Accepted Manuscript

Multi-event oceanic island landslides: New onshore-offshore insights from El Hierro Island, Canary Archipelago

Ricardo León, Luis Somoza, Roger Urgeles, Teresa Medialdea, Mercedes Ferrer, Ander Biain, Jesús García-Crespo, José Francisco Mediato, Inés Galindo Jimenez, Jorge Yepes, Francisco Javier González, Julia Gimenez-Moreno

PII: S0025-3227(16)30137-2
DOI: doi: [10.1016/j.margeo.2016.07.001](https://doi.org/10.1016/j.margeo.2016.07.001)
Reference: MARGO 5487

To appear in: *Marine Geology*

Received date: 21 December 2015
Revised date: 29 April 2016
Accepted date: 14 July 2016

Please cite this article as: León, Ricardo, Somoza, Luis, Urgeles, Roger, Medialdea, Teresa, Ferrer, Mercedes, Biain, Ander, García-Crespo, Jesús, Mediato, José Francisco, Jimenez, Inés Galindo, Yepes, Jorge, González, Francisco Javier, Gimenez-Moreno, Julia, Multi-event oceanic island landslides: New onshore-offshore insights from El Hierro Island, Canary Archipelago, *Marine Geology* (2016), doi: [10.1016/j.margeo.2016.07.001](https://doi.org/10.1016/j.margeo.2016.07.001)

This is a PDF file of an unedited manuscript that has been accepted for publication. As a service to our customers we are providing this early version of the manuscript. The manuscript will undergo copyediting, typesetting, and review of the resulting proof before it is published in its final form. Please note that during the production process errors may be discovered which could affect the content, and all legal disclaimers that apply to the journal pertain.

Multi-event oceanic island landslides: new onshore-offshore insights from El Hierro Island, Canary Archipelago

Ricardo León¹, Luis Somoza¹, Roger Urgeles², Teresa Medialdea¹, Mercedes Ferrer¹, Ander Biain¹, Jesús García-Crespo¹, José Francisco Mediato¹, Inés Galindo Jimenez³, Jorge Yepes⁴, Francisco Javier González¹, Julia Gimenez-Moreno¹

1 Geological survey of Spain (IGME). Ríos Rosas 23, 28003-Madrid, Spain

2 Institut de Ciències del Mar (CSIC), Passeig Marítim de la Barceloneta 37-49, 08003 Barcelona, Spain

3 Geological survey of Spain (IGME). Unit of Canary Islands, Alonso Alvarado, 43, 2ªA, 35003-Las Palmas

4 Department of Civil Engineering, IOCAG, University of Las Palmas de Gran Canaria, 35017 Las Palmas, Spain

* corresponding autor: r.leon@igme.es

Abstract

This paper presents a combined onshore-offshore morpho-structural characterization of the El Golfo giant landslide, island of El Hierro (Canary Archipelago, Spain). Offshore data from multibeam echosounders, chirp sub-bottom profiles and multichannel seismic reflection data and onshore data coming from water wells and galleries have been analyzed. The geomorphology and the internal architecture of this giant flank collapse are defined in order to determine its multi-event nature. The subaerial headscarp shows a non-continuous arcuate profile formed by two nested semi-circular amphitheatres that extend offshore along a smooth chute, suggesting the occurrence of at least two large retrogressive events. Channels/gullies and escarpments developed along the submarine sector of the scar also indicate smaller-scale events and predominance of sediment bypass. In the lower slope, two subunits of submarine mass transport deposits (MTDs), debris avalanche, are identified on multichannel seismic reflection profiles with a total estimated volume of 318 km³: ~84 km³ and ~234 km³, for the younger and older MTDs respectively. Data from wells and galleries show abrasion platforms with beach deposits at sea-level (0 masl) formed after the landslide scar and buried by the post-collapse El Golfo lavas infill, suggesting an age at least older than 23.5-82.5 Ka for the landslide.

Keywords: Submarine landslides, volcanic islands, debris avalanche, Canary Islands

1. Introduction

Flank collapse of ocean islands are among the largest gravity driven processes on Earth, mobilizing hundreds to thousands of km³ of rock and sediments. These processes are inherent to the evolution of large oceanic volcanic islands characterized by periods of rapid volcanic construction. They have been identified in intra-plate archipelagos such as the Hawaiian Islands ([Moore et al., 1989, 1994](#)), Reunion Island ([Labazuy, 1996; Oehler et al., 2008](#)) and Cape Verde ([Day et al., 1999; Le Bas et al., 2007](#)). In the Canary Islands, twenty mega-landslides have been reported so far (Fig. 1a), at less ten occurred during the Pleistocene, most of them concentrated in the western and most active islands of Tenerife, La Palma and El Hierro ([Urgeles et al., 1997, 1999; Masson, 1996; Masson et al., 1998, 2002; Gee et al., 2001; Krastel et al., 2001; Acosta et al., 2003a, 2003b; Dávila Harris et al., 2010](#)). Such flank collapses have been suggested to result from multiple processes such as oversteeping of the slopes due to volcano growth ([Murray and Voight, 1996; Carracedo et al., 1999](#)), seismicity ([Voight and Elsworth, 1997; Tibaldi, 2001](#)), dyke injection ([McGuire et al., 1990; Famin and Michon, 2010; Battaglia et al., 2011; Delcamp et al., 2012](#)) or volcanic spreading ([Van Wyk de Vries and Francis, 1997; Borgia et al., 2000; Münn et al., 2006](#)) among others. However, studies carried out in Hawaii and Reunion island ([Schiffman et al., 2006; Thompson et al., 2008; Oehler et al., 2005](#)) and more recently in Canary islands ([Seisdedos et al., 2012; Ferrer et al., 2010 and 2014](#)), show the importance that the presence in volcanic oceanic islands of a submarine thick layer of hyaloclastites plays in their flank stability conditions; the low strength and high deformability of the hyaloclastite deposits are the main key factors conditioning the failure mechanisms of volcanic flanks, once the volcanic building reaches a critical slope and height. Under these circumstances, the failure initiates in the submarine flank and progresses towards the aerial part of the island, possibly with a multiple or successive failure mechanism.

The volumes involved in the mass transport deposits (MTDs) resulting from these landslides affecting a large part of the subaerial volcanic edifices, and the inferred violent entry to the sea of the debris avalanches imply trans-oceanic tsunamigenic potential. Tsunami deposits from prehistoric mega-landslides have been described in Hawaii ([Moore and Moore, 1984, 1988; McMurtry et al., 2004](#)), Canary Islands ([Ferrer et al., 2013; Pérez-Torrado et al., 2006; Madeira et al., 2011](#)) and Cabo Verde Islands ([Paris et al., 2011](#)). In this regard, the style and phasing of volcanic island flank

collapse, whether the failure is single- or multistage (or even multi-event), retrogressive or progressive, submarine or sub-aerial and disintegrative or not (Harbitz et al., 2006, Ward and Day, 2001; Watt et al., 2012a, 2012b) has critical implications for tsunamigenesis. Particularly, the single- vs. multi-staged and the retrogressive character of island flank collapses (García, 1996, Masson et al., 2002; Hunt et al., 2013) control the volume involved in a landslide and/or the duration and timing over which such volume is mobilized, which is a first order control on the initial tsunami wave amplitude (Murty, 2003; Harbitz et al., 2006). The aim of this study is to bring new light on the failure dynamics and mode of emplacement of a recent major landslide on the volcanic island of El Hierro (Canary archipelago): El Golfo. To achieve such understanding, we analyze both the onshore and offshore geomorphological features of the El Golfo landslide, accurately determine its geometry, internal architecture and volume of the debris avalanche deposits.

2. Geological and geomorphological settings

The Canary archipelago is a NE-SW 1350 km long ridge of volcanic islands and seamounts (from Lars to Tropic seamounts) located 100 km off the northwest African coast (Fig. 1a). The origin of the Canary archipelago is related to a mantle plume (hotspot). A general age progression is observed from the oldest Lars volcanic seamount (68 Ma), to the westernmost La Palma and El Hierro islands, which are both younger than 2 Ma (Geldmacher et al., 2005).

El Hierro is the youngest and most southwesterly of the seven Canary Islands. The oldest subaerial rocks have been dated at 1.12 Ma (Guillou et al., 1996; Carracedo, 2011). The volcanic shield rises about 5000 m from its base on the western flank and reaches 1503 meters above sea-level (masl) at the Malpaso summit. It has an estimated edifice volume of 5500 km³ (Schmincke, 1994). Currently, the island is at a young (shield) stage, characterized by alternating eruptive episodes and major destructive events (Carracedo et al., 1999).

The volcanic edifice is located on Jurassic oceanic crust of about 156 Ma (Roeser, 1982). The sedimentary sequence west of the island has been divided in three units according to their relation to volcano-related processes on the archipelago (Urgeles et

al., 1998). Unit I, the lowermost unit, is a 1-km thick low reflectivity unit interpreted as pre-volcanic sediments. The uppermost ~750 m of sediments form Units II and III and display higher reflectivity. Unit II is associated with distal products of erosion on the older eastern islands, while Unit III represents the most proximal deposits resulting from the erosion of La Palma and El Hierro islands.

El Hierro Island has been considered as a typical example of a three-armed rift system (Fig. 1b), arranged at 120°, radiating from the centre of the volcanic edifice. The rift system is evidenced by the distribution of eruptive centres and dyke network pattern (Fuster et al., 1993; [Navarro and Soler, 1994](#); [Carracedo, 1994](#); [Carracedo et al., 2001](#)).

Three eruptive episodes have been recognized onshore El Hierro Island ([Carracedo et al., 2001](#); [Bellido-Mulas et al., 2012](#)), based on radiometric and paleomagnetism dating ([Guillou et al., 1996](#); [Szérmétya et al., 1999](#)) (Fig. 1b). (i) The Tiñor Edifice (1.2-0.88 Ma) represents the first stage of subaerial growth of the island and outcrops in the NE sector and along the Las Playas scarp; (ii) The El Golfo edifice (0.54-0.176 Ma) fills the depression left by the collapse of the Tiñor edifice and also covers the flanks of the old edifice ([Carracedo et al., 2001](#)); (iii) The Rift Volcanism (0.158 Ma-present), when the island acquired its pseudo-triangular appearance and developed the El Golfo and Las Playas flank collapses. Lithostatic decompression caused by removal of volcanic material by the El Golfo landslide induced a new eruptive phase, post-collapse El Golfo infill volcanism, and lava flow emissions (240 m thick) towards the coast ([Navarro and Soler, 1994](#); [Carracedo 2011](#)). Finally, the most recent emissions are the possible eruption of Lomo Negro in 1973 ([Hernández-Pacheco, 1982](#)) and the submarine eruption of La Restinga in 2011 ([López et al., 2012](#); [Somoza et al., 2012](#)).

2.1. Landslide events

At least five major landslides (Fig. 1b) have been recognized in the El Hierro Island both from onshore geological evidence ([Carracedo et al., 1999, 2001](#)) and offshore deposits ([Holcomb and Searle 1991](#); [Urgeles et al., 1997](#); [Masson et al., 1998, 2002](#); [Masson, 1996](#); [Gee et al., 2001](#); [Krastel et al., 2001](#); [Acosta et al., 2003a, 2003b](#)).

