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Etnean and Hyblean volcanism shifted away from the Malta Escarpment by crustal stresses

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Intraplate volcanism, fault scarp, dike propagation, Malta Escarpment, Hyblean volcanism, Etna.

ABSTRACT

A fraction of the volcanic activity occurs intraplate, challenging our models of melting and magma transfer to the Earth's surface. A prominent example is Mt. Etna, eastern Sicily, offset from the asthenospheric tear below the Malta Escarpment proposed as its melt source. The nearby Hyblean volcanism, to the south, and the overall northward migration of the eastern Sicilian volcanism are also unexplained. Here we simulate crustal magma pathways beneath eastern Sicily, accounting for regional stresses and decompression due to the increase in the depth of the Malta Escarpment. We find non-vertical magma pathways, with the competition of tectonic and loading stresses controlling the trajectories' curvature and its change in time, causing the observed migration of volcanism. This suggests that the Hyblean and Etnean volcanism have been fed laterally from a melt pooling region below the Malta Escarpment. The case of eastern Sicily shows how the reconstruction of the evolution of magmatic provinces may

require not only an assessment of the paleostresses, but also of the contribution of surface loads and their variations; at times, the latter may even prevail. Accounting for these competing stresses may help shed light on the distribution and wandering of intraplate volcanism

1. INTRODUCTION

The distribution of volcanoes on the Earth's surface is determined by two main factors: the availability of melt at depth, generally associated with the decompression of mantle or crustal material, and the presence of a pathway for magma to ascend to the surface, controlled by lithospheric discontinuities or crustal stresses enhancing dike propagation and ascent. Intraplate volcanism is especially challenging to explain. The debate focuses on both the melting mechanisms and magma availability at a more regional scale (Tang et al., 2014), as well as the crustal magma pathways and their controlling features (Shabanian et al., 2012). The Etnean and Hyblean volcanism in eastern Sicily (Fig. 1) provides an exemplary case to investigate these processes at the crustal scale.

Mt. Etna is the site of sustained volcanism whose composition, similar to that of Ocean Island Basalts (OIB; Niu et al., 2011), has traditionally been ascribed to decompression of mantle material upwelling under extension (Barberi et al., 1973) or the presence of a plume (Tanguy et al., 1997). More recently, decompression melting related to suction or toroidal return flow sideways of the subduction underneath the Calabrian Arc has been suggested as an alternative to the plume hypothesis (Gvirtzman and Nur, 1999; Doglioni et al., 2001; Schellart, 2010; Faccenna et al., 2011), or in interaction with it (Schiano et al., 2001), to explain the compositional changes over time. The ascent pathway is widely identified in a lithospheric discontinuity to the east of Etna, the Malta Escarpment (Corsaro et al., 2002; Siniscalchi et al., 2012; Polonia et al., 2016). However, this latter hypothesis does not explain why Mt. Etna has developed on the footwall of, and not directly on, the Malta Escarpment (ME) itself. The earlier Hyblean volcanism, scattered to the south of Etna, is also located in the foreland, further away from the Malta Escarpment. Compositionally, the Hyblean volcanism is similar to OIB and highly similar to the earliest Etnean volcanism, lacking the transitional components to arc volcanism that some researchers identify for Etna (Tanguy et al., 1997); unlike to the Etnean volcanism, the Hyblean volcanism does not show any significant trend of magmatic differentiation (Beccaluva et al., 1998; Schiano et al., 2001; Manuella et al., 2013). The Hyblean volcanism has been generically linked to extensional tectonics (Barberi et al., 1973) or to a deep-rooted mantle plume (Tanguy et al., 1997; Schiano et al., 2001). The Hyblean and Etnean magmas are spatially continuous to the west of the Malta Escarpment, as demonstrated by outcrops, boreholes and seismic reflection lines (Torelli et al., 1998; Branca et al., 2011). However, the compositional and spatial continuity of the Hyblean and Etnean volcanism, as well as its overall northward migration subparallel to, but offset from, the Malta Escarpment, are still unexplained.

Magma is predominately transported through the crust via magma-filled fractures, or dikes, propagating by hydraulic fracturing (Rubin, 1995). Dikes tend to follow trajectories perpendicular to the least compressive principal stress, σ_3 (Dahm, 2000), thus being very sensitive to the crustal state of stress. Heterogeneous crustal stresses may drive dike propagation through curved or anyway complex pathways (Dahm, 2000; Roman and Jaupart, 2014; Sigmundsson et al., 2015;). In particular, the distribution of surface loads may exert a major control on crustal stress (Dahm, 2000; Watanabe et al., 2002). If surface loads are not evenly distributed (e.g. if there are volcanic edifices and fault scarps) magma pathways may bend, reaching the surface offset up to several tens of km from the melt pooling zone (Watanabe et al., 2002; Maccaferri et al., 2014). Here we test the hypothesis that the significantly asymmetric distribution of loads across the Malta Escarpment may have been responsible for bending dike trajectories towards the fault footwall. We combine for the first time two separate notions: that observations of surface volcanism (vent location and fissure orientation) are indicative of paleostresses, and that elastic stresses due to gravitational loads may play a role and at times even dominate over tectonic contributions (e.g. in case of major topographic or bathymetric variations). Based on this approach to evaluate stresses, we constrain the location of Etnean and Hyblean melt source by backtracking the simulated dike trajectories from the observed eruptive locations down to the Moho.

2. TECTONIC SETTING AND VOLCANISM

The Malta Escarpment, a regional NNW-SSE trending fault system offshore eastern Sicily, separates the thinner Ionian oceanic lithosphere subducting beneath the Calabrian Arc (to the east) from the thicker continental crust of the African foreland (to the west) (Fig. 1). The Malta Escarpment probably activated since Mesozoic, even though its main faulting event occurred in the Neogene (Upper Miocene; Argnani and Bonazzi, 2005). The ~350 km^2 wide scattered volcanic activity in the Hyblean region has involved four main cycles since the Upper Triassic, the latter occurring in the Upper Miocene–Lower Pleistocene (1-10 Ma; Patacca et al., 1979; Carbone et al., 1987; Beccaluva et al., 1998; Manuella et al., 2013). Over these cycles, volcanism in the Hyblean foreland migrated northwestward (Fig. 2). Westward migration of the eruptive vents during the Plio-Pleistocene was accompanied by 90° rotations in the alignment of the eruptive fissures, activating NE-SW to ENE-WSW trending dextral to normal faults, (SG and LSG in Fig. 2), suggesting the tectonic regime became compressive (Ghisetti and Vezzani, 1980; Grasso et al., 1991; Beccaluva et al., 1998). Such an overall E-W compression may have continued until ~0.85 Ma (Catalano et al., 2008), or was replaced by neutral tectonic conditions (Cultrera et al., 2015), anticipating the E-W extension from ~0.85 Ma until the present (Catalano et al., 2008; Montone et al., 2012; Cultrera et al., 2015).

Volcanic activity in the Hyblean foreland terminated at ~1.4 Ma (Torelli et al., 1998) and then focused in the Etna area during the last 0.5 Ma. Therefore, this change from compression to extension in the Hyblean foreland seems associated with the northward migration of volcanism towards the Etna area. Such a coexistence of volcanism with compression and its migration at the onset of extension is not straightforward and deserves an appropriate tectono-magmatic model.

