicar la distribución de movimientos en masa en el registro sedimentario de un margen continental. Por eso, a pesar de que las inestabilidades son procesos asociados a casi todos los ambientes, éstos ocurren comúnmente en algunos entornos específicos como fiordos, deltas, sistemas de cañón-abanico, taludes continentales e islas volcánicas.

Inestabilidades submarinas alrededor de la Península Ibérica

Es difícil tener una visión global sobre el papel de los procesos de movimiento en masa en los márgenes ibéricos y sus alrededores (Fig. 2). Esto se debe a la falta de estudios uniformes o una cobertura adecuada que permitan configurar una visión completa en todas las áreas. Existen pocos intentos de realizar inventarios de inestabilidades submarinas y se centran en compilaciones bibliográficas parciales en áreas como los márgenes continentales Mediterráneos, (Camerlenghi et al., 2010; Urgeles and Camerlenghi, 2013). A pesar de ello, una gran cantidad de estudios demuestran que las inestabilidades submarinas son muy abundantes en todos los márgenes ibéricos (Baraza et al, 1990; 1992; Casas et al., 2003; Droz et al, 2006; Urgeles et al., 2006; Lastras et al., 2007; Camerlengi et al., 2009; Alonso et al., 2014; entre otros). Algunos de estos estudios se centran en

las características morfológicas y dinámicas de las inestabilidades mientras que otros lo hacen en su papel en la construcción de márgenes y cuencas. Algunos ejemplos implican grandes volúmenes de sedimentos afectando grandes áreas como la megaturbidita depositada en la Llanura Abisal de Baleares con un volumen de ~ 500 km^3 y un área ~ 6x10^4 km^2 (Fig. 2; Rothwell et al., 1998). Pero un gran número de depósitos más moderados (<200 km^2) están presentes en todos los dominios fisiográficos, afectando a la superficie del fondo del mar o integrados en las secuencias sedimentarias que construyen márgenes y cuencas. La mayoría han sido genéricamente identificados como depósitos de transporte de masa, aunque flujos de derrubios y deslizamientos son muy comunes.

Un fuerte vínculo entre la actividad tectónica e inestabilidades se ha establecido tanto en los márgenes Atlánticos como Mediterráneos. En el Mar de Alborán se han observado una gran cantidad de depósitos de transporte de masa formando parte de la secuencia Plio-cuaternaria (Fig. 3). Estos depósitos se han relacionado directamente con la actividad tectónica cíclica post-Mesiniense, que tiene un importante papel tanto en la configuración fisiográfica como en la arquitectura sedimentaria del área (Vázquez et al., 2013; Alonso et al., 2014). La actividad tectónica también tiene un papel predominante en los márgenes de Galicia y Cantábrico (Figs. 4 y 5). Por otro lado, la actividad de diferentes fallas localizadas en el Golfo de Cádiz y margen S de Portugal (Fig. 2) imprime un importante riesgo geológico sobre las costas de Marruecos, España y Portugal. Esto se debe a que la actividad es suficientemente importante como para detonar no solo inestabilidades de gran entidad sino también tsunamis (Bartolomé et al., 2012). La presencia de fallas y procesos gravitativos también se ha demostrado que juegan un papel interdependiente en la evolución de algunos cañones submarinos (Ercilla et al., 2008b; Pérez-Hernández et al., 2009; Sayago-Gil et al., 2008).

Por otro lado, están las áreas donde la actividad tectónica, aunque no se puede despreciar, no juega un papel importante y se registra una baja sismicidad. En estas áreas las inestabilidades estarán determinadas por otros factores como la compactación diferencial, altas tasas de sedimentación o presencia de gas libre en los sedimentos (Fig. 5). En muchos casos son áreas alimentadas por importantes aportes fluviales (Casas et al., 2003a; Urgeles et al., 2006). También se han observado procesos de inestabilidad asociados a sistemas deposicionales contorníticos (Fig. 6), relacionados tanto a los procesos erosivos producidos por las corrientes de fondo como a la configuración y características de los depósitos (Larberg and Camerlenghi, 2008).

Aunque existe un escaso control de la edad de la mayoría de los movimientos en masa observados en los márgenes de la Península Ibérica, un gran número de ellos se han producido durante el Holoceno (Urgeles and Camerlenghi, 2013). Esto sugiere que además de la actividad tectónica, también existen esfuerzos inducidos por cambios climáticos, es decir cambios en el nivel del mar, en la temperatura del fondo o en la carga sedimentaria, que han tenido un importante efecto en la generación de estas inestabilidades sedimentarias.

Introduction: The significance of submarine mass movements

Submarine mass movements play an important role in the evolution of continental margins as they represent an efficient mechanism of sediment transport from coastal to deep-sea areas. Mass movements occur in all the oceans of the world, and may develop in any physiographic environment, although they commonly occur in areas with thick sedimentary deposits, sloping seafloors and high environmental stresses (Hampton *et al.,* 1996). Submarine mass movements range greatly in size from metre-scale to many kilometres across. According to the observations collected so far, submarine mass movements have similar characteristics to the onshore sedimentary instabilities, with some important exceptions, such as turbidity current flows, which are exclusive to aquatic environments. They can also be much larger, affecting huge areas of the seafloor. A good example of this is the Storegga Slide, which impacts an area of 95 000 km^2 (Canals *et al.*, 2004; Haflidason *et al.*, 2004). Other notable examples are the debris flows in the Canary Islands and Sahara margins with a runout distance of over 600 km (Masson *et al.*, 1997).

In spite of the importance of mass movements in submarine environments their occurrence has never been directly observed except along the coastlines (e.g., Longva *et al.*, 2003; Mulder, 2011) and our knowledge of them is based on observation of the resulting sedimentary products. They have also been detected due to damage to infrastructure resting on or fixed to the seafloor/sub-bottom, such as cables and pipelines. Mass movement processes in marine environments, both in shallow and deep sea areas represent a major geohazard due to their destructive and tsunami-generating potential. Landslide-generated tsunamis also deserve greater attention when evaluating the hazard posed to coastal areas (e.g., Fine *et al.*, 2005; Gisler *et al.*, 2006; Harbitz *et al.*, 2013).

The destructive power of mass movements greatly depends on the location and size of the instability. A statistical regional-scale solution can be developed to determine the probability of a landslide of a particular size, although the statistics to be applied have still to be established (Dussauge *et al.*, 2003; Ten Brink *et al.*, 2006; Casas *et al.*, 2012). The possible solution has generated an interesting controversy because it depends on the landslide model, i.e. if mass movements result from a deterministic model or are contrary to the self-organised criticality model (Turcotte, 1996; Guzzetti et *al.,* 2002; Guthrie and Evans, 2004).

The study of submarine mass movements starts with the analysis of seismic data which offer indirect observations of the tectono-sedimentary framework within which the indicators of mass movements are found, enabling a definition of the slide plane, internal pattern, scale of failure, geometry, runout distances, and so on. However similar sediments may behave differently with respect to slope instability, depending on different petro-physical parameters and stress states. To fully understand the process the geotechnical data, *in situ* or from sediment cores, must be integrated with the morphologic and sedimentary observations.

The aim of this paper is to present the current knowledge on submarine mass movements and their role on the sedimentation of continental margins. An overview of the most significant mass failure processes in deep sea areas around Iberia is also offered.

Variability of marine sedimentary instability processes and deposits

Submarine mass movements, landslides, mass wasting, mass transport and gravity processes, terms widely used herein, are generally synonymous terms that include all types of sedimentary instability processes.

Sometimes, slopes reflect the development of a failure before it is triggered. Slow slope deformations, such as creep, can be a signal that deformation may eventually accelerate to failure. Creep is the long-term deformation of sediments subjected to a constant load on a gentle slope (Nardin *et al.*, 1979). Once the instability is initiated, the process can be classified according to its mechanical behaviour (rheology), particle-support mechanism, concentration and longitudinal changes of the deposits (Mutti and Ricci Lucchi, 1975; Mulder and Cochonat, 1996; Locat and Lee, 2000; Shanmugam, 2000; Mulder and Alexander, 2001; Gani, 2004; Masson *et al.*, 2006). However, it should be noted that the classification of submarine mass movements is very complex, as: 1) it is difficult to observe and monitor instability events, so that their analysis relies principally on the final morphologies of the related features; and 2) most of the terminology applied to submarine mass movements is inherited from that used for subaerial mass movements (e.g., Locat and Lee, 2000; Lee *et al.*, 2009; Hungr *et al.*, 2014), although there are differences between the two processes. The most important types of mass transport processes observed in the marine environment are sediment failures and the extensive and complex "family" of flows (Table 1).

