Deep structure investigations of the geothermal field of the North Euboean Gulf, Greece, using 3-D local earthquake tomography and Curie Point Depth analysis

V.K. Karastathis a,⁎, J. Papouliab, B. Di Fiore a, J. Makris c, A. Tsambas b, A. Stampolidisd, G.A. Papadopouluse a

a National Observatory of Athens, Greece
b Hellenic Centre for Marine Research, Athens, Greece
c GeoPro GmbH, Hamburg, Germany
d Aristotle University of Thessaloniki, Thessaloniki, Greece

Abstract

New findings on the deep origin of the geothermal field and volcanic centres at the North Euboean Gulf (or North Evian Gulf), Central Greece, were obtained by combining a three-dimensional traveltime inversion of microseismic data recorded by an on/offshore local seismic network with a Curie Point Depth analysis based on aeromagnetic data.

A magma chamber was detected from low seismic P-wave velocity values and high Poisson ratios at depths below 8 km also coincident with a Curie surface estimated at 7–8 km depth.

Furthermore, it was also observed that local geothermal anomalies are generated by hydrothermal flux facilitated by NW–SE and NE–SW oriented faults. Microseismic activity is also associated with these fault systems.

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1. Introduction

Along the North Euboean Gulf coast (or North Evian Gulf), central Greece, several significant geothermal sites exist. Representative examples are the thermal springs of Aedipsos, Yaltra, Ilia, Kamena Vourla, Thermopylae and also the Quaternary–Pleistocene volcanoes of Lichades and Vromolimni (Fig. 1). In order to understand the origin of these thermal fields and their relation to the regional tectonics (see the map of Fig. 1) we performed a 3-D local earthquake tomography study and a Curie Point Depth analysis based on aeromagnetic data. The traveltime inversion method of microseismicity data for velocity tomography analysis has been extensively used in the past, to investigate and map volcanic structures (e.g. Thurber, 1984; Lees, 1992; Laigle et al., 2000; Haslinger et al., 2001; Sherburn et al., 2003; Husen et al., 2003, 2004; Lees, 2007; Battaglia et al., 2008; Waite and Moran, 2009; Jousset et al., 2011). Also defining depth points of the Curie temperature surface from aeromagnetic data has been extensively used for studies of magmatic phenomena (e.g. Tsokas et al., 1998; Ates et al., 2005; Dolmaz et al., 2005).

In the frame of the “AMPHITRITE”- project, combined onshore/offshore network consisting of 23 land-stations and 7 Ocean Bottom Seismographs (OBSes) was deployed for a 4 month period (June–October 2003) to record the microseismic activity. The instruments used for both land and OBS stations were of Geopro SEDIS IV type. This type of dataloggers provides 6-channel continuous recording with 24 bit A/D conversion and dynamic range of up to 120 dB.

Papoulia et al. (2006) presented the seismicity results (Fig. 1) based on an 1D P-wave velocity model (Table 1) and the “Hypoinverse” algorithm (Klein, 2002).

More than 2000 earthquakes ranging from Ml 0.7 to 4.5 recorded by a minimum of a 6 stations per event were located. Although most of the observed seismicity was due to an earthquake swarm that occurred at the southern part of the area, there were also many events close to sites of geothermal interest like those of Kamena Vourla and Yaltra.

In the following we present how the seismic traveltime tomography detected velocity anomalies (low Vp and high Vp/Vs ratio) below the volcanic area of the North Euboean gulf that could be related with possible magmatic intrusions. The Curie Point Depth analysis based on aeromagnetic data was complementarily performed to verify that the temperature at the depths of these anomalies was very high (>580°). The methods used and the results are extensively discussed in the next chapters.