The first period of volcanism in the Etna area (600-250 ka) involved mainly tholeiitic and transitional to Na-alkaline magmas, forming both sub-volcanic bodies and lava flows emplaced in submarine/subaerial environment along fissures or from minor cones (Corsaro et al., 2002; Branca et al., 2011). Between 225 and 142 ka, eruptions occurred mostly from fissures forming a ~N-S trending edifice, producing most of the lavas presently outcropping along the Timpe Fault System, the onland surface expression of the Malta Escarpment on the distal eastern flank of Etna (Chiocci et al., 2011; Siniscalchi et al., 2012; Catalano et al., 2013). During the last 110 ka, the fissural volcanic activity became central; the major eruptive axes migrated westward, building strato-volcanoes up to 3800 m high, between 110 and 15 ka. The volcano today, 3,324 m high, uses the same central conduit that has been active over the last 65 ka (Branca et al., 2011). Mt. Etna has been undergoing an overall ~E-W trending extension (Catalano et al., 2013) associated with the activity of the Malta Escarpment (Monaco et al., 1997; Siniscalchi et al., 2012; Mattia et al., 2015) and/or to N-S trending regional compression due to the southward propagation of the Apenninic Chain (Lanzafame et al., 1997; Billi et al., 2010).

3. METHODS

We use a boundary element numerical code for dike propagation (Maccaferri et al., 2011) based on Dahm (2000). Dikes are boundary-element pressurised cracks in 2D (plane strain). Dike trajectories are computed by maximising the elastic and gravitational energy release on incremental elongations of a magma filled fracture in different directions. The dikes propagate upwards due to buoyancy: here the density of magma is set to 2300 kg m^{-3} while the density of the host rock is 2600 kg m^{-3}. We consider three snapshots for the early development of the Etnean and three for the Hyblean magmatism, respectively, based on the geologic history of the area. For each snapshot, we reconstruct the scarp height and distribution of surface loads. We next derive the induced crustal stresses based on analytical formulas for the stress change induced in an elastic half-space by a vertically oriented force (Davis and Selvadurai, 1996; Roman and Jaupart, 2014). Finally, we superpose tectonic extension or compression, as appropriate.

The intensity of the unloading pressure due to the creation of the scarp bathymetric depression, P^U, is $\Delta\rho gh$, where $\Delta\rho$ is the rock density, ρ_r, minus the density of water, ρ_w (this accounts for the load of the water column above the hanging wall of the Malta Escarpment), g is the acceleration due to gravity, and h is the topography height along profiles perpendicular to the Malta Escarpment through Etna (profile 1) or through the Hyblean foreland (profile 2). Finally, we superpose tectonic extension or compression, as appropriate. We adopt a reference frame with the y-axis oriented along the ME, the x-axis perpendicular to it and the z-axis pointing downward. In such a reference frame, the total stress tensor acting within the crust ($\sigma^{tot}{}_{ij}$) is the superposition of an isotropic lithostatic state of stress (with lithostatic pressure $P^L=\rho_r gz$), the stress change induced by unloading forces applied at the Earth's surface ($\sigma^U{}_{ij}$), reported e.g. by Dahm (2000) or Watanabe et al. (2002), and a horizontal tectonic extension (σ^T), acting uniformly through the crust. Thus:

$$\sigma^{tot}{}_{ij} = P^L\delta_{ij} + \sigma^U{}_{ij} + \sigma^T{}_{ij}, \qquad (1)$$

where δ_{ij}=1 if i=j and 0 if i≠j, and the only non-zero component of the tectonic stress tensor, $\sigma^T{}_{ij}$, is $\sigma^T{}_{xx}=\sigma_T$. We use σ^T=5 MPa and -5 MPa for horizontal extension and compression, respectively, and σ^T=1 MPa for weaker.

After calculating magma trajectories, we select those that emerge at locations of observed volcanism, at distance x_{vol} from the ME, and follow them backwards, until we reach a depth of 30 km. There, the trajectory will be at distance x_{mag} from the ME. Since our model is elastic, we cannot extend our trajectory analysis into the Earth's mantle. For Etna, the deepest extent of our domain, the crust-mantle boundary, coincides with a magma storage region, as constrained by geochemistry (Tanguy et al., 1997; Schiano et al., 2001; Doglioni et al., 2001; Clocchiatti et al., 2004). For the Hyblean magma, while there is evidence for a source between 30-80 km of depth, there is no evidence for a magma storage region at the mantle-crust boundary (Beccaluva et al., 1998). This means that for Etna we may be able to locate the magma storage region in (x,y,z) by crossing our trajectories with the petrology information. As for the Hyblean magma, its chemistry does not require a storage region, and the melt source is deeper than our modeling domain. However, crustal pathways simulated down to 30 km depth may still provide important constraints on the melt source.

As for profile 1 (Etna, Fig. 3), we consider three snapshots (t_1=100 ka, t_2=250 ka and t_3=500 ka). We retrieve the correspondent bathymetric profiles by first removing the edifice of Mt. Etna (we keep a dashed topographic profile for Mt. Etna in all figures for reference) and then correcting the current Malta Escarpment bathymetry (height difference ~1700 m, width ~20 km, fault dip ~60^{o}) based on information on recent slip and sedimentation rates, as follows. We consider a Quaternary slip rate (Argnani and Bonazzi, 2005; Mastrolembo Ventura et al., 2014) of 2 mm yr^{-1}, resulting in a subsidence rate of the hanging wall of 1.73 mm yr^{-1}. Considering a sedimentation rate of 0.05-0.5 mm yr^{-1} (A. Argnani, personal communication) and the smaller density of sea floor sediments with regard to average crustal rocks, we correct the subsidence rate to 1.7 mm yr^{-1}. We obtain scarp heights h_1=1530 m, h_2=1275 m, h_3=850 m at t_1=100 ka, t_2=250 and t_3=500 ka, respectively.

We set up a numerical simulation for the best-constrained, latest phase of Hyblean volcanism (Profile 2, Fig. 4) following the general lines used for Etna. However, the bathymetry corresponding to this phase of Hyblean volcanism is less constrained, as the previously used Quaternary rates cannot be extrapolated for earlier periods. Therefore, we consider three snapshots with height difference on the Malta Escarpment of 500, 1500 and 2000 m (the current height difference at the latitude of the Hyblean foreland is ~2400 m) and width of 3, 9 and 12 km (preserving the current slope), that we associate to three earlier phases in the development of the Malta Escarpment.

4. SIMULATIONS FOR DIKE PATHWAYS

The load gradient exerted by the bathymetry of the Malta Escarpment rotates the principal stresses and produces curved dike trajectories in section view, linking the observed volcanism (orange triangles at the surface, x=x_{vol}, in Figs. 3 and 4) to melt sources (orange-filled dikes at 30 km depth, x=x_{mag}) offset to the east, below the Malta Escarpment. The bending of the dike trajectories depends non-linearly on the competition between decompression (due to the activity of the fault, or scarp height) and tectonic stress, as illustrated below.

For the tectonic extension σ^T = 5 MPa, the simulations link the ~100 ka old (h_1=1530 m) volcanism at Etna (orange triangles at -25 km<x_{vol}<5 km), to a source of magma located at -20<x_{mag}<5 (orange-filled dikes), slightly to the west of the Malta Escarpment zone (Fig. 3a). A direct link to a source below the Malta Escarpment requires a weaker tectonic extension (σ^T = 1 MPa), that results in -15 km<x_{mag}<10 km (Fig. 3d). At ~250 ka (h_2=1275 m), the observed volcanism (-10 km<x_{vol}<0 km) is backtracked to 0 km<x_{mag}<15 km, below the Malta Escarpment, for σ^T = 5 MPa (Fig. 3b). At 500 ka (h_3=850 m), for σ^T =5 MPa the subaerial/submarine volcanic activity (-5 km<x_{vol}<5 km) is again linked directly to a magma source below the Malta Escarpment, at 0 km<x_{mag}<20 km.

