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Key Points:

- The topographic anomaly of the Hyblean Plateau (Sicily) is associated to a major gravimetric and magnetic anomaly within the Mediterranean
- Plate flexure induced by the Sicilian orogeny is insufficient to explain the Hyblean Plateau domal topography and geophysical anomalies
- The Hyblean dome-shape relief originated from Late Miocene and Plio-Pleistocene mafic sheet intrusions in the upper crust

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Deep Origin of the Dome-Shaped Hyblean Plateau, Southeastern Sicily: A New Tectono-Magmatic Model

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Abstract Southeastern Sicily is characterized by a prominent topography known as the Hyblean Plateau. It is commonly considered as a flexural bulge formed by the northwest plunging of the Hyblean-Malta platform beneath the Sicilian Fold and Thrust Belt. However its noncylindrical shape and multiphase uplift history correlated to the Late Miocene and Plio-Pleistocene volcanism outline a possible magmatic control on the Hyblean Plateau topographic expression. We argue, using 2-D flexural models, that the flexural bending alone is unlikely to explain the high anomalous elevation of the Hyblean Plateau. Moreover, topographic analysis show that the Hyblean Plateau has a dome shape composed of radially tilted preserved surfaces defining individual blocks. The block edges seem to be related to deep-seated inherited faults. Using an analogue modeling approach, we show that magmatic intrusions beneath the thick Meso-Cenozoic sedimentary cover can reproduce the first-order morphology of the Hyblean Plateau when a structural inheritance is imposed. Accordingly, we propose a tectono-magmatic model that favors mafic sill intrusions as the major uplift source for the topographic anomaly. However, to explain the strong positive Bouguer anomaly, a deeper high-density intracrustal body is needed. Moho uprising related to Mesozoic rifting or mafic lower crust underplating/intraplating since the Late Cretaceous can be evoked. The small wavelength of the topography and the deep intracrustal mass anomaly makes the lithospheric elastic stresses sufficiently high to maintain a nonisostatic state over the geological time scales.

Plain Language Summary A prominent dome-shaped topography known as the Hyblean Plateau overcomes southeastern Sicily. It is commonly considered as originated from foreland flexural bulging of the Hyblean-Malta platform in response to the southeastward migrating Sicilian orogenic wedge. However, its nonlinear shape and its multiphase uplift history correlate to the Late Miocene and Plio-Pleistocene volcanic activity and hence suggest a possible magmatic control on the formation of the Hyblean Plateau. We show that flexural bending alone is unlikely to explain the subcircular topography of the Hyblean Plateau using 2-D flexural models. Moreover, subsurface analysis coupled to an analogue modeling approach shows that the subcircular pattern is subdivided into radially tilted blocks whose edges seem to be related to inherited fault zones. The tectono-magmatic model proposed here minimizes the flexural component to the Hyblean rise and prefers mafic sill intrusions in the upper crust as the major uplift source. However, gravity anomalies suggest a deeper high-density intracrustal body beneath the Hyblean Plateau which can be related to the Meso-Cenozoic history. The lack of isostatic equilibrium in this region can be related to the small length scale of the topography and the inferred intracrustal high-density body which makes the lithosphere rigid enough to support these anomalies.

1. Introduction

The Hyblean-Malta Platform in the Central Mediterranean represents the northernmost promontory of the African continental margin (Patacca et al., 1979) emerging in the southeastern corner of Sicily (Italy), where it forms the so-called Hyblean Plateau (Figures 1a and 2). This anomalous relief is associated to strong gravity and magnetic anomalies and has been affected by a long multiphase volcanic activity up to Quaternary.

To date, the structure of the Hyblean Plateau (HP; Figure 1) has been interpreted as a positive flexural bulge at the front of the Sicilian chain generated by the orogenic load and the slab pull of the subducting Ionian

Figure 1. Tectonic sketch map, geoid, and magnetic anomalies of the Central Mediterranean domain. Main geographic, geologic, and volcanic areas: Sicilian Fold and Thrust Belt (SFTB), Etna (E), Hyblean Plateau (HP), Hyblean-Malta Platform (HMP), Malta Escarpment (ME), Malta (M), Pantelleria island (P), Vulcano (V), Nameless Bank (NB; Lodolo et al., 2012). (a) Main tectonic structures overprinted on the topography-bathymetry compilation from Gutscher et al. (2017). Earthquakes database from <http://earthquake.usgs.gov> including magnitude greater than three events for the 1973–2018 period. Yellow arrows are GPS velocity vectors from www.unavco.org in an African fixed reference frame. (b) Geoid height anomalies from <http://icgem.gfz-potsdam.de> and (c) magnetic anomalies from <http://geomag.colorado.edu>.

slab (Billi et al., 2006; Cogan et al., 1989; Pedley & Grasso, 1992). The flexural assumption is mainly supported by two-dimensional sections across the southeastern front of the chain, the so-called Gela Thrust Front (GTF). However, the peculiar noncylindrical shape of the Hyblean Plateau seems inconsistent with a simple tectonic flexure of the Hyblean crust along the NE-SW trending front of the SFTB, considering flexural mechanics (Turcotte & Schubert, 1982). The Hyblean Plateau has indeed a maximum elevation of ~1,000 m above sea level and its topography appears to progressively plunge radially toward the surrounding Ionian and Sicily Channel seas.

Apart from the widely studied tectonic evolution of the Hyblean region, the magmatic history of eastern Sicily is also a matter of debates in terms of origin and magma pathways. The last cycle of volcanic activity in southeastern Sicily affected the northern flank of the HP (Figure 2) and occurred from the Upper Miocene to Lower Pleistocene (Beccaluva et al., 1998; Carbone et al., 1987; Manuella et al., 2013; Patacca et al., 1979) before shifting further North to feed Mount Etna during the last 0.5 Ma (Branca et al., 2008; Schmincke et al., 1997). The origin of this anomalous anorogenic magmatism has been attributed to extension (Barberi et al., 1973) similarly to the Pantelleria and Linosa rifting volcanism affecting the Sicily Channel or to a deep-rooted mantle plume (Tanguy et al., 1997). More recently, the upper mantle return-flow around the Ionian slab edges, which could explain the dynamic uplift of north-eastern Sicily and Calabria (Faccenna et al., 2005), has been proposed to be the source of this anomalous anorogenic magmatism (Schellart, 2010; Strak & Schellart, 2014).

Earth's subsurface uplift in magmatic provinces is common (e.g., Acocella, 2000; Chen et al., 2015; Klügel et al., 2015; MacLennan & Lovell, 2002; McKenzie, 1984; Perkins et al., 2016; Rohrman & van der Beek, 1996; Wellman, 1986) and numerous processes involving magmatism, directly or indirectly, are suggested to account for surface doming at various scales. Such processes include shallow magmatic intrusions (e.g., Acocella, 2000; Castro et al., 2016; Wellman, 1986), serpentine-bearing clayed diapirs (e.g., Barreca, 2014; De Guidi et al., 2014; Giampiccolo et al., 2017; Manuella et al., 2012), crustal magmatic underplating and/or intraplate (e.g., Chen et al., 2015; Cox, 1993; Klügel et al., 2015; MacLennan & Lovell, 2002; Perkins et al., 2016), upper mantle upwelling (e.g., Hirn et al., 1997; Rohrman & van der Beek, 1996; Wilson et al., 2010), or mantle plumes (e.g., Crough, 1978). The doming of the surface can be a permanent or a transient process whether the region is in isostatic steady state or not (Rohrman & van der Beek, 1996). Although, the cause-and-effect relationship between magmatism and surface uplift of the Hyblean Plateau has been vaguely evoked (Schmincke et al., 1997; Yellin-Dror et al., 1997), no studies have focused on this issue, motivating a reassessment of the Hyblean Plateau origin.

The aim of this study is therefore to analyze jointly all the geological and geophysical data available to provide a new hypothesis concerning the origin of the Hyblean Plateau dome-shape relief, challenging the unlikely flexural hypothesis. The topography of the Hyblean Plateau (HP) was thus investigated by means of a morphostructural analysis, gravimetric forward modeling, and finally, using analogue modeling to validate mechanically our final hypothesis. The outcomes point out that the topographic anomaly of the Hyblean Plateau is most probably related to domal uplift induced by deep viscous intrusions at crustal and lithospheric scales, and that the current tectonic pattern of the area reflects a structural inheritance interacting with the tectono-magmatic processes.

Figure 2. Geological and structural sketch map of the Hyblean Plateau. Bathymetry compilation from Gutscher et al. (2017) and topography data from the Japan Aerospace Exploration Agency (<https://www.eorc.jaxa.jp/ALOS/en/aw3d30/index.htm>). (a) Onshore tectonic structures are from Catalano et al. (2010) and offshore faults are from Gutscher et al. (2017). Main tectonic features: Scordia Lentini, Augusta, and Florida grabens (SLG, AG, and FG) and fault zones: Comiso-Chiaramonte Fault Belt (CCFB), Monterosso-Agnone Fault System (MAFS), Scicli-Ragusa Fault System (SRFS), and Pozzallo-Ispica Fault System (PIFS). Late Tortonian to Quaternary volcanic activity from Behncke (2004). (b) Oblique (20°) view from the south with a vertical exaggeration (2X).

2. Regional Setting

2.1. General Geodynamic Context

The Hyblean Plateau (HP) is bounded by two major active tectonic structures: to the NW, by the Sicilian Fold and Thrust Belt (SFTB) and to the E, by the Malta Escarpment (ME; Figures 1a and 2a). The SFTB is part of the Apennine-Maghrebian orogenic chain (Figure 3) that overthrusts the African Pelagian margin during the Neogene-Quaternary times. Indeed, while the Peloritani-Calabria block started to collide with the Pelagian margin in the Middle-Upper Miocene, the free boundary to the east, constituted by the Ionian realm, allowed further southeastward migration of the Calabrian Arc up to its present location (Figure 1a). Presently, the frontal part of the SFTB is separated from the emerged Hyblean Meso-Cenozoic

Figure 3. Schematic tectono-magmatic scenario of the evolution of the Sicilian collision system (not to scale). From Tortonian to Late Quaternary, the Ionian Paleobasin has been progressively consumed in response to slab roll-back and the Calabria-Peloritani terranes collided with the Pelagian promontory. Meanwhile, over the Hyblean-Malta Platform, the episodic volcanic activity never ceased during that period. The Hyblean Plateau likely uplifted during the Late Miocene and Plio-Pleistocene times. Note that the position of the Hyblean anomalous relief remained stable during that period, whereas the front of the SFTB migrated southeastward for more than 100 km.

carbonates (Ben-Avraham & Grasso, 1991) by a narrow foredeep filled by syntectonic Quaternary deposits (Figure 2) extending from Catania to Gela (e.g., Butler et al., 1992; Ogniben, 1969). On the eastern side of the HP, the Malta Escarpment (ME) has an older geological history inherited from the Late Paleozoic-Early Mesozoic rifting of the Neo-Tethys, and constitutes a major NNW-SSE fault-controlled submarine slope which extends to the south for ~200 km (Scandone et al., 1981).

2.2. Stratigraphic Constraints and Volcanism

The ~8-km-thick Meso-Cenozoic sedimentary cover of the HP has been drilled by deep boreholes up to the Middle-Triassic dolomites at a depth of ~5–6 km (Bianchi et al., 1987). The outcropping foreland units are mainly Late Cretaceous-Late Miocene in age and show carbonate sequences from shallow to deep-water facies (Bianchi et al., 1987; Patacca et al., 1979). Messinian evaporites are missing on the central and northeastern part of the Hyblean Plateau suggesting that this part of the foreland was already emerged ~6 Ma ago (Figure 3; Grasso & Lentini, 1982; Micallef et al., 2018). Then, the Hyblean Plateau edges returned underwater during the Early Pliocene transgression (Pedley et al., 2007), a stage characterized by the deposition of shallow-water marine sediments (locally known as Trubi formation) all around the plateau. Since then, marine to continental Pleistocene deposits mainly filled structural linear depressions on the Hyblean Plateau edges (Figure 2) such as the grabens of the northern and eastern sectors (Scordì-Lentini, Augusta, and Floridia grabens) as well as the Gela-Catania foredeep (Figure 3; Grasso & Lentini, 1982).

During the Meso-Cenozoic, four main volcanic cycles affected the HP (Patacca et al., 1979). The last cycle took place during the Late Miocene-Early Quaternary, emitting alkalic and tholeiitic mafic lavas (Figures 2 and 3) and shifted progressively to the northern sector of the plateau (Neri et al., 2018; Patacca et al., 1979). Eruptive centers in the Late Miocene (Tortonian to Messinian) were mainly shallow submarine or subaerial diatremes and a few NW-SE oriented eruptive fissures (Schmincke et al., 1997). The eruptions then localized along NE-SW trending fissures during the Plio-Pleistocene, firstly under water and lately in subaerial environments (Neri et al., 2018; Schmincke et al., 1997). In the Late Pliocene, a major volcanic phase (~2 Ma; Schmincke et al., 1997), known as the Militello tholeiites lavas (Figure 2), took place in a shallow marine environment, grading upward into subaerial lava flows, and possibly emitted a large volume >120 km³ of lava (Behncke, 2004). Finally, after the emission of shallow water-subaerial alkalic products during the Early Pleistocene, volcanic activity in the Hyblean region ceased around ~1.4 Ma.

