European Geothermal Congress 2013
Pisa, Italy, 3-7 June 2013

Geochemical and Short-Period Rayleigh Wave Dispersion Measurements as evaluation tools to estimate Madeira Island magma chambers depths

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Keywords: volcanic Island, geochemistry, geothermal, geothermometry, seismic tomography.

ABSTRACT

Madeira is an intraplate volcanic island that shows evidences of recent volcanism. In a volcanic geothermal system the heat source comes from magma emplacement at relatively shallow levels, thus knowledge of magma chamber(s) depth(s) is one of the keys to geothermal reservoir assessment. The tentative definition of these depths in Madeira is being studied by the integration of petrographic, petrological, geochemical and geophysical methods.

1. INTRODUCTION

A classical geothermal reservoir is most often found in volcanic regions with historical eruptions where the heat source is related to a volcano or volcanic complex and can be found at relatively shallow depths beneath the surface. On the contrary Enhanced Geothermal Systems (EGS) are usually related to continental areas where the reservoir is at considerable depths and must be stimulated through hydraulic fracturing.

Madeira is a volcanic oceanic island, with no historical eruptions. However the existence of recent volcanism (~ 0,006 Ma, Geldmacher *et al.*, 2000 and this study) with well preserved volcanic cones and thermal evidences, such as the occurrence of hot water rich in CO_2, is suggestive of a heat source enough to induce rocks/water with significant temperatures at depths likely to be exploitable for economic generation of electricity. This setting makes of Madeira an intermediate and innovative case study between classical volcanic geothermal systems and enhanced geothermal systems, where the heat sources can be relatively superficial and deep, respectively.

Intraplate volcanic oceanic islands resulting from a hotspot related to a mantle plume are characterized by submarine swells and high heat flow related to the rising of hot mantle material. In dormant volcanoes the magma emplacement beneath the volcanic edifice is the ultimate provider of heat to the potential volcanic geothermal system, thus it is essential to know the depths and extension of magma chambers related to the most recent eruptions. The body of magma heats up the surrounding rocks as it slowly cools off, while the fractures originated by its emplacement favour the circulation of hot fluids that may migrate to permeable rocks near the subsurface, giving rise to a geothermal reservoir. In average a magma body takes about 10 000 years to cool down, and its heat transfer can go as far as 10 km. However it depends on many variables such as original temperature, depth, size and composition of the magma body, as well as on the characteristics of the surrounding rocks.

Here we present part of the methodology for the evaluation of Madeira island geothermal potential, where geological, geochemical and geophysical methods are being applied, to a tentative determination of the characteristics of the magma body(s) emplaced beneath Madeira Island.

2. GEOLOGICAL OVERVIEW

Madeira is an intraplate volcanic island, located at the eastern North Atlantic Ocean, at the southwest end of a 700 km long volcanic chain with an emerged area of 737 km^2 and maximum altitude of 1861 m (Pico Ruivo). It constitutes, together with Desertas and Porto Santo island, the Madeira Archipelago. A major topographic swell beneath the region consistent with mantle upwelling has been identified by geophysical studies (Cazenave *et al*. 1988, Zhang and Tanimoto, 1992, Hoernle, 1995). Although still disputed the origin of the Madeira Archipelago islands has long been attributed to a hot spot trail resulting from a mantle plume, even if it doesn’t fully agree with the classical plume model (Geldmacher *et al*. 2006). A seamount SW of Madeira, extending to a height of 500 m below sea level is thought to be the present location of the Madeira hotspot (Geldmacher *et al*, 2000).

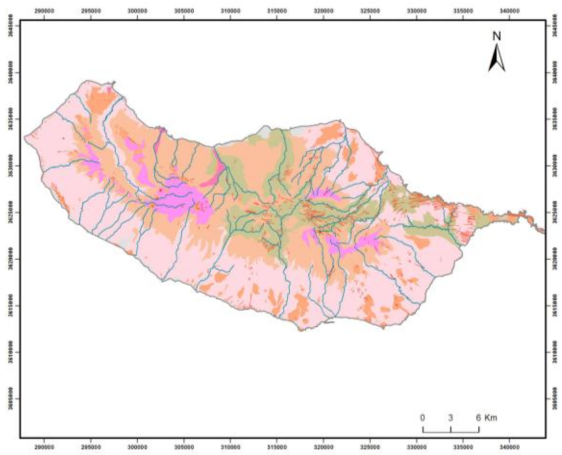

Figure 1 – Geological sketch of Madeira Island (after Ribeiro & Ramalho, 2009). -β6, -β5, -β4, -β 3, -β2, -β1