2. Geological setting – volcanism of the study area

The geographic distribution of volcanic centres in the region of central Greece is illustrated in Fig. 1. Geological and radiometric...
dating have shown that the volcanism of central-east Greek mainland and North Euboea is of Miocene to Holocene age (Pe-Piper and Piper, 2002). In fact, a cluster of volcanic centres concentrated in this area and described by several authors. Most of the volcanic centres are geologically young given that all evidence they were activated during Pliocene and Quaternary (Fytikas et al., 1976; Bellon et al., 1979; Papadopoulos, 1982, 1989; Pe-Piper and Piper, 1989, 2002). The Orio and Oxylithos volcanoes, eastern part of Euboea island, were dated to be of Lower Miocene age. Of questionable age is the volcanic centre at Chronia (Thorio) in NW Euboea. Initially it was considered to be of Neogene (Aronis, 1955) or Quaternary age (Tataris, 1960; Papastamatiou, 1961). Stratigraphic considerations however, of Papayiannopoulou (1971) suggested that this volcano is at least of Pre-Sarmatian age. Katsikatsos et al. (1980) have mapped the outcropping volcanic rocks as altered andesite of Lower Miocene age. They reported that it seems to underlie the fine-grained members of the previous Neogene system.

In addition to volcanism, various fields of hot springs with surface temperatures ranging from 39° to 83° are situated along both coasts of the North Euboean Gulf (Fig. 1): Kamena Vourla (43°), Thermopylae (41°), Yaltra (43.5°), Aedipsos (83°) and Ilia (39°) (see also Papadopoulos, 1982 and Gioni-Stavropoulou, 1983).

In earlier studies (e.g. Fytikas et al., 1976; Bellon et al., 1979) it was suggested that the Plio-Quaternary volcanism in central-east Greek mainland and the island of Euboea belongs to the inner part of the South Aegean volcanic arc. The associated seismicity occurring at intermediate depths of around 150 km results from the subduction of the Ionian oceanic lithosphere below the Hellenic arc. The origin of this volcanism has been disputed by Papadopoulos, 1982, 1989, because the volcanism in central-east Greece has different geophysical and geochemical features from that of the South Aegean Sea. In fact, in the South Aegean volcanic arc the geochemical type of lavas is typical calc-alkaline. In the central-east Greek mainland, chemistry of the lavas ranges from calc-alkaline to weak calc-alkaline and to weak alkaline. Affinity and earthquake foci are much shallower than those of the Cretan Sea.

In the North Euboean Gulf region the Triassic–Jurassic limestones of the Sub-Pelagonian zone overlay the Permo-Triassic volcanosedimentary complex (see the geological map of Fig. 2). This, in turn, is underlain to an ophiolitic nappe tectonically emplaced over the Sub-Pelagonian sediments in the Late Jurassic to Early Cretaceous. The ophiolitic nappe is outcropping at many places of the study area with peridotite or other ultramafic rocks (area between Limni and Mantudi, Vassiliki, SW of Malesina Peninsula, Kallidromo etc.). Younger sediments of Pliocene to Quaternary age are widely extended in this region. The Plio-Pleistocene, Pliocene and Mio-Pliocene sediments consist mainly of sands, clays, marls and conglomerates. Additionally, there are few basins in the Maliakos Gulf, Atalanti and Istria covered by Holocene alluvial deposits. Finally, at the volcanic centres of Lichades, Kamena Vourla and Chronia there is andesitic outcropping due to the past volcanic activity.

![Map of the central-east Greece presenting the volcanic centres and the sites of the most important thermal springs. The dashed line box indicates the study area. The seismograph stations and the initial epicentres (Papoulia et al., 2006) of the recorded seismicity are also depicted in the map.](image)