2.3. Fault Network

The tectonic pattern (Figure 2) of southeastern Sicily is characterized by a diffuse network of variously oriented faults with mostly normal or strike-slip components (Cultrera et al., 2015). The main extensional faults network are subparallel to the NE-SW oriented Gela Thrust Front (GTF), such as the Monterosso-Agnone Fault System to the north, the Pozzallo-Avola Fault System to the SE, and the Comiso-Chiaramonte Fault Belt which downfaulted the western sector by ~4,000 m (Cogan et al., 1989). These faults exhibit significant offsets on the NW and SE boundaries of the Hyblean Plateau and are likely partly related to the extrados extension of the downgoing Hyblean crust (Grasso & Pedley, 1990). By contrast, a second major set of tectonic lineaments, NW-SE oriented, is particularly marked by on-land grabens (e.g., Augusta and Floridia basins) filled by Pliocene to Late Quaternary marine sediments in the NE sector of the HP (Bianca et al., 1999). Furthermore, recent geological mapping shows on the

southwestern side of the HP a NW-SE trending post-Tortonian fault zone along the Tellaro river, whose polyphased tectonic history may reflect a crustal discontinuity between the eastern Syracuse and the western Ragusa domain (Romagnoli et al., 2015). The western sector of the Hyblean Plateau is also affected by an ~70-km-long N-S trending fault zone, known as the Scicli-Ragusa Fault System (Ghisetti & Vezzani, 1980), which likely continues northward for ~40 km and possibly offshore to the south (Gardiner et al., 1995).

Present-day kinematics shows that the Hyblean region is now experiencing a NNW oriented contraction as revealed by geodetic data (Bonforte et al., 2015; Hollenstein et al., 2003; Mastrolembo Ventura et al., 2014), boreholes stress measurements (Ragg et al., 1999), and seismogenic stress tensors (Musumeci et al., 2014). Evidences for fault reactivation over the HP (from pure-normal to strike-slip) come only from the analysis of instrumental seismicity (Cultrera et al., 2015) and are consistent with a NNW direction of crustal compression probably related to the Africa-Eurasia plate convergence. However, most of the well-known faults do not show any evidence of tectonic inversion, testifying of an incipient process whose origin is beyond the scope of this study.

2.4. Uplift History

Vertical motions in southeastern Sicily result from a complex interplay between local subsidence or uplift, eustatic sea-level changes, and sedimentation-erosion processes. Nevertheless, four main vertical motion events were recognized by one-dimensional Airy backstripping technique applied to seven wells located north of the Hyblean Plateau (Yellin-Dror et al., 1997): (1) a Late Triassic rifting subsidence, (2) an Early Jurassic-Late Cretaceous thermal subsidence, (3) an uplift phase during the Paleogene, and finally, (4) the uplift events during the Neogene-Quaternary period (Figure 3). During the later phase, the HP partly remained above sea level during the Pliocene-Early Pleistocene subsidence of its margins (Di Martire et al., 2015; Grasso & Lentini, 1982; Schmincke et al., 1997; Yellin-Dror et al., 1997) and the coeval sea-level rise on the whole Mediterranean (Butler et al., 1995). Following the Early Pleistocene vigorous Militello volcanic phase (Figure 2), a major renewed uplift occurred during the Middle Pleistocene as testified by the presence of Lower Pleistocene calcarenites ~600 m above sea level (near Palagonia village) and raised marine terraces all around the Hyblean Plateau (Schmincke et al., 1997). The uplift rate during this later stage was estimated to about 0.2–0.5 mm/year (Bonforte et al., 2015; Schmincke et al., 1997). Unfortunately, despite no satisfactory explanations for the multistage uplift, and, in particular, the significant mid-Pleistocene central bulging, the complex vertical motion of the Neogene-Quaternary period was regularly attributed to flexural tectonics (Billi et al., 2006; Di Martire et al., 2015; Yellin-Dror et al., 1997).

3. Multidisciplinary Morphostructural and Geophysical Modeling

We adopt a multidisciplinary approach to (1) characterize the anomalous topography of the Hyblean Plateau, (2) outline the weaknesses of the forebulge hypothesis, (3) analyze geophysical data at the crustal scale such as the gravity and magnetic signals, and (4) discuss the results and propose an integrative model. Topographic analysis, 2-D flexural modeling, magnetic data set, and gravity modeling have been used to investigate the morphology and geophysical signature of the HP while analogue models (further described in section 5) allowed to illustrate and test mechanically our proposed tectono-magmatic scenario.

3.1. Geomorphic Analysis

1. Long-wavelength topographic filtering

To investigate the long-wavelength topographic signature of the HP, we processed elevation data (Figures 4a and 4b) from the global DEM ALOSWorld3D-30m (AW3D30, released in 2015 by the Japan Aerospace Exploration Agency). We first removed the drainage network by subtracting to the raw data, the smoothed topography obtained with a 2-km wavelength median filter. The high-frequency topographic noise was also removed by keeping only the topographic information handled by preserved surfaces with slopes <5°. The nearest *neighbor* algorithm from the Global Mapping Tool (GMT) software (Wessel & Smith, 1991) was then used to reconstruct the topography from the masked data set. Finally, a 20-km wavelength low-pass Gaussian filter was applied to extract the long-wavelength signal, highlighting the first-order shape of the Hyblean Plateau.

Figure 4. Topographic analysis of southeastern Sicily. (a) Topography data from the Japan Aerospace Exploration Agency (<https://www.eorc.jaxa.jp/ALOS/en/aw3d30/index.htm>). (b) Reconstructed topography from the masked data set (drainage network and slopes $>5^\circ$ removed) and filtered with a 20-km wavelength low-pass Gaussian filter. The dashed grey circle represents the circumference of the topographic anomaly. (c) Morphological units selected from the masked data set used to extrapolate surfaces. (d) Modeled surfaces and their mean orientation and mean slope (black arrows). (e) Residual between the topographic data (c) and the modeled surfaces (d). (f) Extended modeled surfaces and main inferred structural boundaries (dashed lines). Surface's name: Militello (M), Vittoria (V), Ragusa West (RW), Ragusa East (RE), Avola (A), and Carlentini (C).

2. Subcrustal block analysis

The low local relief associated to preserved surfaces all over the HP also displays a first-order morphologic pattern, where radially tilted blocks appear to shape its topography. We tested this assumption by decomposing the present day topography into six gently deformed panels. To that extent, the masked data described above (drainage network and slopes $>5^\circ$ removed) are clipped into six polygonal surfaces based on our visual assessment and taking into account the major tectonic lineaments outlined in the “Fault network” section (Figure 4c; M = Militello, V = Vittoria plain, RW = Ragusa West, RE = Ragusa East, A = Avola, and

C = Carlentini). Each data set is then modeled by a bilinear trend surface with relatively good fits for the six defined panels (Figure 4e; RMSD rank from ~30 m (Vittoria surface) to ~76 m (Carlentini surface)).

3.2. 2-D Numerical Flexural Modeling

The noncylindrical deformation of the HP was explained either by a heterogeneous locally thick and buoyant Hyblean crust resisting subduction (Ben Avraham et al., 1995) and a doubly plunging foreland due to differential retreating processes of subduction zones between Sicily and the current Calabrian Arc to the east (Billi et al., 2006). However, this hypothesis is not constrained by geophysical data. Thus, we revisited the flexure of the Hyblean foreland model (Cogan et al., 1989) to showcase the limited role of forebulge development on the Hyblean Plateau formation. Cogan et al. (1989) have used four NW-SE cross sections orthogonal to the GTF to constrain the flexural properties of the Hyblean lithosphere. They found that the basement downbending in the foredeep is best fitted with a flexural rigidity $D \sim 1.5\text{--}7.2 \times 10^{21}$ N/m (corresponding to an elastic thickness $T_e \sim 6\text{--}10$ km) which can produce a forebulge of amplitude $W_b \sim 1,000$ m. However, three of the used cross sections were only representing the foredeep and the fourth one, across the HP, was divided in two individual segments analyzed separately. Furthermore, the basement was taken at the top of the Gela formation (Triassic dolomites) which means the deformation, imprinted from Jurassic to Plio-Quaternary time, were also included, whereas the flexure should postdate the Late Miocene-Early Pliocene building of the SFTB.

In the following subsections, the 2-D numerical modeling methodology used to evaluate the flexural hypothesis is presented. First, the construction of a new reference surface constraining the deformation accumulated since the Late Miocene is detailed, and then, the modeling procedure is described.

3. Miocene-Pliocene interface modeling

As the major uplift phase of the Hyblean Plateau is here assumed to be of Upper Pliocene to Pleistocene in age, we selected the base of the Pliocene as a preuplift reference surface. The base of the Pliocene has been determined from the analysis of onshore/offshore well data (DGS-UNMIG; <http://unmig.sviluppoeconomico.gov.it>) and geological maps (Grasso, 1997; Lentini et al., 1987). Then, we computed a smoothed surface of the base of Pliocene units (Figure 5a) using the GMT software (Wessel & Smith, 1991).

4. The 2-D numerical modeling of the flexure

Two NW-SE profiles orthogonal to the GTF have been extracted from the extrapolated surface of the Pliocene base (Figure 5b). One is crossing the Hyblean Plateau summit (A) and the other is taken ~45 km to the SW (B) as far as possible of the Hyblean topographic anomaly. To converge toward realistic and comparable flexural models, a far-field common basement depth is imposed in accordance to the Pliocene base of the southeasternmost well data (Spigola Mare 001, -161 m; Figure 5a). Both profiles are thus extended to the SE and sampled over a length of 155 km with a 1-km step (light red, A, and blue, B, circles in Figure 5b). Profile B on the edge of the Hyblean anomaly is assumed to be the upper bound of the potentially flexed lithosphere. We then try to fit these profiles with a 2-D elastic flexural model assuming a point load at the extremity of a semiinfinite elastic plate of constant thickness T_e overlying an inviscid foundation (Turcotte, 1979). The analytical solution along the x axis of the vertical deflection of the plate $w(x)$ is given by

$$w(x) = Q_0 \frac{\beta^3}{4D} e^{-x/\beta} \cos\left(\frac{x}{\beta}\right)$$

where Q_0 is the load magnitude (N/m) and β is the flexural parameter which depends on the flexural rigidity D of the plate and the density contrast with the foundation (Turcotte, 1979).

For a first-order approximation, we impose a density contrast of 900 kg/m^3 corresponding to an infill material of density 2400 kg/m^3 with respect to the lithospheric mantle density of 3300 kg/m^3 . The applied force corresponds in nature to the conjugated load from the orogenic wedge and the slab pull. Because the contribution of each of these loads is not straightforward, the load magnitude Q_0 is a free parameter in our calculations. Hence, we run the calculation varying Q_0 and the elastic thickness T_e of the plate to fit both profiles A and B (Figure 6).

Figure 5. Reconstructed base of Pliocene units and flexural models. (a) Contour lines show the modeled surface of the base of Pliocene strata extrapolated from onshore and offshore well data (DGS-UNMIG; <http://unmig.sviluppoeconomico.gov.it>) and picked contacts from geological maps (Lentini et al., 1987; Grasso, 1997). Black crosses show the position of far field imposed depth values. They are assumed to be at the same depth as the Pliocene base in the Spigola Mare Well located in the Hyblean-Malta Platform (SMW, -161 m). The 155-km-long subparallel profiles A and B are orthogonal to the Gela Thrust Front (GTF) and 45 km away from each other. (b) Extracted profiles A and B from the modeled surface of the Pliocene base (A, red and B, blue). Base of Pliocene data are stacked within 10 km from each profile. Extended topographic profiles are plotted in the background for each cross sections (A, light red and B, light blue). (c) Best flexural model outputs for profile A and B. The elastic thickness (T_e) and the load (Q_0) are two model parameters.

3.3. Gravity Modeling

1. Gravity data and processing

We analyze the gravity signal of the HP region using the latest available data set from the TOPEX mission (https://topex.ucsd.edu/cgi-bin/get_data.cgi), which includes the Sandwell V24.1 gravity model (Sandwell et al., 2014; Sandwell & Smith, 1997; Sandwell & Smith, 2009) and the V18.1 topographic model (Smith & Sandwell, 1997). The spectral coherence between free-air gravity anomalies and the topography indicates the amount of gravity anomaly caused by the topography (Watts, 2001). The spectral coherence between the two global 1-min grid size satellite data sets shows a sharp coherency drop from ~ 0.5 to 0 for

Figure 6. Inversion results from 2-D flexural modeling. For each profile A and B the elastic thickness (T_e) and the load (Q_0) were varied. Contours represent the 1- σ (83 m), 1.5- σ (125 m), and 2- σ (167 m) levels of uncertainty. For each couple of parameters (T_e and Q_0), the bulge amplitude is calculated and plotted with dashed black lines (200, 500, 1000, and 1500 m).

wavelength $<20\text{--}30$ km. We, thus, remove the high-frequency signal that is not relevant for crustal modeling by filtering the topography and the free-air anomaly data sets with a 25-km wavelength low-pass Gaussian filter (Figure 7a). Then, the Fortran code FA2BOUG (Fullea et al., 2008) is used to compute the complete Bouguer anomaly of the area (Figure 7b) with a crustal and sea-water reduction density of 2700 and 1035 kg/m^3 , respectively.