The oldest event (dated at ~882 ka) involved the eastern sector of the Tiñor edifice and generated the unconformity at the base of the El Golfo edifice ([Carracedo et al., 1999](#),

2001). However, the submarine deposit associated with this flank collapse has not yet been identified ([Masson et al., 2002](#)).

The El Julán landslide (dated at least at 160 ka old; [Guillou et al., 1996](#); [Carracedo et al., 1999](#)) affected the southwest flank of the El Golfo edifice and formed a 15 km wide embayment. The associated submarine deposits have been interpreted as a slump ([Gee et al., 2001](#)) or a complex slump/debris avalanche deposit ([Krastel et al., 2001](#)).

The Las Playas I (545-176 ka) and Las Playas II (176-145 ka) landslides are located on the southeast flank of El Hierro Island ([Masson et al., 2002](#)). Onshore, seaward-dipping normal faults parallel to the shoreline with a downthrow up to 300 m have been associated with the Las Playas I headwall scar ([Day et al., 1997](#); [Klimes et al., 2016](#)). These faults have been interpreted as an aborted flank collapse ([Day et al., 1997](#); [Carracedo et al., 1997, 1999](#); [Krastel et al., 2001](#)), a broad slump analogous to those identified in the Hawaiian islands ([Gee et al., 2001](#); [Masson et al., 2002](#)), and a deep-seated gravitational slope deformation ([Klimes et al., 2016](#)). A narrow (5-km wide) embayment onshore surrounded by ~900 m high cliffs has been identified as the Las Playas II headwall scar. Offshore, a chaotic debris avalanche deposit appears to be superimposed on a wider area of deformed strata according to seismic reflection data ([Gee et al., 2001](#); [Krastel et al., 2001](#); [Masson et al., 2002](#)).

The El Golfo landslide is the most recent flank collapse event of the whole archipelago ([Masson et al., 2002](#)) with recent studies indicating an age between 87 ± 8 and 39 ± 13 ka ([Longpré et al., 2011](#)). The El Golfo embayment was originally interpreted as the remains of a great central caldera ([Fernández-Navarro, 1908](#); [Blumenthal, 1961](#)) and later as a tectonically-controlled erosive structure ([Pellicer, 1977](#)). [Hausen \(1972\)](#) was the first to identify the El Golfo embayment as a flank collapse, while [Masson \(1996\)](#) and [Urgeles et al. \(1997\)](#) mapped the submarine El Golfo debris avalanche (EGDA) deposits. Onshore, the landslide is characterized by a headwall scarp up to 1300 m high ([Carracedo, 2011](#)).

3. Data acquisition and methods

Onshore, a 5-m DEM derived by the Spanish Geographic Institute (IGN) from LiDAR measurements (CNIG, 2012) was used, along with an inventory of data from water

wells and galleries in the El Golfo embayment (IGME, 2012). Offshore, multichannel seismic reflection (MCS) profiles (IGN, 1991), high-resolution sub-bottom profiles (HRS) and multibeam echo-sounder (MBES) data (bathymetry and backscatter) have been used in this study (Fig. 2).

The catalogue of water wells in the El Golfo embayment contains 45 entries, including 30 boreholes, 3 galleries, 5 wells and 7 wells with galleries. Most of them are located in the eastern half of the embayment. They provide detailed information on the evolution of groundwater levels and water chemistry, as well as lithological logs of many of the water wells and their constructive characteristics. The available information has been contrasted with other hydrogeological studies carried out in the vicinity of El Golfo (MOP, 1975; Bravo, 1982; Navarro and Soler, 1994; IGME, 1990) and fieldwork.

MBES data were acquired with a Simrad EM120 and an ATLAS Hydrosweep DS on the *R/V Hespérides* and the *Sarmiento de Gamboa*, respectively. The Simrad EM120 system operates at a frequency of 13 kHz, has 191 beams and a seafloor swath width of 3.5 times the water depth. The ATLAS Hydrosweep DS system operates at a frequency range of 14-16 kHz with a maximum of 345 beams spread over 140°. Bathymetric and backscatter data were processed using the CARISTTM software. Periodic CTD measurements and SIPPICAN XBT casts were made to determine the vertical temperature of the water column (precision ± 0.15 °C) so as to correct for sound refraction problems in oblique beams.

Chirp sub-bottom profiles were acquired using ATLAS Parasound P35 onboard the *R/V Sarmiento de Gamboa*. The ATLAS parametric sounder P35 operated on a secondary lower frequency of 3.5–5 kHz, with a shot interval of 4-10 s, a recording length of 0.17-25 ms and a sampling frequency of 12.2 kHz.

MCS data were obtained in 1988 (CANARIAS_88 cruise) with the Norwegian *R/V TAU* and with *R/V Sarmiento de Gamboa* in 2011 (GAIRE cruise) during the volcanic crisis of La Restiga, El Hierro island. During CANARIAS_88 cruise, the system configuration included an array of airguns with a total capacity of 103 l, which were shot every 50 m. Data acquisition was performed using a 3000 m long streamer with 120 channels. During the GAIRE cruise, the system configuration included two arrays of airguns with a total capacity of 75 l, which were shot every 50 m. Data acquisition

was performed using a 3500 m long streamer with 280 channels. Seismic profiles were imported into Kingdom Suite™ software for visualization and interpretation.

Volume calculation has been developed with the “discrete smooth interpolation” algorithm of Paradigm® SKUA-GOCAD 15.5™.

4. Results

4.1. *Geomorphology of submarine flanks*

The morphological description of collapsed and non-collapsed flanks is essential for the topographic reconstruction of the insular edifice prior to the El Golfo landslide, in order to understand the failure mechanisms, the volume of mobilized material and related hazards. The clearest examples of non-collapsed flanks are the submarine prolongations of the NE and NW rifts that enclose the El Golfo embayment (Figs. 3 and 4). These rift systems and their offshore prolongation border three major collapses: the El Golfo, El Julán and Las Playas. A smaller one, Punta de El Hierro is somewhat unusual as it is located right on the offshore extension of the NE rift (Fig.3). The Punta de El Hierro collapse (Fig.4) has a small headwall scar of 1.5 km long partially covered by Holocene basaltic flows. Offshore it forms a 1.5 km wide, 8 km long, NE-SW oriented chute.

The width of the coastal platform around the El Hierro Island varies from 200 m in Las Playas to 1500 m northeast of Lomo Negro (Fig. 4). The slope of this platform ranges from 0° to 5° and has a smooth and even surface. Beyond the edge of the platform, the slope increases sharply, reaching values higher than 25° (Fig. 5). Around Roques de Salmor, northeast of El Golfo embayment, the edge of the platform shows a concatenated arcuate perimeter mapped as small headwall scars (Fig. 4).

Non-collapsed flanks have an average slope of 14° at 2000 m water depths decreasing to 7° at water depths from 2000 to 3000 m. There is a clear base of the insular slopes at 3000 - 3250 m water depth, where the maximum change in slope takes place (*eg.* the maximum value of the second derivative) and the gradients quickly decline to ~3° (Fig. 4). At this depth the volcanic edifice grades into the marine sedimentary deposits of the NW African continental margin. The base of the insular slope in flank collapse areas is characterized by a concave to convex trend change. Above, the base of the insular slope the isobaths bend landward, while below they bend seaward. This pattern is the result of the passage from landslide erosion to landslide accumulation areas (Fig. 4). Otherwise,

flanks unaffected by landslides (represented in grey in Fig. 3) are characterized by closely spaced, radially aligned volcanic cones (Fig. 4). These volcanic cones range from 150 to 2000 m in diameter and are up to 300 m in height from their base.

On the SE and NW flanks at a water depth between 2000 and 3400 m, just before the base of the volcanic edifice, a series of lobe-type seafloor features are identified (Fig. 3). These lobes have arcuate fronts that reach heights of 100-300 m and slopes around 20-40° (Fig. 3). A field of wavy bedforms parallel to the isobaths (Fig. 3) is also present west of the volcanic edifice, below 3400 m water depth. The absence of bathymetric data impedes the estimation of the westward extension of these structures. They have amplitudes between 10 and 100 m with a mean wavelength of 2 km. Their crests are sinuous in plan view and can be laterally spread approximately 15 km (Fig. 3).

4.2. Main headwall area

The coastal morphology of El Hierro Island is characterized by a well-defined major amphitheater-shaped structure that corresponds to the main scars of the El Golfo landslide (Fig. 6a). The head scarp of the main recent landslide creates an arcuate rocky wall enclosing the El Golfo embayment. Such embayment is 15 km wide (extending from Roques de Salmor to Punta de Arenas Blancas in Fig. 6a) and can be divided into three main morphological sectors:

a) The eastern sector extends from Roques de Salmor to Fuga de Gorreta (Fig. 6a), following a slightly curved NE-SW trend with an average slope of 45° and a maximum elevation of 1200 m. A series of subvertical dykes parallels the scarp with an average N65°E orientation. Within 150 m to the north of Fuga de Gorreta, the dykes are subvertical, resembling a coastal cliff.

b) The central sector extends south of Fuga de Gorreta to the Malpaso peak (1500 m high). Here, the scarp changes its direction and dips westwards defining a curved path or a smaller nested amphitheater (Fig. 6a). The dykes outcropping in the eastern part of this sector maintain the N65°E orientation. Erosion of weaker material around the dykes has left them as prominent ridges. At the base of the escarpment, there is a 15-20 m thick group of heterometric detrital deposits corresponding to post-collapse piedmonts have been over a

slumped block (faulted and rotated) belonging to El Golfo landslide (Fig. 7b). The western part of this sector shows few N110°E-N140°E dykes along the scarp (according to [Becerril et al., 2015](#)), since basaltic tuff cones and lavas cover the base of the escarpment and the scar is only visible in the upper portion, where the slope averages 45°. This central sector appears to be indented within an already existing amphitheatre.

c) The western sector extends from the Malpaso peak to Punta de Arenas Blancas, with a NW-SE curved trace (Fig. 6a). To the west of Malpaso peak, the Tanganasoga volcano (1370 m high) covers most of the slip plane. From Sabinosa to Punta de Arenas Blancas a dyke network also outcrops with an orientation roughly parallel to the scarp (N114°E, according to [Bellido-Mulas et al., 2012](#); [Becerril et al., 2015](#)).

4.2.1 Onshore subsurface characterization

In the Canary Islands, mainly Tenerife, groundwater is extracted from wells (vertical) and galleries (horizontal) excavated on the island flanks, taking advantage of the plastic chaotic brecciated material found on the landslides basal surface, underlying the post-landslide volcanic materials, which acts as an impervious boundary. We have used the valuable lithological information from some wells/galleries in El Hierro, included in the water wells database of the Canary Islands (IGME, 2012) to define the geometry of the onshore scar of the El Golfo landslide in areas where post-collapse lava episodes conceal such scar. Beside, information about sedimentary bodies (piedmont or beach/marine levels) over the landslide scar and structural measures (strike and dip of dykes) inside the El Golfo Edifice near the head-escarp are described below and summarized in Table 1.

