Geomorphic evolution of the Malta Escarpment and implications for the Messinian evaporative drawdown in the eastern Mediterranean Sea.

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ABSTRACT

Carbonate escarpments are submarine limestone and dolomite cliffs that have been documented in numerous sites around the world. Their geomorphic evolution is poorly understood due to difficulties in assessing escarpment outcrops and the limited resolution achieved by geophysical techniques across their steep topographies. The geomorphic evolution of carbonate escarpments in the Mediterranean Sea has been influenced by the Messinian salinity crisis (MSC). During the MSC (5.97-5.33 Ma), the Mediterranean Sea became a saline basin due to a temporary restriction of the Atlantic-Mediterranean seaway, resulting in the deposition of more than one million cubic kilometres of salt. The extent and relative chronology of the evaporative drawdown phases associated to the MSC remain poorly constrained. In this paper we combine geophysical and sedimentological data from the central Mediterranean Sea to reconstruct the geomorphic evolution of the Malta Escarpment and infer the extent and timing of evaporative drawdown in the eastern Mediterranean Sea during the MSC. We propose that, during a MSC base-level fall, fluvial erosion formed a dense network of canyons across the Malta Escarpment whilst coastal erosion developed extensive palaeoshorelines and shore platforms. The drivers of geomorphic evolution of the Malta Escarpment after the MSC include: (i) canyon erosion by submarine gravity flows, with the most recent activity taking place <2600 cal. years BP; (ii) deposition by bottom currents across the entire depth range of the Malta Escarpment; (iii) tectonic deformation in the southern Malta Escarpment in association with a wrench zone; (iv) widespread, small-scale sedimentary slope failures preconditioned by oversteepening and loss of support due to canyon erosion, and triggered by earthquakes.

We carry out an isostatic restoration of the palaeoshorelines and shore platforms on the northern Malta Escarpment to infer an evaporative drawdown of 1800 – 2000 m in the eastern Mediterranean Sea during the MSC. We interpret the occurrence of pre-evaporite sedimentary lobes in the western Ionian Basin as suggesting that either evaporative drawdown and canyon formation predominantly occurred before salt deposition, or that only the latest salt deposition at the basin margin occurred after the formation of the sedimentary lobes.

Keywords: Malta Escarpment; geomorphic evolution; submarine canyon; palaeoshoreline; sea level drawdown; Messinian salinity crisis

Declarations of interest: none.

1. INTRODUCTION

1.1 The Malta Escarpment

Carbonate escarpments are submarine limestone and dolomite cliffs with relief in excess of 1 km; they can be up to 650 km long and have average slope gradients exceeding 40° (Schlager and Camber, 1986; Twichell et al., 1990). Their morphologies are complex and variable, comprising scarps, overhangs, vertical walls, terraces and wide canyons (e.g. Paull et al., 1987; Twichell et al., 1991; 1996; Eberli et al., 2005). Carbonate escarpments have been documented in numerous sites around the world, e.g. offshore Florida (Paull and Dillon, 1980; Paull et al., 1990a), Bahamas (Freeman-Lynde et al., 1981), Yucatan Peninsula (Paull et al., 2014), north-west Australia (Veevers, 1974), Malta (Scandone et al., 1981), and the Balearic Islands (Acosta et al., 2001).

Carbonate escarpments were the focus of several studies in the 1970s and 1980s. Understanding of their formation has been limited by difficulties in assessing their outcrops and the limited resolution achieved by 2D seismic reflection and single beam echo sounding techniques across steep topographies. An origin based on carbonate platform growth and erosion of the escarpments' faces has been put forward to explain the topography of the world's largest carbonate escarpments (Dillon et al., 1987; Paull et al., 1990a; 1991; 2014). This contrasts with the predominantly structural origin attributed to carbonate escarpments in the Mediterranean Sea (e.g. Emile Baudot Escarpment (Mauffret

et al., 2004), Malta Escarpment (Cita et al., 1980; Scandone et al., 1981; Casero et al., 1984; Jongsma et al., 1985; Reuther et al., 1993)).

The Malta Escarpment is a steep, 290 km long submarine cliff with a relief in excess of 3 km. It extends from the eastern margin of Sicily southwards to the Medina Seamounts (Fig. 1). The most detailed surveys of the Malta Escarpment were primarily aimed at constraining the age, stratigraphy and nature of this feature using single-beam echosounders, seismic reflection profilers, gravity coring, dredge sampling and submersible dive observations (Cita et al., 1980; Scandone et al., 1981; Biju-Duval et al., 1983; Casero et al., 1984). The Malta Escarpment is arguably the best sampled of all carbonate escarpments worldwide. Nevertheless, its fine-scale geomorphology is still poorly constrained, and gaps remain in our understanding of the geomorphic evolution of the Malta Escarpment and its key driving forces. In-depth understanding of these aspects is necessary to better assess the hazard that the Malta Escarpment poses to seafloor infrastructure and densely populated coastal settlements in its vicinity, and to refine the tectonic and oceanographic history of the region.

FIGURE 1

1.2 Messinian salinity crisis

The Messinian salinity crisis (MSC) is an extraordinary event in the geological history of the Mediterranean region that had local to global consequences (Ryan, 2008a; Roveri et al.,

2014a; Flecker et al., 2015). Between 5.97 to 5.33 Ma, the Mediterranean Sea became an enormous saline basin due to a temporary restriction of the Atlantic-Mediterranean seaway, resulting in the deposition of more than one million cubic kilometres of salt that can locally exceed 3 km in thickness (Ryan, 2008b). Among the many open questions related to the MSC, the extent and relative chronology of the drawdown phases, following the deposition of the primary gypsum and prior to any significant sea-level change (i.e., the second step *sensu* Clauzon et al. (1996)), is poorly understood and three different scenarios have been proposed:

- (i) Only the shallow Messinian continental shelves were exposed as a consequence of a moderate drawdown (Roveri et al., 2001; 2008; 2014a). Deep basin unconformities in seismic reflection profiles from Mediterranean margins are interpreted as originating from different processes (e.g. cascading brines (Roveri et al., 2014b) or deep water evaporite dissolution (Gvirtzman et al., 2017)). The halite body and the overlying upper evaporites are therefore interpreted as the result of deposition in a deep, saturated, brine-filled basin.
- (ii) A kilometre-scale drawdown occurred following a period of deep-water brine concentration (Ryan, 2008b; Lofi et al., 2011b). According to this scenario, the erosional surfaces are the product of subaerial processes. In the shallow margins, these surfaces pre-date the salt deposition, whereas in the deep margins and basin floors, the erosional surfaces are coeval to the salt deposition during the “evaporative shrinking” of the basin. Evaporite emplacement started in deep water environments and ended in shallow water environments.

(iii) A kilometre-scale drawdown occurred before salt deposition, with the latter occurring in a shallow water environment (Bache et al., 2009; 2015). The main evidence for this scenario is the interpretation of the erosional surface extending below the Messinian evaporites in the Gulf of Lion.

For scenarios (ii) and (iii), attempts at quantifying the maximum drawdown from seismic data around the rim of the Mediterranean that contained the evaporite pinch-out generated are as follows (Fig. 1a):

- Ebro Margin: ~1300 m (Urgeles et al., 2011); ~2000 m (Stampfli and Höker, 1989);
- Valencia Basin: ~1500 m (Maillard and Mauffret, 1993);
- Gulf of Lion: ~2000 m (Ryan, 1976; Lofi, 2002; Steckler et al., 2003);
- Levant Margin: 1000 m (Druckman et al., 1995); ~1500 m (Ben-Gai et al., 2005; Tibor and Ben-Avraham, 2005);
- Nile Delta: 1500-2250 m (Gargani and Rigollet, 2007); 3000-4000 m (Barber, 1981).

Numerical modelling by Blanc (2000) produced drawdown estimates of 1140 m and 1000 m in the western and eastern Mediterranean basins, respectively, while those of Ryan (2008b) suggest a maximum drawdown of ~1800 m and ~2400 m in the western and eastern Mediterranean basins, respectively.

What is conspicuously absent in the above studies are observations from the central Mediterranean Sea. Reconstructions of past sea-level changes (Imbrie et al., 1989) and stratigraphic analyses (Max et al., 1993; Osler and Algan, 1999) suggest that, after the Mediterranean Sea was refilled at the end of the MSC, the central Mediterranean Sea has experienced low sedimentation rates ($\sim 6 \text{ cm ka}^{-1}$) because it has largely remained isolated from inputs of fluvial and littoral sediments. The Malta Escarpment and western Ionian Basin should thus host topographic and sedimentary signatures of the key stages of the MSC in the eastern Mediterranean Basin.