The positive Bouguer anomaly associated to the positive topography in southeastern Sicily reveals a nonisostatic equilibrium over the studied region. Using the *gravfft* program of GMT (Wessel & Smith, 1991) the Airy isostatic contribution to vertical gravity anomaly is calculated by estimating the Moho topography in

Figure 7. Shaded gravity anomaly maps centered on southeastern Sicily (see Figures 1a and 1b for location). (a) Free air gravity anomalies from TOPEX mission (https://topex.ucsd.edu/cgi-bin/get_data.cgi) filtered with a 20-km wavelength low-pass filter. (b) Complete Bouguer anomalies computed with FA2BOUG code (Fullea et al., 2008). (c) Isostatic residual gravity anomalies assuming the local Airy compensation model. Sicilian Fold and Thrust Belt (SFTB), Hyblean Plateau (HP), Calabrian Arc (CA), Malta Escarpment (ME), and Sicily Channel (SC). The dashed grey circles mark the circumference of the topographic anomaly as seen in Figure 4b.

Figure 8. Gravity modeling of the Hyblean positive Bouguer anomaly. (a) Bouguer anomalies filtered with a 40-km wavelength low-pass filter. (b) Synthetic anomaly representing the Hyblean positive Bouguer anomaly. (c) Residual after removing the synthetic anomaly component (b) to the complete Bouguer anomalies (a). The Sicilian Fold and Thrust Belt (SFTB) negative anomaly and the Malta Escarpment (ME) positive anomaly are preserved. (d) Gravity modeling output compared to the synthetic anomaly (~180-km-long E-W profile AB, black line). Four scenarios are tested: a 5-km-thick mantle body at the crust base, alike a 5-km shallower Moho with respect to a 30-km continental crust (blue); an ~1.5-km-thick magnetic intrusive body in the upper crust (~8-km depth) with an excess density of 200 kg/m^3 (red); and the combination of the two first scenarios (purple). A fourth scenario involves a thick high-density lower crustal body (green) with a density contrast of 300 kg/m^3 relative to the surrounding crust and might represent mafic underplating or a partly serpentinized mantle (35 to 40% according to Manuella et al. (2013)). (e) Cross section displaying the geometry of the modeled bodies with their corresponding density contrast relative to the crust of mean density 2700 kg/m^3 .

response to the surface topography according to an Airy local compensation model. A 30-km (continental type) reference crustal thickness is assumed for the calculations. Finally, subtracting this component to the Bouguer anomalies gives the isostatic Airy-type residual gravity anomalies shown in Figure 7c.

2. Gravity modeling

To investigate the potential sources of the gravity anomaly beneath the HP, we used the Matlab code Granom (Hetényi et al., 2007) based on Won and Bevis (1987) algorithm, which calculates the vertical gravity anomaly associated to polygonal density contrasts along 2-D profiles. Because the region surrounding the Hyblean Plateau is highly complex, our forward modeling approach use a synthetic anomaly (map view Figure 8b and black curve Figure 8d) with an amplitude of ~55 mGal and a wavelength of ~180 km. Indeed, this anomaly is well representative of the Hyblean gravity high (Figures 7b and 8a) as evidenced by the residual Bouguer gravity anomaly (Figure 8c) that, without modifying the adjacent SFTB and ME anomalies, is equivalent to the far field ~60-mGal anomaly of the southward Hyblean-Malta platform. We use a reference crustal model with a Moho depth of 30 km (Agostinetti & Amato, 2009) and a crust-

mantle density contrast $\Delta\rho_{moho}$ of 500 kg/m³. The ~55-mGal gravity anomaly of the HP is, thus, compared to the gravimetric signal of dense crustal bodies such as mafic intrusions or mantle uprising.

4. Results

4.1. Geomorphic Structuration of the Hyblean Plateau

The Hyblean Plateau culminates at ~1000 m (Monte Lauro) at the core of the topographic high and gradually plunges to sea level except along the GTF from Catania to Gela where it locally remains at ~500 m above sea level. The results of the large-scale topographic analysis (Figure 4b) show that the HP topography can be, at first-order, defined as a subcircular and long-wavelength topographic dome (amplitude >700 m and ~80-km diameter).

According to the second-order topographic feature analysis (Figure 4d), the dome-shape is constituted by a set of simple radially tilted surfaces (Figure 4e). From the model outputs, we also extracted the mean down-slope direction (D_m) and mean slope (S_m) for each surface (Figure 4d). The tilted surfaces, from Carlentini to Ragusa West (Figure 4d), are dipping gently (~1–2°) and their downslope directions evolve clockwise from the NNE to the SW. The Vittoria and the Militello surfaces, located near the front of the GTF, show opposite downslope directions toward the SW and the NE, respectively. It can be noted that these surfaces should dip toward the NW if they were only originated by the flexural downbending of the foreland beneath the SFTB. Moreover, the orientation of some surface edges, which are compatible with old tectonic lineaments such as the NW-SE Malta Escarpment, the limited number of tilted surfaces, and their low internal deformation are compatible with tectonic inheritance. The localization of the deformation along deep-seated basement faults induced the formation of upper crustal blocks during the uplift phase and thus controlled the morphogenesis of the dome.

4.2. Hyblean Crust Flexural Constraints

The 3-D modeling of the Miocene-Pliocene reference surface (Figure 5a) shows a similar subcircular topographic anomaly of the Hyblean Plateau, with a central uplift >700 m, localized close to the present-day summit. In the southern and northeastern sector, the lack of data could result of emersion during the early Pliocene as mentioned before, and/or of later erosion. The two profiles A and B extracted from this reference surface (Figures 5b and 5c) run parallel and are orthogonal to the orogenic front, one is crossing the center of the Hyblean Plateau and the other one is shifted 45 km to the SW.

Based on the two-parameter space inversion, the best fitting flexural models for both profiles A and B (Figure 6) have low elastic thicknesses, 6 ± 2 and 4 ± 1.5 km, respectively, at the 95% confidence level. However, the load magnitude Q_0 corresponding to each profile is significantly different, $3.75 \times 10^{12} \pm 1.25 \times 10^{12}$ and $7.25 \times 10^{11} \pm 4.75 \times 10^{11}$ N/m for profiles A and B, respectively, at the 95% confidence level. The two best fitting models for profile A and B revealed by the inversion procedure are shown in Figure 5 c. These results show that to account for the Hyblean crust geometry in terms of flexural deformation, a low equivalent elastic thickness ($T_e \sim 5$ km) and a very strong lateral differential loading, about 5 times, are required. Such flexural characteristics seem incompatible with a simple flexure of the lithosphere induced by the Sicilian orogeny.

4.3. Magnetic Anomalies

The magnetic data presented in Figure 1c are from the EMAG2 model, a global 2-arc-minute resolution grid compiled from satellite, marine, aeromagnetic, and ground surveys (<http://geomag.colorado.edu>). The main features and interpretations of the magnetic anomalies are discussed below. However, no magnetic modeling has been performed here because (1) it implies reliable physical parameters such as magnetic susceptibility and Curie depths, (2) the dipolar signature of magnetic bodies makes the 3-D signal difficult to interpret in a straightforward way, and (3) beyond the scope of this study, a significant magnetic signature could be inherited from the long-term tectono-magmatic history of this region.

The data reveal high magnetic susceptibility >2000 ($\times 10^{-5}$ SI) and high magnetic intensity values >200 nT (Figure 1c) with characteristic wavelengths of ~100 km all around the HP. Instead of a unique magnetic signal, centered over the HP topographic high and the main eruptive centers, four interacting magnetic anomalies can be observed to the NW, NE, SE, and SW of the HP (Figure 1c). Such setting renders difficult the

identification of the magnetic sources. However, considering the widespread volcanic activity of the Hyblean Plateau, the intrusion of a large mafic body in the upper crust seems the best hypothesis. The complexity of the magnetic signal can be explained by polyphase intrusion events.

4.4. Gravity Models

One of the striking features revealed by the Bouguer gravity anomaly map is the very strong positive anomaly (>100 mGal) over the Hyblean Plateau region. The analysis of the gravity data reveals that the intrusion of an upper crustal (~ 8 -km depth) mafic, and magnetic body to account for the magnetic anomaly, with a density contrast relative to the crust $\Delta\rho_{intrusion}$ of 200 kg/m^3 and a shape mimicking the uplifted Hyblean topography (~ 60 km wide and 1.5 km thick in the center) produce an ~ 10 -mGal anomaly over a length scale of ~ 80 km (Figures 8d and 8e). Mafic upper crustal sheet intrusions alone are thus unlikely to produce the major Bouguer anomaly in this area unless evoking an unrealistic volume of magma incompatible with the moderate Late Miocene and Plio-Pleistocene volcanism affecting the studied region. However, an ~ 100 -km-wide and 5 -km-thick mantle body replacing lower crustal material can produce an ~ 50 -mGal anomaly considering a density contrast $\Delta\rho_{mantle\ body}$ of 500 kg/m^3 . A thicker (~ 9 km thick) high-density body with a density contrast relative to the crust $\Delta\rho_{lower\ crustal\ high - density\ body}$ of 300 kg/m^3 can also produce an ~ 50 -mGal anomaly. Although a major trade-off between the geometry and the density contrast when modeling gravity signals, the origin of the gravimetric anomaly is likely induced by a strong positive mass anomaly at a lower crustal level.

5. Morphostructural Analogue Modeling

5.1. Hypotheses and Boundary Conditions

To test mechanically the tectono-magmatic scenario that emerged from the morphostructural and geophysical analysis, we performed a series of analogue models assuming that the main process that control the uplift and morphological evolution of the HP is related to viscous magmatic intrusions “pushing from below” the sedimentary cover. We also investigate, although qualitatively, the effects of structural and stratigraphic inheritance (basement faults, mechanical stratigraphy) on the subsurface tectonic pattern as well as the interactions between tectonics and surface processes on drainage network evolution. To simulate the inherited basement fault, basal mechanical discontinuities are used rather than weakened fracture zones, which are challenging to reproduce in analogue models.

5.2. Experimental Setup and Analogue Materials

The experimental apparatus (Figure 9) is made of $90 \times 90 \times 4$ -cm box equipped in its center with a vertical nozzle (inner diameter ~ 1 cm), which injects a pressurized viscous fluid at the base of the overlying analogue model. The intruded viscous material is confined by a 28 -cm radius elastic sheet (3 -mm-thick rubber sheet) at the model base to accommodate upward and lateral viscous flow. To simulate inherited basement faults, delimiting rigid and gently deformable upper crustal blocks outlining the surfaces identified in Figure 4, we used semirigid aluminum plates positioned at the base of the model, on the rubber sheet. These contiguous basal plates are not mechanically linked except in some taped boundaries to damp block differential movements, especially in the far field (Figure 9).

A rainfall apparatus (Figure 9c) is used to induce model erosion, sediment transport, and deposition during surface uplift (Graveleau & Dominguez, 2008). These surface processes give rise to the development of geomorphic patterns, such as drainage networks, interacting with the model surface deformation and allowing a detailed comparison with nature.

Conventional nature/model scaling rules (Ramberg, 1981) were used to find the appropriate analogue materials respecting the geometrical, kinematic, and dynamic similarity criteria. The experimental device and the Hyblean Plateau dimensions imposed a length ratio of $1\text{ cm} - 1\text{ km}$, and analogue materials versus Hyblean Plateau rheological properties constrained the density and stress ratios (table in Figure 9). According to boreholes and seismic data (Bianchi et al., 1987), the Meso-Cenozoic sedimentary cover of the Hyblean Plateau can be split in two main mechanical layers, a 5 – 6 -km-thick massive Triassic dolomite unit covered by a 2 – 3 -km-thick more layered calcareous unit. To reproduce this first-order rheological layering, we used a bilayered model with an ~ 1.5 -cm-thick upper layer made of a granular mixture composed of 50% glass microbeads (diameter 90 – $150\text{ }\mu\text{m}$), 25% of PVC, and 25% of silica powders (MatIV; $C_0 \sim 500\text{ Pa}$, $\phi \sim 40^\circ$;

Figure 9. Modeling apparatus. (a) Side view of the model frame. The analog material is lying on the rigid plates or directly above the elastic sheet. (b) Top view with the used plate geometry when considering structural inheritance and the position of the viscous injection (red dot) as well as the extension of the elastic sheet (dashed red circle). Thin black lines display the taped limits of the plates and thick black lines the dissociated edges. (c) Oblique view of the experimental apparatus in the lab before running the experiment. (d) Values and scaling ratios for the main parameters used in the analogue models. The natural values with an “asterisk” symbol are the one estimated considering the model parameter and the constrained scaling ratio. The accuracy of these values is discussed in the text.