Similarly to Etna, our simulations for the Hyblean region show that dikes nucleated below the Malta Escarpment are deviated westwards and intersect the surface on the footwall (Fig. 4). An extensional tectonic stress cannot explain the last phase of Hyblean volcanism (Fig. 4a), but a compressional tectonic stress with σ^T=-5 MPa reproduces the observed distal fissures perpendicular to the Malta Escarpment (Fig. 2, orange fissures). In fact, σ_3 is here out of plane (Fig. 4d) and NNW-SSE trending dikes are expected to twist by 90^o, intersecting the surface as ENE-WSW trending fissures. Extensional stresses are however consistent with the earlier phases (Fig. 2, cyan and purple fissures) of the Hyblean volcanism, with Malta Escarpment-parallel fissures closer to the coast (Fig. 4b, c).

Distal arrivals require weaker extension (in the younger Etna case) or compression (in the younger Hyblean case) (Fig. 5). The ratio of tectonic extension to decompression, $\sigma^T/P^U = \sigma^T/(\Delta\rho gh)$, controls the bending (rotation around a ~N-S axis) of the pathways and the twisting (rotation around an axis parallel to the dike propagation direction) of the dikes.

If a positive (extensional) σ^T dominates over P^U the trajectories are nearly vertical and offset from the Malta Escarpment only by a few km. If P^U dominates, or σ^T is negative (compressional), the trajectories become inclined or even sub-horizontal, inhibiting the upward propagation of magma, as in the trajectories terminating at depth (see black crosses in Fig. 4d-e-f). If the ratio is such that σ_3 becomes out-of-plane, then dikes are expected to twist and become orthogonal to the Malta Escarpment. Therefore, depending on the σ^T/P^U value, the location of the ascent of magma may be promoted or not, the pathways may change from subvertical to subhorizontal, and the surface fissures may be parallel or perpendicular to the Malta Escarpment.

5. A COMPREHENSIVE MODEL

Our results show that both the Hyblean and early Etnean magma pathways dip towards the Malta Escarpment. For the Etnean case, we locate the 30 km deep magma storage zone inferred from petrology directly below the Malta Escarpment (Fig. 6a-c). For the Hyblean case, arresting our analysis at Moho level does not allow us to reach the inferred melt generation zone (between 30 and 80 km; Beccaluva et al., 1998; Klemme and O’Neill, 2000). Still, we find Hyblean crustal magma pathways to follow the same overall picture of the Etnean pathways (Fig. 6d-f). A melt generation zone below the Malta Escarpment may thus explain both the Hyblean and Etnean volcanism, provided an appropriate amount of regional extension or compression is coupled to the crustal decompression due to the deepening of the seafloor induced by the activity of the Malta Escarpment. Indeed, deepening of the Malta Escarpment alone is not sufficient to explain the recent volcanism in eastern Sicily. In fact, a switch from extension to compression for the most recent Hyblean phase and at least a weakening of extension over the lifetime of the Etnean volcanism are both needed to explain the location, migration and preferred orientation of volcanism. Tectonic studies support our models, showing that the modelled extension and the compression coexisted during the Etnean and Hyblean volcanism, respectively (Ghisetti and Vezzani, 1980; Grasso et al., 1991; Monaco et al., 1997; Beccaluva et al., 1998; Catalano et al., 2008; Cultrera et al., 2015). The weakening of the ~E-W extension in the Etna area may be explained by the repeated emplacement of ~N-S trending dikes, which may even create a local and/or transient compressive stress field (Vigneresse et al., 1999).

Besides being mechanically plausible, the hypothesis of the Hyblean and Etnean magma sourced below the Malta Escarpment is consistent with previous petrological models of the Hyblean volcanism (Beccaluva et al., 1998; Schiano et al., 2001; Manuella et al., 2013). Indeed, Hyblean magmas are relatively undifferentiated, even though lavas cover a large serial affinities, varying from quartz-tholeiites, to alkaline basalts up to nephelinites (Manuella et al., 2015). This wide compositional range results from different degrees of partial melting of a metasomatized mantle (Scribano et al., 2009; Viccaro and Zuccarello, 2017); major and trace element contents are compatible with source depths within spinel peridotite facies lithospheric mantle between 30 and 80 km (Beccaluva et al., 1998; Klemme and O'Neill, 2000).

Our models cannot help constraining the depth of magma generation or magma stalling, nor the time spent by the magma at different depths. However, by following the magma trajectories they return from the surface down to 30 km, which is the deepest we can go with an elastic model, we have proved mechanical consistency of melt generation below the Malta Escarpment rather than directly below Etna. Finally, even if we cannot reach the depth of magma generation for the source feeding the Hyblean volcanism, we still find magma trajectories through the crust to head for the same direction as for Etna, below the Malta Escarpment (Figs. 3, 4 and 6).

We should also note that in the Etna case, mainly a composite volcano made of trachybasalts and trachyandesites, we expect a differentiation from a magma reservoir; conversely, in the Hyblean case the fissure-fed tholeiitic and alkali basalts have virtually no trend of differentiation. These features imply a different feeding system, with a major magmatic reservoir for Etna, probably minor or absent beneath the Hyblean area. Based on our model, it is difficult to put constrains on the development of a magmatic reservoir at the crust-mantle boundary, as this may depend upon different factors beside stress, most importantly the supply rate of magma, its density and viscosity. The higher magmatic supply beneath Etna may originate from other regional processes, including the retreat of the subducting Ionian slab (Gvirtzman and Nur, 1999; Clocchiatti et al., 2004; Faccenna et al., 2011).

In this frame, the overall northward migration of volcanism in eastern Sicily and its spatial continuity may also be related to the progressive increase in the melt production due to the gradual and northward deepening (or increase in the morphological scarp) of the Malta Escarpment (ME).

Based on our simulations, we propose that the distribution and migration of volcanism in eastern Sicily stems from the combination of evolving topographic loads and decompressions and regional stresses varying from compressive to extensional. This study highlights the potential of associating mechanically consistent magma pathways to melting source hypotheses, in order to achieve better constraints on the origin of magmatism. While petrological studies can provide information on the depth of a melt source, here we provide a novel method to reconstruct its lateral position at depth. This method could be used on other complex cases to constrain the actual horizontal location of a melt source at the depth indicated by petrological evidences.

More generally, a comparative stress analysis may be often needed to properly explain intraplate magmatism. In our case, while the source of magmatism may be explained by decompression due to the thinning of the passive margin of the Malta Escarpment, the propagation path of the magma may be ascribed to the intraplate evolution of these stresses and the topographic/bathymetric configuration of the crust. The evolution of magmatic provinces in eastern Sicily provides an outstanding example of how crustal magma transport processes over time scales of millions of years, should be addressed considering both regional, tectonic paleostresses, and the contribution of surface loads and their variations. Such a combination of tectonic and loading stresses has so far explained several magmatic features of relatively simple and general tectonic settings, as narrow rifts, calderas or sector collapses of volcanic edifices (e.g., Maccaferri et al, 2011; Corbi et al., 2015; Maccaferri et al., 2017). In this study we show how we can extend our approach to specific and more complex tectonic contexts, characterized by both tectonic and topographic variations. We suggest that our approach can be applied to other areas worldwide, where a better understanding and definition of the evolution of the tectonic processes and surface elevation may help explain similar apparently enigmatic intraplate volcanism.

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REFERENCES CITED

Argnani, A., and Bonazzi, C., 2005, Tectonics of Eastern Sicily offshore: Tectonics, 24, doi:10.1029/2004TC001656 TC4009.

Barberi, F., Gasparini, P., Innocenti, F., and Villari, L., 1973, Volcanism of the Southern Tyrrhenian Sea and its Geodynamical Implications: J. Geophys. Res., 78, p. 5221–5232.