The Las Casitas water well-gallery is located to the southwest of Las Puntas at the bottom of the El Golfo scarp (Figs. 6a and 7). It is a 60 m deep well with a N150°E trending 170 m long gallery at aprox. 1.5 masl. Both the well and the first 50 m of the gallery cut through the post-collapse El Golfo lavas. At this point (50 m), a thin beach level is found with well-rounded pebbles (30-40 cm in diameter). El Golfo Edifice appears after the beach level. Four dykes have been identified until 70 m, two with N25°E strike and 30-40° dip, the other two dykes are sub-vertical and N65°E strike.

The Los Padrones water well-gallery is located 1 km to the south of Las Casitas (Fig. 6a). The base of the well is 51.5 from surface (at 0 masl) where a N110°E trending horizontal gallery starts that truncates to N130°E after 160 m and extends over 1000 m. The first 195 m of the gallery are post-collapse El Golfo lavas. The contact with the El Golfo Edifice is nearly vertical. Two dykes with N65°E strike have been identified.

The Tincos gallery is the largest in the island. It is located 320 masl, it is 1770 m long and trends N184°E. It cuts materials belonging to the El Golfo Edifice and seventy-six dykes with N40°E and N115°E strike (Fig. 6b and 7). Near the end of the Tincos gallery, a net of E-W trending dykes has been reported.

The Charco de Tincos gallery is located 2 km to the east of Los Padrones (Fig. 6b), at about 550 masl, on the El Golfo escarpment inside El Golfo Edifice. It is 280 m long trending N155°E. A net of 26 dykes with N75°E strike has been identified.

The Tigaday water well-gallery, close to Frontera (Fig. 6a and 7), is a 280 m deep well with a N215°E horizontal gallery 160 m long at -5 masl. The well passes through 200 m of the post-collapse El Golfo lavas. The following 75 m is a piedmont. In the last 5 m, the well crosses highly fractured basalts corresponding to the El Golfo edifice. In the gallery, the basalts are fractured and cut by 9 dykes with ENE-WSW direction.

The Los Llanillos water well-gallery (Fig. 6a and 7) is a 178 m deep well with a horizontal gallery 680 m long at sea level (0 masl) trending N125°E in its first 50 m and N165°E from that point to the end. The well is all excavated inside post-collapse El Golfo lavas, as well as, the first 125 m of the gallery. The following 25 m (from 125 to 150 m) show a coastal gravel deposit, with rounded and polymictic pebbles embedded in a sandy matrix. From 150 to 370 m, the gallery goes through a basaltic lava flow and from 370 to 500 m, followed by another coastal deposit with well-rounded fragments (30 cm) in the last meters. From 500 to 515 m there is a basaltic lava flow similar to the previous one and again, from 515 to 525 m, a new gravel deposit with boulders (1.5-2 m) embedded in sand. From this point to the end of the gallery, only lavas and two N45°E dykes related to El Golfo Edifice are present.

The Chirgo water well-gallery (Fig. 6a) is a 264 m deep well with a N200°E horizontal gallery 30 m long at sea level (0 masl). The first 240 m of the well are crossing the post-

collapse El Golfo lavas. The next 22 m (till 3 masl) are made up of a piedmont deposit over a beach level located at 0 masl. The gallery goes inside the piedmont deposits.

The Lomo de Trébol gallery (Fig. 6b) is located 700 masl, at the base of the escarpment of the El Golfo embayment. It is 300 m long, trends N235°E and cuts entirely across El Golfo Edifice materials. A group of forty-three N110°E dykes and fractures has been reported along the first 150 m of the gallery.

Coastal wells (Fig. 6a), majority of them are about 140 m deep and with the base at about -15 masl, cross post-collapse El Golfo lavas. No dykes or beach levels have been reported.

4.3. Submarine geomorphological features

Two main sectors are identifiable within the submarine El Golfo landslide: a) the chute sector which is the pathway for the mass failure; and b) the MTD lobe sector. The chute sector extends seawards from the coastline for ~6 km following a NW direction. Well-developed sidewall scarps are laterally continuous with the onshore headwall structures (Fig. 8a). The main sidewall scarps cut back to 700 m water depth with slopes up to 20°. They are 900 m high on the south-west side of the Roque de Salmor (Fig. 8c). From this point, the scarp height decreases to 300 m at 1500-2000 m water depth and it completely disappears at 3000 m water depth. The uppermost area of the chute sector shows several V-shaped gullies with arcuate headwalls ~1000 m in diameter (Figs. 8 and 9). The base of these gullies, located at 900-1800 m water depth, is marked by an abrupt decrease in the slope from 30-40° to 12-10° and by the occurrence of fan-shaped sedimentary deposits. These fans are approximately 3 km long and 1.2 km wide. They display transparent internal seismic facies and a smooth upper surface reaching a thickness of ~40 ms Two-Way Travel time (TWT), corresponding to ~35 m (assuming a velocity of 1750 m/s for water-rich sediments) (Fig. 9c).

Within the chute, five major feature types are identified (Figs. 9 and 10): steep sided crests and ridges, central lobe/smooth sedimentary ridge, channels, lateral debris lobe-type and allochthon blocks.

The steep-sided crests and ridges are likely constituted by volcanic material (Fig. 9b). The westernmost volcanic crest has a NNE-SSW direction and extends 3.5 km seawards until 1000 m water depth from the coast east of Sabinosa, where it is partially buried by

the post-collapse volcanism. These volcanic crests appear as elongated features up to 120 m high with lateral slopes up to 25° . Most of these volcanic crests display low backscatter values (-17 to -27 dB; Fig. 8c) and wide hyperbolic echoes on HRS profiles (Fig. 9b).

The middle area of the chute is characterized by an elongated central lobe extending from 1000 to 2800 m water depth (Figs. 8a and 9a). Although some avalanche deposits overlay the central lobe, its proximal area, from 700 to 2900 m water depth, has mainly an erosive character (Fig. 10). The central lobe shows relatively smooth relief, high backscatter strength (-10 to -12 dB; Fig. 8e) and hyperbolic diffractions (Fig. 9). The slope along the chute gradually decreases from 10° at 300 m to 5° at 1500-2000 m water depths (Fig. 5).

Seawards, the central lobe presents smooth sedimentary (likely volcanoclastic) ridges (Figs. 8 and 10). These smooth ridges are elongated erosional features, and have terraced cross-profiles produced by the incision of channels on their flanks. The sedimentary ridges show opaque acoustic facies with hyperbolic diffractions (Fig. 9a). Several blocks, up to 250 m in diameter and belonging to the debris avalanche deposit, can be distinguished on the sedimentary ridges.

Erosive channels run downslope at water depths from 50 m to 3000 m, majority of them range from 500 to 2000 m in width and from 2 to 7 km in length (Figs. 9 and 10). They are mainly located at the base of the scar region and at the margins of the chute and display high backscatter values (-8 to -10 dB) and opaque acoustic facies with multiple small hyperbolae in HRS profiles (Fig. 9a and 9b). They are constrained by gentle escarpments. The largest channel (~7 km long) is located on the western side of the chute. The narrower channels show incised V-shaped profiles, whereas the wider channels have flat-bottomed cross sections. A central “U-shaped” channel, 4.5 km wide and 80 m deep has been identified (Fig. 9b). It shows high backscatter strength (-12 to -14 dB) and goes through the debris avalanche deposits at water depths from 2000 m to 3000 m (profiles c, b and e in Fig. 10) generating a negative relief. This central channel ends at about 3400 m water depth forming a secondary lobe. Smaller scattered blocks (up to 550 m in diameter), partially buried by sediments, are located in the prolongation of the channel at 2900 and 3300 m water depth within the landslide deposit (Fig. 8). We

infer that this central channel is the main path for sediment transport sourced from the island as minor landslides and flows.

Lateral debris lobe-type deposits have been identified between 500 and 1200 m water depth (e.g. Fig. 10e). These fan-shaped mass-transport deposits seem to have mostly derived from the gullies located on the SW arcuate scars of Roques de Salmor and to the north of Punta de las Arenas Blancas (Fig. 8). The gullies originate in water depths between 20 and 50 m and are related to gravitational instabilities. Partially eroded arcuate scars are also identified at depths between 100 and 600 m associated with fan-shaped deposits (Fig. 8a). These lobes likely represent lateral debris avalanche deposits along the slopes of the main headwall.

Allochthonous blocks are interpreted to be large irregular fragments of volcanic rock masses involved in the failure of the volcanic edifice and transported within debris avalanche events. They first appear at water depths between 3000 and 3200 m, at the transition between the proximal and the distal sectors of El Golfo landslide (Fig. 8). In the distal sector, the lateral headwalls disappear and the debris avalanche deposits spread radially over a 1° smooth slope, as they are no longer confined by the headwalls. Nearly a hundred randomly distributed blocks have been recognized within this sector of the El Golfo landslide (Fig. 8a). They are characterized by an irregular positive relief and relatively high backscatter values (-10 to -11 dB) (Figs. 8d and 8e). Their size ranges widely from 200 m to 1.5 km in diameter, with average diameters of 500 m. Most blocks are 50 m high and concentrate in the distal sector of the debris avalanche. The largest blocks, reaching 1.5 km in diameter and heights up to 300 m, are located in the central sector at water depths from 3300 to 3700 m. On the contrary, the smallest blocks have been identified in the upper-central sector of the deposit, between 3000 and 3300 m water depth. In this sector, blocks display an average diameter of 350 m and heights of 10-12 m. Nevertheless, some scattered blocks are also present in relatively proximal parts of the chute.

4. 5. Seismic architecture and depositional features of the submarine MTDs

Based on MBES (bathymetry and backscatter) and seismic data, two main morphological units (upper lobe and main body) have been identified as depositional

units ranging from 1000 to 4000 m water depth (Fig. 8a). The upper lobe unit initiates at the end of the central channel crossing the scar region from 2900 to 3500 m water depth. It is a fan-shaped feature displaying lower backscatter values (3-4 dB of difference) than the surrounding area. This lobe fades into several branches surrounded by large blocks (>700 m in diameter) at 3500 m water depth. The lobe has an upper smooth surface bearing fewer and smaller scattered blocks, 250 to 550 m in diameter, that seem to be partially buried in sediments. The so-called main body unit is directly related to the debris avalanche deposit and, offshore, has been discriminated based on the interpretation of MSC profile B1 acquired during the CANARIAS_88 cruise (Fig. 11) and SM1 and SM2 acquired in 2011 onboard *R/V Sarmiento de Gamboa* during the volcanic crisis in the El Hierro Island. These profiles cross longitudinally and transversally the El Golfo landslide MTDs. Westwards, the seismo-stratigraphic units I, II and III described by [Urgeles et al. \(1998\)](#) are distinguished overlying the acoustic basement (Fig. 11a). These units are wedge-shaped, increasing in thickness towards the flanks of the islands. Below the El Golfo landslide MTDs, the reflections belonging to these units fade out due to increased signal attenuation, possibly caused by the presence of chaotic volcanic material, fluids, dykes or sills.

