

The Methana Volcano – Geothermal Resource, Greece, and its relationship to regional tectonics

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ABSTRACT

Geophysical methods of analysis were applied, in order to investigate the deep structure and the geothermal potential of the Methana Volcano (NE Peloponnese, Greece). The study is based on a re-evaluation and reinterpretation of legacy magnetotelluric (MT) data with modern analysis methods, as well as 3-D inversion of aeromagnetic data constrained by in situ measurements of magnetic susceptibility. Magmatic systems are located in regions of active tectonic processes that often play a controlling role. The MT method is effective in delineating low resistivity functional elements of volcanic systems, such as magma chambers, vents, thermal fluid reservoirs and thermal fluid circulation conduits, the latter two of which are typically associated with active faults. The aeromagnetic data can assist in mapping the configuration, hence emplacement modes of volcanic rocks at depth. Accordingly, the joint interpretation of these lines of evidence, together with structural and geochemical information, is expected to allow insight into the influence of contemporary tectonics on the inception and evolution of the volcano. The contemporary stress field is mainly extensional, NNE-SSW oriented and overall homogeneous; in the area of Methana it allows for the formation of WNW-ESE north-easterly dipping normal faults, W-E faults consistent with the synthetic (dextral) R-shear direction of Riedel's shear theory and NW-SE faults consistent with the antithetic (sinistral) R'-shear direction; all such features have been mapped on Methana Peninsula. The magnetotelluric data imaged a significant geothermal reservoir developing around an intersection of the three active fault zones (normal, R and R') at depths of 1–1.5 km below the centre of the peninsula, as well as elongate *epiphenomenal* conductivity anomalies associated with the circulation of thermal fluids along all three fault zones. The 3-D magnetic susceptibility model strongly suggests that the intrusion and emplacement of magmas were guided by the same active fault zones, with particular reference to the R and R' shears whose influence is imprinted on the configuration of volcanic rocks at depth. The joint interpretation of all lines of evidence indicates that magmatism and volcanism at Methana are almost completely controlled by tectonic activity in a manner analogous to the situation of the large Santorini Volcanic Complex. It also indicates that the reservoir is replenished through the weak permeable zone created by the intersection of the R and R' shears, which is very probably collocated with the main vent of intrusive magmatic activity and may connect with a shallow magma chamber at depths greater than 4.5 km. The apparently common origin and similarities/differences in the circulation paths of thermal fluids may amply explain both the individual characteristics and similarities/differences in the chemical composition of thermal spring discharges, which have been reported by hitherto geochemical investigations.

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1. Introduction

The small Methana Volcano (MV) is located at the northeastern coast of the Argolis Peninsula and within the geologically complex area of north-east Peloponnese (Argolid and Corinthia, [Figs. 1 and 2](#)). The geotectonic setting of the broader area is part of the Corinth Rift (CR) system and bounded by the W-E oriented Gulf of Corinth (GoC) to the north (i.e. the rapidly spreading core of CR), the NW-SE oriented

extensional subsystem (graben) of the Argolic Gulf and Argolic Plain to the west, the complex extensional system of the Saronic Gulf to the east and the Myrtoon Sea to the south. In addition, it is underlain by the Hellenic Subduction System and therefrom straddled by the NW-SE oriented Hellenic Volcanic Arc (HVA), a.k.a. South Aegean Volcanic Arc, which locally includes the Methana volcanic terrain, as well as the local-scale Sousaki volcanics and solfatara at Crommyonia near the southern margin of the GoC ([Figs. 1 and 2](#)). The MV, together with the Crommyonian, Aegina and Milos volcanic fields, comprise the western (older) group of the HVA where typical calc-alkaline, arc-related volcanic rocks of Plio-Pleistocene age predominate ([Pe-Piper and Piper,](#)

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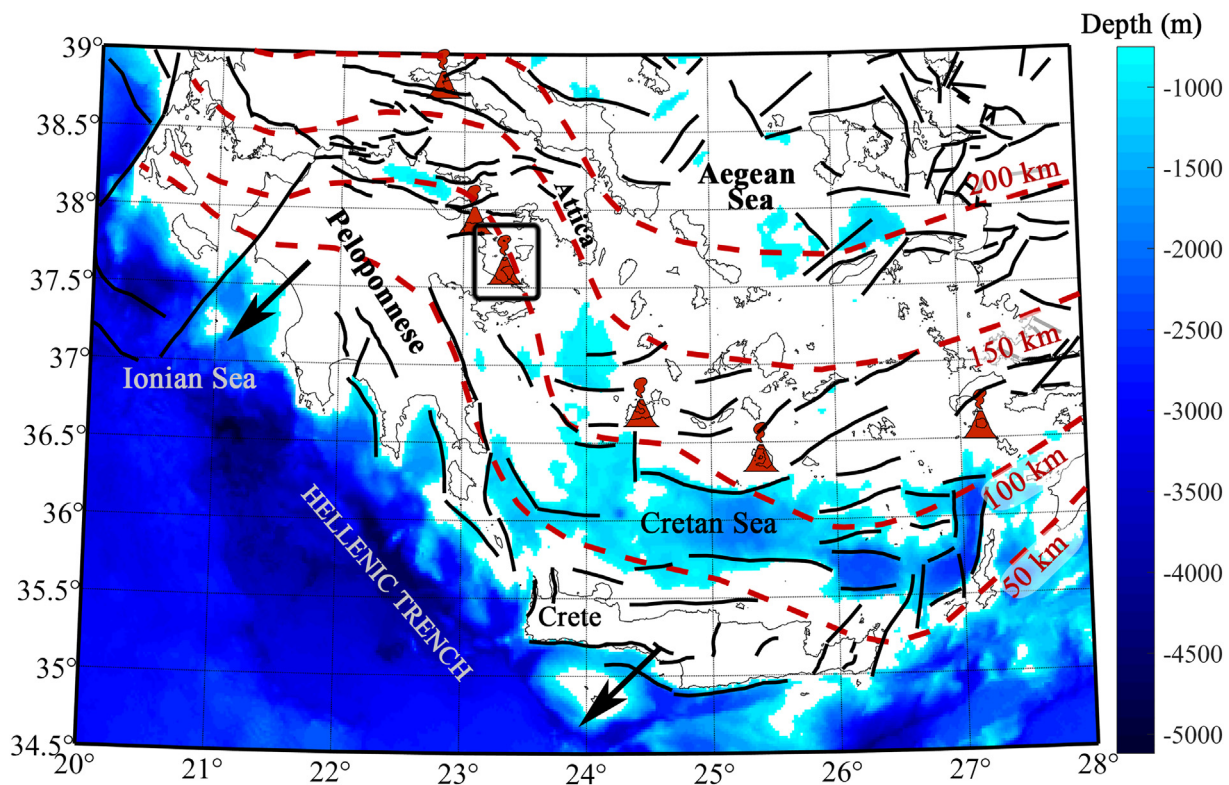


Fig. 1. Location of the Methana Volcano (rounded rectangle) in the Hellenic Subduction System. Volcanic fields are indicated with “smoking volcano” symbols. Black arrows indicate the motion of the Aegean plate relative to the African with respect to Eupore. Red dashed lines indicate the 50, 100, 150 and 200 km isodepths of the subducting slab. Black solid lines indicate main faults. Both data sets were extracted from the SHARE database (Basili et al., 2013). Bathymetry was extracted from the ETOPO1 database (Amante and Eakins, 2009). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

2005; D'Alessandro et al., 2008). The association of the geodynamic setting, regional tectonics and Quaternary plutonic magmatism at NE Peloponnese has been recently examined by Tzanis et al. (2018a), on the basis of seismological, seismotectonic, aeromagnetic and magnetotelluric data.

Although the MV has been subject to continuing academic research (Section 2), the geothermal resource has not attracted considerable attention. Only two significant attempts to explore its potential are hitherto known, both sponsored by the Government of Greece. The first was a geochemical evaluation of the resource by *Geotermica Italiana* (1984 – see Section 2.2). The second was a magnetotelluric survey conducted in 1992 (Section 3), funded by the General Secretariat of Research and Technology Project 90ΠΣ5 and sponsored by the Renewable Energy Directorate of the Public Power Corporation of Greece. The magnetotelluric work reported herein has been part of the latter endeavour (Section 3) and resulted in one- and two-dimensional quantitative interpretations of the measurements, respectively presented by Tzanis and Lagios (1994) and Volti (1999). The previous magnetotelluric work was conducted with methods and techniques that were state-of-the-art during that time frame. The present work reports a qualitative and quantitative re-examination of the magnetotelluric data using improved processing methods, novel hypothetical event analysis techniques to study the spatial configuration of the telluric field (modes of induction), and advanced two-dimensional inversion tools.

A second line of evidence presented herein, is based on three-dimensional quantitative interpretation of aeromagnetic anomaly data using the Li and Oldenburg (1996, 2003, 2010) algorithm. The aeromagnetic data set has been extracted from the uniform and homogeneous aeromagnetic anomaly map of Greece (Chailas et al., 2010; Chailas and Tzanis, 2019) and is part of the data used by Tzanis et al. (2018a). The inversion was performed on upward-continued data so as to yield

a clear image of the intermediate and large scale features of the volcano's interior (Section 4.3).

The Magnetotelluric (MT) method is one of the primary techniques applied to the prospection of volcanic and geothermal systems. Recent reviews can be found in Muñoz (2014), Uchida (2016) and Patro (2017). The method has also been applied to the exploration of all significant volcanoes and geothermal resources of Greece, as in Milos (Galanopoulos et al., 1991), Kos (Lagios et al., 1994; Lagios et al., 1998), Nisyros (Tzanis et al., 2018b), Methana (Tzanis and Lagios, 1994; Volti, 1999) and Santorini (Tzanis et al., 2020). MT is particularly efficient in targeting the low resistivity functional elements of volcanoes and geothermal systems: magma chambers and intrusion vents and thermal fluid reservoirs and circulation conduits. Moreover, owing to the development of *epiphenomenal* electrical conductivity anomalies in response to faulting, the method is suitable for imaging tectonic processes. Faulting generates permeable rock, both within the fault zone (fault gouge, breccia and mylonite) and around it as a result of repeated cycles of elastoplastic deformation (damage). The hydraulic permeability resulting from micro- and mesoscale fracturing is generally aligned with the fault and the more extensive the damage, the higher the fluid content (liquid fraction) and the higher the conductivity. The presence of fluids in the fault zone is very important in tectonic processes as it influences creep and/or stability. In convective hydrothermal systems controlled by concurrent tectonic activity, the circulation of thermal fluids usually takes place along active faults (e.g. Tzanis and Makropoulos, 1999; Tzanis et al., 2020). In the latter case, the electrical conductivity does not only depend on the volume and salinity of the liquid (thermal fluid) fraction, but also on the temperature and the presence of clay minerals that generally increases it by orders of magnitude. The formation of clay minerals is a result of hydrothermal alteration. Consider also that insofar as magmas find it easier to intrude through the weak zones generated by faulting and fault-fault

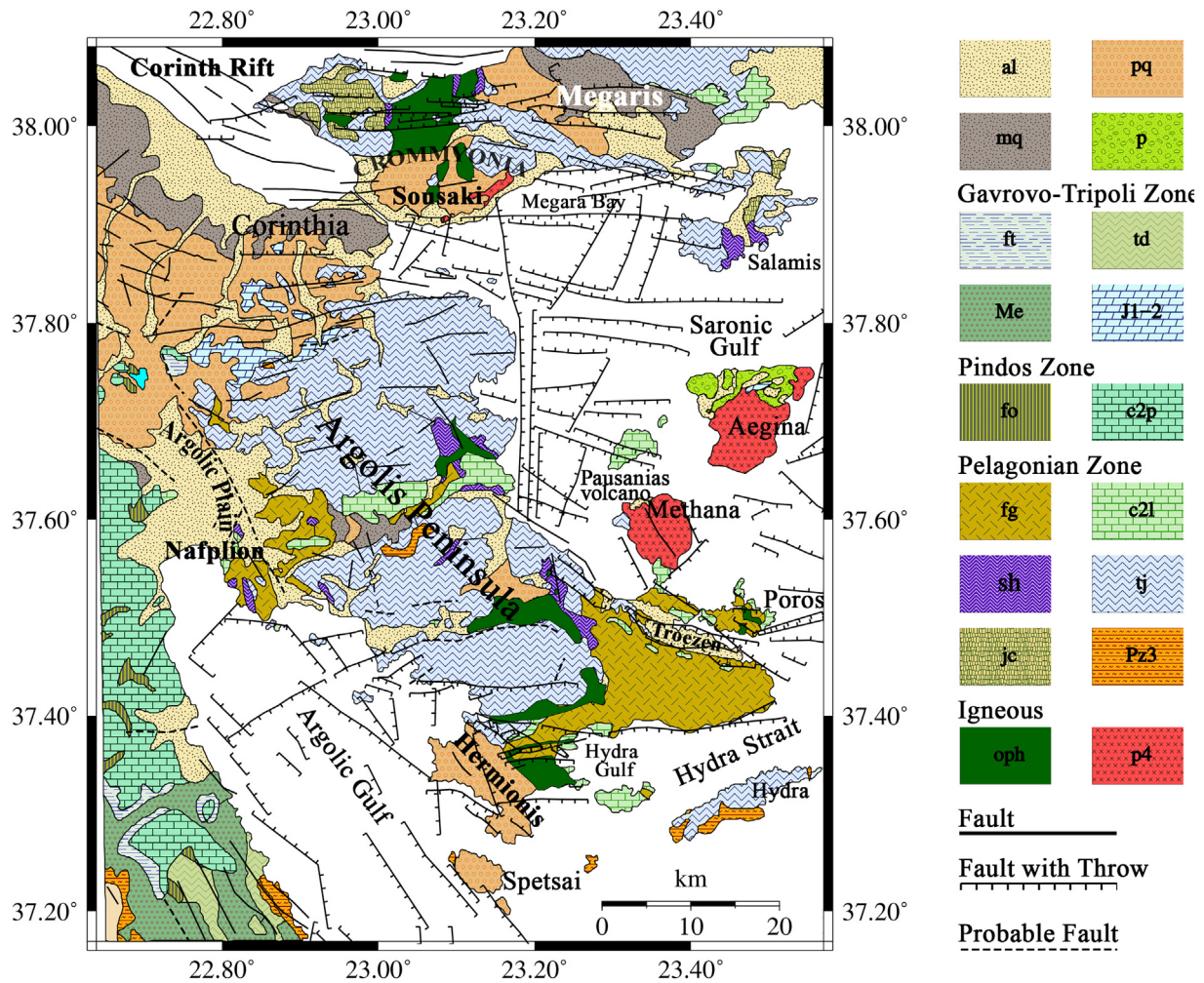


Fig. 2. Geological and location map of the study area. Codenames of geological formations are as follows: *Cainozoic*: **al**, alluvial deposits; **pq**, Plio-Pleistocene lacustrine and marine sediments; **mq**, Pleistocene lacustrine and terrestrial sediments; **p**, Pliocene marine sediments. *Tripoli and Parnassus zone*: **ft**, flysch; **td**, Middle-Late Triassic dolomites; **Me**, Mesozoic–Eocene limestones and dolomites undivided; **J1–2**, Early-Middle Jurassic limestones. *Pindos zone*: **fo**, flysch; **c2p**, Late Cretaceous limestones. *Pelagonian zone*: **fg**, flysch; **c2l**, Late Cretaceous limestones; **sh**, shale-chert formation (ophiolitic-sedimentary mélange); **tj**, Triassic – Jurassic limestones (Didyma–Trapezina Formation); **jc**, Boeotian flysch; **Pz3**, various Late Palaeozoic and Permian–Triassic formations, unclassified. *Igneous*: **oph**, generic classification of ophiolitic rocks (generally harzburgites, serpentinites, gabbros, amphibolites, andesitic lavas in schistose serpentine matrix); **p4**, generic classification of Quaternary calc-alkaline volcanic rocks (andesites to dacites). Fault data collected from Skourtsos and Kranis (2009), Papanikolaou et al. (1988), Vassilopoulou (1999) and Stefatos et al. (2002).

interactions, the influence of tectonics on the development of the volcano can be imprinted in the configuration of volcanic rocks at depth. Moreover, hydrothermal alteration not only will increase electrical conductivity by forming clay minerals in the reservoir and circulation conduits, but will also depress magnetic susceptibility by bleaching iron from ferrous minerals. Thus, the joint evaluation of resistivity and susceptibility models can yield information as to the location, size and configuration of reservoirs and circulation conduits, as well as insight into the origin (localization/inception) and evolution of the Methana Volcano.

