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A Review of the Mt. Amiata Geothermal System (Italy)

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ABSTRACT

Exploration of the Mt. Amiata area started at the beginning of the 50's, and the first overview of the geothermal system was carried out on 1970. After more than 30 years, all the data acquired during the exploration and exploitation activities allowed a significant update of this system. Data from more than 100 drillings (500-4000 m depth), as well as from many geophysical, geological, structural, and hydrogeological studies have been utilized for defining the geothermal model. Two distinct geothermal water dominated reservoirs are present. The first is hosted in Mesozoic, carbonatic formations at depths ranging between 500-1000 m, the second, and deeper one in the Paleozoic, metamorphic basement at depths of 2500-4000 m. These two reservoirs are characterized by different temperatures (150-220°C the shallower one, 300-350°C the deeper one) and, although they are in hydrostatic equilibrium, they are separated by a low permeable layer as evidenced by the temperature profiles of deep wells. These characteristics are common for both the Bagnore and the Piancastagnaio fields presently under exploitation.

An important phreatic aquifer is hosted in the Mt. Amiata volcanic complex that overlies the cap rock of the system, and is separated from the shallow carbonatic geothermal reservoir by a few hundred meters of impermeable shaly formations in flysch facies. Hydrological, structural, and production data demonstrate that no interferences occur between the phreatic aquifer and the geothermal reservoir. Furthermore, the numerical modeling of the whole geothermal system demonstrates that a hypothetic water recharge from the phreatic aquifer to the shallow geothermal reservoir can be at most 0.01 m^3/s, in order to preserve the currently observed thermal anomaly.

1. INTRODUCTION

The Bagnore and Piancastagnaio geothermal fields of the Monte Amiata volcanic area (Figure 1) have been under exploitation for about half a century. Numerous geologic studies have been performed in this area due to its mining and geothermal peculiarity. Recently, environmental interest arose for the presumed interactions between the important shallow aquifer hosted within volcanites and the geothermal reservoir.

In the '60s, stable isotopic determinations demonstrated the prevailing meteoric origin of geothermal fluids (Craig, 1963 and Nuti, 1991 and references therein), proving the renewability of this kind of resource. As a consequence, all the conceptual geothermal models take into account a meteoric recharge. The geothermal monograph on Amiata (Calamai et al., 1970) assumes the occurrence of a recharge area not far from the production zone. In this approach, the outcrops of reservoir formations, or other rocks hydraulically connected with them, should be not too far and of suitable extent to justify production data. The volcanic necks, although not well defined on site, were indicated as likely preferential pathways for local and partial meteoric recharge.

In this paper the main results of recent studies and of the production history of the Amiata fields are summarized in order to revise and update this model.

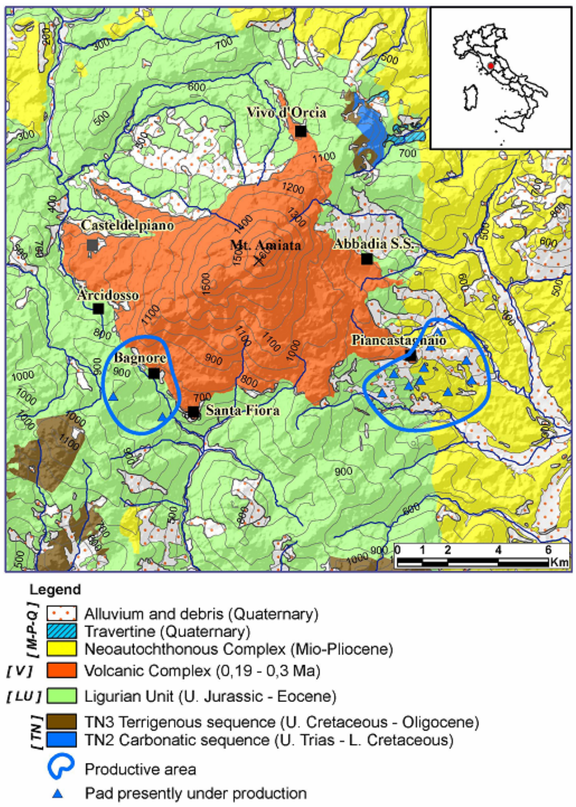

Figure 1: Geological Map of the Amiata area.