**Table 1**

| Velocity model used by Papoulia et al. (2006) for the hypocentral location. |
|-----------------------------|-------------------|
| Local velocity model Vp     | Depth (km)        |
| 4.0                         | 5                 |
| 6.3                         | 20                |
| 6.7                         | 32                |
| 8.0                         | sub-Moho          |
Fault plane solutions of strong earthquakes indicate that in the study and surrounding areas the stress field in the crust is mainly tensional trending NW–SE and also NE–SW. Neotectonic studies showed that the major fault structures in the north Euboean Gulf is that of Atalanti consisting of several segments of normal faults, trending about NW-SE and dipping NE (indicated by AFZ on the map of Fig. 2), with a total length of about 20–30 km (Pavlides et al., 2004). The large earthquakes of 21 and 27 April 1894 (M=6.4 and M=6.9) were associated with this fault system. However, fault plane solutions of smaller earthquakes showed that in the north and central Euboea, as well as offshore Euboea in the Aegean Sea, right-lateral strike slip motions are evident. Geological mapping has not confirmed right-lateral strike slip motion (Hatzfeld et al., 1999; Kiratzi, 2002).

In summary, the up to date geological knowledge about the North Euboean Gulf is limited to the geo-chronological and geochemical study of local volcanic centres and the associated thermal spring. The present study aims the investigation the deep thermal sources (main characteristics and location) and their connection to the surface springs and their hydrothermal flux.

3. Methodology and data processing

3.1. Traveltime inversion of microseismicity data

To build the 3D seismic velocity structure for the investigation area we implemented the algorithm SIMULPS14 (Thurber, 1983; Eberhart-Phillips, 1990; Haslinger and Kissling, 2001). The code performs simultaneous inversion of the model parameters Vp, Vp/Vs and hypocenter locations. The software is actually an advancement of the classical SIMULPS algorithm of Thurber (1983), Eberhart-Phillips (1990) and Evans et al. (1994). Besides the standard “ray-bending” methodology for ray-tracing, an option of a full 3-D shooting ray tracer developed by Virieux and Farra (1991) has also been added (Haslinger and Kissling, 2001). The SIMULPS14 code uses a linearised damped least-square inversion to solve the non-linear problem of the hypocentral location and velocity model. The parameterisation of the velocity model is accomplished by considering of a 3D grid of velocity nodes (Vp and Vp/Vs) with linear interpolation of the velocity values in the intermediate space between the nodes.

Due to the non-linear nature of the problem, the initial velocity model and the initial hypocenter locations in the inversion procedure must be chosen as close as possible to their true values. A good approximation can be derived by adopting the “1-D minimum velocity model” resulted from the VELEST algorithm (Kissling et al., 1994; Kissling, 1995). Thus, both P and S wave arrivals were inverted at first with the algorithm VELEST to calculate the “minimum 1D velocity model” and then with the 3D linearised tomography SIMULPS14 code and determined the 3D velocity structure of the study area.

The P- and S wave phases used by the seismic tomography were picked manually and quality weights were attributed to them. Quality weights of 1.0, 0.75, 0.5, 0.25 were set according to assumed picking uncertainties of 0.03, 0.06, 0.09, 0.12 s respectively. Great effort was put in accurate and consistent picking since, this affects the quality of
the tomographic processing (Diehl et al., 2009; Husen et al., 2009). An example showing the estimation of the picking uncertainty is given in Fig. 3. The picking error in this specific example is not higher than 0.03 s and 0.05 s for the P and S-wave phases respectively. However, in more distant stations picking ambiguity increases. Average picking error for the complete dataset has an RMS value of 0.03 s for the P-waves and the 0.06 s for the S-waves respectively. Simultaneous examination of the three components assisted in separating the pure S-wave from the SP converted phases.