Graveleau et al., 2011). This material is similar to the one used by Graveleau et al. (2011) to model experimentally tectonic-erosion-sedimentation interactions in active mountain foreland. The lower layer representing the massive, brittle, and resistant Triassic dolomites is made of an ~2.5-cm-thick, highly cohesive, silica powder ($C_0 \sim 1,300$ Pa, $\phi \sim 45^\circ$). All the granular materials are water-saturated (~20% of the final weight) and a low percentage of anthracite particles (~1%) is added to enhance syn-tectonic sedimentation layering and to densify the surface texture contrast to improve the accuracy of the model monitoring (strain field measurements and topography).

The viscous material injected to induce model topography uplift is made of a commercial coat with a viscosity $\sim 10^2$ – 10^3 Pa/s that scales to an $\sim 5 \times 10^{12}$ to 5×10^{13} -Pa/s natural viscous fluid, respecting the range of estimated viscosities (10^{11} – 10^{14} Pa/s) for deep magmatic intrusions (~7 km) beneath Mount Etna (e.g., Acocella et al., 2013). The intrusion rate in our experiments is set to 30 ± 10 cm³/min taking into account its viscosity and time-dependent rheology that respect the velocity ratio (table in Figure 9), determined using $\sim 6 \times 10^{-3}$ m/s for the model and 0.1 m/s in nature (Acocella et al., 2013).

To not exceed the pressure capacity of the viscous fluid injector, we slightly reduced to ~4 cm thick (~4 km) the cover representing the 7–8-km-thick Hyblean-Malta platform. The significant rheological contrast that is expected between the thick massive Triassic dolomites and the overlying stratified limestone ($C_0 \sim 80$ MPa for the limestone layer and $C_0 \sim 210$ MPa for the massive dolomites) is taken into account by the cohesion contrast between the MatIV and the silica powder layer. The simulation of inherited basement faults, by using aluminum basal plates, should be considered as an experimental forcing to promote the formation of individual tilted blocks during model deformation.

Figure 10. End-member model 1 without structural inheritance. (a) Local DEM at the end of the experiment from *PhotoScan* software (color scale represents the elevation in mm). The black dot and the red dashed circle are the injection center and the elastic sheet limit, respectively. (b) Oblique side view and (c) top view of the model slightly before the end of the model. (d) Structural interpretation of the model output with the mean orientation and mean slope of the tilted surfaces. A ring-fault pattern is visible with compressional structures outward and extensional faults inward. Some radial fractures are outlined, although they are difficult to pick because the drainage network reused them.

5.3. Modeling Procedure

Two cameras record the model evolution, allowing a 3-D qualitative and a quantitative monitoring (subpixel image correlation) of the model deformation. A third camera is used to measure model topography and reconstruct digital elevation models (DEM) using the *Photoscan* software.

The intrusion rate ranges between 20 and 40 cm³/min, which is sufficiently low to allow a quasi-static equilibrium of the uplifted surfaces under the lithostatic load of 600 ± 50 Pa imposed by the analogue model sedimentary cover. Two types of experiments have been performed, with and without aluminum basal plates to simulate with or without inherited basement faults.

5.4. Morphostructural Outcomes From Analogue Modeling

Two end-member models are presented hereafter, with or without tectonic inheritance, to highlight its impact on the morphology and the structure of the dome. The first experiment (Figure 10) was conducted without the aluminum basal plates (no inherited basement faults and upper crustal blocks). The first deformation stage show a gentle bumping of the surface associated to two normal fault systems, one above the intrusion center forming an ~2-cm-wide crestal graben and the other one on the footwall of a roughly orthogonal fault system ~3 cm from the intrusion center. This early stage of faulting is slightly influenced by desiccation fractures in the lower silica layer. The dome becomes progressively wider and a ring fault pattern appears with internal extension and external compression deformation pattern associated to the outward propagation of the intrusive material at depth. Radial fractures also formed during the domal uplift

Figure 11. End-member model 2 with structural inheritance. (a) Local DEM during the experiment from *PhotoScan* software (color scale represents the elevation in mm). The black dot and the red dashed circle are the injection center and the elastic sheet limit, respectively. The basal plate edges are overprinted (dashed black lines) on the shaded topography. (b) Oblique side view and (c) top view of the model during the experiment. Extensional faults (in red) define the five to six individualized block edges whose orientations are outlined by the T symbols associated to their mean orientation and mean slope.

(Figure 10d) imposing the drainage network geometry. In this end-member model, no structural blocks formed and the dipping of the model surface is perfectly radial (Figure 10d).

The second experiment integrates the aluminum basal plates at its base to simulate subcrustal tectonic blocks delimited by major inherited fault zones. The geometry of the plates (Figure 9b) is in accordance with the main structural features identified in the HP region (Figures 2 and 4f) such as the edges of the tilted topographic surfaces and the SW-NE and SE-NW trending fault systems. The position of the intrusion center is chosen accordingly to the maximum topography of the Hyblean region relative to the tilted surface geometries. This end-member model (Figure 11) shows that the extrados deformation concentrates at the edges of the basal plates defining ~3-cm-wide grabens with conjugated normal faults (~60° dipping) and defining individual tilted blocks. As expected, there is no significant internal deformation affecting the tilted, nearly planar, surfaces. We also note that the two blocks in the top left corner are structurally at lower elevation and began to rise, lately indicating a delayed intrusion in this sector which is not located directly above the intrusion center.

6. Discussion

The radially tilted blocks and multiple trending faults of the Hyblean Plateau suggest a localized punctual uplift. The 2-D flexural models show that such bulging is mechanically possible although the reconciliation in terms of dynamics of plate flexure is problematic. Indeed, given that the two flexural profiles A and B in Figure 5 are relatively close to each other (45 km), we do not expect the corresponding loads to be roughly 1 order of magnitude different (Figure 6). Hence, the diverging results from profiles A and B point out the

difficulty to link the observed topography of the HP to flexural processes alone. The lack of subcylindrical forebulge at the front of the orogenic prism as expected by flexural models (Turcotte & Schubert, 1982) also outlines the limited influence of flexural uplift in southeastern Sicily. Moreover, the early uplift initiation associated to volcanic events in the Late Miocene and its fixed position over time is inconsistent with the south-eastward migration of the SFTB whose deformation front was >100 km away during the Tortonian (Figure 3). Because the flexure is unlikely to produce the first-order high-amplitude (~1000 m) and relatively low-wavelength (>100 km) subcircular topographic anomaly of the Hyblean Plateau, the discussion mainly focused on the tectono-magmatic scenario that arise from the geological and geophysical data.

6.1. Morphology of the Hyblean Dome

The radially uplifted and tilted surfaces shaping the HP topography (Figure 4d) suggest that its subsurface morphology is controlled by deep-seated fault zones (Figure 4f) defining subcrustal blocks with low internal deformation. This structural pattern is most probably inherited from the long-term tectonic evolution affecting the Hyblean region since the Permo-Triassic rifting. Indeed, as proposed by Butler et al. (1992) and Ghisetti et al. (2009), the NW-SE trending Malta Escarpment (ME), and lately the NE-SW extensional faults affecting the foreland, had a significant control on the Late Miocene-Pliocene tectonic pattern of southeastern Sicily.

The Hyblean Plateau is also characterized by a relatively low local relief compared to central Sicily (Figure 4a). This can be explained by the nature of the outcropping Hyblean units, mostly Cenozoic carbonate layers gently deformed in its central part and by the recent emergence of the plateau. Morphometric analysis on the eastern sector have shown that the local relief decreases from WNW to ESE in accord with a Middle-Late Pleistocene ESE tilting of this region, leading to deeper valley entrenchment in the upstream portion of the Anapo River Valley (Bonforte et al., 2015).

The formation of a subcircular relief made of radially tilted blocks can be mechanically explained by a local source of material pushing from below since the Upper Miocene. Indeed, the partial correlation between late Miocene-Pleistocene volcanism and uplift phases points to the possibility that magmatic-driven processes may have significantly contributed to the uplift of the HP. This tectono-magmatic scenario, evaluated mechanically using analog modeling (Figures 10 and 11), imply also a major structural inheritance within the Hyblean crust, as the model requires strong discontinuities to localize the deformation at block edge. If the basement is homogenous and undamaged the dome flanks, show concave down profiles and the deformation pattern is dominated by ring fault systems. This fault pattern is not observed in the Hyblean Plateau. However, when the basement is prefractured (inherited faults) the topographic expression resulting from the punctual uplift defines radially tilted individual surfaces, similarly to what is observed in the southern and eastern sector of the HP (Figures 4d and 4f). The relatively wide grabens at the block limits in the model are the main tectonic features accommodating the extrados extension related to bulging of the subsurface. Such grabens can be compared to the Scordia Lentini Graben or the Florida Graben (Figure 2), although these structures tend to evolve into more diffuse deformation zones like the Comiso-Chiaramonte Fault Belt and Monterosso-Agnone Fault System toward the top of the HP.

Similar tectono-magmatic events involving deep subhorizontal intrusions and associated domal uplift have been already described. For instance, in northwestern Italy, Acocella (2000) identified a 35×50 -km-wide dome, approximatively 2.5 km high resulting from the emplacement of a laccolith intrusion at a depth of ~7 km. Other examples can be found in the literature, such as the ~20-km-wide uplifted zone induced by subhorizontal mafic sill intrusion in the Canary island region (Klügel et al., 2015). Mantle upwelling is also a potential source of uplift, already mentioned in the HP area (Catalano et al., 2010; Schmincke et al., 1997) and to the north, beneath Etna volcano (Hirn et al., 1997), as well as serpentine-bearing clayed diapirs (e.g., De Guidi et al., 2014; Giampiccolo et al., 2017; Manuella et al., 2012). Considering that we cannot decipher between the potential sources of uplift, our model mainly focused on surface deformation resulting from a magmatic-like intrusion. While this preliminary experimental approach do not attempt to give an accurate estimate of the intruded volume in nature, the suggested scenario implies however that the uplifted area should be roughly equivalent to the volume of the intruded material. Accordingly, considering the conical shape of the Hyblean Plateau that is ~1,000 m high with an ~30-km radius, the corresponding volume, ~950 km³, is not unrealistic since large volumes of lava flows have been estimated during the successive Plio-Pleistocene eruptions (>120 km³ for the Militello volcanic phase alone according to Behncke (2004)).

Moreover, giving the uplift rate related to the ~600-m Middle Pleistocene uplift phase, ~0.2–0.5 mm/year, the corresponding intrusion rate would be $\sim 10^{-4}$ km³/year, a value that is consistent with the magmatic activity of the nearby Etna volcano ($\sim 1.6 \times 10^{-4}$ km³/year in average since 330 ka according to Barreca et al. (2018)).

6.2. Geophysical Supports and Open Questions

1. Crustal magnetic signature

Numerous studies on southeastern Sicily already outlined that both gravity and magnetic data (AGIP, 1978, 1982) have different signatures in the Hyblean region and in the Pelagian foreland southward (e.g., Ben-Avraham & Grasso, 1990; Rota & Fichera, 1987; Yellin-Dror et al., 1997). The high magnetic intensity values >200 nT (Figure 1c) recognized over the HP (Bianchi et al., 1987; Rota & Fichera, 1987) was attributed to a dome-shaped magnetic substratum located at depth >6 km (Cassano, 1986).

The presence of mafic intrusive bodies evidenced in seismic lines crossing the Malta Escarpment (Catalano, Franchino, et al., 2000) and by magnetic anomalies in the Sicily Channel (Nameless Bank; Figure 1c; Lodolo et al., 2012) is consistent with mafic intrusions within the Hyblean crust. Indeed, the amplitude and shape of the HP magnetic anomalies (>200 nT) compare well with the one in the Sicily Channel modeled by Lodolo et al. (2012) and interpreted as related to an ~30-km-wide and ~1-km-thick intrusive mafic body located at 4–5-km depth in the Sicily channel. The apparent complexity of the magnetic field in southeastern Sicily (Figure 1c) could be explained as the result of the interferences between the dipole signatures of several adjacent magnetic intrusions inherited from the long-term Meso-Cenozoic magmatic history. Moreover, the basement magnetic signature is poorly known while its signal could be responsible for the main anomaly over the HP. Indeed, the magnetic signature cannot distinguish whether the basement has a basic affinity or if it results from extensive mafic intrusions (Rota & Fichera, 1987). An oceanic-type basement beneath southeastern Sicily is still favored by several authors (e.g., Giampiccolo et al., 2017; Manuella et al., 2012), leading to the interpretation of the magnetic anomalies as the effect of serpentinite seamounts at a depth of 8 to 12 km. Following this hypothesis, the xenoliths from the Hyblean volcanism are compatible with an oceanic affinity, similar to what is found in oceanic core complexes (Scribano et al., 2006) or ultraslow spreading oceanic lithosphere (Manuella et al., 2013). However, the sharp transition along the >150-km-long, 20–30-km-wide Malta Escarpment (ME) from shallow-water carbonates on the Hyblean-Malta promontory to a deep marine (>3000-m depth) environment in the Ionian domain is a typical feature of stretched continental margins (Dellong et al., 2018; Gutscher et al., 2016) that cannot be easily explained by oceanic lithospheric models. Finally, it is not excluded either that along the margin, local crust thinning may have lead mantle at shallower depths, and its partial serpentinization suggested by serpentine-bearing clayed diapirs (Barreca, 2014; Manuella et al., 2012) could also contribute to the magnetic signature of the Hyblean crust and hence reconcile the diverging hypothesis.