In many respects, this paper can be viewed as a continuation of the work by Tzanis et al. (2018a). Those authors have shown that the large-scale Quaternary plutonic magmatism of NE Peloponnese is controlled by regional tectonics and takes place along the volcanic arc, along the axes of major extensional tectonic structures (Argolic Gulf and Plain, Hydra Strait) and along local scale *E–W* transcurrent faults straddling the Argolis Peninsula. Herein we focus on the Quaternary MV and show that the same principles and conditions appear to apply on a local scale and shallower depths: the inception and evolution of MV and geothermal resource appear to be guided by the same tectonic activity that controls the plutonic magmatism of the broader area. The

paper can also be viewed in the context of a more general topical research interest to study magmatic systems and the control exerted by crustal structures and tectonics on crustal magmatism and volcanism. Examples are the hectokilometric-scale study by Comeau et al. (2018) and the decakilometric-scale studies of inland volcanoes by Hill et al. (2015) and Bedrosian et al. (2018). Our work focuses on kilometric-scale, shallow (<3 km) processes in a small insular volcano, situated in one of the most rapidly deforming areas on Earth. In this respect, it is analogous to the analysis of active tectonics in the Santorini Volcanic Complex presented by Tzanis et al. (2020).

2. Geological setting, volcanism and the geothermal resource

2.1. Geodynamic setting

A defining deep dynamic feature and genitor of the Hellenic Volcanic Arc (HVA), is the active subduction of the African oceanic crust beneath the Aegean plate Hellenic Subduction System (HSS). The MV is located above the rapidly subducting segment of the HSS, in which the rate of convergence is estimated at 40 mm/a (Clarke et al., 1998; McClusky et al., 2000). In the vicinity of the study area, the slab is NNW oriented

and plunges to the E-NE. The subducting African plate plunges at an angle of approximately 22° between the depths of 20 km and 50 km and then at an angle of approximately 40° to a depth of 150–160 km below Attica (Hatzfeld et al., 1989; Basili et al., 2013; Sachpazi et al., 2016; Tzanis et al., 2018a). The inflection of the slab is approximately collocated with the 60 km isodepth and its geometry is consistent with the location and orientation of the Argolic Gulf and Plain (Fig. 2), along the axis of which Tzanis et al. (2018a) have found evidence of intrusive plutonic magmatism. The 100 km isodepth is located at a zone consistent with the geographical location of the HVA and directly below Methana Volcano at the NE coast of the Argolid. This is compatible with the depths at which eclogite facies metamorphism drives water into the mantle wedge and initiates the generation of andesitic magmas and arc volcanism (e.g. Grove et al., 2006 and references therein). Finally, Sachpazi et al. (2016) detect segmentation in the slab, caused by down-dip faults in a manner suggestive of mechanical coupling with the overriding (Aegean) plate.

The intensity and distribution of crustal magmatism in the broader study area depends on the thickness of the overriding Aegean plate. According to Drakatos et al. (2005), extensional thinning and the emergence of mantle material limits this thickness to about 20 km in the area of the Saronic Gulf. More recently, Sachpazi et al. (2007), Suckale et al. (2009) and others, detected the Aegean plate Moho at approx. 37 km beneath the west Argolic Gulf and at 30 km beneath the western Saronic gulf. Both lines of evidence suggest significant thinning of the Aegean crust eastward of an important tectonic boundary defined by the Argolic Gulf and Plain. In agreement with these observations, the deep magnetotelluric profile of Galanopoulos et al. (2005) has detected a rising column of relatively conductive material just north of the head of the Argolic Gulf, which penetrates the base of the Aegean crust and spreads laterally beneath the Peloponnese. Tzanis et al. (2018a) attribute this feature to partial melting in the upper mantle and subducting slab, which may be related to the bending of the slab and the thinning of the Aegean crust. Accordingly, the inception and evolution of the Argolic Gulf may be related to deep-rooted processes. Analogous but broader columns of conductive material appear near the ceiling of the subducting slab at a depth of approx. 100 km, and through the mantle wedge rise to the base of the crust beneath Crommyonia (Fig. 2): this is a direct manifestation of partial melting associated with the HVA.

2.2. Regional and local tectonics

As mentioned in the Introduction, the study area is part of the broader Corinth Rift (CR), which is a young graben developing in a compressional regime of continental collision and manifesting high complexity and increased seismicity and deformation rates. The most active contemporary feature of the CR is the W-E oriented Gulf of Corinth (GoC), a composite asymmetric graben with varying geometry along strike (Stefatos et al., 2002; Sachpazi et al., 2003; McNeill et al., 2005). Extensive geological studies (e.g. Doutsos and Piper, 1990; De Martini et al., 2004; Flotté et al., 2005; McNeill et al., 2005), earthquake focal mechanisms (e.g. Rigo et al., 1996) and geodetic investigations (e.g. Billiris et al., 1991; Briole et al., 2000; Avallone et al., 2004), have shown that the GoC experiences N-S extension of 1.0–1.6 cm/a, with higher rates observed at the west and lower at the centre of the graben. Skourtsos and Kranis (2009) found extensional structures with geometrical and kinematic characteristics analogous to GoC as far south as the Mt Mainalon (approx. 37.7°N), 40–50 km west of our study area.

The Argolid and the Argolic and Saronic gulfs are associated with the CR. The Argolis Peninsula is delimited by the Argolic and Saronic gulfs to the west and east respectively (Fig. 2). Papanikolaou et al. (1988) consider the Argolic to be an NW-SE oriented symmetric graben, which during the Pliocene–Early Pleistocene formed a channel that connected the GoC with the south Aegean Sea and separated the peninsula from the rest of the Peloponnese. Van Andel et al. (1993) consider the Argolic Gulf to be a half-graben whose eastern marginal faults are a brittle

response to downwarping. They also assess that along its western margin the Holocene subsidence rates increase from 50 cm/ka near Argos to 100 cm/ka near the mouth of the gulf and along its eastern margin from zero near Nafplion to 150 cm/ka near Spetsai Island. Drakatos et al. (2005) provide seismological evidence that the Saronic Gulf is divided into two basins by a central platform that may comprise the extension of a large thrust belt which dominates the adjacent onshore areas of Attica. The western basin exhibits higher seismic activity. It is further distinguished into a northern and southern blocks by a well-defined W-E fault zone which is located between Aegina and Salamis islands, extends westwards into the Peloponnese and may be part of the Corinth Rift system.

Papanikolaou et al. (1988) and Papanikolaou and Lozios (1990) present evidence that the orientation of normal faulting changes in two discrete steps, from W-E in the GoC to WNW-ESE in the Saronic (on average $\text{N}110^\circ$) to and NW-SE at the Cyclades in the Aegean Sea. This indicates rotation of the stress field. Tzanis et al. (2018a) confirmed this by formal inversion of all theretofore known digital-era earthquake focal mechanisms. As illustrated in Fig. 3, the principal compression axis σ_1 is oriented at $\text{N}64^\circ$ and plunges at 77° (open square), while the principal extensional axis σ_3 is oriented at $\text{N}210^\circ$ and plunges at 10° (open circle). The stress field is mainly extensional, NNE-SSW oriented and overall homogeneous. The “typical” focal mechanism generated by this field is shown with thick black lines and a schematic depiction of the faulting pattern predicted by Riedel's shear theory is superimposed. The expected principal direction of normal faulting is WNW-ESE (approx. $\text{N}290^\circ$) for north-easterly dipping faults and NW-SE (approx. $\text{N}130^\circ$) for south-westerly dipping faults. The former is evident in the hanging cliffs of the north-eastern coast of Argolis peninsula, as well as in normal faults mapped by Papanikolaou et al. (1988) in the Saronic Gulf and by Dietrich and Gaitanakis (1995) on Methana peninsula. It is also congruent with a similarly oriented lineament of earthquake foci between Methana peninsula and Aegina Island detected by Drakatos et al. (2005) and collocated with faulting structures mapped by Papanikolaou et al. (1988). The latter orientation is observed mainly offshore, along the northern margin of the Argolic, between Spetsai Island and the head of the Gulf (Papanikolaou et al., 1988; Van Andel et al., 1993).

Now, consider that NW-SE normal faults are mainly observed along the coasts of Argolis peninsula. Within the peninsula, the orientation of major terrain features and morphological discontinuities is W-E and the orientation of neotectonic faulting zones is mainly WNW- ESE to W-E (Vassilopoulou, 1999, 2010). All these form a series of local horsts and grabens with particular reference to Mt Adheres and Hermionis Peninsula; they are most active at the south and SE, where they contribute to the formation of the tectonic basin of Hydra Strait and Gulf, as well as toward the Argolic Gulf (Fig. 1). The contemporary stress field determined by Tzanis et al. (2018a) predicts the existence of W-E faults consistent with the synthetic (dextral) R-shear direction of Riedel's shear theory and NW-SE (approx. $\text{N}330^\circ$) faults consistent with the antithetic (sinistral) R'-shear direction. The configuration of the stress field allows for a small ENE-WSW compressional component which may easily account for local-scale faulting with significant lateral heave. Indeed, active, oblique-normal (transtensional) faulting consistent with the expected R and R' directions is observed in a few focal mechanisms of small earthquakes at the southern Argolid and the Argolic Gulf (Tzanis et al., 2018a). Accordingly, the W-E- faults mapped by Dietrich and Gaitanakis (1995) on Methana and inferred by Drakatos et al. (2005) to exist between Aegina and Salamis islands, should correspond to the R-shear. Tzanis et al. (2018a) proposed a geotectonic model of Argolis Peninsula, according to which the strain differential due to the disparate extensional trends of the Argolic and Saronic gulfs is accommodated by block motion associated with the R-shear and igneous intrusive activity along major block boundaries. Moreover, the approx. $\text{N}340^\circ$ escarpments of segmented offshore faulting observed by Papanikolaou et al. (1988), with particular reference to the vicinity of Methana Peninsula,

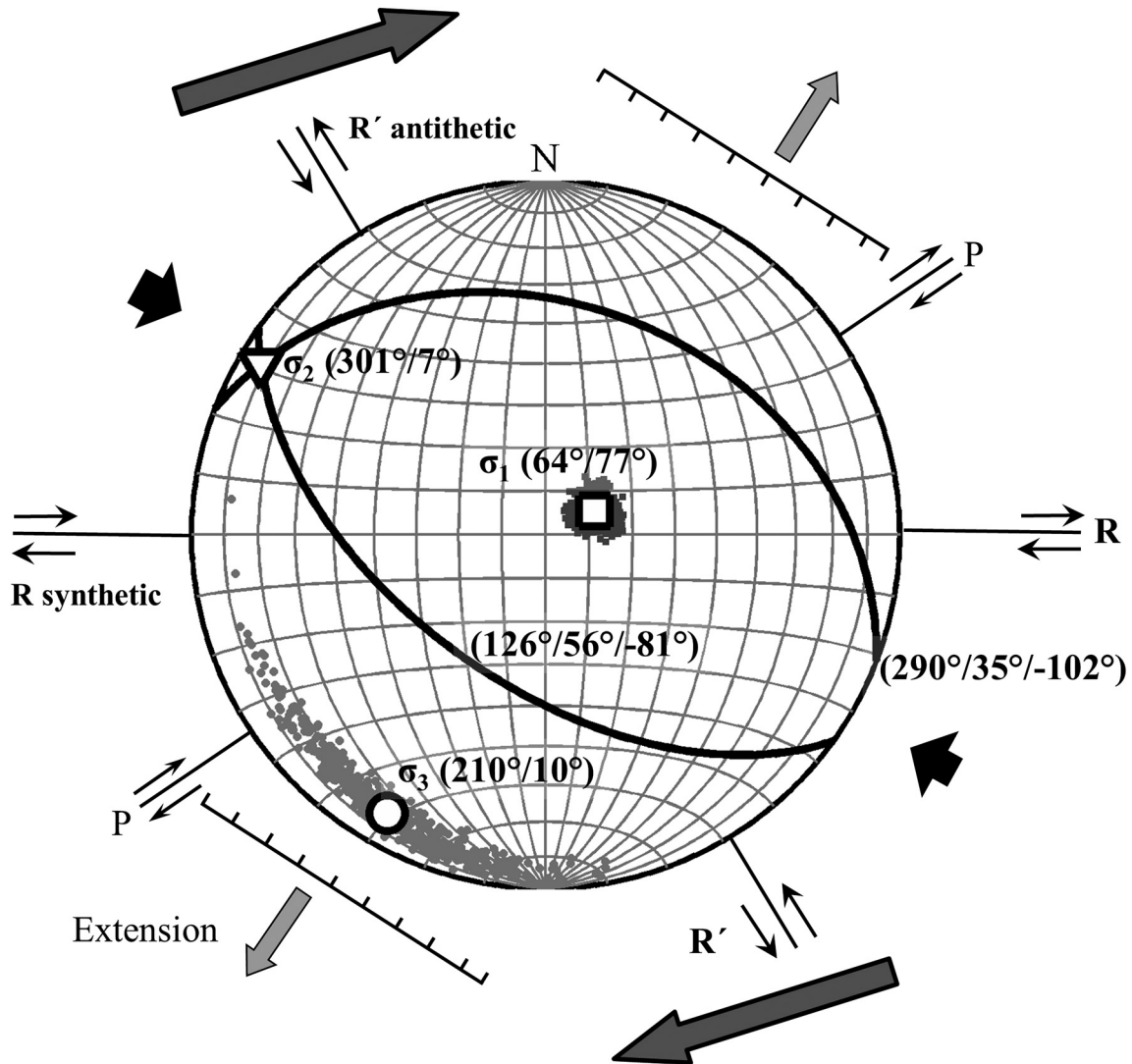


Fig. 3. Schmidt projection of the stress field of the broader Saronic Gulf, computed with formal inversion of digital-era earthquake focal mechanisms. Grey squares and circles indicate the expanse of the 95% confidence area for σ_1 and σ_3 respectively. The “typical” expected faulting mechanisms and the schematics of the faulting pattern predicted by Riedel shear theory are also projected.

Figure adapted from Tzanis et al. (2018a).

could comfortably be explained as expressions of the R' shear. Finally, they noted that this stress configuration requires the existence of crustal-scale WSW-ENE ($N70^\circ$ – $N80^\circ$) regional shearing deformation which was assumed to be accommodated in the regional-scale SW-ward translation of the Aegean plate.