Features characteristic of volcanic rift zones, like dense dike swarms, normal faults, graben structures and series of pyrolastic cones parallel to the major axis of the Madeira (E-W) and Desertas Islands (NNW-SSE) point to a two-arm rift system. Porto Santo (NNE-SSW) Island located to the NE of Madeira can correspond to an aborted third-arm rift zone. Normally these kind of rift zones are the result of pressure induced by magma upwelling. These features are symptomatic of a favourable fracture framework to geothermal exploitation.

Generally three major volcanic stages and six volcanic complexes (β1 to β6) are attributed to Madeira (e.g. Zbyszewski *et al*. 1976, 1975; Geldmacher, 2000; Mata *et al*. 1989; Ribeiro & Ramalho, 2009). The first building phase corresponds to a transitional early shield stage that is materialized by the β1 Complex (2.5 to 5 Ma), which is the oldest, probably from the Miocene – Pliocene, and mainly made up of pyroclastic rocks, volcanic breccias and minor lava flows, which indicates an origin related to a particularly explosive phase during the formation of the island. A voluminous post-shield phase (1 – 2.5 Ma) comprising the β2, β3 and β4 volcanic complexes consisting of variable volumes of lava flows interlayered or alternating with pyroclastic material. While β2 and β3 are abundant throughout the island, β4 complex occupies a much more limited area. Finally, the post-erosional phase comprises the β5 and β6 complexes, consisting of some valley constrained lava flows in the north-northwest part of the island and to recent (0.006 Ma) to scoria cones and pyroclastic rocks and volcanic plugs and dykes (see fig.1).

The definition of the main directions of faults and dikes that may constrain fluid/heat circulation is important to identify the preferable zones for ascending fluids and is another factor to have in consideration.

3. PETROGRAPHY

Petrographic characteristics of magmatic rocks can provide important information concerning the processes they went through until their complete crystallization as well as the eventual post-magmatic processes. Petrographic, geochemical and mineral chemistry data are fundamental to know the order of crystallization of minerals, the physical-chemical conditions in which they crystallized and to which conditions magmas were subjected during their evolution.

Petrographic studies on Madeira rocks were aimed at characterizing the volcanic complexes, determining the crystallization history, detecting aspects suggestive of hydrothermal fluids circulation, and selecting samples, and, in these, the suitable crystals to perform geothermobarometric calculations.

The wide prevalence of basaltic rocks, either in effusive or in explosive phases, which gives a certain lithological monotony the island, as repeatedly referenced and already suggested by macroscopic observation, was also evidenced by petrographic analysis.

Amongst the diverse volcanic complexes there is no remarkable petrographic diversity. Holocrystaline rocks of porphyritic texture prevail, with phenocrysts (resulting from slow intra-telluric crystallization) of olivine, clinopyroxene and sometimes plagioclase in a matrix composed of plagioclase microliths, clinopyroxene, oxides and often small olivine crystals, confirms a polibaric crystallization, as is usual in these kinds of rocks.

Most olivine phenocrysts are euhedral to subeuhedral (Figure 2) being common to present corroded rims, occasionally showing corrosion gulfs resulting from reaction with the liquid. This points out to variations in the liquid composition or in pressure, temperature, or oxidation state conditions. The characteristic rim alteration into idingsite (goethite + smectite) in some cases is a symptom of a deuteric alteration under oxidizing conditions. A few olivine phenocrysts are almost completely filled with opaque fine grains, coalescent aligned (Figure 3). Although the alignment may suggest a 'melt path' type origin (Augustithis, 1978), i.e. the inclusions derived from the penetration of liquid in the growing crystals, in this case they may correspond to oxides generated by hydrothermal alteration within the crystals internal fractures.