Sediment failures are the movement of sediment or rock along a shear plane with relatively low shear resistance. They can be divided into slumps and slides: rotational movements are called slumps while translational movement defines a slide. Sediment failures usually form complex structures such as retrogressive failures associated with multiple phases of movement that propagate the failure upslope (e.g., Prior and Suhayada, 1979; Mulder and Cochonat, 1996). Sometimes considered to be a variant of slides (Hutchinson, 1988), spreads indicate the movement of sediment or blocks of consolidated sediment, in very low slopes, due to the presence of a weak layer or liquefying underlying material, and not a basal shear plane. Liquefied flows result from the destruction of the sediment fabric due to an increase in interstitial pressure and displacement of interstitial fluid (Nardin *et al.*, 1979). The term liquefication (the transformation of a solid into a liquid) includes *fluidization* when the transformation results from pore fluid movement and *liquefaction* when it is caused by grain agitation during cyclic shear stress across sands and silty sands or sensitive clays, which have the tendency to change from a relatively stiff condition to a nearly liquid mass if disturbed (Sultan *et al.*, 2003).

A wide range of flow types can occur as a result of the interplay of rheology, grain size composition, concentration and consequently the particle support mechanism (Table 1). Flows in general have viscoplastic behaviour and the amount of clay or fine-grained matrix is a key factor in defining the threshold between two groups: *cohesive* and *non-cohesive* (granular) flows (Mulder and Alexander, 2001). *Cohesive* flows have a matrix strength that imparts *a pseudoplastic rheology,* resulting from cohesion between clay and fine silt particles. Depending on the percentage of the silt/clay, a range from mudflows (> 40% clay) to siltflows (<25% clay) may be differentiated. Sometimes reduced resistance at the flow/seafloor interface due to the hydroplaning process favours long runouts and low erosional power of the flows. As a consequence of plastic behaviour, deposition occurs through the "freezing" of the flow, when the applied shear stress falls below the yield stress (Mulder, 2011). A very common cohesive flow is the so-called debris flow, although this term is sometimes used in the literature for indistinct flows. Debris flows are defined as

Process	Rheology/ transport mechanism	Sedimentary structures	Seismic features
Slide	Elastoplastic/Coulomb Shear failure along discrete shear planes	Undeformed continuous bedding	The deposits show little internal deformation and pre-existing bedding is preserved. Plastic deformation can occur at the base of the failed deposit.
Slump	Elastoplastic/Coulomb Shear failure with rotation along discrete shear surface	Plastic deformation at the toe, folds, tension faults, rotational blocks.	Compressional ridges, irregular upper bedding contacts, contorted layers.
Debris flow	Viscoplastic Cohesive flow: Strength is principally from cohesion due to clay content.	Generally a poor grading and fabric. Massive beds with some blocks at the top of the flow. Typical hummocky surface on the seafloor.	Convex-up shape with low amplitude to transparent facies. The presence of blocks generates hyperbolic reflectors.
Rock/ debris avalanches	Non cohesive flow: Strength is principally from grain-to grain interaction	Poorly sorted ungraded to normally graded breccia or conglomerate with little matrix. Finer grained tail over the coarse grained head	Widespread, hummocky depositional lobes.
Turbiditic flow	Newtonian Supported by fluid turbulence	Normal size grading, sharp basal contacts, gradational upper contacts. Bouma sequence	Lobate or laterally continuous reflectors.

Table 1. The most common types of mass movements described in the marine environment and the characteristics of their deposits, compiled from Moscardelli *et al.* (2008) and Mulder *et al.* (2011).
Tabla 1. *Inestabilidades sedimentarias más comunes en el medio marino y características de sus depósitos. Compilado de Moscardelli et al. (2008) y Mulder et al. (2011).*

plastic (Bingham), poorly sorted flows in which clasts float in a fine-grained matrix. Debris flows may transport boulder-size clasts floating close to the upper surface of the flow.

Non-cohesive or cohesionless flows are essentially grain flows, made of discrete grains with a limited amount of cohesive material. Grain-to-grain interaction supports the particles in these flows (Iverson, 1997). This process requires high energy levels and steep slopes are needed for it to be maintained (Jaeger *et al*., 1996). The high energy explains the erosive character of these cohesionless flows. Erosion over the seafloor can produce an increased concentration of fine-grained material and, finally, collapse of the flow. Deposition is also assumed to occur via the freezing of the flows due to the interlocking of the grains (Prior *et al*., 1982). A debris/rock avalanche is one type of cohesionless flow. It involves large volumes of failing masses (fragmented bedrock or consolidated sediment) enabling clasts to move and segregate. Large clasts (blocks from metres to hundreds of metres in size) can accumulate in different parts of the flows, including even running ahead of the flows (Prior and Doyle, 1985; Blikra and Nemec, 1998). This type of flow usually originates from deep rotational failures on high gradient slopes (>10°), which transform into debris avalanches through a process of shearing fragmentation and dilation. The avalanching of rock or consolidated material mostly occurs in volcanic environments and is characterised by high velocities (10-100 m s^{-1}) and long runouts (10-40 km) (Masson *et al*., 2002).

Another important type of gravity flow with a significant impact on deep-sea sedimentation are turbidity currents. These are flows with a Newtonian rheology of mixed sediment in which the particles are maintained in suspension due to fluid turbulence. Besides the turbulence, it is accepted that there may be other particle-support processes acting near the bed (Shanmugam, 2002; Mulder, 2011). Turbidity currents show variation in their vertical and longitudinal structure. Velocity and density decrease upwards, and turbulent flows consist of head, neck, body and tail. The coarser grains tend to concentrate in the bulge-shaped head. This explains why the head is mainly erosional while the body is mainly depositional; the tail is the diluted thinner back part of the flow. These differences are important in explaining flow spilling and stripping on obstacles such as sedimentary levees, and for meander formation (Mulder, 2011).

Understanding the dynamics of failures. The need to know the physical/geotechnical properties of marine sediments

The general view of submarine mass movements is that their occurrence is a multivariable problem expressed as a complex equilibrium between applied and resisting stresses. The loss of equilibrium is related to an increase in environmental loads or to a decrease in the strength of the sediment, or a combination of both (e.g., Hampton *et al.*, 1996; Leynaud *et al.*, 2004; Mulder *et al.*, 2009; Zitter *et al.*, 2012). Once the stability is lost and mass movement is initiated, for example as a coulomb failure, a wide range of factors can define the behaviour of the sediments until they are deposited or transformed into another type of movement as debris flow or turbidity current (Fisher, 1983; Locat and Lee, 2000). To understand submarine landslides and their pre/post failure behaviour it is necessary to take into account all the factors considered in geological and geotechnical approaches.

Often an external trigger (e.g., an earthquake) is needed to initiate a landslide, but the location of the failure surface within the sedimentary column is predetermined by the physical and geotechnical properties of the materials involved, particularly the shear strength. The shear strength of marine sediments depends on their intrinsic properties, which are mainly determined by mineralogy and grain size. Thus, poor-sorted sediments present high internal friction angles and, consequently, greater shear resistance than well-sorted sediments. Moreover, cohesion depends on the type of clay within the sediments, which in turn determines the plasticity and therefore the shear strength. Furthermore, the presence of organic matter increases the plasticity of the sediment and dramatically reduces its resistance (e.g., Skempton, 1970; Locat and Lee, 2000).

Porosity and permeability are other key parameters that influence the development of pore-water overpressure. Sediment weakens mainly in response to pore-water overpressure affecting the effective stress.

Shear strength increases with burial depth due to dewatering and compaction of the sediment; moreover, sediments that have previously been under a higher load (i.e., over-consolidated, such as sediments below an erosional unconformity) tend to resist failure better than identical sediments under a state of normal consolidation. Sediments can become over-consolidated (strengthened) as a result of bioturbation (e.g., Perret *et al.*, 1995), erosion (e.g., Skempton, 1970), cementation (e.g., Bryan and Bennett, 1988), and repeated seismic loading (if sediments are able to drain overpressures generated by the earthquake, Boulanger *et al.*, 1998). All these factors govern the existence of so-called “weak layers” (e.g., O’Leary, 1991; L’Heureux *et al.*, 2012; Locat *et al.*, 2014), which may become failure surfaces. Weak layers often meet unconformities or other key surfaces (onlaps, downlaps and erosional surfaces, ooze horizons, tephra layers, etc.), thus highlighting the importance of the pre-failure architecture, the sedimentology of the deposits and their geotechnical properties, in understanding slope stability along continental margins.