2.3. Quaternary volcanism and the geothermal resource

The MV (Fig. 4) consists of 32 domes, calc-alkaline to alkaline in composition. Potassium-Argon dating from the older part of the complex (Stavrolongos, Chionessa, Chelona, Malisa) indicate ages between 0.9 and 0.5 Ma, although activity may have begun during the late Pliocene (Dietrich and Gaitanakis, 1995; Pe-Piper and Piper, 2002, 2013; Smet, 2014). A younger phase of activity took place between 0.035 and 0.025 Ma, forming a series of lava domes and flows not only at the centre of the peninsula, but also at its eastern (Kossona, Tsonaka and Kypseli) and western (Agios Andreas, Kammeni Chora, Makrylongos, Palea Loutra) halves, with the latter being the most recent. Thus, the pyroclastic flows of Agios Andreas are dated to approximately 200 BCE and the andesitic domes of Palea Loutra and Makrylongos to 0.3–0.1 Ma, while their dacitic base extends to the

Pleio-Pleistocene boundary. The youngest dome is Kammeni Chora, located on the NW of Methana Peninsula. This eruption took place circa 238 BCE and effused lava that flowed up to 500 m north of the coastline. It was mentioned by Pausanias (who also described the post-eruptive thermal activity), Strabo and Ovid.

Peripheral to the volcano magmatic activity appears both on- and offshore. Evidence of onshore activity is found in Poros Island, in the form of a small (ca. 1 km²) andesitic outcrop with a composition similar to that of the Aegina volcanics (Mitropoulos, 1987; Francalanci et al., 2005). Evidence of offshore activity was found a few kilometres to the NW of Kammeni Chora, where andesitic lavas have been dredged, with an age of approx. 1 Ma estimated by the thickness of the sedimentary cover (Papanikolaou et al., 1988). This is the site of the “Pausanias” submarine volcano and locus of a minor eruption that took place in 1700 CE (Foutrakis and Anastakis, 2018).

The igneous rocks, debris and breccia that almost completely cover the peninsula, are mostly dacitic-andesitic in composition. Small occurrences of Quaternary alluvial deposits are also present, while a Mesozoic basement of sedimentary rocks (limestones) outcrops only at the north-west and south of the peninsula. The Argolid mainland, to the south of Methana Peninsula, is almost completely covered by flysch; a small

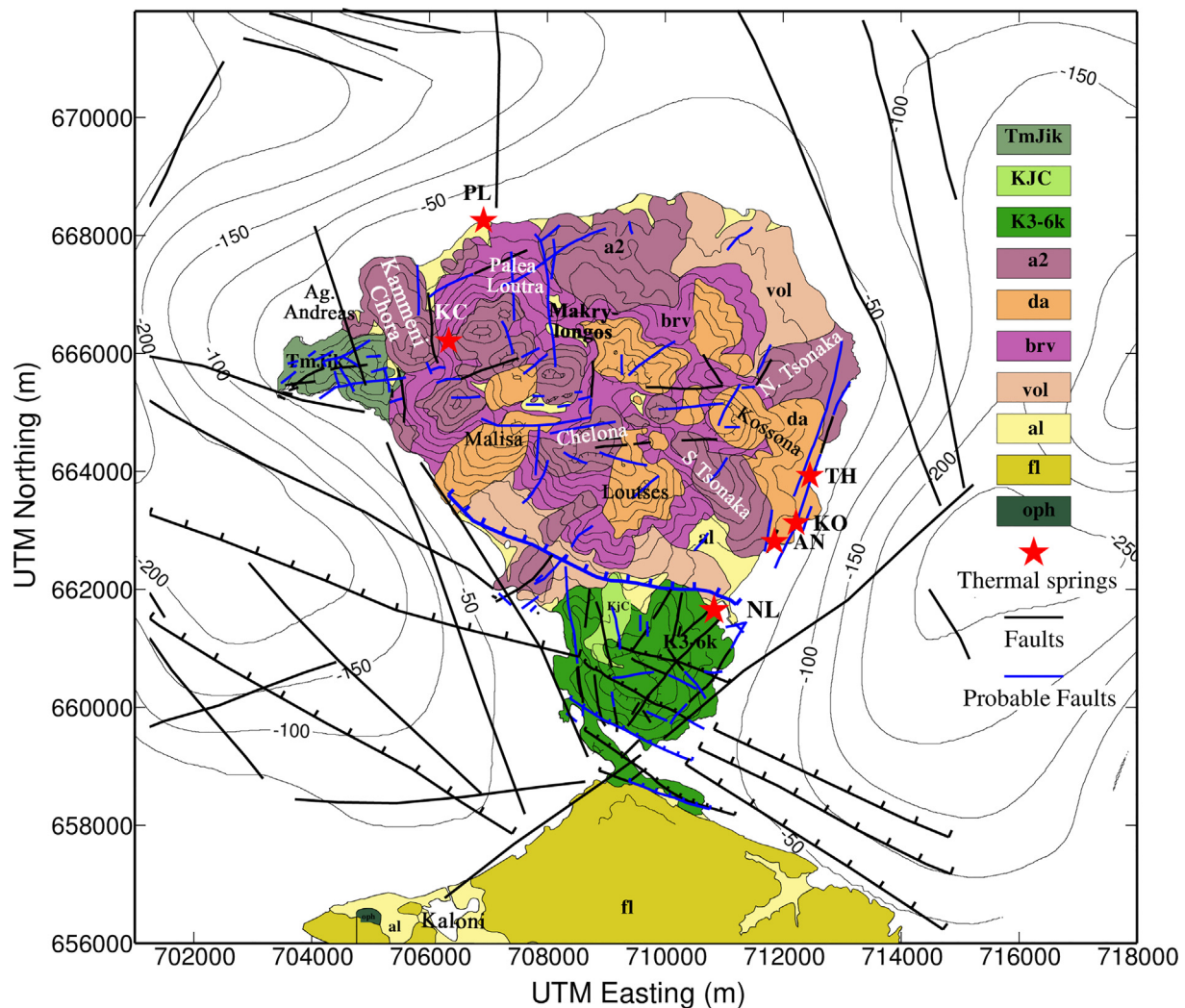


Fig. 4. Map of the Methana Peninsula illustrating geological formations, faulting, topography, bathymetry and thermal springs. Geological formations are coded as follows: **al**, alluvial deposits; **a2**, andesites/andesitic dacites; **da**, dacites; **brv**, undifferentiated volcanic rocks, volcanoclastic flows/breccia, mostly andesitic and dacitic; **vol**, volcanic debris; **TmJik**, Triassic-Lower Jurassic limestone; **KJC**, Jurassic-Cretaceous limestone; **K3-6k**, Cretaceous limestone; **fl**, flysch; **oph**, ophiolites. Thermal springs and fumaroles: **NL**, Nea Loutra; **PL**, Palaia Loutra; **AN**, Agios Nikolaos; **KO**, Kokkinopetra; **TH**, Thiafi; **KC**, Kammeni Chora.

occurrence of heavily altered ultrabasic formations can also be observed at the area of Kaloni.

The relationship between tectonics and volcanism at MV, or more precisely how the geochemistry and the eruptive style of volcanic products is influenced by the regional tectonics, has been examined in a remarkable paper by Pe-Piper and Piper (2013). These authors refined the volcanic stratigraphy by using radiometric dating, litho-geochemistry and field observations that included recording deformational structures and enclave abundance and distinguished eight phases in the evolution of MV. In Phase A small N-S-striking Pliocene domes and a central volcano of uncertain type developed, that was related to magma transport along N-S-striking listric faults associated with Pliocene E-W regional extension. These edifices eroded to produce a widespread volcanoclastic apron in Phase B. In Phase C an explosive central volcano with flank eruptions of basaltic andesite developed in the early Quaternary. These eruptions were (presumably) controlled by crustal-scale strike-slip NE-SW faulting. Afterwards, E-W-striking faults, mapped from offsets of the late Pliocene volcanoclastic apron and erosional terraces, controlled the location of volcanism. Phases D-H are characterized by dacite domes and small andesitic stratovolcanoes formed throughout the mid and late Quaternary. The volcanism of Phases F and G was voluminous and synchronous with the onset of steep normal faulting in the Gulf of

Corinth. Overall, the authors concluded that changes in regional fault patterns, as in Phases C, F and G, have the greatest influence on the eruptive style and abundance of volcanic products. As will be shown below, the inferences made by Pe-Piper and Piper (2013) are remarkably accurate for their methodology and approach, but simultaneously limited by the (at that time) incomplete understanding of the stress field and permissible faulting patterns, as well as the absence of images of faulting structures in the volcano's interior. Accordingly, these topics will be revisited and discussed in Section 5 (Synthesis and discussion).

Post-eruptive activity is manifested with thermal springs along the coastline: the Nea Loutra (NL) and Agios Nikolaos (AN) springs are located on the east coast, while the Palaia Loutra (PL) or Pausanias spring on the north coast. Recent, albeit presently ceased, hydrothermal activity has been found at the area of Thiafi Bay (D'Alessandro et al., 2008) and Kokkinopetra (Dietrich and Gaitanakis, 1995), both located at the east coast of the peninsula. Finally, we found and documented evidence of recent but defunct fumaroles at Kammeni Chora, right at the southern boundary of the 258 BCE activity.

The water temperatures of the thermal springs have been recorded since the mid-19th century and have generally not exceeded 40 °C (D'Alessandro et al., 2008). Specifically, at the Nea Loutra spring (NL) the temperature was 27.0–34.4 °C in the period 1835–1975, at the

Palea Lourta (PL) spring 32.8–38.8 °C in the period 1835–1931 and at the AN spring reached 41.2 °C in 1922 and up to 40 °C in the early '70s. For the period 2004–2007, the temperatures observed by D'Alessandro et al. (2008) did not exceed 38 °C and were consistent with those determined by Geotermica Italiana (1984).

All geothermal manifestations are controlled by local active faulting. According to Dotsika et al. (2010) and Dietrich and Gaitanakis (1995), the NL and AN springs are controlled by an NNE-SSW fault system while the PL spring is controlled by a WNW-ESE fault system. On the east coast, the neighbouring NL and AN springs would be expected to be chemically similar and therefore controlled by the same faults. Nonetheless, chemical and isotopic compositions are correlated between PL and NL; this suggests that these springs may be linked with a more complex than anticipated active fault network. The existence and approximate location of an NNE-SSW fault linking the two springs was discussed by Tzanis and Lagios (1994).

The interpretation of the geochemical analysis of fluids from the hot springs of Methana town is hindered, or even misguided by the extensive mixing of geothermal fluids with seawater (Geotermica Italiana, 1984). Some results, however, indicated that these fluids may have had an initial temperature of 100–120 °C, provided that before discharging they circulated in a substratum of carbonate rocks, or were stored in a reservoir within carbonate rocks. The same source questioned the existence of a geothermal reservoir but did not preclude the concentration of hot fluids in faulting structures at depth. The presence of saline fluid concentration beneath the central volcanic domes and at a depth of 1.5–3 km was observed by geophysical (Magnetotelluric) sounding and was interpreted to be a geothermal reservoir (Tzanis and Lagios, 1994; Volti, 1999). Later independent studies by gas geothermometry, however, have estimated temperatures of the order 227–274 °C for the reservoir (Nicholson, 1993; Goff and Janik, 1999). D'Alessandro et al. (2008) pointed out that these may be overestimated due to the loss of CO₂ through the interaction of the gases with shallow aquifers and evaluated the reservoir temperature to 200–220 °C.

3. The magnetotelluric survey

The magnetotelluric (MT) data were collected in the autumn of 1992, in cooperation with the Department of Geophysics of the University of Edinburgh. Measurements were carried out in the nominal bandwidth 128 Hz–40s with Pb/PbCl₂ electrodes, CM11E induction coils and the MkIIb model of the Short Period Automatic Magnetotelluric (SPAM), developed in the U. of Edinburgh by Dawes (1984). Given that SPAM enabled simultaneous two-station data acquisition, the Telluric – Magnetotelluric field procedure (Hermance and Thayer, 1975) was implemented, for which information is provided in Section S1 of the Supplementary Information. The data were acquired using a 5-component Magnetotelluric configuration at a Base site (denoted by the suffix 'b') and a 2-component Telluric configuration at the Remote site (denoted by the suffix 'r'). This enabled calculation of impedance tensors at base and satellite stations and induction vectors at the base stations. The survey comprised a total of 18 soundings, base and remote. Only 14 are used herein, the locations of which are shown in Fig. 5. Of the remaining four, three have yielded unreliable response functions due to intense anthropogenic noise and one was uninterpretable due to extreme natural distortion of the telluric field. The estimation of impedance tensors and magnetic transfer functions is outlined in Section S1 of the Supplementary Information. Examples of the quality, general properties and spatial variation of the impedance tensors and magnetic transfer functions are presented in Section S3 of the Supplementary Information. A 1-D interpretation of the same data has previously been presented by Tzanis and Lagios (1994), while an analysis of the complete data set together with 2-D inversion was reported by Volti (1999).

3.1. Spatial analysis and determination of geoelectric strike

The spatial analysis of the magnetotelluric Earth response endeavours to extract information about the configuration of the induced natural EM fields, which in turn depend on the geometry, size and configuration of lateral geoelectric inhomogeneities. Herein, the spatial analysis of impedance tensors implements the *Anti-symmetric Singular Value Decomposition* (ASVD) proposed by Tzanis (2014). The ASVD is a reformulation of the equivalent Canonical and Singular Value Decompositions of Yee and Paulson (1987) and LaTorraca et al. (1986) respectively. These are *symmetric* and apply when the electric and magnetic fields are measured in different orthogonal coordinate frames associated by a rotation of 90°. Tzanis (2014) has shown that they constitute proper rotations in 3-D space, based on the topology of the SU(2) rotation group and result in a characteristic state – characteristic value (generalized eigenstate – eigenvalue) formulation of the magnetotelluric induction problem. He has also shown that they can be reformulated into an anti-symmetric decomposition, suitable for the analysis of *practical* magnetotelluric measurements in which the electric and magnetic fields are referred to the *same* coordinate frame. An introduction to the ASVD is provided in Section S2 of the Supplementary Information and only a very brief account of its geometrical characteristics is given herein.

According to the ASVD, at any location on the surface of the Earth, the magnetotelluric induction problem can be formulated as

$$\begin{bmatrix} E_1(\theta_E, \Phi_E, \omega) \\ E_2(\theta_E, \Phi_E + \frac{\pi}{2}, \omega) \end{bmatrix} = \begin{bmatrix} 0 & \zeta_1(\omega) \\ -\zeta_2(\omega) & 0 \end{bmatrix} \cdot \begin{bmatrix} H_1(\theta_H, \Phi_H, \omega) \\ H_2(\theta_H, \Phi_H + \frac{\pi}{2}, \omega) \end{bmatrix},$$

where θ and Φ are rotation angles, $\{E_1(\theta_E, \Phi_E), H_1(\theta_H, \Phi_H)\}$ comprises the *maximum characteristic state* of the magnetotelluric field, $\{E_2(\theta_E, \Phi_E + \pi/2), H_2(\theta_H, \Phi_H + \pi/2)\}$ comprises the *minimum state*, E_1 and E_2 are the eigenvalues of the telluric field and H_1, H_2 the eigenvalues of the total magnetic field. With reference to the experimental coordinate axes $\{x, y, z\}$, the angles (θ_E, Φ_E) define a *characteristic* coordinate frame $\{x_E, y_E, z_E\}$ of the electric field such that x_E is rotated Φ_E clockwise with respect to the x -axis and the plane $\{x_E, y_E\}$ is tilted by an angle θ_E clockwise with respect to the horizontal $\{x, y\}$. Likewise, the angles (θ_H, Φ_H) define the characteristic frame $\{x_H, y_H, z_H\}$ of the magnetic field such that x_H is rotated by Φ_H clockwise with respect to the x -axis and the plane $\{x_H, y_H\}$ is tilted by θ_H clockwise with respect to $\{x, y\}$. Each characteristic frame contains orthogonal, *linearly polarized* components. For 2-D geoelectric structures, $\Phi_E = \Phi_H$ and $\theta_E = \theta_H = 0$. In 3-D structures it is possible that $\Phi_E \neq \Phi_H$ and/or $\theta_E \neq \theta_H \neq 0$: the electric and magnetic eigen-fields may not be orthogonal. Moreover, the electric and magnetic characteristic frames are not horizontal because the 3-D magnetotelluric field may be associated with significant gradients. Accordingly, the tilt angles θ_E and θ_H are measures of the local landscape of the telluric and magnetic field. The projection of the eigenstates on the horizontal plane yields elliptically polarized components: the normalized projected field vectors will have a major axis equal to $\cos\theta$ and a minor axis equal to $\sin\theta$ so that $b = \tan\theta$ is the *ellipticity* with $\theta > 0$ implying a counter-clockwise, and $\theta < 0$ a clockwise sense of rotation: ellipticity on the horizontal plane is defined in terms of a rotation in higher dimensional space. It is not straightforward to see in this thrifty presentation, but the essence of this analysis is that it approaches the geoelectric structure as a material birefringent at low frequencies and large scales.