2. GEOLOGICAL AND HYDROGEOLOGICAL SETTING.

The Amiata stratigraphic sequence (Figure 2) has been reconstructed on the base of the well data and geological studies (Batini et al., 2003). From top to bottom it is:

- Mount Amiata Volcanic Complex [V] consisting of trachi-dacitic to olivin-latitic lavas, 0.19 – 0.3 Ma old (Ferrara and Tonarini, 1985, Ferrari et al. 1996);
- Neoautochthonous and Quaternary Complex (Upper Miocene – Pliocene) [M-P-Q] represented by continental and marine sediments;

- Ligurian Unit (Upper Jurassic – Eocene) [LU] represented by remnants of the Jurassic oceanic basement and its pelagic sedimentary cover;
- Tuscan Nappe Unit (Upper Trias – Oligocene) characterized from top to bottom by terrigenous [TN3], carbonate [TN2] and evaporitic [TN1] sequences;
- Tuscan Metamorphic Complex (Upper Palaeozoic – Middle Trias), only encountered by deep boreholes and referred to the Monticiano-Roccastrada Unit [MRU] (Bertini et al., 1991), comprises:
    - [MRU3] Triassic Verrucano Group, made of continental metapelites, metasandstones and metaconglomerates;
    - [MRU2] Palaeozoic Group, made of graphitic phyllites and metasandstones of probable Carboniferous age, Devonian (?) hematite-rich and chlorite phyllites, metasandstones with dolostone layers, and Late Permian fusulinid-bearing crystalline limestones and dolostones with intercalations of graphitic phyllite (Elter and Pandeli, 1991);
    - [MRU1] Micaschist Group and the Gneiss Complex [GC]. Their occurrence at depth has been indirectly documented by xenoliths in the Quaternary lavas (Van Bergen, 1983);
- Magmatic Rocks (Quaternary) [MR] assumed to be located beyond 6000 m depth on the basis of geophysical interpretations, and of granite lodes, crossed by a few deep wells.

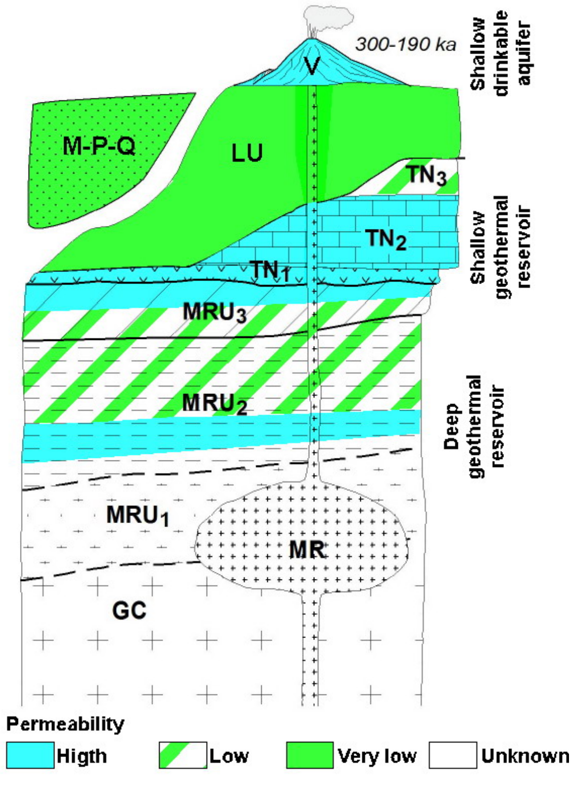

Figure 2: Geologic and hydrogeologic sketch (from Batini et al., 2003, modified).

In the same Figure 2 the hydraulic properties of the geological formations are evidenced by colours. The main water bodies are, from top to bottom, a phreatic aquifer hosted in the Volcanic Complex, and a water dominated geothermal system that comprises shallow and deep reservoirs, hosted in the Tuscan Nappe and in the Metamorphic Complex respectively.

3. THE PHREATIC AQUIFER

The Mt. Amiata Volcanic Complex has a maximum elevation of 1738 m a.s.l. and spreads over a total area of 80 km^2. The mean annual precipitation is about 1200 mm and the average temperature is 10°C. The snow persists for about 100 days/y (Barazzuoli et al. 2004).