The determination of shear strength in marine sediments often includes the assumption that these materials behave as a Mohr-Coulomb elastic, a perfectly plastic material. This Mohr-Coulomb model requires two parameters that can be obtained from basic *in situ* tests or from samples tested in the laboratory: ϕ: the Friction angle (°); c: Cohesion (kPa). To determine the stress state of the material it is also necessary to measure the pore pressure (u). On the other hand, to study the stress-strain behavior elastic parameters should be obtained: E: Young’s modulus (kPa); ν: Poisson’s ratio (-).

Examples of *in situ* testing include the Cone penetrometer test with pore pressure measurement (CPTU) providing cone tip (q_c), friction sleeve (f_s) and excess pore pressure (u) measurements continuously along the length of the probe. In the laboratory, the parameters for modelling sediments can be obtained from continuous tests along sediment cores or from discrete samples. In the first case, Vane Shear Tests or Penetrometers provide peak and residual undrained strength data ($S_{u\ peak}$ and $S_{u\ res}$). In the second case, deformational properties are obtained from the oedometer test, while the shear strength is obtained from triaxial or direct shear tests (Fig.1).

Once the geotechnical parameters have been obtained, numerical modelling is often carried out based on an assumption of plane deformation. Stability is established in terms of a numerical value known as the safety factor (SF) which is usually obtained by limit equilibrium methods, so that SF = resisting forces/driving forces. Finite element methods (FEM), which consist of computing the maximum displacement for various values of SF, can also be used.

Forces controlling the triggering of submarine mass movements

Many factors have been suggested as probable or possible contributors to the initiation of submarine landslides; they range from sudden impacts operating on a timescale of minutes (short-term triggers, such as shaking due to earthquakes), to geological processes operating on timescales of tens or hundreds of thousands of years (long-term or predisposing factors, such as climate change and sedimentation processes). The main factors are:

- *Rapid sedimentation rates.* This contributes to slope failure through two main mechanisms: a) an

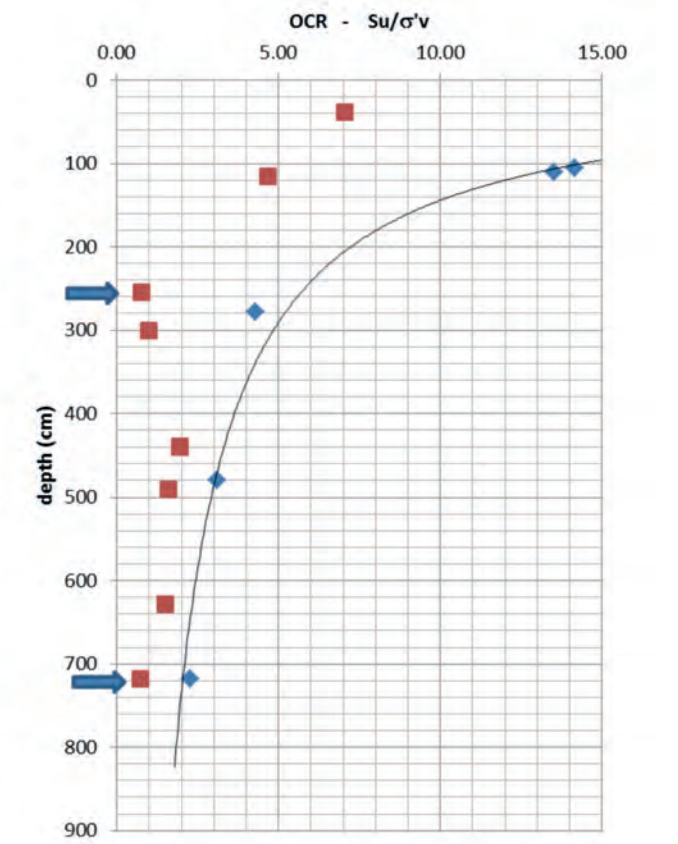

Figure 1. Example of deformational and strength properties obtained from sediment samples. The blue dots correspond to the variation of the over-consolidation ratio (OCR) parameter with the depth. The red squares correspond to the relationship between the undrained shear strength (Su) and the vertical effective stress (σ'_v). In the designated areas (arrows) the undrained shear strength is lower than the vertical effective stress and, consequently, these are areas of lower resistance prompted the development of failure surfaces in the event of an external trigger. Modified from Yenes *et al.* (2012).

Figura 1. *Ejemplo de propiedades de deformación y resistencia obtenidas en muestras de sedimento. Los puntos azules corresponden a la variación de OCR (over-consolidation ratio) respecto la profundidad. Los cuadrados rojos corresponden a la relación entre la resistencia a la cizalla no drenada (Su) y el esfuerzo vertical efectivo (σ'_v). Los intervalos (flechas) donde la resistencia a la cizalla es menor al esfuerzo vertical efectivo corresponden a zonas de baja resistencia donde se puede desarrollar una superficie de rotura ante la acción de un detonante. Modificado de Yenes et al. (2012).*

increase in shear stress due to slope over-steepening, and b) retarded strength development. This is explained by the fact that a rapid lithostatic load is mainly carried by pore-water pressure, producing an under-consolidated state of sediment due to excess pore pressure (Sultan *et al.*, 2003).

- *Gas and gas hydrates.* Free gas and gas from dissociation of gas hydrates in the pore spaces mainly act as a predisposing factor for slope failure by decreasing the shear strength of the sediment. This can be explained by the fact that gas bubbles exert a pressure on the surrounding water that subsequently becomes over-pressured. Any change in the equilibrium parameters controlling the stability field of gas hydrates (e.g., pressure, temperature; Sultan *et al.*, 2003) may trigger the conversion of hydrate to free gas and water, generating an increase in pore pressure and a significant weakening of the sediment.
- *Erosion* can act as a predisposing/triggering factor by decreasing shear resistance in the slope. This process is observed, for example, on submarine canyon sidewalls, as the base can be undercut by erosive gravity flows leading to a progressive slope over-steepening and decrease of stability.
- *Groundwater seepage* from coastal aquifers on the continental shelf and slope has been proposed as a possible trigger during periods of lowered sea levels (Budillon *et al.*, 2011). A similar process occurs in coastal areas during low tides or when tsunamis approach the shoreline generating a sudden lowering of sea level (e.g., Morgenstern, 1963; Kulikov *et al.*, 1996; Seed *et al.*, 1988; L'Heureux *et al.*, 2011). When water levels fall rapidly, the pore pressure in the subaerial zone does not have time to reach steady state conditions. This situation generates an accelerated seepage of ground water seawards, which can act as a driving stress and/or as excess pore pressure, reducing the effective stress.
- *Tectonic activity* contributes to slope failure through: a) an increase in shear stress due to tectonic deformations resulting in steepened seabed surfaces, and b) a decrease in shear strength close to or at the faults due to shearing, dilatancy and possible sediment creep.
- *Earthquakes* are considered the most common trigger of landslides (e.g., Locat and Lee, 2000). They contribute to slope failures both by increasing the shear stress on the slope and by decreasing shear strength through the development of pore overpressure. This is mainly due to the cyclic loading exerted on the sediment by an earthquake. In this case, the sediment dynamics are controlled by the intensity and duration of the cyclic loading, as well as by the state of the sediment (i.e., grain size distribution, and the presence or absence of clay fraction).
- *Storm-waves* induce slope failures by increasing shear stresses and pore pressures in sediments through cyclic loading, in a similar way to that described for earthquakes (Henkel, 1970; Seed and Rahman, 1978). The storm-wave stress varies with

the characteristics of the waves, i.e., water depth and the depth below the seafloor (e.g., Yamamoto, 1981), but it is unlikely to affect depths of greater than 100 m (Wright and Rathje, 2003). The repeated reversals of shear strain in the sediment can also cause remoulding of the sediment and thus a reduction in shear strength.

- *Volcanic activity* has a strong impact on sedimentary environments around submarine volcanoes, because of the potential to supply large amounts of lavas and tephras in a very short period of time, thus altering the "normal" basin-margin sedimentation and stress environment. Moreover, due to the steepness of volcanic slopes, volcanic settings are very prone to the development of landslide processes on their flanks, ranging from small rockfalls (with volumes of a few thousands of cubic metres) up to large debris avalanches, mobilising tens or thousands of cubic kilometres of material (McGuire, 2006).
- *Human activity* can also play a role in the development of slope failures, as human-constructed facilities along the coastline or on the seafloor can modify the stress conditions within the slope and increase pore-water pressures. Possible examples of anthropic-induced landslides are the 1977 Gioia Tauro landslide which occurred during the enlargement of Gioia Tauro harbour (Colantoni *et al.*, 1992), the 1979 Var landslide which occurred during the enlargement of Nice airport (Assier-Rzadkiewicz *et al.*, 2000; Dan *et al.*, 2007), the 1994 Skagway landslide in Alaska (Rabinovich *et al.*, 1999), and the 1996 Finneidfjord slide in Norway (Longva *et al.*, 2003).