In order to build a reliable minimum 1D velocity model we based only on the most reliable events depending on the number of phases. We selected 481 events observed at a minimum of 8 stations per event. The azimuthal gap (GAP) was chosen to be lower than $180^\circ$ (GAP $<$ 180$^\circ$). The maximum GAP selection criterion is usually used in the application of VELEST (e.g. Husen et al., 2003; Raffaele et al., 2006; Battaglia et al., 2008; Maurer et al., 2010). We also selected the Vp/Vs initial values to be $1.78 \pm 0.02$ based on the Wadati diagrams of the present dataset and previous studies from the wider region (Hatzfeld et al., 1999). For the Wadati diagrams we selected events with the best fitting ($\text{rms error} \leq 0.6$, correlation coefficient $\geq 0.95$). The VELEST algorithm simultaneously inverted the hypocentral parameters and the velocity model in order to calculate the model that gives the optimum solutions (1-D minimum model). The initial and final model of the 1-D inversion is presented in Fig. 4a. To select a proper initial input model of VELEST we tested a variety of models (Fig. 4b), some based on results of seismic surveys previously contacted in the area (Makris, 1978; Makris et al., 2001) and some others used in previous seismological and microseismicity studies in the region (Panagiotopoulos et al., 1985; Papoulia et al., 2006). The figure also shows the respective results for each model. The model “mod1” gave the lowest rms residual and it was very close to the model “mod3”, resulted with the initial model based on the seismic profiles of Makris et al. (2001). By the use of the algorithm “HypoInverse” (Klein, 2002) we relocated the events on the basis of “mod3” velocity model. The relocation shifted the events to concentrate in better shaped clouds (see supplementary data — Fig. S1).

The procedure and parameters followed for the 1-D inversion were according to the guidelines given by Kissling et al. (1994). In fact, the procedure was materialized through some steps as the establishment of an a-priori model, the setting up of the geometry of the models, the selection of the events, the calculation of 1D minimum velocity model and finally the evaluation of the solution. The station corrections used in the inversion were in agreement with the local geology (For technical details see Fig. S2 of the electronic supplement).

The three-dimensional inversion required as basic inputs the initial velocity model, the traveltime P-wave and S-wave phases of the recorded microseismicity and the stations characteristics. The number of the events increased in this procedure to 540 with the selection of those with minimum number of P-wave phases equal to 6. In the selection criteria the azimuthal gap (GAP) was accepted to be lower than 180$^\circ$.

The grid adopted was coarse enough to ensure dense ray coverage in most of the cells (Fig. 5, Fig. S3). The interval between the nodes in the SW-NE direction was 9 km, in SE-NW direction it was 13 km and 4 km were chosen for the vertical, except the first layer that was 3 km. A finer grid was also tested with horizontal spacing between the nodes to be half that of the coarser grid. The final decision about the grid was taken after examination of the hits count distribution (Fig. S3, S5) and derivative weighted sum (DWS) (Figs. S4, S6), parameters that reflect the ray coverage and the robustness of the solution. Actually, the distribution of hit counts shows the number of

\[\text{Fig. 3. An example of a seismic record of a local event. The picks of P-wave and S-wave arrivals are presented with pink coloured lines. The time intervals of 8–10 s and 12–14 s of z and n components have been enlarged and presented as embedded pictures. The three red lines present from left to right the earliest possible pick, the most probable pick and the latest possible pick. The picking error of the P-wave phase was small, less than 0.3 s. The respective error for the S-wave phase was higher but less than 0.5 s.}\]
rays passing through each grid node (hit count is the sum of the number of rays that contribute to the solution at one node). On the other hand, the derivative weighted sum (DWS) shows the ray density in the neighbour area of a model node. Ray density is a measure of the normalized total length of rays in a unit volume (Koulakov et al., 2010). The comparison between the two grids showed that the finer grid presented a heterogeneous pattern of the derivative weighted sum (DWS) and poor hits count distribution. Therefore, the interpretation of the results was based on the coarse grid.

The damping factor parameter value was chosen with the aid of the trade-off curve between the model variance and data variance (Eberhart-Phillips, 1986). For the damping factor of the Vp the curve pointed to the value 50 and for the damping factor of Vp/Vs the respective curve pointed to 15. These values decreased the data variance adequately without causing considerable increase to the model roughness. The inversion was more stable when the station corrections were included in the procedure. The instability was caused mainly because the distances between some stations were significantly smaller than the 9 km distance of the grid spacing (i.e. the 10 pairs of stations from the 870 ones have distances between 5 and 6 km). Since the geology of these station sites was rather different, the delays were too high for so short distances causing instability to the inversion (Husen et al., 2003; Husen S. personal communication, 2010).