1.3 Objectives

In this study we analyse geophysical and sedimentological data from the Malta Escarpment and adjacent western Ionian Basin to: (i) reconstruct the geomorphic evolution of the Malta Escarpment since the MSC, and (ii) infer the extent and timing of evaporative drawdown during the MSC in the eastern Mediterranean Sea.

2. REGIONAL SETTING

The Malta Escarpment is the north-eastern boundary of the Pelagian Platform, which consists of continental crust forming part of the African continental plate (Finetti, 1982). The north-eastern Pelagian Platform comprises the 100-150 m deep, gently sloping Malta Plateau (Micallef et al., 2011; 2016) (Fig. 1b). Here, parallel-bedded units of Plio-Pleistocene terrestrial, pelagic and hemipelagic sediments, which are up to 300 m thick,

overlie >4500 m thick older sedimentary sequences (Jongsma et al., 1985; Max et al., 1993; Torelli et al., 1995; Osler and Algan, 1999). These older sequences outcrop across the Malta Escarpment and were dredged during a number of expeditions (Cita et al., 1980; Scandone et al., 1981; Casero et al., 1984) (Fig. 2). They comprise Triassic to Cretaceous shallow platform carbonates and Cretaceous to Miocene shelf edge carbonate build ups (Scandone et al., 1981; Pedley et al., 1993). The upper Malta Escarpment is covered by Tortonian to Recent pelagic deposits (Biju-Duval et al., 1982). The Mesozoic succession of the Malta Escarpment is punctuated by extensive depositional hiatuses, as well as tuffs and pillow lavas deposited during a number of volcanic episodes (Scandone et al., 1981; Pedley et al., 1993; Bosellini, 2002).

FIGURE 2

The Malta Escarpment was thought to have originated by rifting in the Upper Permian-Triassic, followed by spreading from the Jurassic until the Upper Cretaceous-early Tertiary (Ben-Avraham and Grasso, 1991; Grasso, 1993; Catalano et al., 2000a). More recently, Catalano et al. (2000b) suggested that continental rifting took place from the pre-Triassic till the Early Cretaceous, and was followed by spreading in the Early Cretaceous. Adam et al. (2000) propose that, since the onset of plate convergence between Africa and Europe during the Late Cretaceous, the Malta Escarpment was transformed from a passive margin into a mega-hinge fault system with an additional sinistral strike-slip component. For this reason, the Malta Escarpment exhibits both Mesozoic normal block faulting and more recent sinistral strike-slip movement (Scandone et al., 1981; Reuther, 1990; Grasso, 1993;

Pedley et al., 1993; Reuther et al., 1993; Catalano et al., 2000b). The current tectonics of the Malta Escarpment are likely controlled by differences in the convergence rates of the Ionian oceanic crust and the adjacent continental crust of the Pelagian Platform, and passive sinking and lateral tear of the dense Ionian oceanic lithosphere along a slab transfer edge propagator fault (Adam et al., 2000; Argnani and Bonazzi, 2005; Govers and Wortel, 2005; Gutscher et al., 2016).

The Pelagian Platform is traversed by the Sicily Channel Rift Zone (Fig. 1b), which comprises three principal NW-SE trending grabens (Pantelleria, Malta and Linosa) that are bound by sub-vertical normal faults and that have been active from Late Miocene to Recent times (Jongsma et al., 1985; Dart et al., 1993; Civile et al., 2008). The grabens are up to 1700 m deep and filled by up to 2000 m thick Lower Pliocene-Pleistocene turbidites (Reuther and Eisbacher, 1985; Civile et al., 2010). Three explanations have been put forward for the origin of the Sicily Channel Rift Zone. The first is that the grabens are pull-apart basins developed along a major dextral wrench zone, which shows divergent, strike-slip and convergent sections from west to east (Jongsma et al., 1985; 1987; Reuther and Eisbacher, 1985; Finetti and Del Ben, 2005). The eastern extremity of this wrench zone coincides with the Medina Seamounts, which have been interpreted as upthrust blocks formed during the Plio-Quaternary (Groupe Escarmé, 1982). The second explanation is that the rifting is associated with mantle convections developed during the roll-back of the African lithospheric slab beneath the Tyrrhenian Basin (Argnani, 1990; Reuther et al., 1993). The third explanation is intraplate rifting related to the north-eastern displacement of Sicily away from the African continent (Illies, 1981; Winnock, 1981).

220

221 The Ionian Basin, to the east of the Malta Escarpment, is a deep and narrow oceanic basin
222 considered to be the remnant of the Mesozoic Tethys Ocean (de Voogd et al., 1992;
223 Catalano et al., 2000a). It is characterised by ~10 km thick crystalline oceanic crust overlain
224 by 5-7 km of sedimentary cover (Speranza et al., 2012). The age of these sediments ranges
225 from the Jurassic to Recent (Cernobori et al., 1996). Since the Tortonian, the Ionian Basin
226 has been the site of abundant accumulation of Messinian evaporites and Plio-Pleistocene
227 hemipelagic sediments (Hsü et al., 1978; Casero et al., 1984; Polonia et al., 2011; Gallais
228 et al., 2013; Gutscher et al., 2016). The Ionian lithosphere has been subducting underneath
229 Calabria to the north-west, giving rise to an accretionary wedge that involves the
230 sedimentary cover and that has shallow detachments in the Messinian evaporitic sequences
231 (Finetti, 1982; Maesano et al., 2017).

232

233 The central Mediterranean Sea is the pathway for the Modified Atlantic Water (MAW) -
234 Atlantic water entering through the Gibraltar Strait (Millot and Taupier-Letage, 2005) –
235 which flows from the western to the eastern Mediterranean Sea (Fig. 1b). The upper 100-
236 120 m of the water column over the Pelagian Platform consists of the MAW flowing in a
237 south-easterly direction at velocities of up to 25 cm s^{-1} (Lermusiaux and Robinson, 2001).
238 The more saline Levantine Intermediate Water (LIW) originates in the eastern
239 Mediterranean Basin and splits into two flows - westwards on the Pelagian Platform and
240 below the MAW, and northwards along the Malta Escarpment. In the latter case, the LIW
241 flow reaches velocities of up to 20 cm s^{-1} (Lermusiaux and Robinson, 2001).

242

3. MATERIALS AND METHODS

3.1 Data

This study is based on data acquired from the Malta Escarpment and the western Ionian Basin between 1969 and 2015 (Fig. 2), including:

- a. Multibeam echosounder (MBES) data: An area of ~8000 km², extending from offshore Siracusa to the Medina Seamounts and covering a depth range of 120 – 4200 m, was surveyed during the CUMECS, CIRCEE-HR, CUMECS-2 and CUMECS-3 oceanographic surveys using Kongsberg-Simrad EM-710 and EM-302 systems and Reson SeaBat 7150 and 8111 systems. After sound velocity corrections and basic editing, bathymetry and backscatter grids with 15 m to 50 m grid resolution were derived. The backscatter data were processed with PRISM (Processing of Remotely-sensed Imagery for Seafloor Mapping) software (Le Bas and Hühnerbach, 1998).
- b. Multi-channel seismic reflection profiles from the Malta Escarpment region were compiled from the following oceanographic surveys:
 - MS, 1969-1973: Finetti and Morelli (1973).
 - CROP, 1988-1995: Finetti and Del Ben (2005).
 - CA-99, 1999: Source included an air gun array (total volume of 36.9 l). Receiver included a 6000 m long streamer with average hydrophone spacing of 12.5 m.

- MEM-07, 2007: Source included an air gun array (total volume of 36.9 l). Receiver included a 7200 m-long streamer with average hydrophone spacing of 12.5 m.
 - CIRCEE-HR, 2013: Gutscher et al. (2016).
 - CUMECS-3, 2015: Source included one GI-Gun and one Mini GI-Gun (total volume of 4.4 l). Receiver included a 300 m-long GEOMETRICS GeoEel digital streamer with average channel spacing of 3.125 m.
- c. Gravity cores: Six cores were recovered using a 3 m-long gravity coring system during the CUMECS-2 oceanographic survey (Table 1).