Volcanic rocks are highly permeable because of fractures and host an important phreatic potable aquifer that represents the main water resource for a wide area in southern Tuscany. The aquifer geometry is delimited at the bottom by the impermeable formations of the Ligurian Unit (Figure 3). The base of volcanites has been mapped out upon available data (Dini et al., 2010).

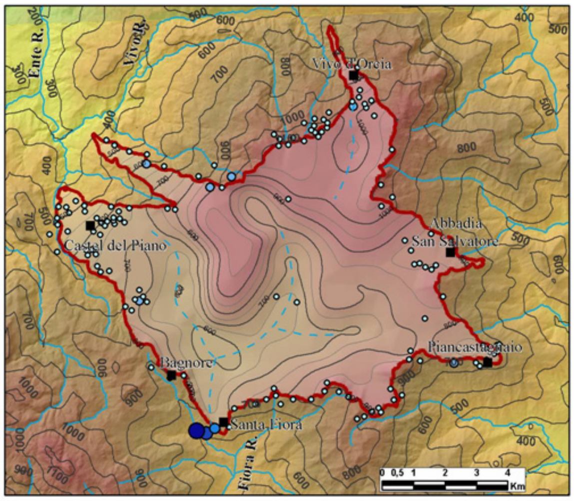

Figure 3: Base of the volcanic products (inside the red line) external to the red line DEM of the topographic surface; blue points represent springs (from Dini et al., 2010, mod.).

The phreatic aquifer outflow is represented by the over 150 springs occurring all around the boundary between volcanites and the Ligurian Unit, at elevations ranging from about 1000 m on the northern slope, to 600-800 m on the southern one. The total spring yield is of about 2 m^3/s.

Several hydrologic balances have been carried out for this aquifer. The estimated inflow varies between $69 \cdot 10^6$ m^3/y (Bazzurro et al., 1986) and $53 \cdot 10^6$ m^3/y (Barazzuoli et al., 2004), while outflow ranges from $66 \cdot 10^6$ m^3/y (Bazzurro et al., 1986) to $56 \cdot 10^6$ m^3/y (Barazzuoli et al., 2004). Discrepancies (about ±5%) fall in the approximation for the applied methods.

Other Authors (Celico et al. 1988) calculated an outflow surplus ($59 \cdot 10^6$ m^3/y) with respect to inflow ($51 \cdot 10^6$ m^3/y). The resulting difference (about 15%) is deemed to be not too high, taking into account the uncertainties of the data, especially as for evapotranspiration. In this hydrologic balance, a deficit or a surplus in the order of a few million m^3/y falls in the range of the standard uncertainty.

As for the presumed water shortage in the Mt. Amiata phreatic aquifer, it is worth noting that in the period 1940-1987 the decreasing trend is consistent with the decline in local precipitations (Barazzuoli et al., 1994, 1995). The effect of this decline is probably emphasized by the aquifer overexploitation and by water losses due to an aged water pipe network (Papalini, 1989).

4. THE GEOTHERMAL SYSTEM

The two water dominated fields (Bagnore and Piancastagnaio) of the Mt. Amiata geothermal system (see Figure 1) correspond to Tuscan Nappe structural highs that host the first and shallow geothermal reservoir at depth of 500-1000 m. Due to their particular structural and hydrological conditions, these reservoirs present gaseous caps. In both the fields, a second and deeper geothermal reservoir was discovered at a depth greater than 2000 m in the metamorphic basement. The shallow and deep reservoirs are in piezometric equilibrium as shown by the hydrostatic pressure distribution (Figure 4).

The shallow reservoir temperatures range between 150°C (Bagnore) and 230°C (Piancastagnaio) while the deep reservoir, temperatures are more uniform, usually greater than 300°C, (Bertini et al. 1995).

The shallow and the deep reservoirs are separated by 500-1000 m of low permeable metamorphic rocks, as proved by the typical thermal profile of Figure 5. This profile clearly shows the occurrence of an interval characterized by conductive regime that separates the two convective zones which are the characteristic of a geothermal reservoir. The separation between these reservoirs is hydraulically significantly effective since no pressure interference has been observed in the shallow aquifer during the exploitation of the deep one.

The Bagnore and Piancastagnaio shallow reservoirs are located in two dome shaped structures ("traps") that have a thickness of 300-500 m and an extension of 5-10 km^2.