2. Gravity significance of subcrustal bodies

The strong negative Bouguer anomaly (<−100 mGal) to the west in the SFTB is widely accepted to result from the accretion of allochthonous sedimentary units (e.g., Ben-Avraham & Grasso, 1990; Catalano et al., 2013) over a thickness of 15–20 km as illustrated by the SIRIPRO seismic line crossing central Sicily (Catalano et al., 2013). However, even if sometimes evoked, the major positive Bouguer anomaly (>100 mGal) over southeastern Sicily has never been explained convincingly.

A positive Bouguer anomaly at the front of orogenic belts may result from the upward flexure of the foreland inducing underlying mantle rise (Karner & Watts, 1983; Moretti & Royden, 1988). In such a case, the maximum gravity anomaly associated to a crustal uplift of $\Delta z = 1000$ m is given by the Bouguer plate anomaly solution $\Delta g = 2\pi G \Delta z \Delta \rho$ (where $\Delta \rho$ is the crust-mantle density contrast and G is the gravitational constant; Turcotte & Schubert, 1982). Considering a crust-mantle density contrast $\Delta \rho$ of 500 kg/m³, a positive anomaly of ~20 mGal is estimated. This upper bound value is much lower than the one observed in the Hyblean region and shows, once again, that the flexure alone cannot explain the high positive Bouguer anomaly (>100 mGal). Moreover, if the gravity anomaly was related to foreland dynamics, it should run subparallel to the orogenic front like in the Pelagian foreland (Moretti & Royden, 1988). By contrast, voluminous crustal mafic intrusions previously evoked for the Hyblean region (Atzori et al., 1999; Schmincke et al., 1997; Yellin-Dror et al., 1997) and evidenced nearby in the Sicily Channel (Civile et al., 2008; Lodolo et al., 2012) should

also be associated to positive gravity anomalies as observed in the SW Iberia margin (Neres et al., 2018) or in the Adriatic region (Mancinelli et al., 2015).

Crustal thickness heterogeneities are also common along stretched continental margin and their gravimetric signature is highly controlled by the Moho geometry. For instance, a strong and narrow positive Bouguer anomaly >140 mGal overlap the Malta Escarpment (ME). This NNW-SSE elongated anomaly is certainly associated to the previously discussed local stretching of the margin (Dellong et al., 2018), combined with a decrease of the gravity anomaly to the east in the accretionary wedge of the Calabrian Arc. However, the Bouguer gravity high over the Hyblean Plateau extends further west and might not belong to the same context.

Different deep crustal scenarios can be evoked to explain the main positive Bouguer gravity anomaly. For instance, an ~ 100 -km-wide and 5-km-thick mantle body lying at the base of the crust (in blue; Figures 8d and 8e) with a density contrast $\Delta\rho_{\text{mantle body}}$ of 500 kg/m^3 producing an ~ 50 -mGal density anomaly. This scenario is equivalent to a Moho uprising beneath the Hyblean Plateau (~ 25 -km depth) compare to the Hyblean-Malta Platform to the south (~ 30 -km depth). Another scenario (in green; Figures 8d and 8e) could involve the emplacement of a thicker (~ 9 km thick) high-density body in the lower crust with a density contrast $\Delta\rho_{\text{lower crustal high - density body}}$ of 300 kg/m^3 representing mafic underplated/intraplating material or partly serpentinized (~ 35 to 40%) mantle as suggested by Manuella et al. (2012) and Giampiccolo et al. (2017).

There is obviously no uniqueness solution for this problem but we can still raise some major conclusions from these simulations. On one hand, the gravity anomalies associated to flexural bulging or reasonable mafic intrusions are unlikely to cause the >50 -mGal anomaly over the Hyblean Plateau (but magmatic bodies are likely needed to explain the strong magnetic anomalies). On the other hand, these two processes might have a gravity signature hidden by the large-amplitude signal to which they could contribute. Finally, a density anomaly located at the base of the crust might be the most likely origin of the positive Bouguer gravity anomaly. Different scenarios are proposed here and should be the subject of future investigations: (1) an uprised Moho beneath the Hyblean Plateau inherited from crustal thinning during the early phases of Mesozoic rifting; (2) similarly to the preceding scenario, a thicker serpentinized mantle body, also inherited from the Mesozoic history; and (3) a long-term magmatic history of the Hyblean region, at least since the Late Cretaceous, leading to voluminous mafic underplating/intraplating beneath the Hyblean Plateau. The latter scenario has the advantage to link the gravity anomaly, the magnetic anomaly, and the subsurface Late Cenozoic uplift with a common long-term process while the others are in favor of independent processes occurring coincidentally at different times in the same area and with the same characteristic length scale.

3. Load support

A clear positive residual isostatic gravity anomaly >80 mGal over the HP is visible in Figure 7c, and correlates well with the unexpected strong positive Bouguer anomaly (>100 mGal) centered on the positive relief. The remaining question concerns the mechanism leading to a maintained uncompensated mass anomaly. To support uncompensated crustal mass excess two conditions can be complete: (1) the rigidity of the lithosphere is sufficiently high to support the load by elastic stresses and (2) an upward force dynamically push from below. Although our flexural analysis (Figure 8) does not converge toward a single solution, we found that the elastic thickness over the Hyblean region could be as low as 5 km. However, it has been shown using the gravity-topography transfer function relationship (or “isostatic response function”), that for wavelength <100 km, an elastic thickness >5 km is sufficient to support an uncompensated load (Watts, 2001). If the topography is maintained by lithospheric elastic support in the studied region, the lower bound of the Hyblean lithosphere rigidity could be constrained by a >5 -km elastic thickness.

Nevertheless, a partial dynamic support of this region adjacent to the Calabrian Arc could also be expected. Effectively, the Mediterranean Sea is well known for its microscale upper mantle dynamics associated to slab retreat (e.g., Faccenna et al., 2001, 2005; Schellart, 2010; Strak & Schellart, 2014). Mantle upwelling has been evoked to account for the dynamics of deformation along the Malta Escarpment and tomography highlights low V_p (Scarfi et al., 2018) that could reflect such mantle flow.

The geoid-to-topography ratio has been used to infer the depth of the compensating mass over oceanic swells (Cadio et al., 2012; Crough, 1978; Haxby & Turcotte, 1978) and decipher whether the compensation mechanism is supported by the crust itself, deeper in the lithospheric mantle by thermal rejuvenation or within the asthenosphere by dynamic uplift. While the geoid-to-topography ratio is usually discussed with

compensated scenarios, and gives an estimate of the compensating mass depth, its use with uncompensated scenarios (as the Hyblean Plateau) can lead to a maximum “compensation” depth. Hence, this approach is used to decipher whether or not the link between the geoid and the topography is due to sublithospheric or crustal mechanisms.

As shown in Figure 1b, the geoid anomaly mimics the HP topography motivating the use of this ratio to investigate whether or not the Hyblean region is dynamically supported. We first determined the spectral coherence between the geoid and the topography at various wavelengths to select the relevant length scale on which the geoid-to-topography ratio (also referred as the “admittance”) can be interpreted. Using the wavelet method of Cadio et al. (2012), high coherence values are found (>0.7) at wavelengths ~ 100 km, which corresponds to the characteristic length scale of the Hyblean Plateau topography, and relatively low admittance values ~ 1.6 m/km are associated to that wavelength range (Appendix B). Using the Pratt compensation model of Haxby and Turcotte (1978), a geoid-to-topography ratio of ~ 1.6 m/km is consistent with a compensation depth of ~ 28 km using a reference column density of 2700 kg/m^3 . Based on these calculations, the support mechanism is therefore unlikely to be driven by sublithospheric/asthenospheric upwelling. The low admittance value of ~ 1.6 m/km indicates that the topography is supported by a subcrustal source consistently with previous low, <2 m/km, geoid-to-topography ratios (Crough, 1978). Moreover, using the equations from McKenzie (2010) which connects thermal-induced convection to dynamic topography for various flexural rigidities, we note that, at the characteristic wavelength of the Hyblean Plateau (~ 100 km) and for elastic thicknesses ~ 5 km, less than 100 m of dynamic uplift can be produced. Hence, the support by elastic strength of the lithosphere appears as the most likely scenario, although minor dynamic support contribution is also suggested by geodynamical modeling (Schellart, 2010; Strak & Schellart, 2014).

6.3. Tectono-Magmatic Model of the Hyblean Plateau

Multiple trending faults and radially tilted blocks characterize the dome-shaped Hyblean Plateau and are likely due to a local uplift of the Hyblean platform in relation to a multiphase magmatic history initiated during the Late Miocene. Our tectono-magmatic model (Figure 12) outlines a major uplift component associated to magmatic (and magnetic) sheet intrusions beneath the prefractured Meso-Cenozoic carbonatitic sedimentary cover. The Late Miocene and Plio-Pleistocene volcanism correlate with the HP uplift events during these periods, suggesting a magmatic-controlled uplift, although a minor component from flexural bulging during the Quaternary is not excluded. Compressive tectonic features have been found on the Plateau edges (Romagnoli et al., 2015) and might result from block bending external compression as suggested by analogue models. However, the relatively minor strike-slip kinematics found over the Hyblean Plateau should be linked to regional tectonic stresses in response to the southeastward migration of the Peloritani-Calabria terranes.

The origin of the anorogenic magmatism in eastern Sicily is possibly related to mantle flow as suggested by low V_p values beneath the Hyblean Plateau (Scarfì et al., 2018) and geodynamic models of the central Mediterranean (e.g., Faccenna et al., 2001, 2005; Schellart, 2010). The cessation of the volcanic activity during the Pleistocene in southeastern Sicily could be due to the progressive change of tectonic regime from extension to compression as well as the shift of magma pathways toward the Etna volcanic system (Neri et al., 2018). The Etna central-type volcanism likely diverged from the Hyblean fissural volcanism because of its interaction with major crustal to lithospheric fault zones.

Finally, the high-density body in the Hyblean crust also sign possible voluminous mafic underplating associated to the long-term magmatic evolution of southeastern Sicily. The last episode started in the Late Miocene and is likely responsible of the actual topographic anomaly. Although the Hyblean lithosphere is apparently rigid enough to support the uncompensated density heterogeneity, minor dynamical support is expected, particularly during the major volcanic phases of the Late Cenozoic.

7. Conclusion

The morphology analysis of southeastern Sicily revealed a first-order dome-shape pattern characterizing the Hyblean Plateau. Its subsurface deformation outlines a subdivision into radially tilted blocks with low internal deformation showing preserved surfaces. The edges of this fragmented topography might be related to deep-seated fracture zones inherited from the Meso-Cenozoic tectonic evolution of the Central Mediterranean Sea and reactivated since the Late Miocene-Quaternary times. The lack of dipping blocks

Figure 12. Schematic tectono-magmatic model of southeastern Sicily during the Quaternary period (not to scale; vertical exaggeration $\times 4$). Magmatic intrusions mainly end beneath the thick Triassic dolomite unit where they laterally spread. In response to mafic sheet intrusions the Meso-Cenozoic sedimentary cover is progressively uplifted and reactivates subcrustal weak zones inherited from the long-term tectonic history of the Hyblean-Malta platform. Inverse faulting is not clearly identified in the field, although some folding has been found at the southeastern end of the Tellaro valley (Figure 2; Romagnoli et al., 2015). Compressive structures at the “dome” edges can be related to blocks bending during uplift and potentially to NW oriented compressive regional tectonic stress (σ_1) associated to the convergence. The late arrival of the Sicilian Fold and Thrust Belt (SFTB) downbended the northwestern part of the Hyblean-Malta platform and HP, possibly reactivating slightly inherited faults and contributing to a minor uplift of the HP.

toward the Sicilian Fold and Thrust Belt (SFTB), the noncylindrical shape of this topographic high, and the need of significant load variations in space to account for the observed post-Miocene deformation makes the flexural hypothesis insufficient to explain the formation of the Hyblean “Dome.” Although some flexural uplift might have occurred during the Plio-Quaternary, tectono-magmatic processes involving intracrustal mafic sill intrusions are more likely to produce (1) the domal morphology of the Hyblean Plateau, (2) its early uplift during the Late Miocene while the SFTB deformation front was located far away (>100 km) from the HP, and (3) the volcanic and magnetic signature characterizing the HP region. Moreover, the complex interplays between magmatic intrusions and inherited subcrustal weak zones can explain the multiple trends of faults all over the area better than for forebulge extrados deformation. Nevertheless, the recent strike-slip kinematics is more likely related to a regional compressive setting within the NNW oriented collision (Cultrera et al., 2015).