Typical examples of impedance tensors processed with the ASVD are provided in Section S3 of the Supplementary Information. Analogous studies of all impedance tensors show that the geometry of the geoelectric structure is essentially 2-D with particular reference to the longer period band ($T > 1$ s). However, significant lateral variation of electric field strikes is observed at short periods and is attributed to *low-contrast* local/shallow 3-D resistivity perturbations of the

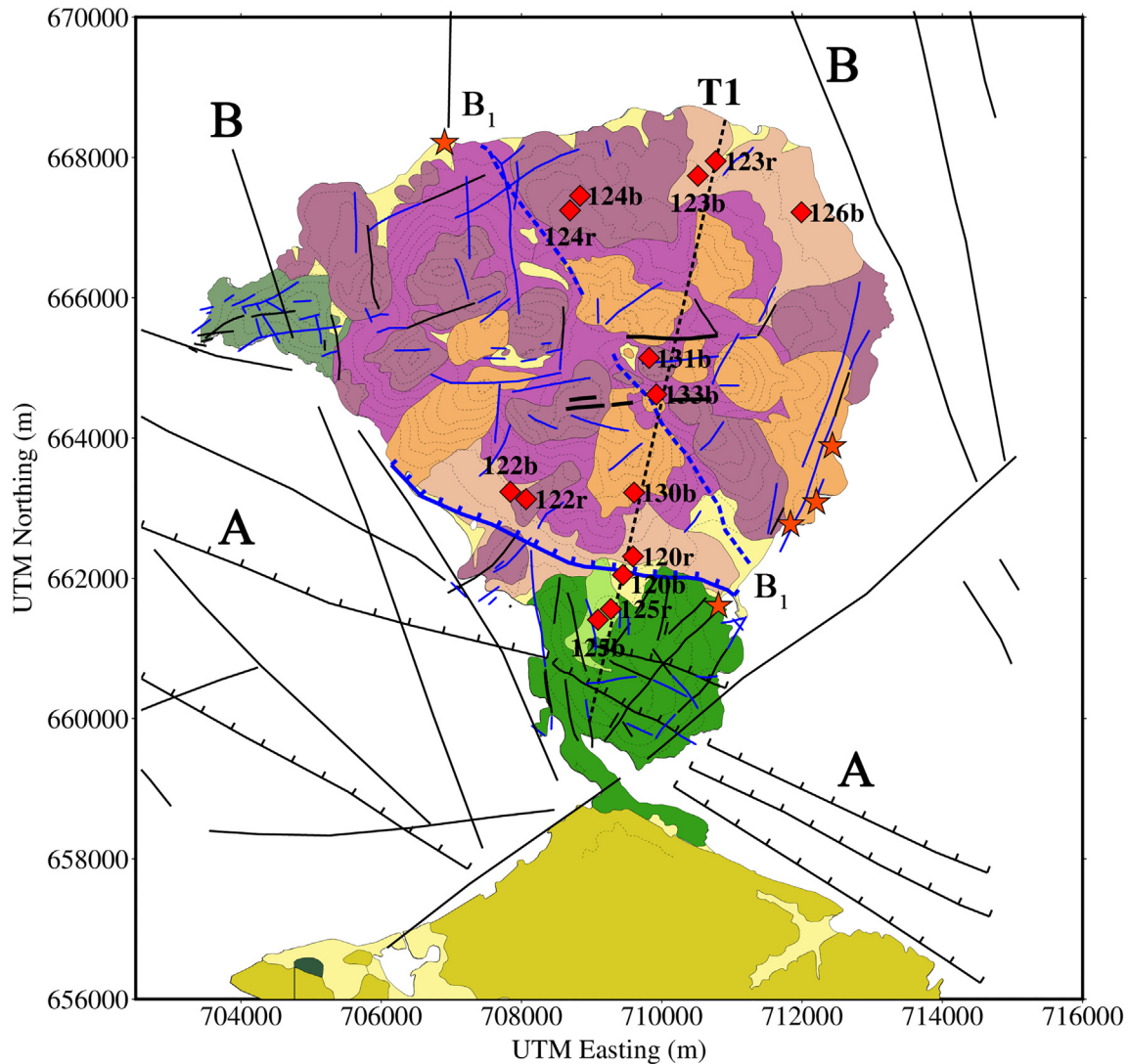


Fig. 5. The distribution of MT soundings on Methana Peninsula. T1 is the profile along which 2-D inversion has been performed. For the geological classification of rock formations see Fig. 4. A denotes the WNW-ESE, north-easterly dipping normal faults of the Argolis – Saronic Gulf boundary; the contact between the country carbonate and volcanic rocks is assumed to have developed in association with this family of faults. B denotes faults presumably associated with the R'-shear direction (see Section 2.2); B1 denotes a particularly significant member of the R'-shear family detected by the present analysis.

background 2-D structure. In what follows, we shall concentrate on the configuration of the deeper and larger-scale 2-D structure which is of particular interest as it is expected to contain the primary functional elements of the geothermal system (reservoir, fluid circulation conduits etc.).

Fig. 6 illustrates the configuration of the polarization ellipses of the maximum electric field (maximum eigenvalue) averaged over the bandwidth 10s–40s. The maximum electric field appears to be systematically associated with NE-SW and NW-SE strikes. The former averages to N51.8° and are observed at sites 123b, 123r and 126b at the northeast, as well as sites 120b, 122b, 122r, 125r and 130b at the southern half of the peninsula. The latter averages to N320.7° and are observed at sites 124b, 124r, 130b and 133b at the centre and north of the peninsula; it is also absolutely comparable to the strike of the R' faulting direction. Thus, while the orientation of individual polarization ellipses varies due to experimental errors and local weak 3-D effects associated with the transition from shallower to deeper 2-D structures (for example see Section S3 of the Supplementary Information), the average orientation of the maximum electric field between the central-north and south/northeast sectors of the peninsula is approximately orthogonal,

as expected of 2-D geometries. Such orthogonality implies profound differences in the mode on induction between the central-north sector and the rest of the peninsula. The ambiguity as to which of the NE-SW or NW-SE orientations represents the true strike of the deeper 2-D structure can be resolved by utilizing Induction Vectors.

In the interest of brevity and simplicity, and given the 2-D nature of the deeper structure, we shall limit our inquiry to the examination of Real Induction Vectors (RIVs). Fig. 7 illustrates the configuration of RIVs averaged over the bandwidth 10s–40s and drawn in the Parkinson convention, i.e. pointing toward current concentration: they consistently have NW-SE orientations. The RIVs are particularly strong at the northern sites 123B, 124B and 126B. Volti (1999) noted a similar behaviour which, after thin sheet modelling she attributed to a “coast effect”. This is may *not* be a likely interpretation, because the geometry of the IVs is unimodal across the peninsula, Saronikos Gulf is rather shallow, especially in the vicinity of Methana, and the overall resistivity contrasts at and below sea-level are so low, as to *not* generate a coast effect (see Sections S3 and S5 of the Supplementary Information, as well as Section 3.2 below). In our view, the strong vertical field observed at the three northern sites is more likely an effect of “normal” induction.

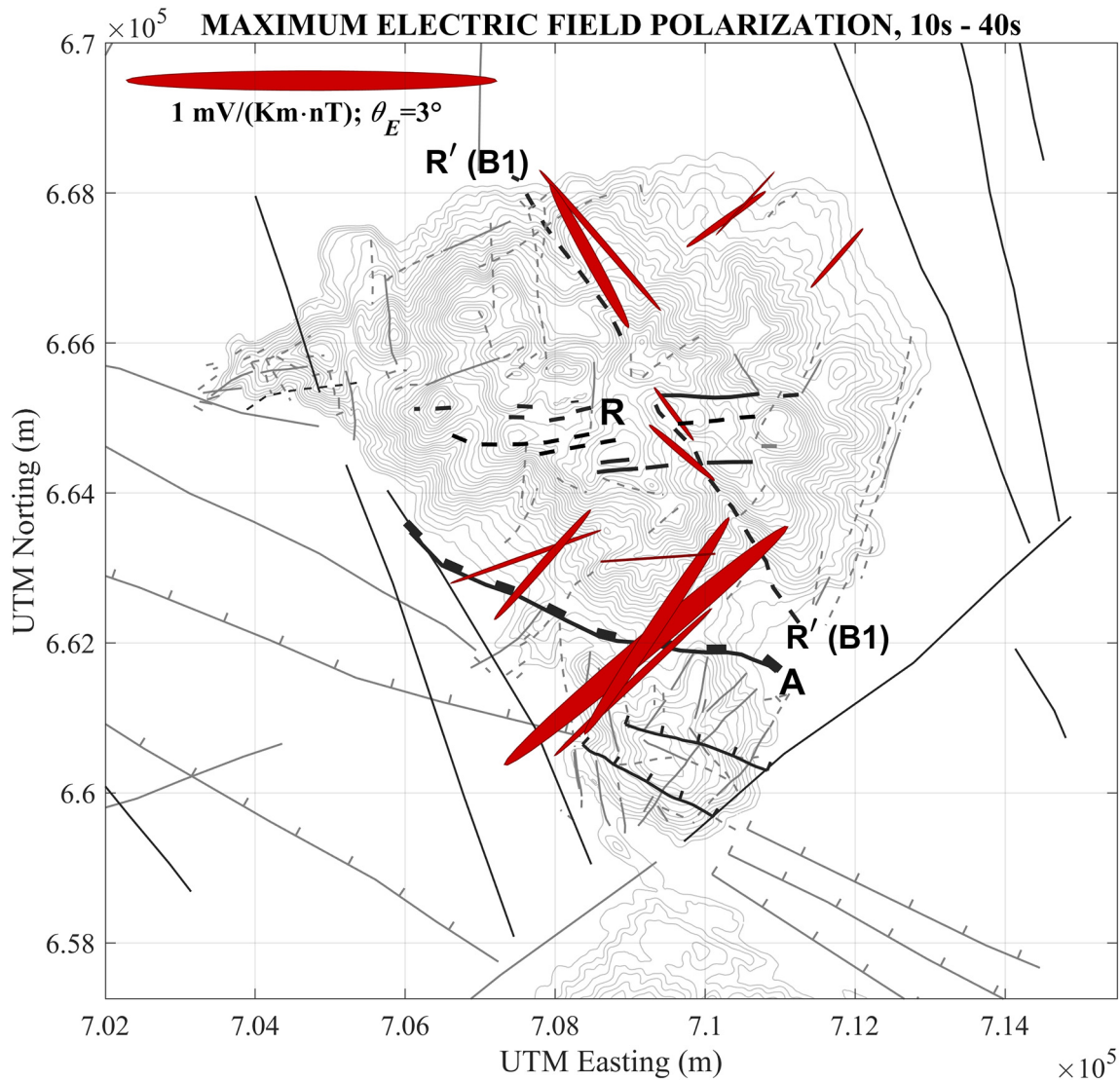


Fig. 6. Polarization ellipses of the maximum electric field (*maximum eigenvalue* of the impedance tensor) averaged over the bandwidth 10s–40s (responses from deeper and broader structures). R indicates the E–W R-shears discussed in Section 2.2. R' or B1 is the most significant R'-shear discussed in Sections 2.2 and 3.1; A are the WNW–ESE, north-easterly dipping normal boundary faults of Argolis peninsula; the contact between the country carbonate and volcanic rocks is assumed to have developed in association with family A.

The only RIV pointing SW, possibly to current concentrations along the normal boundary faults of the Saronic Gulf, is the one observed at site 125b. All the rest point to the NE and have a mean azimuth of $N49.2^\circ \pm 15.6^\circ$; they also appear to be almost perpendicular to the normal faults and R'-shears and therefore may point to current concentrations associated with elongate good conductors along their strike.

When the geoelectric strike determined from impedance tensors and RIVs are jointly examined, it becomes clear that the orientation of the 2-D background geoelectric structure is NNW–SSE, specifically $N330^\circ$ to $N350^\circ$ and prospectively associated with elongate good conductors developing along R' faulting direction. Although the impedance and IV responses measured at a given period interval do not normally originate at the same depth range, the mutual configuration of the induced electric and magnetic fields is tell-tale of the principal mode of induction in the vicinity of a sounding (which is absolutely necessary prior to application of 2-D inversion). Accordingly:

a) At the southern and northeast sectors of the peninsula (sites 120b/r, 122b/r, 123b/r, 125b/r, 126b and 130b), the approximate collinearity (parallelism) of the maximum electric field and RIVs is

consistent with TM induction above the resistive side(s) of $N320^\circ$ – $N330^\circ$ lateral conductivity interfaces.

b) At the central-north sector of the peninsula (sites 124b/r, 131b and 133b), the approximate orthogonality of the maximum electric field and RIVs is consistent with TE induction over a $N320^\circ$ – $N330^\circ$ conductive dyke, presumably associated with a significant local expression of the R' faulting direction and designated B1 in Figs. 5–7.

The surface expression of B1 has been tentatively linked to an approx. $N330^\circ$ alignment of indentations in the topography of the peninsula, located in the immediate vicinity of sites 124b/r, 131b and 133b and joining the areas of Palaea Loutra (*Antique Baths*) and Vromolimni (*Malodorous Pond*) at the north of Methana town, both of which are appropriately associated with past hydrothermal activity. Clear evidence of past hydrothermal activity can also be observed on the southwest cliffs of Mt South Tsonaka, between Vromolimni and site 133b. If our interpretation is correct, it implies that at least in Methana and at depth, the R' shear is important for the ascent, circulation and discharge of thermal fluids (that epiphenomenally generate high conductivities).