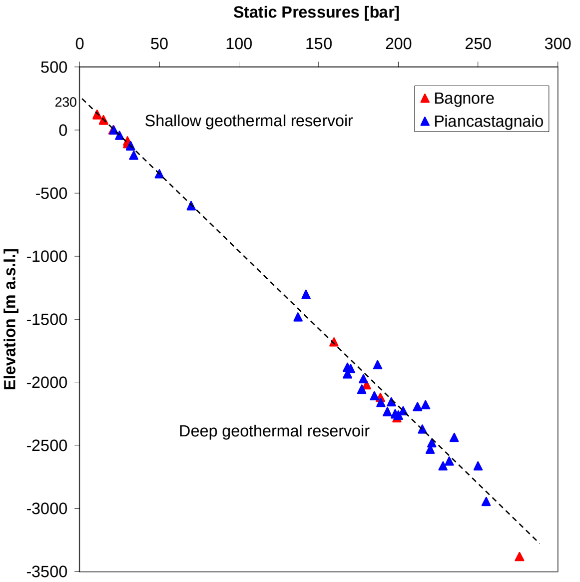

Figure 4: Static pressures vs. elevation in the Mt. Amiata geothermal system.

These traps (Figure 6) allowed steam and gas accumulation; their pressure is a function of the difference between the piezometric level and the elevation of the liquid-gas interface. At geological times, this interface was established by the spill-point elevation.

A gas generation at depth and the occurrence of a permeable layer below an impermeable cover are the necessary conditions for the gas traps to exist. The gaseous phase saturates the geothermal aquifer and builds up in the upper part of the structure, pushing down the interface. The gaseous phase pressure is determined by the hydrostatic pressure value existing at the interface elevation. Therefore, the maximum pressure that can build up in the trap corresponds to the hydraulic head at the most elevated saddle (spill point). In case of additional gas inflow from depth, the "spill point" acts as a valve which allows an outward migration of the gaseous phase, keeping the gas cap pressure constant

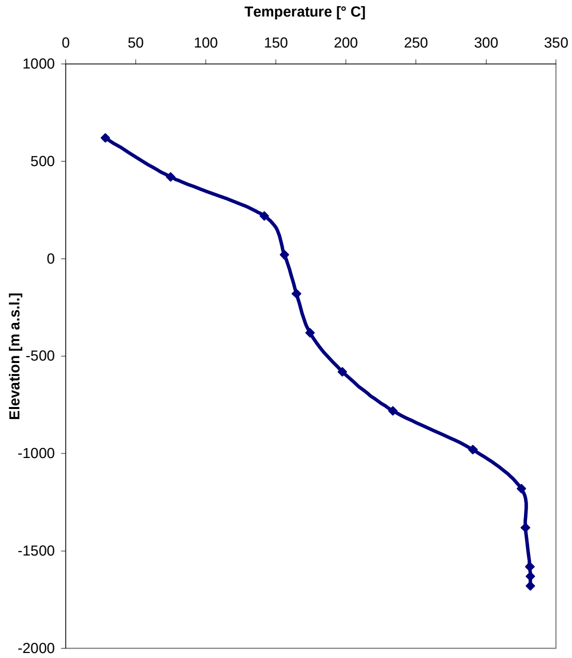

Figure 5: Typical temperature vs. elevation log in a geothermal well of the Mt. Amiata system.

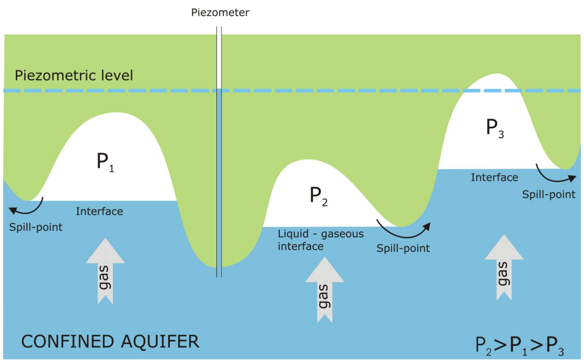

Figure 6: Theoretical sketch of "gas traps".