The main body unit is characterized by an upper irregular morphology bounded to the east by a 120 m high headwall scarp that extends 6 km in a NE-SW direction. This headwall scarp has been previously interpreted as a possible relict scarp of the Canary Islands Debris Flow (CDF, [Urgeles et al., 1997](#); [Masson et al., 1998](#)). The western boundary of the avalanche is not well differentiated because of the high degree of material deformation and faulting (Fig. 11a). Two main MTD subunits can be differentiated within the main body:

MTD A

The MTD A is the shallowest unit. It is characterized by high-amplitude and discontinuous reflections (Fig. 11a). Hyperbolic diffractions are also observed, which could indicate the presence of buried blocks (Figs. 11b and 12c). The base is represented by a group of continuous reflections at ~220 ms TWT below the seafloor (a thickness corresponding to ~192 m assuming water-rich deposits with a mean wave velocity of 1750 m/s ([IGN, 1991](#))). These basal reflections extend laterally about 14 km

from the NE-SW headwall scarp to a large, 150 m high, block (Fig. 11b). However, from this block to ~10 km westwards, the basal reflectors are not so clear, possibly due to masking by the block or to erosion by the emplacement of MDT A. In the proximal area, it has 25-70 ms TWT in thickness (Fig. 12c) and its base fade with hyperbolae.

In the chute – proximal area (Fig.12c), MTD A shows a variable thickness ranging from 50 to ~200ms TWT. Besides, its base fades (Fig.12b) because of the intensive fluid flow during the volcanic crisis in November, 2011. In the upper area of the SM1 seismic profile, seafloor and MTD A are lightly uplifted, drawing possible initial surfaces of rupture.

MTD B

The MTD B is characterized by discontinuous and wavy reflections interbedded with transparent acoustic areas (Fig. 11b). Compared to MTD-A, the boundaries of the MTD B are more clearly marked in the eastern sector. This unit shows a maximum thickness of 170 ms TWT (corresponding to 148 m at 1750 m/s). The basal reflection located at 440 ms TWT below the seafloor is rather continuous and displays amplitude anomalies (Fig. 11b). These acoustic anomalies may be due to the presence of fluid rich sediments confined by the avalanche deposit. These fluid rich sediments correspond to the seafloor (or near to the seafloor) sediments before the avalanche took place. The interstitial water in these fluid-rich sediments remained at depth after avalanche emplacement. Alternatively gas-charged emissions from El Hierro could be trapped in the lower portion of the debris avalanche.

The base of MTD B pinches out 15 km further west than that of MTD A. The westernmost part of MTD B, from the seafloor to 270 ms TWT, consists of continuous and deformed high-amplitude reflections. They are interpreted here as caused by the remobilization and incorporation of ocean sediments due to the rapid emplacement of the avalanche deposit in a similar way to that proposed by Brunet et al. (2015). The deformation reaches the seafloor generating a wavy morphology, where several blocks are also recognized (Fig. 11b).

The western boundary of the debris avalanche deposit is defined by the presence of surface blocks. However, the area affected by the emplacement of the avalanche is larger and marked by faulting that makes its boundary unclear. The area affected by the

debris avalanche is also characterized by several faults, some of which also disrupt the debris avalanche deposit. These faults can be traced down to depths of 600 ms TWT (525 m at 1750 m/s) below seafloor, affecting the units II and III defined by Urgeles et al. (1998). The largest structure generates a 120 m high positive relief and its crest extends N-S along 5 km (Figs. 3 and 11b).

In the SM1 MCS profile (Fig. 12b), the base of MTD B is cut by MTD A at about 2200 m water depth (~ 3 s TWT). Besides, the base of MTD B shows, in this profile, weak and non-continuous reflections mainly due to the intensive fluid migration during the volcanic crisis in November, 2011. In the middle and to the East of the MCS profile SM1, transparent seismic facies and amplitude anomalies (red arrows in Fig.12b) are found. Just in this area, the epicenters of shallow earthquakes (< 3 km) are located, near of pre-existing or incipient fractures that might control the MTD B thickness.

5. Discussion: nature, origin and multievent evidences

It was previously hypothesized that the El Golfo debris avalanche occurred as a consequence of a single major catastrophic event (Urgeles et al., 1997; [Masson et al., 2002](#)). However, the analysis of seismic profiles and the geomorphological features identified on-shore and off-shore point to at least two main large landslides events taking place.

5.1. Landslide-scarp area, morphological features

The most relevant geomorphological feature in the on-shore area is the steep headwall scarp. This subaerial arcuate scarp surrounding the El Golfo embayment is a composite scarp constituted by two superposed amphitheater headwalls (near Fuga de Gorreta and Malpaso in Figs. 4, 6a and 8). This morphology suggests a first event generating the main and larger headwall scarp and a subsequent smaller episode, which formed the second scarp and enlarged landward the El Golfo embayment.

This multi-event nature of the El Golfo landslide is also supported by both subaerial and submarine morphologies of the landslide scar area. Although post-collapse lavas filled the slip area in the El Golfo embayment, the headwall scars present spatial continuity

landward. Offshore, lateral escarpments of the chute are the landward prolongation of the headwall scarps in the Roques de Salmor and Punta Arenas Blancas. Besides, the landward prolongation of the second superposed amphitheater headwall from Fuga de Gorreta and Malpaso matches with the second lineal escarpments inside the chute (purple escarpments in Fig. 8). In this way, we suggest that the sedimentary ridges identified on the offshore uppermost sector of the El Golfo landslide scar might represent levees formed during transport of the secondary debris event (Fig. 8).

The lateral limits of the subaerial huge headwall scarp seem to be controlled by the networks of dykes emplaced according the structural configuration of the island. Along the El Golfo scarpment, the orientation of the dykes identified in the galleries (Fig. 6) agrees with the triaxial rift configuration of the island, and probably exerted a structural control of the landslide. The network of N110°E dykes and fractures in the western sector (Lomo de Trébol) and of N60°E dykes in the eastern sector would have determined the location of the landslide limits and the arcuate-shape of the head area.

The multi-event character of the El Golfo flank collapse could also explain the controversial relative age of the submarine NE-SW scarp (Figs. 8 and 11b) to the east of the EGDA, related to headwall scar of CDF (Urgeles et al., 1997). A series of works identified the CDF as an event synchronous to the debris avalanche ([Masson, 1996](#); [Masson et al., 1998, 2002](#); [Canals et al., 2004](#)); the idea was that excess pore pressure induced by sudden emplacement of the EGDA on the continental margin sediments might have triggered a giant debris flow such as the CDF (Roberts and Cramp 1996; Urgeles et al., 1999). Because of the multi-event character of the El Golfo flank collapse, the scars and scarps related to the CDF could also be buried by younger debris avalanche events. The idea that the CDF took place before the El Golfo landslide could be, in a multi-event context, justified because the headwall scarps of CDF are covered and/or eroded by the EGDA (Urgeles et al., 1997).

Regarding the load by the EGDA as the trigger mechanism for the CDF, it has been shown in other contexts that debris avalanches may involve the deformation of basinal sediments, resulting in a significant increase of the landslide volume ([Crutchley et al., 2013](#); [Watt et al., 2012a, 2012b](#)). In these cases, the seafloor expression of the landslide scars has been clearly recognized and, the transport of the sediment incorporated into the induced minor or secondary landslides is restricted to a few tens of kilometers (Watt

et al., 2012a). However, the case of El Golfo landslide is quite different: the MTDs from the El Golfo landslide and the CDF are clearly differentiated (Fig. 3) and this latter, acquires great mobility translating along slopes of less than 1° over 500 km.

5.2. Depositional units: nature and multi-event character

Offshore, three sedimentary features have been detected: lateral debris lobe-type, upper lobe and main body (Fig. 8).

The main body is a MTD, debris avalanche deposit, directly related to the El Golfo flank landslided. The identification of a group of continuous reflections in the MCS profiles (Fig. 11b) located within the main body of the debris avalanche (between MTDs A and B) indicates that a lapse of unspecified time occurred between the emplacement of both landslide deposits. Besides, SM2 MCS profile (Fig. 12c) shows as the base of lateral debris lobe-type could be interbedded with the base of MTD A. This clearly evidences the occurrence of at least two large superposed recent landslides.

The multi-event character of the El Golfo landslide was already suggested based on cores collected in the Agadir Basin (Wynn and Masson, 2003). They distinguished three subunits within the volcanoclastic turbidite “b”, which is apparently related to the formation of the El Golfo landslide (Masson, 1996), deposited in a relatively short time span with no pelagic/hemipelagic sedimentation in between. The occurrence of these subunits within this turbidite has been interpreted as an evidence of a multistage collapse (possibly retrogressive) associated with a single collapse event (Wynn and Masson, 2003; Hunt et al., 2013). A similar interpretation has been proposed by García (1996) for landslides on the Hawaiian Islands based on the relationship between landslide units interbedded with turbidite deposits. Based on our work, we suggest that the El Golfo debris avalanche resulted from at least two large flank collapses separated by an unknown amount of time (Fig. 11), as attested by the set of continuous reflectors separating MTD A from MTD B. Due to the still outstanding morphological features onshore and offshore created by the (at least) two events, we can also state that the time lapsus between them should not be long, probably some few thousands of years.

The time elapsed between MTD A and MTD B is difficult to calculate without borehole data. Firstly, because the set of reflections between MTDs A and B present an uncertain

nature. They could represent a period of pelagic sedimentation, pyroclastic flow, or that the base of the MTD A is showing on seismic profiles high amplitude events mainly due to an increase in density due to increased friction and compaction at the time of deposition (Moscardelli et al., 2006). In this way, taking into account the quite variable sedimentation rates on the submarine volcanic flanks (ranging from 3 cm/ky to 110 cm/ky; Le Friant et al., 2004), the set of reflections between MTDs A and B could represent several thousand years in the case of pelagic sedimentation. In the case of being a pyroclastic flow deposits emplaced during an eruptive activity accompanying or not the landslide, then the time span might be shorter. Or they may be geologically simultaneous, in the case of being high amplitude seismic events in the base of MTDs. According to the elapsed time observed from distal turbidites (hours to months; Wynn and Masson, 2003), the above continuous reflectors could both represent rapid sediment gravity flows, or high amplitude seismic events in the base of MTDs.

On the contrary, the difference in the number of events between the Madeira abyssal plain (three) and that at the base of the islands (two) could then point to a complex nature of the gravitational movements in volcanic islands. Fast landslides evolving along the island flank could bypass the upper slope region generating deposition only on the distal parts of the slope and on the abyssal plain.