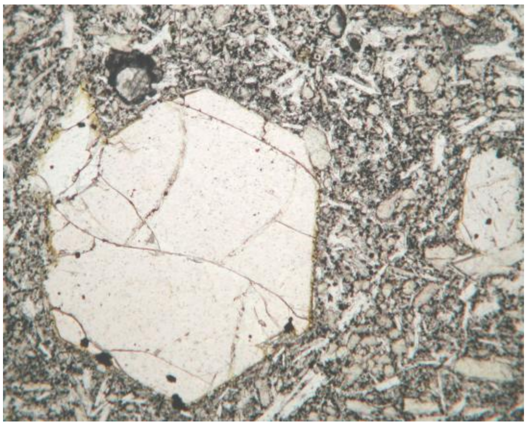

Fig. 2 – Euhedral crystal of olivine from a basaltic dike. Nicols //, 25x

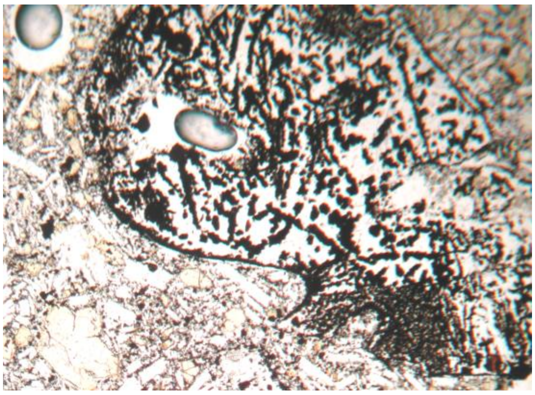

Fig. 3 – Olivine phenocrystal filled with oxide aligned coalescent grains from a b6 basalt. Nicols //, 25x

Some clinopyroxene phenocrysts centers or interior areas are replaced by matrix (Figure 4), which may correspond to a filling of pre-existing cavities in the crystals, caused by crystallization under conditions of supercooling or by partial dissolution.

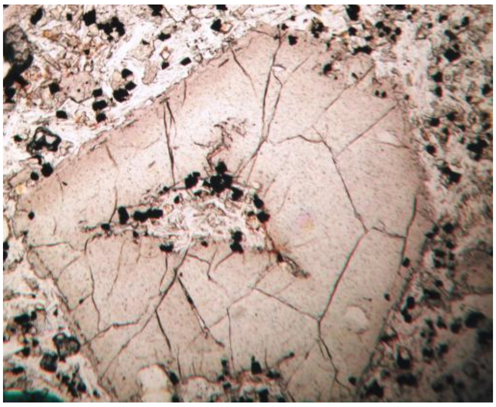

Fig. 4 – Clinopyroxene crystal with a center replaced by matrix in a basaltic rock. Nicols //, 160x

Although some samples of β1 and β3 complexes include what appear to be xenocrysts, there are no mantle xenoliths in the rocks studied, although they are reported in the literature for rocks of the complexes β2 and β3 (e.g. Munhá *et al*. 1990). Thus, there seems to be an absence of xenoliths and xenocrysts in the more recent complexes, favoring the hypothesis of the existence of crustal magma chambers (superficial), since these can act as filters to the xenoliths when transported by the magmas, preventing their ascent to the surface.

4. MINERAL CHEMISTRY

Whole rock analysis and mineral chemistry are a helpful tool, to know the crystallization temperatures and pressures/depths of magmas. Through the application geothermobarometric methods, the physical and chemical crystallization history of rocks can be unveiled, and the depths of possible levels of magma storage constrained..
In order to identify the physical-chemical parameters of mineral crystallization of the rocks related to the latest stages of Madeira volcanism, a set of mineral chemical analyses was performed, by an EPMA (JEOL JXA8200) at the FCUL (Sciences Faculty, University of Lisbon), on olivine and clinopyroxene phenocrysts from rocks of the three most recent volcanic complexes (β4, β5 and β6) of Madeira Island.