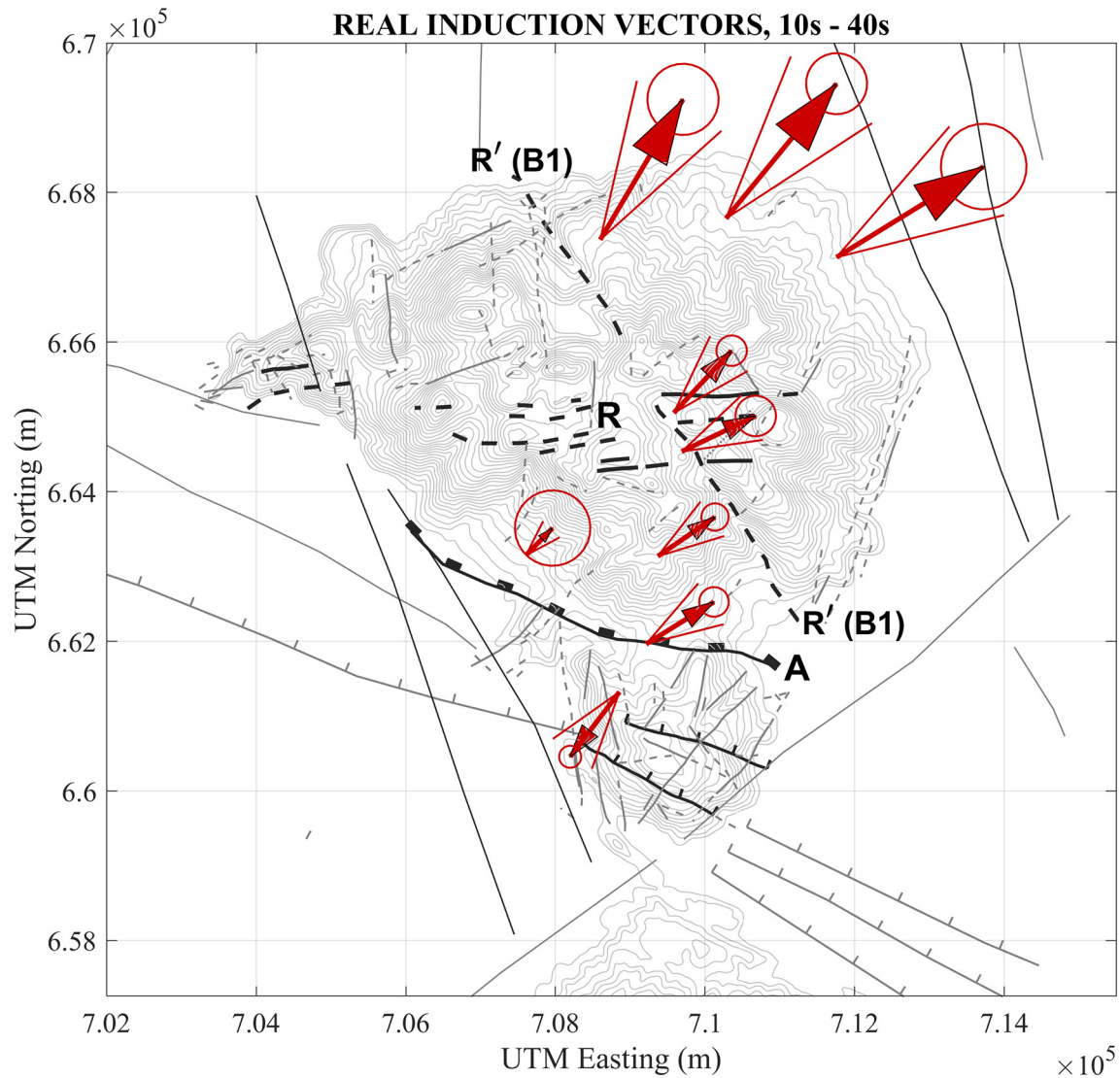


Fig. 7. Real Induction Vectors averaged over the bandwidth 10s–40s containing responses from deeper and broader structures. Angular uncertainties are indicated by thin red lines bracketing the vectors; magnitude uncertainties are indicated by circles drawn at the tips of the vectors. As per Fig. 6, R indicates the E-W R-shears, R' or B1 is the most significant R'-shear and A are the WNW-ESE normal boundary faults of Argolis peninsula. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

3.2. Quantitative interpretation

Two-dimensional inversion was carried out with the algorithm of Rodi and Mackie (2001), along the transect T1 shown in Fig. 5. The TE and TM modes were simultaneously inverted assuming the TE/TM mode configuration specified above. A discretized homogenous half-space was used as a starting model, with topography, bathymetry and the presence of seawater all taken into account. The discretization scheme is apparent in Fig. 9. An error floor of 5% was used for the data uncertainty. Several inversions with different regularization factors were carried out before a final model was declared. The aggregate RMS error of the final model is only 1.14, and with an expectation value of 376, the aggregate χ^2 is 417, of which approximately 25% derives from only one sounding (131b). The quality of the final solution can be evaluated in Section S4 of the Supplementary Information, where detailed graphs of the observed vs. predicted maximum and minimum response apparent resistivities and phases are shown in Figs. S5–S13. As it turns out, the final model successfully reproduces the responses observed at sites 125b, 125r, 120r, 133b, 123b and 123r. On the other hand, the model successfully reproduces the shape of the

responses observed at sites 120b, 130b and 131b. Thus it resolves the geometry of nearby resistivity contrasts but fails to reproduce an accurate magnitude of the contrasts because it cannot precisely fit the maximum apparent resistivity at sites 120b/130b and the maximum and minimum phase at site 131b. Overall, it is easy to assert that the inversion has been successful and that the final resistivity model is an adequate representation of the geoelectric structure.

The final model is shown in Figs. 8 and 9. Additional 3-D renderings of the resistivity model are presented in Section S5 of the Supplementary Information (Figs. S14–S16). Let us begin by pointing out that between kilometres 5 and 6.5 the model *cannot* be interpreted with confidence due to the absence of direct measurements. In the southern half of section T1 a massive conductive domain ($\rho < 5 \Omega\text{m}$) can be observed between soundings 120r (location Throni) and 131b (location Stavrolongos). It extends to a depth of approx. 1.6 km bmsl and appears to be bounded, on one hand by an interface dipping approximately 60° to the north beneath site 120r, and on the other by a sub-vertical conductor in the area of sites 131b/133b. Both of the latter two conductors rise very close to the surface. The former (north-dipping) conductor can be directly associated with the interface between the limestones of the

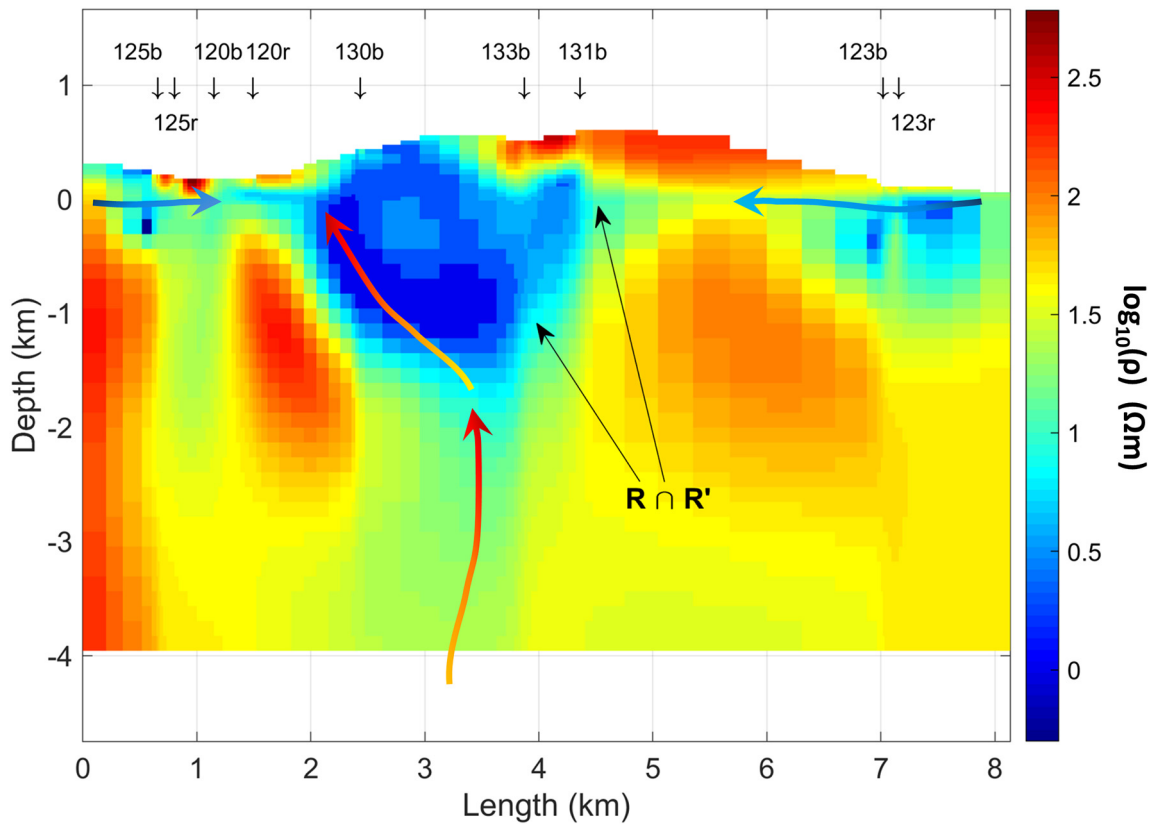


Fig. 8. Resistivity cross-section along the transect T1 of Fig. 5, obtained by 2-D inversion with the algorithm of Rodi and Mackie (2001). White arrows (red in online version) indicate the ascent of thermal fluids. Black arrows (blue in online version) indicate the infiltration of sea water. $R \cap R'$ indicates the area of interaction between the R and R' shears (see Section 2.2 and Figs. 6 and 9). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

local Alpine basement and the Quaternary volcanic rocks of the MV (Fig. 9). The latter (sub-vertical) conductor is apparently collocated with the intersection of R and R' faults (Fig. 9). It also appears to extend to depths of at least 4 km albeit with resistivity increasing to 25–30 Ωm . At depths between 0.5 and 1.5 km bmsl, the geometry of the “massive conductor” has attributes suggestive of a geothermal reservoir and also appears to rise close to the surface in the vicinity of site 130b, between Dritseika and Megalochori. Inasmuch as there is no clear evidence of present or past hydrothermal activity there, this could be an artefact due to the absence of nearby soundings to constrain the inversion. A second sub-vertical conductor ($10 \Omega\text{m} < \rho < 40 \Omega\text{m}$) is imaged beneath site 125r; it extends to depths greater than 2.5 km and is sharply bounded by relative resistors between sites 125b and 120r (Fig. 8). This feature may be related to the normal boundary faults of the Saronic Gulf (designated by ‘A’ in Figs. 5 and 9). Good conductors extending to depths greater than 1 km bmsl are detected in association with the north-western sites 123b/r – this could be an effect of seawater intrusion along the R'-related coastal faults, as evident by the configuration of local RIVs (Fig. 7). Throughout the section, one may observe a conductive layer at sea level, indicating lateral thermal fluid diffusion and pervasive seawater infiltration throughout the peninsula.

The apparent association of faults and conductors at the south-east and centre of the peninsula (between sites 120b and 131b) can be interpreted as an epiphenomenon of thermal fluids ascending via the permeable zone generated by the intersection of R and R' faults at the centre of the peninsula and charging a geothermal reservoir at depths 0.5–1.6 km beneath the central domes (Chelona, Chionessa and Loutsas). Therefrom, thermal fluids ascend and circulate, firstly through the interface between the Alpine basement and volcanic rocks discharging at Nea Loutra (*New Baths*) in Methana town, and secondly through the B1 fault discharging at Vromolimni and Agios Nikolaos to

the south and Palaea Loutra (*Antique Baths*) to the north. The common source of the thermal fluids may easily account for the chemical and isotopic similarities observed by Dotsika et al. (2010) between the thermal springs of Nea Loutra, Agios Nikolaos and Palaea Loutra.

Finally, thermal fluids are inevitably mixed with cold seawater, presumably via the permeable zones created by faulting, which amply explains the chemical composition and rather low temperatures of the thermal springs (see Section 2.3).

4. Three-dimensional aeromagnetic data analysis

4.1. The digital aeromagnetic anomaly model

The present study utilizes data from the uniform and homogeneous aeromagnetic anomaly map of Greece (Chailas et al., 2010; Chailas and Tzanis, 2019). In our study area, the compilation was based on the 1:50,000 map series produced by Hunting Geology and Geophysics Ltd. under contract by the Institute of Geological and Mineral Exploration (IGME), for area C3 (Eastern Central Greece). Measurements were conducted in the year 1977. Hunting has generally used a measurement spacing of 200–250 m along-track and has flown in a NE-SW direction (approx. N45°) at a nominal clearance of 300 m above ground level and nominal distance of 800 m between tracks. Connection lines were also flown in a SW-NE direction with 10 km spacing between tracks. The IGRF correction was based on the IGRF model for the epoch 1977.3. The original 1:5000 maps were converted to high-resolution raster images and the contour lines were then digitized to vector form in image coordinates. Using the corners of each map as control points, the digitized contour coordinates were geo-located and transformed to the UTM projection. The resulting set of digital data lines was finally

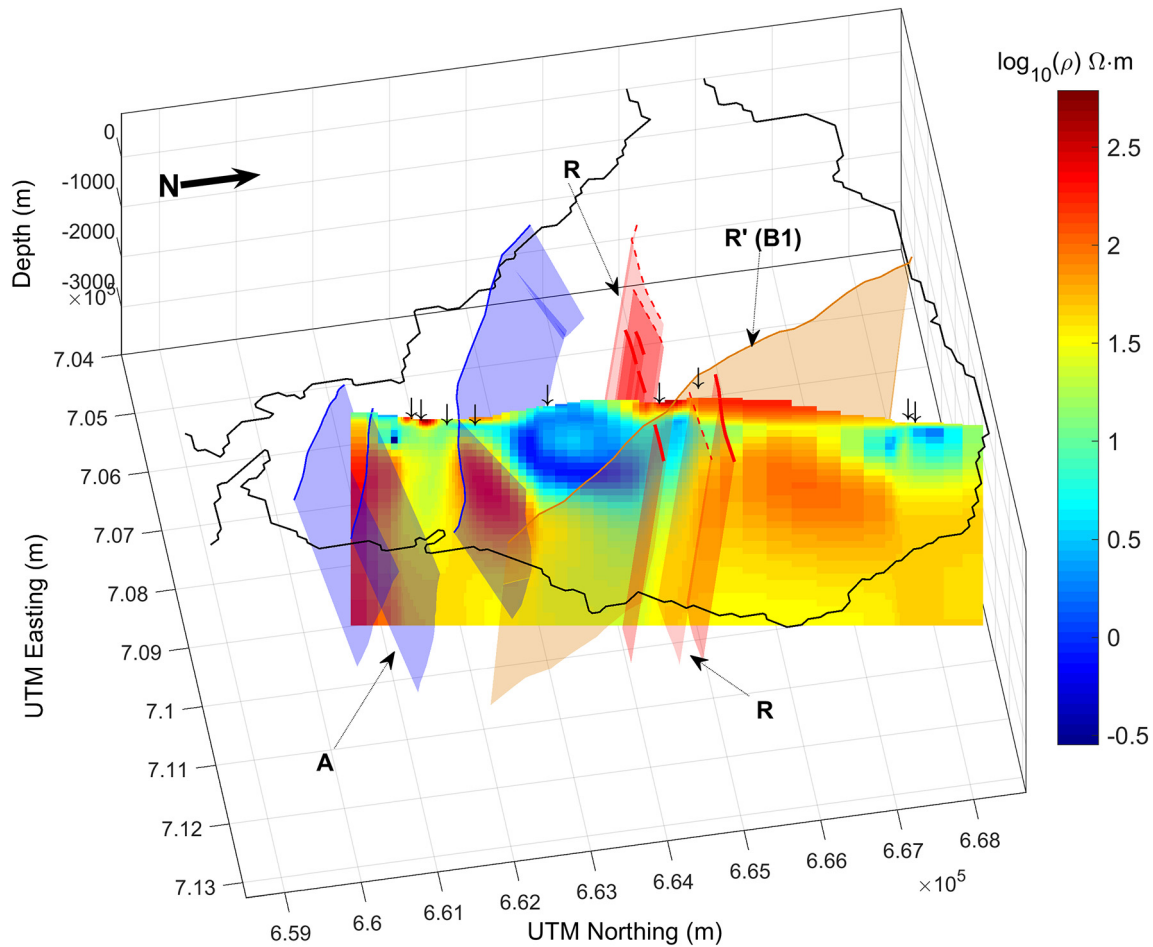


Fig. 9. Three-dimensional rendering of the resistivity cross-section T1 with the most significant contemporary active faults: R (red) are the W-E R-shears discussed in Section 2.2. R' or B1 (orange) is the most significant R'-shear discussed in Sections 2.2 and 3.1; A (blue) are the WNW-ESE, north-easterly dipping normal boundary faults interfacing the Argolis peninsula and the Saronic Gulf; the contact between the country carbonate and volcanic rocks is assumed to have developed in association with this family of faults. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

interpolated to a rectangular grid of 250×250 m spacing, which is optimal given the orientation and distance between track lines.