Blocks inside debris avalanche deposit show differences in size and spatial density (Fig. 8). In the MTD B blocks are smaller and are more scattered (~ 0.5 km in average) than in the MTD A (~ 1 km in average), and rarely existent over the upper lobe. Block size, or the degree of disintegration, depends on the type and mechanism of failure, transport dynamics, nature of the source area, slide velocity and the shape and width of the chute where the blocks will be forced to go through, interact and crack (Hurlimann et al., 1999; Mitchell et al., 2002; Moore et al 1994). Considering that the source of the deposits is practically the same in both MTDs and that the difference in size does not reflect difference in the composition of the source area (Masson et al., 2002); we argue that this difference in size and density could point to differences in the geometry of the landslide (longitude of the headwall scar respect to the width of the chute) of the two events related to MTD A and MTD B. In this way, the higher degree of disintegration in the MTD B could be related to the fact that the width of chute is narrower than the headwall amphitheater. On the other hand, the blocks could be bigger in MDT A due to

the width of chute being similar to its related secondary superposed amphitheater headwall (Fig. 8).

Younger instabilities and sedimentary units described in this paper, lateral debris lobe-type and upper lobe, are thinner and cover a smaller cartographic surface than the debris avalanche deposits, reflecting a clear energy decrease.

Lateral debris lobe-type deposits are depicted in the proximal area at ~1000 m water depth (Figs. 8, 9 and 12c). They are recent mass transport deposits directly related to the activity of gullies and instabilities derived from arcuate scars in the edge of the platform. These lateral submarine landslides could be one of the possible sources of deposits of the upper lobe, transported along the channels of chute. These channels are active and show a clear erosive character with seismic hyperbolae and high backscatter value.

Upper lobe is located at the end of the central channel with two possible origins related to flows through this channel. It could be linked to the sediment transfer from the proximal (gullies and lateral debris lobe-type) to the distal areas and also could be active at present. Instead, it could be related to pyroclastic flows or volcano-sedimentary deposits linked to the post-collapse El Golfo infill eruptions. These large volcanic eruptions, that occupied the offshore area with possibly submarine events, could generate seaward a huge pyroclastic fan, similar to the lower apron detected during the La Restiga volcanic crisis in 2011 (Somoza et al., 2012).

During the volcanic crisis in La Restinga in 2011, several earthquakes were registered in the El Golfo area (Fig. 12a). MCS SM1 profile shows fluid migration anomalies along pre-existing or incipient fractures and the epicenters of shallow earthquakes (<3km) match with this area. Taking into account this situation in 2011, we could hypothesize that later and recent sedimentary units and/or gravitational instabilities could be related with the multiple volcanic crises from post-collapse infill eruption to nowadays.

5.3. Age of El Golfo landslide

The age of the multievent El Golfo landslide is uncertain. Onshore $^{40}\text{Ar}/^{39}\text{Ar}$ radiometric dating using for the age range the youngest lavas outcropping on the scarp

and the oldest lavas concealing the scar, indicate an age in between 87 ± 8 ka and 39 ± 13 ka ([Longpré et al., 2011](#)). The occurrence of a marine abrasion platform beneath the post-collapse El Golfo infill lavas ([Carracedo et al., 1999, 2001](#)), as shown from boreholes, in El Golfo embayment can be used to constrain further the age of the landslide. This abrasion platform is located at present sea-level and extends ~ 1 km inland from the coastline. The occurrence of this platform at the same elevation than the present sea-level has been corroborated in this study by beach deposits identified in the Los Llanillos water well-gallery. Its development before emplacement of the post-collapse lavas indicates a period of volcanic and landslide inactivity between the El Golfo collapse and the onset of the post-collapse volcanism. Furthermore, the Los Llanillos gallery displays three levels of beach deposits overlying the abrasion platform. These beach deposits outcrop 25, 130 and 10 m along the gallery, and appear interbedded with thick lava flows corresponding to post-collapse volcanism, possibly suggesting phases of inactivity between the eruptive events.

Taking into account the erosion rates in marine abrasion platforms for oceanic volcanic islands, ranging from 2 to 7 mm a⁻¹ ([Quartau et al., 2010](#); [Le Friant et al., 2004](#)), the total 165 m length of the abrasion platform in Los Llanillos gallery (Figs. 7 and 13b) would have taken between 23.5 and 82.5 ka to develop. Due to the fact that the abrasion platforms were developed after the flank collapse and over the landslide scar, the age of the landslide have to be older than the above range.

On the other hand, these abrasion platforms must have formed in a period with a sea level similar to present (0 masl; Figs. 7 and 13b), which leads to suggest a Tyrrhenian II age (the last interglacial event, ca. 130 ka BP) assuming that no subsidence or uplift processes occurred in the volcanic edifice in recent times ([Carracedo et al., 1999](#)). However, the occurrence of uplifted beach levels at 5 m above mean sea level (mamsl) in Roques de Salmor (Bravo, 1982) and 12 mamsl in El Golfo (Bellido-Mulas et al., 2012) suggest general recent uplift of the El Hierro and therefore the beach deposits identified in the wells and galleries at present sea level would have been deposited during a lower sea level period. A Tyrrhenian III age (ca. 80 ka) fits better the $^{40}\text{Ar}/^{39}\text{Ar}$ ages and the inferred uplift rates of the island.

In any case, our proposed range of ages are for the El Golfo landslide derived from geomorphologic analysis of beach deposits ratifies the range of ages proposed by

Longpré et al., (2011). Nevertheless, they are still contradictory with ages correlated from the Madeira Abyssal Plain suggesting a much younger age of avalanche emplacement at 10-17 ka as deduced from turbidite “b”(Masson, 1996). In this sense, we argue that the turbidite “b” deposits could be related with another landslide event different than the El Golfo landslide.

5.3. Data limitation, constrains and uncertainty assessment of the mass transport deposits volume calculation

The material described as piedmont deposits in Tigaday and Chirgo wells, could correspond to debris avalanche deposits from the landslide, connecting the slumped block at the base of the El Golfo scarp (Fig. 7b) with the seismic facies interpreted as MTDs (~200 ms TWT; Fig. 12c). Besides, Guillou et al. (1996) also suggest the presence of debris avalanche deposits at the bottom of a borehole in Frontera, at a depth of approximately 200 m or more. Finally, majority of the lithological description in water wells and galleries was done during 70th and 80th decades. In this period, the geological model of landsliding in Canary Island was not developed, and possibly any of piedmont descriptions could correspond with debris avalanche deposits. Taking into account this assumption, we hypothesize that debris avalanche deposits could be present in an uncertain area (green interrogation in Fig. 8b) below the post-collapse infill El Golfo lavas with a possible thickness ranging from 20 to 100 m (Fig. 13b).

Respect on seismic profiles in volcanic islands, the interpretation of MCS profiles in volcanic flanks implies a degree of potential ambiguity. The presence in the flanks of volcanic substrate and debris avalanche deposits with huge blocks, return high hyperbolae, low penetration and discontinuous reflections in the seismic signal. Besides, the MCS profiles registered during the volcanic crisis in 2011 show several amplitude anomalies and blanking seismic facies due to the intensive fluid migration in the island at that time. Nevertheless, the set of continuous reflections that we have established as the base of the debris avalanche deposits presents a lateral continuity in three MCS profiles, as well as continuous reflections between MTD A and MTD B in at least, two MCS profiles, B1 and SM1 (Figs. 11 and 12b). Hence, we argue that this group of continuous reflections is showing a coherent evidence of a real compositional and structural change.

In order to assess the reliability of seismic interpretation and confirm the concomitance of the interpreted debris avalanche deposits with the El Golfo scar, a 3D model of both debris avalanche deposits and a paleo-topographic reconstruction of the El Golfo flank have been developed. A comparison of the volume of failure rocks over the El Hierro Island with the volume of the EGDA is the most intuitive way to evaluate the transfer of rocks and sediments involved in a mass wasting event. Both, the 3D geological model of the El Golfo landslide and the paleo-topographic reconstruction has been carried out taking into account the geomorphological features and the geological elements shown in this paper.

Based on seismic data, the two main reflectors related to MTD-A and MTD-B have been modelled (Figs. 11 and 12b). The boundary of modeled sediments has been taken from the mapping (Fig. 8) and as fixed point of thickness of the three interpreted seismic profiles (Figs 11 and 12). With this data, the basal surfaces of MTDs have been interpolated and two 3D geological models generated using the “discrete smooth interpolation” algorithm of Paradigm SKUA-GOCAD 15.5®. An error of ± 7 m has been established in the calibration of the seafloors inferred from MBES and MCS profiles, and a possible ± 50 m/s in the calculation of the velocity of the debris avalanche deposits (1750-1800 m/s TWT; IGN, 1991; [Canals et al., 2004](#)). Taking into account these values, the calculated error for the debris avalanche volume will be of about 7%. In this way, the debris avalanche deposits have a total volume of $318 \pm 22 \text{ km}^3$ (Fig. 14): $84 \pm 6 \text{ km}^3$ and $234 \pm 16 \text{ km}^3$ for each of the MTDs, A and B respectively.

As this volume largely exceeds the previously published in $150\text{--}180 \text{ km}^3$ by [Masson et al. \(2002\)](#), the volume of rock and sediment involve in the El Golfo flank collapse has been calculated to assess the reliability of our calculation. It has been carried out in three steps: (i) reconstruction of the paleo-topography of the El Golfo collapsed flank, (ii) reconstruction of the landslide scar and (iii) calculation of expanded volume of fallen rock mass.

The paleo-topography reconstruction (Fig. 14) of the El Golfo flank has been based on the geomorphological mapping (Fig. 8). It has been extrapolated from the morphology of non-landslided flanks. From the mapped evidences, a paleo-flank that already recorded failures prior to occurrence of the EGDA (maybe [Tiñor landslide event; Klimes et al., 2016](#)) has been considered, rather than of a perfect conic shape. This

assumption is supported by the pre-existing or incipient fractures observed below EGDA (Fig. 11b) and the relict scars observed around the Roque de Salmor area (Figs. 4 and 8), which were previous to the El Golfo flank collapse. Paleo-scars surfaces and the base of the insular slopes have been extrapolated along hardlines (isobathymetry lines) digitized under expert criteria. Finally, the height of El Hierro Island has been estimated at about 2000 m, considering the dip of lavas in the headwall scar of the El Golfo (Carracedo et al., 1997, 2001; Navarro and Soler, 1994). With all the above fix point, the hypothetic paleo-topography was interpolated with the “discrete smooth interpolation” algorithm of Paradigm® SKUA-GOCAD 15.5™.

The failure surface has been modeled in two steps. The first step was to model the surface in the accumulation area below the offshore debris avalanche deposits. It has been developed from the MCS interpretation and removing the above sediments. The second step took place in the erosional area. The relief belonging to post-collapse infill El Golfo lavas was erased from the geological boundaries mapped, and the underlay failure surface was extrapolated. The final failure surface was the union of the result of the above two steps (Fig. 13). The volume of the island involved in the failure (volcanic rock, lavas) has been estimated in 247 km³. This volume is ~10% larger than that calculated by Masson et al. (2002) of 225 km³, and less than the EGDA deposits estimated in 318 km³ in this paper.

The error involved in the paleo-topographical reconstruction is difficult to calculate because of we have not ground terrain control point and the calculation has been based in expert criteria. For the error assessment, we have compared our result with a hypothetic landslide in a conic flank. Considering a conic flank without paleo scars and the small volcanic cones (1.5 km long and 300 m high) covering the flanks, it would yield a volume of 280 km³. This will mean an uncertainty of $\pm 12\%$ in our calculation.