The application of the empiric geothermobarometer of Soeeso (1997), which is well calibrated for basaltic magmas and only requires the clinopyroxene chemical composition, to the obtained mineral chemistry results indicated that: a) Almost all crystallization temperatures vary between 1150 ºC and 1200 ºC; b) most clinopyroxene crystallization pressures are between 2 and 5 kb, but many fall in the fields of 6 to 10 and 11 to 15 kb; c) almost all crystallization pressures calculated for β6 complex fall in the 2-5 kb field; d) pressure of crystallization for β4 and β5 phenocrysts is more variable, with values from < 2 kb to 11-12 kb (see figs.). These results suggest fractional crystallization at various depths and can be interpreted as a consequence of the existence of at least two magma chambers. One, 20 to 40 km depth (6-12 kb) in the lithospheric mantle, and another between 7 and 16 km (2 a 5 kb) in the oceanic crust. These values partially overlap with the ones found by Schwartz *et al.* (2004) for the magma stagnation levels beneath Madeira Island where two main fractionation depths of 15 to 28 km and 8 to 10 km were calculated for the early and late Madeira rift phases. The application of other geothermobarometers based on olivine-liquid and clinopyroxene-liquid equilibrium Putirka (2004), Putirka (1999) and Putirka *et al*. (2003) will allow validating or otherwise constraining the crystallization levels already obtained, so that these parameters, important to the geothermal assessment, will be as accurate as possible.

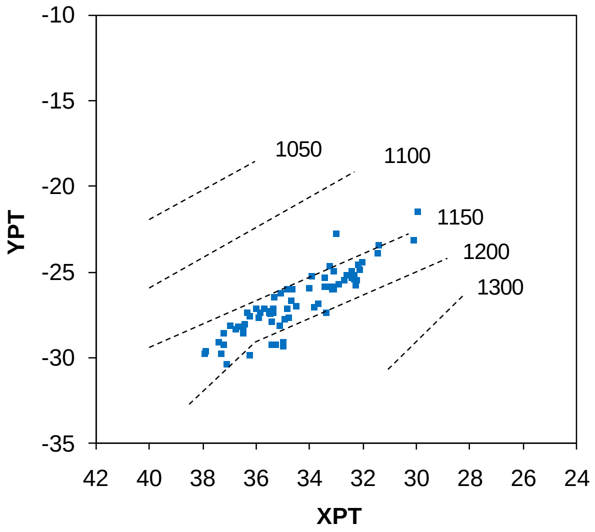

Figure 5 – Soeeso (1997) Temperature XPT – YPT Diagrams for the clinopyroxenes of β4, β5 and β6 complexes of Madeira. para a cristalização das clinopiroxenas analisadas.

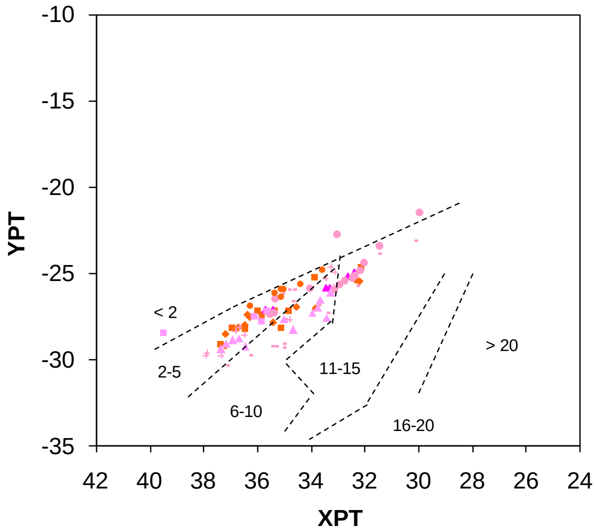

Figure 6 – Soeeso (1997) pressure XPT – YPT diagrams for the clinopyroxenes of β4, β5 and β6 complexes of Madeira. Symbols: β6 - orange -, β5 – dark rose e β4 – light rose.

5. RAYLEIGH DISPERSION MEASUREMENTS

On the last decades ambient noise surface-wave tomography has been widely used to map the crustal structure in areas of low or practically no seismicity, which is the Madeira island situation. Also, this method allows to image regions with a lateral and depth resolution that mainly depends on the size and aperture of the seismic network. In our case, and taking into account Madeira dimension (Figure 7) and the distance between station pairs, our results will be limited to the first 10 km, with an higher sensivity on the first 5 km. Therefore, it will enable to image the shallower structures.