Characterisation of a continental margin and register of submarine mass movement

Following the above arguments, the understanding of submarine landslides takes into account several factors. The analysis of all these factors can explain the distribution of landslides in the sedimentary record of a continental margin. Sediments are prone to failure depending on their composition, geometry, stress history and, ultimately, their location. They can fail where they form deposits that may experience low shear strength (under-consolidation, excess pore pressure, etc.) and/or are subjected to processes that can increase the applied stress (cyclic loading from earthquakes, storm-waves, etc.). Because of this, although submarine landslides are widespread processes that can be found in almost all settings, they commonly occur in certain specific environments: fjords, active river deltas, canyon-fan systems, open slopes and oceanic volcanic islands (Hampton *et al.*, 1996).

Rapid sedimentation rates are used to explain the widespread mass-wasting features recognised in fjords and delta environments, such as on the Mississippi Delta front (e.g., Coleman *et al.*, 1993 and references therein). Gas-charging is also one of the main triggering mechanisms for slope failures in fjords and on deltas, where a large amount of gas is generated through the decay of organic matter transported by rivers (e.g., Locat and Lee, 2000). Storm-wave loading is another important trigger in these environments and is recognised, for example, as the major factor responsible for causing submarine landslides which in turn, led to the failure or damage of several offshore drilling platforms when Hurricane Camille struck the Mississippi Delta in 1969 (Bea *et al.*, 1983).

Examples of landsliding processes at the canyon head and on sidewalls have been widely documented in various tectonic and physiographic settings (e.g., McAdoo *et al.*, 2000; Greene *et al.*, 2002; Arzola *et al.*, 2008; Paull *et al.*, 2010 and 2013; Casalbore *et al.*, 2011;). Baztan *et al.* (2005) also showed how axial incision can trigger slides in the sedimentary fill of the canyon; those destabilisations can, in turn, induce instabilities and retrogressive slumps along the main flanks of the canyons. Moreover, landslide deposits from sidewalls often dam the underlying canyon floor, so that subsequent sedimentary gravity flows are diverted and erode away part of the dam forming a new talweg and/or meander. These new meanders can lead to further erosion and second-generation landslides (e.g., Baztan et al., 2005).

Open slope landslides are related to several factors such as sedimentation rates, tectonic activity and earthquakes. Examples of interaction between tectonics and slope failures are numerous and can be found, for example: a) on the Californian continental slope, where widespread mass-wasting features are related to the continual Quaternary uplift of the San Pedro tectonic escarpment (Bohannon and Gardner, 2004); b) on the western Ligurian margin, where a close correlation between slide scars and faults affecting the base of the slope is observed (Migeon *et al.*, 2011), and c) in the Gulf of Mexico, relating to salt diapiric deformations (Silva *et al.*, 2004). Examples of earthquake-induced submarine failures are numerous and include the famous 1929 Grand Banks event (Piper *et al.*, 1999), and the catastrophic 1998 Papua New Guinea landslide that caused a tsunami that killed over 2,000 individuals (Tappin *et al.*, 2001).

The destabilisation of entire flanks of volcanoes, generating huge debris avalanche deposits, is commonly observed around several volcanic islands, such as Hawaii (Moore *et al*., 1989, 1994; McMurtry *et al*., 2003), the Canary Islands (Urgeles *et al*., 1999; Masson *et al*., 2002), the Cape Verde Islands (Elsworth and Day, 1999; Masson *et al*., 2008), La Rèunion (Oehler *et al*., 2008), Tristan da Cunha (Holcomb and Searle, 1991), Lesser Antilles (Boudon *et al*., 2007), Aleutian Arc (Coombs *et al*., 2007), Sandwich Arc (Leat *et al*., 2013); Bismark Arc (Silver *et al*., 2009); Oshima-Oshima (Satake and Kato, 2001), Ischia (Chiocci and DeAlteriis, 2006) and Stromboli Island (Casalbore *et al*., 2010). It should also be taken into consideration that repeated large-scale instability processes often occur on the same volcanic flank due to feedback effects between collapses, magma upwelling and dyke emplacement (e.g., Tibaldi, 2001).

Driving mechanisms, variety of landslides and controlling factors vary according to the environment, but environmental conditions are not constants through time. This fact entails different probabilities of failure for different time periods. The type and rate of sediment delivered to the continental margins can be correlated to climatically-controlled changes in sedimentation style from glacial to interglacial conditions. In this sense there are certain differences between glaciated and non-glaciated margins (e.g., Owen *et al*., 2007; Leynaud *et al*., 2009) and there are also differences in the timing of submarine slope failures. In low latitudes, submarine slope failures preferentially develop during glacial conditions with low sea-levels, when depocentres move over the continental slope and rapid sediment deposition occurs directly onto the upper continental slope. In high latitudes, slope failures preferentially develop during the relatively fast transition from glacial to interglacial conditions (i.e., during sea level rises). Clayey units deposited during interglacials are rapidly loaded by large amounts of coarse-grained glacial sediments in glacial periods, promoting the development of high pore-pressures in clay units. This fact, together with continental uplift and increased seismic activity due to isostatic rebound, results in a greater probability of failure (Bryn *et al*., 2003; Lee, 2009). It is worth mentioning that similar cyclic conditions may contribute to repeated landslides in the same area (Bryn *et al*. 2005; Casas *et al*., 2013).

An overview of submarine mass-failure processes around Iberia

Iberian continental margins and adjacent basins or abyssal plains are the result of a complex geologic evolution. As a result of this complexity, diverse areas with peculiar morpho-sedimentary and morpho-tectonic features can be differentiated surrounding the Iberian Peninsula (Maestro *et al*., 2013). Dominant tectonic control is observed in the Alboran, Gulf of Cadiz, Portuguese, Galicia and Cantabrian margins. Sedimentary processes prevail on the Valencia-Catalan margin and the Balearic Promontory (Fig. 2). The bottom currents also have an important role throughout the Iberian margins (e.g., Hernández-Molina *et al*., 2011).

With these diverse geologic settings it is difficult to construct a comprehensive picture of the role of submarine landslide processes around the approximately 23 million km^2 that involve the Iberian margins (Maestro *et al*., 2013). This is also because the distribution of known submarine landslides is not well understood due to an incomplete coverage and a lack of uniform studies in all areas. Even so, the aim of this section is to give a broad overview of the distribution of submarine mass movements in the above mentioned geological contexts. There have only been a few attempts to provide complete inventories to with help this objective. Information from the scientific literature has been compiled in a GIS-based framework for the continental margins of the Western Mediterranean but not for the Atlantic margins (Camerlenghi *et al*., 2010; Urgeles and Camerlenghi, 2013). In spite of this, there are a huge number of studies demonstrating that submarine landslides are ubiquitous on the Iberian margins and adjacent deep sea areas (e.g., Baraza *et al*., 1990; 1992; Acosta *et al*., 2002; Casas *et al*., 2003a; Droz *et al*., 2006; Urgeles *et al*., 2006; Lastras *et al*., 2007; Urgeles *et al*., 2007; Camerlenghi *et al*., 2009; Cattaneo *et al*, 2010).

Several authors have focused their efforts on the study of the morphological and dynamic characteristics of particular individual landslide systems (e.g., Urgeles *et al*., 2006; Iglesias *et al*., 2010; Casas *et al*., 2011; Lafuerza *et al*., 2012). Other authors have focused on their role in the stratigraphical architecture of margins and basins (Ercilla *et al*., 2008a; 2011a; Vázquez *et al*., 2013; Alonso *et al*., 2014; among others). Some striking cases involve huge volumes of sediment filling large areas in different stratigraphic positions, meaning a significant contribution to the sedimentary architecture. The largest example is the megaturbidite deposited on the Balearic Abyssal Plain at ~22 ka cal BP, with a volume of ~500 km^3 and an area of ~6x10^4 km^2 (Fig. 2; Rothwell *et al*., 1998). The Balearic megaturbidite remains an enigmatic event as the source area has not yet been identified. Another important example is the Western Gulf of Lions debris flow (Fig. 2) (Canals, 1985; Alonso *et*