Neither mafic sheet intrusions, nor flexural bulging, can reasonably produce the >50 -mGal residual positive Bouguer anomaly (after removing the Malta Escarpment Bouguer anomaly trend) over the Hyblean Plateau. Interestingly, the gravity anomalies superimpose to the anomalous Hyblean high, making the main gravimetric signal anticorrelated to the topography. Although upper crustal mafic intrusions may contribute to a positive mass anomaly, a deeper and thicker high-density body is needed to account for the major positive Bouguer anomaly over the Hyblean Plateau. Two types of scenario leading to the observed gravity high can be suggested, on one side, a Moho uprising of 5 to 10 km depending on the degree of serpentinization (from 0 to 40%). This scenario involves two different processes: crustal thinning during the Early Mesozoic rifting, and then upper crustal doming in the Late Cenozoic due to magma intrusions. On the other side, magmatic underplating/intraplating since, at least, the Late Cretaceous could also produce the main positive gravity anomaly. In such a case, the Late Miocene and Plio-Pleistocene subsurface uplift would be special events, with mafic intrusions stacking mainly beneath the Meso-Cenozoic sedimentary cover. A deeper understanding of the Hyblean lower crust structure and composition is therefore needed and should be the focal point of further works.

Finally, the nonisostatic state of this area is probably maintained by the rigidity of the lithosphere rather than dynamically supported. Indeed, the relatively low length scale (~ 100 km) of the gravity and topographic anomalies makes the lithospheric elastic stresses sufficient enough to support such crustal-scale anomalies, even considering a low lithospheric rigidity ($D \sim 10^{21}$ N/m, $T_e \sim 5$ km).

Appendix A: Base of Pliocene Data and Modeling

Table A1
Wells and Geological Data Used to Constrain the Base of the Pliocene

	Longitude	Latitude	Elevation (m) relative to sea level	Uncertainties (m)
Well's name				
Simento_001	14,995573	37,426882	−839	NaN
Dittaino_001	14,910914	37,415967	−1056	NaN
S,Demetrio_001	15,030410	37,360085	NaN	NaN
Ramacca_001	14,696809	37,448113	−991	NaN
Naftia_001	14,678612	37,327315	NaN	NaN
Naftia_002	14,707500	37,285788	122	NaN
Sigona_grande_001	14,763195	37,322313	−331	NaN
Margi_001	14,672361	37,313565	−362	NaN
Mineo_001	14,644862	37,280928	−516	NaN
Mineo_002	14,607501	37,279401	−1000	NaN
Piazza_armerina_001	14,529862	37,270930	−1686	NaN
Francofonte_001	14,714583	37,253428	NaN	NaN
Francofonte_002	14,714304	37,226762	NaN	NaN
Caltagirone_001	14,417919	37,218435	−2323	NaN
Ursitto_001	14,370976	37,180104	−2554	NaN
Ponte_olivo_001	14,348028	37,170105	−2087	NaN
Troitta_001	14,533416	37,161629	−342	NaN
Vizzini_001	14,710691	37,145932	NaN	NaN
Settefarine_001	14,255422	37,122052	−2026	NaN
Contrada_ulmo_001	14,434028	37,111494	−991	NaN
Buccheri_001	14,855685	37,117041	NaN	NaN
Giarratana_1	14,794054	37,047844	NaN	NaN
Monterosso_001	14,748188	37,099544	NaN	NaN
Biddine_001	14,506875	37,059678	−117	NaN
Rabbito_001	14,171812	37,101222	−2745	NaN
Acate_001DIR	14,349166	36,993445	−844	NaN
Acate_002	14,337361	37,002889	−822	NaN
Acate_003	14,316528	37,020111	−1008	NaN
Valle_torta_001DIR	14,342918	37,034832	−953	NaN
Vittoria_ovest_001	14,493746	36,945943	NaN	NaN
Vittoria_003	14,459997	37,005665	−212	NaN
Vittoria_004	14,542218	37,017051	60	NaN
Palazzolo_001	14,936234	37,029543	NaN	NaN
Rigolizia_001	14,932344	37,017599	NaN	NaN
Siracusa_001	15,099006	37,001485	NaN	NaN
Avola_001	15,064839	36,971154	NaN	NaN
Cassibile_001	15,186782	36,980928	NaN	NaN
Maddalena_001	15,312614	37,015370	−623	NaN
Capo_negro_001	15,174214	36,942450	NaN	NaN
S,croce_camerina_001	14,513466	36,859556	NaN	NaN
S,croce_camerina_002	14,509994	36,853862	17	NaN
Scicli_001	14,748458	36,774554	NaN	NaN
Scicli_002	14,707070	36,782610	NaN	NaN
Bimmisca_001	15,079557	36,809549	−71	NaN
Cozzo_scalia_001	15,021390	36,775188	NaN	NaN
Carrubo_001	14,963476	36,767448	NaN	NaN
Marzamemi_001	15,114832	36,743576	−142	NaN
Pachino_004	15,058559	36,698143	−45	NaN

Table A1
(continued)

	Longitude	Latitude	Elevation (m) relative to sea level	Uncertainties (m)
Rosa_001	15,158167	36,860656	−140	NaN
Giada_001	14,246653	36,902167	−892	NaN
Plinio_Sud_001	14,270569	36,893570	−833	NaN
Pancrazio_sud_001	14,361510	36,827650	−203	NaN
Pellicano_ovest_001	14,317380	36,791121	−515	NaN
Delfino_001	14,382126	36,770197	−267	NaN
Merluzzo_mare_001	14,463852	36,656567	−173	NaN
Mila_001	14,506964	36,738722	NaN	NaN
Archimedes_001	14,640037	36,635120	−395	NaN
Lucata_001_x	14,665678	36,633439	NaN	NaN
Polpo_001	15,265810	36,623398	NaN	NaN
Vera_001	15,048726	36,566128	−218	NaN
Spada_mare_001	14,827143	36,583578	−208	NaN
Gabbiano_001	14,750757	36,523079	−346	NaN
Eva_001	14,673462	36,446761	−640	NaN
Aretusa_001	14,949141	36,494428	−237	NaN
Ulisse_001	14,833456	36,380686	−325	NaN
Pilade_est_001	15,148035	36,344615	NaN	NaN
Cernia_001	15,134243	36,309438	NaN	NaN
Oreste_001	15,301574	36,321672	NaN	NaN
Spigola_mare_001	15,061331	36,270570	−161	NaN
Patty_est_001	13,971124	36,995442	−2431	NaN
Genziana_001_dir_a	13,847052	36,998283	−2317	NaN
Pm (from geological map)				
	14,508700	36,921660	109	28
	14,694162	37,193091	541	17
	14,989633	36,885973	218	94
	14,980575	36,892963	215	71
	14,721412	37,213044	582	57
	14,734978	37,245206	479	36
	14,733163	37,262941	405	65
	14,736975	37,282817	340	79
	14,743418	37,293808	244	54
	14,474104	36,878790	22	39
	14,503575	36,900282	88	47
	14,515821	36,910791	46	14
	14,514427	36,925070	116	23
	14,527147	36,960237	175	7
	14,703155	36,785905	190	30
	14,698864	36,776377	165	19
	15,066650	36,678473	21	10
	15,067887	36,699564	38	15
	15,063263	36,712593	56	8
	14,977391	36,754393	38	17
	14,982145	36,757196	42	18
	15,012670	36,774852	54	13
	15,014851	36,773643	48	13
	15,017640	36,768240	34	11
	15,052085	36,776913	33	14
	15,057385	36,776972	30	6
	15,066574	36,804284	41	12
	15,073111	36,816041	37	8
	15,060132	36,821190	21	13
	15,075709	36,826710	23	15
	15,047084	36,845486	20	13
	15,024539	36,867201	129	22
	15,018153	36,859384	81	25
	15,035711	36,869497	127	19

Table A1
(continued)

	Longitude	Latitude	Elevation (m) relative to sea level	Uncertainties (m)
	15,023645	36,883023	158	32
	15,027427	36,894287	169	16
	15,037597	36,894689	153	32
	15,067983	36,902485	174	19
	15,060294	36,900981	136	43
	15,081357	36,897479	117	43
	15,198498	36,956049	14	3
	15,197946	36,961736	23	8
	14,686464	37,280878	230	52
	14,687232	37,132966	392	56
	14,680455	37,138176	338	33
	14,648604	37,164806	444	19
	14,677103	37,050825	385	12
	14,687713	37,052482	407	21
	14,701616	37,063316	467	20
	14,684294	36,812979	262	15
	14,453528	36,841871	2	17
	14,441859	36,889348	44	18
	15,205420	36,962944	11	9
Pb (from geological map)				
	14,824315	37,076674	796	47
	14,797369	37,089624	797	57
	14,830154	37,110932	871	87
	14,761856	37,152799	521	36
	14,856977	37,137529	715	67
	14,912533	37,139540	783	23
	14,748367	37,108888	608	42
	14,748862	37,124657	560	49
	14,752068	37,157158	487	26
	14,915965	37,179347	380	63

Figure A1. Comparison between measured and predicted elevation (in m) for the base of the Pliocene units. Circles are well data and diamond and triangles are contacts picked from geological maps (see Figure 5 for details).

Appendix B: Geoid-Topography Coherence and Admittance

Figure B1. (a) Local wavelet topography, (b) geoid, and the associated (c) coherence and (d) admittance from East of the Hyblean Plateau to the West (y axis; latitude = 36.8°N) at each analyzing spatial scale (x axis; from 50 to 500 km). Wavelet method calculation from Cadio et al. (2012).