The inversion was sufficiently effective since after four iterations led to a data variance reduction for the Vp model was 70% and for the Vp/Vs solution 40%.

The final P-wave velocity model and the ratio Vp/Vs are shown in Figs. 6 and 7 respectively. Vertical sections of the resulted Vp model are presented in Fig. 8. The Poisson ratio as derived from the calculated Vp/Vs ratio (Eq. (1)) is also shown in the electronic supplements, Fig. S7. Although Poisson ratio (ν) is absolutely equivalent to the Vp/Vs ratio, we included these sections in the supplementary data, since the majority of the Earth scientists and engineers are more familiar with the former especially when it is considered as an indicator of the plastic condition of materials. Figs. 9 and 10 show the respective Resolution Diagonal Elements (RDE) distributions for Vp and Vp/Vs models.

\[
\frac{V_p}{V_s} = \sqrt{\frac{1-\nu}{\frac{1}{2}-\nu}} \quad \text{or} \quad \nu = \frac{\left(\frac{V_p}{V_s}\right)^2 - 2}{2\left(\frac{V_p}{V_s}\right)^2 - 2} \quad (1)
\]
The yellow lines in Figs. 6 and 7 imply the reliability limit that we finally considered based on the comparison between the graphical representations of the diagonal elements of the resolution matrix (RDE) and the recovery ability of synthetic models resembling the real earth model (Husen et al., 2000). More specifically we produced synthetic traveltime data for all the seismic events of our dataset with a given synthetic model similar to the "expected" earth structure according to the processing results of the real data. Subsequently, starting with a 1D model we tried to recover the synthetic input model by applying inversion with the same parameters as those in the real case (see the electronic supplement Fig. S8a, b). Artificial Gaussian distributed noise was added to the synthetic travel times with a standard deviation proportional to the observational weight. Comparison between the synthetic input and the resulted model delineated areas of good and poor resolution. After comparing the results with the diagonal elements of the resolution matrix we realized that beyond the 0.1 curve there was no substantial information. Although in the space defined by the 0.1 resolution curves the model was in general well resolved, there were specific areas that were not well recovered in the synthetic tests (the poorly resolved areas are defined by yellow dashed line in Fig. S8a).

Similar evaluation was also undertaken for the Vp/Vs inversion (Fig. S9). In this case the limit was set quite higher, at a value of 0.2, since the resolution values were also higher. Nevertheless, since RDE values are strongly influenced by the selection of the damping factor, it is difficult to define a general limit of reliability. The smaller the damping factor is, the higher are the RDE values. This does not imply however, that the reliability of the solution has been improved. The reliability depends mainly on the geometrical distribution and density of the rays (ray coverage) (Husen et al., 2003).

In general the resolution was adequate for all the sections above the one of the 20 km depth. At this depth the resolution was poor since only few deep events have been recorded (see also Fig. 5).

3.2. Aeromagnetic data

The Curie point (approximately 580 °C for magnetite at atmospheric pressure) is the temperature at which spontaneous magnetization vanishes and magnetic minerals exhibit paramagnetic susceptibility. Because paramagnetism is much smaller than spontaneous magnetization, rocks are essentially non-magnetic at temperatures greater than the Curie point of magnetite. We used “Curie Point Depth” (CPD) to describe the depth to the inferred Curie point transition of magnetite. Magnetic data have been used at various parts of the world in order to estimate Curie point depths (Okubo et al., 1985, 1989; Tsokas et al., 1998; Stampolidis and Tsokas, 2002; Stampolidis et al., 2005; Ross et al., 2006).
The computation of the depth to the bottom comprises one of the most difficult problems in potential field inversion (Blakely, 1995). The signal from the top of a magnetized body dominates the signal from the bottom at all wavelengths. This was recognized from the fundamental work of Spector and Grant (1970), who established the well known method for estimating the average depths to the top of magnetized bodies, based on the slope of the log power spectrum. However, the shape of the radially averaged spectrum also offers means to obtain an estimate of the depth to the bottom.