To estimate the expanded volume that the 247 km³ of island rocks represent in the debris avalanche, we have followed two methodologies. The first one involved applying a swelling coefficient for igneous rocks of 64% and then a compacting factor of 30% (Bustillo et al., 2014). With these values the expanded volume of the failed portion of the island is 331 km³. The second method involves calculating the expanded volume of a failed volcanic block that undergoes density changes as it is transformed into a debris avalanche. The initial mass of volcanic island undergoing landslide processes is the

calculated volume, 247 km^3 , multiplied by the estimated mean density for the volcanic rocks on El Hierro island: i.e. 3000 kg/m^3 (Mironov, 1997). It implies a total mass of $7.41 \cdot 10^{12} \text{ kg}$ that transformed to a debris avalanche deposit, with approximately a mean density of $2100\text{-}2200 \text{ kg/m}^3$ (Iverson, 1997, 2013; Mironov, 1997), results in a volume of 337 km^3 . From these two methodologies, we infer an error of $\pm 7 \text{ km}^3$ in the estimation of expanded volume of island rocks.

Finally, taking into account a maximum uncertainty in our calculations of $\pm 12\%$ (estimated for the paleo-topographic reconstruction), both the volume of debris avalanche deposits and the expanded volume of the island failure are inside on the confidence interval. In this way, we argue that the calculated volume for the debris avalanche deposits is coherent and founded.

6. Conclusions

The combination of onshore (topography, geology and water wells and galleries observations) with offshore data (multichannel and high-resolution seismic profiles, bathymetry and backscatter data) covering the whole area affected by the El Golfo landslide has allowed us to map in further detail the landslide features and the corresponding debris avalanches with the following conclusions:

- Onshore and offshore data indicate that the El Golfo is a multiple landslide event.
- Onshore, the sub-aerial part of the El Golfo landslide is constituted by two superposed amphitheater headwalls (up to 1000 m in height) and suggests the presence of two collapse events: a collapse that generated the main part of the headwall, and another subsequent smaller collapse.
- The submarine proximal area of the El Golfo landslide is characterized by erosive structures (escarpments, scars and channels/gullies) in spatial continuity with the onshore scars and also suggesting two collapse events.
- The distal deep-water area is characterized by multiple allocthonous blocks up to 1.5 km in diameter and 300 m high, randomly distributed on the seafloor. Seismic and geomorphological analysis of the distal area reveals at least two major MTDs A and B, that are interbedded with continuous reflectors.

- Based on analysis of seismic data a volume of 318 km³ has been calculated for the debris avalanche deposit: 84 km³ and 234 km³, for the MTD A and MTD B, respectively. This calculation matches the estimated volume of island failure.
- Later events have been observed in the proximal area (lateral debris lobe-type) and deep-sea distal area (upper lobe).

ACKNOWLEDGEMENTS

This work has been supported by the Spanish research project SUBVENT (CGL2012-39524-C02-02) and Project for the Extension of the Continental Shelf West to Canary Islands (CTM2010-09496-E). This research has been also developed within the framework of the IGCP-640 S⁴SLIDE project (Significance of Modern and Ancient Submarine Slope LandSLIDES). We thank Lorena Moscardelli and Neil Mitchell for thorough reviews and assessments which helped to improve the article.

REFERENCES

- Acosta, J., Uchupi, E., Muñoz, A., Herranz, P., Palomo, C., Ballesteros, M., Working, Z. E. E., et al., 2003a. Geologic evolution of the Canarian Islands of Lanzarote , Fuerteventura , Gran Canaria and La Gomera and comparison of landslides at these islands with those at Tenerife , La Palma and El Hierro. Marine Geophysical Researches 24, 1–40.
- Acosta, J., Uchupi, E., Smith, D., Muñoz, A., Herranz, P., Palomo, C., Llanes, P., et al., 2003b. Comparison of volcanic rifts on La Palma and El Hierro, Canary Islands and the Island of Hawaii. Marine Geophysical Researches 24, 59–90.
- Battaglia, M., Di Bari, M., Acocella, V., Neri, M., 2011. Dike emplacement and flank instability at Mount Etna: Constraints from a poro-elastic-model of flank collapse. Journal of Volcanology and Geothermal Research, 199, 153–164.
- Becerril, L., Galindo, I., Martí, J., Gudmundsson A., 2015. Three-armed rifts or masked radial pattern of eruptive fissures? The intriguing case of El Hierro volcano (Canary Islands). Tectonophysics 647-648, 33-47.
- Bellido-Mulas, F., Gómez Sainz de Aja, J.A., Barrera, J.L., 2012. Mapa Geológico Digital continuo E. 1: 50.000, Zona Canarias - El Hierro (Zona-2916). In: GEODE. Mapa Geológico Digital continuo de España [on line]. Sistema de Información Geológica Continua: SIGECO. IGME. Editor: J. Navas [Date of query: 20/10/2012]. Available in: <http://cuarzo.igme.es/sigeco/default.htm>.
- Biaín, A., León, R., Urgeles, R., Somoza, L., Medialdea, T., Ferrer, M., González, F.J., 2016. Onshore and offshore geomorphological features of the El Golfo debris avalanche (El Hierro, Canary Islands). In: G. Lamarche et al. (Eds.), Submarine Mass Movement and their Consequences, Advances in Natural and Technological Hazards Research 41, 83-92.
- Blumenthal, M.M., 1961. Rasgos principales de la geología de las Islas Canarias con datos sobre Madeira. Boletín del Instituto Geológico y Minero de España. Madrid.
- Borgia, A., Delaney, P.T., Denlinger, R.P., 2000. Spreading volcanoes. *Annual Review of Earth Planetary Sciences* 28, 539–70.
- Bravo, T., 1982. Formaciones geológicas de la isla de El Hierro. *Instituto de Estudios Canarios* 50, 85-96.
- Brunet, M., Le Friant, A., Boudon, G., Lafuerza, S., Talling, P., Hornbach, M., Ishizuka, O., Lebas, E., Guyard, H., IODP Expedition 340 science party, 2015.

- Composition, geometry, and emplacement dynamics of a large volcanic island landslide offshore Martinique: From volcano flank-collapse to seafloor sediment failure. *Geochemistry, Geophysics, Geosystems* 17. doi:10.1002/2015GC006034
- Bustillo M., Durán A., Fueyo L., 2014. *Manual de Aridos*. Fueyo Editores: 590 pp..
- Canals, M., Lastras, G., Urgeles, R., Casamor, J. L., Mienert, J., Cattaneo, a., De Batist, M. et al. 2004. Slope failure dynamics and impacts from seafloor and shallow sub-seafloor geophysical data: case studies from the COSTA project. *Marine Geology* 213, 9–72.
- Carracedo, J.C., 1994. The Canary Islands: an example of structural control on the growth of large oceanic-island volcanoes. *Journal of Volcanology and Geothermal Research* 60, 225–241.
- Carracedo, J.C., Day, S.J., Guillou, H., Perez-Torrado, F.J., 1997. International workshop on volcanism and volcanic hazards in immature intraplate oceanic islands. Excursion Guidebook. El Hierro 22-24 September 1997.
- Carracedo, J.C., Day, S.J., Guillou, H., Perez-Torrado, F.J., 1999. Giant Quaternary landslides in the evolution of La Palma and El Hierro, Canary Islands. *Journal of Volcanology and Geothermal Research* 94, 169–190.
- Carracedo, J.C., Badiola, E.R., Guillou, H., Nuez, J., Pérez-Torrado, F.J., 2001. Geology and volcanology of La Palma and El Hierro, western Canaries. *Estudios Geológicos* 57, 175–273.
- Carracedo, J.C., Guillou, H., Rodríguez Badiola, E., Pérez-Torrado, F.J., Rodríguez González, A., Paris, R., Troll, V., 2009. La dorsal NE de Tenerife: hacia un modelo del origen y evolución de los rifts de islas oceánicas. *Estudios Geológicos* 65, 5–47.
- Carracedo, J.C., 2011. *Geología de Canarias: origen, evolución, edad y volcanismo*. Ed Rueda. 398 pp.
- CNIG, 2012. Centro Nacional de Información Geográfica, centro de descargas. Instituto Geográfico Nacional. Modelo digital del terreno MDT05-LIDAR. Isla de El Hierro, Canarias. Hojas 1105/1108. Query: 20-mars-2012. (<http://centrodedescargas.cnig.es/CentroDescargas/buscadorCatalogo.do?codFamilia=MDT05>).
- Crutchley, G. J., Karstens, J., Berndt, C., Talling, P. J., Watt, S. F. L., Vardy, M. E., Hühnerbach, V., et al., 2013. Insights into the emplacement dynamics of volcanic landslides from high-resolution 3D seismic data acquired offshore Montserrat, Lesser Antilles. *Marine Geology* 335, 1–15.

- [Dávila Harris, P., Branney, M.J., Storey, M., 2011. Large eruption-triggered ocean-island landslide at Tenerife: Onshore record and long-term effects on hazardous pyroclastic dispersal. *Geology* 39, 951-954.](#)
- [Day, S.J., Carracedo, J.C., Guillou, H., 1997. Age and geometry of an aborted rift flank collapse: the San Andres fault system, El Hierro, Canary Islands. *Geological Magazine* 134, 523–537.](#)
- [Day, S.J., Heleno da Silva, S.I.N., Fonseca, J.F.B.D., 1999. A past giant lateral collapse and present-day flank instability of Fogo, Cape Verde Islands. *Journal of Volcanology and Geothermal Research* 94, 191–218.](#)
- [Delcamp, A., Troll, V.R., Van Wyk de Vries, B., Carracedo, J.C., Petronis, M.S., Pérez-Torrado, F.J., Deegan, F.M., 2012. Dykes and structures of the NE rift of Tenerife, Canary Islands: a record of stabilisation and destabilisation of ocean island rift zones. *Bulletin of Volcanology* 74, 963–980.](#)
- [Famin, V., & Michon, L. \(2010\). Volcano destabilization by magma injections in a detachment. *Geology* 38, 219–222.](#)
- Fernández Navarro, L., 1908. Observaciones geológicas de la isla de Hierro (Canarias). *Memorias de la Real Sociedad Española de Historia Natural* 5, 49-92.
- [Ferrer, M., Seisdedos, J., González de Vallejo, L.I., 2010. The role of hyaloclastite rocks in the stability of the volcanic island flank of Tenerife. In: Olalla et al. \(Eds\), *Volcanic rock Mechanics*, Taylor and Francis Group, London, 167-170. ISBN 978-0-415-58478-4.](#)
- [Ferrer, M., González de Vallejo, L., Seisdedos, J., Coello, J.J., García, J.C., Hernández, L.E., Martín, C., et al., 2013. Güímar and La Orotava mega-landslides \(Tenerife\) and tsunamis deposits in Canary Islands. In: *Landslide Science and Practice. Vol. 5: Complex Environment*. Margottini, Canuti and Sassa \(Eds\) Springer-Verlag, 27-34. ISBN 978-3-642-31426-1.](#)
- [Ferrer M., González de Vallejo, L., González, S and Jiménez, E., 2014. Stability and failure mechanisms of large landslides in the volcanic island flanks of the Canary Islands. In: *Engineering Geology for Society and Territory. Vol. 2: Landslide processes*. Lollino et al. \(Eds\). Springer, 915-920. ISBN 978-3-319-09056-6.](#)
- Fuster, J., Hernán, F., Cendrero, A., Coello, J.J., Cantagrel, J.M., Ancochea, E., Ibarrola, E., 1993. Geocronología de la Isla de El Hierro (Islas Canarias). *Boletín de la Real Sociedad Española de Historia Natural* 88, 85–97.