Fig. 10a shows the digital aeromagnetic anomaly model (DAAM) of the study area and Fig. 10b the same reduced to the North Pole so as to be centred directly above the magnetic sources (RTP-DAAM). The strongest anomalies are observed above the major domes of the centre of the peninsula, indicating that the bulk of extrusive activity has taken place in that area. It is also apparent that magmatic activity develops primarily in the W-E direction in association with the subaerial central (Chelona, Chionessa, Choni, Chiroma) and eastern (Kossona, North Tsonaka) domes, and secondarily in the NW-SE direction in association with the southeastern subaerial (South Tsonaka) and northwestern subaerial (Kammeni Chora) and submarine (Pausanias) domes. Rather clear W-E indentations in the intensity of the total magnetic field are also observed at the area of the central domes, between Chelona and Chionessa to the south and Choni/Chiroma to the north, apparently coincident with the W-E R-shear faults mapped in that area by Dietrich and Gaitanakis (1995). A discussion of the possibility of a direct association between the “magnetic field indentation” and faulting will be attempted below. Finally, one may observe an elongate W-E to WNW-ESE magnetic anomaly south of Methana peninsula, which is attributed to (known) intrusive activity beneath Poros Island and along an important W-E fault with significant right-lateral offset crossing the Argolid (Tzanis et al., 2018a). This fault was also mapped by Pe-Piper and Piper (2013) and Dietrich and Gaitanakis (1995).

4.2. Three-dimensional inversion

We implemented a 3-D inversion algorithm developed by the Geophysical Inversion Facility of the University of British Columbia (Li and Oldenburg, 1996, 2003, 2010). This method is computationally efficient and flexible in including diverse a priori information to physically constrain the range of acceptable models. The algorithm assumes that the magnetic response is entirely due to induced magnetization (negligible remanent magnetization), thus avoiding complications caused by geological and thermo-chemical processes. It solves the inverse problem on the basis of a discrete susceptibility model defined on a rectangular coordinate system and comprising a three-dimensional array of rectangular prism cells (mesh). Details about the construction of the mesh are provided in Section S7 of the Supplementary Information. A priori information admitted by the algorithm includes surface geology, borehole data, structural information from geophysical cross-sections etc. In our case, only surface information was available in the form of boundaries between different rock formations and in situ magnetic susceptibility measurements. A brief presentation of these measurements and their usage in the construction of an initial (starting) susceptibility model (mesh) and variation bounds is given in Section S6 of the Supplementary Information.

It is well known that the magnetic data have no inherent depth resolution. A numerical consequence of this is that when an inversion is performed, the reconstructed susceptibility tends to concentrate near the observation point. The rapid decay of field intensity with distance

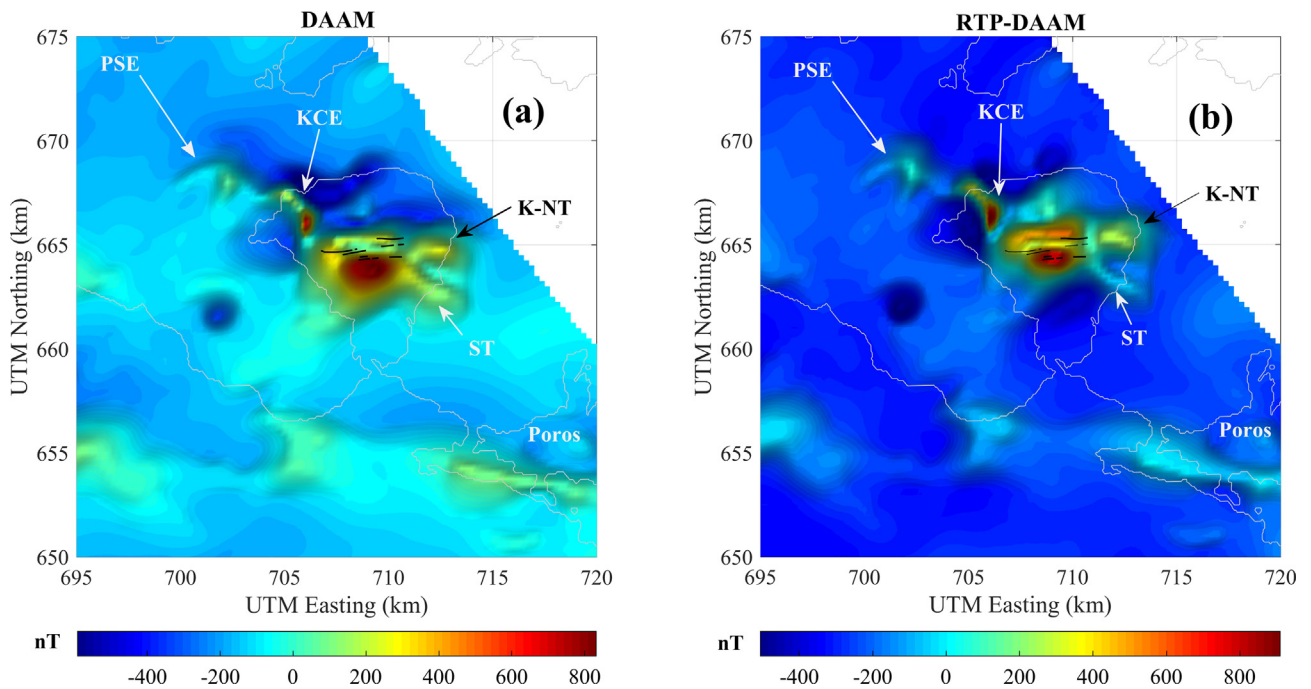


Fig. 10. (a) DAAM is the digital aeromagnetic anomaly model (total field aeromagnetic anomaly) of the Methana VC as observed at a constant ground clearance of 300 m. (b) RTP-DAAM is the same DAAM reduced to the North Pole. KCE: Kammeni Chora Eruption; PSE: Pausanias Submarine Eruption; ST: South Tsonaka (dome); K-NT: Kossona–North Tsonaka (dome).

between model cells and observation points prohibits the reconstruction of a model function that possesses significant structure at locations far from observations. This limitation can be overcome with weighting schemes that counteract this natural decay by applying a geometrical correction approximately equal to the inverse of the geometrical decay and give all cells equal probability to contribute with non-zero susceptibility. The weighting function has two forms: depth weighting and distance weighting, the latter being more appropriate in case of irregularly spaced or sparsely distributed data, or the presence of highly variable observation heights (Li and Oldenburg, 1996, 2003, 2010; Williams, 2008). Now, consider that herein we are interested in studying larger-scale structures, located at sub-kilometric to kilometric scale depths and associated with kilometric scale tectonic and magmatic activity. The original DAAM (Fig. 10a) is quite detailed and having been acquired in constant ground clearance mode, comprises measurements obtained at highly variable observation heights. Given the notorious non-uniqueness of magnetic data inversion and regardless of how robust/efficient the distance-weighting scheme may be, the reconstructed susceptibility model might contain structural variability that would not serve the above objective. Accordingly, we decided to use depth weighting and invert a smoother version of the original DAAM by upward continuing it to a flat observation surface above the highest observation point.

The DAAM of Fig. 10a was upward continued to a unique constant altitude of 1100 m, approx. 350 m above the highest topographic point of the study area and somewhat above the highest possible observation point (approx. 100 m). Because the observation surface was not flat, thereby requiring continuation between arbitrary surfaces, we adopted an equivalent sources approach and implemented the highly accurate and computationally efficient method of Xia et al. (1993). Prior to continuation, the *unmeasured* void nodes of the northwest quadrant of the original DAAM were filled with low-amplitude white noise. The upward-continued DAAM is illustrated in Fig. 10b. Finally, and as required by the inversion algorithm, an uncertainty of 5% plus a random, small amplitude perturbation was assigned to each element of the upward continued DAAM. This is *arbitrary* because a nominal observation error was not given in the surviving report of Hunting Ltd. and

was not included in the map sheets from which the DAAM was reconstructed. Nevertheless, hitherto international experience indicates this is typical of the error level associated with “recent” aeromagnetic measurements and could rise to twice that order for older data sets. Although we are dealing with “old” measurements, we are inverting upward-continued versions of the original data which has lower dynamic range and should be associated with lower-than-measured uncertainties: the assumption of 5% error is realistic. The inversion was performed with main field elements corresponding to epoch 1977.3, namely $I = 55.6^\circ$, $D = 1^\circ$ and a total intensity equal to 45000 nT. Remanent magnetization was considered to be a minor issue because the bulk of the activity that built the volcanic edifice occurred during the current epoch of “normal” polarity.

The inversion produced an “exceptionally” good model. Fig. 11b illustrates the residuals after subtraction of the theoretical anomalies generated by the inversion, from the observed anomalies of Fig. 11a. As can clearly be seen, the model reproduces the observations almost faithfully, generating a maximum residual of less than 10 nT at the centre of the peninsula, near the peak of the main anomaly; The goodness of fit is $R^2 = 0.997$, the RMS error 1.29 and, within the arbitrariness of the 5% error assumption, the observed χ^2 misfit is 408.3 given an expectation value of 4233.

4.3. Results and interpretation

Fig. 12 shows four N-S sections across the model, designed so as to cross the major features of the MV: Pausanias Submarine volcano (a), Kammeni Chora volcano (b), the central domes (c) and the eastern domes (d). The SE quadrant of the study area appears to comprise of weakly magnetized rocks with susceptibilities generally lower than 15×10^{-3} . In the area between coordinates 698E–702.5E and 660N–664N, this can be partially attributed to a peculiar negative local anomaly (Fig. 10) and because it may also be degraded by edge artefacts, it should be approached with much caution. In the area between coordinates 703E–706E and 660N–664N the structure can be attributed to south-westward branching of the main/central anomaly (Fig. 10) and should be considered reliable.

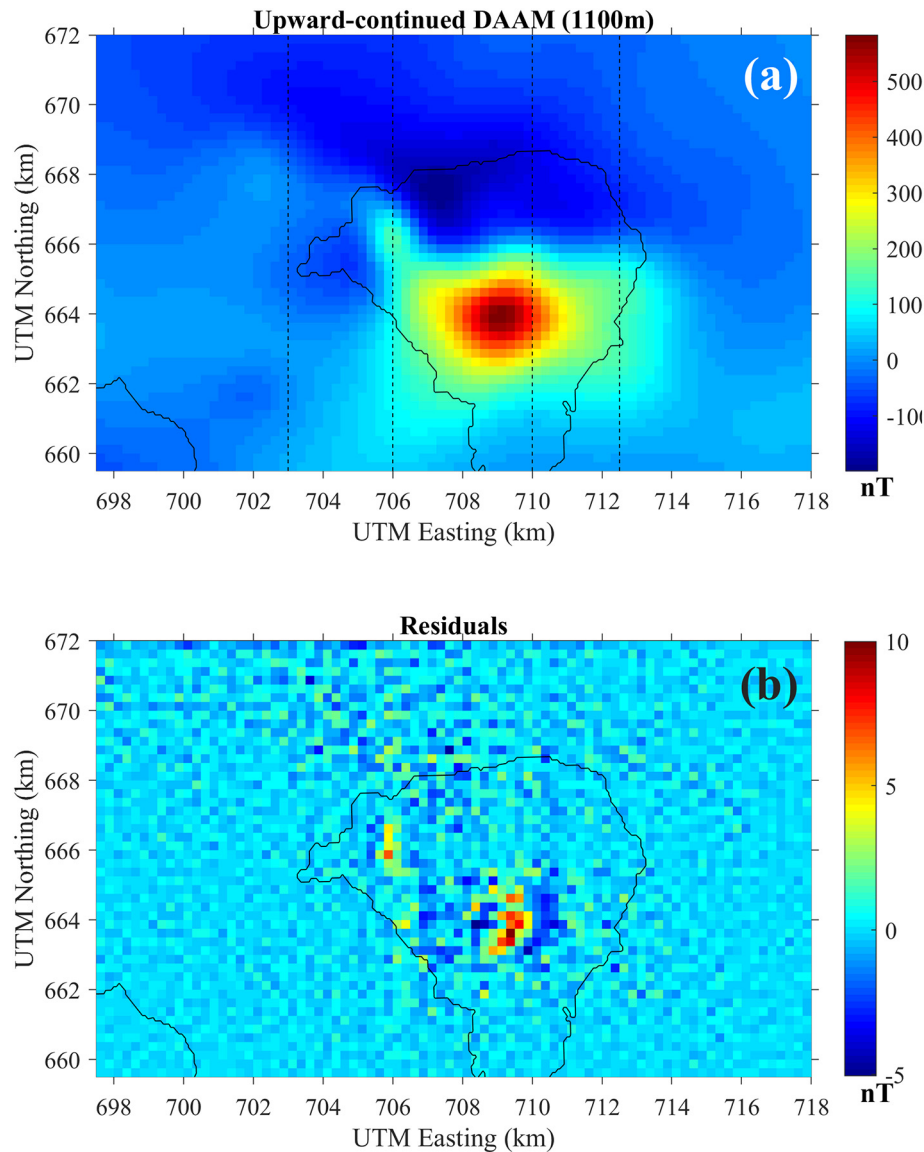


Fig. 11. (a) The DAAM of Fig. 10a upward continued to a unique constant altitude of 1100 m. This is the data used for 3-D susceptibility inversion. The dashed vertical lines indicate the location of the model slices shown in Fig. 12. (b) The residuals of the 3-D inversion i.e. the observed anomaly of Fig. 11a minus the calculated anomaly.

At depths greater than a few hundred metres beneath the eastern/central domes, Kammeni Chora and Pausanias volcanics, one observes susceptibilities of the order of 40×10^{-3} – 80×10^{-3} , considerably higher than those measured at the surface. Beneath the eastern and central domes the high susceptibility domain extends down to at least 4 km, while beneath Kammeni Chora and Pausanias it extends to approx. 2 km. This implies the presence of igneous rocks with intermediate to mafic compositions, as opposed to those of the extrusive formations at the surface. Pe-Piper and Piper (2013) discuss the abundance of mafic enclaves in some volcanic rocks on Methana and we have observed dark andesitic xenolites in freshly cut sections of the Kammeni Chora volcanics; these are rounded, meaning that they originated relatively far/deep in the crust. Thus, there is compelling petrographic evidence for the presence of mafic, or at least intermediate composition rocks at depth.