Okubo et al. (1985, 1989) provided a method to obtain the depth to the bottom, \( z_b \), of the deepest magnetic sources in two steps: first, obtain the depth to the centroid, \( z_c \), and second, determine the depth to the top, \( z_t \), of the deepest magnetic sources. The depth to the bottom, \( z_b \), is then calculated from those two depths: \( z_b = 2z_c - z_t \).

The magnetic dataset for the study area came from two different aeromagnetic surveys, which were carried out in 1966 and 1978 by two different surveying companies ABEM-Elektrisk Malmetning and Hunting Ltd, on behalf of IGME (Greek Institute of Geology and Mineral Exploration). ABEM in 1966 covered the Euboea’s island area with flight lines flown at constant terrain clearance of 275±75 m above ground level, with a data resolution not more than 1 nT and flight directions NW–SE oriented. Hunting Ltd in 1978 covered a complementary wider region with flight lines positioned at a constant altitude of 300 m above the ground level, with a data resolution less than 1 nT and the directions of flights WSW–ENE.

From the aeromagnetic datasets, the International Geomagnetic Reference Fields (IGRF) 1966.5 and 1978.5 respectively were removed, before to be sutured in one dataset and then gridded using a minimum curvature algorithm (Webring, 1982) with a constant spacing equal to 0.3 km. The residual dataset we obtained, was a rectangular area of about 110 km × 220 km NW–SE oriented. To avoid the effects of the irregular topographic surface the residuals were upward continued to the horizontal level of 3 km (a.s.l.) employing the chessboard method of Cordell (1985). Fig. 11 shows the total upward continued anomaly field of the area. Finally to emphasize the deep magnetic anomalies a band pass filter (10–50 km) was applied on the residual dataset that removed the long-wavelength topographic components and the high frequency noise (Fig. 12).

To estimate the CPD, we followed the procedures described by Stampolidis et al. (2005) and we considered five overlapping squared study areas of 100 × 100 km with centres shown in Fig. 12 as C1 up to C5. The overlapping of these subregions was 70%. In each subregion the radially averaged power spectra was computed using the program MFINIT from Phillips (1997). The depth to the top \( z_t \) and the centroid
depth $z_0$ of the deepest magnetic sources for each subregion were computed. An example concerning the procedure for the C1 square area is presented in Fig. 13.

Using the formula $z_b = 2z_0 - z_t$ (Bhattacharyya and Leu, 1975, 1977; Okubo et al., 1985) the Curie Point Depth values were derived. Actually using the MFINIT program (Phillips, 1997) we calculated the log power spectrum, obtained $z_0$ and $z_t$ as described in Stampolidis et al. (2005) and finally presented them in Fig. 13. The calculated top and centroid depths have to be reduced to the sea level surface by subtracting 3 km, as these values correspond to the upward continued field. The estimated CPD values are reported in km below the sea level (Table 2).

4. Interpretation of the results

The most important structural feature detected by the three-dimensional local earthquake tomography was a low velocity anomaly (relative to the initial 1D reference model) below the region of North Euboean Gulf at depths of 8–12 km. The negative velocity anomaly is obvious in both sections of 8 and 12 km depth (Fig. 6). The resolution at the 16 km depth slice does not allow a clear picture on the continuation of this anomaly at depth. The data indicate however, that its continuation to a greater depth is very probable.

In Fig. 14a, b we correlated the 12 km depth anomaly with the locations of the volcanic and geothermal centres. It can be easily seen that a low velocity zone is closely confined below these centres having its local minimum at the site of “Chronia” volcano. The anomaly zone has also high $V_p/V_s$ ratios, in the order of 1.82 or higher and Poisson ratios in the order of 0.28 or higher — (see also Fig. S7). Local maxima are at the volcanic centres of “Vromolimni” and “Chronia”, where the values of $V_p/V_s$ exceed 2 or 0.33 Poisson ratio.