TABLE 1

We integrated the above data with existing multibeam data from Gutscher et al. (2017) (60 m grid resolution) and EMODnet bathymetry (<http://www.emodnet-bathymetry.eu>) (220 m grid resolution), core logs from Deep Sea Drilling Program Site 374 (Hsü et al., 1978), well data published in VIDEPI (<http://unmig.sviluppoeconomico.gov.it/videpi/>), and gravity core and dredge samples from Scandone et al. (1981) and Casero et al. (1984) (Figs. 1, 2).

3.2 Methods

Standard GIS tools were used to generate morphometric attributes from the multibeam data and map landforms across the Malta Escarpment (Micallef et al., 2007). Local mean slope

gradients were calculated based on circular footprints with a diameter of 500 m. The boundaries and thalwegs of submarine canyons were extracted using flow direction and flow accumulation routines (Micallef et al., 2014a; 2014b). The resulting maps were validated by a thorough qualitative inspection.

The gravity cores were visually logged, photographed, x-radiographed, and analysed in terms of sediment colour, magnetic susceptibility, P-wave velocity, and gamma density using a Geotek® Multi-Sensor Core Logger (MSCL) at the Irish Sediment Core Research Facility (ISCORF) of Maynooth University. Samples from selected intervals were analysed quantitatively for grain size distribution with a Malvern Mastersizer 3000™ at the Tullow Oil Core Facility of UCD. AMS ¹⁴C dating of four samples from three cores (CU14_01, CU14_06 and CU14_07) was carried out by the Poznan Radiocarbon Laboratory. For three of the samples, foraminiferal assemblages of mixed planktonic species were used, as there were not enough individuals for monospecific picking. The fourth sample was an intact *Cavolinia* shell.

In order to derive the original depth of features identified across the Malta Escarpment, we have performed an isostatic restoration similar to a 1-step back-stripping analysis used for sedimentary basins (Allen and Allen, 1990). For the thin plate flexural calculations, we used a standard thin plate approach and the TISC code (Garcia-Castellanos, 2002). A lithospheric elastic thickness of 50 km was adopted as reference, with a Young modulus of 7×10^{10} N m. Values of 2000 and 3200 kg m⁻³ were used for sediment and asthenospheric densities, respectively (Lachenbruch and Morgan, 1990). We first unloaded the

sedimentary units above the evaporites identified in the western Ionian Basin in the seismic reflection profiles (section 5.1) and calculated the corresponding flexural-isostatic rebound. The decompaction of the underlying sedimentary units was then estimated. This estimation is limited by the poor knowledge of the depth to the oceanic crystalline basement. However, we found that the contribution of this depth to isostatic rebound is almost an order of magnitude smaller than that of the other parameters. Decompaction was carried out by adopting a near-surface porosity of 0.5 (applicable for a material with porosity between that for shales and sandstone) and a decay constant of $5.1 \times 10^{-4} \text{ m}^{-1}$ (average value for shales) (Allen and Allen 1990). Subsequently, we unloaded the weight of the seawater layer and calculated the corresponding flexural-isostatic response. Because our study region includes one of the oldest oceanic lithospheres preserved on Earth (Granot, 2016), we use a large lithospheric equivalent elastic thickness of 50 km for the flexural calculations (Watts, 2001). To determine the water unloading effect, the sea level before the MSC is considered to have been close to the present one (Miller et al., 2011). The cumulative error inherent to these approaches is estimated to be around 20%, or +/- 120 m along the escarpment.

4. RESULTS

We divide the study area into four sections – northern Malta Escarpment, central Malta Escarpment, southern Malta Escarpment, and western Ionian Basin (Fig. 3a) – and present our main observations accordingly (Fig. 4).

FIGURE 3

FIGURE 4

4.1 Northern Malta Escarpment

Our dataset covers a 120 km-long section of the northern Malta Escarpment (Fig. 3a). Here, the escarpment is oriented north-northwest to south-southeast and faces east-northeast. The total relief on the escarpment face increases to the south, reaching a maximum of 3200 m. The depth of the top and base of the escarpment increases southwards from 100 m to 600 m, and from 2600 m to 3700 m, respectively. The maximum overall slope gradient of the escarpment is 11° , although the highest local mean slope gradient is 74° . The steepest slopes are characteristically found along the lower escarpment.

The seismic signature of the escarpment is predominantly comprised of acoustically transparent seismic facies topped by a high amplitude and irregular reflector (facies C) (Figs. 5, 6). On the steeper sections of the escarpment ($>10^\circ$), facies C is overlain by a thin (<0.05 s Two Way Travel Time (TWTT)) package of high amplitude, sub-parallel and continuous reflectors (facies A) (Fig. 6a). Gentler slope gradients are draped by sequences of sub-parallel, convex upward, continuous reflectors (facies B) (Figs. 5, 6b). The thickness of this drape is up to 0.3 s TWTT. The backscatter response is low where the escarpment is covered by facies A and B (Fig. 3b).

FIGURE 5

FIGURE 6

4.1.1 Submarine canyons

(a) Fine-scale morphology

We define submarine canyons as steep-sided valleys on the continental shelf and slope. Submarine canyons are the most pervasive landforms across the northern Malta Escarpment (Fig. 4). The escarpment hosts more than two hundred canyons (60 canyons and 140 tributaries). The canyons have V-shaped cross-sections, sharp interfluves and pointed heads (Fig. 7a); their morphometric attributes are listed in Table 2. The heads of the canyons connect, or are located close, to the top of the northern Malta Escarpment (Fig. 7a). The canyon thalwegs are characterised by high backscatter, whereas their walls and interfluves have a low backscatter response (Fig. 3b). Some canyons host individual or series of crescent-shaped steps along their thalwegs (Fig. 7a). In general, the shapes of the canyon thalweg profiles are concave (Fig. 7b). This is in contrast with the convex longitudinal profile of the escarpment face that is not associated with canyons (Fig. 7b).

FIGURE 7

TABLE 2

The northern Malta Escarpment includes two larger canyons – Noto Canyon in the north, which is 27 km long, 15 km wide, and 1.5 km deep (Fig. 7c), and Cumecs Canyon in the

381 south, which is 37 km long, 39 km wide, and 3 km deep (Fig. 7e). The morphology of Noto
382 Canyon is unusual for the Malta Escarpment as it has a U-shaped cross-section and a
383 theatre-shaped head that is up to 700 m high and 70° in overall slope gradient (Fig. 7d).
384 Upslope of Noto Canyon, the seafloor has a mean slope gradient of 7° and hosts a dendritic
385 network of tributary canyons. The alignment of the main tributary of Noto Canyon changes
386 by 90° at a distance of 8 km upslope from its mouth. The northernmost tributaries of
387 Cumecs Canyon are oriented northwest to southeast (Fig. 7e), which contrasts with the
388 southwest to northeast orientation of the other tributaries along the same canyon and the
389 majority of the canyons nearby. The seafloor on the bed or downslope from the mouths of
390 these large canyons exhibits series of ellipsoidal depressions up to 1.5 km in width (Figs.
391 7c, e).

392
393 A distinctive feature on the northern Malta Escarpment is a promontory that extends 8 km
394 northeast from the general trend of the escarpment's base (Fig. 3a). The promontory is the
395 only section of the escarpment not cut by canyons.

396 397 **(b) Seismic facies**

398
399 The canyon thalwegs are characterised by acoustically chaotic to transparent seismic facies
400 (facies D) that are up to 0.2 s TWTT thick and that overlie strata that are acoustically
401 transparent and topped by a high amplitude and irregular reflector (facies C) (Fig. 7d). The
402 mouth of Cumecs Canyon is buried beneath 0.4 s TWTT of high amplitude, sub-parallel
403 and continuous reflectors (facies A) (Fig. 7f). Along the northern flank of Cumecs Canyon,

a package of sub-parallel, convex upward continuous reflectors (facies B) is slightly offset at three sites (Fig. 6b).

(c) Sediment cores

One 0.4 m-long core was retrieved from Cumecs Canyon (CU14_04; Fig. 8a). It predominantly consists of mud, whereas the core catcher contained gravel, pebbles and a ~10 cm-long limestone cobble (Fig. 8b).

FIGURE 8

Five cores were retrieved from Noto Canyon – two from upslope of the tributary canyons (CU14_06 and _07), one from below the head (CU14_08), and two from the mouth (CU14_01 and _02). The two cores from upslope of the tributary canyons (CU14_06 and _07), taken from water depths of 133 m and 230 m, contain a distinct coarse layer composed of an aggregate of shells, sand and mud, which is overlain by a fining-upwards sequence (Fig. 8a). The radiocarbon date for the shell in this layer is 19,811 cal. years BP. Since that the shell is intact, it must have either been deposited in situ or it has not been transported far from its living position. We therefore consider the age of the coarse layer as <19,811 cal. years BP.