Beneath the central domes, the susceptibility diminishes at depths greater than 4.5 km and drops to under 10×10^{-3} below 5–5.5 km. This may indicate thermal demagnetization due to the presence of a shallow magma chamber at depths below 5 km. A palaeomagnetic analysis of Aegina dacites indicates unblocking temperatures of the order of

580 °C (Morris, 2000). Unblocking temperatures are lower than the Curie point albeit close to it, but still allow the material to be magnetized along the direction of the geomagnetic field. It is generally thought that the Aegina igneous formations originate at the same deep magma source, and have similar compositions to those of Methana (e.g. Pe-Piper and Piper, 2005, 2013). If so, the temperature should not be higher than 550–600 °C at 4–5 km, but might rise above this level and even above the Curie point at greater depths; this is compatible with the depths expected for shallow magma chambers below arc volcanoes.

To the extent that there is no evidence of appreciable susceptibility below the depth of 2 km, the recent Kammeni Chora and Pausanias formations do not appear to have been fed by vents directly beneath them. Thermal demagnetization should be ruled out because high temperatures at shallow depths should also be associated with onshore/offshore thermal activity of which there is absolutely no indication. It follows that if a central magma source is indeed located beneath the central domes, the recent Kammeni Chora and Pausanias formations may have been fed laterally, possibly through faults, and for this reason are confined to depths shallower than 2 km. This possibility will further be explored below.

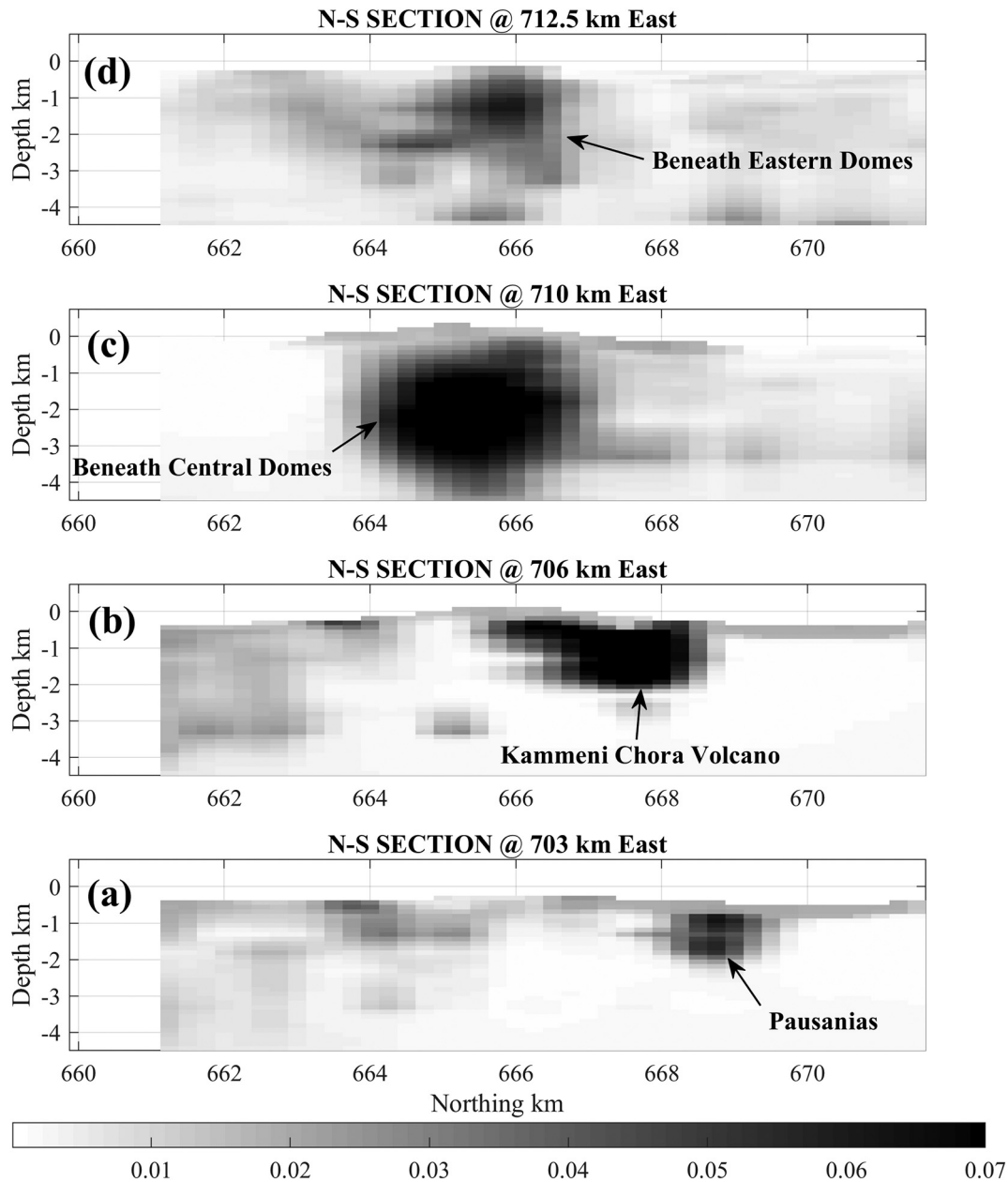


Fig. 12. N-S sections through the 3-D magnetic susceptibility model and across the major features of the MV: (a) the Pausanias submarine 'volcano'; (b) the Kammeni Chora eruption; (c) the central domes and, (d) the eastern domes. The exact locations of the sections are shown in Fig. 11a with dashed vertical lines.

Figs. 13–14 present the 3-D susceptibility structure beneath the MV in the form of an isometric surface (isosurface), i.e. a hull of equal susceptibility whose interior comprises values greater than that of the hull. The external to the hull structure (with susceptibilities smaller than that of the hull), is shown as empty space. In all cases, structures near the mesh boundaries are excluded from consideration as they may be “contaminated” by inversion artefacts. Fig. 13 is a plan view of the 25×10^{-3} isosurface and Fig. 14 is the same but viewed from an angle high above the SE corner of the study area (observer's azimuth/elevation is $150^\circ/50^\circ$ respectively). Both figures include 3-D renderings of the most significant contemporary active faults hitherto recognized. Additional images of the 3-D susceptibility structure viewed from the east ($90^\circ/0^\circ$), north ($0^\circ/0^\circ$) and northwest ($225^\circ/50^\circ$) are presented in Section S8 of the Supplementary Information (Figs. S20, S21 and S22 respectively). We note that the 25×10^{-3} isosurface, which by no coincidence corresponds to the average susceptibility of relatively healthy

andesite, affords the most meaningful and informative representation of the model. At lower values, significant features are partially or totally masked by the halo of the volcano's core structures; at higher values information is purged and when $k = 4 \times 10^{-3}$, all that remains is a featureless bean-like shape right in the middle of the graph.

As a general observation, the andesitic core of the volcano (beneath central and eastern domes) develops in a clear W-E direction and aligns with the orientation of the R-shear. In fact, and although this should be taken with a large grain of salt, a number of W-E striations on the isosurface appear to coincide, or lie in the extension of W-E fault lines mapped at the surface. It is also apparent that the area between the central and eastern domes exhibits secondary NNW-SSE and NE-SW structural features; a particularly steep NNW-SSE slope (approximate coordinates 7.104×10^5 E, 6.661×10^5 W) is also collocated with the fault line of a NW-SE fault. All the above features are particularly evident in Fig. 13.

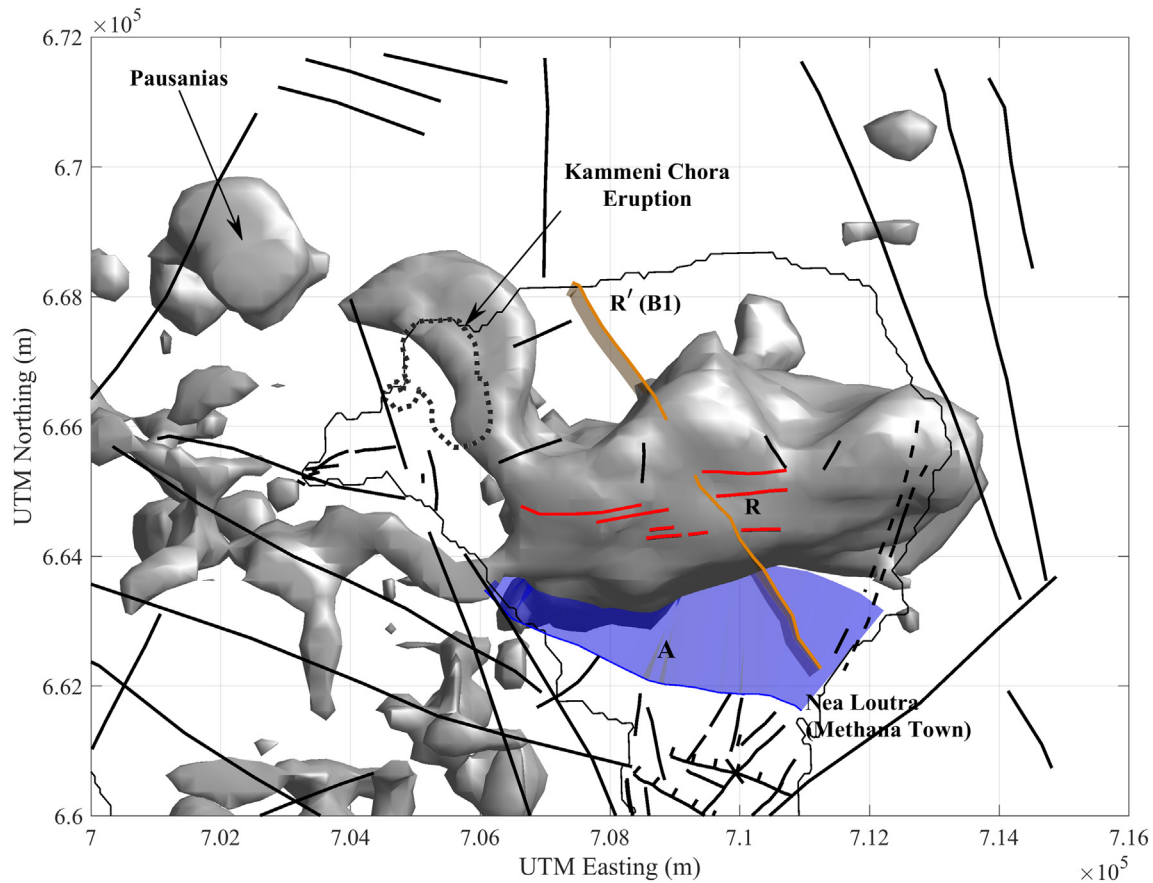


Fig. 13. Plan view of the 3-D susceptibility structure beneath the MV as presented in the form of 25×10^{-3} isometric surfaces enclosing higher susceptibility values. The most significant faults hitherto recognized are also rendered as follows: R are the E-W R-shears discussed in Section 2.2; R' or B1 is the most significant R'-shear discussed in Sections 2.2 and 3.1; A are the WNW-ESE, north-easterly dipping normal boundary faults interfacing the Argolis peninsula and the Saronic Gulf.

At the western half of Methana peninsula the core branches heavily to the NW: the Kammeni Chora and Pausanias intrusions align in a N320°–N330° orientation. Moreover, the outcropping Kammeni Chora extrusive volcanics are offset to the west with respect to what appears to be the main body of this eruption (Fig. 13). This indicates that the extrusion may have taken place through an inclined surface, quite possibly a fault. It also appears the Kammeni Chora extrusion may have been partially modulated by N-S structural discontinuities; in fact, Dietrich and Gaitanakis (1995) have mapped a N-S fault running along the boundary of the extrusion.

West of the Methana peninsula and between eastings 6.97×10^5 – 7.06×10^5 m and northings 6.62×10^5 – 6.66×10^5 m, one observes an intricate complex of tubular magnetized rock formations, approximately aligned in the NW-SE direction. Moreover, between eastings 7.01×10^5 – 7.08×10^5 m and northings 6.60×10^5 – 6.63×10^5 m, one observes an additional complex of “blobs” with susceptibilities higher than 25×10^{-3} . At first sight, these may appear to be artefacts. Nevertheless, closer inspection of the both the original (Fig. 10a) and upward continued (Fig. 11a) DAAMs demonstrates that low-intensity anomalies exist in these areas, generated by buried formations of ostensibly igneous origin; these are more clearly observed in the colour maps of the online version. It is conceivable that due to the small amplitude of their anomalies, these formations cannot be rigorously constrained by the inversion and appear patchy in the model. The former (tubular) complex may comprise small scale intrusions associated with the NE-dipping normal boundary faults of the Saronic Gulf. If so, the web of $k > 25 \times 10^{-3}$ tubes may well be the inversion algorithm's attempt to cope with a web of magnetized and thermo-chemically demagnetized volumes of igneous rocks. Their recovered characteristics should not be taken at face value, but their existence should also not be dismissed.

5. Synthesis and discussion

To begin with, let us point out that the spatial and quantitative analysis of the magnetotelluric data and the 3-D susceptibility model are consistent in two important ways: a) Both lines of evidence indicate that the main vent of extrusive activity (and possibly the magma chamber) is located beneath the central domes and apparently coincides with the intersection of the main contemporary active faults, namely the normal faults of the Argolis Peninsula – Saronic Gulf boundary, the W-E R-shear and the NNW-SSE R'-shear; these faults also appear to form the principal path for the ascent, circulation and discharge of thermal fluids. b) Both lines of evidence indicate that the placement of intrusive and extrusive volcanic rocks and the circulation of thermal fluids are directly associated (controlled) by the main contemporary active faults. These inferences are also evident in Fig. 15 which combines a 3-D rendering of the magnetotelluric cross-section T1 with the 25×10^{-3} isometric surface of the 3-D susceptibility model shown in Figs. 13–14. In the case of Methana Volcano, the control of tectonics on magmatic and thermal activity would appear to address the “granite space problem”. The problem concerns the mechanism by which space is created for viscous magma to intrude and occupy large volumes of the crust (e.g. Hutton, 1996). In reality, it is more or less understood that magma can easily and quickly flow up vertical cracks and through small incremental intrusions, form large igneous bodies (e.g. Stevenson, 2009; Petford et al., 2000). It thus appears that the solution of the problem is that magma ascends through the space created by tectonic activity, and more effectively so, through the weak zones generated by the interaction of the main active faults. In this respect, the localization