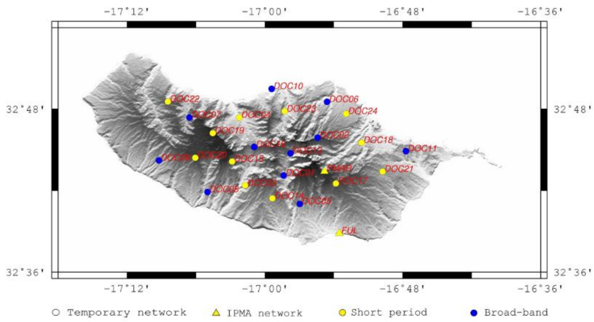

Figure 7 – Topographic map of Madeira and location of seismographic-stations. Symbols: Short period – yellow, broad-band – blue. IPMA stations are represented with a triangle and temporary stations with a circle.

To accomplish this task, 23 seismic stations have been deployed from May 2011 to September 2012. This temporary network was composed of 11 broad-band seismic stations and 12 short-period stations. It was complement with two short period seismic stations belonging to the IPMA Portuguese seismic network (see Figure 7).

Data processing has essentially followed the Bensen et al. (2007) methodology. This first step consisted of a data quality control and instrumental response correction. After removing the mean and the trend from the signal, we applied a spectral whitening to broaden the frequency band of the ambient noise signal and reduce the effect of persistent monochromatic sources.

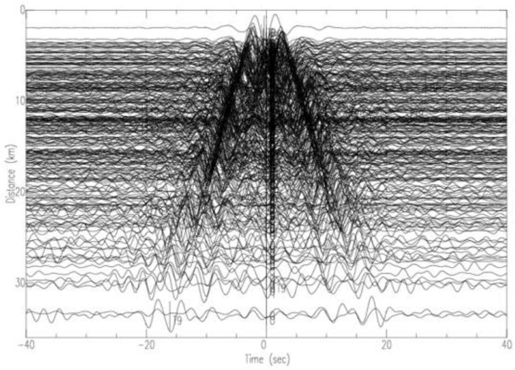

Figure 8 – Short period record section (0.5 – 5 s) of the correlations computed between 300 stations pairs.

To obtain Green’s functions, one-hour segments of the data set were cross-correlate and stacked for the entire recording period to improve signal-to-noise ratio. Figure 8 shows a record section for the correlations computed between 300 stations pairs. The presence of dispersive Rayleigh wave trains is visible in both causal and acausal lags. We can perceive the move-out of the wave trains as a function of distance, having an average velocity of ~2.0 km s-1. Final Green functions for each pair were obtained by summing both causal and acausal parts of the crosscorrelograms

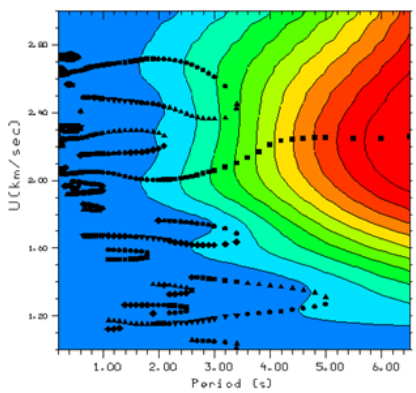

Figure 9 – Energy diagram computed for DOC11-DOC20 station pair. Dispersion can be measure by picking the maximum for each period (frequency)

Group velocity dispersion on the Green’s functions was measured using the time-frequency analysis developed by Herrmann and Ammon (2002). Figure 9 presents, as example, an energy diagram for the pair DOC11-DOC20, where the fundamental-mode group velocity can be identified.

Group velocity curves for selected paths have then been inverted to obtain a S-wave velocity profile as a function of depth. We used a linearized method (Herrmann and Ammon, 2002) in the 1D inversion with an initial velocity model consisting of horizontally homogeneous layers. Throughout the inversion, the thickness and Vp/Vs for each layer remained fixed and the density was estimated from the P-velocity using Birch law. By iteratively perturbing the initial model, the final model is obtained when a good fit to the dispersion curve is achieved. In the absence of a starting model for area, we used a 1D S-velocity model from the Azores. A smoothness constrain was included to prevent large velocity variations between adjacent layers.