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References

- Acocella, V. (2000). Space accommodation by roof lifting during pluton emplacement at Amiata (Italy). *Terra Nova*, 12(4), 149–155. <https://doi.org/10.1046/j.1365-3121.2000.00286.x>
- Acocella, V., Neri, M., & Norini, G. (2013). An overview of experimental models to understand a complex volcanic instability: Application to Mount Etna, Italy. *Journal of Volcanology and Geothermal Research*, 251, 98–111. <https://doi.org/10.1016/j.jvolgeores.2012.06.003>
- AGIP (1978). Carta gravimetrica della Sicilia. Anomalie di Bouguer. Map scale 1:500.000. *Bollettino del Servizio Geologico d'Italia*, 99.
- AGIP (1982). Carta magnetica d'Italia. Anomalie del campo magnetico residuo. Italy: San Donato Milanese.
- Agostinetti, N. P., & Amato, A. (2009). Moho depth and Vp/Vs ratio in peninsular Italy from teleseismic receiver functions. *Journal of Geophysical Research*, 114, B06303. <https://doi.org/10.1029/2008JB005899>
- Atzori, P., Mazzoleni, P., Punturo, R., & Scribano, V. (1999). Garnet-spinel-pyroxenite xenoliths from Hyblean Plateau (south-eastern Sicily, Italy). *Mineralogy and Petrology*, 66(4), 215–226. <https://doi.org/10.1007/BF01164494>
- Barberi, F., Gasparini, P., Innocenti, F., & Villari, L. (1973). Volcanism of the southern Tyrrhenian Sea and its geodynamic implications. *Journal of Geo-Physical Research*, 78(23), 5221–5232. <https://doi.org/10.1029/JB078i023p05221>
- Barreca, G. (2014). Geological and geophysical evidences for mud diapirism in southeastern Sicily (Italy) and geodynamic implications. *Journal of Geodynamics*, 82, 168–177. <https://doi.org/10.1016/j.jog.2014.02.003>
- Beccaluva, L., Siena, F., Coltorti, M., Di Grande, A., Lo Giudice, A., Macciotta, G., et al. (1998). Nephelinitic to tholeiitic magma generation in a transtensional tectonic setting: an integrated model for the Iblean volcanism, Sicily. *Journal of Petrology*, 39(9), 1547–1576. <https://doi.org/10.1093/ptro/39.9.1547>
- Behncke, B. (2004). Late Pliocene volcanic island growth and flood basalt-like lava emplacement in the Hyblean Mountains (SE Sicily). *Journal of Geophysical Research*, 109, B09201. <https://doi.org/10.1029/2003JB002937>
- Ben Avraham, Z., Lyakhovsky, V., & Grasso, M. (1995). Simulation of collision zone segmentation in the central Mediterranean. *Tectonophysics*, 243(1–2), 57–68. [https://doi.org/10.1016/0040-1951\(94\)00191-B](https://doi.org/10.1016/0040-1951(94)00191-B)
- Ben-Avraham, Z., & Grasso, M. (1990). Collisional zone segmentation in Sicily and surrounding areas in the Central Mediterranean. *Annales Tectonicae*, 4(2), 131–139.
- Ben-Avraham, Z., & Grasso, M. (1991). Crustal structure variations and transcurrent faulting at the eastern and western margins of the eastern Mediterranean. *Tectonophysics*, 196(3–4), 269–277. [https://doi.org/10.1016/0040-1951\(91\)90326-N](https://doi.org/10.1016/0040-1951(91)90326-N)
- Bianca, M., Monaco, C., Tortorici, L., & Cernobori, L. (1999). Quaternary normal faulting in southeastern Sicily (Italy): a seismic source for the 1693 large earthquake. *Geophysical Journal International*, 139(2), 370–394. <https://doi.org/10.1046/j.1365-246x.1999.00942.x>
- Bianchi, F., Carbone, S., Grasso, M., Invernizzi, G., Lentini, F., Longaretti, G., et al. (1987). Sicilia orientale: profilo geologico nebrodi-iblei. *Meteorological Society Geology Italian*, 38, 429–458.
- Billi, A., Porreca, M., Faccenna, C., & Mattei, M. (2006). Magnetic and structural constraints for the noncylindrical evolution of a continental forebulge (Hyblea, Italy). *Tectonics*, 25, TC3011. <https://doi.org/10.1029/2005TC001800>
- Bonforte, A., Catalano, S., Maniscalco, R., Pavano, F., Romagnoli, G., Sturiale, G., & Tortorici, G. (2015). Geological and geodetic constraints on the active deformation along the northern margin of the Hyblean Plateau (SE Sicily). *Tectonophysics*, 640–641, 80–89. <https://doi.org/10.1016/j.tecto.2014.11.024>
- Branca, S., Coltelli, M., De Beni, E., & Wijbrans, J. (2008). Geological evolution of Mount Etna volcano (Italy) from earliest products until the first central volcanism (between 500 and 100 ka ago) inferred from geochronological and stratigraphic data. *International Journal of Earth Sciences*, 97(1), 135–152. <https://doi.org/10.1007/s00531-006-0152-0>
- Butler, R., Grasso, M., & La Manna, F. (1992). Origin and deformation of the Neogene–recent maghrebian foredeep at the Gela nappe, SE Sicily. *Journal of the Geological Society*, 149(4), 547–556. <https://doi.org/10.1144/gsjgs.149.4.0547>
- Butler, R. W., Lickorish, W. H., Grasso, M., Pedley, H. M., & Ramberti, L. (1995). Tectonics and sequence stratigraphy in messinian basins, Sicily: constraints on the initiation and termination of the Mediterranean salinity crisis. *Geological Society of America Bulletin*, 107(4), 425–439. [https://doi.org/10.1130/0016-7606\(1995\)107<0425:TASSIM>2.3.CO;2](https://doi.org/10.1130/0016-7606(1995)107<0425:TASSIM>2.3.CO;2)
- Cadio, C., Ballmer, M. D., Panet, I., Diamant, M., & Ribe, N. (2012). New constraints on the origin of the Hawaiian swell from wavelet analysis of the geoid to topography ratio. *Earth and Planetary Science Letters*, 359–360, 40–54. <https://doi.org/10.1016/j.epsl.2012.10.006>
- Carbone, S., Grasso, M., Lentini, F., & Pedley, H. (1987). The distribution and palaeoenvironment of Early Miocene phosphorites of Southeast Sicily and their relationships with the Maltese phosphorites. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 58(1–2), 35–53. [https://doi.org/10.1016/0031-0182\(87\)90004-6](https://doi.org/10.1016/0031-0182(87)90004-6)
- Cassano, E. (1986). *Pianura Padana: interpretazione integrata di dati geofisici e geologici*. Roma: Agip.
- Castro, J. M., Cordonnier, B., Schipper, C. I., Tuffen, H., Baumann, T. S., & Feisel, Y. (2016). Rapid laccolith intrusion driven by explosive volcanic eruption. *Nature Communications*, 7, 13585. <https://doi.org/10.1038/ncomms13585>
- Catalano, R., Franchino, A., Merlini, S., & Sulli, A. (2000). A crustal section from the Eastern Algerian basin to the Ionian ocean (Central Mediterranean). *Meteorological Society Geology Italian*, 55, 71–85.
- Catalano, R., Valenti, V., Albanese, C., Accaino, F., Sulli, A., Tinivella, U., et al. (2013). Sicily's fold-thrust belt and slab roll-back: the SI.RI. PRO. seismic crustal transect. *Journal of the Geological Society*, 170(3), 451–464. <https://doi.org/10.1144/jgs2012-099>
- Catalano, S., Romagnoli, G., & Tortorici, G. (2010). Kinematics and dynamics of the Late Quaternary rift-flank deformation in the Hyblean Plateau (SE Sicily). *Tectonophysics*, 486(1–4), 1–14. <https://doi.org/10.1016/j.tecto.2010.01.013>
- Chen, Y., Xu, Y., Xu, T., Si, S., Liang, X., Tian, X., et al. (2015). Magmatic underplating and crustal growth in the Emeishan Large Igneous Province, SW China, revealed by a passive seismic experiment. *Earth and Planetary Science Letters*, 432, 103–114. <https://doi.org/10.1016/j.epsl.2015.09.048>
- Civile, D., Lodolo, E., Tortorici, L., Lanzafame, G., & Brancolini, G. (2008). Relationships between magmatism and tectonics in a continental rift: The Pantelleria Island region (Sicily Channel, Italy). *Marine Geology*, 251(1–2), 32–46. <https://doi.org/10.1016/j.margeo.2008.01.009>
- Cogan, J., Rigo, L., Grasso, M., & Lerche, I. (1989). Flexural tectonics of southeastern Sicily. *Journal of Geodynamics*, 11(3), 189–241. [https://doi.org/10.1016/0264-3707\(89\)90007-0](https://doi.org/10.1016/0264-3707(89)90007-0)
- Cox, K. G. (1993). Continental Magmatic Underplating. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 342(1663), 155–166. <https://doi.org/10.1098/rsta.1993.0011>
- Crough, S. T. (1978). Thermal origin of mid-plate hot-spot swells. *Geophysical Journal of the Royal Astronomical Society*, 55(2), 451–469. <https://doi.org/10.1111/j.1365-246X.1978.tb04282.x>
- Cultrera, F., Barreca, G., Scarfi, L., & Monaco, C. (2015). Fault reactivation by stress pattern reorganization in the Hyblean foreland domain of SE Sicily (Italy) and seismotectonic implications. *Tectonophysics*, 661, 215–228. <https://doi.org/10.1016/j.tecto.2015.08.043>

- De Guidi, G., Imposa, S., Scudero, S., & Palano, M. (2014). New evidence for Late Quaternary deformation of the substratum of Mt. Etna volcano (Sicily, Italy): Clues indicate active crustal doming. *Bulletin of Volcanology*, 76(5), 816. <https://doi.org/10.1007/s00445-014-0816-8>
- Dellong, D., Klingelhoefer, F., Kopp, H., Graindorge, D., Margheriti, L., Moretti, M., et al. (2018). Crustal structure of the Ionian basin and Eastern Sicily margin: Results from a wide-angle seismic survey. *Journal of Geophysical Research: Solid Earth*, 123, 2090–2114. <https://doi.org/10.1002/2017JB015312>
- Di Martire, D., Ascione, A., Calcaterra, D., Pappalardo, G., & Mazzoli, S. (2015). Quaternary deformation in SE Sicily: Insights into the life and cycles of forebulge fault systems. *Lithosphere*, 7(5), 519–534. <https://doi.org/10.1130/L453.1>
- Faccenna, C., Becker, T. W., Lucente, F. P., Jolivet, L., & Rossetti, F. (2001). History of subduction and back arc extension in the central Mediterranean. *Geophysical Journal International*, 145(3), 809–820. <https://doi.org/10.1046/j.0956-540x.2001.01435.x>
- Faccenna, C., Civetta, L., D'Antonio, M., Funicello, F., Margheriti, L., & Piromallo, C. (2005). Constraints on mantle circulation around the deforming Calabrian slab. *Geophysical Research Letters*, 32, L06311. <https://doi.org/10.1029/2004GL021874>
- Fuller, J., Fernandez, M., & Zeyen, H. (2008). FA2BOUG a fortran 90 code to compute bouguer gravity anomalies from gridded free-air anomalies: Application to the Atlantic-Mediterranean transition zone. *Computers & Geosciences*, 34(12), 1665–1681. <https://doi.org/10.1016/j.cageo.2008.02.018>
- Gardiner, W., Grasso, M., & Sedgeley, D. (1995). Plio-Pleistocene fault movement as evidence for mega-block kinematics within the Hyblean-Malta Plateau, central Mediterranean. *Journal of Geodynamics*, 19(1), 35–51. [https://doi.org/10.1016/0264-3707\(94\)00006-9](https://doi.org/10.1016/0264-3707(94)00006-9)
- Ghisetti, F., & Vezzani, L. (1980). The structural features of the Iblean Plateau and of the mount Judica area (Southeastern Sicily); a microtectonic contribution to the deformational history of the calabrian arc. *Bollettino della Società Geologica Italiana*, 99(1-2), 57–102.
- Ghisetti, F. C., Gorman, A. R., Grasso, M., & Vezzani, L. (2009). Imprint of foreland structure on the deformation of a thrust sheet: The Plio-Pleistocene Gela nappe (southern Sicily, Italy). *Tectonics*, 28, TC4015. <https://doi.org/10.1029/2008TC002385>
- Giampiccolo, E., Brancato, A., Manuella, F. C., Carbone, S., Gresta, S., & Scribano, V. (2017). New evidence for the serpentinization of the Palaeozoic basement of southeastern Sicily from joint 3-D seismic velocity and attenuation tomography. *Geophysical Journal International*, 211(3), 1375–1395. <https://doi.org/10.1093/gji/ggx369>
- Grasso, M. (1997). *Carta geologica del settore centro-meridionale dell'Altopiano Ibleo (provincia di Ragusa, Sicilia sudorientale), scale 1: 50.000*. SELCA, Florence: Università di Catania, Istituto di Geologia e Geofisica.
- Grasso, M., & Pedley, H. M. (1990). Neogene and Quaternary Sedimentation Patterns in the North Western Hyblean Plateau. *Rivista Italiana di Paleontologia e Stratigrafia*, 96, 219–240.
- Grasso, M. t., & Lentini, F. (1982). Sedimentary and tectonic evolution of the eastern Hyblean Plateau (southeastern sicily) during Late Cretaceous to Quaternary time. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 39(3-4), 261–280. [https://doi.org/10.1016/0031-0182\(82\)90025-6](https://doi.org/10.1016/0031-0182(82)90025-6)
- Graveleau, F., & Dominguez, S. (2008). Analogue modelling of the interaction between tectonics, erosion and sedimentation in foreland thrust belts. *Comptes Rendus Geoscience*, 340(5), 324–333. <https://doi.org/10.1016/j.crte.2008.01.005>
- Graveleau, F., Hurtrez, J.-E., Dominguez, S., & Malavieille, J. (2011). A new experimental material for modeling relief dynamics and interactions between tectonics and surface processes. *Tectonophysics*, 513(1-4), 68–87. <https://doi.org/10.1016/j.tecto.2011.09.029>
- Gutscher, M. A., Dominguez, S., De Lepinay, B. M., Pinheiro, L., Gallais, F., Babonneau, N., et al. (2016). Tectonic expression of an active slab tear from high-resolution seismic and bathymetric data offshore Sicily (Ionian Sea). *Tectonics*, 35, 39–54. <https://doi.org/10.1002/2015TC003898>
- Gutscher, M. A., Kopp, H., Krastel, S., Bohrmann, G., Garlan, T., Zaragosi, S., et al. (2017). Active tectonics of the Calabrian subduction revealed by new multi-beam bathymetric data and high-resolution seismic profiles in the Ionian Sea (Central Mediterranean). *Earth and Planetary Science Letters*, 461, 61–72. <https://doi.org/10.1016/j.epsl.2016.12.020>
- Haxby, W. F., & Turcotte, D. L. (1978). On isostatic geoid anomalies. *Journal of Geophysical Research*, 83(B11), 5473–5478. <https://doi.org/10.1029/JB083iB11p05473>
- Hetényi, G., Cattin, R., Brunet, F., Bollinger, L., Vergne, J., Nabelek, J. L., & Diamant, M. (2007). Density distribution of the India plate beneath the Tibetan Plateau: Geophysical and petrological constraints on the kinetics of lower-crustal eclogitization. *Earth and Planetary Science Letters*, 264(1-2), 226–244. <https://doi.org/10.1016/j.epsl.2007.09.036>
- Hirn, A., Nicolich, R., Gallart, J., Laigle, M., & Cernobori, L. (1997). Roots of Etna volcano in faults of great earthquakes. *Earth and Planetary Science Letters*, 148(1-2), 171–191. [https://doi.org/10.1016/S0012-821X\(97\)00023-X](https://doi.org/10.1016/S0012-821X(97)00023-X)
- Hollenstein, C., Kahle, H.-G., Geiger, A., Jenny, S., Goes, S., & Giardini, D. (2003). New gps constraints on the Africa-Eurasia plate boundary zone in Southern Italy. *Geophysical Research Letters*, 30(18), 1935. <https://doi.org/10.1029/2003GL017554>
- Karner, G., & Watts, A. (1983). Gravity anomalies and flexure of the lithosphere at mountain ranges. *Journal of Geophysical Research*, 88(B12), 10,449–10,477. <https://doi.org/10.1029/JB088iB12p10449>
- Klügel, A., Longpré, M. A., Garcia-Canada, L., & Stix, J. (2015). Deep intrusions, lateral magma transport and related uplift at ocean island volcanoes. *Earth and Planetary Science Letters*, 431, 140–149. <https://doi.org/10.1016/j.epsl.2015.09.031>
- Lentini, F., Grasso, M., & Carbone, S. (1987). Introduzione alla geologia della Sicilia e guida all'escursione.
- Lodolo, E., Civile, D., Zanolli, C., & Geletti, R. (2012). Magnetic signature of the Sicily Channel volcanism. *Marine Geophysical Research*, 33(1), 33–44. <https://doi.org/10.1007/s11001-011-9144-y>
- MacLennan, J., & Lovell, B. (2002). Control of regional sea level by surface uplift and subsidence caused by magmatic underplating of Earth's crust. *Geology*, 30(8), 675–678. [https://doi.org/10.1130/0091-7613\(2002\)030<0675:CORSLB>2.0.CO;2](https://doi.org/10.1130/0091-7613(2002)030<0675:CORSLB>2.0.CO;2)
- Mancinelli, P., Pauselli, C., Minelli, G., & Federico, C. (2015). Magnetic and gravimetric modeling of the central Adriatic region. *Journal of Geodynamics*, 89, 60–70. <https://doi.org/10.1016/j.jog.2015.06.008>
- Manuella, F. C., Brancato, A., Carbone, S., & Gresta, S. (2013). A crustal upper mantle model for southeastern Sicily (Italy) from the integration of petrologic and geophysical data. *Journal of Geodynamics*, 66, 92–102. <https://doi.org/10.1016/j.jog.2013.02.006>
- Manuella, F. C., Carbone, S., & Barreca, G. (2012). Origin of saponite-rich clays in a fossil serpentinite-hosted hydrothermal system in the crustal basement of the Hyblean Plateau (Sicily, Italy). *Clays and Clay Minerals*, 60(1), 18–31. <https://doi.org/10.1346/CCMN.2012.0600102>
- Mastrolombo Ventura, B., Serpelloni, E., Argnani, A., Bonforte, A., Bürgmann, R., Anzidei, M., et al. (2014). Fast geodetic strain-rates in eastern Sicily (southern Italy): New insights into block tectonics and seismic potential in the area of the great 1693 earthquake. *Earth and Planetary Science Letters*, 404, 77–88. <https://doi.org/10.1016/j.epsl.2014.07.025>
- McKenzie, D. (1984). The generation and compaction of partially molten rock. *Journal of Petrology*, 25(3), 713–765. <https://doi.org/10.1093/ptrology/25.3.713>