The other cores display intervals of the canyon's most recent activity. CU14_01 features a homogeneous sandy layer within the top 0.5 m of the core. This layer, deposited after ~2,600 cal. years BP (age derived from dating of foraminiferal assemblages of mixed

planktonic species), is characterised by a sharp erosional base with parallel lamination. These features are evident in the X-ray and coincide with an increase in P-wave velocity, gamma ray density and magnetic susceptibility. This layer is not present in CU14_02. However, the physical property logs of CU14_02 show a relative increase of magnetic susceptibility at nearly the same depth below the seafloor (marked by a dashed line in fig. 8a). Likewise, we see a P-wave velocity peak in cores CU14_06 and _07 at a similar depth below the seafloor. About one metre below this event, core CU14_01 contains a fining-upward sandy layer that contains some shell fragments and that caps a layer of mixed mud clasts. The latter is separated by a sharp, straight boundary from an interval of thinly laminated mud and silt below, which is occasionally punctuated by thin, coarser, sandy layers (laminated facies). The sharp boundary at the top of the laminated facies is also very clearly identified in core CU14_02, but it lacks the clast-rich interval and fining up sand layer above it. The laminated facies is fully penetrated in core CU14_02, where it is 2 m thick and shows thicker and coarser sandy intervals punctuating it downcore.

CU14_08, which was retrieved from the base of the Noto Canyon head, contains an interval that is nearly 0.5 m thick, comprising contorted and sheared mud layers (Fig. 8a). The core does not penetrate through the base of the layer, so this can only be inferred to be the minimum thickness. It is difficult to determine whether this layer correlates with the sandy layer and clast-rich layer of core CU14_01.

We estimate an indicative sedimentation rate of 4.85 cm ka^{-1} based on the radiocarbon ages obtained from core CU14_06. Based on this sedimentation rate, the contorted interval in

CU14_08 is 7,800 yr old. The homogeneous sandy layer and the fining upward sandy layer in CU14_01 are 2,600 yr and 20,000 yr old, respectively.

Attempts at coring the smaller canyon thalwegs proved unsuccessful, and the corer returned clean and undamaged (Table 1).

4.1.2 Re-entrants

A number of re-entrants cluster along the base of the southern half of the northern Malta Escarpment (Fig. 4). Re-entrants have U-shaped cross-sections and theatre-shaped heads; their morphometric attributes are listed in Table 2 (Fig. 7g). The base of the re-entrants is characterised by low backscatter, whereas their walls have a high backscatter response (Fig. 3b).

4.1.3 Scars

More than two hundred scars have been identified across the northern Malta Escarpment (Fig. 4). The headwalls of the scars are generally linear to arcuate, steep (10-25°), and characterised by high backscatter. At their distal limit, most scars connect to either another scar or a canyon. The seabed within these scars is typically relatively smooth, planar and characterised by low backscatter. The smaller scars are located along the heads and walls of submarine canyons (Fig. 9f). They have headwalls <50 m high and an average width of

1.3 km. The larger scars are located along the top of the escarpment (Fig. 7e), which are up to 23 km wide, with headwalls up to 200 m in height.

FIGURE 9

4.1.4 Breaks of slope and terraces

Thirteen breaks of slope were identified across the northern Malta Escarpment (Fig. 4). We define a break of slope as a change in profile curvature of $> 0.4^\circ \text{ m}^{-1}$ (concave) or $< -0.4^\circ \text{ m}^{-1}$ (convex), occurring along a length of $> 2 \text{ km}$ and not associated with a scar. The longest of these breaks of slope are concave in profile and include a 70 km terrace at a mean depth of $2478 \text{ m} \pm 119 \text{ m}$ (1 S.D.) extending between Noto Canyon and Cumecs Canyon, and a 25 km-long terrace occurring within Cumecs Canyon at a mean depth of $2545 \text{ m} \pm 70 \text{ m}$ (1 S.D.) (Fig. 9d). The surfaces of these terraces are smooth, up to 1 km in width, and slope at $< 5^\circ$ (Fig. 9a). In seismic profiles, the terraces consist of a high amplitude and irregular reflector marking the top of an acoustically transparent seismic facies (facies C) (Fig. 9c). The terraces are generally draped by a mounded facies with sub-parallel, convex upward, continuous reflectors (facies B). A similar facies with comparable thickness occurs along the top of the escarpment.

The remaining eleven breaks of slope are convex in profile, $< 7 \text{ km}$ long, are not associated with a terrace, and occur in 350 – 2000 m water depths (Figs. 4, 9b, 9d).

4.2 Central Malta Escarpment

The central Malta Escarpment is 90 km long, up to 16 km wide, faces northeast and is arcuate-shaped in plan view (Fig. 3a). The top and base of the escarpment occur at an almost invariable mean depth of 800 m and 3400 m, respectively. The maximum relief of the central Malta Escarpment is 2600 m and the maximum overall slope gradient is 8°, with local mean slope gradients of 50°. The fine-scale morphology is difficult to characterise in view of the lower resolution of the available bathymetric grid covering this section. However, the central Malta Escarpment hosts at least 35 canyons that share similar characteristics to those of the northern Malta Escarpment (Fig. 7h). The morphometric attributes of these canyons are listed in Table 2.

4.3 Southern Malta Escarpment

4.3.1 Fine-scale morphology

The southern Malta Escarpment is dominated by Heron Canyon, which is the largest submarine canyon along the entire Malta Escarpment (Fig. 10a). Heron Canyon is 100 km long, up to 10 km wide and 1 km deep. Its depth ranges from 450 m at the head to 4000 m at the mouth, and its orientation changes from west-northwest to east-northeast, to southwest to northeast, at 40 km from its mouth. The longitudinal profile of Heron Canyon is linear and characterised by three steps in its lower half; these are up to 30° in slope gradient and have heights of 130 m, 110 m and 110 m (Fig. 10b). Heron Canyon hosts 20

tributary canyons along its southern flank, which share similar morphologic characteristics to those in the northern and central Malta Escarpment; morphometrics are shown in Table 2. The upper half of Heron Canyon is characterised by three linear escarpments oriented northwest to southeast, which are up to 50 km long and 250 m high (Fig. 10a).

FIGURE 10

Along most of its length, Heron Canyon has an asymmetric cross-section with a steeper northern flank. The downstream half of the northern flank is contiguous with a 20 km wide upland that has an average depth of 1500 m. Here, the flank is characterised by an escarpment that is up to 1200 m high and has an overall slope gradient of 25°. The eastern flank of the upland features a 13 km-long concave break of slope, which is associated with a 3 km wide terrace and which occurs at a mean depth of $2165 \text{ m} \pm 67 \text{ m}$ (1 S.D.)

More than 60 scars were mapped on Heron Canyon walls (Figs. 4, 10a). These scars have arcuate headwalls that are up to 100 m high, 25° steep, 7 km wide and characterised by high backscatter.

4.3.2 Seismic facies

We identify three facies in the multichannel seismic reflection profiles located upslope of Heron Canyon (Figs. 10c, d):

- (i) Facies with high amplitude, sub-parallel and continuous reflectors (facies A), which has a maximum thickness of 0.2 s TWTT. Reflectors are clearly offset vertically to sub-vertically by up to 0.1 s TWTT in places. These offset seismic reflectors are often associated to a step in the seafloor.
- (ii) Facies with low amplitude, sub-parallel and discontinuous reflectors (facies G); the latter are vertically to sub-vertically offset down to 2 s TWTT below facies A.
- (iii) Acoustically chaotic to transparent seismic facies, (facies D), which has a maximum thickness of 0.7 s TWTT and fills a buried depression.