Beccaluva, L., Siena, F., Coltorti, M., Di Grande, A., Lo Giudice, A., Macciotta, G., Tassinari, R., and Vaccaro, C.,1998, Nephelinitic to tholeiitic magma generation in a transtensional tectonic setting: An integrated model for the Iblean volcanism, Sicily: J. Petrol., 39, p. 1547– 1576.

Billi, A., Presti, D., Orecchio, B., Faccenna, C., and Neri, G., 2010, Incipient extension along the active convergent margin of Nubia in Sicily, Italy: Cefalù-Etna seismic zone: Tectonics, 29, TC4026, 10.1029/2009TC002559.

Branca, S., Coltelli, M., Groppelli, G., and Lentini, F., 2011, Geological map of Etna volcano, 1:50,000 scale: Ital. J. Geosci., 130, p. 265–291, doi: 10.3301/IJG.2011.15.

Carbone, S., Grasso, M., and Lentini, F., 1987, Lineamenti geologici del Plateau Ibleo (Sicilia SE): presentazione delle carte geologiche della Sicilia Sud-Orientale: Mem. Soc. Geol. It., 38, p. 127–135.

Catalano, S., Bonforte, A., Guglielmino, F., Romagnoli, G., Tarsia, C., and Tortorici, G. 2013, The influence of erosional processes on the visibility of Permanent Scatterers Features from SAR remote sensing on Mount Etna (E Sicily): Geomorphology, 198, 128-137, 10.1016/j.geomorph.2013.05.020.

Catalano, S., De Guidi, G., Romagnoli, G., Torrisi, S., Tortorici, G., and Tortorici, L., 2008, The migration of plate boundaries in SE Sicily: influence on the large-scale kinematic model of the African promontory in southern Italy: Tectonophysics, 449, p. 41–62.

Chiocci, F.L., Coltelli, M., Bosmanba A., and Cavallaro, D, 2011, Continental margin large-scale instability controlling the flank sliding of Etna volcano: Earth Planet. Sci. Lett., 305, 57-64, doi:10.1016/j.epsl.2011.02.040.

Clocchiatti, R., Condomines, M., Guénot, N., Tanguy, J.C., 2004, Magma changes at Mount Etna: the 2001 and 2002-2003 eruptions, Earth Planet. Sci. Lett., 226, 397-414.

Corbi, F., Rivalta, E., Pinel, V., Maccaferri, F., Bagnardi, M., and Acocella, V., 2015, How caldera collapse shapes the shallow emplacement and transfer of magma in active volcanoes. Earth Plan. Sci. Lett., 431, 287-293.

Corsaro, R.A., Neri, M., and Pompilio, M., 2002, Paleo-environmental and volcano-tectonic evolution of the south-eastern flank of Mt. Etna during the last 225 ka inferred from volcanic succession of the «Timpe», Acireale, Sicily: J. Volcanol. Geotherm. Res., 113, p. 289-306, doi:10.1016/S0377-0273(01)00262-1.

Cultrera, F., Barreca, G., Scarfi, L., and Monaco, C., 2015, Fault reactivation by stress pattern reorganization in the Hyblean foreland domain of SE Sicily (Italy) and seismotectonic implications: Tectonophysics, 661, p. 215-228.

Dahm, T., 2000, Numerical simulations of the propagation path and the arrest of fluid-filled fractures in the Earth: Geophys. J. Int., 141, p. 623–638.

Davis, R., and Selvadurai, A., 1996, Elasticity and Geomechanics (Cambridge Univ. Press, 1996).

Doglioni, C., Innocenti, F., and Mariotti, G., 2001, Why Mt Etna?: Terra Nova, 13(1), p. 25–31, doi:10.1046/j.1365-3121.2001.00301.x.

Faccenna, C., Molin, P., Orecchio, B., Olivetti, V., Bellier, O., Funiciello, F., Minelli, L., Piromallo, C., and Billi, A., 2011, Topography of the Calabria subduction zone (southern Italy): clues for the origin of Mt. Etna: Tectonics, 30, TC1003, doi:10.1029/2010TC002694

Gvirtzman, Z., and Nur, A., 1999, The formation of Mount Etna as the consequence of slab rollback: Nature, 401, p. 782–785, doi:10.1038/44555.

Finetti, I., 1982, Structure, stratigraphy and evolution of Central Mediterranean. Boll. Geof.

Teor. Appl., XXVI, 96, 247-312.

Ghisetti, F., and Vezzani, L., 1980, The structural features of the Iblean plateau and of the Monte Iudica area (South Eastern Sicily): A microtectonic contribution to the deformational history of the Calabrian Arc., Boll. Soc. Geol. It., 99, p. 57–102.

Grasso, M., Pezzino, A., Reuther, C.D., Lanza, R., and Miletto, M., 1991, Late Cretaceous and Recent tectonic stress orientations recorded by basalt dykes at Capo Passero (southeastern Sicily): Tectonophysics, 185, p. 247-259.

Klemme, S., and O'Neill, H., 2000, The near-solidus transition from garnet lherzolite to spinel lherzolite: Contrib. Mineral. Petrol., 138, p. 237-248, doi:10.1007/s004100050560.

Lanzafame, G., Leonardi, A., Neri, M., and Rust, D., 1997, Late overthrust of the Appenine - Maghrebian Chain at the NE periphery of Mt. Etna, Sicily: C. R. Acad. Sci. Paris, t.324, serie II a, p. 325-332.

Maccaferri, F., Bonafede, M., and Rivalta, E., 2011, A quantitative study of the mechanisms governing dike propagation, dike arrest and sill formation: J. Volcanol. Geotherm. Res., 208, p. 39–50.

Maccaferri, F., Richter, N., Walter, T.R., 2017, The effect of giant lateral collapses on magma pathways and the location of volcanism. Nature Communications, 8, 1097, DOI: 10.1038/s41467-017-01256-2.

Maccaferri, F., Rivalta, E., Keir, D., and Acocella, V., 2014, Off-rift volcanism in rift zones determined by crustal unloading: Nature Geoscience, 7, p. 297-300.

Manuella, F.C., Brancato, A., Carbone, S., and Gresta, S., 2013, A crustal–upper mantle model for southeastern Sicily (Italy) from the integration of petrologic and geo-physical data: J. Geodyn., 66, p. 92–102.

Mastrolembo Ventura, B., Serpelloni, E., Argnani, A., Bonforte, A., Bürgmann, R., Anzidei, M., Baldi, P., and Puglisi, G., 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 Planet. Sci. Lett., 404, p. 77–88, doi:10.1016/j.epsl.2014.07.025.

Mattia, M., Bruno, V., Caltabiano, T., Cannata, A., Cannavò, F., D'Alessandro, W., Di Grazia, G., Federico, C., Giammanco, S., La Spina, A., Liuzzo, M., Longo, M., Monaco, C., Patanè, D., and Salerno, G., 2015, A comprehensive interpretative model of slow slip events on Mt. Etna's eastern flank, Geochem. Geophys. Geosyst., 16, 635-658, doi:10.1002/2014GC005585.

Monaco, C., Tapponnier, P., Tortorici, L., and Gillot, P.Y., 1997, Late Quaternary slip rates on the Acireale–Piedimonte normal faults and tectonic origin of Mt. Etna (Sicily): Earth Planet. Sci. Lett., 147, p. 125–139, doi:10.1016/S0012-821X(97)00005-8.

Montone, P., Mariucci, M.T., and Pierdominici, S., 2012,, The Italian present-day stress map: Geophys. J. Intern., 189, p. 705-716.

Morelli, C., Gantar, C., and Pisani, M., 1975, Bathymetry, gravity and magnetism in the Strait of Sicily and in the Ionian Sea. Boll. Geof. Teor. Appl., 17: 39-58.