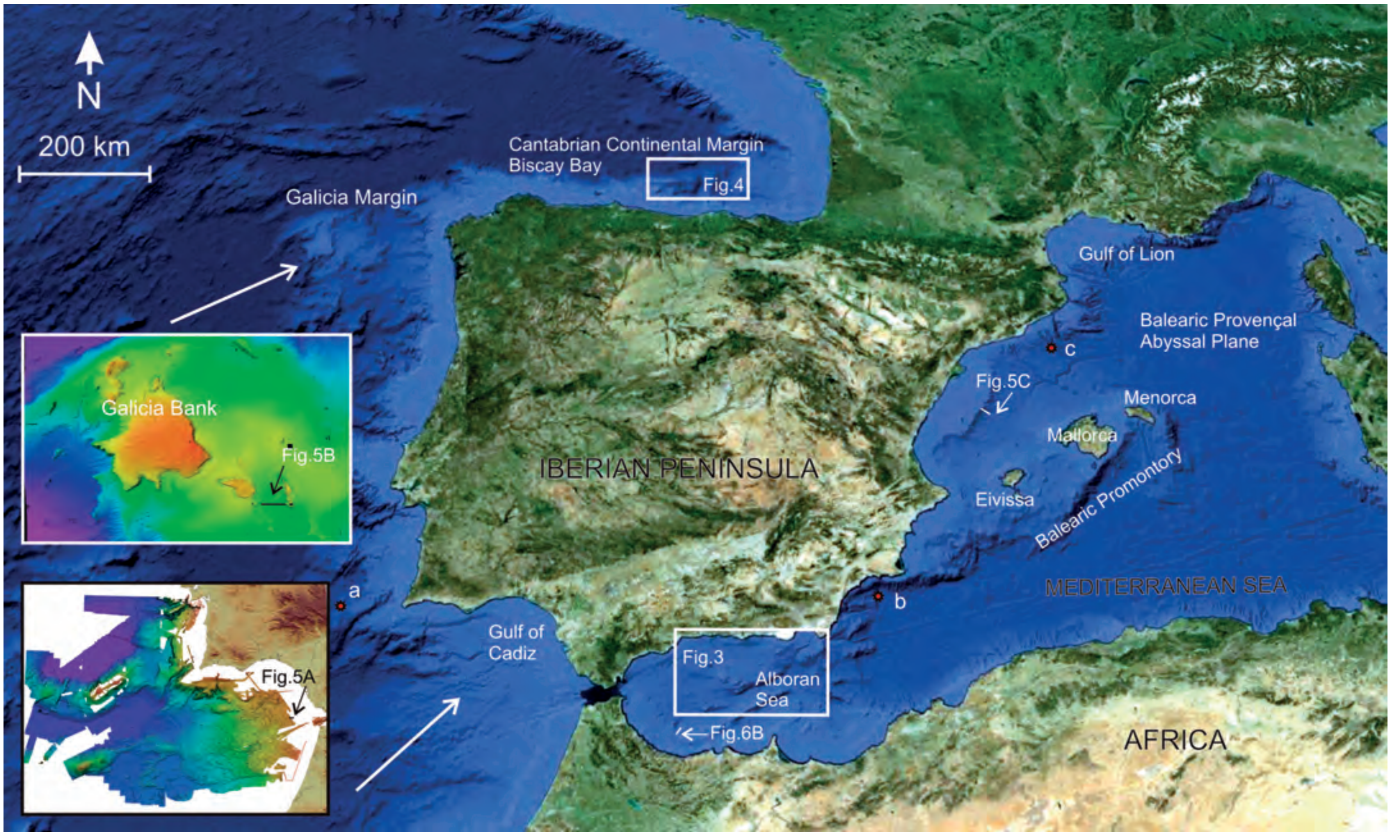

Figure 2. General bathymetry of the continental margins around the Iberian Peninsula (extracted from Google Earth). The location of Figs. 3, 4, 5 and 6 are also displayed. The red dots a, b and c correspond to the Marques de Pombal fault, the Aguilas canyon (Palomares margin) and the Catalan margin respectively.
Figura 2. *Batimetría general de los márgenes continentales alrededor de Iberia (extraído de* Google Earth*). En el mapa se localizan las Figs. 3, 4, 5 y 6. Los puntos rojos a, b y c corresponden a la localización de la falla Marques de Pombal, cañón de Águilas (margen de Palomares) y margen Catalán respectivamente.*

al., 1991) which involves at least 260 km^3 of material (Gaullier *et al.*, 1998). But more moderate deposits (<200 km^2) are widely distributed on the seafloor, in all physiographic domains, and embedded in the sedimentary sequences making up margins and basins. Most submarine landslides have been generically identified as Mass Transport Deposits, but debris-flow deposits, shallow- and deep-seated failures and slumps are also common failure styles. In the majority of cases, tectonic activity, sedimentary load or climate-induced stress such as sea level changes have played a major role as triggering mechanisms.

The compilation undertaken by Urgeles and Camerlenghi (2013) established that most exposed landslides in the western Mediterranean originate in water depths exceeding 2 000 m on slopes of 2° and the majority arrest only at slightly deeper water depths. This is also shown by the relatively short vertical displacement that landslide deposits exhibit, with most of the events (44%) displaying vertical displacements not exceeding 100 m, while 85% of the landslides have a vertical displacement of shorter than 500 m. This illustrates that: a) the landslides in the database are relatively small, but also b) that the continental rise is a place of higher slope instability compared to the continental slope, and c) that limited energy is available for down-slope sediment transport, with most failures arresting shortly after being triggered and/or producing almost null sediment transport. The lack of systematic data for the other margins means it is not possible to corroborate these observations as a general rule for instabilities around Iberia.

Landslides in tectonically active areas

A strong link between tectonic activity and landslides has been established for both the Atlantic and Mediterranean Iberian margins and surroundings. The margins that are most tectonically active display the largest number of slope failure events, although the landslides tend to be smaller. This relationship has also been observed in other margins such as the Algerian margin where 146 submarine landslides have been reported with a mean area of 23.37 km^2 (Camerlenghi *et al.*, 2010). For this margin, Cattaneo *et al.* (2010) indicates that the distribution of landslides appears to be related to present day morphologic sectors and seafloor structures of tectonic origin. However, the relationship between recent earthquake epicentres and submarine landslides is not readily apparent.

In the Alboran Sea (Fig. 2) the reported failures are similar to each other in terms of deposit surface area (30.82 km^2 on average). Post-Messinian tectonic activity has played an important role in the physiographic configuration and stratigraphical architecture of the Alboran basin (e.g., Vázquez *et al.*, 2013). This tectonic activity has been established as the main trigger mechanism for most of the instabilities observed throughout the Alboran Sea (Fig. 3). This fact in turn provides evidence of repeated fault activity where successive landslide deposits appear embedded in the sedimentary sequences (e.g. Vázquez *et al.*, 2013; Alonso *et al.*, 2014).Throughout the South Alboran Basin, tectonism is strongly associated with the Alboran Ridge and the Cape Tres Forcas promontory. Linked to this promontory is the largest exposed deposit observed in the area, the Montera Slide (Fig. 3). It has an average thickness of 50 ms (maximum 180 ms) and covers an area of around 90 km^2 (Vázquez *et al.*, 2013). Several mass movements have been also described from the Pliocene-Quaternary sequence in the area. At least 8 mass-transport deposits (with recurrence intervals of 0.56 to 0.18