- García, M.O., 1996. Turbidites from slope failure on Hawaiian volcanoes. In: McGuire, W.J., Jones, A.P., Newberg, J. (Eds) Volcano instability on the Earth and other planets. Geological Society of London, Special Publication 110, 281–294.
- Gee M.J., Martin J.R., Watts, A. B., Masson, D. G., Mitchell, N. C. 2001b. Landslides and the evolution of El Hierro in the Canary Islands. *Marine Geology* 177, 271–293.
- Geldmacher, J., Hoernle, K., Bogaard, P., Duggen, S., Werner, R., 2005. New $^{40}\text{Ar}/^{39}\text{Ar}$ age and geochemical data from seamounts in the Canary and Madeira volcanic provinces: Support for the mantle plume hypothesis. *Earth and Planetary Science Letters* 237, 85–101.
- Guillou, H., Carracedo, J.C., Torrado, F.P., Badiola, E.R., 1996. K-Ar ages and magnetic stratigraphy of a hotspot-induced, fast grown oceanic island: El Hierro, Canary Islands. *Journal of Volcanology and Geothermal Research* 73, 141–155.
- Harbitz, C.B., Løvholt, F., Pedersen, G., Masson, D.G., 2006. Mechanisms of tsunami generation by submarine landslides: a short review. *Norwegian Journal of Geology* 86, 255–264.
- Hausen, H., 1972. Outlines of the geology of El Hierro. *Commentationes Physico-Mathematicae* 43, 65–167.
- Hernández Pacheco, A., 1982. Sobre una posible erupción en 1793 en la isla de El Hierro (Canarias). *Estudios Geológicos*, 38, 15-25.
- Holcomb, R., Searle, R., 1991. Large landslides from oceanic volcanoes. *Marine Geotechnology*, 10, 19–32.
- Hunt, J.E., Wynn, R.B., Talling, P.J., Masson, D.G., 2013. Multistage collapse of eight western Canary Island landslides in the last 1.5 ma: Sedimentological and geochemical evidence from subunits in submarine flow deposits. *Geochemistry Geophysics Geosystems* 14, 2159–2181.
- Hürlimann, M., Ledesma, A., Marti, J., 1999. Conditions favouring catastrophic landslides on Tenerife (Canary Islands). *Terra Nova* 11, 106-111.
- IGME, 1990. Proyecto para la actualización de la infraestructura hidrogeológica de la isla de El Hierro. Sistema Acuífero 87. Madrid. Ministerio de Industria y Energía.
- IGME, 2012. Base de datos de aguas de Canarias. Query: 21-Mars-2012. (<http://info.igme.es/bdaguas/>).
- IGN, 1991. Estudio geofísico de la Cuenca oceánica al oeste del Archipiélago Canario. Publicacion Tecnica nº 25: 46pp.

- Iverson, R.M., 1997. The physics of debris flows. *Reviews of Geophysics* 35, 245-296.
- Iverson, R.M. 2013. Mechanics of debris flows and rock avalanches. In: Harindra, J., Sherma, F. (Eds.), *Handbook of environmental fluid dynamics*. Vol 1. CRC Press/Taylor & Francis Group, LLC, Boca Ratón, Florida, pp. 573-587
- Klimes, J., Yepes, J., Becerril, L., Kusák, M., Galindo, I., Blahut, J., 2016. Development and recent activity of the San Andrés megalandslide on El Hierro, Canary Islands, Spain. *Geomorphology* 261, 119-131.
- Krastel, S., Schmincke, H.U., Jacobs, C.L., Rihm, R., Le Bas, T.P., Alibés, B., 2001. Submarine landslides around the Canary Islands. *Journal of Geophysical Research*, 106, 3977-3997.
- Labazuy, P., 1996. Recurrent landslides events on the submarine flank of Piton de la Fournaise volcano (Reunion Island). In: McGuire WJ, Jones AP, Neuberg J (eds) *Volcano instability on the Earth and other planets*. Journal of the Geological Society of London. Special publication 110, 295–306.
- Lebas, E., Masson, D.G., Holtomb, R.T., Grevemeyer, I., 2007. Slope failures of the flanks of the southern Cape Verde islands. In: Lykousis, V., Sakellariou D. and Locat J. (Eds.), *Submarine Mass Movements and Their Consequences* 27. Springer, Dordrecht, The Netherlands, pp. 337–345.
- Le Friant, A., Harford, C.L., Deplus, C., Boudon, G., Sparks, R.S.J., Herd, R.A., Komorowski, J.C., 2004. Geomorphological evolution of Montserrat (West Indies): importance of flank collapse and erosional processes. *Journal of the Geological Society* 161,147–160.
- Longpré, M.A., Chadwick, J.P., Wijbrans, J., Iping, R., 2011. Age of the El Golfo debris avalanche, El Hierro (Canary Islands): New constraints from laser and furnace $^{40}\text{Ar}/^{39}\text{Ar}$ dating. *Journal of Volcanology and Geothermal Research* 203 (1-2), 76–80.
- López, C. Blanco, M.J., Abella, R., Brenes, R., et al., 2012. Monitoring the volcanic unrest of El Hierro (Canary Islands) before the onset of the 2011–2012 submarine eruption. *Geophysical Research Letters* 39, L13303.
- Madeira, J., Ferrer, M., Gonzalez de Vallejo, L.I., Andrade, C., Freitas, M.C., Lomoschitz, A., 2011. Agaete revisited: new data on the Gran Canaria tsunamiites. *Geophysical Research Abstracts*, vol. 13, EGU2011-2292-2. EGU General Assembly 2011.

- [Masson, D.G., 1996. Catastrophic collapse of the volcanic island of Hierro 15 ka ago and the history of landslides in the Canary Islands, *Geology*, 24, 231-234.](#)
- [Masson, D.G., Canals, M., Lonsoà, B.A., Urgeles, R., Huhnerbach, V., 1998. The Canary Debris Flow : source area morphology and failure mechanisms. *Sedimentology* 45, 411-432.](#)
- [Masson, D., Watts, A.B., Gee, M.J., Urgeles, R., Mitchell, N., Le Bas, T., Canals, M., 2002. Slope failures on the flanks of the western Canary Islands. *Earth-Science Reviews* 57, 1–35.](#)
- [McGuire, W.J., Pullen, A.D., Saunders, S.J., 1990. Recent dyke induced large-scale block movement at Mount Etna and potential slope failure. *Nature* 343, 357-359.](#)
- [McMurtry, G.M., Watts, P., Fryer, G.J., Smith, J.R. and Imamura, F., 2004. Giant landslides, mega-tsunamis and paleo-sea level in the Hawaiian Islands. *Marine Geology* 203, 219-233.](#)
- [Mitchell, N.V., Masson, D.G., Watts A.B., Gee, M.J.R., Urgeles, R., 2002. The morphology of the submarine fanks of volcanic ocean islands. A comparative study of the Canary and Hawaiian hotspot islands. *Journal of Volcanology and Geothermal Research* 115, 83-107.](#)
- Mironov, V.S., 1977. Curso de Prospección Gravimétrica. Ed. Nedra, sección de Leningrado. Ed. Reverté, S.A., Barcelona, 525 pp.
- [Moore, J.G., Moore, G.W., 1984. Deposit from a giant wave on the island of Lanai, Hawaii. *Science* 226, 1312-1315.](#)
- [Moore, J.G., Moore, G.W., 1988. Large scale bedforms in boulder gravel produced by giant waves in Hawaii, U.S. Geological Society of America Special Papers, 1988, 229, 101-110,](#)
- [Moore, J.G., Clague, D.A., Holcomb, R.T., Lipman, P.W., Normark, W.R., and Torresan, M.E., 1989. Prodigious submarine landslides on the Hawaiian Ridge: *Journal of Geophysical Research* 94, 17465–17484.](#)
- [Moore, J.G., Normark, W.R., Holcomb, R.T., 1994. Giant Hawaiian landslides. *Annual Review of Earth and Planetary Sciences* 22, 119– 144.](#)
- MOP, 1975. Estudio científico de los recursos de agua en las Islas Canarias (SPA/69/515). Volumen III, 2: parte. Minist. Obras Públ, Dir. Gral. Obr. Hidr. UNESCO. Las Palmas de Gran Canaria, Madrid. 3 vol.+ mapas. 262 pp.

- Moscardelli, L., Wood, L., Mann, P., 2006. Mass-transport complexes and associated processes in the Offshore Area of Trinidad and Venezuela. *AAPG Bulletin*, 90, 1059-1088.
- Münn, S., Walter, T.R., Klügel, A., 2006. Gravitational spreading controls rift zones and flank instability on El Hierro, Canary Islands. *Geological Magazine* 143, 257-268.
- Murray, J.B., Voight, B., 1996. Slope stability and eruption prediction on the eastern flank of Mount Etna. In: McGuire, W.J., Jones, A.P., Neuberg, J. (Eds.), *Volcano Instability on the Earth and Other Planets*. Geological Society of London, Special Publication, pp. 111-114.
- Murty, T.S., 2003. "Tsunami Wave Height Dependence on Landslide Volume", *Pure and Applied Geophysics* 160, 2147-2153.
- Navarro, J.M., Soler, C., 1994. El agua en El Hierro. Resumen del Avance del Plan Hidrológico. Public. Cabildo de El Hierro and Consej. O.P. Gob. Canarias, pp. 1–59.
- Oehler, J.F., Van Wyk de Vries, B., Labazuy, P., 2005. Landslides and spreading of oceanic hot-spot and arc shield volcanoes on Low Strength Layers (LSLs): an analogue modeling approach. *Journal of Volcanology and Geothermal Research* 144, 169–189.
- Oehler, J.F., Lénat, J.F., Labazuy, P., 2008. Growth and collapse of the Reunion Island volcanoes. *Bulletin of Volcanology* 70, 717–742.
- Paris, R., Giachetti, T., Chevalier, J., Guillou, H., Frank, N., 2011. Tsunami deposits in Santiago Island (Cape Verde archipelago) as possible evidence of a massive flank failure of Fogos volcano. *Sedimentary Geology* 239, 129–145.
- Pellicer, M., 1977. Estudio volcanológico de la Isla de El Hierro (Islas Canarias). *Estudios Geológicos* 33, 181-197.
- Pérez-Torrado, F.J., Paris, R., Cabrera, M.C., Schneider, J.L., Wassmer, P., Carracedo, J.-C., Rodríguez-Santana, A., et al., 2006. Tsunami deposits related to flank collapse in oceanic volcanoes: The Agaete Valley evidence, Gran Canaria, Canary Islands. *Marine Geology* 227, 135– 149.
- Quartau, R., A.S. Trenhaile, N.C. Mitchell, Tempera, F., 2010. Development of volcanic insular shelves: Insights from observations and modelling of Faial Island in the Azores Archipelago. *Marine Geology* 275, 66-83.