The Amiata geothermal shallow aquifers share a common piezometric level at +230 m (see Figure 4) that has been constant all over the exploitation history. In 1958, at the beginning of exploitation, the pressure in the two gas caps was 23 bar at Bagnore and 40 bar at Piancastagnaio. This difference was due to the different spill-point elevations that are zero m.a.s.l. in Bagnore, and at -170 m in Piancastagnaio (Figure 7A).

The exploitation caused the gas cap pressure to decrease and the interface rise; the peripheral wells were watered out while the shallower wells continued producing steam and gas. Figure 7B shows the typical situation under exploitation conditions, relevant to the '70-'90s, when steam pressures were 4 bar at Bagnore and 22 bar at

Piancastagnaio. The interfaces had risen to +190 m and to +10 m respectively.

The shallow reservoir exploitation was stopped at Bagnore in 1994 and was significantly curtailed in Piancastagnaio in the ’80s. As a consequence, a progressive increase in pressure and a water table fall has been ongoing.

Figure 7C represents the 2008 situation, with pressure values of 12 bar at Bagnore and 24 bar at Piancastagnaio, with the interfaces at +110 m and at -10 m respectively.

It is noteworthy to stress that, regardless of the pressure changes recorded in the gas caps, the piezometric level of the water table has remained unchanged from 1958 through the present.

The current geothermal exploitation of deep reservoirs has been inducing a pressure draw-down at depth, but it doesn’t affect the shallow aquifer thus demonstrating their effective separation. This can be stated on the basis of the pressure measurements in the shallow reservoir wells that show no change after the onset of the deep reservoir exploitation.

5. CAN THE PHREATIC AQUIFER INTERACT WITH THE GEOTHERMAL SYSTEM?

Recently, environmental concerns have been raised about the possible interactions between the potable aquifer and the geothermal system in spite of the presence of a well established impermeable (clayey and shaly) cap rock between these two water bearing bodies.

5.1 Physical and Hydrogeological Considerations

The evaluation of the possible interactions involves exclusively the shallow geothermal reservoir because of its geological positions and its separation with respect to the deep one. The possible interaction concerns just the two gas caps as they were the only exploited structures of the shallow geothermal reservoirs.

This interaction can be easily ruled out considering the following evidences:

- gas caps are very small with respect to the volcanic complex and located at its margins; furthermore, if there would have been even a small hydraulic connection with the environment, gas caps would have not formed over geological times and kept constant at industrial times;
- a loss of potable water into the geothermal shallow reservoir is highly improbable due to the presence of a cap rock;
- the maximum potable water inflow to the geothermal aquifer is probably null and however much smaller than 0.005 m^3/s; this is the result of a conservative heat flow balance that has been made on the geothermal shallow aquifer. This procedure consists in equating the net conductive heat flow, i.e. the difference between the incoming and the exiting heat flow (< 3.2 MW over a surface of 10 km^2) to the convective heat subtracted from the aquifer by the possible cold water inflow. If the water inflow would be larger than the above said figure, the geothermal reservoir would have been cooled down in geological times and would not exist anymore;
- the above mentioned natural inflow could not be increased by the geothermal exploitation since the geothermal shallow aquifer did not change its pressure over the exploitation history both in Piancastagnaio and in Bagnore;
- the potable aquifer has a piezometric level which is at least 400 m higher than the geothermal aquifer. This means that no pollution of the potable water by the geothermal waters is possible.