At the shallower depth of 8 km the picture is very similar. The anomaly extends to the Malesina Peninsula. The recovery tests showed that the north-western part of the investigated area covering the regions of Kamena Vourla, Vromolimni, Lichades and Yaltra was not adequately resolved (Fig. S8). This part of Fig. 14c was therefore kept shadowed.

It is also interesting that with decreasing depth from the 12 km level to that of 8 to 7 km, the areal extend of the high Poisson ratio values decreases. This indicates that at 7 to 8 depth we have mapped the top of the magnetic intrusion. This is also confirmed by the magnetic evaluation that placed the top of the Curie surface (580°) at the same depth. High Poisson ratio values at shallower
levels than the magma top are of local significance and indicate zones of hydrothermal circulation or magma intrusions in tectonic faults.

As Benz and Smith, 1984; Miller and Smith, 1999 and Husen et al., 2004 have demonstrated, magmatic intrusions in crustal levels are usually associated with depressed Vp velocity values and high Poisson ratios. The results of our tomography and the aeromagnetic data that permitted to locate the Curie surface are in agreement with above mentioned references and confirm our conclusion of a magmatic intrusion below the north Euboean gulf.

The very anomalous Vp and Poisson values located below Aedipsos area and also north of Chronia areas, indicate very high temperature fields and magma intrusives that have significantly influenced the rigidity of the lithological units. Particularly the area of Aedipsos seems to be of great geothermal interest.

Hatzis et al., 2008 evaluated the geothermal capacity of the Aedipsos region from temperatures of springs and at shallow boreholes and showed that the geothermal gradient at this area was seven times higher than normal. A temperature of 80°C was reached at a depth of 350 m only. During the same project of IGME (c.n.

**Fig. 8.** Vertical sections of the resulted 3-dimensional P-wave velocity model. The locations of the vertical sections are shown in the embedded maps.
also exposed onshore and seen in the geological map of Fig. 2. At this level we also see irregularities of Vp and Vp/Vs distribution beneath the gulf area that follow the orientation of the main faults zones like those of Atalanti–Kallidromon (AFZ–KDFZ in Fig. 2) and Kamena Vourla (KVFZ in Fig. 2) (see also Ganas and Papoulia, 2000; Palyvos, 2001; Pavlides et al., 2004; Karastathis et al., 2007; Tzanis et al., 2010) and the faults of the northern part of Euboea island, at the Lichades peninsula (see Fig. 2 and Galanakis et al., 1998). Some of these faults are also seismic active as seen e.g. in Fig. 15 along profiles LL′, PP′, BB′, and FF′. This indicates, that the anomalous values of Vp and Vp/Vs mapped in the 4 km slice which is dominated by the velocity values of the limestones (4–4.8 km/s), have developed as a consequence of fracturing and deformation of major faults.

5. Discussion

A better resolving of the tectonic features at the top slices of the seismic tomography would require a denser grid. However extensive testing on the grid selection showed that this was not possible with the given network and the recorded seismicity. The depth frequency plot of the relocated seismicity (see also Fig. S1) showed a linear increase for the first 6–7 km and an exponential decay after the depth of 7–8 km (Fig. 16). This indicates

![Fig. 9. Resolution Diagonal Elements (RDE) distributions for Vp.](image)
that the brittle–ductile transition zone is at this depth. Therefore after this depth the seismicity is controlled mainly by the temperature. The seismicity practically stops after the depth of 22 km. This distribution of the hypocentral depths is in accordance with the results of the seismic tomography and Curie Point Depth analysis as far as the depth of the thermal source is concerned.