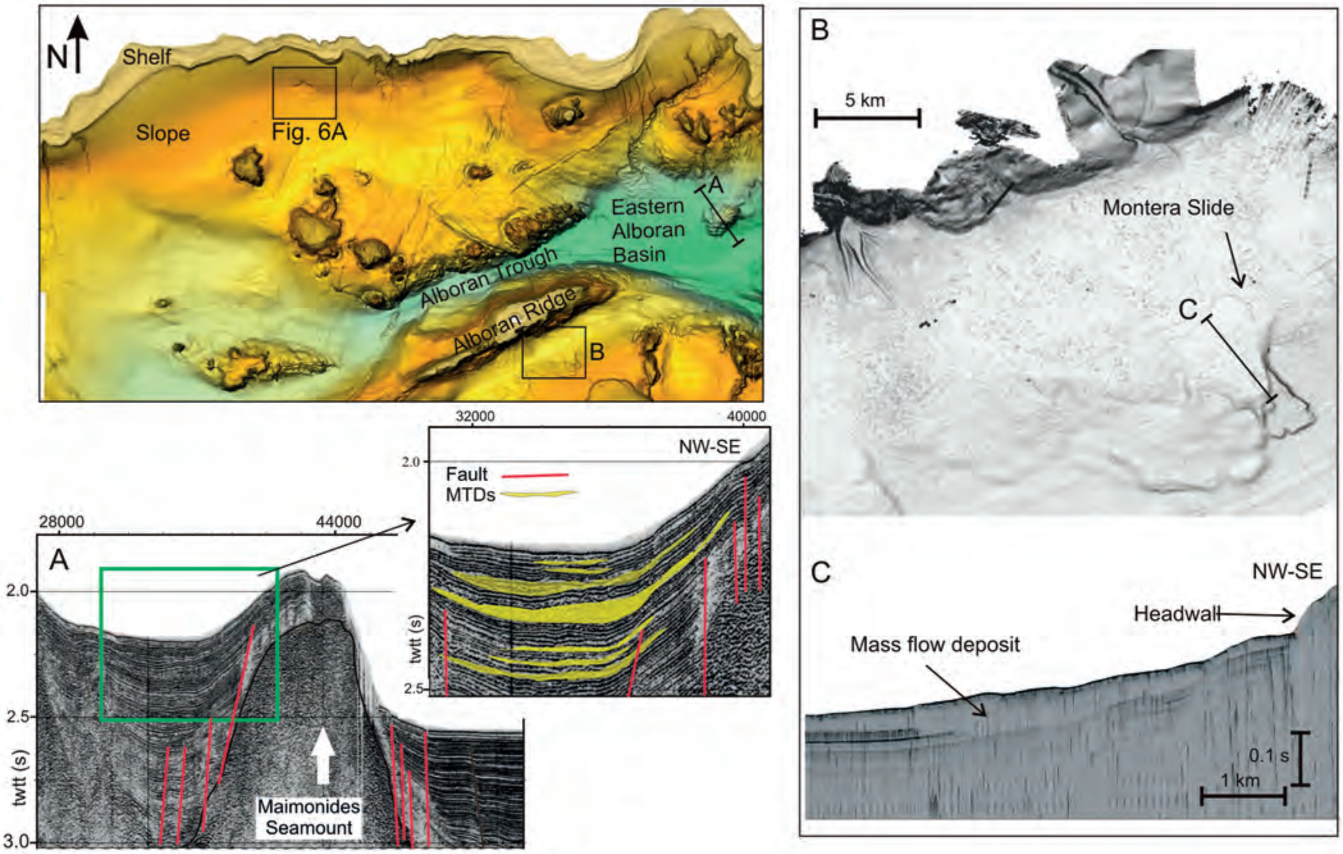

Figure 3. Bathymetry of the Alboran Sea in the westernmost Mediterranean Sea provided by the Spanish Ministry of the Environment and Rural and Marine Affairs (see location in Fig. 2).The location of Fig. 6A is displayed. (A) The staked mass transport deposits (MTDs) defined in the Quaternary sequences around the Maimonides Seamount, modified from Alonso *et al.* (2014). (B) Shaded relief of the Montera slide and (C) seismic-parametric profile showing the headwall and deposit of the Montera slide, modified from Vázquez *et al.* (2013).

Figura 3*. Batimetría del Mar de Alborán cedida por el Ministerio de Medio Ambiente y Medio Rural y Marino (ver localización en Fig. 2). En el mapa también se localiza la Fig. 6A. (A) Depósitos de transporte en masa definidos en la secuencia cuaternaria alrededor del monte Maimonides. Modificado de Alonso et al. (2014). (B) Batimetría sombreada del deslizamiento Montera y (C) perfil sísmico paramétrico mostrando la cabecera y depósito asociados al deslizamiento Montera. Modificado de Vázquez et al. (2013).*

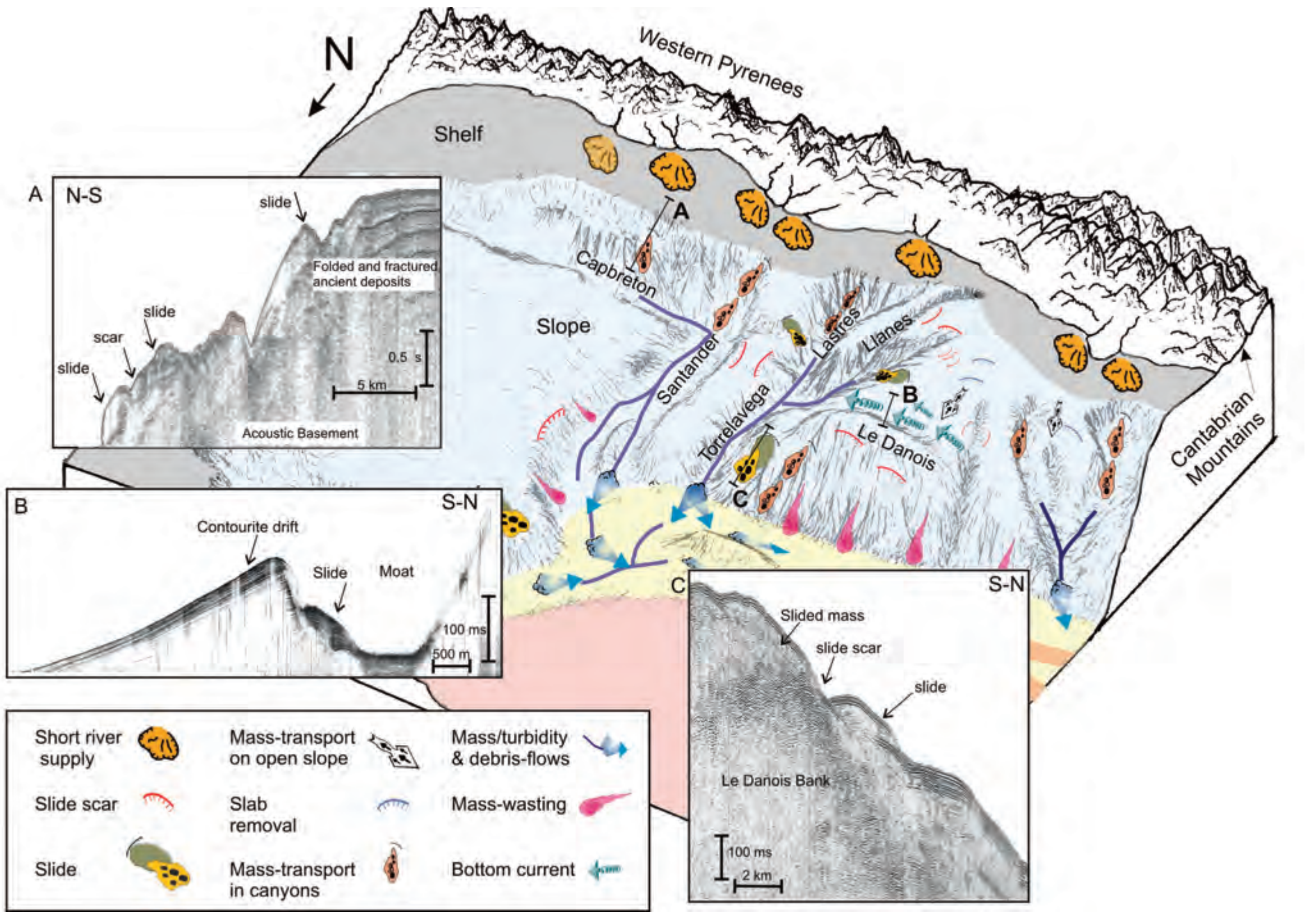

Figure 4. Sketch of the sedimentary instability processes affecting and characterising the Cantabrian Continental margin (see location in Fig. 2), modified from Ercilla *et al.* (2008 b). (A) Airgun profile showing multiple failures on the Capbreton canyon wall. (B) TOPAS profile showing sedimentary failure affecting a contouritic deposit and (C) airgun profile showing sediment failures on the slope of the Le Danois Bank.

Figura 4*. Esquema de la distribución de las inestabilidades sedimentarias que afectan al margen continental del Cantábrico (ver localización en Fig. 2). Modificado de Ercilla et al. (2008b). (A) Perfil airgun mostrando múltiples inestabilidades sedimentarias en un margen del cañón Capbreton. (B) Perfil TOPAS mostrando un deslizamiento afectando a depósitos contorníticos y (C) Perfil airgun mostrando inestabilidades sedimentarias en el talud de Le Danois Bank.*

Myr) have been correlated with two major tectonics phases in the area. The first occurred during the Lower Pliocene and relates to the main uplift of the basin margins; and the second is from the upper part of the Quaternary (0.92 My to present) and is characterized by a reactivation of tectonic structures (Vázquez *et al*., 2013). Similar conditions have been observed in the Eastern Alboran Basin (Fig. 3). At least 53 stacked mass transport deposits are embedded in the Quaternary sequences, around Pollux and Sabinar Banks and Maimonides and Adra Ridges. The deposits exceed 5 km in length and 18 ms thick and have a recurrence period of between 40 to 373 ka (Alonso *et al*., 2014).