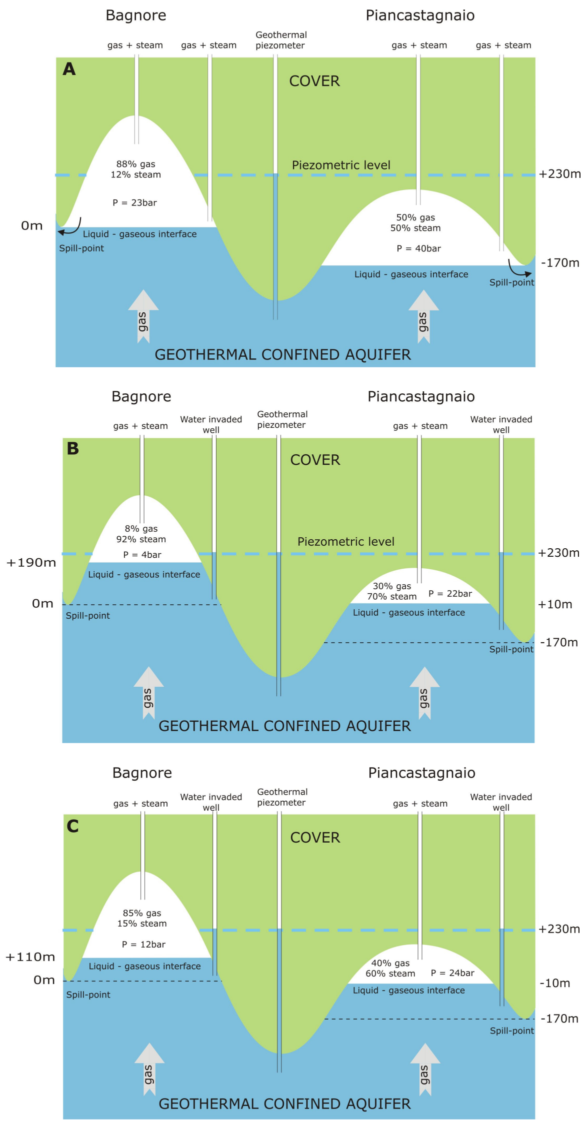

Figure 7: Status of the shallow geothermal reservoirs: A) prior to exploitation; B) after some 40 years of exploitation; C) current status (consequent to the stop of production at Bagnore and to its reduction at Piancastagnaio).

5.2 Numerical Modelling

Possible interactions have been evaluated also by means of a 3D numerical modelling performed with the TOUGH2 software (Pruess, 1991).

The modelled volume has a surface of more than 1100km^2 and a thicknesses ranging between the Mt. Amiata top (1738 m a.s.l.) and -4000 m a.s.l. (maximum geothermal well depth). This volume has been subdivided into thousands of cells in order to enter physical parameters (Figure 8). A constant thermal heat flow (mean value of 400 mW/m^2 and peaks of 600-700 mW/m^2 in correspondence of the geothermal fields) has been imposed at the lower model boundary. Permeability of each cell has been given in accordance with the geological features. The system has been considered completely closed.

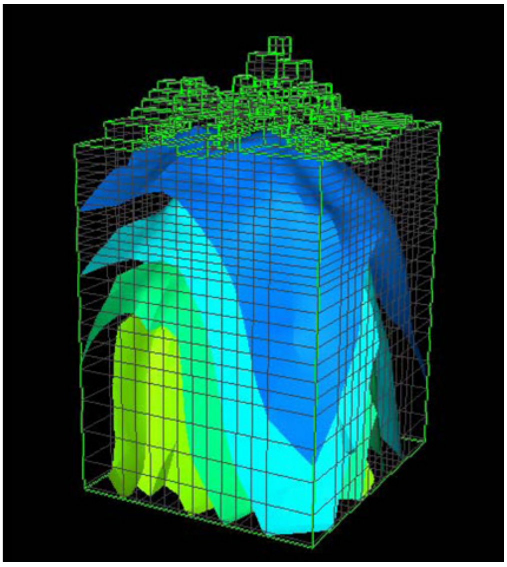

Figure 8: 3D view of the simulation grid (vertical scale exaggeration). In colours the isotherm surfaces.

The Figures 9, 10 and 11 are referred to elevation of -50m a.s.l. which can be considered representative of the shallow geothermal reservoir top.

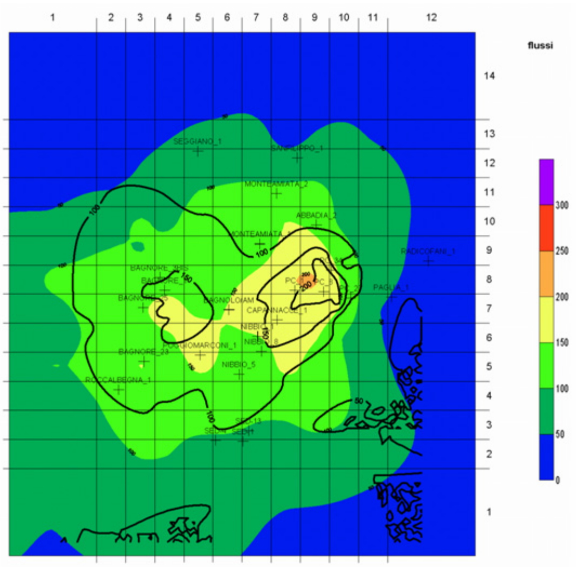

Figure 9: Natural thermal state at -50 m a.s.l: calculated temperatures (coloured scale) and measured ones (isolines).