Although the present study revealed the main characteristics of the structure of the geothermal field of Central Greece, it did not succeed to give its complete picture. A more sedulous study of the geothermal field would require the extent of the investigation area northern (to Thessaly) and eastern, in order to include in the study some additional volcanic centres. This could assist the correlation of the volcanism of N. Euboean Gulf with the volcanoes of Thessaly which are of Pleistocene age. The study showed that the investigation of the deep structure, which is related with the volcanism of central Greece, is possible with the local earthquake tomography techniques and therefore it would be useful a similar study in Thessaly. It would be of great interest to investigate if the detected anomalies of North Euboean Gulf are continuing to the North.

6. Conclusions

We combined the local earthquake tomography method with evaluation of aeromagnetic data and obtained the depth and the characteristics of the thermal source of the North Euboean Gulf geothermal field.

The analysis of the aeromagnetic data provided us with the Curie temperature isotherm (580 °C). This surface compared within accuracy limits with velocity Vp values and Vp/Vs ratio developed from the 3D seismic velocity inversion. We have shown that the magnetic and the seismic results defined a zone of high temperature at 7 to 8 km depth which we have interpreted to be the top of a magmatic intrusion that is the source of the volcanic phenomena of this region.

We could also show that hydrothermal activity which is widely distributed of the northern part of the Euboean Gulf is clearly linked to faults closely linked to the magmatic intrusives.

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Fig. 11. Total magnetic field after International Geomagnetic Reference Field (IGRF), reduction to the pole (RTP) and CHESS corrections (the data were continued to a common observation surface of 3000 m above ground using the chessboard method of Cordell (1985); program CHESS (Phillips, 1997)). The four white “corners” define the area where local earthquake tomography was performed.

Fig. 12. Aeromagnetic map of the wider region. The estimated CPD depths have been also depicted in their positions. The four white “corners” define the area where local earthquake tomography was performed.
Appendix A. Supplementary data

Supplementary data to this article can be found online at doi:10.1016/j.jvolgeores.2011.06.008.

References


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Table 2

<table>
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<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
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<td>7.3</td>
<td>8.0</td>
<td>7.80</td>
<td>8.1</td>
</tr>
<tr>
<td>Error (km)</td>
<td>±0.5</td>
<td>±0.7</td>
<td>±0.5</td>
<td>±0.4</td>
<td>±0.5</td>
</tr>
<tr>
<td>Quality</td>
<td>A</td>
<td>A</td>
<td>B</td>
<td>B</td>
<td>C</td>
</tr>
</tbody>
</table>

Fig. 13. Example of CPD estimated depths for Cell 1 over band-pass filtered (10–50 km) data set.

Table 3

| ZB estimates for each subregion with quality value assigned to the estimates (A = excellent, B = good, C = fair). The assigned quality index is related to the original data density in each subregion. |
|---|---|---|---|---|---|
| Subregions | 1 | 2 | 3 | 4 | 5 |
| $z_i$ (km) | 7.0 | 7.3 | 8.0 | 7.80 | 8.1 |
| Error (km) | ±0.5 | ±0.7 | ±0.5 | ±0.4 | ±0.5 |
| Quality | A | A | B | B | C |


Fig. 14. Correlation of the detected anomaly with the position of the local volcanic and geothermal centres, namely K: Kamena Vourla V: Vromolimni L: Lichades G: Agios Georgios Port of Lichades Y: Yaltra A: Aedipsos I: Ilia C: Chronia. The individual images show a. The P-wave velocity slice at 12 km b. The Vp/Vs ratio slice at 12 km c. The P-wave velocity slice at 8 km d. The Vp/Vs ratio slice at 8 km. e. The P-wave velocity slice at 4 km. f. The Vp/Vs ratio slice at 4 km.


Ph. D. Thesis, University of Thessaloniki, (in Greek).


Fig. 15. Relation between the relocated seismicity and the estimated velocity structure.

Fig. 16. The distribution of the depth values estimated for the relocated seismicity. For the example only the events with a minimum of 8 stations and azimuthal gap (GAP) lower than 180° (GAP < 180°) were taken into account. The relocation was accomplished by the program Hypoinverse that used the minimum velocity model.