Niu, Y., Wilson, M., Humphreyes, E.R., and O’Hara, M., 2011, The Origin of Intra-plate Ocean Island Basalts (OIB): the Lid Effect and its Geodynamic Implications: J. Petrol. 52, p. 1443-1468, doi: 10.1093/petrology/egr030.

Patacca, E., Scandone, P., Giunta, G., and Liguori, V., 1979, Mesozoic paleotectonic evolution of the Ragusa zone (southern Sicily): Geologica Romana, 18, p. 331–369.

Polonia, A., Torelli, L., Artoni, A., Carlini, M., Faccenna, C., Ferranti, L., Gasperini, L., Govers, R., Klaeschen, D., Monaco, C., Neri, G., Nijholt, N., Orecchio, B., and Wortel, R.,2016, The Ionian and Alfeo - Etna fault zones: New segments of an evolving plate boundary in the central Mediterranean sea?: Tectonophysics, 675, p. 69-90 doi:10.1016/j.tecto.2016.03.016.

Roman, A., and Jaupart, C., 2014, The impact of a volcanic edifice on intrusive and eruptive activity: Earth Planet. Sci. Lett., 408, p. 1-8, doi:10.1016/j.epsl.2014.09.016.

Rubin, A., 1995, Propagation of magma-filled cracks: Annu. Rev. Earth Planet. Sci., 23, p. 287–

336.

Schiano, P., Clocchiatti, R., Ottolini, L., and Busà, T., 2001, Transition of Mount Etna lavas from a mantle-plume to an island-arc magmatic source: Nature, 412, p. 900–904.

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, p. 691-694.

Scribano, V., Viccaro, M., Cristofolini, R., Ottolini, L., 2009, Metasomatic events recorded in ultramafic xenoliths from the Hyblean area (Southern Sicily, Italy). Miner. Petrol. 95, 235-250.

Shabanian, E., Acocella, V., Gioncada, A., Ghasemi, H., and Bellier, O., 2012, Structural control on magmatism in intraplate collisional settings: extinct example from NE Iran and current analogues: Tectonics, 31, TC3013, doi: 10.1029/2011TC003042.

Sigmundsson, F., et al. 2015, Segmented lateral dyke growth in a rifting event at Bárðarbunga volcanic system, Iceland: Nature, 517(7533), p. 191–195, doi:10.1038/nature14111.

Siniscalchi, A., Tripaldi, S., Neri, M., Balasco, M., Romano, G., Ruch, J., and Schiavone, D., 2012, Flank instability structure of Mt Etna inferred by a magnetotelluric survey: J. Geophys. Res., 117, B03216, doi:10.1029/2011JB008657.

Tang, J., Obayashi, M., Niu, F., Grand, S.P., Chen, Y.J., Kawakatsu, H., Tanaka, S., Ning, J., and Ni, J.F., 2014, Changbaishan volcanism in northeast China linked to subduction-induced mantle upwelling: Nature Geoscience, 7, p. 470-475.

Tanguy J.C., Condomines M., and Kieffer, G., 1997, Evolution of the Mount Etna magma: Constraints on the present feeding system and eruptive mechanism: J.Volcanol. Geotherm. Res., 75, 3-4, p. 221–250.

Tarquini, S., Isola, I., Favalli, M., Mazzarini, F., Bisson, M., Pareschi, M. T., and Boschi, E., 2007, TINITALY/01: a new Triangular Irregular Network of Italy: Ann. Geophys., 50, p. 407 – 425.

Torelli, L., Grasso, M., Mazzoldi, G., and Peis, D., 1998, Plio–Quaternary tectonic evolution and structure of the Catania foredeep, the northern Hyblean Plateau and the Ionian shelf (SE Sicily): Tectonophysics, 298, p. 209–221.

Viccaro, M., and Zuccarello, F., 2017, Mantle ingredients for making the fingerprint of Etna alkaline magmas: implications for shallow partial melting within the complex geodynamic framework of Eastern Sicily. J. Geodyn., 109, 10-23.

Vigneresse, JL, Tikoff, B., and Ameglio, L., 1999, Modification of the regional stress field by magma intrusion and formation of tabular granitic plutons: Tectonophysics, 302, p. 203-224, doi:10.1016/S0040-1951(98)00285-6.

Watanabe, T., Masuyama, T., Nagaoka, K., and Tahara, T., 2002, Analog experiments on magma-filled cracks: Competition between external stresses and internal pressure: Earth Planets Space, 54, p. 1247–1261.

FIGURE CAPTIONS

Figure 1. Simplified tectonic map of Sicily. P1 and P2 (white lines) are the location of the profiles illustrated in Figs. 3 and 4, respectively. The image of the Italian territory is obtained from the TINITALY DEM (Tarquini et al., 2007) re-sampled to 100 m resolution.

Figure 2. Etnean and Hyblean volcanism in eastern Sicily. Positions of eruptive centers (dots) and fissures (colored lines) in eastern Sicily (see Fig. 1 for location). Volcanic outcrops are in light gray. The arrows in the legend represent the more recent and better constrained direction of regional extension (white) or compression (black). SG=Simeto Graben; LSG=Lentini-Scordia Graben; ME=Malta Escarpment (main faults=black dotted lines). Data sources: Carbone et al., 1987; Grasso et al., 1991, Lanzafame et al., 1996; Monaco et al., 1997; Beccaluva et al., 1998; Corsaro et al., 2002, Branca et al., 2011.

Figure 3. Simulations for dike path in Etnean area. Principal stresses and simulated dike pathways along the EW profile P1 (location in Fig. 1); z is depth below sea level, x is horizontal coordinate along an ~E-W direction through Etna, perpendicular to the Malta Escarpment (ME). The dashed profile of Etna is included as a reference (its load is not included in the stress model). The left and right columns are for 5 and 1 MPa extensional stress, respectively; the different rows correspond to different fault scarp heights and epochs, as indicated in the inlets. Grey segments are σ_3 directions. Black curves are simulated dike pathways which link starting dikes (vertical ellipses at z=30 km and $x=x_{mag}$) to a surface triangle (z=0 km, $x=x_{vol}$). Surface volcanism observed in the individual epochs is indicated by orange-filled triangles; Orange ellipses correspond to feeder dikes at their nucleation depth. Color shading indicates intensity of