- McKenzie, D. (2010). The influence of dynamically supported topography on estimates of Te. *Earth and Planetary Science Letters*, 295(1-2), 127–138. <https://doi.org/10.1016/j.epsl.2010.03.033>
- Micallef, A., Camerlenghi, A., Garcia-Castellanos, D., Cunarro Otero, D., Gutscher, M. A., Barreca, G., et al. (2018). Evidence of the Zanclean megaflood in the eastern Mediterranean Basin. *Scientific Reports*, 8, 1078. <https://doi.org/10.1038/s41598-018-19446-3>
- Moretti, I., & Royden, L. (1988). Deflection, gravity anomalies and tectonics of doubly subducted continental lithosphere: Adriatic and Ionian Seas. *Tectonics*, 7(4), 875–893. <https://doi.org/10.1029/TC007i004p00875>
- Musumeci, C., Scarfi, L., Palano, M., & Patan'e, D. (2014). Foreland segmentation along an active convergent margin: New constraints in southeastern Sicily (Italy) from seismic and geodetic observations. *Tectonophysics*, 630(C), 137–149. <https://doi.org/10.1016/j.tecto.2014.05.017>
- Neres, M., Terrinha, P., Custodio, S., Silva, S. M., Luis, J., & Miranda, J. M. (2018). Geophysical evidence for a magmatic intrusion in the ocean-continent transition of the SW Iberia margin. *Tectonophysics*, 744, 118–133. <https://doi.org/10.1016/j.tecto.2018.06.014>
- Neri, M., Rivalta, E., Maccaferri, F., Acocella, V., & Cirrincione, R. (2018). Etnean and Hyblean volcanism shifted away from the Malta Escarpment by crustal stresses. *Earth and Planetary Science Letters*, 486, 15–22. <https://doi.org/10.1016/j.epsl.2018.01.006>
- Ogniben, L. (1969). Schema introduttivo alla geologia del confine Calabro-Lucano. *Meteorological Society Geology Italian*, 8, 453–763.
- Patacca, E., Scandone, P., Giunta, G., & Liguori, V. (1979). Mesozoic paleotectonic evolution of the Ragusa zone (Southeastern Sicily). *Geologica Romana*, 18, 331–369.
- Pedley, M., & Grasso, M. (1992). Miocene syntectonic sedimentation along the western margins of the Hyblean-Malta platform: A guide to plate margin processes in the central Mediterranean. *Journal of Geodynamics*, 15(1-2), 19–37. [https://doi.org/10.1016/0264-3707\(92\)90004-C](https://doi.org/10.1016/0264-3707(92)90004-C)
- Pedley, M., Grasso, M., Maniscalco, R., & Esu, D. (2007). The Monte Carrubba Formation (Messinian, Sicily) and its correlatives: New light on basin-wide processes controlling sediment and biota distributions during the Palaeomediterranean-Mediterranean transition. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 253(3-4), 363–384. <https://doi.org/10.1016/j.palaeo.2007.06.009>
- Perkins, J. P., Ward, K. M., De Silva, S. L., Zandt, G., Beck, S. L., & Finnegan, N. J. (2016). Surface uplift in the central Andes driven by growth of the Altiplano Puna magma body. *Nature Communications*, 7, 13185. <https://doi.org/10.1038/ncomms13185>
- Ragg, S., Grasso, M., & Muller, B. (1999). Patterns of tectonic stress in Sicily from borehole breakout observations and finite element modeling. *Tectonics*, 18(4), 669–685. <https://doi.org/10.1029/1999TC900010>
- Ramberg, H. (1981). *Gravity, Deformation and the Earth's Crust* (p. 452). New York: Academic Press.
- Rohrman, M., & van der Beek, P. (1996). Cenozoic postrift domal uplift of north Atlantic margins: an asthenospheric diapirism model. *Geology*, 24(10), 901–904. [https://doi.org/10.1130/0091-7613\(1996\)024<0901:CPDUON>2.3.CO;2](https://doi.org/10.1130/0091-7613(1996)024<0901:CPDUON>2.3.CO;2)
- Romagnoli, G., Catalano, S., Pavano, F., & Tortorici, G. (2015). Geological map of the Tellaro River Valley (Hyblean Foreland, southeastern Sicily, Italy). *Journal of Maps*, 11(1), 66–74. <https://doi.org/10.1080/17445647.2014.944878>
- Rota, F. A., & Fichera, R. (1987). Magnetic interpretation related to geomagnetic provinces: the Italian case history. *Tectonophysics*, 138(2-4), 179–196. [https://doi.org/10.1016/0040-1951\(87\)90038-2](https://doi.org/10.1016/0040-1951(87)90038-2)
- Sandwell, D. T., Müller, R. D., Smith, W. H., Garcia, E., & Francis, R. (2014). New global marine gravity model from CryoSat-2 and Jason-1 reveals buried tectonic structure. *Science*, 346(6205), 65–67. <https://doi.org/10.1126/science.1258213>
- Sandwell, D. T., & Smith, W. H. (1997). Marine gravity anomaly from geosat and ers-1 satellite altimetry. *Journal of Geophysical Research*, 102(B5), 10,039–10,054. <https://doi.org/10.1029/96JB03223>
- Sandwell, D. T., & Smith, W. H. (2009). Global marine gravity from retracked geosat and ers-1 altimetry: Ridge segmentation versus spreading rate. *Journal of Geophysical Research*, 114, B01411. <https://doi.org/10.1029/2008JB006008>
- Scandone, P., Patacca, E., Radoicic, R., Ryan, W., Cita, M., Rawson, M., et al. (1981). Mesozoic and Cenozoic rocks from Malta escarpment (central Mediterranean). *AAPG Bulletin*, 65(7), 1299–1319.
- Scarfi, L., Barberi, G., Barreca, G., Cannavo, F., Koulakov, I., & Patanè, D. (2018). Slab narrowing in the central Mediterranean: the Calabro-Ionian subduction zone as imaged by high resolution seismic tomography. *Scientific Reports*, 8, 5178. <https://doi.org/10.1038/s41598-018-23543-8>
- Schellart, W. P. (2010). Mount Etna-Iblean volcanism caused by rollback-induced upper mantle upwelling around the Ionian slab edge: An alternative to the plume model. *Geology*, 38(8), 691–694. <https://doi.org/10.1130/G31037.1>
- Schmincke, H.-U., Behncke, B., Grasso, M., & Raffi, S. (1997). Evolution of the northwestern Iblean mountains, Sicily: uplift, Pliocene/Pleistocene sea-level changes, paleoenvironment, and volcanism. *Geologische Rundschau*, 86(3), 637–669. <https://doi.org/10.1007/s005310050169>
- Scribano, V., Sapienza, G., Braga, R., & Morten, L. (2006). Gabbroic xenoliths in tuff-breccia pipes from the Hyblean Plateau: Insights into the nature and composition of the lower crust underneath Southeastern Sicily, Italy. *Mineralogy and Petrology*, 86(1-2), 63–88. <https://doi.org/10.1007/s00710-005-0092-6>
- Smith, W. H., & Sandwell, D. T. (1997). Global sea floor topography from satellite altimetry and ship depth soundings. *Science*, 277(5334), 1956–1962. <https://doi.org/10.1126/science.277.5334.1956>
- Strak, V., & Schellart, W. P. (2014). Evolution of 3-D subduction-induced mantle flow around lateral slab edges in analogue models of free subduction analysed by stereoscopic particle image velocimetry technique. *Earth and Planetary Science Letters*, 403, 368–379. <https://doi.org/10.1016/j.epsl.2014.07.007>
- Tanguy, J. C., Condomines, M., & Kieffer, G. (1997). Evolution of the Mount Etna magma: Constraints on the present feeding system and eruptive mechanism. *Journal of Volcanology and Geothermal Research*, 75(3-4), 221–250. [https://doi.org/10.1016/S0377-0273\(96\)00065-0](https://doi.org/10.1016/S0377-0273(96)00065-0)
- Turcotte, D. L. (1979). *Flexure in: Advances in Geophysics* (Vol. 21). New York: Elsevier.
- Turcotte, D. L., & Schubert, G. (1982). *Geodynamics: Applications of continuum physics to geological problems* (p. 450). New York: John Wiley.
- Watts, A. B. (2001). *Isostasy and Flexure of the Lithosphere*. Cambridge: Cambridge University Press.
- Wellman, P. (1986). Intrusions beneath large intraplate volcanoes. *Exploration Geophysics*, 17(3), 135–139. <https://doi.org/10.1071/EG986135>
- Wessel, P., & Smith, W. H. (1991). Free software helps map and display data. *Eos, Transactions American Geophysical Union*, 72(41), 441–446. <https://doi.org/10.1029/90EO00319>
- Wilson, D. J., Peirce, C., Watts, A. B., Grevemeyer, I., & Krabbenhoef, A. (2010). Uplift at lithospheric swells - I: Seismic and gravity constraints on the crust and uppermost mantle structure of the Cape Verde mid-plate swell. *Geophysical Journal International*, 182(2), 531–550. <https://doi.org/10.1111/j.1365-246X.2010.04641.x>

- Won, I., & Bevis, M. (1987). Computing the gravitational and magnetic anomalies due to a polygon: Algorithms and fortran subroutines. *Geophysics*, 52(2), 232–238. <https://doi.org/10.1190/1.1442298>
- Yellin-Dror, A., Grasso, M., Ben-Avraham, Z., & Tibor, G. (1997). The subsidence history of the northern Hyblean plateau margin, southeastern Sicily. *Tectonophysics*, 282(1-4), 277–289. [https://doi.org/10.1016/S0040-1951\(97\)00228-X](https://doi.org/10.1016/S0040-1951(97)00228-X)