We present on Figure 10 the final S-velocity models for two station-to-station paths, superimposed on the starting model. The fit between the observed and the computed group velocity is also shown.

The final 1D velocity model for both paths display a large discontinuity at a depth of ~2 km. For the path DOC08-DOC11, we observe the presence of a high velocity layer at depths between 0.8 and 1.4 km. This same layer seems to be present on path DOC11-DOC20, but with lower Vs.

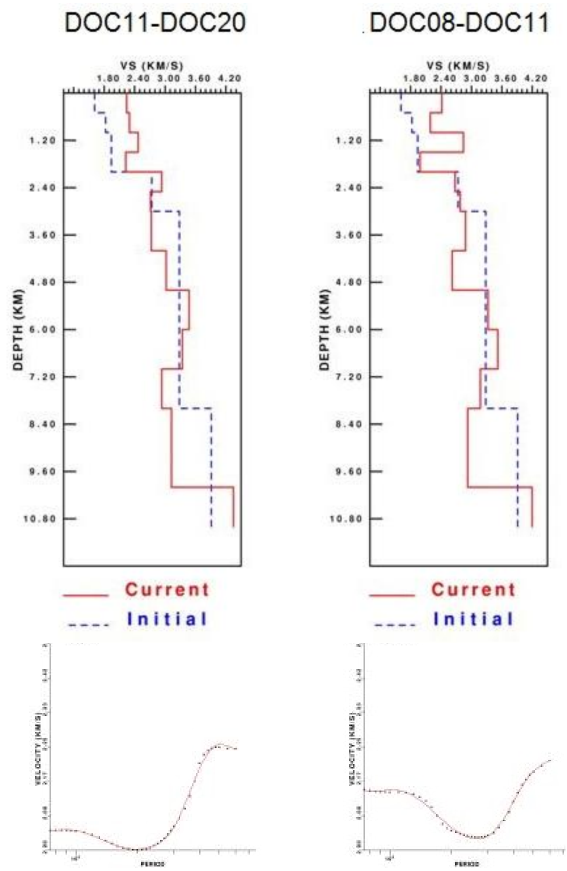

Figure 10 – TOP - S-wave velocity profiles paths between DOC11-DOC20 (29 km) and DOC08-DOC11 (28 km); Bottom – Observed dispersion curve (dots) and dispersion curve computed from final model (line)

Taking into account the azimuthal coverage of the area (Figure 11), a 2D group velocity inversion to obtain maps with velocity lateral variations as a function of the period is underway. This will enable to produce 1D S-wave velocity profiles in each grid node, as a function of depth, towards a quasi-3D model. Perturbations on S-wave velocities will be interpreted taking into account the geology and other geophysical results to produce an image of the subsurface structure of Madeira Island.

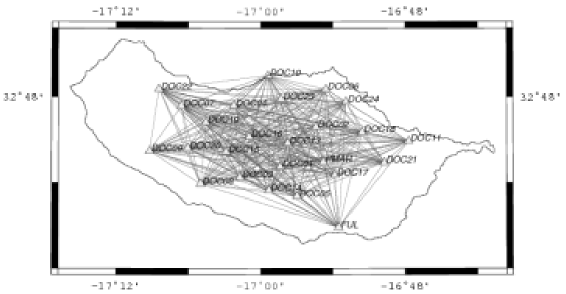

Figure 11 – Possible ray-paths between all stations

6. CONCLUSIONS

The methodologies described above are being applied to the Madeira Island in order to locate a magma body that might be considered as a geothermal system heat source. Some of the depths indicated by geobaromethric calculations are partially within the limits of the application of short-period Rayleigh wave dispersion measurements to the Madeira Island (up to 10 km). The combination of these two methods will be essential to geographically constrain the location of the heat source.

Both methods, together with petrological, lithological, geochronological, rock physics, structural, tectonic, thermometric and magnetometric ones, will contribute to assess and characterize the geothermal potential of the Island.

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Acknowledgements

The authors which to thank EEM – Madeira Electricity Enterprise for all their financial and logistic support during field work at Madeira island.