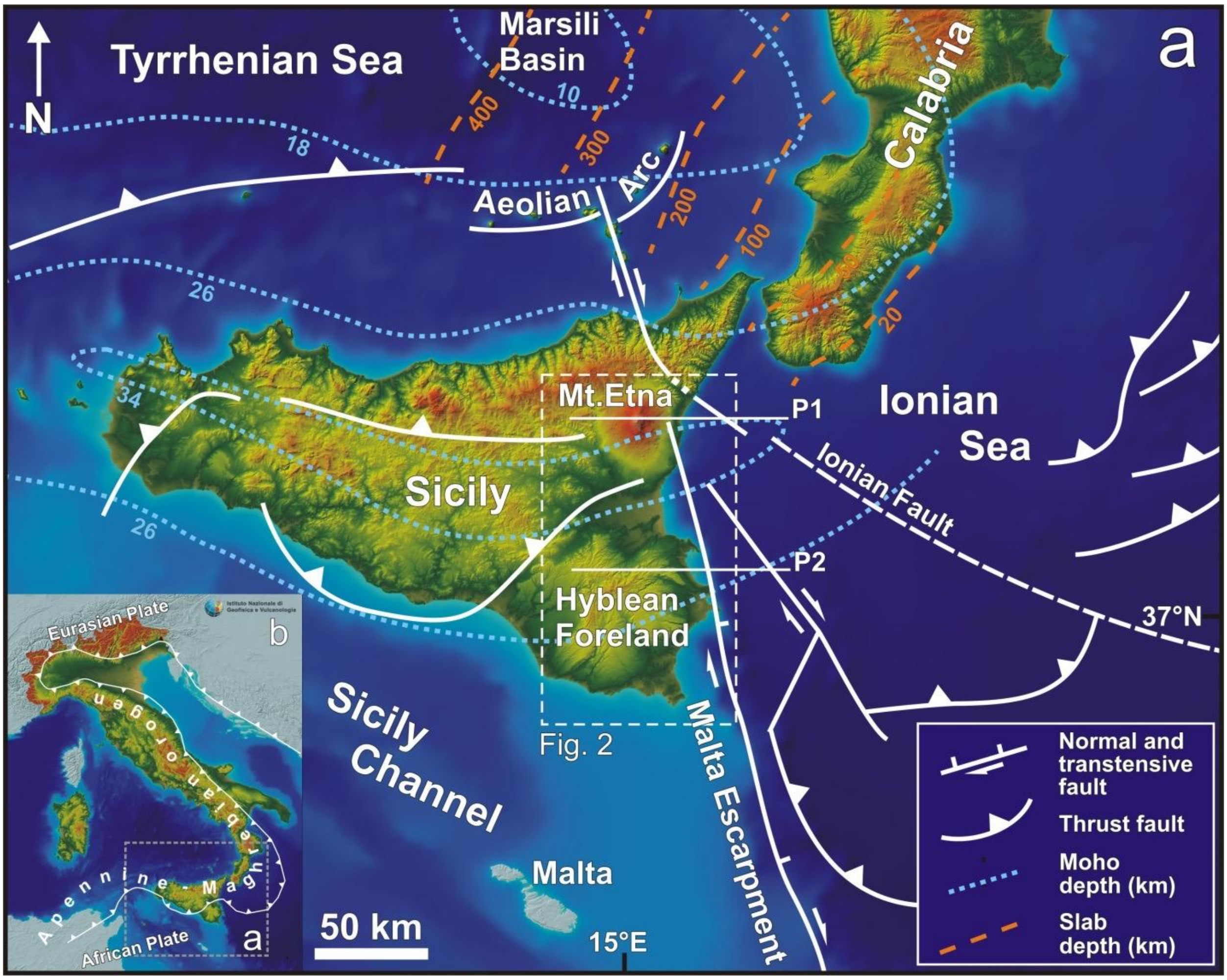
isotropic stress $\sigma^I=(1+\upsilon)(\sigma_{xx}+\sigma_{zz})/3$ (where plane strain is assumed), resulting from the combination of tectonic and unloading stresses. Stress is positive if extensional (see color bars to the right). Dikes more likely reach the surface for higher extensional stresses (darker colours). Parameters used are: crustal shear modulus is 20 MPa, Poisson's ratio is 0.25, rock density is 2600 kg m^{-3}, initial volume of the dikes is $2.5\cdot10^{-2}$ km^3 and magma buoyancy is 300 kg m^{-3}.

Figure 4. Simulations for dike path in Hyblean area. Same as Fig. 3, but relative to the ~E-W profile P2 through the Hyblean foreland (location in Fig. 1). The simulation refers to the fault scarp heights indicated in the inlets, developed during Upper Miocene–Lower Pleistocene (1-10 Ma). Grey segments are σ_3 directions; a circle indicates out of plane σ_3 (perpendicular to the page), inducing dikes to twist and align to the page. Pathways of dikes that are halted on their way without reaching the surface terminate with a cross (x). Color shading indicates intensity of isotropic stress $\sigma^I=(1+\upsilon)(\sigma_{xx}+\sigma_{zz})/3$, resulting from the combination of tectonic and unloading stresses. Stress is positive if extensional, negative if compressional (see color bars to the right).

Figure 5. Migration of volcanism in eastern Sicily due to variation of the regional tectonic stresses and crustal decompression. Distance d from the Malta Escarpment of surface fissures generated by dikes starting beneath the Malta Escarpment as a function of the tectonic/unloading stress ratio, σ^T/P^U. Crosses indicate dikes that turned perpendicular to the Malta Escarpment and extend laterally up to a distance *d*. The colored inlets on the right-hand side indicate observed distances from the Malta Escarpment for surface vents in the different epochs. The colors refer to those used in Fig. 2.

Figure 6. A synthetic tectono-magmatic model for eastern Sicily. Schematic E-W crustal cross-sections below Etna (a, b, c) and the Hyblean Foreland (d, e, f), along profiles P1 and P2,

respectively (location in Fig. 1). In both cases, the magma storage zones are imaged at ~30 km depth, i.e the starting depth of our models. Question marks show the uncertainties in the limits of the magma storage regions, especially for Hyblean case. Volcanism in the Etna region has been active since ~500 ka, and migrated westward under E-W and WNW-ESE tectonic extension. It has been possibly fed by a larger magma storage zone below the Malta Escarpment, also supplied at deeper levels from the asthenospheric window at the western edge of the subducting oceanic Ionian lithosphere (not shown here). Volcanism in the Hyblean Foreland was fed by a possible smaller magma storage zone located below the Malta Escarpment. Up to Miocene, dikes followed NNW-SSE oriented paths (e, f), driven by ENE-WSW tectonic extension. During Pliocene-Early Pleistocene (d), the tectonic regime switched to ENE-WSW compression and the dike trajectories rotated towards NE-SW to E-W paths. Crustal profiles from Morelli et al. (1975), and Finetti, (1982), modified.



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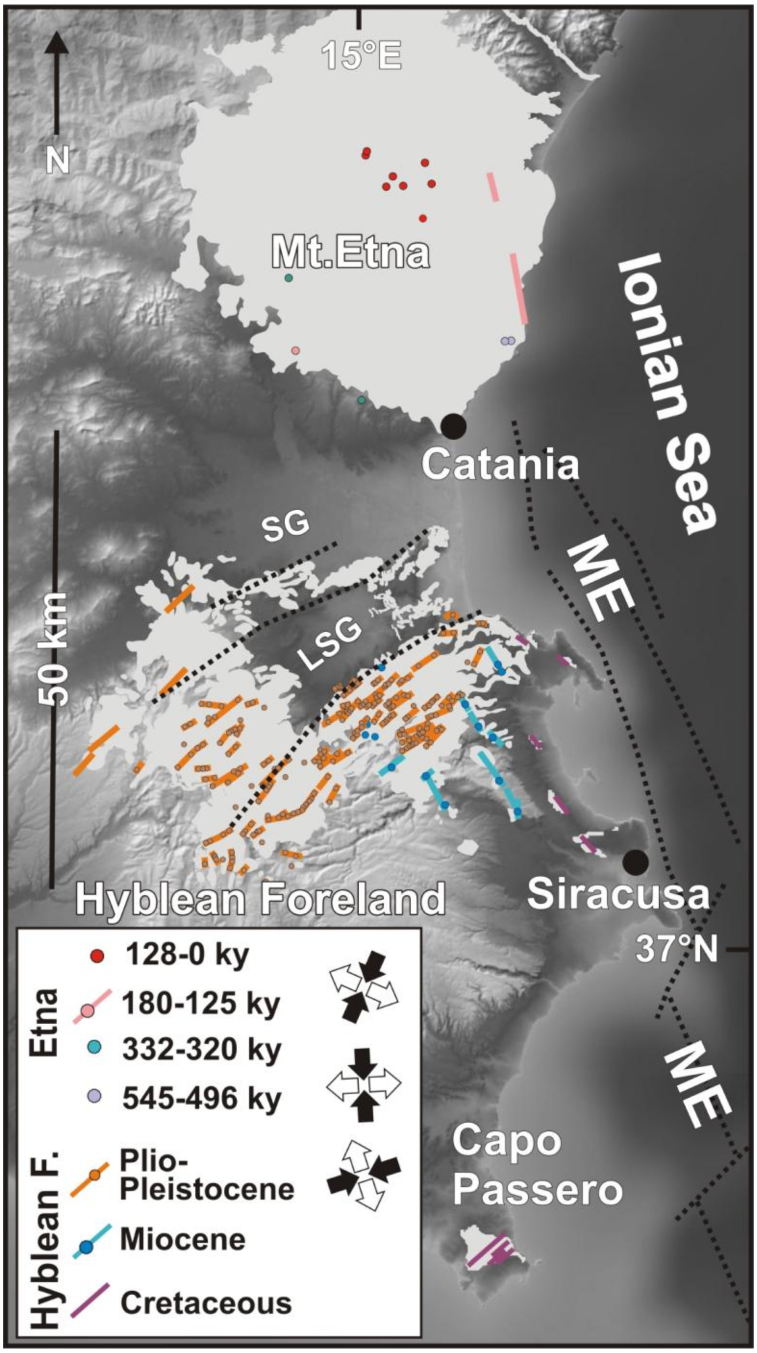