4.4 Western Ionian Basin

4.4.1 Bedforms

The surface of the Ionian Basin seafloor at the foot of the Malta Escarpment is generally smooth with a south-southeast slope of 0.6° (Fig. 3a). Twenty ellipsoidal depressions, up to 2 km wide and 125 m deep, occur along the base of the northern Malta Escarpment, mostly in proximity to the promontory and near the base of the Noto Canyon walls (Figs. 4, 9e). The depressions generally coincide with the mouths of submarine canyons. One unusually large depression (10 km wide and 40 m deep) occurs at the tip of the promontory (Fig. 9e). Seismic reflection profile CIR-04 shows that the high amplitude, sub-parallel and continuous reflector package (facies A) within the large depression is 0.07 s TWTT thinner at the top in comparison to the adjacent seafloor (Fig. 11a). Along 100 km of the base of the northern Malta Escarpment, the western Ionian Basin seafloor is also characterised by

a series of ridges and troughs oriented perpendicular to the escarpment (Figs. 3, 4). The ridges are up to 8 km long, have an average height of 30 m and are spaced 2-3 km apart.

FIGURE 11

4.4.2 Seismic facies

Five seismic facies characterise the western Ionian Basin (Figs. 5, 11). The shallower part comprises a sequence of high amplitude, sub-parallel and continuous reflectors (facies A) that reaches a thickness of up to 0.5 s TWTT. The shallower part of facies A features undulating, sub-parallel seismic reflector sequences underneath ridges and troughs on the seafloor (Fig. 11b). Along the base of the northern Malta Escarpment, facies A initially onlaps on, and subsequently drapes over, elongated mounded packages of sub-parallel, convex upward, continuous reflectors (facies B) that reach a maximum thickness of 0.3 s TWTT (Fig. 11a). Between Siracusa and Cumecs Canyon, an acoustically chaotic to transparent seismic facies (facies D) with a maximum thickness of 0.6 s TWTT occurs underneath facies A and B (Fig. 11a). Facies D has a horizontal, planar top along parts of its WSW boundary (Fig. 11a). Facies E underlies facies A, B or D, and consists of two sub-facies that reach a total maximum thickness of 1.5 s TWTT - E1 (highly reflective, discontinuous facies) and E2 (reflector-less facies with a basin-fill geometry) (Figs. 11b-c). Lenses of intermediate amplitude and sub-parallel reflectors (facies F) underlie facies E and are up to 0.2 s TWTT thick, 8 km wide and located downslope of the large canyon

mouths (Figs. 11b-c). The top of these lenses, which is marked by a high amplitude reflector, is sealed by facies E2.

4.5 Isostatic modelling

By implementing the isostatic modelling methodology outlined in section 3.2, the original depth of the breaks of slope and terraces on the northern Malta Escarpment are estimated to be 500-700 m higher during the MSC than their present position (Fig. 12). Subsidence since 5.33 Ma results from a combination of the weight of water refill above -2500 m, the weight of post-MSC sedimentation (facies A, B and D), and the compaction of pre-Tertiary sediment. The eastward tilting predicted by the model is the result of the water load being concentrated along the eastern side of the Malta Escarpment. Smaller values of equivalent elastic thickness predict a steeper slope of the subsidence pattern, but conveniently the hinge of this tilting roughly coincides with the escarpment itself. For example, an equivalent elastic thickness of 40 km results in the 700 m subsidence isoline remaining very close to the Malta Escarpment, which is in relative proximity to the same isoline generated by an equivalent elastic thickness of 50 km. Hence, reasonable values for equivalent elastic thickness exert little effect on the subsidence predicted along the escarpment (<150 m).

FIGURE 12

5. DISCUSSION

5.1 Stratigraphic framework

We interpret the seismic facies across the study area based on correlation with published seismic reflection profiles, core and well data (Fig. 5):

(i) Facies A: In the western Ionian Basin, this facies correlates to units of Plio-Pleistocene terrestrial, pelagic and hemipelagic sediments (Hsü et al., 1978; Polonia et al., 2011; Gallais et al., 2013). Similar sequences also occur across the outer Malta Plateau (Max et al., 1993; Osler and Algan, 1999; Micallef et al., 2016).

(ii) Facies B: This facies is very similar to that reported along the outer Malta Plateau (Micallef et al., 2016) and is typical of contouritic deposits (Rebesco and Stow, 2001; Stow et al., 2002).

(iii) Facies C: This facies correlates with carbonate bedrock sequences (limestone and dolomite), the top of which corresponds to the Messinian Salinity Crisis (MSC) erosional surface (Scandone et al., 1981; Casero et al., 1984; Finetti, 1984; Micallef et al., 2011; 2016).

(iv) Facies D: In the western Ionian Basin, this facies has been interpreted as a deposit associated to the passage of the Zanclean megaflood via a gateway located in south-eastern Sicily at the end of the MSC (Micallef et al., 2018). Facies D upslope of Heron

Canyon is interpreted as infill by high energy syn-rift deposition in the Early Pliocene (according to Civile et al., 2010).

(v) Facies E: This facies coincides with evaporites deposited during the MSC (Hsü et al., 1978; Gallais et al., 2011; Polonia et al., 2011; Gutscher et al., 2016). We interpret facies E1 as gypsum and marls from the Upper Unit, and E2 as halite from the Mobile Unit that infills a pre-existing topography (following nomenclature for MSC units in Lofi et al., 2011a).

(vi) Facies F: This facies does not correlate with well data from the western Ionian Basin. Previously interpreted as Tortonian deposits (Polonia et al., 2011), we interpret facies F as sedimentary lobes because of the lenticular geometry and spatial association with submarine canyon mouths.

(vii) Facies G: This facies also correlates with carbonate bedrock sequences. The seismic signature of this facies is different from that of facies C because a higher penetration seismic acquisition system was used to acquire the MS profiles.

5.2 Geomorphic processes shaping the Malta Escarpment since the MSC

5.2.1 Canyon erosion

In view of their morphology and connection to buried channels on the outer Malta Plateau, we propose that the submarine canyons on the Malta Escarpment were formed by mechanical erosion by downslope surface flows (Gerber et al., 2009; Pratson et al., 2009; Micallef et al., 2014b; 2016). Two observations lead us to infer that such erosion primarily took place during the MSC. First, the mouth of Cumecs Canyon is buried by a thick cover (i.e. 0.4 s TWTT) of Plio-Pleistocene sediments, indicating that it was excavated before the deposition of these sediments (Fig. 7e). Secondly, there is an absence of canyon mouth fans or mass transport deposits in the Plio-Pleistocene sequence along the western Ionian Basin, which indicates that the canyons were eroded before the Pliocene. We suggest that these submarine canyons were incised by subaerial fluvial drainage systems when the Malta Escarpment was partially exposed during a MSC base-level fall; unexposed areas would have been eroded by coarse-grained gravity flows sourced by the fluvial systems. Such a scenario has been reported in other margins of the western (Lofi et al., 2005; Garcia et al., 2011) and eastern (Barber, 1981) Mediterranean Sea. It would also explain why the Malta Escarpment exhibits the highest drainage density of any carbonate escarpment, despite it being located the farthest from terrestrial sources of sediment (Micallef and Paull, in prep.).

Submarine gravity flows of terrestrial, pelagic and hemipelagic sediments appear to have played a role in eroding and transporting sediment across the Malta Escarpment canyons after the end of the MSC. This inference is supported by the occurrence of: (i) crescent-shaped steps along the canyon thalwegs (Fig. 7a), which we interpret as upslope-migrating sedimentary bedforms or knickpoints associated to the passage of gravity flows flows (e.g. Mitchell, 2006; Paull et al., 2011; Micallef et al., 2014a); (ii) individual depressions

downslope of canyons mouths (Fig. 9e), which we interpret as plunge pools formed by gravity flow scouring (e.g. Principaud et al., 2015); (iii) series of ellipsoidal depressions within or downslope of the larger canyons (Figs. 7c, e), which we interpret as scour marks; (iv) debrites and turbidites reported in sediment cores from these depressions (Casero et al., 1984) and from Cumecs Canyon and its tributaries (Scandone et al., 1981) (Fig. 2).

Our sediment cores also show evidence of gravity flows. We interpret the laminated interval in cores CU14_01 and CU14_02 (Fig. 6a) as the deposit(s) of dilute weak turbidity currents. We interpret the fining upward sandy layer capping the clast-supported, contorted interval in core CU14_01 as a turbidite overlying a debrite. Another debrite is also found in core CU14_08, as evidenced by the contorted, sheared mud layers in the lower half of the core. The thin sandy layers in CU14_01 and CU14_02 are interpreted as sandy turbidites. Gravity flow activity across the Malta Escarpment canyons appears to have occurred episodically since at least the Last Glacial Maximum, with the youngest sampled gravity flow deposit (CU14_01) having an age <2600 cal. years BP. This, combined with the nature of the deposits, leads us to conclude that the gravity flows would have been sourced by slope failures on the outer Malta Plateau (Micallef et al., 2016) and the Malta Escarpment canyon walls (section 5.2.3). During sea-level lowstands, when the coastline would have been located close to the upper northern Malta Escarpment (Fig. 3a), subaerial fluvial discharge and possibly hyperpycnal flows, associated to wave or tidal suspension, could have generated gravity flows.