In other tectonically active areas, such as the Gulf of Cadiz and south Portuguese margins (Fig. 2), a characterisation of the mass-wasting deposits indicates that they are associated with active faults (Gràcia and Lo Iacono, 2008; Lo Iacono *et al*., 2012). In some cases this activity is capable of generating earthquakes of great magnitude (M>= 8.0) with the potential to trigger slope failures and tsunamis (Bartolomé *et al*., 2012). For instance, a large (260 km^2) translational landslide and debris flow is associated to the Marques de Pombal Fault (Gràcia *et al*., 2003). The most recent slide is from about 230 yr BP and may have been triggered by the 1755 Lisbon earthquake (Gràcia *et al*., 2010), possibly the most destructive event in western Europe in recent history (Gràcia *et al*., 2003). By dating previous slide deposits a recurrence period of 2,000 yr has been obtained (Vizcaino *et al*., 2006), suggesting cyclic activity of the

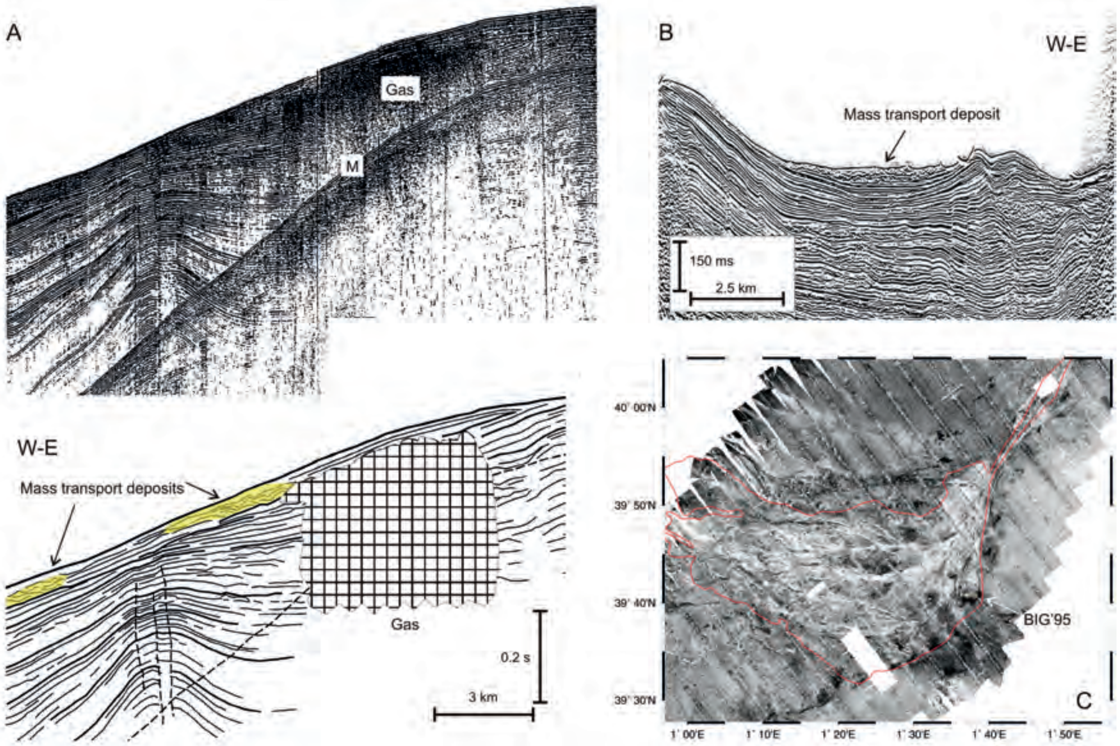

Figure 5. (A) Geopulse seismic profile and line drawing from the upper continental slope of the central Gulf of Cadiz showing acoustic masking caused by gassy sediment and surficial mass-transport deposits immediately downslope. (M=multiple), modified from Baraza *et al.* (1999). (B) Airgun seismic profile illustrating mass-transport deposits mapped in the Galicia Bank region, modified from Ercilla *et al.* (2011). (C) Backscatter and outline (red line) of the BIG 95 debris flow located on the Ebro margin. Modified from Urgeles *et al.* (2003). See locations in Fig. 2.

Figura 5. *(A) Perfil Geopulse y esquema localizados en el talud superior del Golfo de Cádiz mostrando el enmascaramiento acústico causado por la presencia de gas en los sedimentos y las inestabilidades asociadas. (M=múltiple). Modificado de Baraza et al. (1999). (B) Perfil sísmico airgun ilustrando depósitos de transporte en masa localizados en la región del Banco de Galicia. Modificado de Ercilla et al. (2011). (C) Reflectividad y delimitación (línea roja) del flujo de derrubios BIG'95 localizado en el margen del Ebro. Modificado de Urgeles et al. (2003). Ver localizaciones en Fig. 2.*

Marques de Pombal Fault (Fig. 2). Consequently, it can be seen that there is a significant geohazard affecting the coasts of Portugal, Spain and Morocco.

Tectonic activity also plays a predominant role in causing the submarine mass movements of the northern Iberian Peninsula margins (Figs. 2, 4 and 5). Sedimentary instability represents, for example, one of the most widespread elements in the Galicia Bank region (Ercilla *et al.*, 2008a; 2011a). Sedimentologically, the Galicia Bank is considered a seamount far from continental sediment sources. In this context, the sediment sources are linked to relict or relatively recently active structural scarps. Sediment is delivered as a result of tectonic, chemical and physical (disintegrative) wasting processes (Ercilla *et al.*, 2008a). The exhumation and erosion of the scarps have favoured the frequent occurrence of mass transport and turbidite deposits (Fig. 5; Alonso *et al.*, 2008; Ercilla *et al.*, 2008a; Casadei, 2012).

Mass movement deposits of variable dimensions also characterise the near-surface sediments in the canyons and continental slope of the Bay of Biscay and Le Danois Bank (Fig. 4; Ercilla *et al.*, 2008b). Although it is unclear what the main triggering mechanism in the area is, it probably related to seismicity, oversteepening, and particularly faulting (Ercilla *et al.*, 2008b). In the same way as was observed for the western Mediterranean, mass-wasting deposits in the area, especially those located on the scarps and open slope, are characterised by short runout distances (of up to hundreds of metres) due to their deposition down the lower steep and narrow continental slope (Fig. 4; Ercilla *et al.*, 2008b).

Sometimes the presence of faults and slumps plays an interdependent role in the evolution of submarine canyons. This is the case of the Aguilas canyon, located on the northern Palomares margin, W Mediterranean (Fig. 2). High-angle faults (Águilas

Escarpment) and the uplifting of the margin are responsible for the diverse small-scale, retrogressive landslides that cover 180 km^2 from the lower slope to the continental rise, favouring the evolution of the canyon (Pérez-Hernández *et al.*, 2009). Other canyons observed such as Portimao, Fado (Portuguese margin); Llanes, Torrelavega and Santander canyons (Fig. 4; Bay of Biscay); Torreblanca, Peñiscola and Francolí (Ebro margin) or the Menorca canyon (Balearic margin), all appear to have been affected by failures (Alonso *et al.*, 1991; Canals *et al.*, 2000; Casas *et al.*, 2003a; Ercilla *et al.*, 2008b; Sayago-Gil *et al.*, 2008; Lo Iacono *et al.*, 2014). In most of the cases individual or multiple, and sometimes retrogressive, failures begin on the upper margins of the canyon walls, being one of the drivers for canyon upslope migration. The sediment removed travels along the walls down to the canyon floor, evolving to mass flow processes and turbidity currents.

Landslides in other settings. Sedimentary, oceanographic and fluid-related processes.

There are areas around the Iberian Peninsula with lower seismicity, where tectonic activity cannot be disregarded but does not play a leading role as a trigger. This is the case of the Valencia-Catalan margin or the Balearic Promontory where sedimentary processes are dominant. Some parts of these margins such as the Ebro margin are fed by large river systems and sediment instability is linked mainly to over-steepening and/or a reduction in shear strength induced by differential compaction, high sedimentation rates or the presence of free gas in the sediments (Baraza *et al.*, 1990; Farran *et al.*, 1990; Urgeles *et al.*, 2006).