- Roberts, J.A., Cramp, A., 1996. Sediment stability on the western flanks of the Canary Islands. *Marine Geology* 134, 13–30.
- Roeser, H.A., 1982. Magnetic anomalies in the magnetic quiet zone off Morocco. In: von Rad, U., Hinz, K., Sarnthein, M. and Seibold E. (Eds.), *Geology of the Northwest African Continental Margin*. Springer, Berlin, pp. 61–68.
- Schiffman, P., Watters, R.J., Thompson, N., Walton, A.W., 2006. Hyaloclastites and the slope stability of Hawaiian volcanoes: Insights from the Hawaiian Scientific Drilling Project's 3-km drill core. *Journal of Volcanology and Geothermal Research* 151, 217–228.
- Schmincke, H.U., 1994. *Geological Field Guide of GranCanaria*. Edit. Pluto, Kiel, Germany, 219 pp
- Seisdedos, J., Ferrer, M., González de Vallejo, L.I., 2012. Geological and geomechanical models of the pre-landslide volcanic edifice of Güímar and La Orotava mega-landslides (Tenerife). *Journal of Volcanology and Geothermal Research* 239-240, 92–110.
- Somoza, I., Vázquez, J.T., González-Aller, D., Dañobeitia, J.J., Medialdea, T., Moya, A.J., Fernández-Salas, L.M., León, R., Jurado, M.J., Paredes, M., Palomino, D., Vargas, I., Goñzalez, F.J., López-Rodríguez, M., Corbalán, A., Gonzalez, P., Hernández-Puentes, M.P., Martín-Jiménez, D. 2012. Estudio geofísico e hidroacústico del volcanismo submarino activo de la Isla de El Hierro (Archipiélago Canario). *Geo-Temas* 13, 601-604.
- Széréméta, N., Laj, C., Guillou, H., Kissel, C., Mazaud, A., Carracedo, J.C., 1999. Geomagnetic paleosecular variation in the Brunhes period , from the island of El Hierro (Canary Islands). *Earth and Planetary Science Letters* 165, 241–253.
- Thompson, N., Watters, R.J., Schiffman, P., 2008. Stability analysis of Hawaiian Island flanks using insight gained from strength testing of the HSDP core. *Journal of Volcanology and Geothermal Research* 171, 163–177.
- Tibaldi, A., 2001. Multiple sector collapses at Stromboli volcano, Italy: how they work. *Bulletin of Volcanology* 63, 112–125.
- Urgeles, R., Canals, M., Masson, D. 1997. The most recent megalandslides of the Canary Islands: El Golfo debris avalanche and Canary debris flow. *Journal of Geophysical Research* 102, 305–320.

- Urgeles, R., Canals, M., Baraza, J., Alonso, B., 1998. Seismostratigraphy of the western flanks of El Hierro and La Palma (Canary Islands): a record of Canary Islands volcanism. *Marine Geology* 146, 225–241.
- Urgeles, R., Canals, M., Roberts, J., SNV “Las Palmas” Shipboard Party, 1999. Fluid flow from pore pressure measurements off La Palma, Canary Islands. *Journal of Geophysical Research* 94, 305–321.
- Van Wyk de Vries, B., Francis, P.W., 1997. Catastrophic collapse at stratovolcanoes induced by gradual volcano spreading. *Nature* 387, 387–390.
- Voight, B., Elsworth, D., 1997. Failure of volcano slopes. *Géotechnique* 47, 1-31.
- Ward, S.N., Day, S., 2001. Cumbre Vieja Volcano -- Potential collapse and tsunami at La Palma , Canary Islands. *Geophysical Research Letters* 28, 3397–3400.
- Watt, S.F.L., Talling, P.J., Vardy, M.E., Heller, V., Hühnerbach, V., Urlaub, M., Sarkar, S., et al. 2012a. Combinations of volcanic-flank and seafloor-sediment failure offshore Montserrat, and their implications for tsunami generation. *Earth and Planetary Science Letters* 319-320, 228–240.
- Watt, S.F.L., Talling, P.J., Vardy, M.E., Masson, D.G., Henstock, T.J., Hühnerbach, V., Minshull, T., et al. 2012b, Widespread and progressive seafloor-sediment failure following volcanic debris avalanche emplacement: Landslide dynamics and timing offshore Montserrat, Lesser Antilles. *Marine Geology* 323-325, 69–94.
- Wynn, R.B., Masson, D.G., 2003. Canary island landslides and tsunami generation: can we use turbidite deposits to interpret landslide processes. In: Locat J. and Mienert J. (Eds.), *Submarine mass movements and their consequences*, Dordrecht, Netherlands: Kluwer Academic Publishers, 325–332..
- Wynn, R.B., Masson, D.G., Stow, D., Weaver, P.P., 2000. Turbidity current sediment waves on the submarine slopes of the western Canary Islands. *Marine Geology* 163, 185–198.
- Wynn, R.B., Weaver, P.P.E., Masson, D.G., Stow., 2002. Turbidite depositional architecture across three interconnected deep-water basins on the north-west African margin. *Sedimentology* 49, 669–695.

Figure captions

Figure 1. (a) Geological schema of the Canary Islands area (Modified from Wynn et al., 2000, 2002; Masson et al., 2002). Ages of volcanic edifices are taken from Geldmacher et al. (2005). (b) El Hierro Island geological map: Main constructive units (modified from Bellido-Mulas et al., 2012), offshore debris avalanche boundaries (modified from Masson et al., 2002); Tiñor landslide event (modified from Carracedo et al., 2001).

Figure 2. Dataset presented in this paper and location of figures. Onshore: shaded relief map of topography at 5 m resolution (data from Instituto Geográfico Nacional; CNIG, 2012); red stars are water well locations (IGME, 2012). Offshore: colored multibeam data from ZEEE-2000 and GAIRE-2011 cruises. Black thick lines are high-resolution bathymetric profiles. White lines are high-resolution seismic data.

Figure 3. Map of the submarine morphology of the El Hierro Island (Modified from Biain et al., 2016).

Figure 4. (a) 3D view of the subaerial and submarine El Hierro Island. (b) Geomorphological interpretation of the El Golfo landslide.

Figure 5. Maps of local maximum slope gradient and second derivative derived from the elevation and bathymetry data shown in Figure 2.. Bathymetry vs second derivative profiles of the non-collapsed flanks in which black arrows show the maximum change in gradient.

Figure 6. (a) Location of wells, drainage galleries and boreholes over the onshore geological map (modified from Bellido-Mulas et al., 2012). (b) Rose diagrams of orientations of dykes observed in galleries. In the Tincos diagram, light grey represents dike azimuths in the first 550 m, whereas the remaining is shown in black.

Figure 7. Onshore view and non-scaled schemas of the lithological description of the main groundwater extractions used in the paper: water well-galleries of Los Llanillos, Tigaday and Las Casitas; and the water gallery of Tincos. **a** Surficial aspect of the northeast side of the headwall scarp. Note the positive relief of dykes and the erosion of the paleo-coastal cliff over the landslide scar and post-collapse piedmont. **b** Slumped block (faulted and rotated) belonging to El Golfo landslide located at the base of the headwall scarp. Heterometric detrital deposits corresponding to Upper Pleistocene (Bellido-Mulas et al., 2012) post-collapse piedmonts. **c** Aerial view and surficial aspect of the post-collapse infill El Golfo lavas and the main geomorphological features of the El Golfo landslide.

Figure 8. (a) Geomorphological interpretation of the El Golfo landslide showing the main features: Debris avalanche deposits (MTD A and MTD B), Upper lobe, lateral debris lobe-type and Canary Debris Flow (CDF). (b) Detail of the chute and headwall sectors (c) Detailed backscatter mosaic of the uppermost area; (d) High-resolution bathymetry; and (e) Backscatter image .

Figure 9. Chirp profiles crossing the chute sector showing the main geomorphological features. See Figs. 2 and 8 for location.

Figure 10. High-resolution bathymetric profiles from proximal (a) to distal (e) areas of the chute sector. The main geomorphological features are also identified. See Fig. 2 for location.

Figure 11. (a) Multichannel seismic profiles (Line B, Canarias-88 Cruise). The acoustic basement is located between 6400 and 7500 ms TWT. Seismic units I, II and III defined by Urgeles et al. (1998). (b, c and d) Details of the seismic profile showing the architecture of the MTDs.

Figure 12. Multichannel seismic profiles recorded in 2011 (Lines SM1 and SM2, GAIRE cruise) during the volcanic crisis of La Restinga (El Hierro island). (a) Situation of seismic profiles and earthquakes. (b) Longitudinal seismic profile SM1. The base of MTD B does not appear in the proximal area of the El Golfo landslide. (c) Transversal seismic profile SM2.

Figure 13. (a) Geological cross-section of the El Golfo landslide integrating subaerial and submarine observations. (b) Detail of the subaerial headwall area. Two possible detachment planes are related with the two sub-units of submarine MTDs. The drainage gallery of Los Llanillos cuts three levels of paleo-beaches deposited after the El Golfo debris avalanche.

Figure 14. 3D Geological model showing the thickness of the El Golfo landslide MTDs. Isopachs are every 20 m. In red, paleo-topographical reconstruction of the El Golfo collapsed flank, paleo-isobaths in white. Headwall scar has been modelled below post-collapse lavas (uppermost black lines). The total volume of MTDs is $\sim 318 \text{ km}^3$ and the volcanic rocks failure $\sim 247 \text{ km}^3$.

Table 1. Summary of the lithological and structural information of the groundwater extractions (water wells, galleries and well-galleries) used in this paper.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12

Figure 13

Figure 14

Table 1

Location	Groundwater extraction		Depth/Length	dykes (headwell)		beach levels
lat/long	Name	Type	(m)	num	direction	num/masl
27.79°N/17.98°W	Las Casitas	well - gallery	60/170	(4): 2/2	N25°E/N65°E	1/~1.5
27.78°N/17.99°W	Los Padrones	well - gallery	51.5/1160	(2)	N65°E	<i>not found</i>
27.75°N/17.99°W	Tincos	gallery	0/1770	(76)	N40-115°E/(E-W)*	<i>not found</i>
27.75°N/17.98°W	Charco de Tincos	gallery	0/280	(26)	N75°E	<i>not found</i>
27.75°N/18.01°W	Tigaday	well - gallery	280/160	(9)	ENE-WSW	<i>not found</i>
27.75°N/18.03°W	Los Llanillos	well - gallery	178/680	(2)	N45°E	3/0
27.75°N/18.04°W	Chirgo	well - gallery	265/30	<i>not found</i>		1/0
27.74°N/18.10°W	Lomo de trébol	gallery	0/300	(43)	N110°E	<i>not found</i>
See Fig. 6a	Coastal wells	well	~140/0	<i>not found</i>		<i>not found</i>