We assign the high backscatter signature of the canyon thalwegs (Fig. 3b) to bedrock exposures and/or coarse sediment (Collier and Brown, 2005). These observations, combined with the thin Plio-Pleistocene sedimentary cover along canyon thalwegs (Fig. 6b), indicate that post-MSC pelagic and hemipelagic sedimentation within the canyons has been low, likely a result of transport and erosion by gravity flows.

The above inferences do not apply to Noto Canyon, however. The unusual morphology of Noto Canyon is similar to that reported for bedrock canyons that were rapidly eroded by megafloods (Lamb et al., 2008; 2014). A buried 4 km wide and ~400 m deep erosional channel is located upslope of Noto Canyon, whilst the megaflood deposit (facies D) is located downslope of Noto Canyon (Micallef et al., 2018). For these reasons, Noto Canyon has been attributed to erosion of the Malta Escarpment during the passage of the Zanclean megaflood at the end of the MSC (Micallef et al., 2018).

5.2.2 Palaeoshoreline and shore platform development

The concave breaks of slope and associated terraces, which correspond to the MSC erosional surface (Fig. 9c), are interpreted to have formed during the MSC and subsequently became draped by Plio-Pleistocene contouritic deposits. In view of their morphology and the low standard deviation associated with their depths, we propose that the concave breaks of slope and terraces are palaeoshorelines and shore platforms formed by wave erosion during a Messinian sea-level lowstand. The high salinity contrast between a brine-filled western Ionian Basin and freshwater discharge along the palaeocoastline

could have enhanced dissolution processes (Plummer, 1975; Hanshaw and Back, 1980). The shoreline recession rate is estimated to be 0.003 m/a (based on a 3000 m recession in 100 ka (Gargani and Rigollet, 2007)), which is comparable to, or lower than, rates measured for coastal limestone cliff erosion (e.g. 0.0013 m/a in a coral atoll in the Indian Ocean (Trudgill, 1976); 0.018 m/a in Sweden (Rudberg, 1967); 0.07 m/a in Wales (Williams and Davies, 1987); 0.5 m/a in the Black Sea coast (Zenkovich et al., 1965)).

An alternative explanation for the concave slope breaks and terraces may be the occurrence of lithologies with contrasting erosional resistances. However, neither the boreholes nor the dredge samples (Scandone et al., 1981; Casero et al., 1984) show significant changes in lithology at the depths at which the breaks occur. A lithological change is also unlikely to maintain a similar depth over a distance of almost 100 km, particularly if we consider that the eastern Pelagian Platform strata are not horizontal, as indicated by the 500 m change in bathymetry along the top of the northern Malta Escarpment. Thus, a lithological contrast does not explain the development of concave slope breaks and terraces at a quasi-uniform depth along the escarpment.

5.2.3 Slope failures

The numerous scars on the escarpment are evidence for slope instability. The style of deformation is either translational sliding, as indicated by the linear to arcuate steep headwalls and the smooth and planar scars (Locat and Lee, 2002), or debris flow, because most of the failed material has been evacuated from the scars. The slope failures occurred

in stratified, fine-grained contouritic or hemipelagic/pelagic sediments deposited across the Malta Escarpment canyon walls and heads during the Plio-Pleistocene. Micallef et al. (2016) have shown that slope failures across the eastern Malta Plateau occur along specific stratigraphic horizons in the Plio-Pleistocene sediment package; this is likely to also be applicable to mass movements across the Malta Escarpment, because the sedimentological and stratigraphic characteristics of the material that has failed are similar. The most likely preconditioning factor for slope failure is over-steepening and loss of support as a result of canyon erosion. Seismicity, on the other hand, is likely to be the main trigger, because ground shaking associated with distal earthquakes of EMS-98 intensities >VIII has the potential of initiating slope failures in the eastern Malta Plateau (Micallef et al., 2016). Slope failures have played a key role in widening and extending canyons upslope and facilitating tributary development.

5.2.4 Bottom current activity

The contouritic drift deposits in the western Ionian Basin (facies B in fig. 11a) are indicative of the establishment of strong and long-term bottom currents, which flowed parallel to the Malta Escarpment, after the Mediterranean Sea was refilled at end of the MSC. The horizontal truncation of the underlying Zanclean megaflood deposit (facies D) to the WSW of the contouritic drift deposit (Fig. 11a) is indicative of erosion by the bottom currents. The ridges and troughs located at the seafloor along the base of the northern Malta Escarpment and the underlying undulating sub-parallel seismic reflectors, have been interpreted as sediment waves (Gutscher et al., 2016; Munari et al., 2016) (Figs. 3a, 11b).

The occurrence of the above features, together with contouritic drift deposits on the escarpment (Figs. 4-6) and the outer Malta Plateau (Micallef et al., 2016) suggest that bottom currents flowing parallel to the escarpment have been active across the entire depth range of the Malta Escarpment. The unusually large depression located at the tip of the promontory (Fig. 9e) may be either a zone of non-deposition between the escarpment and the sediment wave field, or an erosional zone excavated by bottom current flow that was locally modified by the promontory.

5.2.5 Structural and tectonic processes

The southern Malta Escarpment features long escarpments and offset seismic reflectors that provide clear evidence of recent fault activity. We interpret the several vertical and sub-vertical offsets of reflectors upslope of Heron Canyon (Fig. 10c-d) as normal faults that displace sequences of Plio-Pleistocene pelagic/hemipelagic material and carbonate bedrock, and occasionally bound half-graben structures. Evidence of strike-slip deformation, on the other hand, is provided by twenty six earthquakes, which were recorded in the region since 1985 and which have a strike-slip focal mechanism (Figs. 2; 10a). We therefore infer that the interpreted faults are associated to both extensional and strike-slip kinematics. Faulting appears to have had a direct control on the morphology of Heron Canyon, in particular its dimensions, orientation, cross-sectional asymmetry, and linear longitudinal profile (as in Covault et al., 2011; Micallef et al., 2014a). We therefore infer that Heron Canyon constitutes the eastern extremity of the Sicily Channel Rift Zone, and that the latter likely corresponds to a wrench zone.

791

792 The palaeoshoreline and shore platform identified in the southern Malta Escarpment are
793 335 m shallower than those recorded in the northern Malta Escarpment. This difference is
794 similar to the total height of the steps measured along the bed of Heron Canyon (350 m),
795 which we interpret as knickpoints formed due to changes in base level (e.g. Mitchell, 2006;
796 Micallef et al., 2014a). We propose that the difference in depth is a response to tectonic
797 deformation. Namely, regional base level was lowered due to uplift of the southern Malta
798 Escarpment in association to wrench faulting (Jongsma et al., 1987). The vertical
799 displacement was at least 335 m higher than that for the northern and central Malta
800 Escarpment during the last 5.3 Ma. Other explanations may include differential subsidence
801 due to lower sedimentation rates in the southern Malta Escarpment, or base level
802 fluctuations in the eastern Mediterranean during the MSC and the formation of
803 palaeoshorelines and shore platforms at multiple depths.

804

805 The section of the Malta Escarpment between Siracusa and the southern Malta Escarpment
806 lacks obvious topographic and seismic evidence of widespread, recent fault activity,
807 confirming inferences made previously that this is concentrated north of Siracusa (Argnani
808 and Bonazzi, 2005). Evidence of recent fault activity in our seismic reflection data is
809 restricted to offsets in the Plio-Pleistocene sedimentary cover in the lower northern Malta
810 Escarpment and north of Cumecs Canyon (Fig. 6b), which we interpret as normal faults.
811 Our data do provide evidence of the geomorphic response to the occurrence of fault
812 structures, however. The perpendicular change in orientation of Noto Canyon (Fig. 7c), as
813 well as the linear, NW-SE oriented tributary of Cumecs Canyon (Fig. 7e), coincide with

faults inferred from seismic data by Casero et al. (1984) (Fig. 2). We therefore propose that such abrupt changes in canyon orientation are fault-controlled (e.g. Noda et al., 2008; Micallef et al., 2014a). The shallow, convex breaks of slope identified across the northern Malta Escarpment may also be seafloor expressions of faults.