The catalogue of submarine landslides for this geological setting reports 41 landslides in the Ebro margin, with a mean surface of 187.97 km^2 (Urgeles and Camerlenghi, 2013). The largest exposed landslide in the Ebro margin is the BIG'95 debris flow (Fig. 5) that affected 2 200 km^2 of the seafloor and mobilised 26 km^3 of sediment on the slope and base of slope (Lastras *et al.*, 2002; Urgeles *et al.*, 2003). The location of BIG'95 is affected by overloading due to the input of the Ebro River through time and large overpressure that must be present at depth (Urgeles *et al.*, 2006). However, the trigger for the particular case of BIG'95 was a consequence of growth pulses in the volcanic Columbretes Islets (Lastras *et al.*, 2007). The evolution of the volcanic dome favoured an oversteepening of the margin and enhanced seismic activity. In the same way that the Ebro River leads to high sedimentation rates, other rivers like the Llobregat play a similar role, explaining the frequent landslide deposits in the intercanyon and open slope areas (Figs. 2; 5) of both the Ebro and Catalan margins (Lastras *et al.*, 2007).

Submarine landslides are also quite common in sediment starved margins, far from any river input, where sediment mainly originates from local coastal erosion processes and then environmental stresses from high sedimentation rates are absent. This is the case of the continental slope of the Balearic Promontory and the channels between islands such as Menorca, Mallorca and Eivissa (Acosta *et al.*, 2002; Lastras *et al.,* 2004; Acosta *et al.*, 2013; Lo Iacono *et al.*, 2014). The Menorca Channel is the sector connecting the Menorca and Mallorca islands (Fig. 2), where widespread mass movement features have been described. The preponderant role of instability features in shaping this insular margin has been associated to steep gradients, the presence of weak layers and the action of major storms during lowstand stages (Lo Iacono *et al.*, 2014). On the other hand, part of the Eivissa Channel is occupied by a series of small landslides, the largest affecting 16 km^2 (Lastras *et al.*, 2004). Contrary to Menorca, fluid-related features (e.g. pockmarks) suggest the gas seepage as the responsible for the sediment weakness in this area (Berndt *et al.*, 2012; Panieri *et al.*, 2012). Sediment instability associated with the presence of gas has also been described in other areas such as the Gulf of Cadiz (Fig. 5). Multiple slumps, occupying up to 147 km^2 and mainly concentrated in the continental slope, are genetically related to the presence of free gas in the sediment (Baraza *et al.*, 1999; Casas *et al.*, 2003b).

The bottom currents have a great impact in deep-sea morphodynamics; in this sense water-masses have a critical role governing the sedimentary outbuilding of deep-sea areas throughout the Iberian margins from Mediterranean to Cantabrian Seas (Hernández-Molina *et al.*, 2011; Hernández-Molina *et al.*, 2014 amongst others). Sediment instability has also been observed associated with different features of contouritic depositional systems. Local intensification of bottom currents is responsible for erosive processes, undermining slopes and causing instability (Fig. 4). But gravitational instability in contouritic sediments also occurs frequently (Fig. 6). Contouritic sediments can be prone to failure because of their composition (i.e., well-sorted), geometry and location. They often develop excess pore pressure due to high sedimentation rates and/or relatively high organic-carbon content (Larberg and Camerlenghi, 2008).

The best known example around the Iberian Peninsula is located in the Gulf of Cadiz which is influ-

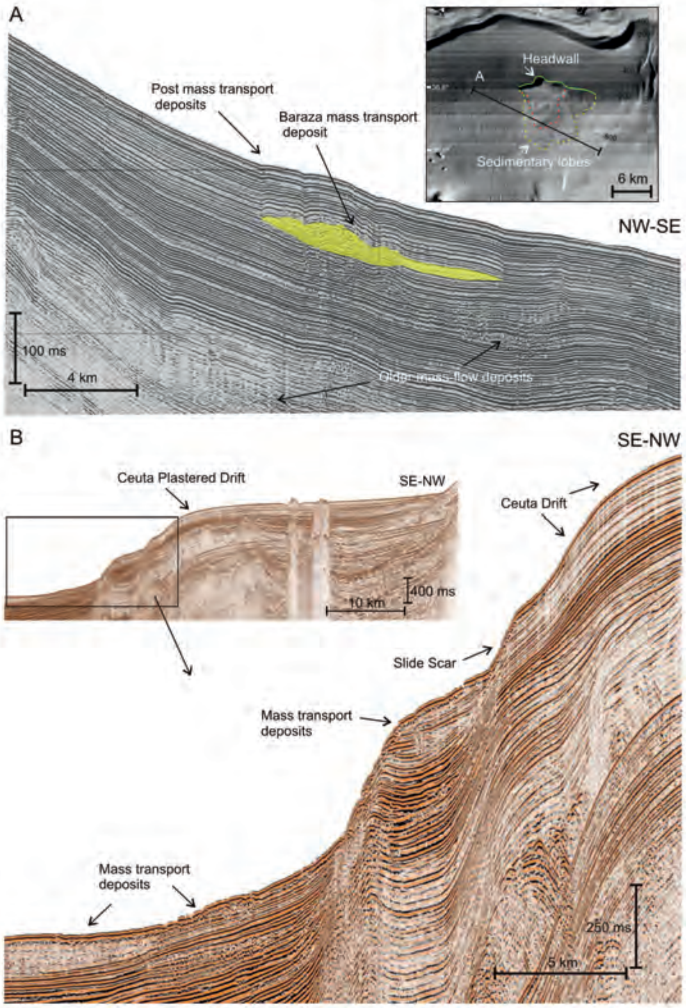

Figure 6. (A) Map showing the morphology of the Baraza Slide affecting the slope- sheeted drift defined in the Alboran margin (see location in Fig. 3). The sparker profile illustrates the occurrence of this mass-transport feature, modified from Casas *et al.* (2011). (B) Airgun profile showing mass movement features affecting the contouritic deposits known as the Ceuta drift (Moroccan slope; SW Alboran Sea). The trigger suggested for the particular case of these failures is a consequence of the diapiric activity in the area, modified from Ercilla *et al.* (2014). See location in Fig. 2.

Figura 6. *(A) Mapa mostrando la morfología del deslizamiento Baraza que afecta al* drift *contornítico definido en el talud continental N del Mar de Alborán (ver localización en Fig. 3). Perfil* Sparker *ilustrando el depósito asociado al deslizamiento Baraza. Modificado de Casas* et al. *(2011). (B) Perfil* airgun *mostrando inestabilidades sedimentarias afectando a los depósitos contorníticos del* Drift *de Ceuta (margen marroquí del Mar de Alborán). El detonante sugerido para estas inestabilidades en particular está relacionado con la actividad diapírica observada en la zona. Modificado de Ercilla et al. (2014). Ver localización en Fig. 2.*

enced by the Mediterranean outflow water (MOW) responsible for the formation of a contourite depositional system. Deformation and instability of sediments associated with the contouritic feature is evidenced in the form of slide scars and multiple slumps (Lee and Baraza, 1999; Hernández-Molina *et al*., 2003; Mulder *et al*., 2003; García *et al*., 2014).

Recently, a new depositional model that characterizes the margins and sub-basins of the Alboran Sea identifies the ubiquitous contourite features in this area. The model includes depositional (plastered, sheeted, channel-related, mounded and separated drifts), erosive (moats, channels and furrows) and mixed (terraces and scarps) features (Ercilla *et al*., 2011b; Ercilla *et al*., 2014; Juan *et al*., 2014). In this context, mass transport deposits ranging from hundreds to a few kilometres in scale, are found locally within slope sheeted and plastered drifts (Fig. 6; Casas *et al*., 2011; Ercilla *et al*., 2011b; *Ercilla et al*., 2014).

Submarine landslides can, in turn, significantly alter the relief of the sea-floor and thereby initiate a realignment of the prevailing current pattern causing the creation of new drifts. This fact has been observed along the SW Mallorca slope where mass wasting appears to be associated with drift deposits (Ludman *et al*., 2012).

Timing of occurrence and hazard

With regard to the age of the failure events little is known so far. Only a few of the huge number of landslides recognised have reasonably accurate age determinations. The majority of the remainder are simply assigned to a geological epoch, which induces a large margin of error and makes it difficult to establish a relationship with triggering mechanisms and environmental factors. Nevertheless, it is worth mentioning the large number of events that are reported as Holocene (for example, 53 events in only the Ebro, Gulf of Lions and Ligurian margins). This fact reinforces the idea that as well as tectonic activity, climate-induced stress (e.g., sea level changes) or sedimentary load have played a major role in triggering slope failures.

Although it is accepted that submarine landslides represent a major geohazard due to their destructive power, the lack of accurate knowledge about the time of occurrence of most of the failures makes difficult to assess properly the hazard in the areas and domains described around the Iberia Peninsula. For a given distribution of failures on the seafloor, abundant but old landslides features might represent a low hazard potential, whereas a sparse distribution of features in an area of high sedimentation rates could represent a relatively high hazard potential.

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