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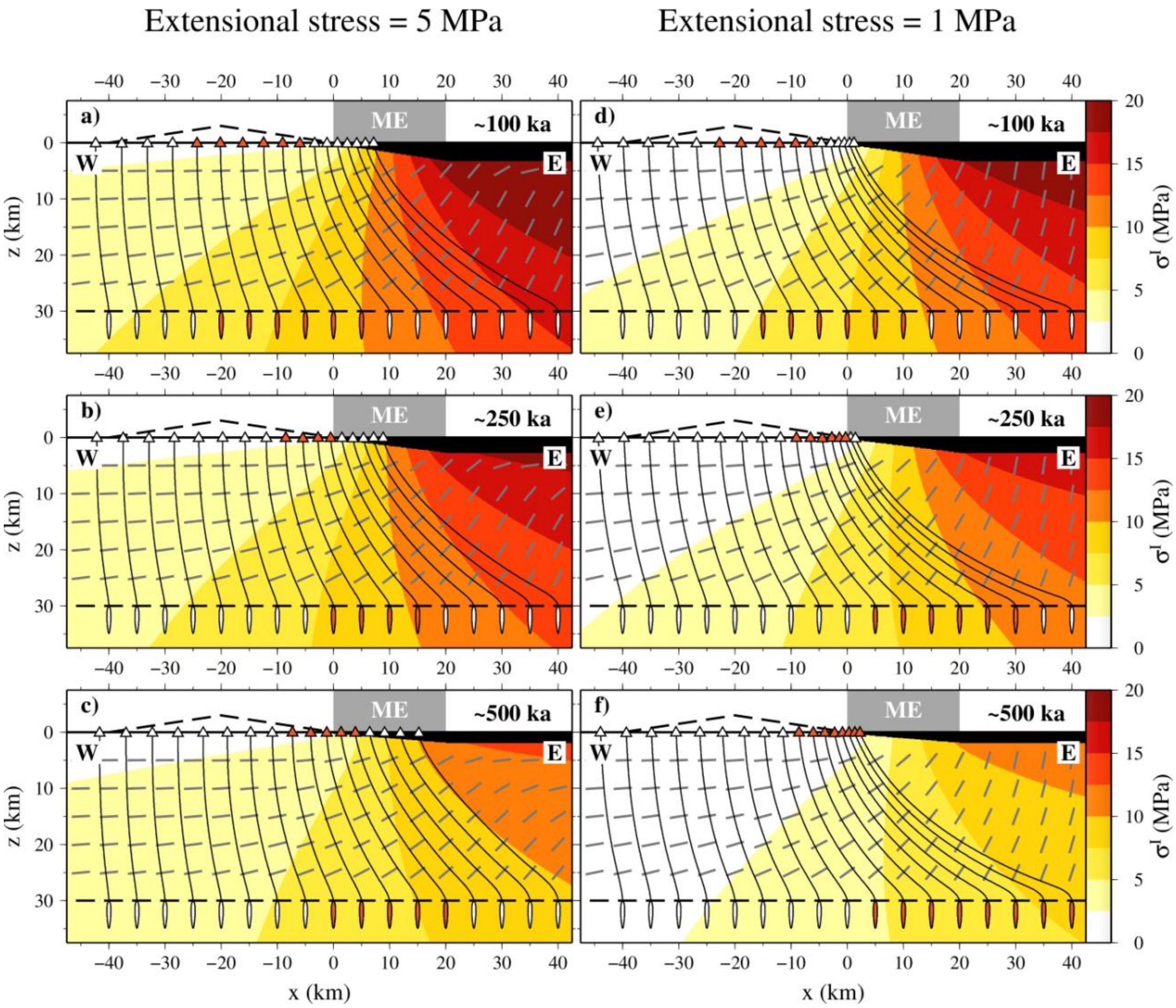

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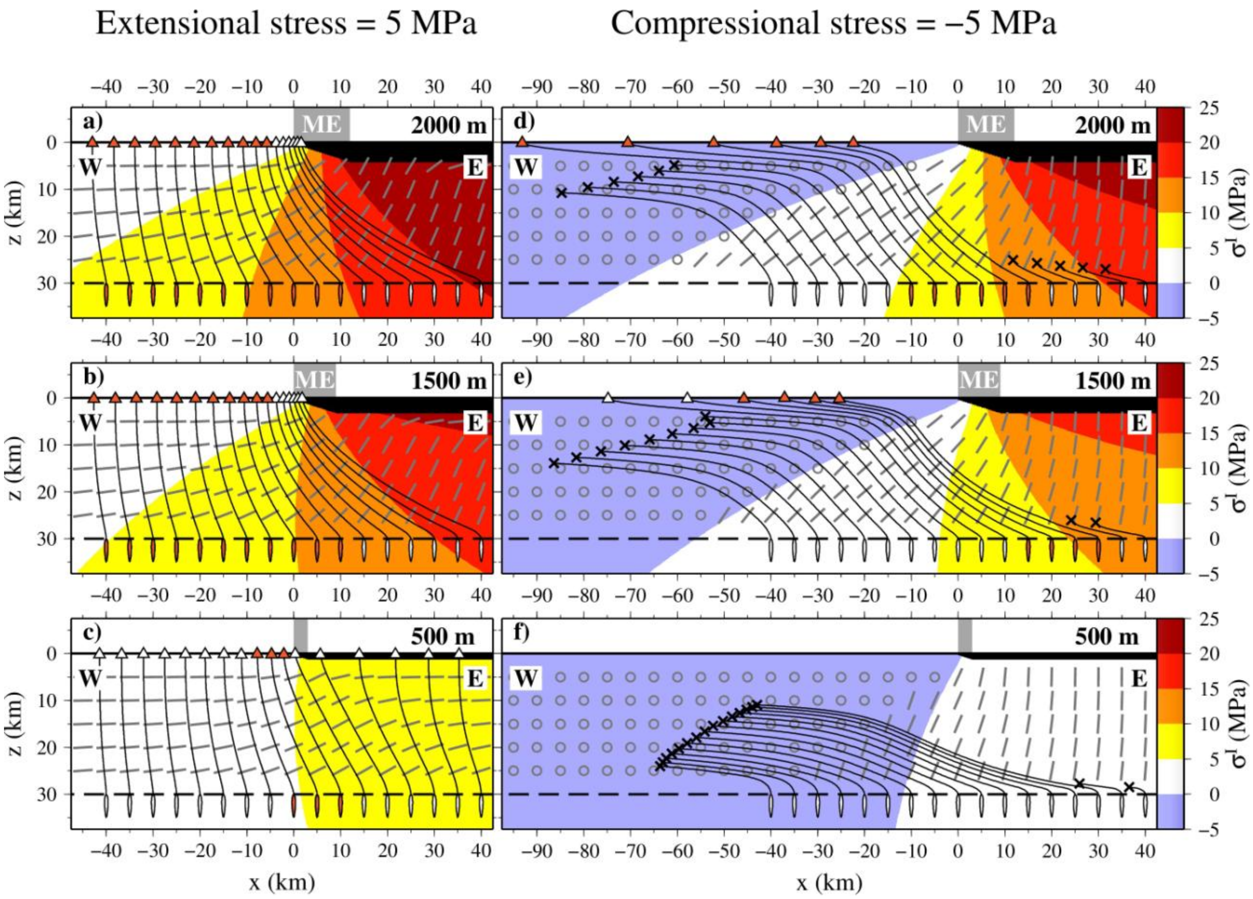