5.2.6 Hydrogeological processes

The morphologic and backscatter characteristics of the re-entrants are not consistent with the action of either sediment gravity flows or subaerial fluvial incision. Rather, they share their characteristics with canyons reported on the Florida Escarpment and New Jersey margin. Because of their similarity to subaerial box canyons, such canyons are thought to have been formed by groundwater seepage (Paull and Neumann, 1987; Dunne, 1990; Paull et al., 1990b; Robb, 1990; Twichell et al., 1990; Nagihara, 1996). According to this model, canyons are retrogressively eroded, both mechanically and chemically, into a seepage face. Corrosion, which is enhanced by mixing of solutions of different salinities (Plummer, 1975; Hanshaw and Back, 1980), is thought to be the most important erosional process associated with groundwater seepage in bedrock (Twichell et al., 1990). Groundwater seepage at the base of carbonate escarpments is mainly a result of fluid circulation driven by lateral thermal/salinity differences (e.g. Kohout, 1965; Paull and Neumann, 1987; Chanton et al., 1991; Hughes et al., 2007).

Groundwater seepage may be an important agent forming re-entrants across the northern Malta Escarpment. Erosion by groundwater corrosion would explain the general absence

of debris aprons at the mouths of the re-entrants. Fluid circulation across the escarpment is plausible, either due to a combination of excess pore pressure (due to overburden) and faulting, or geothermal convection. There are a number of uncertainties associated with the role of groundwater seepage in eroding re-entrants, however. First, we do not have direct or indirect evidence of fluid seepage at the base of the Malta Escarpment. Second, and perhaps more important, is the fact that the efficacy of fluid seepage as an erosive agent, both in subaerial and submarine contexts, still needs to be validated (Irwin et al., 2006; Lamb et al., 2006).

5.3 Implications for the MSC in the eastern Mediterranean Sea

5.3.1 Extent of evaporative drawdown in the eastern Mediterranean Basin

The submarine canyons, palaeoshorelines and shore platforms across the Malta Escarpment are all indicative of base-level fall in the eastern Mediterranean Basin during the MSC. By taking into consideration the results of the isostatic restoration, and assuming that the eastern Pelagian Platform has not been tectonically displaced since the Late Miocene (Argnani and Bonazzi, 2005), the palaeoshorelines point to a drawdown of 1800 – 2000 m in the eastern Mediterranean. This value is in line with the observations by Gargani and Rigollet (2007) from the Nile Delta, and by Ryan (1976), Stampfli and Hoker (1989), Lofi (2002), and Steckler et al. (2003) for the western Mediterranean Basin.

5.3.2 Timing of evaporative drawdown in the eastern Mediterranean Basin

860

861 The thick lobes in the western Ionian Basin only occur in facies F, pre-date the evaporite
862 sequence and spatially correlate with canyon mouths. We interpret these lobes as sediment
863 deposits sourced by canyon erosion of the Malta Escarpment during the second step of the
864 MSC, thus forming what has been called a Complex Unit (Clauzon et al., 1996; Lofi et al.,
865 2011a). The fact that the complex palaeo-seafloor topography associated with these lobes
866 is infilled and smoothened by subsequent deposition of the salt layer (facies E2; Fig. 11c)
867 may suggest that the large-scale drawdown occurred before the salt deposition, in
868 agreement with scenario (iii) described in section 1.2 and proposed by Bache et al. (2009;
869 2015). However, the salt layer at this location is thin, possibly deposited in a marginal
870 setting, not necessarily in lateral continuity with the thicker and deeper basin salt body
871 buried below the Ionian Basin. A similar situation is present on the lower slope of the steep
872 Emile Baudot Escarpment on the southern margin of the Balearic Promontory in the
873 western Mediterranean Sea (Dal Cin et al., 2016; their fig. 10). In this case, the deposition
874 of the marginal, shallower salt deposition occurred above the clastic lobes deposited
875 simultaneously with the evaporative shrinking in the deeper basin, in accordance with the
876 models by Ryan (2008b) and Lofi et al. (2011b).

877

878 **5. CONCLUSIONS**

879

880 We used high resolution geophysical and sedimentological data from the central
881 Mediterranean Sea to reconstruct the geomorphic evolution of the Malta Escarpment and

infer the extent and timing of Messinian evaporative drawdown in the eastern Mediterranean Basin. Our main conclusions are the following:

(i) Submarine canyons are the most common landform across the Malta Escarpment and were predominantly driven by subaerial fluvial erosion during the MSC. After refilling of the Mediterranean Basin at the end of the MSC, submarine canyons continued to be eroded by submarine gravity flows. The latter were sourced by slope instability in the outer Malta Plateau and canyon walls and heads on the Malta Escarpment. During sea-level lowstands, subaerial fluvial discharge and possibly hyperpycnal flows could have also generated gravity flows. Episodic gravity flow activity has taken place as recently as <2600 cal. years BP.

(ii) The submarine canyons, palaeoshorelines and shore platforms are indicative of base-level fall during the MSC. Isostatic restoration of the palaeoshorelines points to a drawdown of 1800 – 2000 m in the eastern Mediterranean Sea.

(iii) The thick pre-evaporite sedimentary lobes in the western Ionian Basin margin are interpreted as part of a Complex Unit. The occurrence of this unit would suggest either that evaporative drawdown and canyon formation occurred predominantly before salt deposition (in agreement with scenario (iii) proposed by Bache et al., 2009; 2015), or that only the latest salt deposition at the basin margin, near the lateral pinch-out of the evaporites, occurred after the formation of the sedimentary lobes.

(iv) The post-MSC sedimentary environment in the Malta Escarpment and western Ionian Basin is dominated by hemipelagic sedimentation and deposits associated

905 with bottom current activity, such as sediment waves and contouritic drift deposits.

906 The latter involved strong and long-term bottom currents that flowed along the
907 entire depth range of the Malta Escarpment after the end of the MSC.

908 (v) Slope failures – predominantly translational slides and debris flows - are
909 widespread and affect Plio-Pleistocene sediments. They are preconditioned by
910 over-steepening and loss of support due to canyon erosion, and triggered by
911 earthquakes. However, they are small in scale and not a significant escarpment
912 formation process.

913 (vi) Recent tectonic activity is restricted to the southern Malta Escarpment and was
914 associated with extensional and strike-slip kinematics. Heron Canyon likely
915 comprises the eastern extremity of a wrench zone. We estimate that, during the last
916 5.3 Ma, uplift in the southern Malta Escarpment could have been at least 335 m
917 higher than the northern and central Malta Escarpment. This would explain the
918 shallower depth of palaeoshorelines in the southern Malta Escarpment, although
919 differential subsidence and base-level fluctuations may also have played a role.
920 Evidence of recent tectonic activity in the section between Siracusa and the
921 southern Malta Escarpment is limited to minor normal faults displacing the Plio-
922 Pleistocene sedimentary cover in the lower northern Malta Escarpment and north
923 of Cumecs Canyon.

924 (vii) Re-entrants located along the base of the Malta Escarpment are inferred to
925 be associated with groundwater seepage erosion/weathering on the basis of
926 geomorphology alone, in the absence of other supporting evidence.

This study has demonstrated that erosional and depositional processes have played an important role in shaping the Malta Escarpment. Some of these processes are widespread and still active today, and they need to be taken into consideration when assessing risk associated to marine geohazards in the near future. A more in-depth study of the contouritic drift deposits and sediment waves, on the other, will yield valuable information on the oceanographic evolution of the western Ionian Basin after the MSC.

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7. DATA AVAILABILITY

The multibeam echosounder data, multichannel seismic reflection profiles (from MS, CROP, CIRCEE-HR, CUMECS-3), and gravity core data are available from the authors upon reasonable request.

The multichannel seismic reflection profiles from CA-99 and MEM-07 are available from SPECTRUM but restrictions apply to the availability of these data, which were used under license for the current study, and so are not publicly available. Data are thus available from the corresponding author upon reasonable request and with permission of SPECTRUM.