The modelling output without any inflow from the volcanic aquifer shows a good fitting between the simulation results and the experimental data for the natural state (Figure 9).

A hypothetical hydraulic connection between the volcanic base and the geothermal reservoir has been modelled to evaluate the possible downward inflow. This connection was located along the SW-NE volcano feeding structure.

From the modelling, an inflow of only 0.03 m^3/s could completely cancel the thermal anomaly of Bagnore shallow reservoir (Figure 10).

Even an the inflow of 0.01 m^3/s shows an appreciable cooling even if a residual 150°C area still could exist (Figure 11).

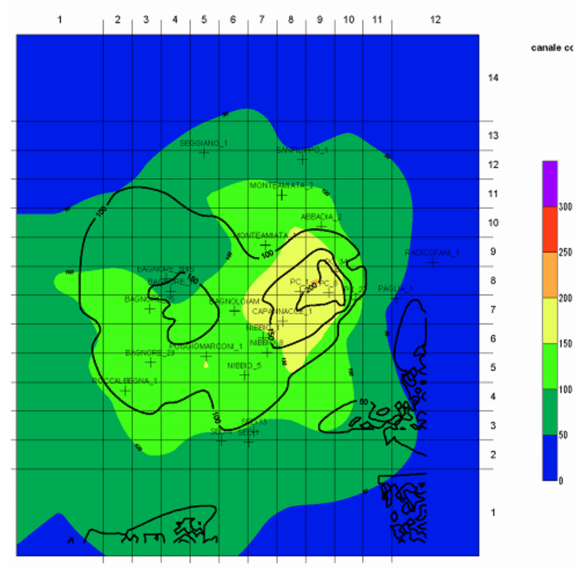

Figure 10: Thermal state at -50 m a.s.l considering an inflow of 0.03 m^3/s: calculated temperatures (coloured scale) and measured ones (isolines).

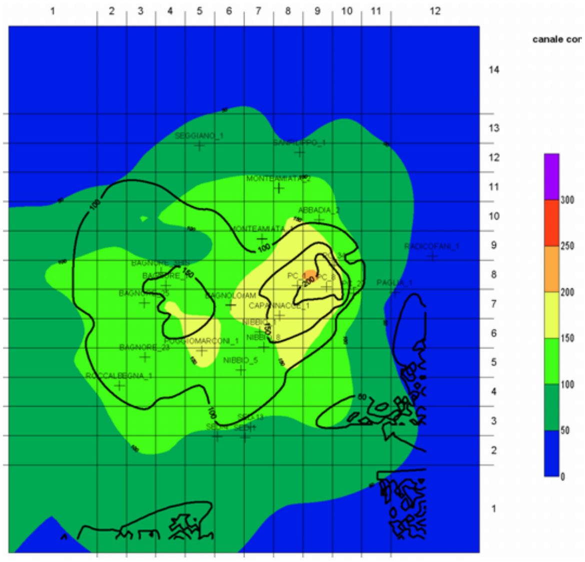

Figure 11: Thermal state at -50 m a.s.l considering an inflow of 0.01 m^3/s: calculated temperatures (coloured scale) and measured ones (isolines).

The model could not sustain a cool water inflow higher than 0.01 m^3/s in accordance with the previously discussed heat flow balance.

6. CONCLUSION

The whole exploitation history of the Mt. Amiata geothermal system has been analyzed to provide an updated review of the system itself and to highlight hypothetical relationships between phreatic and geothermal aquifers. The main results of this study are here summarized:

Deep and shallow geothermal reservoirs are separated by a low permeability layer although in hydrostatic equilibrium.

Deep geothermal reservoir exploitation doesn’t affect the shallow geothermal aquifer pressure.

Phreatic aquifer and geothermal system display different hydrostatic level.

A possible phreatic inflow, if present, should not be more than 0.01 m^3/s to allow the existence of the existing geothermal fields. This flow should be compatible with the natural state conditions and cannot be affected by the geothermal exploitation since no pressure change has been ascertained in the shallow geothermal aquifer.

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