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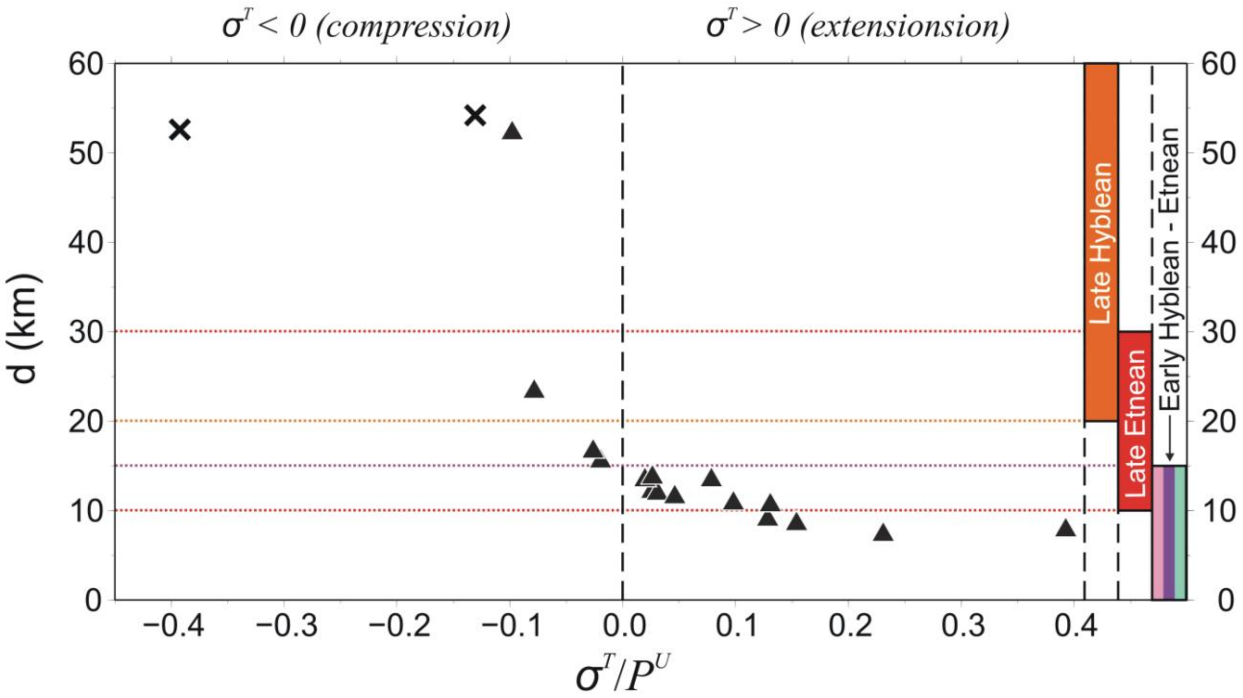

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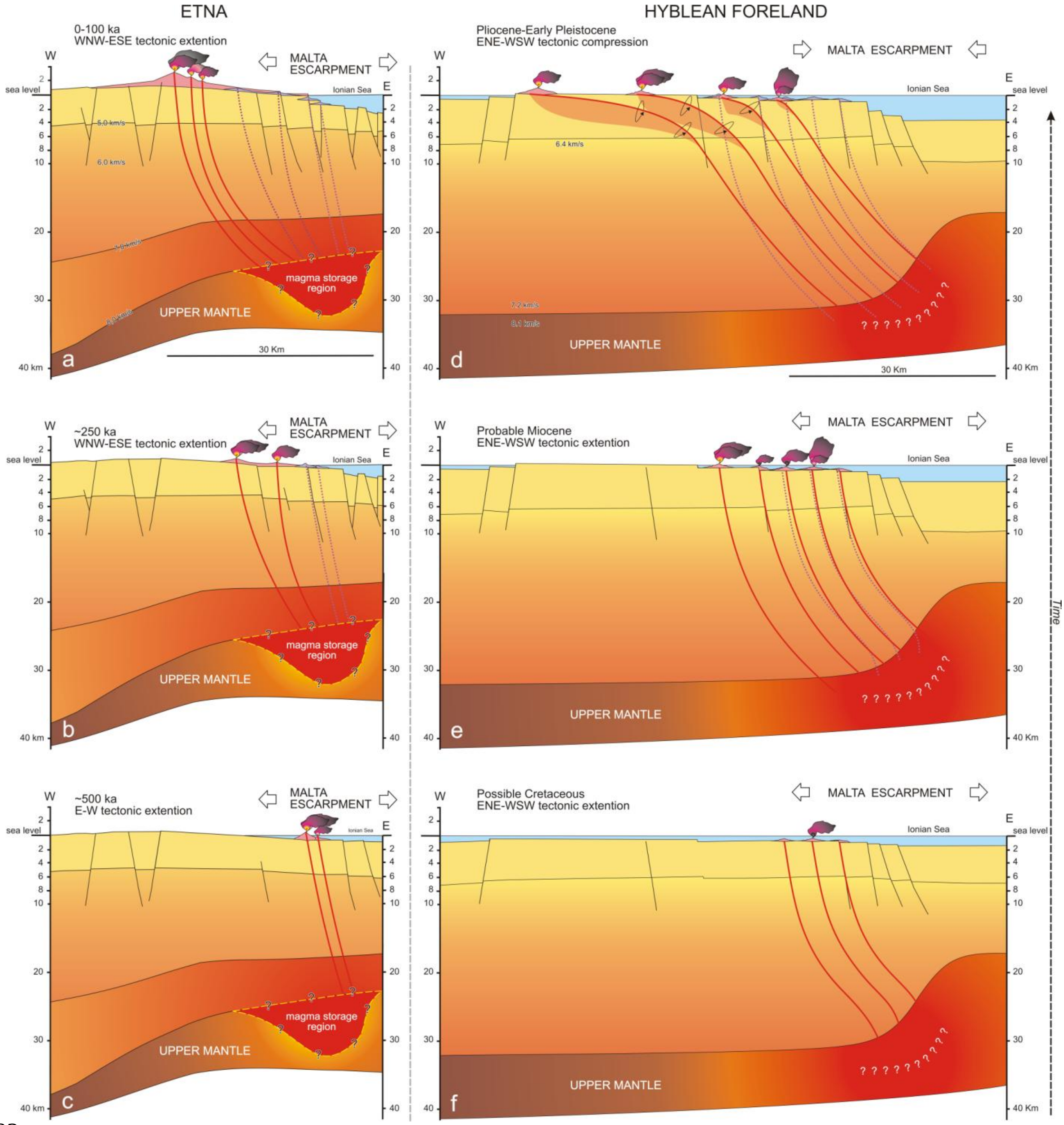

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