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1393 **9. FIGURES**

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1396 Figure 1: (a) Location map of the central Mediterranean Sea. 1: Valencia Basin, 2: Ebro
 1397 Margin, 3: Gulf of Lions, 4: Nile Delta, 5: Levant Margin. (b) Bathymetric map of the
 1398 eastern margin of the Pelagian platform and western Ionian Basin. The map displays the

1399 principal morphological features, accretionary wedge thrust structures (solid white lines,
1400 with white teeth at deformation front), structures proposed as surface expressions of STEP
1401 fault and Alfeo-Etna fault (yellow line), and the Ionian fault (orange line) (Gutscher et al.,
1402 2016; Polonia et al., 2016). The main pathways of Modified Atlantic Water (MAW) and
1403 Levantine Intermediate Water (LIW) in eastern Sicily Channel are denoted by purple lines
1404 (Béranger et al., 2004; Ciappa, 2009). Background bathymetry is from Gutscher et al.
1405 (2017) and EMODnet bathymetry (<http://www.emodnet-bathymetry.eu>).
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1408 Figure 2: Map of the Malta Escarpment showing the spatial coverage of new and published
 1409 multibeam echosounder, multi-channel reflection seismics, gravity cores and dredged
 1410 samples used in this study. Location of figures 6, 11 and Oreste_001 well is indicated. Map

1411 includes location of earthquakes with a strike-slip focal mechanism recorded since 1985
1412 (<http://info.terremoti.ingv.it/>). The background bathymetry is from Gutscher et al. (2017)
1413 and EMODnet bathymetry (<http://www.emodnet-bathymetry.eu>). The mapped faults
1414 (black lines) are from Lipparini et al. (2009), Gardiner et al. (1995) and Casero et al. (1984).
1415

Figure 3: (a) Compilation of multibeam bathymetry data from the study areas. The pink line indicates the coastline during the Last Glacial Maximum (22 ka BP). The background bathymetry is from Gutscher et al. (2017) and EMODnet bathymetry

1420 (<http://www.emodnet-bathymetry.eu>). (b) Multibeam backscatter data from the northern
1421 Malta Escarpment. It includes enlarged sections of backscatter maps of examples of
1422 canyons and re-entrants. The isobaths, denoted in red, are at 1000 m intervals. The location
1423 of figures 7, 9 and 10 is indicated.
1424

Figure 4: Map of principal morphologic elements and seismic facies and features identified across the study area, draped on a shaded relief map of the seafloor. Figure 4a displays the north-western part of the study area; its location roughly coincides with that of figure 3b. Figure 4b shows the south-eastern continuation.

Figure 5: Multichannel seismic reflection profile MEM07-104 correlated with Oreste 001 well (189 m water depth) and DSDP 374 (4088 m water depth; 100 km to the south-east of profile MEM07-104). (Facies A = high amplitude, sub-parallel, continuous reflectors; facies B = sub-parallel, convex upward reflectors; facies C = transparent facies; facies D = chaotic to transparent facies; facies E1 = highly reflective, discontinuous facies; facies E2 = reflector-less facies with basin-fill geometry; facies F = lenses of intermediate amplitude and sub-parallel reflectors). Location of profile MEM07-104 is in the middle of the northern Malta Escarpment.

Figure 6: (a) Multichannel seismic reflection profile CIR-03 intersecting the northern Cumecs Canyon, showing thin parallel seismic reflector packages (facies A), or absence thereof, on the steeper parts of the northern Malta Escarpment. On gentler gradients, these packages have a mounded morphology (facies B) and reach a thickness of 0.3 s TWTT. Location of profile CIR-03 is shown in figure 2.

Figure 7: (a) Multibeam bathymetric map of canyons. Inset shows enlarged section of crescent-shaped scars along the bed of a canyon. (b) Profiles of canyon thalwegs (location of profile CIR-03 is shown in figure 2).

in figures a and g). (c) Multibeam bathymetric map of Noto Canyon. (d) Profile CIR-07, which provides a cross-section of Noto Canyon. Location of profile in figure 7c. (e) Multibeam bathymetric map of Cumecs Canyon. (f) CROP profile M-23A, which provides a cross-section of the mouth of Cumecs Canyon. Location of profile in figure e. (g) Multibeam bathymetric map of re-entrants at the base of the northern Malta Escarpment. (h) Multibeam bathymetric map of the central Malta Escarpment. Locations of figures a, c, e, g and h are shown in figure 3. (Facies A = high amplitude, sub-parallel, continuous reflectors; facies C = transparent facies; facies D = chaotic to transparent facies).

Figure 8: (a) CUMECS-2 gravity core correlation panels, including core photographs, x-radiographs, description, and P-wave velocity, gamma ray density and magnetic susceptibility plots. (b) Gravel, pebbles and a cobble retrieved from the core catcher at CU14_04.

Figure 9: (a) Slope gradient map of the promontory in the northern Malta Escarpment showing a concave break of slope (denoted by black arrows) and associated terrace below. Isobaths at 1 km interval shown as black lines. (b) Slope gradient map of a convex break of slope (denoted by white arrows) in the northern Malta Escarpment. Isobaths at 1 km interval shown as white lines. (c) Multichannel seismic reflection profile CIR-04

intersecting the break of slope (denoted by dotted black line) and associated terrace. (Facies B = sub-parallel, convex upward reflectors; facies C = transparent facies). Location in (a). (d) Histogram of the depth of breaks of slope in the northern Malta Escarpment. (e) Depressions (denoted by black arrows) located at the base of the central part of the promontory. (f) Shallow arcuate scars (denoted by black arrows) on the northern interfluvium of Cumecs Canyon. Location of figures is shown in figure 3.

Figure 10: (a) Multibeam bathymetry map of the southern Malta Escarpment. The strongest earthquake recorded on the northern flank of Heron Canyon (M4.4; 30-10-2016) (<http://info.terremoti.ingv.it/>) is shown. (b) Longitudinal profile along the lower half of Heron Canyon, showing the three morphological steps along the canyon axis. Location in (a). (c) Interpreted multichannel seismic reflection profile MS-20 located upslope of Heron Canyon. (d) Interpreted multichannel seismic reflection profile MS-14 located upslope of Heron Canyon. Location of MS profiles in figure 2. (Facies A = high amplitude, sub-

1489 parallel, continuous reflectors; facies D = chaotic to transparent facies; facies G = low
 1490 amplitude, sub-parallel and discontinuous reflectors).

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Figure 11: (a) Interpreted multichannel seismic reflection profile CIR-04, intersecting the base of the central Malta Escarpment and the western Ionian Basin. (b, c) Interpreted multichannel seismic reflection profile CUMECS-3, located in the western Ionian Basin and parallel to Malta Escarpment. Location of profiles in figures 2 and 9e. (Facies A = high amplitude, sub-parallel, continuous reflectors; facies B = sub-parallel, convex upward reflectors; facies C = transparent facies; facies D = chaotic to transparent facies; facies E1 = highly reflective, discontinuous facies; facies E2 = reflector-less facies with basin-fill geometry; facies F = lenses of intermediate amplitude and sub-parallel reflectors). Figure c is redrawn from Micallef et al. (2018).

Figure 12: Map showing isolines of post-MSC subsidence (in metres) of the seafloor across the study area, based on a flexural isostatic compensation of the weight of post-MSC sediment and the water column, and using an equivalent elastic thickness (EET) of 50 km and 40 km. The tilting towards the east reflects the deeper water column loaded in that region relative to the Pelagian Platform in the west. Isobaths at 1 km interval shown as black lines.

10. TABLES

Table 1:

Station	Latitude (°N)	Longitude (°E)	Water depth (m)	Recovery (m)
CU-01	36.823872	15.370824	2307	2.25
CU-02	36.814108	15.373017	2281	2.85
CU-03	36.716713	15.473141	2835	0
CU-04	36.043433	15.556455	2578	0.42
CU-05	36.150213	15.588011	2915	0
CU-06	36.841167	15.210168	133	1.38
CU-07	36.84200	15.232400	230	2.16
CU-08	36.83330	15.31000	1830	1.11

Table 2:

	Mean length (km)	Mean length:width ratio	Mean spacing (km)	Mean bifurcation angle (°)	Mean wall gradient (°)
Northern Malta Escarpment - canyons	7.9	2.38	0.7	20	22
Northern Malta Escarpment – re-entrants	1.5	0.75	5.0	45	52
Central Malta Escarpment - canyons	8.7	3.21	2.2	18	18

Southern Malta Escarpment - canyons	9.1	3.31	2.18	21	18
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1517 **11. TABLE LEGENDS**

1518

1519 Table 1: Location, depth and recovery of the gravity cores collected during the CUMECS-
1520 2 oceanographic survey.

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1522 Table 2: Morphometric parameters of submarine canyons and re-entrants.