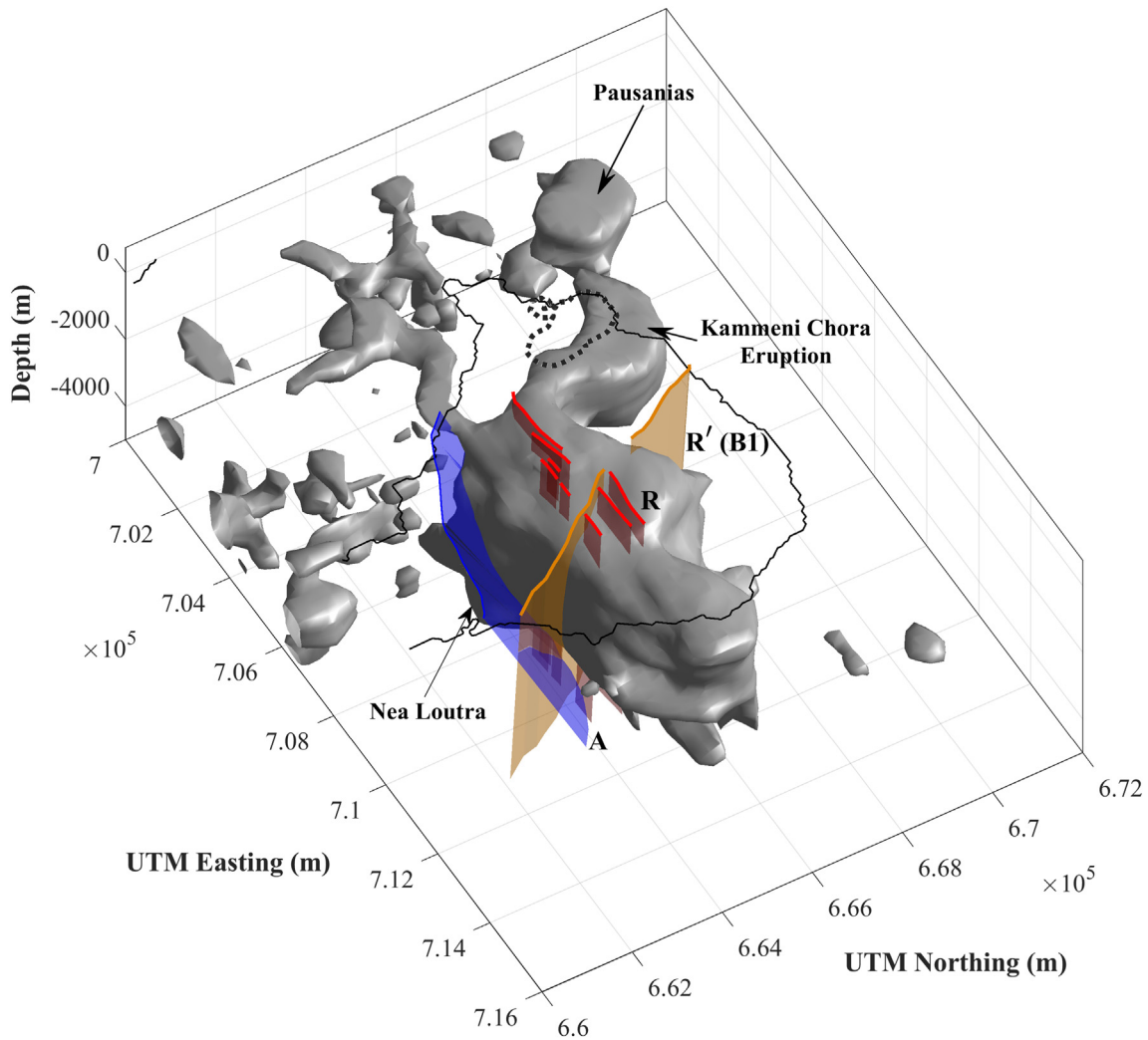


Fig. 14. 3-D susceptibility structure beneath the MV presented in the form of 25×10^{-3} isometric surfaces enclosing higher susceptibility values and viewed from the southeast (observer azimuth/elevation is $150^\circ/50^\circ$). The figure includes the most significant faults hitherto recognized: R are the E-W R-shears discussed in Section 2.2; R' or B1 is the most significant R'-shear discussed in Sections 2.2 and 3.1; A are the WNW-ESE, north-easterly dipping normal boundary faults interfacing the Argolis peninsula and the Saronic Gulf.

and evolution of Methana Volcano are completely analogous to those of the much larger and meaner Santorini Volcanic Complex at the middle of the Hellenic Volcanic Arc (Tzanis et al., 2020).

Our findings and conclusions are very similar to those of Pe-Piper and Piper (2013), albeit more complete and comprehensive due to information not available to those authors at that time. To begin with, let us note that quite obviously, the origin of the contemporary crustal stress field and associated faulting patterns should 'predate' in a cause and effect sense, the inception of the Saronic Gulf which is Pliocene of age. In addition to the R (W-E) and R' (NNW-SSE) shears, the stress field allows for tertiary NE-SW synthetic (dextral) P-shears (Fig. 3) and almost exactly N-S antithetic (sinistral) P'-shears. The latter two faulting directions are auxiliary features accommodating deformation at local scales. Almost exactly N-S local-scale faults have been mapped on Methana peninsula, particularly at the north where they have been associated with the effusion and emplacement of the Kammeni Chora dacites. To the best of our knowledge, exactly N-S large-scale normal faults do not appear in any direct observations of the crust obtained by seismological or geophysical imaging available through the international literature. On the other hand, almost N-S ($N330^\circ$) large-scale normal faults form the east coast of East Peloponnesus and the Argolic Gulf as well as the east flank of the Argolic plain. The latter extensional structures (grabens) almost exactly trace the location of the inflection of the

subducting slab (Fig. 1) and define the local strike of the associated plutonism along their axis (Tzanis et al., 2018a and references therein). The strike of the contemporary R'-shears at Methana is exactly parallel to these faults and to the local strike of the 100 km isodepth of the subducted slab, which also underlies all manifestations of volcanic and plutonic magmatism along the western volcanic arc (e.g. Tzanis et al., 2018a). Accordingly, it stands to reason that the R'-shears may be inherited features genetically related to the $N330^\circ$ normal faults of east Peloponnesus, that have locally adapted to the crustal stress field of East Peloponnesus – Saronic Gulf areas. In this respect, they could be closely related to the ascent and emplacement of Pliocene magmas, albeit in the context of a tectonic regime that is far more complex than hitherto appreciated.

In the broader study area, large-scale NE-SW oblique-normal and presumably right-lateral faulting are apparently responsible for the formation of the Hydra – Spetsai strait; they may represent the continuation and diffusion of analogous transcurrent faults descending from the Aegean Sea (Fig. 1; also see Sakellariou and Tsampouraki-Kraounaki, 2019 and Tzanis et al., 2010). These are exactly parallel to NE-SW faults mapped on Methana peninsula, which appear to affect the emplacement of magmas beneath the eastern domes (e.g. Fig. 13). Accordingly, the NE-SW faults may again represent inherited features that have locally adapted to an auxiliary (P-shear) role within the

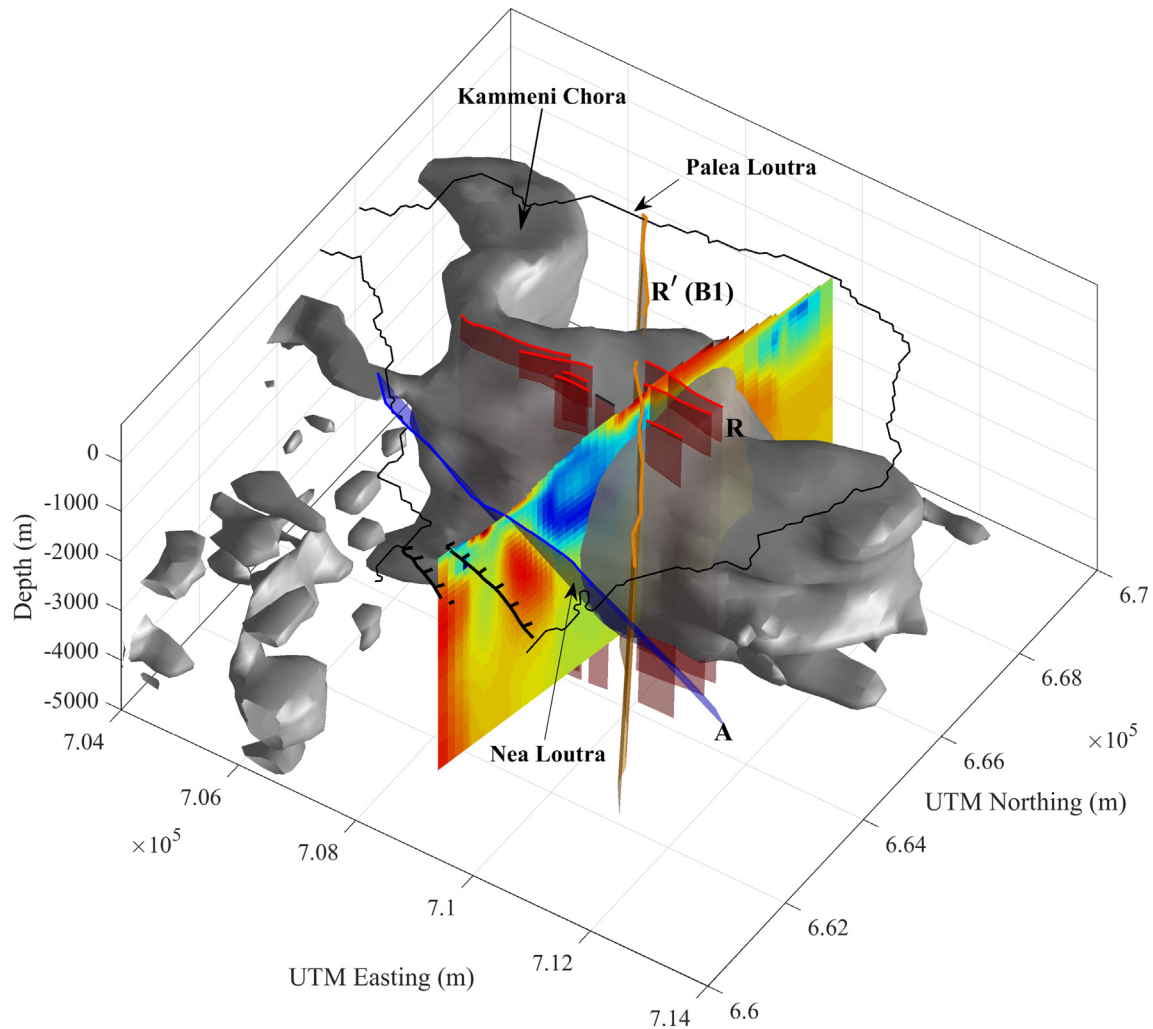


Fig. 15. Combined 3-D rendering of the magnetotelluric cross-section T1, and the 3-D susceptibility model again presented in the form of the 25×10^{-3} isometric surface enclosing higher susceptibility values. The most significant faulting features are superimposed as follows: R are the E-W R-shears discussed in Section 2.2; R' or B1 is the most significant R'-shear discussed in Sections 2.2 and 3.1; A are the WNW-ESE, north-easterly dipping normal boundary faults interfacing the Argolis peninsula and the Saronic Gulf.

context of the operative crustal stress field. However, their “downgraded” role indicates that their overall contribution to the evolution of Methana Volcano may not be as significant as appraised by Pe-Piper and Piper (2013). Overall, the crustal stress field probably predated the inception of MV in a cause and effect sense. Thus, instead of changes in regional tectonics, the evolution of geochemical characteristics, eruptive style and abundance of volcanic products discussed by Pe-Piper and Piper (2013) could be explained in terms of opportunistic exploitations of time-local adaptations of the same stress field, by magmas seeking the easiest way to the surface.

Focusing now on the geothermal resource, we point out that the magnetotelluric data detected what appears to be a geothermal reservoir with resistivity of the order of $0.3 \Omega\text{m}$, located underneath the central domes (Chelona, Chionessa and Loutsas in particular) at a depth of approximately 1 km below sea level. This is presumably replenished through the weak permeable zone created by the intersection of the R and R' shears and branches upwards, to the south through the interface between the Alpine country and Quaternary volcanic rocks, and to the north through the R and R' shears. The former branch appears to be more voluminous and conductive ($0.3\text{--}1 \Omega\text{m}$) than the latter ($\sim 3 \Omega\text{m}$). As evident in Fig. 15, the reservoir is apparently hosted by volcanic rocks with relatively low magnetic susceptibilities, of the order of $0.003\text{--}0.01$. In the case of andesitic–dacitic rocks, the combination of such susceptibilities and resistivities would indicate highly porous/

permeable zones and hydrothermal alteration (argilization) that has significantly depressed the magnetic susceptibility. In the particular setting of this geothermal reservoir, the high permeability can be understood as a result of high fragmentation by cooling during the emplacement of the volcanic host rocks and, mainly, by the persistent tectonic (faulting) activity.

From a geothermal resource point of view, our analysis confirms previous work that anticipated the existence of thermal fluids with temperatures greater than 120°C at significant depths (e.g. Geotermica Italiana, 1984; D'Alessandro et al., 2008); this reservoir may have been located at depths of approximately 700–1500 m beneath the Loutsas dome. We have also confirmed the prediction of Geotermica Italiana, that prior to discharging the fluids have circulated in a substratum of carbonate rocks, or were stored in a reservoir within carbonate rocks: it is apparent in both Figs. 10 and 15 that the interface between the Alpine carbonate and Quaternary volcanic rocks (fault A) extends to the ‘bottom’ of the reservoir, and that through this interface thermal fluids ascent in direct contact with carbonate rocks and circulate discharging at the significant thermal springs of Nea Loutra. The second most important fluid circulation zone appears to be a $320^\circ\text{--}330^\circ$ faulting structure designated as B1 in Figs. 8, 9 and 13–15, which belongs to the local group of R'-shear faults; this appears to facilitate direct permeable communication between the geothermal reservoir and the Palea Loutra thermal springs at the north of the peninsula on one hand, and the Vromolimni, Agios Nikolaos and

(possibly) Kokkinopetra thermal springs at the east of the peninsula on the other. The common source of the thermal fluids beneath the central domes would amply explain the chemical and isotopic correlations between the thermal springs of Nea Loutra, Agios Nikolaos and Palaea Loutra, observed by Dotsika et al. (2010). The magnetotelluric data also demonstrates how seawater infiltrates at sea level and mixes with the ascending thermal fluids thus depressing their temperature to no higher than 40 °C (see Section 2.3). These findings are also consonant with the geochemical analyses of surface thermal waters by *Geotermica Italiana* (1984) which determined that prior to discharging thermal fluids have undergone extensive mixing with cold seawater.

In a final step, we note that owing to the temperature of the reservoir and the composition of the host rock, the argilization can be expected to correspond to the phyllic alteration zone at the lower levels of the reservoir, where the main products should be less hydrophile sericite-quartz assemblages; it is also entirely possible that argillic alteration products are to be found at the lower-temperature upper levels of the reservoir and within the circulation zones where the more hydrophile illite-montmorillonite-kaolinite assemblages should predominate and significantly depress the effective resistivity. Without any means of estimating the volumetric distribution of clay minerals in the reservoir and circulation zones, it is not possible to evaluate the effective porosity, hence the capacity of the reservoir. It is, however, possible to provide an upper limit to a fair approximation. Based on the geochemical analyses, it is safe to assume a liquid fraction with salinity at least equal to that of seawater and having a temperature dependence of $\rho_w = (3 + T^{\circ}\text{C} / 10)^{-1}$ (Lee et al., 1983). Assuming that the average effective resistivity ρ_e of the reservoir formations is 1 Ωm (including the effect of clay minerals), the effective formation factor $F = \rho_e / \rho_w$ is approximately 15 for $T = 120^{\circ}\text{C}$ and 20 for $T = 170^{\circ}\text{C}$. This ensures effective porosities of the order 14%–10% according to the experimental evidence of Keller et al. (1978). Albeit a gross approximation, this still indicates that the geothermal system of Methana Volcano may provide a sustainable source of energy, at least where low-intensity applications are concerned (e.g. heating), and therefore comprises a target worthy of further investigation.

CRedit authorship contribution statement

A. Tzanis: Conceptualization, Methodology, Software, Validation, Formal analysis, Investigation, Writing - original draft, Writing - review & editing, Visualization, Supervision, Project administration. **A. Efsthathiou:** Methodology, Formal analysis, Investigation, Writing - original draft. **S. Chailas:** Methodology, Formal analysis, Investigation, Data curation, Writing - original draft, Visualization. **E. Lagios:** Methodology, Validation, Investigation, Writing - original draft, Supervision. **M. Stamatakis:** Methodology, Validation, Investigation, Writing - original draft.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary Information

Supplementary Information to this article can be found online at <https://doi.org/10.1016/j.jvolgeores.2020.107035>.

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